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Coseismic damage and softening of fault rocks at seismogenic depths

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ABSTRACT

Elastic stiffness, a critical property for stress-orientation, propagation of earthquake ruptures and associated seismic waves, and the capability of crustal rocks to store strain energy, is expected to be highly variable throughout the seismic cycle due to complex sequences of damage and healing. Post-seismic healing and exhumation-related alteration render it impossible to assess how well rock stiffness as measured in the laboratory on samples collected from fault zones represents *in situ*, coseismic rock stiffness at seismogenic depths. Here we estimate the *in situ*, coseismic stiffness of fault rocks from the pseudotachylyte-bearing Gole Larghe Fault Zone (Italian Southern Alps), using aspect ratio measurements of pseudotachylyte injection veins and numerical simulations. Aspect ratios of injection veins cutting across tonalite and cataclasite exhibit a maximum vein aperture positively correlating with vein length. To model vein opening, fault and injection veins are assumed to be filled with pressurized melt. Consistent with recent results from studies of melt lubrication we assume that the magnitude of the melt pressure is in equilibrium with the *fault-normal* stress and the fault vein approximately maintains constant thickness during slip. The numerical simulations of injection vein opening due to pressurized frictional melt indicate that the average *in situ* coseismic stiffness of the wall rocks is 5–50 times smaller than the stiffness obtained from laboratory measurements on the same rocks in their present-day state. The disagreement between laboratory measurements and simulations brings into question the appropriateness of using laboratory-derived values for rock stiffness to model coseismic processes at depth.

1. Introduction

The bulk elastic stiffness of faulted crust constrains the amount of elastic strain energy that can be stored during tectonic loading and released during earthquakes and also exerts a direct control on the earthquake rupture velocity (Biegel et al., 2008). Near-fault stiffness changes can result in significant stress rotation, allowing faults to slip under less-than-optimal far-field stress states (Rice 1992; Faulkner et al., 2006). Therefore evolution of rock stiffness during earthquakes is critical to understanding seismic rupture nucleation, propagation, and arrest. Fracture and pulverization damage surrounding crustal scale faults can result in stiffness reductions of $\sim$40% over large fault-normal distances ($>$ 1 km) and up to 8 km depth, and this softening has been shown to persist 1 ky or more (Cochran et al., 2009; Li et al., 1994, 1998, 2006). At critical seismogenic depths, pressure, temperature, and fluid flow tend to increase the degree and rate of healing of damaged rocks (e.g. Brantley et al., 1990; Moore et al., 1994; Morrow et al., 2001; Tenthorey and Cox, 2006; Tenthorey and Fitz Gerald, 2006). Several studies document significant coseismic drops in velocity, believed to be associated with rupture-induced damage, followed by time dependent increases in velocity over timescales of 2–10 yr (Brenguier et al., 2008; Hiramatsu et al., 2005; Vidale and Li, 2003). Direct and indirect observations constrain *in situ* rock mechanical properties along seismogenic faults to $\sim$3 km (i.e., San Andreas Fault Observatory at Depth drilling project, Zoback et al., 2010) and $\sim$8 km depth (e.g., Cochran et al., 2009; Li et al., 2006).

The fact that fault rock stiffness evolves throughout the seismic cycle presents a conundrum for scaling laboratory-derived rock properties which can vary with damage-related elastic stiffness changes (e.g., hydraulic conductivity, compressibility, and thermal expansion coefficient) to natural fault processes. For example, theoretical models of coseismic processes, such as sliding friction and heat and pore-fluid flow, are based on laboratory measurements of rock properties (e.g. Lachenbruch, 1980; Noda and Shimamoto, 2005; Rice, 2006). These measurements in the laboratory are performed on specimens collected

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from exhumed fault zones, or from boreholes much shallower than typical seismogenic depths (e.g., Faulkner et al., 2006). Given that significant healing has taken place during exhumation (e.g., Gratier et al., 2003; Faulkner et al., 2006), use of such laboratory results in modeling relies on potentially biased estimates of rock properties.

Our study addresses the questions: (1) "What is the *maximum* reduction in rock stiffness near a fault at seismogenic depths?" and (2) "How does this estimate compare to laboratory measurements on rocks collected from exhumed fault zones?". Answering (1) requires one to eliminate effects of the poorly-constrained healing process at depth. Given that the damage source is fault rupture and slip, one thus has to find the *in situ*, *coseismic* rock stiffness. We estimate this *in situ*, *coseismic* rock stiffness by simulating the formation of injection veins, interpreted to represent coseismic fault structures found along pseudotachylyte (solidified melt derived from frictional heating) bearing faults. Presumably, simulations matching our field observations best constrain stiffness at the time of vein formation. Question (2) can be answered by comparing our modeled values to laboratory data for samples from the analyzed outcrops.

2. Geologic background: The Gole Larghe Fault Zone

During the last decade, abundant evidence has been reported for seismic slip along the Gole Larghe Fault Zone (GLFZ), Southern Alps, Italy (Di Toro and Pennacchioni, 2004, 2005). The GLFZ is a right-lateral strike slip fault-zone with a minor reverse slip component that cuts east-to-west across the Adamello tonalitic batholith, and comprises hundreds of individual fault strands across a zone ~500 m wide (Di Toro and Pennacchioni, 2005). Fault strands hosting pseudotachylytes, generally accepted as evidence of seismic slip (Sibson, 1975), nucleated on pre-existing joints at depths of 9–11 km at ambient temperatures of 250–300 °C approximately 30 Ma ago (Di Toro and Pennacchioni, 2004; Pennacchioni et al., 2006). In the GLFZ, fault strands cut and offset abundant aplite dikes, and fault rocks on individual strands consist either of cataclasite, cataclasite overprinted by pseudotachylyte, or, in rare cases, only pseudotachylytes. Pseudotachylyte veins are easily distinguished from the host tonalite and cataclasite in polished outcrops at the base of the Lobbia glacier (Fig. 1a and b), and occur as abundant millimeter to centimeter thick fault and injection veins, or less commonly as pseudotachylyte breccias. Pseudotachylyte injection veins intrude both cataclasite and tonalite and range in length from sub-millimeter to approximately one meter.

3. Injection veins

3.1. Field observations

In general, pseudotachylyte injection veins within the GLFZ intrude the host tonalite, emanating from pseudotachylyte veins running along the fault surface (fault veins). Di Toro et al. (2005b) measured orientations of a large number of injection veins, and found that (1) > 70% occurred on the south side of the fault veins, and (2) two sets dominate: one making a small (15–25°) angle with the main fault veins (Set I), and one almost orthogonal to the fault veins (Set II). Set II injection veins were likely formed as virgin fractures near propagating earthquake rupture tips whereas Set I injection veins are parallel to and intrude pre-existing minor cataclasite faults persisting throughout the GLFZ (Di Toro et al., 2005b). Some isolated pseudotachylyte injection veins and reservoirs are also located at bends and stepovers along faults (e.g., Di Toro et al., 2005a).

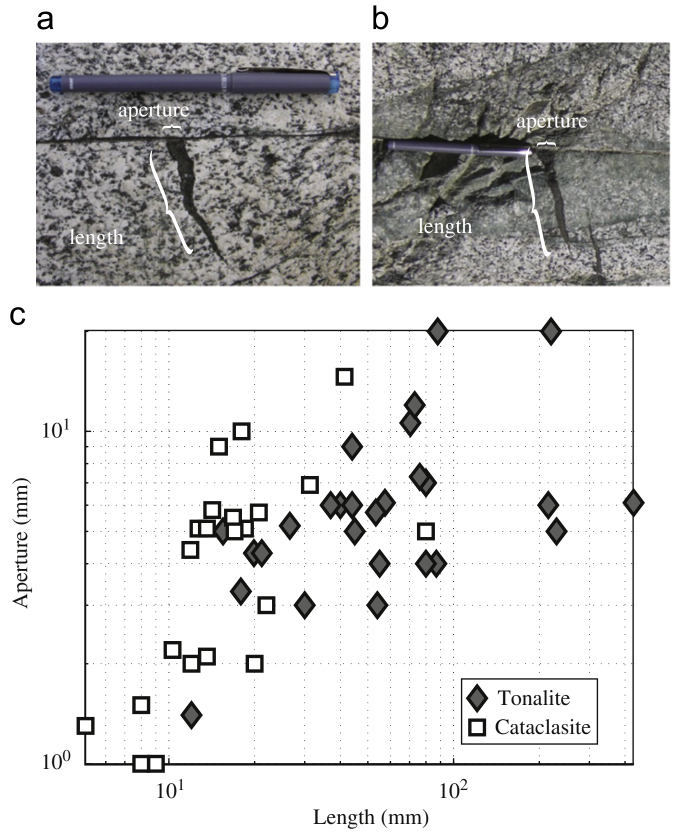

Fig. 1. Injection vein field photos and data. (a) Injection vein cutting tonalite, (b) injection vein cutting cataclasite, and (c) aperture and length of pseudotachylyte injection veins which cut cataclasite (empty squares) and tonalite (gray diamonds) in the Gole Larghe Fault Zone.

Displacement across the injection veins is principally opening, and the maximum aperture is typically at their base (i.e., the point of intersection with the fault vein, Fig. 1). In general, injection vein opening distribution is that of a half-bell, tapering rapidly from the vein base and less rapidly toward the tip. Injection veins extending variable distances into the wall rock are most commonly truncated by other fractures, and less-frequently terminate at discrete tips in otherwise monolithic rock.

Measurement of > 50 injection veins reveal a positive correlation between maximum aperture and length (Fig. 1c). The outcrop surface in most cases is glacially-polished allowing us to view the injection veins in great detail. However, the hummocky nature of the outcrop surface means that exposures are not exactly perpendicular to the injection veins potentially biasing aperture measurements. We took care to correct for this distortion geometrically, so aperture measurements should reflect the true aperture closely. The errors are acceptable since most injection veins do not deviate far from being perpendicular to the exposure surface and the apparent aperture (A_{ap}) as exposed in the field differs from the true aperture (A) by only 10% for exposures that deviate from vein-perpendicular by as much as $\alpha = 25°$ because $A_{ap} = A/\cos\alpha$.

3.2. Microstructural observations

Based on field observations revealing that the displacement across veins is primarily opening, pseudotachylyte injection veins have been interpreted as fluid-filled mode I cracks (e.g., Sibson, 1975; Grocott, 1981; Swanson, 1992; Di Toro et al., 2005a,b).