Deep structure of lithospheric fault zones

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We calculate the cumulative width $w$ of ductile shear zones accommodating plate motion in continental lithosphere, based on the assumptions that (1) the flow stress is controlled by the yield strength of intact rock at any given depth; (2) the yield strength profile through the crust can be constrained from observations in exhumed shear zones; and (3) strain localization is primarily caused by grain-size reduction leading to a switch to grain-size-sensitive creep. We use a mid-crustal stress-temperature profile measured in the Whipple Mountains, California, and calculate stress profiles at depth from published flow laws for feldspar and olivine. We conclude that $w$ for a plate boundary shear zone accommodating 50 mm/yr displacement (comparable to the San Andreas Transform) can reach 180 km in the quartz-rich mid-crust, depending on water fugacity and thermal gradient. It narrows to a few meters in feldspathic lower crust and in the uppermost mantle, and then widens rapidly with depth in the lower lithosphere. We explore the effects of differences in crustal thickness and composition, thermal gradient, and water activity. Citation: Platt, J. P., and W. M. Behr (2011), Deep structure of lithospheric fault zones, Geophys. Res. Lett., 38, L24308, doi:10.1029/2011GL049719.

1. Introduction

[2] There is no consensus at present on the width and structure of plate-boundary fault zones below the seismogenic layer. Estimates for fault-zone widths in the lower crust range from <10 km [e.g., Henstock et al., 1997], to several tens of km [e.g., Vaucoux and Tomassi, 2003; Wilson et al., 2004], and in the lithospheric mantle estimates vary from <100 km [e.g., Hervet et al., 1999; Titus et al., 2007] to several hundreds of km [Baldock and Stern, 2005]. Fault-zone width at any depth should be a function of the constitutive relationship between stress and strain-rate in the shear zone, and therefore should be calculable [Platt and Behr, 2011b]. The difficulties with doing this stem from uncertainties both in shear-zone rheology and the level of deviatoric stress in the lithosphere. We propose three concepts that can help resolve these problems, and allow calculation of fault zone width as a function of stress and temperature. These concepts, discussed in more detail below, are as follows.

[3] 1. The deviatoric stress in a plate-boundary shear zone at any given depth approximates the yield stress of the surrounding rock at that depth [Platt and Behr, 2011b].

[4] 2. Following from (1), stress-temperature data from exhumed ductile shear zones provide proxies for strength profiles through the lithosphere [Behr and Platt, 2011].

[5] 3. Strain localization to form ductile shear zones is primarily a result of grain-size reduction causing a switch to grain-size-sensitive creep. This allows the assignment of a rheology to the shear zone material [Platt and Behr, 2011a]. Using these concepts, we calculate the minimum cumulative width of a plate-boundary ductile shear zone as a function of relative plate velocity, rock rheology, temperature, and water fugacity, and we explore the effects of these variables.

2. Ductile Shear Zones Are Constant Stress Experiments

[7] Ductile shear zones reflect strain localization, which result from microstructural changes produced by deformation. These changes require the surrounding rock to deform, so the ambient stress must reach and remain at the yield stress $\sigma_y$ of intact rock [Platt and Behr, 2011b]. For an imposed plate velocity $V$, the cumulative width $w$ of the shear zones accommodating that motion will stabilize at a value given by $w = V/\dot{\varepsilon}$. The strain rate $\dot{\varepsilon}$ in the shear zone is controlled by the rheology of the shear zone material, so that $\dot{\varepsilon} = A \sigma_y^n$, where $A$ and $n$ are material parameters. The shear zones act as a self-organizing system: $w$ cannot exceed the value given above, or the strain-rate will drop, and hence the stress will too, preventing any further widening at the expense of the surrounding rock. If $w$ is less than this value, the stress will exceed $\sigma_y$, and the surrounding rock will deform, widening the shear zone. The velocity-boundary condition imposed by plate motion is therefore converted to a stress-boundary condition within the shear zone, buffered by the yield stress of the surrounding rock. This concept leads to the conclusion that shear zone width is related to the material properties of the shear zone by:

$$w = V/A \sigma_y^n$$

(1)

Shear zones that are reactivated, or that experience a drop in the imposed relative plate velocity, may not obey this constraint. In contractional and extensional shear zones rocks change depth and hence temperature, so that their rheology changes, but the stress in the shear zone should still approximate the strength of the surrounding rock at any given depth [Platt and Behr, 2011b].

3. Measuring Stress Profiles Through the Crust

[8] Exhuming shear zones preserve a record of the stress and temperature profile through the deforming crust. We constructed such a profile for an exhumed normal-sense ductile shear zone underlying the Whipple detachment fault in SE California, using the dynamically recrystallized grain-size piezometer for quartz [Stipp and Tullis, 2003, and the Ti-in-quartz thermobarometer [Thomas et al., 2010]. By
modelling the cooling history during exhumation, we converted this to a stress-depth profile \[ \text{Behr and Platt, 2011} \]. The Whipple Mountains accommodated Basin and Range extension during early Miocene time at a rate of \( /C245 \text{ mm/yr} \) \[ \text{Stockli, 2005} \]. The stress profile from the Whipple Mountains therefore provides a proxy for a strength profile through the ductile crust of the Cordillera to a depth of 25 km. Because the yield strength of intact rock is independent of the strain-rate in the shear zone, this profile can be applied to shear zones such as the San Andreas Transform system that move at different rates from the Whipple Mountains shear zone.

4. Controls on the Rheology of Ductile Shear Zones

The most potent cause of weakening in shear zones is likely to be dynamic recrystallization, resulting in grain-size reduction and a switch to grain-size-sensitive creep \[ \text{Platt and Behr, 2011a} \]. The change in deformation mechanism occurs because at small grain-sizes the grain-size sensitive creep mechanism can accommodate a higher strain-rate at any given stress. At low temperature and high stress, dynamic recrystallization occurs primarily by grain-boundary migration driven by lattice strain energy \( /\sigma \text{GBM} \). \( /\sigma \text{GBM} \) may be the dominant mechanism of recovery during dislocation creep under these conditions \[ \text{Tullis and Yund, 1985; Fliervoet and White, 1995} \]. This leads to a form of grain-size-sensitive dislocation creep (DRX creep) with a flow law of the form:

\[
\dot{\varepsilon} = \frac{A_d D_{\text{gb}} \sigma^3}{d_r},
\]

where \( A_d \) incorporates lattice-dependent material parameters, \( D_{\text{gb}} \) is the diffusion coefficient for grain-boundary migration, and \( d_r \) is grainsize \[ \text{Platt and Behr, 2011a} \]. The grain size in materials undergoing dynamic recrystallization depends on the flow stress raised to the power \( -p \), where \( p \) has a value in the range 1.26 for quartz \[ \text{Stipp and Tullis, 2003} \] to 1.33 for olivine \[ \text{Van der Wal et al., 1993} \]. The effective stress exponent in this flow law is therefore a little over 4, which corresponds well to experimental and observational constraints \[ \text{Hirth et al., 2001; Hirth and Kohlstedt, 2003} \]. We constrained the material parameters in this flow law for quartz using published experimental data and our measurements of differential stress, strain-rate, and shear-zone width in the Whipple Mountains shear zone \[ \text{Platt and Behr, 2011b} \] (see also Table S1 in the auxiliary material).

5. Width of Plate-Boundary Shear Zones

We apply the relationship between shear zone width and rock mechanics expressed in equation (1) to calculate the cumulative width \( w \) of a transform plate boundary with a slip rate of 50 mm/yr (comparable to the San Andreas Transform) through the lithosphere below the brittle-ductile

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**Figure 1.** (left) Differential stress profile and (right) cumulative shear zone width \( w \) through continental lithosphere with a composition and thermal gradient comparable to southern California, cut by a plate boundary shear zone with a displacement rate of 50 mm/yr. Curves shown are calculated for wet (blue) and dry (red) rheologies. The stress profile was calculated using Byerlee’s Law for strike-slip faulting in the seismogenic layer, stress-temperature data from paleopiezometry in the Whipple Mountains for the mid-crust after \text{Behr and Platt, 2011}, and dislocation creep laws for feldspar \[ \text{Rybacki et al., 2006} \] and olivine \[ \text{Hirth and Kohlstedt, 2003} \]. \( w \) is calculated for quartz-dominated middle crust using the DRX creep law (P&B DRX creep) and climb-assisted dislocation creep laws (P&B climb creep) from \text{Platt and Behr, 2011a}, and the creep law from \text{Hirth et al., 2001} (green). For feldspar dominated lower crust we used the diffusion creep laws for feldspar \[ \text{Rybacki et al., 2006} \] and olivine \[ \text{Hirth and Kohlstedt, 2003} \] (see text for further explanation). Data are presented in tabular form in Table S2 in the auxiliary material.
transition, and we explore a variety of possible shear zone rheologies, lithospheric structures and geotherms. To do this, we need to estimate the yield stress of the lithosphere as a function of temperature, depth, composition, and water content. For quartz-rich middle crust with a composition and thermal gradient comparable to that in southern California, we use the stress-temperature profile from the Whipple Mountains determined by Behr and Platt [2011]. Naturally constrained stress profiles do not exist for the lower crust or upper mantle, however, and yield strengths for rocks under geological conditions are not well known. To estimate stress profiles through these parts of the lithosphere we assume that in the absence of strain localization, rocks at these depths deform by dislocation creep at $\dot{\varepsilon} \approx 10^{-15} \text{ sec}^{-1}$. This rate corresponds to non-localized deformation (equivalent to distributing Pacific / North America plate motion over the entire width of the North American Cordillera). We then use published flow laws for feldspar and olivine to calculate stress profiles for various crustal thicknesses, compositions, thermal gradients and water contents (shown on the left side of the plots in Figures 1 and 2). We assume that strain localization takes place at constant stress for any given depth, following the stress profile, as a result of a switch to grain-size-sensitive creep caused by dynamic recrystallization. For feldspar and olivine, the deformation switches from dislocation to diffusion creep, and for quartz the deformation switches to the DRX creep law discussed above.

[12] In all models we arbitrarily assume the brittle part of the transform zone in the upper crust has a cumulative width of 1 km – this does not affect the calculations. We do not allow for stress transfer between different levels in the lithosphere [e.g., Roy and Royden, 2000], transient effects associated with the seismic cycle [e.g., Freed et al., 2010], or the effects of stress concentrations as modeled by Regenauer-Lieb et al. [2006], for example. These effects are real and important, but they are likely to be small relative to the effects of uncertainties in the rheology, which is the primary focus of this paper. We also accept that real rock materials, such as peridotite, gabbro, and granite, are polyphase systems, and that these are likely to differ significantly in their rheology from pure quartz, feldspar, and olivine [Dell'Angelo and Tullis, 1996; Dimanov and Dresen, 2005].

[13] To outline the effect of variations in the parameters, we show in Figures 1 and 2 the following combinations. In Figure 1 we take a lithospheric column and geotherm that is appropriate to southern California, where the San Andreas Transform cuts through continental crust similar in age, composition, and thickness to that exposed in the Whipple Mountains. Rather than assume the water content, we assume our stress profile is correct, and explore the consequences of using wet (blue in Figure 1) and dry (red) rheologies for quartz, feldspar and olivine. Wet rheologies assume water activity equal to unity, and the water weakening effect is linearly dependent on the water fugacity [Gleason and Tullis, 1995]. Dry rheologies assume a constant and low water fugacity. The effective water fugacity during deformation is difficult to constrain, as water present in the system may not in fact be able to enter the crystal lattice [Post and Tullis, 1998]. Neither of the two alternatives described above is likely to be correct in general: while water fugacity is likely to increase with depth due to the effects of temperature and pressure [Paterson, 1986], this will be counteracted by hydration reactions in crystalline rocks, which consume water [Yardley and Valley, 1997]. The wet and dry rheologies we assume therefore bracket the likely range of natural water fugacities.

[14] To explore the effects of thermal gradient and tectonic setting, we take a lithospheric column and geotherm that is appropriate to southern California, where the San Andreas Transform cuts through continental crust similar in age, composition, and thickness to that exposed in the Whipple Mountains. Rather than assume the water content, we assume our stress profile is correct, and explore the consequences of using wet (blue in Figure 1) and dry (red) rheologies for quartz, feldspar and olivine. Wet rheologies assume water activity equal to unity, and the water weakening effect is linearly dependent on the water fugacity [Gleason and Tullis, 1995]. Dry rheologies assume a constant and low water fugacity. The effective water fugacity during deformation is difficult to constrain, as water present in the system may not in fact be able to enter the crystal lattice [Post and Tullis, 1998]. Neither of the two alternatives described above is likely to be correct in general: while water fugacity is likely to increase with depth due to the effects of temperature and pressure [Paterson, 1986], this will be counteracted by hydration reactions in crystalline rocks, which consume water [Yardley and Valley, 1997]. The wet and dry rheologies we assume therefore bracket the likely range of natural water fugacities.

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competition between the effects of increasing temperature (which increases strain-rate at any level of stress) and decreasing stress following our stress profile. The results are extremely sensitive to the precise level of stress, and to the flow law used to calculate strain-rates. For comparison we show \( w \) calculated on the same assumptions, but using the quartz flow-law estimated from experimental and natural data by Hirth et al. [2001]: this predicts \( w \) increasing with depth to nearly 200 km. The strain rates predicted by this flow law appear to be more than an order of magnitude less than those we observe at corresponding stresses and temperatures in the Whipple Mountains. Our predictions for \( w \) are therefore likely to be minima. In the feldspathic lower crust we predict an extremely narrow shear zone, with \( w \leq 200 \) m. This reflects the effect of the switch to diffusion creep in fine-grained feldspar [Rybacki et al., 2006], combined with the high stresses needed to develop the shear zone. Note that the strain-rates in such narrow ductile shear zones will be very high, and at high stress, shear heating will be significant, and may lead to melting and seismogenic faulting [John et al., 2009]. In the mantle, \( w \) is 850 m at the Moho, due to the activity of grain-size sensitive creep [Hirth and Kohlstedt, 2003], but it widens downwards, reaching over 300 km at 42 km depth (850°C). Below 45 km depth (900°C) the dominant flow law changes to dislocation creep, and our model predicts non-localized strain. In practice, processes that we have not taken into account, such as ingress of water, development of crystallographic preferred orientation, or the effect of lateral contrasts in temperature or composition, may cause some degree of localization.

[16] In the “California dry” model (red curve in Figure 1) \( w \) increases to 156 km at the base of the quartz-rich layer. This results from the much greater strength of dry quartzite, so that using our measured stress values from the Whipple Mtns, strain-rates are lower, and hence the shear zone has to be wider to accommodate the imposed plate velocity. Calculated stresses in the felspathic lower crust are so high that we predict brittle failure in this region. In the upper mantle the shear zone is 50 m wide at the Moho, widening to over 1100 km at 54 km depth (1000°C), below which we predict no strain localization.

[17] For our “dry craton” model we calculated a stress-temperature profile through the upper crust using the quartz flow law determined by Rutter and Brodie [2004], which may be a good approximation to the behavior of unweakened quartz, and in the feldspathic lower crust and upper mantle we used published flow laws for the Whipple Mountains strain-rates are lower, and hence the shear zone has to be wider to accommodate the imposed plate velocity. Calculated stresses in the felspathic lower crust are so high that we predict brittle failure in this region. In the upper mantle the shear zone is 50 m wide at the Moho, widening to over 1100 km at 54 km depth (1000°C), below which we predict no strain localization.

[18] For a hot wet orogen we use the Whipple Mountains stress profile to 550°C (blue curve in Figure 2), but we recognize that this may overpredict the stress (and hence underpredict \( w \)). Above 550°C we use a stress profile calculated from the quartz flow law determined by Rutter and Brodie [2004]. Our calculations predict a mid-crustal shear zone widening downwards to 177 km wide at 46 km depth (800°C). At temperatures above 550°C we lack an observational stress profile, so that values for \( w \) at depths > 29 km, where the shear zone is 7 km wide, are highly uncertain. We predict a shear zone 1.8 km wide in feldspathic lower crust at 46 km depth (800°C), which widens rapidly downwards to 46 km at the Moho (60 km depth, 1000°C). Below the Moho at these temperatures the dominant flow law in olivine is dislocation creep, and hence we predict no strain localization.

6. Conclusions

[19] Our exploration of shear zone width for wet and dry rheologies in active tectonic and cratonic crustal environments covers a reasonable range of parameter space, and allows us to draw some general conclusions.

[20] 1. Cumulative widths for plate boundary ductile shear zones below the seismogenic layer may reach 180 km in quartz-rich crystalline continental crust, but depend on both water content and the yield strength of undeformed crust, which controls the ambient stress. Shear zone width results from the interplay between the effects of deformation mechanism switches, increasing temperature, and decreasing stress with depth. In strong, dry, crustatic shear zones may be <1 km width. Our predicted widths are within the range reported in geological studies of green schist facies, quartz-rich shear zones exhumed from the middle crust [e.g., West and Hubbard, 1997; Whitmeyer and Simpson, 2003; Faleiros et al., 2010], and are consistent with the general observation that strain becomes increasingly distributed at depth within the crust [Vauzech and Tomassi, 2003]. Our prediction that \( w \) may reach 180 km beneath a San Andreas type transform zone suggests that the upper crust may be largely decoupled from the underlying lower crust and mantle lithosphere. Upper crustal faults may be kinematically linked at depth, as suggested by Platt & Becker [2010], and the narrow shear zones we predict in the lower crust and upper lithospheric mantle may not have a direct spatial connection to upper crustal faults, as suggested by Roy and Royden [2000]. Note that the wide shear zones we predict in the mid-crust do not directly relate to the concept of channel flow at these levels [e.g., Beaumont et al., 2004]. We predict wide shear zones where there is a relatively low level of strain-related weakening compared to undeformed rock. By contrast, channel flow and related phenomena result from the presence of inherently weak zones within the lithosphere, such as zones of partial melt, and their thickness reflects the thickness of the thermal perturbation or weak compositional layer that controls the process.

[21] 2. In feldspar-dominated lower crust and in the uppermost mantle, stresses are high, resulting in very narrow shear zones, and in dry crust they may exceed the brittle fracture strength of rock. Even where the stresses are low enough for ductile shear zones to develop, the effect of shear heating may result in thermal runaway, melting, and seismogenic faulting [John et al., 2009]. These predictions are consistent with the occurrence of earthquakes in the lower crust, particularly of cratonic regions [Maggi et al., 2000].

[22] 3. Below the Moho, shear zones widen dramatically with depth. This results from the progressively decreased effectiveness of grain-size-sensitive creep as a weakening and strain localization mechanism with increasing temperature.
Strain cease to be localized at depths ranging from 46 km (in wet warm lithosphere) to >180 km (in cool dry lithosphere): this transition generally occurs within the thermal boundary layer, rather than at the base of the lithosphere. Several recent geophysical and geochemical studies are consistent with large-scale upper mantle flow [Ballock and Stern, 2005; Freed et al., 2007; Sol et al., 2007] beneath transform plate boundaries, supporting our prediction that deformation in the lower part of the lithosphere may be very broadly distributed in actively deforming regions. The abruptness of shear zone widening and its proximity to the Moho may affect the likelihood of large-scale decoupling between the crust and lithospheric mantle, including, for example, the development of seismically active mega-detachments at Moho depths [e.g., Davis et al., 2006; Wernicke et al., 2008], or the foudering and delamination of lithospheric mantle.

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