Influence of shear heating on microstructurally defined plate boundary shear zones

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Abstract

In plate-boundary scale ductile shear zones defined by microstructural weakening, shear heating may lead to a temperature increase over 5 m.y. of up to 80 °C just below the brittle ductile transition, up to 120 °C just below the Moho, and to thermal boundary zones tens of km wide on either side of the shear zone. Where rock strength is highest, shear zones are narrow (~1 km), and thermal gradients within the shear zone itself are low, so there is no tendency for increased localization. Heating results in thermal weakening, but this is partly offset by grain growth related to the decrease in stress. In shear zones of the order of 10 km width, shear stress, strain rate, and hence heat generation are lower, and thermal gradients are insufficient to cause additional strain localization. Temperature increases in the mid-crust are of the order of 10 °C, insufficient to cause partial melting or an increase in metamorphic grade. In the upper mantle, shear zones may be 50 km or more wide, and the temperature increase is less than 20 °C in 5 m.y., but temperature differences between center and margin may enhance the strain rate at the center by up to 18%.

1. Introduction

Shear heating during ductile deformation can produce significant temperature changes over the likely lifetime of plate-boundary deformation zones (e.g., Brun and Cobbold, 1980; Fleitout and Froidevaux, 1980; Molnar and England, 1990; Thatcher and England, 1998; Kaus and Podladchikov, 2006; Takeuchi and Fialko, 2012), particularly in the middle crust close to the brittle–ductile transition, where shear stresses of the order of 100 MPa have been documented (e.g., Kohlstedt and Weathers, 1980; Hirth et al., 2001; Behr and Platt, 2014), and in actively subducting oceanic lithosphere, where bending stresses approaching 1 GPa are predicted from rock-mechanical data (e.g., Hirth and Kohlstedt, 2003; Thielmann and Kaus, 2012). Deformational processes in crystalline rocks show strong temperature dependence, so it is reasonable to suggest that shear heating could amplify strain rates and contribute to strain localization. Several workers have suggested that shear heating could be an important mechanism in localizing deformation in plate boundary shear zones (e.g., Yuen et al., 1978; Kameyama et al., 1999; Leloup et al., 1999; Schott et al., 2000; Bercovici and Karato, 2002; Regenauer-Lieb et al., 2006; Thielmann and Kaus, 2012; Duretz et al., 2014), and hence be a factor favoring plate tectonics. It has also been suggested that under some circumstances shear heating in an otherwise ductile environment could lead to thermal runaway and the initiation of earthquakes (e.g., Kelemen and Hirth, 2007; John et al., 2009). There is little observational evidence to support significant shear heating in major fault zones in the upper crust, however (e.g., Lachenbruch and Sass, 1992), and some classic examples of metamorphism and magmatism that were attributed to shear heating at depth, such as the Red River fault zone (Leloup and Kienast, 1993), the Himalayan Main Central Thrust (England and Molnar, 1993), and the Vincent fault in southern California (Graham and England, 1976) have more recently been re-evaluated in terms of alternative heat sources or structural models (e.g., Gilley et al., 2003; Chapman et al., 2011; Kohn, 2014).

The premise of this paper is that shear zones are initiated by weakening caused by the microstructural changes that accompany deformation, and that shear heating contributes to the evolution of the rheology of the shear zone as deformation proceeds. We need to examine this premise, however, because numerical modeling studies suggest that under certain specific conditions shear heating may play a major role in initiating ductile shear zones. High stress, a constant stress boundary condition, plastic behavior (i.e., involving a yield stress), and pre-existing mechanical boundaries are all
factors that appear to favor this process. Under low stress conditions, and in linear viscous materials, conductive heat loss tends to prevent substantial temperature rises. In a detailed analysis, Kaus and Podladchikov (2006) explored the conditions under which shear heating would cause strain localization in a visco-elasticplastic material subject to a temperature perturbation. They show that under a constant velocity boundary condition, which is most appropriate for plate-boundary shear zones, localization is strongly sensitive to the scale of the initial temperature perturbation and to the background strain-rate \( \dot{\epsilon}_0 \) in the material. They provide a scaling law for localization, which for a linear viscous material can be expressed as:

\[
W_0 \geq \frac{1.4}{\dot{\epsilon}_0} \sqrt{\frac{R k T_0^2}{\mu_0 Q}}.
\]

where \( W_0 \) is the width of the initial temperature perturbation, \( R \) is the gas constant, \( k \) is thermal conductivity, \( T_0 \) is the initial temperature, \( \mu_0 \) is the initial viscosity, \( Q \) is the activation energy for viscous flow, and \( \dot{\epsilon}_0 \) is the background strain rate. This relationship predicts that for a background strain rate of \( 10^{-15} \) s\(^{-1} \) (corresponding to non-localized strain accommodating normal relative plate velocities), a shear stress of \(-1 \) GPa is required within a 10 km wide perturbation for heat generation to be able to overcome conductive heat loss. If the perturbation is only 1 km wide, the shear stress has to be \(-10 \) GPa, because of the much greater rate of conductive heat loss. The relationship between perturbation width and stress predicted by this relationship is plotted in Fig. 1, using \( k = 3.0 \) W m\(^{-1}\) K\(^{-1}\), \( Q = 5.3 \times 10^5 \) J/mole. In general, the levels of shear stress required for the initiation of strain localization by shear heating alone appear to be outside the likely range predicted by experimental rock-mechanics data within the continental lithosphere (e.g., Kohlstedt et al., 1995; Bürgmann and Dresen, 2008), or the range determined by paleopiezometric measurements on mylonites (Behr and Platt, 2014).

This conclusion appears to be contradicted by some numerical analyses, such as that of Duretz et al. (2014), who predict that shear zones triggered by shear heating can stabilize at only 3–7 km width for stresses in the range 100–1000 MPa. The apparent discrepancy arises because these authors used a background strain-rate of \( 5 \times 10^{-14} \) s\(^{-1} \) for their experiments. This is equivalent to pre-localizing the strain: at this strain-rate, a relative plate velocity of 50 mm/yr, say, can be accommodated within a zone 16 km wide. Hence, although useful, their models do not correspond to the initiation of shear zones by shear heating in the absence of other localization processes, at least under conditions prevailing on Earth.

There is now widespread agreement that the pronounced changes in grain size, mineral assemblage, and fabric commonly found in natural shear zones provide an adequate explanation for the required weakening and strain localization (White and Knipe, 1978; Rutter and Brodie, 1988; Wintsch et al., 1995; Jaroslow et al., 1996; Jin et al., 1998; Imber et al., 2001; Drury, 2005; Holyoke and Tullis, 2006; Jefferyes et al., 2006; Warren and Hirth, 2006; Platt and Behr, 2011b). Based on this consensus, Platt and Behr (2011a) proposed that the primary mechanism of strain localization within the lithosphere is microstructural damage, and that the deviatoric stress during this process is buffered by the requirement that sufficient strain must accumulate on the timescale of shear-zone initiation to produce microstructural weakening. This concept, described in more detail in the next section, provides a straightforward way of predicting the stress, strain rate, and cumulative width of lithospheric shear zones in terms of their rheology (Fig. 2). Shear heating accumulates over time, however, and will change the rheology: once a shear zone has formed, increasing temperature will induce a drop in the stress required to drive the imposed strain rate, and may also induce microstructural changes. The purpose of this paper is to investigate and quantify the
changes induced by shear heating in plate-boundary scale shear zones with realistic and self-consistent widths, deviatoric stress levels, and strain-rates, and to discuss the implications of these changes for the mechanics of plate boundaries.

2. Mechanics of shear zones initiated by microstructural weakening

The mechanical description for the formation of plate-boundary scale ductile shear zones proposed by Platt and Behr (2011a) forms the basis for the calculations in this paper, so I summarize and discuss it here for the convenience of the reader.

For a shear zone to initiate by microstructural weakening, sufficient strain has to accumulate to cause the required microstructural changes, and this has to happen on a time-scale that is consistent with the rate of change in plate boundary configurations (~1 m.y.). In the absence of strain localization, the lithosphere would have to deform at a strain-rate of $10^{-15}$ s$^{-1}$ in order to accommodate relative plate motions. This rate produces 3% strain per million years, which is probably sufficient to initiate strain-rate decreases in the surrounding rock. Rate of microstructural weakening around grain-boundaries and other natural heterogeneities on this time scale. I therefore use

$$\dot{\varepsilon} = \frac{V}{2\varepsilon_0} = \frac{V}{2f(\sigma_{\sigma - 15})},$$

where $f$ is the shear zone rheology. If we can determine a rheology (flow law) for the material in the shear zone, this provides a way of determining self-consistent values of shear zone width, stress, and strain-rate.

The concept outlined above predicts that the stress $\dot{\sigma}_{\sigma - 15}$, the tensor strain rate $\dot{\varepsilon}_0$, in the shear zone (equal to half the rate of simple shear $\dot{\gamma}_0$), the cumulative width $w$ of shear zone material, and the imposed plate velocity $V$, will be related by

$$w = \frac{V}{2\varepsilon_0} = \frac{V}{2f(\sigma_{\sigma - 15})},$$

where $f$ is the shear zone rheology. If we can determine a rheology (flow law) for the material in the shear zone, this provides a way of determining self-consistent values of shear zone width, stress, and strain-rate.

Probably the most powerful, and certainly the best quantified, mechanism of microstructural weakening is grain-size reduction induced by dynamic recrystallization, leading to a switch to grain-size sensitive creep. Flow laws for a variety of grain-size-sensitive creep mechanisms have been proposed, including solid-state grain-boundary diffusion creep (e.g., Coble creep, Poirier, 1985), grain-boundary sliding accommodated by either diffusion creep or dislocation creep (Langdon, 2006; Warren and Hirth, 2006; Hansen et al., 2012), and dislocation creep where the recovery mechanism is dynamic recrystallization (DRX creep, Platt and Behr, 2011b). To utilize these flow laws we need to specify a grain-size. It has long been accepted in the materials science community that dynamic recrystallization during dislocation creep results in a mean grain size $d$, that is inversely related to differential stress by a relationship of the form $d = K\sigma^{-p}$, where $K$ is a material constant and $p$ is $-1$. This is known as the piezometric relationship.

A number of alternative grain-size evolution laws have been proposed (e.g., De Bresser et al., 1998; Shimizu, 1998; De Bresser et al., 2001; Bercovici and Karato, 2002; Austin and Evans, 2007; Rozel et al., 2010; Austin, 2011), based on the idea that the grain size in a deforming material is controlled by an equilibrium between grain-size reduction by dynamic recrystallization and grain growth (Derby, 1992; De Bresser et al., 1998). Platt and Behr (2011b) showed that these two processes involve incompatible grain-boundary topologies, however, and cannot normally act together in the same material at the same time. Establishment of an equilibrium between the two processes is, therefore, unlikely. The equilibrium models also predict a temperature dependence for the resulting grain size that is not observed.

The related concept is the “mineral flow law” introduced by Austin and Evans (2007) and Austin (2011). This is based on the partitioning of the rate of mechanical work between shear heating and microstructural damage, governed by a parameter they call $\lambda$. The value of $\lambda$ is chosen arbitrarily, and is assumed to be constant at 0.1. The microstructural energy sinks involved are the storage of elastic strain in the form of lattice defects, and the creation of new grain-boundary surface area. The build up of dislocation density is likely to be largely complete by the time about 20% strain is reached, after which it will reach steady state or decrease. The surface energy stored in grain boundaries will build up more slowly, but both terms are extremely small compared to the total energy budget. For example, olivine deformed at a differential stress of 200 MPa recrystallizes to a grain-size of 13 microns (van der Wal et al., 1993), which at a surface energy of 10.1 m$^{-2}$ (Tasaka and Hiraga, 2013) results in a total grain-boundary energy of $3.06 \times 10^{07}$ J m$^{-2}$. The dislocation strain energy at this stress is $9.38 \times 10^{07}$ J m$^{-2}$ (see, for example, Platt and Behr, 2011b for this relationship and the parameter values). The microstructure is likely to reach steady state after a shear strain of about $5$ (Heilbronner and Tullis, 2006), equivalent to a total mechanical work of $2.5 \times 10^{07}$ J m$^{-3}$. This process may take some time: at a strain rate of $10^{-12}$ s$^{-1}$, for example, it takes 80,000 years to reach a $\lambda$ of 5. $\lambda$ will have an average value of 0.007 during that period, and will decrease as microstructural steady state is approached. The evolution of the microstructure (and hence the rheology) during the period of shear-zone initiation is an interesting problem (Herwegh et al., 2014), but is not addressed in this paper.

Although the theoretical basis for the grain-size–stress relationship remains unclear (see Platt and Behr, 2011b for discussion), it has been calibrated experimentally for a variety of metals and rock-forming minerals (Twiss, 1977), and it has been shown to be independent of strain-rate, temperature and water content within the experimental uncertainties for both quartz (Stipp and Tullis, 2003; Stipp et al., 2006) and olivine (van der Wal et al., 1993).

For the reasons discussed above we proceed in this paper to use the experimentally calibrated grain-size–stress (piezometric) relationships for quartz, feldspar, and olivine. If the differential stress is known, the calculated grain size can then be used to provide a grain-size sensitive flow law for fully recrystallized material.

The rheological assumptions underlying this description are clearly oversimplified. Real rocks are polypehase, and our understanding of the rheology of polypehase systems is still in its infancy (e.g., Tullis et al., 1991; Ji et al., 2001, 2004; Dimanov and Dresen,
3. Shear heating in zones created by microstructural weakening

The issue addressed in this paper is how shear heating will modify a plate-boundary scale ductile shear zone that has been defined by a microstructural weakening process. To answer this, I have investigated the behavior of a strike-slip shear zone accommodating 50 mm/yr of displacement (comparable to the San Andreas Transform system), which varies in width, shear stress, and strain-rate with depth, following the concept discussed in the previous section. This provides reasonable and self-consistent values of these parameters, and allows exploration of the effects of shear heating. The calculations were carried out in two stages. The first stage is to determine the variation of the initial width, shear stress, and strain-rate in the shear zone as a function of depth, temperature, and composition (rheology). The second stage is to calculate the effect of shear heating as a function of time within the shear zone at different depths.

Stage 1: Definition of shear zone width by microstructural weakening.

This stage of the calculation is based on the following assumptions, discussed in more detail above. The step-by-step procedure for the calculations is set out in the Appendix, and the results shown in Fig. 2.

i. The rheology of the crust to a depth of 27 km is assumed to be governed by quartz, the crust below 27 km by feldspar, and the mantle (below 35 km depth) by olivine. The stress $\sigma_{15}$ required for shear-zone initiation is calculated for each depth using dislocation creep flow laws for the relevant mineral and a strain rate of $10^{-15}$ s$^{-1}$. These flow laws have the general form:

$$\dot{\varepsilon} = A_e \sigma^3 f(H_2O) \exp(-Q_e/RT)$$

where $A_e$ is a material-dependent pre-factor, $f(H_2O)$ is either water fugacity (for quartz and feldspar) or the water content in ppm H/Si for olivine, $Q_e$ is the activation energy, $R$ is the gas constant, and $T$ is temperature in degrees K. Flow law parameters used for the calculations are listed in Table 1. Water fugacity or water content was calculated assuming water saturation, as this is most consistent with the levels of stress documented in areas of active tectonics (e.g., Hirth and Kohlstedt, 2003; Bürgmann and Dresen, 2008; Behr and Platt, 2014).

ii. Shear zones are initiated by a switch in dominant deformation mechanism from dislocation creep to grain-size sensitive creep, as a result of grain-size reduction driven by dynamic recrystallization. The dynamically recrystallized grain-size is related to the stress by experimentally determined paleopiezometers for quartz (Stupp and Tullis, 2003; Holyoke and Kronenberg, 2010), feldspar (Post and Tullis, 1999) and olivine (van der Wal et al., 1993). I assume the switch takes place when the stress at a depth of 25 km, within the lithospheric mantle, stress is low, the dynamically recrystallized grain-size is large, and the switch does not occur. Localization by grain-size reduction below this depth is therefore not predicted (Fig. 2).

iii. Once the shear zone has been formed, I assume the grain size continues to be controlled by the piezometric relationship, even after the dominant deformation mechanism has switched to grain-size sensitive creep (Platt and Behr, 2011b). This is because dislocation creep, which controls the grain-size, remains active at a rate controlled by the stress -- the switch in the dominant mechanism is a result of the increase in the rate of grain-size sensitive creep.

iv. The strain-rate in the shear zone at any depth is calculated from the stress $\sigma_{15}$ and the dominant flow law.

v. At any depth the cumulative width $w$ of the shear zones making up the plate boundary zone, the strain-rate in the shear zone $\dot{\varepsilon}$, the stress $\sigma_{15}$, and the imposed plate-boundary velocity $V$ are related by $w = V/2\dot{\varepsilon}$, where the strain-rate is the sum of the background dislocation creep rate and a rate given by a grain-size sensitive flow law

### Table 1

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
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<tr>
<td>$A$ (Pa m$^{-3}$ s$^{-1}$)</td>
<td>2.75E-34</td>
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<tr>
<td>$Q$ (J/mole)</td>
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<td>$V^*$ (m$^3$/mole)</td>
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<tr>
<td>$n$</td>
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<tr>
<td>$K$ (m Pa$^{-1}$)</td>
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<td>$\rho$</td>
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<tr>
<td>$q$</td>
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</tr>
<tr>
<td>$r$</td>
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</tr>
<tr>
<td>$k$ (W m$^{-1}$ K$^{-1}$)</td>
<td>3.0</td>
</tr>
<tr>
<td>$\kappa$ (m$^2$ s$^{-1}$)</td>
<td>1.0E-6</td>
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</table>

Note that the prefactor for DRX creep in quartz has small stress and temperature sensitivities (see reference 1). The value shown here is for 99 MPa differential stress and 320°C.
where $d$ is the grain-size, $q$ the grain-size exponent, and the other variables are as in Eqn (3). Flow law parameters used for the calculations are listed in Table 1.

The calculated stress and width profile for the shear zone is shown in Fig. 2. The profile is modified slightly from Platt and Behr (2011a), as I have used revised flow laws for olivine (Freed et al., 2012) and a lower geothermal gradient. I also corrected an error in the paper, in which they used the rate of simple shear instead of the second invariant of the strain-rate tensor for the flow-law calculations.

Stage 2: Shear heating

The assumptions underlying this stage of the calculations are discussed below. The step-by-step procedure is set out in the Appendix.

As a result of shear heating, the temperature in the shear zone increases, and the shear zone rheology undergoes thermal weakening. Heat loss to the surroundings limits the temperature increase, and will cause some thermal weakening in the country rock. As a result the shear zone may get wider. This effect is limited because thermal weakening of the shear zone results in a stress drop (see below), and any widening of the shear zone will also contribute to the stress drop; as a result there is unlikely to be much further microstructural weakening in the country rock. I therefore assume for simplicity that the shear zone remains the same width. The only situation in which the shear zone width should change significantly is if the drop in stress due to thermal weakening results in a sufficient increase in grain size to reverse the switch to grain-size sensitive creep. At that point strain localization would be controlled by the temperature perturbation only, and the shear zone would widen substantially. This situation did not occur in any of the calculations presented here.

It is important to note that if the plate boundary zone is divided up into a number of separate shear zones separated by distances of tens of km, as is the case in southern California, for example, then conductive heat loss from each shear zone is much more effective, and the resulting temperature increase is substantially reduced. I have not attempted to model this situation.

I use a constant velocity boundary condition for the deformation, corresponding to an imposed relative plate motion $V$, and for the calculations presented here, I assume that there is a single shear zone with width $w$, and that all deformation takes place within the shear zone. This assumption is justified by the common observation that plates do not deform significantly outside the plate boundary zones. Under the assumption of constant $w$ and $V$, the average strain rate in the shear zone at any particular depth remains constant, and hence the stress progressively decreases as temperature increases. This sets up a negative feedback: decreasing stress results in an increase in the stress-related grain size, which lessens the weakening effect of the grain-size sensitive flow law.

The thermal calculations were carried out in 1-D. The purpose of this exercise is to quantify the effect of shear heating and lateral heat loss on the shear-zone rheology, and the effects of the transfer of heat and stress from one level to another in the shear zone would complicate and obscure that investigation. The disadvantage of 1-D thermal models is that there is no steady-state solution, as there is no constant temperature in the far-field in the limit of infinite time. A 2-D model that allows heat loss to the surface can in principle reach steady state (e.g., Thatcher and England, 1998). Heat loss to the surface is likely to become important once the horizontal scale of the perturbation is comparable to its depth below the surface, which for depths of >15 km requires times >5 m.y. The calculations here are carried out for over a time period of 5 m.y., which in any case is close to the upper limit for the time scale of many plate tectonic processes. After 5 m.y., the rate of temperature change in the center of the shear zone is fairly constant, and relatively low (a maximum of 6°C/m.y.), as can be seen from the center column of Fig. 3, which shows the evolution of temperature with time. The rate of temperature change over the final 1 m.y. is shown in Table 2, as an aid to seeing how it is likely to evolve on longer time scales.

The heat equation at each step was solved using a Crank–Nicolson finite difference scheme. Domain boundaries for the calculations were set at least 50 km from the shear zone margin, and the temperature at the boundaries was fixed. The strain-rate within the shear zone at each point was determined from the rheology, temperature, and grain size, using the velocity boundary condition of 50 mm/yr imposed across the calculated cumulative width. This then determines the stress in the shear zone, which decreases as the temperature increases.

I have assumed that all mechanical work done in the shear zone is dissipated as heat. As discussed above, the proportion of the work energy involved in microstructural damage is likely to be <1%, is not constant, and will decline as the shear zone approaches microstructural steady state.

4. Results

Results of the calculations for shear zones at various depths in the crust and mantle are shown in Figs. 3–5 and Table 2, and summarized below. Note that stress in Figs. 3–5 and in the discussion below is expressed as the maximum shear stress for a plane stress environment, appropriate for a shear zone. This is the stress I have used to calculate shear heating. It is related to the differential stress, measured under uniaxial conditions and used in experimental flow laws and for paleopiezometry, by the factor $1/\sqrt{3}$ (see Behr and Platt, 2013, Appendix 1).

1. Narrow high-stress sections of the shear zone (1 km or less) generate significant heat, and the temperature may rise by as much as 120 °C over a 5 m.y. period. This can occur just below the brittle–ductile transition (e.g., at 15 km depth, top panel in Fig. 3), near the top of the feldspathic lower crust, and just below the Moho in the uppermost mantle (Fig. 4a). Much of the heat generated is lost by conduction to the surroundings, however, and after 5 m.y. the shear zone develops a conductive boundary zone >20 km wide on either side (Fig. 3). As a result, thermal gradients in the shear zone itself are negligible, and there is little tendency for strain to localize further within the zone.

2. Shear heating in these narrow, high-stress sections causes a significant drop in the shear resistance, particularly just below the brittle–ductile transition, and just below the Moho in the uppermost mantle, where it drops from 170 to 55 MPa (Fig. 3, right column, and Fig. 4b). This acts as a strong negative feedback on the shear heating process.

3. Shear zones with an initial width of several km have lower stresses and strain-rates (e.g., 26.9 km depth, at the base of the middle crust, and 41 km depth, in the upper mantle, shown in the two middle panels in Fig. 3). They generate heat less rapidly, but the temperature may still rise by as much as 10 °C/m.y., resulting in some thermal weakening. Conductive heat loss is significant, and on time-scales of a few million years this may result in a temperature difference of a few degrees between the center and margin of the shear zone (middle column of Fig. 3), but this is not sufficient to produce a significant difference in strain rate (Fig. 4e).

4. Shear zones at depths of >40 km in the mantle may be 50 km wide or more (e.g., 45 km depth, lower panel in Fig. 3), as a result of the limited degree of weakening produced by grain-size
sensitive creep at high temperature, low stress, and hence relatively large grain size. These wide shear zones experience relatively low stresses and strain-rates, and the temperature increase over a 5 m.y. period is small. Because of their width, a small difference in temperature between the center and the margin of the shear zone develops, resulting in an enhancement of the strain rate at the center by up to 18% (Fig. 4e).

Fig. 3. Shear heating in a microstructurally defined shear zone at depths of 15, 27, 41 and 45 km. Shear zone width, initial temperature, initial stress, and rheology from Fig. 2 and Table 1 (note that stress in Fig. 2 is differential stress, stress in this figure is shear stress). Panels arranged from left to right: temperature distribution across shear zone and surroundings after 5 m.y. (shear zone highlighted in yellow); evolution of temperature with time at center and margin of the shear zone; evolution of shear stress with time (shear stress is constant across shear zone). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

I carried out a total of 33 analyses covering depths from the brittle–ductile transition to the depth in the lithosphere at which weakening by grain-size reduction no longer takes place (~52 km for the chosen thermal gradient), the results of which are summarized in Fig. 4. The dependence of the temperature increase $\Delta T$ on the initial stress and on the shear zone width is illustrated in Fig. 5. The dependence of $\Delta T$ on the initial stress is very clear

### Table 2

Results for representative depths in the shear zone. Output variables are shown in italics in left column. Note that the water fugacity is a function of temperature and pressure, assuming maximum water activity. $\Delta T$ center-background: total increase in temperature at center of shear zone. $\Delta T$ center-margin: difference in temperature between the center and the margin of the shear zone. *For depths 15 and 26.9 km, no background rheology is assumed; shear stress is calculated from measurements in crustal mylonites by Behr and Platt (2011). Abbreviations as in Table 1.

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<th>27</th>
<th>35</th>
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<td>Background rheology</td>
<td>*</td>
<td>*</td>
<td>Fsp wet dis</td>
<td>Ol wet dis</td>
<td>Ol wet dis</td>
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<tr>
<td>Shear zone rheology</td>
<td>Qtz DRX</td>
<td>Qtz DRX</td>
<td>Fsp wet diff</td>
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<td>Final shear stress, MPa</td>
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<td>56.8</td>
<td>16.8</td>
</tr>
<tr>
<td>$\Delta T$ center-margin</td>
<td>1.0</td>
<td>0.9</td>
<td>0.8</td>
<td>0.8</td>
<td>3.8</td>
<td>8.5</td>
</tr>
<tr>
<td>$dT/dt$ last 1 my., °C/m.y.</td>
<td>4.9</td>
<td>1.1</td>
<td>4.4</td>
<td>4.4</td>
<td>6.4</td>
<td>5.2</td>
</tr>
</tbody>
</table>

(Fig. 5a), as we might expect. The relationship is non-linear, because of two competing effects. Firstly, rocks that sustain high stresses undergo a greater degree of grain-size reduction, weakening, and localization, so that the strain-rate increases as well as the stress, enhancing heat generation. Secondly, narrow shear zones lose heat more rapidly than wide ones. Lithology has little effect, except in the mantle lithosphere (blue symbols in Fig. 5), where localization effectively ceases at low stresses, limiting heat generation.

The dependence of ΔT on shear zone width is illustrated in Fig. 5b. The effect is again non-linear, because of the competing effects of increased heat generation and heat loss in narrow shear zones. Lithology also exerts a control: olivine generates much more heat for a given shear zone width than quartz, because of the higher stress required to deform it at a given rate. Note that the scale for feldspar (green symbols) in Fig. 5b is in meters, rather than km. Shear zones in feldspar-dominated rocks are predicted to be very narrow, because of the intense weakening caused by the grain-size reduction and the switch to diffusion creep. Conductive heat loss from these m-scale shear zones severely limits ΔT.

Thermal weakening in the shear zone causes a drop in stress, and the dynamically recrystallized grain-size may adjust accordingly. Because the rheology is grain-size sensitive, this will partly counteract the thermal weakening resulting from shear heating.
This effect was built into the calculations. Initial and final stresses and grain sizes are shown in Fig. 4.

5. Conclusions and implications

Shear heating can produce significant temperature increases in pre-existing shear zones created by microstructural weakening. Narrow shear zones in which both stress and strain-rate are high, may experience temperature increases up to 80 °C (e.g., just below the brittle–ductile transition) or 120 °C (just below the Moho) over a 5 m.y. period, but conductive heat loss results in a thermal anomaly extending >20 km on either side, so that thermal gradients within the shear zone itself are negligible. Thermal weakening lowers the stress in the shear zone by up to 50%, which may result in an increase in the dynamically recrystallized grain size, and together with conductive heat loss limits the increase in temperature.

Shear zones of the order of 10 km wide in the mid crust and upper mantle experience heating by several tens of degrees over 5 m.y., but thermal gradients within them are low. Heating in the mid-crust is unlikely to be sufficient to cause a perceptible change in metamorphic grade or partial melting. Shear zones in feldspathic lower crust are predicted to be very narrow (50m or less), but shear heating is limited by rapid conductive heat loss to <65 °C.

Shear zones ~50 km wide, which may develop at greater depths within the subcontinental lithospheric mantle, experience low stresses and strain-rates, and temperature increase is limited to ~20 °C, but because of their width, there may be a significant difference in temperature between the center and the margin of the shear zone, causing an enhancement of the strain rate at the center by up to 18%.

Note that all these conclusions apply to shear zones with displacement rates of 50 mm/yr. The rate of heat generation scales with the displacement rate, so shear zones with lower displacement rates will have correspondingly lower rates of heat generation. Shear zones with displacement rates of 50 mm/yr are unusual: even the San Andreas Fault has displacement rates of <30 mm/yr over most of its trajectory (e.g., Platt and Becker, 2010), because the relative plate motion is taken up on a set of sub-parallel faults spaced several tens of km apart.

A number of workers have suggested that shear heating has influenced the thermal structure of major faults in the continental lithosphere. These include the Ailao Shan–Red River fault zone in SE Asia (Leloup and Kienast, 1993), the Vincent fault in southern California (Graham and England, 1976), and the Main Central Thrust in the Himalayas (Molnar and England, 1990). The results presented here suggest that temperature increases caused by shear heating in the crust are unlikely to exceed 80 °C over a 5 m.y. period, and 5 °C/m.y. thereafter, for reasonable values of rock strength and slip rate. Shear heating over this period would also create thermal perturbations on the scale of tens of km on either side of the fault. In the mid-crust, shear stress is likely to be <10 MPa, and heating over a 5 m.y. period will be 10 °C or less, insufficient to cause partial melting or any detectable change in metamorphic grade. Techniques now exist for determining both the stress and the temperature in ductile shear zones (Kohn and Northrup, 2009; Thomas et al., 2010), and strain-rates can be determined from the shear zone width and the displacement rate. Appropriate field and microstructural observations should therefore allow a more precise assessment of the significance of shear heating in the evolution of these and other faults.

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Appendix

The step-by-step procedures for the calculations in this paper are set out below. Parameters values are given in Table 1 of the main text.

Shear zone width

- Depths to boundaries in the crust are: brittle–ductile transition (BDT), 15 km; transition from quartz-dominated to feldspar-dominated crust (QF), 27 km; transition to olivine dominated mantle (Moho), 35 km; base of lithosphere, 80 km. Vertical grid spacing is 100 m.
- Thermal gradients and densities are: surface to BDT, 20 °C/km, 2600 kg/m³; BDT to QF, 20 °C/km, 2700 kg/m³; QF to Moho, 16.25 °C/km, 2800 kg/m³; Moho to LB, 10 °C/km, 3260 kg/m³.
- Pressure (P) and temperature (T) are calculated on the grid.
- Water fugacity as function of P and T is loaded with values determined from the fugacity calculator at http://www.geo.umn.edu/people/researchers/withe012/fugacity.htm (Pitzer and Sterner, 1994). Water content of olivine is calculated from fugacity using expression from Freed et al. (2012).
- Differential stress was calculated as follows: surface to BDT, linear gradient from zero to 135 MPa; BDT to QF, stress array from Behr and Platt (2011), adjusted to the thermal gradient; QF to LB: differential stress calculated from wet dislocation creep flow laws (Eq. (3) in the main text) for feldspar from Rybacki et al. (2006) and for olivine from Freed et al. (2012) at a strain rate of 10⁻¹⁵ s⁻¹.
- Dynamically recrystallized grain sizes for quartz, feldspar, and olivine are calculated from differential stress using the piezometric relationships (Table 1).
- Strain-rates in the shear zone are calculated from the differential stress using grain-size sensitive creep flow laws (Eq. (4) in the main text): DRX creep flow law for quartz from Platt and Behr (2011b), and wet diffusion creep flow laws for feldspar from Rybacki et al. (2006) and olivine from Freed et al. (2012). See Table 1 for parameter values.
- Shear zone widths are calculated from Eq. (2) in the main text, using a plate velocity of 50 mm/yr and the calculated strain rates.

Shear heating

- Initial values for T and differential stress σ₁₅, rheological parameters, strain rate ε, in the shear zone, and shear zone width w, as functions of depth are passed from the previous step. Grain size is incorporated into the stress exponent nd, which assumes that grain size adjusts continuously with stress following the piezometric relationship (e.g., Montési and Zuber, 2002).
- Inner spatial domain set up for shear zone with width LS, and outer domain extending at least 60 km on either side of the shear zone boundaries, so as to isolate shear zone from the constant temperature boundary condition.
- At each time step the following calculations are carried out.
  1. The heat flow equation:
\frac{dT}{dt} = \alpha \frac{\Delta T}{\Delta x^2} \quad \text{where } x \text{ is normal to the shear plane, is}
\text{solved across the whole domain using a Crack Nicholson matrix with constant T boundary condition equal to the initial T.}

2. Velocities in the shear zone are calculated as follows. Assuming constant volume, the flow law is expressed using a rheological parameter \( B \) such that \( \eta_B = v_B^e/2B \) where \( v_B \) is the deviatoric stress tensor. For simple shear this means that the velocity gradient
\[ \frac{\partial v_B}{\partial x} = \frac{v_B}{B}, \]
where \( x \) is normal to the shear plane and \( y \) is the direction of shear. \( v_B \) is taken as \( \sigma_{ij}/3 \). Force balance dictates that \( v_B \) is constant in the \( x \) direction, hence:
\[ \frac{\partial (v_B/\text{Bouy}/\partial x)} = 0. \]
This is solved by discretization using boundary conditions of zero and \( V \) (plate boundary velocity, set to 50 mm/yr) on either side of the shear zone.

3. \( B \) is calculated from \( T \); the velocities are solved from \( B \); strain-rates are calculated at the nodes from the velocity; stress is calculated from strain-rate, \( B \) and \( nd \) (stress should be constant across shear zone); heat generation is calculated from stress and strain-rate; and a new temperature solution is calculated across the whole domain. \( T \), \( B \), velocities, and heat generation are interpolated between grid nodes for these calculations.

• Calculations are repeated for 5000 time steps of 1000 yr.
• The operation is repeated for each depth at 1 km intervals from 15 to 45 km.

References


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