Shear heating during distributed fracturing and pulverization of rocks

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ABSTRACT

We provide estimates of temperature changes produced in fault damage zones during brittle deformation associated with distributed cracking and pulverization. In contrast to localized faulting accompanied by significant frictional weakening, the relatively high friction coefficient on the multitudinous small cracks generated in the fracturing process can lead to significant shear heating. Simple calculations with parameter values constrained by laboratory experiments and simulations indicate that the temperature can increase during the generation of rock damage and pulverization by 100 °C or more. The results can help explain signatures of elevated temperature observed in geometrically complex fault zone sections with significant rock damage and regions with broad distributed deformation.

INTRODUCTION

Heat generation during faulting processes has been the subject of considerable research (e.g., McKenzie and Brune, 1972; Mase and Smith, 1987; Kanamori and Heaton, 2000; Rice 2006). Most studies have focused on highly localized deformation, motivated by the inverse relation between temperature change per heat production and width of the actively deforming region. This expectation of high temperatures in the fault core assumes implicitly that the resistance to shear motion is independent of the localization width. Efforts to observe signatures of frictional heat in highly localized sections of large faults have been generally unsuccessful (e.g., Brune et al., 1969; Lachenbruch and Sass, 1992; d’Alessio et al., 2003; Tanaka et al., 2006). This has led to the so-called “heat-flow/fault-strength paradox.” Proposed solutions for this paradox commonly involve strong dynamic weakening during slip (e.g., Ben-Zion, 2001; Rice, 2006).

Laboratory experiments with rocks and rock-like materials show indeed that progressive localization during brittle deformation is accompanied generally by significant weakening (e.g., Tsutsumi and Shimamoto, 1997; Di Toro et al., 2004; Han et al., 2007; Yuan and Prakash, 2008). This suggests that heat generation may be higher in locations where faulting is associated with distributed rather than localized deformation. In particular, faulting processes involving brittle rock damage in the form of distributed fracturing and pulverization may produce a larger temperature rise than frictional sliding in a highly localized but dynamically weak zone.

Recent observations from the geometrically complex trifurcation area of the San Jacinto fault zone (California, United States), show mineral changes in the fault damage zone that are consistent with temperatures of 100–150 °C (Morton et al., 2012). These elevated temperatures may have been produced by circulation of hot fluids through the fractured rocks (Morton et al., 2012), but they could also have been generated (at least partially) by shear heating during rock fracturing and pulverization. In the present paper we explore this possibility with basic analytical estimates calibrated by laboratory measurements. The results indicate that episodes of distributed brittle deformation with gradual extended weakening, of the type seen in laboratory experiments and simulations, can produce temperature changes of 100 °C or more. This may explain field evidence of elevated temperature in regions of distributed deformation.

AVAILABLE ENERGY

The strain energy, $E$, available for rock fracturing and heat generation is given by the area under the stress-strain curve:

$$E = \int \sigma \, d\varepsilon,$$

where $\sigma$ is stress and $\varepsilon$ is strain.

The shape of the stress-strain curve beyond the elastic limit depends on the post-failure mechanics of the material. If strain localizes onto a single slip plane or a narrow granular zone, the stress is controlled by friction. At low loading rates, the post-failure shear strength of both planar rock surfaces and granular layers is ~0.6 times the normal stress across the localization zone, i.e., the coefficient of friction ($\mu$) is ~0.6.

Figure 1 presents measurements of the friction coefficient for shear deformation of granular layers. Two cases are shown from experiments using a double-shear apparatus (Biegel et al., 1989). The dashed curve marked “All 710 µm” began with a layer of angular particles all having a diameter near 710 µm. The solid curve marked “Fractal” began with a power-law distribution of particle sizes ranging from 10 to 710 µm. Because the fractal distribution is packed more efficiently, it had a significantly lower initial porosity than the all–710 µm distribution. This difference in initial porosity is evident in the evolution of $\mu$ with strain in Figure 1. The all–710 µm distribution behaves like a typical underconsolidated soil in that the coefficient of friction gradually increases to the equilibrium value of $\mu = 0.6$. In contrast, the initially fractal layer behaves like an overconsolidated soil: friction rises quickly to a value in excess of $\mu = 0.6$ and then decreases gradually to the equilibrium value.

Figure 1 also shows the evolution of friction calculated by Abe and Mair (2009) using a discrete element model. Their model allows for multiple fragmentations of particles and gives one of the most realistic simulations of the evolution of friction with strain in a granular layer. Each of their initial particles is composed of 10,000 subparticles so they are able to produce a wide range of fragment sizes. Their simulation follows an underconsolidated approach to equilibrium similar to the all–710 µm experiment because they started with a distribution of same-sized particles. After the early evolution stage, the friction coefficient oscillates...
around \( \mu = 0.6 \). Both the all–710 \( \mu \) experiment and the discrete element simulation produce a fractal distribution of fragment sizes with a dimension near 2.6 (in three dimensions).

As noted, various mechanisms can produce significant weakening in highly localized faulting. For instance, there is evidence that the friction coefficient on sliding surfaces decreases significantly with increasing sliding rate, possibly due to a combination of flash heating at the contacts and thermal pressurization (e.g., Rice, 2006, and references therein). For slip on bimaterial interfaces in large-displacement localized faults, the juxtaposition of different elastic bodies across the interface leads to strong dynamic reduction of the normal stress (e.g., Ben-Zion, 2001, and references therein). These and other dynamic weakening mechanisms can explain the lack of evidence for significant frictional heat on highly localized fault sections.

Earthquake faults are geometrically complex and have regions (e.g., near stepovers and branching) associated with distributed brittle deformation. In such places the strain does not localize into a narrow zone. Rather, distributed strain produces a myriad of smaller faults and there is no expectation that the coefficient of friction on these small distributed fractures will decrease. Shear deformation of the fractured or pulverized rocks during earthquake failure episodes can therefore generate significant heat. Strain markers are not usually observed in field studies of pulverized rocks (e.g., Dor et al., 2006; Mitchell et al., 2011). One interpretation is that they did not experience significant shear (e.g., Brune, 2001). This may be the case if most of the original grains are highly fractured but still identifiable. However, laboratory experiments and computer simulations indicate that only some of the original grains maintain their identity. Most grains experience shear and size reduction in a progressive fragmentation process that destroys shear markers. As examples, Biegel et al. (1989) produced a fractal grain size distribution in experiment with shear strain of 3 and found no macroscopic evidence of the strain in the fragmented layer, and Abe and Mair (2009) produced a fractal gouge in discrete element simulations at strain near 1 with no remnant structural evidence of shear deformation.

For guidance to the post-failure behavior of rock pulverized at high loading rates, we refer to recent Hopkinson split-bar experiments on granite (Yuan et al., 2011). Figure 2 shows selected stress-strain curves from their results for two values of the confining stress: 60 MPa provided by a copper confining sleeve (solid curves) and 132 MPa provided by a brass sleeve (dashed curves). For both confining pressures, and in earlier experimental results of Doan and Gary (2009), pulverization did not occur at low strain rates. Rather, strain localized in such cases onto a primary slip plane and the axial stress fell precipitously beyond failure. The horizontal lines show the minimum stress at which pulverization began (solid line for 60 MPa, dashed for 132 MPa confinements). For the cases leading to pulverization, the stress falls to the post-failure stress \( \sigma_p \), while for extended weakening, \( E = \frac{1}{2} (\sigma_e - \sigma_p) \varepsilon_p \).

\[
E = \frac{\sigma_e \varepsilon_e}{2} + \frac{(\sigma_e - \sigma_p)\varepsilon_e}{2} + \sigma_p (\varepsilon_e + \varepsilon_p).
\]

Because the elastic strain is typically small, we can ignore the first term and calculate two end-member cases: instant strain weakening with \( \varepsilon_e = 0 \), where the stress falls immediately from \( \sigma_e \) to \( \sigma_p \); and extended strain weakening, in which \( \varepsilon_e = 0 \) and the stress decreases from \( \sigma_e \) toward \( \sigma_p \) during the entire deformation process. For instant weakening,

\[
E = \sigma_p \varepsilon_p,
\]

while for extended weakening,

\[
E = \frac{1}{2} (\sigma_e - \sigma_p) \varepsilon_p.
\]

Some of the strain energy \( E \) is radiated as seismic waves and some is used in the fragmentation process. Seismological, computational, and laboratory studies indicate that the radiated seismic energy during earthquake processes is just a few percent of the strain energy reduction (e.g., Kanamori et al., 1998; McGarr, 1999; Shi et al., 2008; Lockner and Okubo, 1983). The following simple calculation shows that the energy required to pulverize the rock is also very small. Consider a 1 m cube that is pulverized into an array of 1 \( \mu \)m cubes. The surface area will increase to \( 6 \times 10^9 \) m\(^2\). Assuming a density of 3000 kg/m\(^3\), the surface area density is 2

![Figure 2. Selected stress-strain curves at two values of confining stress from Hopkinson split-bar experiments on granite (Yuan et al., 2011). The maximum stress increases generally with strain rate. Horizontal lines (solid line for 60 MPa, dashed for 132 MPa confinements) indicate the boundary between samples that failed along a localization zone and those that were pulverized.](image1)

![Figure 3. Idealized representation of stress-strain curves for the pulverization regime in Figure 2. Maximum elastic stress is \( \sigma_e \), and elastic strain is \( \varepsilon_e \); \( \varepsilon_t \) is the transitional strain over which the stress falls to the post-failure stress \( \sigma_p \). Additional permanent strain \( \varepsilon_p \) occurs at \( \sigma_p \).](image2)
ties in given fault sections and nearby faulting that produce compressive stress can increase locally the confining pressure from the values assumed in Figure 4. Our assumption of a flow stress that is 0.6 times the confining stress on the myriad of internal failing planes may also be a conservative estimate. For the highest strain rates in Figure 2, the axial flow stress (σr) is 900 MPa when the radial confining stress (σt) is 132 MPa. From a standard Mohr’s circle construction, the coefficient of friction on an optimally oriented slip plane with no cohesion is

\[ \mu = \tan \left[ \sin^{-1} \left( \frac{(\sigma_t - \sigma_r)}{(\sigma_t + \sigma_r)} \right) \right] \]  

(6)

For the experimentally observed values of σt and σr, this gives μ = 1.1, which is nearly twice the value we assumed.

Because the expected temperature rise associated with a given strain is proportional (in the Druker-Prager approximation) to the mean stress, higher temperatures are expected at greater depths. Therefore, the discussed mechanism combined with the fact that fault zones have relatively high permeability for fault-parallel flow can generate hot fluids at depth that may then move toward the surface. Our results explain the signatures of elevated temperatures in a section of the San Jacinto fault zone with a broad damage zone and pulverized rocks (Morton et al., 2012), and are likely to be relevant to shear heating in other fault locations where geometrical complexities lead to distributed brittle damage rather than macroscopic localization and frictional weakening. On a larger scale, Devèes et al. (2011) used nonlocalized strain at high stress to explain important topographic features of the Dead Sea region. They further suggested that shear heating in areas of distributed deformation may be important for metamorphism and magma generation. These suggestions are consistent with our basic estimates.

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Figure 4. Estimated temperature rise, ΔT, for the two end-member cases of instant and extended weakening (σe — elastic stress; σf — post-failure flow stress). Solid lines are based on failure strength of granite at 5 km depth (σf = 1000 MPa), while dashed lines are based on failure strength of granite at 1 km depth (σf = 500 MPa). For both cases, we assume post-failure flow stress of σf = 0.6σe. Intersections with horizontal dashed line indicate shear strain required to produce a 100 °C rise in temperature.
ADDENDUM

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The Devès et al. (2011) reference in the last paragraph of the paper should be accompanied by the following related references:


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The minus sign in Equation 4 is a typographical error that should be replaced with a plus sign, as followed from Equation 2.