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Microstructural evidence for seismic and aseismic slip along clay-bearing, carbonate faults

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Key points

- Fault microstructures record fast (seismic) to slow (aseismic) deformation stages consistent with some geophysical observations
- A few % of clays in natural fault gouge may strongly affect frictional behavior
- Shear experiments suggest that clay-bearing layers have shorter weakening distance than carbonate-pure gouges during seismic slip.

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Abstract: In this multi-methodological study, microstructural observations of fault rocks are combined with micromechanical property analyses (Contact Resonance Atomic Force Microscopy, CR-AFM) and with rotary friction experiments (SHIVA apparatus) to find evidence of seismic to aseismic slip and understand the nanoscale rheology of clay-bearing, carbonate-hosted faults. Fluidized structures, truncated clasts, pores and vesicles, and phyllosilicate nano-sized spherules and tubes suggest fast deformation events occurred during seismic slip, whereas clay-assisted pressure-solution processes, clumped clasts, foliation surfaces, and mantled clasts indicate slow deformation events occurred during postseismic/interseismic periods. CR-AFM measurements show that the occurrence of ~5 wt.% of clay within the carbonate-hosted gouges can significantly reduce the fault core stiffness at nanoscale. In addition, during high-velocity friction experiments simulating seismic slip conditions, the presence of ultra-thin phyllosilicate-bearing (≤ 3 wt.%) layers within calcite gouges, as those observed in the natural fault, show faster dynamic weakening than that of pure calcite gouges. The weak behavior of such layers could facilitate the upward propagation of seismic slip during earthquakes, thus possibly enhancing surface faulting. Microstructural observations and experimental evidence fit some well-known geophysical and geodetic observations on the short- to long-term mechanical behavior of faults such as post-seismic/interseismic aseismic creep, interseismic fault locking, and seismic slip propagation up to the Earth's surface.

1. Introduction

Understanding the short and long-term mechanical behavior and deformation mechanisms of seismogenic faults is the prerequisite to properly model seismic cycles of active faults and eventually contribute to physically-based probabilistic earthquake forecasting [e.g., *Dieterich et al.*, 1994; *Hainzl et al.*, 2010]. Detailed fault mechanical

models require indeed, among other parameters, the complete understanding of rheology, frictional behaviors, fluid pressure fluctuations, and deformation mechanisms acting along faults during the seismic cycle [e.g., *Li et al.*, 2011; *Hadizadeh et al.*, 2012; *Ben-Zion and Sammis*, 2013; *Tesei et al.*, 2014].

Crustal deformation related to seismic cycles has been recorded and studied with paleoseismological, geophysical, and geodetic techniques (GPS and InSAR; e.g., *Harris*, 2017 and reference therein); however, the short-term records of instrumental seismology (less than one century) and geodesy (about 25 years), as well as the non-continuous record of paleoseismology have limited the long-term understanding of the seismic cycle. For these reasons, the direct study (*in situ*) of the geological record preserved within fault zones is still fundamental to understand the spectrum of deformation mechanisms and mechanical parameters that can be useful to better model the long-term behavior of active faults during the seismic cycle [e.g., *Siman-Tov et al.*, 2013; *Bullock et al.*, 2015; *Boulton et al.*, 2017]. Results from these studies are complementary to and consistent with some geophysical and geodetic observations [e.g., *Smith et al.*, 2011; *Fondriest et al.*, 2015; *Demurtas et al.*, 2016] and can be accomplished studying seismogenic faults through deep boreholes [e.g., *Solum et al.*, 2006; *Gratier et al.*, 2011] or through the analysis of exhumed faults [e.g., *Balsamo et al.*, 2013; *Tesei et al.*, 2014].

In this paper, we present a multi-methodological research starting from the natural case of the exposed extensional seismically-active Tre Monti Fault (TMF) within the carbonate-dominated seismic domain of the central Apennines, Italy [*Smith et al.*, 2011; *Benedetti et al.*, 2013; *Smeraglia et al.*, 2016a]. The TMF has been exhumed from shallow depths [≤ 3 km; *Smeraglia et al.*, 2016a] and can hence be considered as an exposed analogue for shallow seismogenic faults. In particular, we focus our study on the ~1 m thick clay bearing, carbonate-hosted Tre Monti principal fault core, which accommodated up to ~2 km of total

displacement [Smeraglia *et al.* 2016a]. To do so, microstructural, chemical, and mineralogical analyses with laboratory micromechanical analyses were combined with low- to high-velocity rotary shear experiments that simulate the natural setting under laboratory-controlled conditions [Di Toro *et al.*, 2010]. New results are integrated with those from previous studies on this fault [Smith *et al.*, 2011; Benedetti *et al.*, 2013; Smeraglia *et al.*, 2016a] to elaborate a synthetic model of deformation mechanisms acting during seismic cycles. Fault rock microstructures from this study recorded indeed evidence of both seismic and aseismic slip events. In particular, ≤ 5 wt.% of clay within a carbonate-hosted fault zone can strongly affect fault rheology at the nanoscale, thus influencing fault rock stiffness and deformation mechanisms during the co-seismic and postseismic/interseismic phases. These results fit some well-known geophysical and geodetic observations on the short- to long-term mechanical behavior of faults such as post-seismic/interseismic creep, interseismic fault locking, and seismic slip propagation up to the Earth's surface.

2. Geological setting

The Central Apennines are a Late Oligocene to Present fold-and-thrust belt related to the westward directed subduction and eastward rollback of the Adriatic plate under the European plate [e.g., Doglioni, 1991]. Shortening was characterized by the northeastward migration of NW-SE-oriented thrust faults, which scraped off and piled-up the sedimentary succession of the Adriatic plate [e.g., Cosentino *et al.*, 2010]. Thrust faults juxtaposed 4-5 km thick Late Triassic-Middle Miocene shallow-water carbonate rocks above ~2 km thick Late Miocene syn-orogenic deposits (hemipelagic marls and siliciclastic sandstones with clayey and marly interbeds; Fig. 1a; Cosentino *et al.*, 2010). Seismic reflection profiles show that imbricated carbonate deposits occur from the surface down to a depth of ~7-8 km [e.g., Billi *et al.*, 2006].

Since Early Pliocene time, the axial part of the Central Apennines belt experienced uplift and extensional tectonics in connection with the development of the Tyrrhenian backarc basin [Doglioni, 1991]. This process is still active and has generated intra-mountain grabens bounded by NW-SE-oriented extensional faults and less frequent ENE-WSW-oriented strike-to oblique-slip transfer faults (Fig. 1a). This fault network has produced large historical and instrumental earthquakes up to M_w 7 (Fig. 1a; e.g., Avezzano, 1915, M_w 7.0 earthquake, Galli *et al.*, 2008; L'Aquila, 2009, M_w 6.3 earthquake, Valoroso *et al.*, 2013; Amatrice, 2016, M_w 6.0 and Norcia, 2016, M_w 6.5 earthquakes, Anzidei and Pondrelli, 2016]. The TMF belongs to this active tectonic setting representing an extensional-to-transensional transfer fault between two NW-SE-oriented extensional faults [e.g., Morewood and Roberts, 2000].

Although in the Central Apennines several seismic mainshocks nucleated at depths < 10 km, probably in crystalline rocks, it is well documented that many foreshocks and aftershocks as well as some mainshocks nucleated in and propagated upward through the carbonate sedimentary succession [e.g., Valoroso *et al.*, 2013]. In particular, seismic slip often propagate up to the surface, causing ground displacements and surface faulting, as documented for the recent cases of the 2009 L'Aquila and 2016 Amatrice-Norcia earthquakes [e.g., Anzidei and Pondrelli, 2016; Pucci *et al.*, 2017] and as shown by paleoseismological observations [e.g., Galli *et al.*, 2008]. For instance, during the 2016 Amatrice-Norcia seismic sequence, two mainshocks (M_w 6.0 and M_w 6.5) nucleated at ~7-8 km depth and seismic slip propagated upward through carbonate rocks causing a ~15 km long and up to ~60 cm high surface rupture [Chiaraluce *et al.*, 2017]. Moreover, many aftershocks of the 2016 Amatrice-Norcia mainshocks nucleated at shallow depths (< 5 km) along carbonate-hosted faults and also the maximum slip patches (up to ~2.5 m) of the mainshocks were located at ~3-5 km depth [Anzidei and Pondrelli, 2016; Tinti *et al.*, 2016; Liu *et al.*, 2017] probably within carbonate rocks. This evidence suggests that is likely that the shallow fault zone modulated

the observed earthquake-related deformation processes and slip recorded at shallow depths and at the surface.

The TMF is a SSE-dipping and ~8 km long extensional/transtensional fault, which delimits the north-western boundary of the Fucino basin, an intramountain half-graben filled by Late Pliocene–Quaternary continental deposits [Fig. 1b; *Morewood and Roberts*, 2000; *Cavinato et al.*, 2002; *Smith et al.*, 2011; *Smeraglia et al.*, 2016a]. Paleoseismological investigations suggest that the TMF has been active between Early Pliocene and recent-present times [*Benedetti et al.*, 2013]. The TMF has been exhumed from depths ≤ 3 km accumulating up to ~2 km of cumulative displacement along the principal fault [*Cavinato et al.*, 2002; *Smith et al.*, 2011; *Smeraglia et al.*, 2016a]. This displacement is at least in part transferred from the NW-striking extensional Venere Fault that is longer (i.e., ~20 km) than the TMF [e.g., *Cavinato et al.*, 2002]. The principal surface of the TMF is exposed along the foothills of the Tre Monti Hills and juxtaposes carbonate deposits in the footwall with both syn-orogenic clay-rich deposits and post-orogenic continental sediments (i.e., lacustrine, fluvial, and slope debris deposits) in the hangingwall along a series of ENE-WSW-oriented fault scarps in Mesozoic limestones [Fig. 1c; *Benedetti et al.*, 2013; *Smeraglia et al.*, 2016a]. Synthetic secondary faults are hosted within the carbonate bedrock of the footwall and juxtapose Mesozoic carbonates and Plio-Quaternary continental deposits in the hangingwall with Mesozoic carbonates in the footwall [*Smith et al.*, 2011; *Smeraglia et al.*, 2016a].

Using carbon and oxygen stable isotopes, *Smeraglia et al.* [2016a] showed that the cataclastic rocks of the TMF principal fault developed at ~2-3 km depth. Therefore, in this work, the inferred deformation mechanisms acted along the TMF principal surface at depths ≤ 3 km.

3. Methods

(1) Rock slabs were cut perpendicular to the fault surface and parallel to the slip direction from oriented samples collected in the field. The slabs were then impregnated with a low-viscosity epoxy resin (PEC 1 Logitech) and 15 thin sections were produced by thinning the slabs down to $\sim 30\ \mu\text{m}$ in thickness using an Al-free abrasive paste and a diamond-bearing $1\ \mu\text{m}$ -thick powder for polishing. Al-free abrasive paste was used to avoid Al contamination and to better detect the presence of phyllosilicates.

(2) Optical and electron microscopy (FE-SEM Zeiss Auriga at the CNIS Laboratory, Sapienza University, Rome, and FE-SEM Zeiss Merlin at the CERTEMA Multidisciplinary Laboratory, Grosseto) was used to investigate micro- and nano-deformation mechanisms on twenty oriented thin sections of fault rocks.

(3) X-ray powder diffraction (XRPD, at the Padova University laboratory) and Energy Dispersive X-ray analysis (EDS, using FE-SEM Zeiss Auriga at the CNIS Laboratory, Sapienza University, Rome, and FE-SEM Zeiss Merlin at the CERTEMA Multidisciplinary Laboratory, Grosseto) were used to investigate both natural and experimental fault rock mineralogy and chemistry.

(4) Contact Resonance Atomic force measurements (CR-AFM; *Passeri et al.*, 2013; Text S1 of SI) on a $\sim 30\text{-}\mu\text{m}$ -thick thin-section were realized to measure in situ mechanical properties (Young's modulus and viscoelasticity) within natural samples (at the CNIS Laboratory, Sapienza University, Rome). For these measurements, one thin section was prepared without epoxy impregnation and was analyzed after removal of abrasive paste and powder to perform a realistic analysis of mechanical properties of rocks without contamination. In CR-AFM, the sample is coupled with a piezoelectric transducer, which excites the sample surface through out-of-plane oscillations. A cantilever records these oscillations through a tip that is in contact with the sample surface. The contact resonance

frequencies (CRFs) of the tip-sample system are measured and used to calculate the indentation modulus, which is proportional to the Young's modulus, of an elastic material [e.g., *Passeri et al.*, 2007; 2013]. In addition, if a material is viscoelastic, CR-AFM measurements can be used to calculate the elastic part (i.e., the storage modulus, E') and the viscous part (i.e., the loss modulus, E'') of the elastic modulus. The ratio between the elastic and viscous moduli E''/E' (namely the loss tangent, $\tan\delta$) quantifies the relative impact of the viscous behavior, which promote energy dissipation during mechanical deformation [e.g., *Meyers and Chawla*, 2009].

(5) Pulse-echo ultrasonic technique (at the CNIS Laboratory, Sapienza University, Rome; Text S2 of SI) to measure the elastic properties (i.e., Young's modulus and Poisson ratio) of host carbonate rock and to calibrate the CR-AFM measurements.

(6) Low to high velocity friction experiments using SHIVA (Slow- to High-Velocity rotary-shear friction Apparatus, at INGV HP-HT Laboratory, Rome; *Di Toro et al.*, 2010; *Smith et al.*, 2013; Text S3 of SI) to investigate frictional properties of natural fault rocks. During these experiments we recorded also gas emission through a mass spectrometer. Moreover, to quantitatively constrain the behavior of shear and normal stress, we performed a zero-lag cross-correlation analysis to quantify the correlation between the imposed normal stress and the recorded shear stress.

4. Field- to nano-scale observations on the Tre Monti principal fault

The architecture of the Tre Monti principal fault has been partially documented in previous studies [*Smith et al.*, 2011; *Smeraglia et al.*, 2016a]. In particular, *Smith et al.*, [2011] and *Smeraglia et al.*, [2016a] showed that the principal fault (Fig. 2a) accommodated tectonic deformation across a ~1 m thick fault core. Such fault core consists of two different fault rocks: ultracataclasite and foliated cataclasite [Figs. 2b and 2c; *Smeraglia et al.*, 2016a].

Smeraglia et al., [2016a] showed that the ultracataclasite consists of calcite (~98.5 wt.%) and phyllosilicates (~1.5 wt.%, mainly illite/smectite and kaolinite) and that the foliated cataclasite consists solely of calcite (Table S1). In addition, *Smeraglia et al.*, [2016b] showed that phyllosilicates between the ultracataclasite are mainly derived from incorporation, smearing and mixing along the principal fault of phyllosilicate-rich syn-orogenic deposits in the hangingwall (Fig. 1c).

Hereafter, we describe new field and microstructural observations (i.e., fluid-like structures, Figs. 2d and 2e; ultra-thin phyllosilicate-bearing layers, Figs. 2c and 2f; pressure-resolution features, Figs. 2h, 2i, and 3b; foliation surfaces Figs. 2g and 2i; mantled clasts, Fig. 2i; clumped clasts, Figs. 3a and 3g; pores and vesicles, Figs. 3c and 3h; and phyllosilicate nano-size spherules and tubes, Fig. 3i) within the Tre Monti principal fault (Fig. 1c) and then discuss such observations in terms of seismic/aseismic deformation mechanisms.

The ultracataclasite forms patches and plagues, up to ~10 cm thick, above the foliated cataclasite (Fig. 2b). The boundary between these two fault rocks is often sharp and continuous (Fig. 2c); however, in places, the two fault rocks are mixed together showing fluid-like structures at the centimeter scale (Figs. 2d and 2e). These structures show irregular/convolute margins with load-like or intruded/injected structures and bulbous protrusion of the ultracataclasite into the foliated cataclasite. Moreover, fragments from the foliated cataclasite are, in places, completely enclosed within the ultracataclasite (Fig. 2d). These fluidized structures develop at distances up to 10 cm from the principal slip surface (Figs. 2d and 2e). Similar fluid-like structures have been previously documented along the TMF only at the millimeter-scale [*Smith et al.*, 2011]. In this latter case, the micrometer-thick bulbous protrusions were intruded only for small distances (i.e., a few microns) between the different cataclastic layers [*Smith et al.*, 2011].

The foliated cataclasite (100 wt.% calcite) is characterized by calcite clasts derived from the host rock (Mesozoic limestones; Fig. 2g). The foliation is defined by slip-parallel shear surfaces bounding relative intact host rocks sigmoids (Fig. 2g). The ultracataclasite (~98.5 wt.% calcite and ~1.5 wt.% phyllosilicates) consists of large calcite clasts (centimeter to hundreds of micrometer in size, Figs. 2c, 2f, 2h, and 2i) derived from the underlying foliated cataclasite, scattered within an ultra-fine matrix (Fig. 3a). Clasts are sub-angular to well-rounded and, in places, are in reciprocal contact showing indented and teeth-shaped margins (Figs. 2h and 2i). Such margins do not show any preferred orientation with respect to the Tre Monti principal fault. The ultracataclasite is characterized, in places, by foliation surfaces roughly parallel to the slip direction (Fig. 2i). Within these foliation surfaces, large calcite clasts are characterized by mantled-like structures [e.g., Passchier and Trouw, 2005] showing elongate trails of smaller calcite clasts aggregates, stretched parallel to the foliation (Fig. 2i). These small calcite clasts are in reciprocal contact showing indented and teeth-shaped margins roughly parallel to the Tre Monti principal fault (Fig. 2i).

The ultra-fine matrix consists of micro- to nano-meter-thick sub-angular to well-rounded calcite clasts and phyllosilicates (~1.5 wt.% of the bulk ultracataclasite; Figs. 3a and 3b). Contacts between calcite clasts show indented and teeth-shaped geometries down to the microscale, defining a highly compact and low-porosity texture, also due to ultra-thin phyllosilicate lamellae (a few hundreds of nm thick) often interposed between clast margins (Fig. 3b). Such teeth-shaped margins do not show any preferred orientation with respect to the Tre Monti principal fault. In places, clumps of small calcite clasts are compacted together to form larger and distinct clasts (Figs. 3a and 3b). In places, calcite clasts are affected by microdefects such as twinning and microfractures (Fig. 3c).

Field and microstructural observations show that phyllosilicates concentrate within ultra-thin (up to ~1 mm-thick) and localized layers (hereafter “phyllosilicate-bearing layers”,

Figs. 2c, 2f, and 3d-f). Quantitative XRPD analyses (sampling sites in Fig. 2c) show that the average phyllosilicate content within these layers is ~5 wt.% (Table S1 in SI), thus higher than the phyllosilicate content within the ultracataclasite (~1.5 wt.%). Phyllosilicate-bearing layers occur along the sharp boundaries between the ultracataclasite and the foliated cataclasite (Figs. 2c, 3e, and 3f), along some riedel-shear surfaces cutting across the ultracataclasite (Fig. 3d) or, in places, cutting across fluid-like structures (Fig. 2f), delineating, at least in the study outcrop, a pervasive network within the Tre Monti principal fault-zone.

Phyllosilicate-bearing layer margins are sharp and continuous, truncating calcite clasts (Figs. 3d-f). In these layers, carbonate clasts are well rounded and scattered within a phyllosilicate-bearing matrix. Calcite clasts that are in reciprocal contact show indented and teeth-shaped margins (Fig. 3g) and, in places, form clumped clasts (Fig. 3g). Calcite clasts show pores and vesicles (dark rounded features) arranged, in places, in linear arrays probably along clast microdefects such as twinning and cleavage planes (Fig. 3h). Down to the nanoscale, phyllosilicates are characterized by fibrous or massive morphologies, which, in detail, consist of 50 nm-thick clumped and chained nano-sized spherules and tubes (Fig. 3i). We found these peculiar phyllosilicate nanostructures only within the phyllosilicate-bearing layers.

5. Micromasurements of elastic property

Measurements of elastic property (i.e., Young's modulus) of fault rocks are important to understand fault rock stiffness, which can influence fault rock deformation mechanisms and the capability of a rock to store or release the energy during seismic cycles [e.g., Gudmundsson 2004; Fondriest *et al.*, 2017]. Moreover, recent studies showed that rheological heterogeneity and fault morphology, at nanoscale, could strongly affect the bulk fault zone rheology and nanoscale frictional processes [e.g., Chen *et al.*, 2013; Omori *et al.*,

2015; Gerbi *et al.*, 2016]. In this section, we provide micromasurements of elastic property (i.e., indentation modulus and viscoelasticity) and then we discuss how such measurements can impact fault rock stiffness at the nanoscale and associated deformation mechanisms.

Figures 4a and 5a are topographic CR-AFM maps of two areas ($16\ \mu\text{m}^2$ and $\sim 2\ \mu\text{m}^2$, respectively) within the phyllosilicate-bearing layer (Fig. 3g). In these maps, several calcite clasts (few microns to some tens of nanometers in size) are scattered within a phyllosilicate-bearing matrix. Phyllosilicates are characterized by first contact resonance frequency (f_1) and by first quality factor (Q_1) values lower than those of calcite clasts (Figs. 4b, 4c, 5b, and 5c). This setting is confirmed by second contact resonance frequency (f_2) and by second quality factor (Q_2) values (Figs. 4d, 4e, 5d and 5e). From these values, the indentation modulus, which is proportional to the Young's modulus of the material (see Text S1 in SI), was calculated.

Quantitative maps of indentation modulus (i.e., Young's modulus; Figs. 4f, and 5f) at ambient pressure show that calcite clasts and phyllosilicates are characterized by indentation moduli of 49 ± 8 GPa and 7 ± 3 GPa, respectively. The indentation modulus of phyllosilicates is consistent with that previously calculated with Atomic Force Acoustic Microscopy [Prasad *et al.*, 2002]. On the contrary, the indentation modulus of calcite clast is smaller than Young's modulus of the TMF carbonate host rock (80 ± 5 GPa; Text S2 in SI) and of other intact carbonate rocks (75 ± 7 GPa; e.g., Ciccotti *et al.*, 2004). Any effect of phyllosilicates (i.e., wrapping the calcite clasts) in lowering the indentation modulus of calcite clasts is excluded as the indentation modulus of clasts has been measured within clasts larger than the CR-AFM tip in order to avoid potential buffering effect of phyllosilicates.

Calcite clasts are characterized by loss tangent and loss modulus values of 0.01 ± 0.002 and 450 ± 150 MPa, respectively (Figs. 4g, 4h, 5g, and 5h). The same maps show that phyllosilicates are characterized by a loss tangent and loss modulus values of 0.023 ± 0.004

and 160 ± 30 MPa, respectively. In particular, the higher is the loss tangent value the more viscous is the sample behavior. Therefore, phyllosilicates are characterized by a more viscous behavior (i.e., capacity to dissipate energy during deformation; e.g., *Meyers and Chawla*, 2009) than calcite clasts, which are characterized by a more elastic behavior (i.e., capacity to store energy during deformation; e.g., *Meyers and Chawla*, 2009).

Overall, the analyzed areas from the phyllosilicate-bearing layers can be considered as a two-phase composite, within which phyllosilicates roughly represent ~50% of the total volume (Figs. 4a and 5a). Thus, the effective composite modulus (i.e., indentation modulus of phyllosilicates together with calcite clasts) of these areas calculated using the Voight Model [e.g., *Hua et al.*, 2010] is about 28 GPa.

6. Low to high velocity friction experiments

6.1. Experimental procedure

Field and microstructural observations show the occurrence of low percentage (~5 wt.%) of phyllosilicates concentrated along ultra-thin and discrete layers. Therefore, rotary-shear experiments reproducing the natural setting observed in nature were conducted to understand the frictional behavior and unravel the origin of such layers within calcite-bearing fault rocks.

The starting gouge material was made from rocks collected in the field that were crushed and sieved ($<125 \mu\text{m}$). Pure calcite powder was made from foliated cataclasite (100% calcite, Table S1 in SI), and phyllosilicate-rich powder from the syn-orogenic deposits (~38% of phyllosilicates content; Table S1 in SI). 4.5 g of gouge were used for each experiment creating ring-shaped layers with 35/55 mm external/internal diameter and initial thicknesses of ~3 mm (Fig. 6b; e.g., *Smith et al.*, 2013).

Starting gouge material was prepared using natural rocks collected in the field. The foliated cataclasite (100% calcite, Table S1 in SI) and the syn-orogenic deposits (~38% of phyllosilicates content; Table S1 in SI) were both crushed and sieved ($<125\ \mu\text{m}$) to obtain a pure calcite powder and a phyllosilicate-rich powder, respectively. For each experiment was used 4.5 g of gouge and was created a ring-shaped gouge layer with initial thickness of ~3 mm (Fig. 6b; e.g., *Smith et al.*, 2013).

Experiments were carried out using three different gouge settings (Fig. 6b). Mixed gouge (Fig. 6b, left panel) consists of 10% of phyllosilicate-rich powder mixed with 90% of pure calcite powder (total phyllosilicates content of $\leq 3\ \text{wt.}\%$). Layered gouge (Fig. 6b, middle panel) consists of a ~1 mm-thick layer made of 10% of phyllosilicate-rich powder mixed with 90% of calcite powder overlain by a ~2 mm-thick layer of pure calcite powder. This configuration simulates the ultra-thin phyllosilicate-bearing layers observed along the TMF (Figs. 2c, 2f, 3f, and 3g). Calcite gouge (Fig. 6b, right panel) consists of pure calcite powder used as an end member and representing the bulk composition of the TMF.

Experiments were performed at $0.001\ \text{ms}^{-1}$ (i.e., almost subseismic slip velocity) and $1\ \text{ms}^{-1}$ (i.e., typical average seismic slip velocity, *Heaton*, 1990) slip-rates under water-damped (hereafter wet) and room-humidity (hereafter dry) conditions. Water-damped gouges were saturated with 0.5 ml (i.e., 10 wt.%) of distilled water. Gouges were sheared holding the normal stress constant at 8.5 MPa and 5 MPa for dry and wet experiments, respectively, for a total displacement of 0.5 meter. For the high-velocity experiments, these conditions crudely represent the expected seismic slip occurring along a shallow-seated (depth $<0.5\ \text{km}$) fault patches during the M_w 6.0-6.5 earthquakes [e.g., *Anzidei and Pondrelli*, 2016].

The experimental procedure consists of the following steps: (1) samples were initially pre-sheared at $1\ \text{cm s}^{-1}$ for 10 cm of slip at 1 MPa of normal load, both in dry and in wet conditions. Gouges were pre-sheared to properly simulate our natural analogues, within

which the fault rocks are naturally pre-sheared due to fault displacement accumulation; (2) samples were loaded to the target normal stress and were then sheared with acceleration and deceleration rates of 24 ms^{-2} , up to the total displacement; (3) during the experiment, gas emission was measured with a Pfeiffer Quadstar quadrupole mass spectrometer (see Text S3 in SI); (4) after experiments, compact fragments of sheared samples were collected and impregnated under vacuum using low-viscosity epoxy. The impregnated gouges were cut perpendicularly to the surface and approximately parallel to slip direction. Then, petrographic thin sections $\sim 30 \text{ }\mu\text{m}$ thick were prepared for microstructural observations.

6.2. Results

At 0.001 ms^{-1} , both in dry and in wet conditions, the effective friction coefficient (i.e., note that, in the wet experiments, the pore fluid pressure cannot be measured; hence the term "effective"; e.g., *Chen et al.*, 2017) achieves a peak of ~ 0.6 at slip initiation and then increases almost monotonically towards a constant value of 0.7-0.8 that is eventually achieved after $\sim 0.5 \text{ m}$ of slip (Fig. 6c). At seismic slip rates (1 ms^{-1}), gouges in the three configurations (both in dry and in wet conditions) show a strong dynamic weakening (i.e., friction drop) after a strengthening phase, which is defined here as the distance to the onset of dynamic weakening, D_{ow} (Figs. 6d and 6e; Smith et al., 2015). Overall, both in dry and in wet conditions, a general trend was recognized: layered gouges weaken faster than mixed gouges, which, in turn, weaken faster than calcite pure gouges (Fig. 6d).

To test the reproducibility of this trend, high velocity experiments (with layered, mixed, and calcite gouges) were replicated both in dry and in wet conditions. The major relevant difference from the previous experiments is observed in the weakening stage of wet experiments, which can suffer from slightly different ambient conditions, sample saturation, and holding time. However, despite the large heterogeneity of natural materials used for

gouges, the replicated experiments are in good agreement with the previously observed mechanical trends: layered gouges weaken faster than mixed gouges, which, in turn, weaken faster than calcite pure gouges (Fig. 6e).

During high velocity dry experiments, a normal stress increase was observed in correspondence with the shear stress increase (i.e., during the strengthening phase; Fig. 7a), showing a positive correlation with zero-lag cross-correlation analyses (Fig. 7e, S1, and S2 in SI). Such stress increase can be considered as an apparatus artifact (i.e., due to a slightly misalignment between the rotary and the stationary columns), unrelated to gouge behavior.

In contrast, high velocity wet experiments displayed an increase in the normal stress in correspondence with the shear stress decrease (i.e., during the weakening phase; Fig 7b). This negative correlation in the zero-lag cross-correlation analyses (Fig. 7f, S1, and S2 in SI) is relevant for the experiment mechanics and is further explained in the discussion section.

Low velocity experiments under dry conditions are characterized by normal stress oscillations in phase with shear stress oscillations (Fig. 8a), whereas low velocity experiments under wet conditions show a constant normal stress (Fig. 8b).

Significant gas emissions (CO_2) were detected only during high velocity experiments under dry conditions (compare Fig. 7c with Fig 7d) and in correspondence with the weakening phase (Fig. 7c). Instead, gas emissions were not detected (i.e., or gas emissions were below the instrumental detection limit) in the low velocity experiments performed both under dry and under wet conditions (Figs. 8c and 8d).

6.3. *Experimental microstructures*

Overall, gouge microstructures are characterized by grain comminution towards the principal slip surface regardless of the gouge configuration (i.e., layered and mixed), slip velocity, and saturation conditions (Figs. 9a-h). However, the $\sim 200\text{ }\mu\text{m}$ -thick principal slip

zone of gouges from high velocity dry experiments is characterized by a compact and welded groundmass (Figs. 9e and 9f). EDS analyses revealed that this groundmass consists of a barely resolvable (or unresolvable) mixture of phyllosilicates and calcite (Figs. 9e and 9f). At the nano-scale, the groundmass texture shows slip-parallel ~50-nm-thick fibers composed by clumped nano-sized spherules and tubes (Figs. 9e and 9f). In places, individual nanospherules clump together to form granular zones (Figs. 9e and 9f). These nanostructures are very similar to phyllosilicate nanostructures observed within natural phyllosilicate-bearing layers along the TMF (Fig. 3i). These nano-sized spherules and tubes are absent in the principal slip zones of gouges from low velocity experiments, both in wet and in dry conditions (Figs. 9a and 9b).

Element distribution maps (from EDS analyses) show that gouges from low velocity experiments are characterized by mixed distribution of phyllosilicate and calcite particles, regardless of the gouge configuration (i.e., layered and mixed) and saturation conditions (Figs. 9a-d). Instead, gouges from high velocity experiments both under dry and under wet conditions are characterized by phyllosilicate segregation and concentration along slip-parallel micrometer-thick layers (Figs. 9e-h). These layers consist of calcite clasts scattered within phyllosilicates, with thin clay lamellae wrapping calcite clasts, and show microstructures similar to those observed within phyllosilicate-bearing layers along the TMF (Fig. 3h).

7. Discussion

7.1. Synthetic model of the seismic cycle

Co-seismic phase. Fluid-like structures (Figs. 2d-f), truncated clasts (Figs. 3d-f), clay nano-sized spherules and tubes (Fig. 3i), pores and vesicles (Fig. 3h), and phyllosilicate-bearing layers (Figs. 2c, 2f, 3d, and 3e) are interpreted as indicators of deformation at high

strain-rate and high temperature owing to seismic slip occurring during co-seismic phase (Fig. 10a).

(1) Fluid-like structures (Figs. 2d-f) suggest that both the ultracataclasite and the foliated cataclasite behaved, in part, as a mobile, cohesionless, and fluidized granular material. Such fluid structures have been previously documented within the TMF [Smith *et al.*, 2011; Smeraglia *et al.*, 2016a] and in other carbonate-hosted faults [e.g., Fondriest *et al.*, 2012; Demurtas *et al.*, 2016] and are interpreted as the result of fast and instantaneous fluidization of granular materials triggered by pressurized fluids trapped or injected within the fault gouge [e.g., Fondriest *et al.*, 2012; Rowe *et al.*, 2012] during seismic slip. Therefore, we suggest that fluid-like structures within the TMF are consistent with seismic-related fluid thermal pressurization [e.g., Rice, 2006; Chen *et al.*, 2017].

Overpressured fluids may have occurred within water-saturated fault gouges as suggested by experimental data. High velocity wet experiments are characterized by negative correlation of normal/shear stress (i.e., increase in normal stress during coeval decrease of shear stress) during the weakening phase. The increase in normal stress would result from the contribution of the normal stress applied by the piston, on one side of the load cell, and from a spurious pressure occurring within the gouge and exerted on the other side of the load cell (Fig. 6a; i.e., contrasting the normal stress applied on gouge; see discussion in Ferri *et al.*, 2010). We propose that such spurious pressure can be related to fluid (i.e., water) pressure increase due to water thermal expansion [e.g., Wibberley and Shimamoto, 2005; Chen *et al.*, 2017] or to impulsive and confined fluid pressurization related to rapid undrained gouge compaction [e.g., Kameda *et al.*, 2010]. Fluid pressure increase would indeed expand the slipping zone generating gouge dilatancy. However, in our high-velocity experiments, both in dry and in wet conditions, we did not record any gouge dilatancy (Fig. S3 in SI). This

evidence can be explained by the occurrence of a slight gouge extrusion during experiments (i.e., and consequently gouge compaction), thus buffering the gouge dilatancy.

(2) Truncated clasts along ultra-thin phyllosilicate-bearing layer margins (Figs. 3d-f) suggest high strain-rate and fast (i.e., seismic) slip localization along layer margins. The presence of truncated clasts, to our knowledge, are indeed a peculiar microstructure observed along sharp slip surfaces only during experiments conducted at seismic slip rates [$\sim 1 \text{ ms}^{-1}$; *Fondriest et al.*, 2013] and are not observed within experiments conducted at sub-seismic slip rates, at least in carbonate-hosted gouges [e.g., *Fondriest et al.*, 2015].

(3) Separation of phyllosilicates from the calcite and their segregation into distinct layers formed only during high velocity experiments both under dry and under wet conditions (Figs. 9e-h). Moreover, microstructures of experimental phyllosilicate-bearing layers (Figs. 9e-h) are very similar to natural phyllosilicate-bearing layers within the TMF (Fig. 3g). This evidence suggests that the natural phyllosilicate-bearing layers observed in nature probably formed at seismic slip rates. Phyllosilicate-bearing layers cutting through fluidized structures (Fig. 2f) and ultracataclasite (Fig. 3d) indicate multiple events of cataclasis, fluidization, and slip localization affecting, in alternating moments, the TMF principal surface during subsequent seismic cycles.

(4) Pores and vesicles found in calcite clasts within phyllosilicate-bearing layers (Fig. 3h) suggest the occurrence of thermo-chemical processes, such as calcite decarbonation, activated at high friction temperatures during seismic slip [e.g., *De Paola et al.*, 2011]. In fact, these features are similar to those documented within other carbonate-bearing fault [e.g., *Bullock et al.*, 2014; *Collettini et al.*, 2013, 2014] and during high velocity rotary shear experiments on carbonate-bearing gouges [e.g., *Violay et al.*, 2013; *Bullock et al.*, 2015] and interpreted as the results of calcite thermal decomposition at temperature above 600°C , developed during seismic slip [e.g., *Collettini et al.*, 2013].

(5) Nano-sized spherules and tubes observed within the TMF are similar to those observed only within the principal slip zone of other phyllosilicate-rich seismogenic faults (i.e., the Chelungpu Fault, Chi-Chi earthquake, *Hirono et al.*, 2014, Fig. S4c in SI) and differ markedly from the typical nanostructures of detrital phyllosilicates (e.g., *Setti et al.*, 2009; Fig. S4f in SI). In addition, nano-sized spherules and tubes are similar to those observed within phyllosilicates affected by hydrothermal processes at high temperatures (at ~200-250 °C, *Fiore et al.*, 1995; *Setti et al.*, 2009; Figs. S4d and S4e in SI) and to those observed within phyllosilicate-bearing gouges after experimental deformation at seismic slip velocities ($\geq 1 \text{ ms}^{-1}$) and at high temperatures (up to ~250-570 °C, *Ujiie et al.*, 2011; Fig. S4b in SI). In this study, nano-sized spherules and tubes along the TMF (Figs. 3i and S4a in SI) are similar to those developed during high velocity experiments only in dry conditions (Figs. 9e-f). In particular, during high velocity experiments under dry condition, CO₂ emission (Fig. 7b) occurred. CO₂ emission can be related to high friction temperatures that activated thermochemical processes within the gouge, such as calcite decarbonation [e.g., *De Paola et al.*, 2011]. Moreover, nano-sized spherules and tubes along the TMF occur together with calcite clasts characterized by pores and vesicles, which suggest high temperatures developed during seismic slip (i.e., see the above point 4).

These evidence suggest that the observed nano-sized spherules and tubes within the TMF probably developed by both seismic shear heating [e.g., *Rice*, 2006] and grain comminution [e.g., *Hirono et al.*, 2014] associated with seismic slip along the phyllosilicate-bearing layers.

Postseismic/interseismic phase. Teeth-shaped clasts margins (Figs. 3a-c and 3g), clumped clasts (Figs. 3a, 3b, and 3g), foliation surfaces (Figs. 2g and 2i), and mantled clasts (Fig. 2i) imply deformation at low strain-rate owing to aseismic slip occurring during postseismic/interseismic phases (Fig. 10b).

(1) Teeth-shaped clast margins are identical to those produced by intergranular pressure solution of carbonate grains during slow compaction experiments [e.g., *Zhang et al.*, 2010]. Moreover, intergranular pressure solution has been widely documented within other carbonate-hosted faults exhumed from depths < 3 km [e.g., *Billi*, 2003; *Agosta and Aydin* 2006] and interpreted as the product of deformation at low strain-rate occurring during postseismic/interseismic phases [e.g. *Renard et al.*, 2000; *Gratier et al.*, 2014]. Pressure solution processes may be enhanced by the occurrence of phyllosilicates between clast contacts (Fig. 3b; e.g., *Renard et al.*, 2001; *Aharonov and Katsman*, 2009).

(2) We propose that foliated cataclasite (Fig. 2g) and foliation surfaces within the ultracataclasite (Fig. 2i) were generated through slow and permanent (i.e., aseismic) fault creep along the TMF during postseismic/interseismic phases. Slow and permanent fault creep is also consistent with mantled structures of calcite clasts (Fig. 2i), which are similar to those observed in mylonitic rocks deformed at low strain-rate within the lower crust [e.g., *Passchier and Trouw*, 2005]. In particular, mantled structures roughly parallel to the Tre Monti principal fault and teeth-shaped margins within trails of small calcite clasts aggregates (Fig. 2i) are consistent with the interplay of fault creep along the Tre Monti principal fault contemporaneously with pressure solution processes acting perpendicularly to the normal stress applied on the fault surface. The above inferences are consistent with geological evidence and laboratory investigations showing that foliated fault rocks typically develop along faults through slow and stable slip [i.e., aseismic creep; e.g., *Collettini et al.*, 2009; *Tesei et al.*, 2014]. However, recent experimental investigations showed that foliated carbonate fault rocks could form also at seismic slip rates [*Smith et al.*, 2016]. In this case, foliation surfaces are often associated with sharp slip zones. In our case, in contrast, the lack of any sharp slip surface along or near the foliation surfaces (Fig. 2g) or mantled clasts (Fig. 2i) suggests aseismic slip along such features.

7.2 Implications

Microstructural evidence of seismic and aseismic microstructures together with the support of mechanical data and microstructures fits several geophysical and geodetic observations.

(1) The occurrence of fluidized fault rock textures preserved along the TMF (Figs. 2d-e), probably testifying for episodic and impulsive fault rock fluidization during seismic slip, is consistent with the evidence of fluid overpressures at hypocentral depths in the Italian Central Apennines based on focal mechanism tomography [Terakawa *et al.*, 2010]. In fact, seismic slip under these fluid-rich conditions would result in the possible activation of seismic-related fluid thermal pressurization [e.g., Rice, 2006; Chen *et al.*, 2017].

(2) Slow and continuous deformation along foliation surfaces (Fig. 2g) and mantled clasts (Fig. 2i) can promote tectonic elastic strain energy dissipation during interseismic phases through permanent aseismic creep [e.g., Collettini *et al.*, 2009]. In addition, CR-AFM measurements show that phyllosilicates behave more viscously (i.e., higher loss tangent and lower loss modulus values, Figs. 4 and 5) than calcite clasts. The viscous behavior of phyllosilicates within the phyllosilicate-bearing layers can enhance energy dissipation during inelastic mechanical deformation [e.g., Meyers and Chawla, 2009] within the interseismic phase. On the contrary, clay-assisted intergranular pressure solution processes, such as those observed within the ultracataclasite (Figs. 2h, 3a, and 3b), can promote transient aseismic creep during the postseismic phase [e.g., Gratier *et al.*, 2014] and fault healing during the interseismic phase (i.e., by dissolution contacts flattening and gouge compaction/cementation; e.g., Renard *et al.*, 2000). In particular, fault healing is responsible for strength recovery, fault locking, and tectonic elastic strain energy accumulation.

These deformation mechanisms are consistent with geodetic observations suggesting that active faults can move slowly (i.e., dissipating elastic strain energy) through dominant aseismic creep during the postseismic/interseismic phase [e.g., *Wilkinson et al.*, 2010; *Harris et al.*, 2017 and reference therein] or can be locked (i.e., storing elastic strain energy) between two earthquakes [e.g., *Johanson and Burgman*, 2005; *Harris et al.*, 2017]. Moreover, the interplay between frictional sliding through foliation surfaces and intergranular pressure solution processes occurring within the same fault segment, as observed along the TMF fault core, is consistent with geodetic observations of both locked and creeping portions being present along the same fault segment [e.g., *Johanson and Burgman*, 2005; *Tong et al.*, 2013; *Harris et al.*, 2017].

Although active especially within principal fault cores, the above-mentioned frictional-viscous deformation mechanisms can occur also in the damage zones of carbonate-hosted faults [e.g., *Salvini et al.*, 1999; *Agosta and Aydin*, 2006; *Hausegger et al.*, 2010]. Therefore, further contribution to the total displacements observed by geodesy due to the same deformation mechanisms but acting within the damage zone cannot be excluded.

(3) Our experimental results show that calcite gouges with ≤ 3 wt.% of clay content localized in ultra-thin layers weaken faster than mixed and calcite gouges. Therefore, we propose that the occurrence of such peculiar phyllosilicate-bearing layers within calcite gouges, as observed in nature (Figs. 2c, 2f, and 3d-f), could facilitate seismic slip propagation through the shallow part of fault zones (depths < 0.5 km; Figs. 6d-e; *Smeraglia et al.*, 2017). This inference is consistent with paleoseismological and direct observations in the Italian Central Apennines, showing that moderate to large earthquakes can easily propagate up to the Earth's surface causing surface faulting [e.g., *Galli et al.*, 2008; *Anzidei and Pondrelli*, 2016].

(4) The occurrence of ~ 5 wt.% of clay content, as observed within ultra-thin phyllosilicate-bearing layers (Table S1 in SI), is sufficient to strongly reduce the fault rock

stiffness at the nanoscale. The indentation moduli (i.e., Young's modulus) of phyllosilicates (7 ± 3 GPa) and of the whole phyllosilicate-bearing layers (28 GPa) are indeed far lower than Young's modulus of the comminuted calcite clasts (49 ± 8 GPa) and of the carbonate host rock (80 ± 5 GPa). The different indentation moduli of calcite clasts and intact calcite host rock can be due to the development of very comminuted clasts characterized by microfractures, junctions between clumped clasts (Figs. 3a, 3b, 3g, and 5), and/or by nanodefects (i.e., nanoscale grain/crystal boundaries, pores/vesicles, twinning and cleavage planes, Figs. 3c and 3h) developed during faulting [e.g., *Siman-Tov et al.*, 2013; *Collettini et al.*, 2014]. This is consistent with CR-AFM measurements showing softening of nanoscale mechanical properties in presence of nano-size grain boundaries, fractures, dislocations, and voids [e.g., *Yamanaka*, 1996; *Parlak et al.* 2008; *Hurley et al.*, 2006; *Passeri et al.*, 2016] and with measurements on macroscopic samples of brecciated and fractured fault rocks, which are characterized by Young's moduli smaller than intact rocks [e.g., *Kahraman and Alber*, 2006].

The strong decrease of the bulk fault rock stiffness (down to 28 GPa) due to only ~5 wt.% of clay content indicates the occurrence of a more compliant fault core that can facilitate tectonic elastic strain energy dissipation due the onset of ductile and permanent deformation within phyllosilicates [i.e., due to their viscous behavior; e.g., *Meyers and Chawla*, 2009] preventing elastic strain energy storage.

8. Conclusions

The study of the clay-bearing, carbonate-hosted Tre Monti Fault, exhumed from depths < 3 km, allowed us to:

- (1) Document microstructures developed during different stages of the seismic cycle. In particular, fluidized fault rocks, truncated calcite clasts, pores/vesicles in calcite clasts, and ultra-thin phyllosilicate-bearing layers are interpreted as the result of seismic deformation. On

the contrary, clay assisted intergranular pressure solution processes and mantled clasts/foliation surfaces are interpreted as the result of slow transient creep and permanent creep, respectively, during postseismic/interseismic phases;

(2) Link microstructures and micromechanical properties to geophysical and geodetic observations. In particular, fluidized fault rocks are consistent with seismic-related fluid thermal pressurization. Clay content ≤ 3 wt.% (i.e., as that used in some of our high velocity rotary shear experiments) localized along ultra-thin layers could facilitate earthquake slip propagation up to the Earth's surface, consistently with surface displacement observations from both direct and paleoseismological studies. Clay assisted intergranular pressure solution and foliation surfaces can contribute to fault locking/re-strengthening and aseismic sliding, respectively, consistently with geodetic observations on active faults showing locked patches and/or transient/permanent aseismic creep along fault segments;

(3) Constrain fault rock stiffness heterogeneity at nanoscale and unravel the contrasting mechanical behavior of phyllosilicates. In particular, phyllosilicates can facilitate both (a) aseismic creep by dissipation of elastic strain energy and/or fault healing and re-strengthening by enhancing pressure solution processes within calcite clasts during the post/seismic interseismic phase and (b) seismic slip propagation if sheared at seismic slip velocity.

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Figure captions

Figure 1. Geological setting and structure of the Tre Monti Fault. **(a)** Geological map of the central Apennines, Italy, showing active faults. Stars indicate the epicenters of the 1915, M_w 7.0, Avezzano, 2009, M_w 6.3, L'Aquila, and 2016, M_w 6.0, Amatrice earthquakes. **(b)** Geological map of the northern boundary of the Fucino basin, including the Tre Monti Fault (TMF) that is studied in this paper. **(c)** Geological map of the TMF showing principal and secondary faults and sampling location ($42^\circ 4'33''N$, $13^\circ 29'57''E$ from Google Earth). Simplified cross-section (inset) through the TMF, showing the principal fault outcrop.

Figure 2. Rocks of the Tre Monti fault. **(a)** Overview of the Tre Monti principal fault, which juxtaposes, in the study outcrop, Plio-Quaternary continental deposits in the hangingwall with carbonate deposits in the footwall. **(b)** Detail of the principal fault surface showing patches and plagues of ultracataclasite (with 1.5 wt.% of phyllosilicates) above the foliated cataclasite (100 wt.% calcite). **(c)** High resolution scan of sample cut perpendicular to the principal fault surface, showing the sharp contact between the ultracataclasite and the foliated cataclasite marked by the phyllosilicate-bearing layers laying between the two fault rocks. **(d, e)** High resolution scans of samples cut perpendicular to the principal fault surface,

showing fluid-like structures characterized by irregular margins and bulbous/convolute structures with injection of the ultracataclasite into the foliated cataclasite. **(f)** High resolution scan of sample cut perpendicular to the principal fault surface showing a fluid-like injection of the ultracataclasite into the foliated cataclasite cut by the phyllosilicate-bearing layer. **(g)** Texture of the foliated cataclasite showing foliation surfaces defined by cataclastic shear zones encasing host rock sigmoids. **(h)** Texture of the ultracataclasite showing calcite clasts scattered within an ultra-fine reddish phyllosilicate-bearing matrix. Clasts that are in contact with one another show indented and teeth-shaped margins. **(i)** Texture of the ultracataclasite showing foliation surfaces defined by shear zones rich in ultra-fine reddish phyllosilicate-bearing matrix. Large clasts in these foliation surfaces are characterized by mantled structures showing an elongate trail of smaller calcite clasts.

Figure 3. Microphotographs of thin sections showing representative fault rock textures along the TMF under optical and field emission scanning electron microscope. **(a, b)** FESEM images of the texture of the ultra-fine matrix from the ultracataclasite. Calcite clasts that are in contact with one another show indented and teeth-shaped margins. In places, an ultra-thin phyllosilicate film is interposed between clasts. Small calcite clasts are welded together to form larger clasts. **(c)** Detail of calcite clasts within the ultra-fine matrix showing serrated grain boundaries, twinning and nanofractures. **(d)** High resolution scan of a sample cut perpendicular to the TMF, showing phyllosilicate-bearing layers along a riedel shear surfaces

and the boundary between the ultracataclasite and the foliated cataclasite. Inset shows a FESEM image of the sharp boundary with truncated clasts between the phyllosilicate-bearing layer (i.e., along the riedel shear surfaces) and the ultracataclasite. **(e, f)** FESEM image of the phyllosilicate-bearing layer between the ultracataclasite and the foliated cataclasite showing a sharp boundary, which truncates calcite clasts. **(g)** Detail of the phyllosilicate-bearing layer showing calcite clasts scattered within phyllosilicate layers. Calcite clasts that are in contact show indented and teeth-shaped margins. **(h)** Detail of calcite clasts within phyllosilicate-bearing layers showing pores and vesicles (dark rounded features), in places arranged in linear arrays probably along clasts microdefects such as twinning and cleavage planes. **(i)** Detail of phyllosilicate nanostructures (i.e., nano-sized spherules and tubes) within the phyllosilicate-bearing layer.

Figure 4. In situ mechanical properties within a $16\ \mu\text{m}^2$ area from the phyllosilicate-bearing layer. **(a)** Topographic maps showing calcite clasts (black) scattered within phyllosilicates (red). Maps of the **(b)** first and **(d)** second contact resonance frequency (f_1 and f_2 , respectively) and of the **(c)** first and **(e)** second quality factor (Q_1 and Q_2 , respectively) values showing that calcite clasts are stiffer than phyllosilicates. **(f)** Map of the Young's modulus values showing that calcite clasts have higher Young's modulus than that of phyllosilicates. Maps of **(g)** loss tangent ($\tan\delta$) and **(h)** loss modulus (E'') values showing that phyllosilicates behave in a less viscous manner (i.e., higher loss tangent and lower loss modulus values than calcite) than calcite clasts, which have an elastic behavior.

Figure 5. In situ mechanical properties within a $\sim 2 \mu\text{m}^2$ area from the phyllosilicate-bearing layer. **(a)** Topographic maps showing calcite clasts (black) scattered within phyllosilicates (red). Maps of the **(b)** first and **(d)** second contact resonance frequency (f_1 and f_2 , respectively) and of the **(c)** first and **(e)** second quality factor (Q_1 and Q_2 , respectively) values showing that calcite clasts are stiffer than phyllosilicates. **(f)** Map of the Young's modulus values showing that calcite clasts have higher Young's modulus than that of phyllosilicates. Maps of **(g)** loss tangent ($\tan \delta$) and **(h)** loss modulus (E'') values showing that phyllosilicates behave in a more viscoelastic manner (i.e., higher loss tangent and lower loss modulus values than calcite) than calcite clasts, which show an elastic behavior.

Figure 6. Experimental apparatus, gouge configurations, and mechanical results (evolution of friction coefficient versus slip) of low to high velocity shear experiments (the list of experiments presented in this study can be found in Table S2 in SI). Note that all the data are characterized by a wavy pattern of the friction coefficient. This is commonly expected due to wobble of a rotary system in the case of the not perfect alignment of the so-called stationary and rotary columns [e.g., *Di Toro et al.*, 2010; *Smith et al.*, 2013]. **(a)** SHIVA apparatus with main components labeled (modified from *Smith et al.*, 2015). **(b)** Gouge configurations showing the three different setups: mixed, layered, and pure calcite gouges. **(c)** Evolution of friction coefficient versus slip during experiments at subseismic (0.001 ms^{-1}) slip rate. Friction coefficient reached the 0.7-0.8 values and remained stable during all experiments. **(d, e)** Evolution of friction coefficient versus slip during experiments at seismic (1 ms^{-1}) slip rate. Layered gouges are weaker (i.e., they have a lower distance to the onset of dynamic weakening, D_{ow}) than mixed gouges, which are, in turn, weaker than pure calcite gouges, both in dry and in wet conditions.

Figure 7. Mechanical and mass spectrometer data for high velocity experiments both in dry (red curves, left column) and in wet (blue curves, right column) conditions. **(a, b)** Evolution of shear stress versus normal stress. Wet experiments show an increase in the normal stress in correspondence of friction drop (i.e., during the weakening phase). On the contrary, dry experiments show an increase in the normal stress in correspondence of the strengthening

phase. **(c, d)** CO₂ emissions were recorded only during dry experiments. **(e, f)** Evolution of cross-correlation (Lag0) coefficient. Dry experiments show a positive correlation during most experiments, indicating an in-phase oscillation between normal and shear stress. On the contrary, wet experiments show a negative correlation during the central part of the experiment (i.e., the weakening phase), indicating an increase of normal stress during a decrease in shear stress (i.e., out of phase oscillation).

Figure 8. Mechanical and mass spectrometer data for low velocity experiments both in dry (red curves, left column) and in wet (blue curves, right column) conditions. **(a, b)** Evolution of shear stress versus normal stress. Wet experiments show constant normal stress, whereas dry experiments show normal stress oscillations, which are in phase with shear stress oscillations. **(c, d)** No CO₂ emissions were recorded both in dry and in wet conditions.

Figure 9. FESEM images (left) and EDS maps (right) of experimental microstructures. **(a-b)** Wet gouges (layered and mixed configurations) sheared at subseismic velocity (0.001 ms^{-1}) showing grain size reduction toward the principal slip zone. EDS maps show no concentration of phyllosilicates along distinct layers. **(c-d)** Dry gouges (layered and mixed configurations) sheared at subseismic velocity (0.001 ms^{-1}) showing grain size reduction toward the principal slip zone. Inset shows a detail of phyllosilicate lamellae characterized by no nano-sized spherules and tubes. Grain comminution is more intense than that observed

within wet gouges. EDS maps show no segregation and concentration of phyllosilicates along distinct layers. **(e, f)** Wet gouges (layered and mixed configurations) sheared at seismic velocity (1 ms^{-1}) showing grain size reduction toward the principal slip zone. EDS map shows segregation and concentration of phyllosilicates along micrometer-thick layers both for layered and for mixed configurations. Insets show calcite clasts within phyllosilicate-bearing layers scattered within phyllosilicates. These microstructures are identical to those observed along the Tre Monti Fault (Fig. 3h). **(g, h)** Dry gouges (layered and mixed configurations) sheared at seismic velocity (1 ms^{-1}) showing grain size reduction toward the principal slip zone. Grain comminution is more intense than that observed within wet gouges. Inset shows the nanostructures from the principal slip zone both for layered and for mixed configurations. These structures consist of clumped and chained nano-sized spherules and tubes similar to those observed within phyllosilicate layers along the TMF (Fig. 3i).

Figure 10. Synthetic model for the seismic cycle within clay-bearing, carbonate-hosted fault cores in the brittle upper crust (depths < 2-3 km). **(a)** During the co-seismic phase, earthquake-related slip is facilitated along phyllosilicate-bearing layers, which can promote seismic slip propagation up to the Earth's surface, possibly enhancing surface faulting. Seismic slip can also be assisted by overpressured fluids as testified by fluid-like fault rock structures. **(b)** During the interseismic phase, slow deformation processes are accommodated by transient creep and/or fault healing and re-strengthening through clay-assisted pressure-solution processes and by permanent creep through slow sliding along foliation surfaces and phyllosilicates (i.e., promoting tectonic elastic strain energy dissipation).