Fault structure, frictional properties and mixed-mode fault slip behavior

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ABSTRACT

Recent high-resolution GPS and seismological data reveal that tectonic faults exhibit complex, multi-mode slip behavior including earthquakes, creep events, slow and silent earthquakes, low-frequency events and earthquake afterslip. The physical processes responsible for this range of behavior and the mechanisms that dictate fault slip rate or rupture propagation velocity are poorly understood. One avenue for improving knowledge of these mechanisms involves coupling direct observations of ancient faults exhumed at the Earth’s surface with laboratory experiments on the frictional properties of the fault rocks. Here, we show that fault zone structure has an important influence on mixed-mode fault slip behavior. Our field studies depict a complex fault zone structure where foliated horizons surround meter- to decameter-sized lenses of competent material. The foliated rocks are composed of weak mineral phases, possess low frictional strength, and exhibit inherently stable, velocity-strengthening frictional behavior. In contrast, the competent lenses are made of strong minerals, possess high frictional strength, and exhibit potentially unstable, velocity-weakening frictional behavior. Tectonic loading of this heterogeneous fault zone may initially result in fault creep along the weak and frictionally stable foliated horizons. With continued deformation, fault creep will concentrate stress within and around the strong and potentially unstable competent lenses, which may lead to earthquake nucleation. Our studies provide field and mechanical constraints for complex, mixed-mode fault slip behavior ranging from repeating earthquakes to transient slip, episodic slow-slip and creep events.

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1. Introduction

A traditional interpretation of tectonic faults is that stress is relieved either as earthquakes resulting from sudden frictional instabilities or by continuous aseismic frictional sliding and fault creep (e.g., Scholz, 1998). In this view, crustal faults have a stable region near the surface (0–3 km), owing to the presence of loosely consolidated material (Marone and Scholz, 1988), and at depth (15–20 km), owing to the onset of viscous deformation of fault rocks with increasing temperature (Brace and Kohlstedt, 1980). Between these depths, within the seismogenic zone, fault slip is envisaged to occur primarily by earthquakes. Frictional processes and the parameters that dictate the stability of frictional sliding are therefore critical for the physics of earthquakes. Several lines of evidence suggest that most seismically active faults are statically strong structures with friction, μ, in the range 0.6–0.85 (e.g. Byerlee, 1978; Scholz, 2000). Geological evidence along seismic faults suggests: 1) that earthquake ruptures are localized within principal slip zones less than a few cm thick, with evidence in some cases of extreme localization within zones mm or less in thickness (e.g., Power and Tullis, 1989; Sibson, 2003) and 2) that dynamic processes may result in a dramatic strength loss from pre-slip, high static friction, to co-seismic, low dynamic friction (e.g. Di Toro et al., 2011; Kitajima et al., 2010). Alternative situations exist in at least some cases, such as the San Andreas in central California, where recent work shows that extreme fault weakness (μ≈0.1) occurs within a 3 m wide creeping fault core (Zoback et al., 2010) due to the presence of weak clay minerals (Carpenter et al., 2011).

In the last decade, the development of highly sensitive surface and borehole seismometers and the improvement of geodetic networks have resulted in a re-evaluation of how crustal faults accommodate plate motions (e.g., Peng and Gomberg, 2010). Tectonic faults, in fact, appear to fail by a continuous spectrum of slip modes. Details of aseismic fault creep were first highlighted along the San Andreas fault (e.g. King et al., 1973) and then recognized along other crustal-scale structures (Cashman et al., 2007; Chen et
The Zuccale fault is a regionally-important low-angle normal fault exposed on the isle of Elba in Central Italy. Geological data suggest that the fault has accommodated a total shear displacement of 6–8 km and that the fault zone structure and fault rocks formed at less than 8 km crustal depth (Collettini et al., 2009a). The present day fault structure is the final product of several deformation processes superposed during fault zone exhumation (Smith et al., 2007, 2011). Here, we focus on a series of highly foliated and phyllosilicate-rich fault rocks that represent the basal horizon of the low-angle fault zone. These phyllosilicate-rich fault rocks, formed by dissolution and precipitation processes in carbonate rocks (Collettini et al., 2009a), surround lenses of more competent, non-foliated fault rock materials (Figs. 1 and 2) with observed dimensions ranging from centimeters to tens of meters (though larger lenses may occur). Within the foliated horizons the deformation is continuous and accommodated predominantly by frictional sliding along phyllosilicate lamellae (Collettini et al., 2009a). Frictional sliding is accompanied by the development of foliation-parallel extension veins filled with calcite (Fig. 2b); the crack-and-seal texture of the calcite veins implies cyclic fluid pressure build-up during fault activity (Collettini et al., 2006). The competent lenses are made of intact carbonates, which are not significantly affected by dissolution processes (Fig. 2a), and mafic materials that have been sheared off the wall rocks and incorporated into the fault zone (Fig. 2c–e). In the competent lenses, part of the deformation is localized along discrete minor faults (Fig. 2a and c) with associated cataclastic fault rocks: the presence of calcite veins suggests fluid-driven brittle processes (Fig. 2b and d). In some outcrops, it is possible to observe minor faults that cross both the competent lenses and part of the foliated horizons (Fig. 2e). Minor footwall normal faults, with meter-scale displacements, merge into the base of the low-angle fault zone (Fig. 1 and Smith et al., 2007). Field evidence indicates that low-angle slip within the foliated, phyllosilicate-rich horizons containing the competent lenses occurred broadly synchronously with slip along the minor footwall normal faults. The foliated and phyllosilicate-rich horizons presumably formed continuous layers in the early stages of fault activity, but during fault zone evolution, growth of the footwall normal faults led to thickening of the phyllonitic fault core and eventual dismemberment of the phyllosilicates into a series of isolated fault rock units: the present day thickness of the phyllonites is in the range 0–8 m (details of fault slip history and fault zone evolution are presented in Smith et al., 2007; 2011).

Our field studies document a fault zone structure characterized by a mixture of materials and fault slip behavior. To quantify these observations, we investigated the frictional strength and slip stability of the fault rocks via detailed laboratory experiments.

3. Frictional properties of the fault rocks

3.1. Frictional strength

We selected fault rocks representative of both the foliated phyllosilicate-rich horizons, zones 1 and 2 (Figs. 1 and 2 and Table 1),
and the competent lenses, zones 3 and 4 (Figs. 1 and 2 and Table 1). In zones 1 and 2 the microstructure involves an interconnected and phyllosilicate-rich network (smectite, talc and chlorite) surrounding clasts of stronger mineral phases (calcite and tremolite). Zone 2 is characterized by a higher content of tremolite (Table 1), which gives it a more greenish color compared to zone 1 which is whitish in color. The mineralogy of zone 3 is similar to zone 2, but the deformation is localized along small brittle faults.

Table 1

<table>
<thead>
<tr>
<th>Exp.</th>
<th>Sample</th>
<th>Normal stress (MPa)</th>
<th>Sliding velocities (µm/s)</th>
<th>Mineralogy</th>
</tr>
</thead>
<tbody>
<tr>
<td>P2041</td>
<td>4 powder</td>
<td>10, 20, 35, 50</td>
<td>1, 3, 10, 30, 100, 300</td>
<td>Bt 33, Chl 12, Hbl 55</td>
</tr>
<tr>
<td>P2044</td>
<td>4 powder</td>
<td>75, 100, 150</td>
<td>1, 3, 10, 30, 100, 300</td>
<td>Bt 33, Chl 12, Hbl 55</td>
</tr>
<tr>
<td>P2048</td>
<td>1 foliated</td>
<td>10, 20, 35, 50</td>
<td>1, 3, 10, 30, 100, 300</td>
<td>Cal 30, Tlc 14, Sm 20, Tr 26, Chl 1</td>
</tr>
<tr>
<td>P2052</td>
<td>1 foliated</td>
<td>75, 100, 150</td>
<td>1, 3, 10, 30, 100, 300</td>
<td>Cal 30, Tlc 14, Sm 20, Tr 26, Chl 1</td>
</tr>
<tr>
<td>P2053</td>
<td>2 foliated</td>
<td>75, 100, 150</td>
<td>1, 3, 10, 30, 100, 300</td>
<td>Cal 43, Tlc 6, Sm 14, Tr 36, Chl 1</td>
</tr>
<tr>
<td>P2054</td>
<td>2 foliated</td>
<td>10, 20, 35, 50</td>
<td>1, 3, 10, 30, 100, 300</td>
<td>Cal 43, Tlc 6, Sm 14, Tr 36, Chl 1</td>
</tr>
<tr>
<td>P2056</td>
<td>3 powder</td>
<td>10, 20, 35, 50</td>
<td>1, 3, 10, 30, 100, 300</td>
<td>Cal 43, Tlc 6, Sm 14, Tr 36, Chl 1</td>
</tr>
<tr>
<td>P2066</td>
<td>3 powder</td>
<td>75, 100, 150</td>
<td>1, 3, 10, 30, 100, 300</td>
<td>Cal 43, Tlc 6, Sm 14, Tr 36, Chl 1</td>
</tr>
<tr>
<td>P2057</td>
<td>2 foliated/wet</td>
<td>10, 20, 35, 50</td>
<td>1, 3, 10, 30, 100, 300</td>
<td>Cal 43, Tlc 6, Sm 14, Tr 36, Chl 1</td>
</tr>
</tbody>
</table>
affected by cataclastic processes (Fig. 2a). Zone 4 is derived from a mafic protolith and is composed of hornblende, biotite and chlorite. To take into account the role of foliation in frictional properties we conducted friction experiments on: a) intact fault rocks of the foliated samples sheared under their in situ geometry, i.e. zones 1 and 2 (Collettini et al., 2009b) and 2) powders prepared by crushing fault rock materials from the competent lenses, i.e. zones 3 and 4. The competent lenses are cross-cut by minor faults characterized by cataclasis and grain size reduction (Fig. 2a, c and e), suggesting that experiments on both powders rather than on solid samples, are more appropriate to investigate the frictional properties of these fault rocks. Powders were obtained by crushing and sieving (150 μm) fault rocks from the competent lenses. For experiments on intact, foliated fault rocks we used natural samples 0.8–1.2 cm thick and 5 × 5 cm in area and followed procedures documented in Collettini et al. (2009b) and Ikari et al. (2011).

Frictional sliding experiments were conducted in the double direct shear configuration (Fig. 3a inset) at room temperature and humidity (~25 °C and ~25% relative humidity), over a range of normal stresses from 10 to 150 MPa. The foliated fault rocks from zones 1 and 2 exhibit sliding friction coefficients in the range 0.25–0.31 (Collettini et al., 2009b), implying that the phyllosilicate-rich fault rock horizons are weak (Fig. 3b). In contrast, fault rocks from the competent lenses of zones 3 and 4 exhibit sliding friction coefficients in the range 0.55–0.62 (Fig. 3b), suggesting that the competent lenses are strong fault portions.

### 3.2. Constitutive modeling for friction parameters

In order to investigate possible modes of fault slip-behavior, we measured the velocity dependence of sliding friction of the fault rocks. For all samples, after a ‘run-in’ phase (Fig. 3a) at a constant sliding velocity of 10 μm/s, we applied a velocity-stepping sequence consisting of 5 steps from 1–300 μm/s, with displacements during individual steps of 500 μm. Velocity step tests (Fig. 3c) involved imposing step changes in sliding velocity from V₀ to V (e.g., Dieterich and Kilgore, 1979; Marone, 1998; Ruina, 1983; Scholz, 1998). For each velocity step, the instantaneous change in friction coefficient scales as the friction parameter a ln(V/V₀), and is always of the same sign as the velocity change. The decay to a new steady state value of sliding friction scales as the friction parameter b ln(V/V₀), and is usually of the opposite sign of the velocity change. a and b are empirical constants (unitless) defined as the direct effect and the evolution effect respectively (e.g., Ruina, 1983). Frictional velocity dependence is defined as:

\[
(a-b) = \frac{\Delta \mu_{ss}}{\ln(V/V_0)}
\]

where \(\Delta \mu_{ss}\) is the change in the steady state coefficient of friction due to a change in sliding velocity from \(V_0\) to \(V\). Positive values of \((a-b)\) indicate velocity-strengthening frictional behavior which results in stable fault sliding and is associated with aseismic fault creep. Negative values of \((a-b)\) indicate velocity-weakening frictional behavior, which is a prerequisite for unstable, stick-slip behavior associated with earthquake nucleation (e.g., Dieterich and Kilgore, 1994; Marone, 1998; Scholz, 1998).

We modeled friction data from velocity-step tests using the rate and state variable friction equations (Fig. 4 and Ruina, 1983; Saffer and Marone, 2003):

\[
\mu = \mu_0 + a \ln\left(\frac{V}{V_0}\right) + b \ln\left(\frac{V_0}{V}\right)
\]

Fig. 4. a) Modeling results for constitutive friction parameters a, b, d₁, and a–b for experiment p2048, zone 1 foliated. b) For experiment on zone 3 powders, the frictional decay is better represented by two evolution parameters b₁ and b₂ (where b = b₁ + b₂) and two associated critical slip distances d₁ and d₂. The black line is evolution of friction upon a velocity step, the red line is the modeling result. Friction parameters are presented in Appendices A and B.
\[ \frac{d\theta}{dt} = \left( 1 - \frac{V}{V_{lp}} \right) \text{Dieterich evolution equation} \]

where \( \mu_c \) is a reference friction value, \( \theta \) is a state variable and \( d_i \) is the critical slip distance. The state variable is inferred to be the average lifetime of asperity contacts that control friction, and the critical slip distance is the displacement over which the population of asperity contacts are renewed (Marone and Kilgore, 1993; Scholz, 1998). The interaction of the sample with its elastic surroundings is incorporated according to the equation:

\[ \frac{d\mu}{dt} = k(V_{lp} - V) \]

where \( k \) is the stiffness of the fault surroundings (in the laboratory this represents the testing apparatus and the sample blocks), normalized by the normal stress \( \left( k = -3 \times 10^{-3} \mu \text{m}^{-1} \text{ at } 25 \text{ MPa normal stress} \right) \). \( V_{lp} \) is the load point velocity, and \( V \) is the true slip velocity. We solve Eqs. (2) and (3) simultaneously, with Eq. (1) as a constraint, using a fifth-order Runge–Kutta method. The constitutive parameters (presented in Appendices A and B) are obtained as solutions to a nonlinear inverse problem using an iterative least squares method (e.g., Blanpied et al., 1998; Reinen and Weeks, 1993; Saffer and Marone, 2003). For some samples, zone 3, the decay of friction to a new steady state condition following a velocity step is better modeled using two state variables, leading to two evolution parameters, \( b_1 \) and \( b_2 \) \( (b = b_1 + b_2) \) associated with two critical slip distances, \( d_{i1} \) and \( d_{i2} \) respectively. Fig. 4 shows examples of one- and two-state variable behavior.

### 3.3. Friction parameters for foliated horizons and microstructures

In our experiments the intact foliated rocks exhibit positive \( (a - b) \) values for the entire range of normal stresses \( (10 - 150 \text{ MPa}) \) and sliding velocities (Fig. 5). The degree of velocity strengthening increases with sliding velocity, with \( (a - b) \) values up to 0.009 and 0.007 for zones 1 and 2 respectively (Fig. 5a and b). Analysis of the friction parameters \( a \) and \( b \) shows that the foliated rocks are characterized by a distinct direct effect that is independent of the applied normal stress and that increases with increasing sliding velocity (Fig. 6), whereas the evolution effect is very small, in some cases close to zero. In these samples, deformation is mainly accommodated within the phyllosilicate-rich horizons by foliation-parallel sliding surfaces, interlayer delaminations and micro-folds (Fig. 7a–d), whereas cataclasis and grain size reduction is rare. These microstructural observations are consistent with the observed low \( b \) values, confirming the concept of “contact saturation”, in which mineral surfaces are in complete contact and thus the real area of contact does not evolve when the velocity is perturbed (e.g. Ikari et al., 2009; Niemeijer et al., 2010; Saffer and Marone, 2003).

### 3.4. Friction parameters for the competent lenses and microstructures

For the fault rocks derived from the competent lenses, we observe that zone 3 shows an evolution from velocity strengthening, \( (a - b) \approx 0.003, \) to velocity-weakening behavior, \( (a - b) \approx 0.005, \) with increasing sliding velocity (Fig. 5b). Zone 4 shows a similar evolution towards velocity-weakening with increasing sliding velocity, but some velocity-weakening is also measured at low sliding velocities (Fig. 5a). For both zones 3 and 4 the friction parameters \( a \) and \( b \) are independent of the applied normal stress and they increase with increasing sliding velocity (Fig. 6). A distinct evolution effect is documented at high sliding velocities and in particular for zone 4 (Fig. 6). The microstructure of the powdered materials from the fault lenses is generally characterized by two fine-grained slipping zones located near the boundaries of the shear zone and a central zone not affected by grain size reduction (e.g. Fig. 7e–i). Microstructural observations suggest that deformation is mainly accommodated by cataclasis with grain size reduction and localization along B, R1 and Y shear planes (e.g. Beeler et al., 1996; Cowan et al., 2003; Logan, 1978; Marone, 1998).

### 4. Discussion

The structure of the Zuccale fault zone seems to be quite similar to other mature, phyllosilicate-rich fault zones (e.g. Gueydan et al., 2003; Holdsworth, 2004; Moore and Rymer, 2007; Wintsch et al., 1995). For example, the Carboneras fault zone in Spain is a c.1 km wide strike-slip fault zone composed of continuous strands of foliated phyllosilicate-rich horizons that bound lenses of more competent materials cross-cut by discrete brittle faults (Faulkner et al., 2003). A similar fault zone structure is proposed for the Median Tectonic Line in Japan (Jeffries et al., 2006; Wibberley and Shimamoto, 2003). Additionally, field studies of exhume subduction-related shear zones depict a thick fault zone in which overall shearing is accommodated by foliated horizons that surround rock lenses with different mechanical properties.
This picture of fault zone structure, characterized by distributed deformation within a thick phyllosilicate-rich shear zone, differs from the typical geological interpretation of seismic faults, where deformation is localized within thin principal slip surfaces (Chester and Chester, 1998; Sibson, 2003). Instead, in weak fault zones the deformation is distributed within a thick shear zone composed of varied fault rocks with different frictional properties.

Here we propose a model for slip behavior of heterogeneous fault zones, like the Zuccale fault, that is consistent with our laboratory observations. In particular, the model is based on two end members: the weak foliated horizons (zones 1 and 2) and the strong competent lenses (zones 3 and 4). To mimic frictional sliding along the foliated and phyllosilicate-rich horizons, we used the frictional properties obtained from solid foliated wafers from zones 1 and 2. Because deformation in the competent lenses is characterized by slip on small faults affected by cataclastic processes (e.g. Fig. 2a, c and e), we use results from our experiments on powders from zones 3 and 4 to represent these. At the same time, it is worth noting that a contribution to deformation is likely to come from pressure solution creep (Bos and Spiers, 2002; Chester, 1995; Gratier et al., 2009; Niemeijer and Spiers, 2007; Niemeijer et al., 2009; Smith et al., 2011).

Our experiments have been conducted on dry samples and at room temperature: in the following we discuss the applicability of our results to fluid saturated faults at temperatures expected for the shallow crust. For the phyllosilicate-rich horizons, our experiments conducted on wet samples sheared in their in-situ geometry show that the presence of fluid weakens the fault rock: we observe a decrease in friction of about 0.2 for the same lithology, but the velocity-strengthening behavior does not change with the added fluid (Fig. 8). We note only a more pronounced evolution effect on fluid-saturated samples in comparison to dry ones (Fig. 8). Because recent experimental studies have shown that phyllosilicates like talc and smectite remain weak and velocity-strengthening with increasing temperature (Moore and Lockner, 2007, 2008), we infer that our foliated fault rocks, where slip is accommodated by the phyllosilicate foliation (e.g. Fig. 7a–d), will remain weak and velocity-strengthening with increasing temperature. For the strong competent lenses very few frictional data are available on powders composed of calcite, tremolite and phyllosilicates (zone 3) or amphibole and phyllosilicates (zone 4). Experiments conducted on water-saturated calcite powders, at lower displacements (<2.5 mm) and sliding velocities (0.12–1.22 μm/s) compared to our experiments, show that with increasing temperature calcite evolves from velocity strengthening between 25–50 °C, to velocity-weakening at 150 °C (Verberne et al., 2010). This suggests that, in addition to the increase in sliding velocity and shear localization as documented by our data, an increase in temperature can promote frictional instabilities in carbonate-rich rocks. Preliminary data on powders collected from the Zuccale fault with a mineralogical composition similar to zones 3 and 4 show that an increase in fluid pressure and temperature favor frictional instabilities (Niemeijer et al., 2011).

Following a discussion of the applicability of our laboratory results to fluid-rich faults in the shallow crust, we proceed with illustrating the details of the model. During tectonic loading, the shear stress rises until it reaches the frictional strength of the weak, foliated horizons (Fig. 9a), and the fault will begin to slip (Fig. 9b point 1). Because the weak regions exhibit velocity-strengthening frictional behavior (e.g. Figs. 9a and 5), the fault should undergo stable sliding and aseismic fault creep during this phase of slip. With continued shear, however, the tectonic stress may increase due to a combination of geometric factors (related to interaction between weak and strong regions within the fault zone) and variations in fluid pressure within the fault system (see discussion below). When the shear stress reaches the frictional strength of the strong lenses (Fig. 9b point 2),...
fractures and/or a frictional instability in the strong lenses may nucleate. The localization of deformation along discrete brittle surfaces observed along the strong lenses (Fig. 2) and the velocity-weakening frictional behavior of the material from the strong lenses (e.g. Figs. 9d and 5) suggest that fault slip may become unstable, leading to seismic rupture.

Fault creep within the phyllosilicate-rich horizons may also increase shear stress on the minor faults that merge into the base of the main fault zone (e.g. Fig. 1), which would tend to promote seismic rupture nucleation along these structures. Both the strong lenses and the minor normal faults may remain in broadly the same location for long time periods, extending over several seismic cycles. This could result in repeated rupture nucleation at a given position on the fault (e.g. repeating earthquakes, Chen et al., 2009; Chiaraluce et al., 2007; Waldhauser et al., 2004).

The Zuccale fault is an ancient and exhumed low-angle normal fault within the Apenninic extensional system. In the currently active area of the Apennines, the Altotiberina low-angle normal fault is accommodating...
extension by microseismicity and aseismic creep at rates of $\approx 2.5$ mm/yr (Chiaraluce et al., 2007; Hreinsdottir and Bennett, 2009). Along this fault, clusters of seismicity have been identified including small, repeating earthquakes with $M_L$ from 0.4 to 1.6 and rupture dimension of radius ~10–70 m (Chiaraluce et al., 2007). We note that Zuccale fault (Fig. 1) contains decameter-sized lenses of competent material, which is consistent with the dimensions of the repeating earthquakes recorded along the Altotiberina fault. Larger earthquakes could be induced by larger lenses not documented in the exposed outcrops, clusters of small competent lenses along the fault zone (e.g. Fagereng, 2011), and/or rupture propagation from the strong competent lenses into the foliated horizons which might become velocity-weakening at higher sliding velocities (Boutareaud et al., 2008; Niemeijer and Spiers, 2007).

Recently by using a dense seismic array in South African mines, it has been shown that mining-induced earthquakes follow the Gutenberg–Richter relation at moment magnitude $M_w = -1.3$ (Boettcher et al., 2009), implying a nucleation zone size for these earthquakes smaller than 25 cm (Boettcher et al., 2009; Fagereng, 2011). These nucleation zones, or asperities, are of the same length scale as numerous competent lenses documented by our field and laboratory data.

In the following, we discuss factors that might promote the episodic occurrence of failure within the strong and competent lenses. One possible cause of repeated failure is geometric factors that result from competency contrast between the strong lenses and the weak and phyllosilicate-rich horizons. Material heterogeneity results in differences in strain rate and finite strain magnitude between
competency domains that must be accommodated along the domain boundaries (e.g. Goodwin and Tikoff, 2002). These incompatibilities, if small, can be accommodated by domain-boundary deformation mechanisms, but if sufficiently large, they produce mechanical instabilities resulting in the nucleation of faults (Fagereng and Sibson, 2010; Goodwin and Tikoff, 2002). We start by considering a slip rate of 2.5 mm/yr, \( v \) (the slip rate of the Altotiberina fault measured by Hreinsdottir and Bennett, 2009) and noting that the deformation is accommodated by homogeneous simple shear along the weak and phyllosilicate-rich horizons (thickness, \( w \)). Then by considering the distribution of competent vs. incompetent material within the fault zone, it is possible to estimate the fluctuation in strain rate, \( \gamma = \frac{v}{w} \). For example, along transect A–A' (Fig. 1) the thickness of the weak material is 0.5 m and the resulting strain rate is \( \gamma = 1.6 \times 10^{-10} \text{ s}^{-1} \). Along transect B–B' there are two weak portions with a thickness of 2.5 m each, with \( \gamma = 1.6 \times 10^{-11} \text{ s}^{-1} \); lower strain rates can be related to thicker portions of phyllonites. In general, variable strain rates are strongly related to the amount of competent vs. incompetent materials within the shear zone (e.g. Fagereng and Sibson, 2010). The proposed analysis shows that the value of \( \gamma \) varies within heterogeneous fault zones like the Zuccale fault, suggesting that strain rate magnitudes can be sufficiently large to produce mechanical instabilities and faulting within the competent lenses.

Another cause that might trigger failure within the competent lenses is fluid pressure within the fault system. Here we investigate with Mohr circles the stress conditions required to induce slip within the weak phyllosilicate-rich horizons and reactivation of small faults within the strong lenses. We study this 2D mechanical problem (for a 3D analysis, see Collettini and Trippetta, 2007; Lisle and Srivastava, 2004; Morris et al., 1996), where the fault planes lie parallel to the \( \sigma_2 \) direction (e.g. Sibson, 1985), at 5 km of crustal depth, a representative exhumation depth of the investigated fault rocks (Collettini et al., 2009a). We assume an extensional environment characterized by a vertical, effective maximum principal stress, \( \sigma_1^e \) (Collettini and Holdsworth, 2004), and low differential stress \( (\sigma_1^e - \sigma_3^e) = 10 \text{ MPa} \). We note that similar low values of differential stress have been inferred for frictional reactivation of phyllosilicate-rich faults oriented at high angles to the maximum compressive stress (Fagereng et al., 2010). Our field data (e.g. Fig. 2a and Smith et al., 2008) constrain fault planes oriented at reaction angles, \( \theta_r \), to \( \sigma_1^e \) of about 70° and 45° for weak and strong fault portions respectively (Fig. 10a). The reactivation, or re-shear condition (e.g. Sibson, 1985), results in a friction coefficient of 0.25 and 0.62 for the weak and strong fault portions respectively (Fig. 10b). Under the boundary conditions described so far, during fault loading re-shear of the phyllosilicates occurs only for high values of fluid pressure, e.g. \( \lambda = 0.89 \) (Fig. 10b), where \( \lambda \) is the ratio of fluid pressure to lithostatic load. The widespread hydrofractures system documented with the Zuccale fault core is in agreement with fluid overpressure during fault activity (Fig. 2b and Collettini et al., 2006; Smith et al., 2008). During frictional reactivation of the phyllosilicates, velocity- and shear-strengthening (Fig. 10c) make the condition for reactivation more severe (e.g. \( \mu_s = 3.0 \), Fig. 10d); further reactivation can be achieved by either increases in fluid pressure (e.g. leftward shift of Mohr circle) or by a decrease of the minimum effective principal stress (Fig. 10d), that can eventually promote reactivation of faults in the stronger lenses (Fig. 10d). Several factors like crustal depth, differential stress, fluid pressure and cohesive strengthening of the fault zone (e.g. Fagereng et al., 2010; Muhuri et al., 2003; Streit and Cox, 2001; Tenthorey et al., 2003) influence the stress values of our mechanical analysis, however two main points can be highlighted. First, at crustal depth \( >4\text{–}5 \text{ km} \) the reactivation of a weak fault that is oriented at high angles to the maximum compressive stress can be achieved only by fluid overpressure. Second, both along well-oriented strong faults and misoriented weak faults (Fig. 10a), a small variation in differential stress or fluid pressure can trigger a switch from re-shear of strong to weak fault portions and vice versa (Fig. 10d).
Our results on the mixed-mode fault slip behavior are akin to findings from studies involving experiments on rock analogs (Niemeyer and Spies, 2005) and field studies on the fault zone structure (Fagereng and Sibson, 2010; Faulkner et al., 2003). Here, by integrating field data on the fault zone structure with frictional properties of the fault rocks sheared under their in situ geometry, we provide further constraints for the mixed-mode fault slip behavior of crustal faults. We suggest a better understanding of the complex fault slip behavior, documented in recent works using high-resolution seismological and GPS investigations, can be achieved by the integration of field studies of complex fault zone structures with the determination of the heterogeneous frictional properties of the fault rocks.

Acknowledgments

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Appendix A

Constitutive parameters for least-squares fit of the Dieterich Law. The standard deviation in $a$, $b$, and $a - b$ values is in the order of $10^{-4}$, the standard deviation in $d_c$ is generally less than 10% of $d_c$.

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