Beyond Byerlee Friction, Weak Faults and Implications for Slip Behaviour

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Abstract

Some faults are considered strong because their strength is consistent with the Coulomb criterion under Byerlee's friction, $0.6 < \mu < 0.85$. In marked contrast, numerous studies have documented significant fault weakening induced by fluid-assisted reaction softening that generally takes place during the long-term evolution of the fault. Reaction softening promotes the replacement of strong minerals with phyllosilicates. Phyllosilicate development within foliated and interconnected fault networks has been documented at different crustal depths, in different tectonic regimes and from a great variety of rock types, nominating fluid-assisted reaction softening as a general weakening mechanism within the seismogenic crust. This weakening originates at the grain-scale and is transmitted to the entire fault zone via the interconnectivity of the phyllosilicate-rich zones resulting in a friction as low as $0.1 < \mu < 0.3$. Collectively, geological data and results from laboratory experiments provide strong supporting evidence for structural and frictional heterogeneities within crustal faults. In these structures, creep along weak and rate-strengthening fault patches can promote earthquake nucleation within adjacent strong and locked, rate-weakening portions. Some new frontiers on this research topic regard: 1) when and how a seismic rupture nucleating within a strong patch might propagate within a weak velocity strengthening fault portion, and 2) if creep and slow slip can be accurately detected within the earthquake preparatory phase and therefore represent a reliable earthquake precursor.

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Highlights

Fault friction drops from 0.6 to 0.2 when interconnected networks of phyllosilicates are present.

Fluid-assisted reaction softening is a general weakening mechanism within the seismogenic crust.

The integration of geological data and results from laboratory experiments depicts structural and frictional heterogeneities within crustal faults.

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1. Introduction

For years, there have been lines of evidence that inform the weak vs. strong fault debate. Robust evidence exists that indicates that some crustal faults are strong whereas others are weak. However in the last ten years, the classification of faults as strong or weak seems to have been replaced by the idea that faults are structurally and frictionally heterogeneous. Fault heterogeneities have been mainly proposed on the grounds of field geology, frictional measurements of different natural fault rocks, seismicity distribution, frequency content of seismic waveforms and geodetic imaging of active faults. In this review we begin by presenting supporting evidence for strong faults and their associated internal structure. Then we will show examples of weak faults together with the physico-chemical processes at the origin of fault zone weakness, and report on the frictional properties as measured in the laboratory. In the discussion we merge observations from the internal structure of strong and weak faults with their frictional properties to: a) derive an integrated view of a structural and frictional heterogeneous crustal scale fault; and b) discuss how heterogeneous fault patches might interact during tectonic loading.

An important point worth mentioning in the introduction is that by fault zone weakening processes we mean processes occurring mainly during the entire fault history (hundreds to millions of years), and for fault weakness we refer to a very low steady-state fault frictional strength. This low frictional strength is generally measured in laboratory experiments at low sliding velocities, i.e. $0.01\ \mu m/s < v < 100\ \mu m/s$, and can be used as a proxy to evaluate the fault strength during the interseismic or pre-seismic phases of the seismic cycle. Therefore, the important dynamic weakening mechanisms that occur during the earthquake slip and induced by temperature rise at high slip velocities are not considered in our analysis.

2. Anderson-Byerlee frictional fault mechanics and strong faults

2.1 Anderson-Byerlee frictional fault mechanics

The strength evaluation of faults contained within the crust requires both a measure of the resolved stress on the fault plane and a quantifiable model for the failure threshold. E. M. Anderson in his seminal paper of 1905 and in his memoirs of 1951 developed groundbreaking research on this topic. He identified three tectonic regimes together with the orientation of the faults within these regimes, laying the foundations for fault strength evaluation. The Andersonian theory of faulting is based on three main assumption: a) the crust is homogeneous and isotropic; b) one principal stress is vertical since the Earth's surface is a free surface; and c) brittle faults form in accordance with the Coulomb criterion for shear failure:

$$\tau = C + \mu_i \sigma'_n = C + \mu_i (\sigma_n - P_f) \quad (1)$$

where τ and σ_n are, respectively, resolved shear and normal stresses on the failure plane, C is the cohesive strength, μ_i is the coefficient of internal friction, (C and μ_i are rock material properties) and Pf is the pore fluid pressure. However with increasing displacement the cohesive strength of a fault is very small compared to the shear and normal stresses to be neglected and the internal friction coefficient is replaced by a sliding friction coefficient, μ_s, resulting in the Amontons' law:

$$\tau = \mu_s (\sigma_n - P_f) \quad (2)$$

At this point, assuming hydrostatic fluid pressure, a characterization of the sliding friction coefficient is required for fault strength evaluation. In 1978 J. Byerlee published an extensive dataset of laboratory friction measurements showing that friction is nearly independent of the rock type and is in the range $0.6 < \mu_s < 0.85$. This experimental friction range is commonly known as the Byerlee's rule of friction and the near 4000 citations in Google Scholar received so far by the 1978 paper testifie to the great impact and widespread use of the Byerlee's rule.

2.2 Supporting evidence for Byerlee friction

Once provided experimental support for μ_s another important question is whether such friction coefficients, $0.6 < \mu_s < 0.85$, obtained from laboratory experiments using centimetric or millimetric samples, also hold for large-displacement faults in the crust with dimensions of several kilometres or larger. Two lines of evidence support the applicability of Byerlee friction to crustal faults.

The first evidence is produced by the dip distribution of moderate-to-large ruptures in extensional and compressional environments that seem to be controlled by frictional fault reactivation theory under Byerlee friction (Sibson, 1985; Collettini and Sibson, 2001). For the two-dimensional case in which an existing fault containing the σ_2 axis lies at a reactivation angle, θ_r, to σ_1, equation 2 may be written in term of principal stresses as:

$$R = \frac{(\sigma_1 - P_f)}{(\sigma_3 - P_f)} = \frac{(1+\mu_s \cot\theta_r)}{(1-\mu_s \tan\theta_r)} \qquad (3)$$

defining the relative ease of reactivation for faults at varying angles to σ_1 (Sibson, 1985), where σ_1, σ_2 and σ_3 are the maximum, intermediate and minimum principal stresses respectively. The optimal orientation for frictional reactivation is given by $\theta^*_r = 0.5\tan^{-1}(1/\mu_s)$. As θ_r increases or decreases away from this optimal position, the stress ratio required for reactivation increases. Frictional lock-up ($R \rightarrow \infty$) occurs when $\theta^*_r = 2\ \theta^*_r \tan^{-1}(1/\mu_s)$. For Byerlee's friction range, optimal reactivation occurs when $\theta_r = 25°–30°$ and frictional lockup is expected at $\theta_r = 50°–59°$.

Figure 1 shows the dip distribution of reverse and normal fault ruptures obtained from focal mechanisms of shallow, intracontinental earthquakes (M > 5.5; slip vector raking 90° ± 30° in the fault plane) where the rupture plane is unambiguously discriminated. On the same figure, under the assumption of vertical and horizontal σ_1 trajectories for extensional and compressional regimes, respectively, the dip distributions, δ, of ruptures are plotted also as functions of the reactivation angle, θ_r. The cut-off at $\theta_r \approx 60°$ for both varieties of dip-slip faults is consistent with frictional lockup for $\mu_s = 0.6$, at the bottom of the Byerlee range. Lower coefficients are possible, but $\mu_s = 0.6$ is also consistent with the dominant peak at $\theta_r \approx 30°$ in the reverse-slip distribution representing the optimal orientation for reactivation. The absence of this peak for normal faults is explicable by: a) the different reactivation curves for the two faulting modes with a more acutely defined minimum at optimal orientation for reverse faults than for normal faults (Collettini and Sibson, 2001); and b) the significant number of normal faults ruptures obtained from earthquakes occurring in central Italy and nucleating on faults forming with dip angles of 45° as ductile shear zones following planes of maximum shear stress (Collettini et al., 2009a). The dataset and frictional analysis presented in figure 1 provide strong evidence for seismogenic faults possessing sliding friction coefficient similar to those measured in the laboratory by Byerlee, but see also contradictory observations for earthquakes occurring in the oceanic lithosphere (Craig et al., 2014; Tesei et al., 2018) or microseismicity along low-angle normal faults (Collettini, 2011). The fact that some faults are reactivated close to frictional lock-up also implies that localized fluid overpressure may be needed for reactivation.

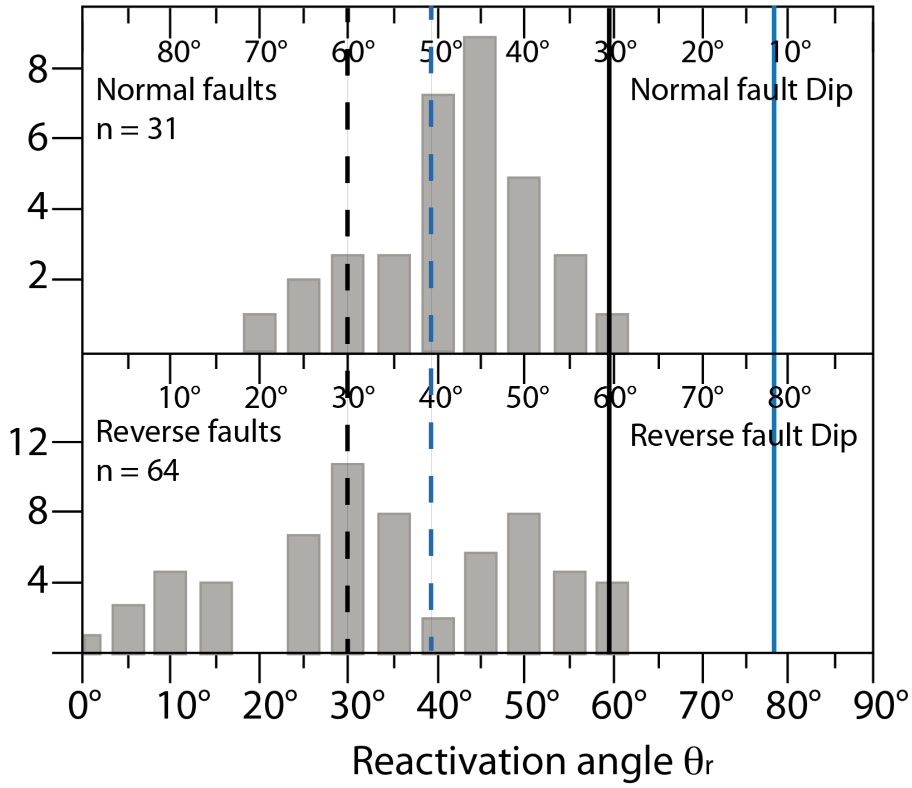

Figure 1. A constraint on Byerlee friction from the dip of the earthquake ruptures. Dip distribution of normal and reverse fault ruptures obtained from focal mechanisms of shallow, intracontinental earthquakes where the rupture plane is unambiguously discriminated (from compilations of Jackson and White 1989; Sibson and Xie, 1998; Collettini and Sibson 2001; Sibson 2009), with the addition of 6 extensional earthquakes from Italy occurring during the L'Aquila 2009 and central Italy 2016-2017 seismic sequences (details in Chiaraluce, 2012; Chiaraluce et al., 2017; cnt.rm.ingv.it). Within an Andersonian stress field, for normal faults, $\theta_r = 90° - \delta$, and for reverse faults, $\theta_r = \delta$, where δ is the fault dip angle. The vertical lines mark optimal orientation (dashed line) and frictional lock-up (solid line) for a friction coefficient of $\mu_s = 0.6$ (black lines) and $\mu_s = 0.2$ (blue lines).

The second evidence for the applicability of Byerlee friction to crustal faults derives from in-situ measurements of the state of stress in the crust. One of the first places where frictional faulting was demonstrated to be clearly applicable to faulted crust was the Yucca Mountain area in Nevada. Here the magnitudes of the least principal stress, measured at different depths, are consistent with frictional fault reactivation, summarized in equation 2, for a friction coefficient of $\mu_s \approx 0.6$ (Zoback and Healy, 1984). On the same line of evidence, a comprehensive compilation of stress measurements, in relatively deep boreholes from different tectonic environments and through different rock types, shows that the measured stress magnitudes are consistent with the values predicted by the Coulomb criterion for hydrostatic fluid pressure and for friction coefficients within the Byerlee's experimental range (Townend and Zoback, 2000). In normal faulting stress fields, borehole stress measurements in sedimentary basins are consistent with the Coulomb criterion for a friction of $\mu_s \approx 0.6$

(Zoback, 2010). Examples are documented in Texas within different rock types like sandstone, siltstone, shale and limestones, in the North Sea within chalk and in the Gulf of Mexico within sand reservoir. More recently the induced seismicity crisis in Oklahoma and southern Kansas have provided a unique opportunity to better characterize the reactivation of dormant faults under anthropogenic forcing. The widespread occurrence of seismicity despite very modest pressure changes in the basement and the observation that the activated faults are well oriented within the contemporary stress field (Walsh and Zoback, 2016; Schoenball and Ellsworth, 2017) provide strong support for the hypothesis of a critically stressed strong crust with hydrostatic fluid pressure and Byerlee friction (e.g. Townend & Zoback, 2000).

2.3 Structural and mechanical characteristics of strong faults

The localization of strain along crustal faults produces a fault structure that generally consists of a fault core, where most of the deformation is localized, surrounded by a damage zone formed by distributed fractures and subsidiary, small displacement faults (e.g. Chester et al., 1993). Cataclastic deformation (brittle fragmentation by macroscopic fracturing and grain comminution) intensifies toward the fault core that consists of one or more tabular zones of ultracataclasite, within which bands of intense grain-size reduction define principal slip zones (e.g. Sibson, 1977; Chester and Chester, 1998). A comprehensive characterization of fault zone structure is beyond the scope of the present manuscript and it is well described in several excellent review papers such as Caine et al., (1996), Ben-Zion and Sammis, (2003), Sibson (2003), Wibberley et al., (2008) and Faulkner et al., (2010). Here we try to link structural and mechanical data in order to summarize the main mechanical characteristics of strong faults.

Several faults hosted within crystalline rocks show cataclastic deformation with grain size reduction and localization along discrete slipping surfaces (e.g. Fig. 2a and Smith et al., 2013). The preservation of pseudotachylites within some of these shear zones (Sibson 1977; Swanson 2006; Di Toro and Pennacchioni 2005; Spray, 2010) testifies to earthquake occurrence along these structures. The San Gabriel fault, within the San Andreas fault system, has an heterogeneous fault structure that in some localities is constituted by a fault core with significant slip localization along ultracataclasite layers made of quartz and feldspars (Evans and Chester, 1995). A classical example of extreme localization has been documented for the Punchbowl fault (Chester and Chester, 1998), with a 1 mm thick principal slip zone consisting of < 100 nm particles (Chester et al., 2005). Dynamic weakening mechanisms (e.g. Rice, 2006; Di Toro et al., 2011) favoured by extreme localization have been invoked (Chester and

Chester, 1998) to explain the weakness of the San Andreas: in this view the fault would be statically strong yet dynamically weak (Rice et al., 2009). Some faults that cross-cut silica-rich sediments, like the Corona Heights fault, in San Francisco, show a mirror-like finish due to the presence of 1-3 mm thick zone of vitreous silica formed during earthquake slip (Kirkpatrick et al., 2013). Faulting within sandstones results in one or more through-going slip surfaces where the great part of the displacement is accommodated by quartz grain-size reduction and localization along principal slip surfaces (Shipton and Cowie, 2001). The Pretorius fault in Tau Tona mine, South Africa, during its Archaean activity, experienced multiple slip events along a quartz-rich principal slip surface and in 2004 it was reactivated by a M = 2.2 earthquake. The mapped earthquake rupture at 3.6 km of depth reveals that the seismic slip produced 1-5 mm thick fault gouge along four quasi-planar segments of the ancient fault-zone (Heesakkers et al., 2011). Faulting within massive carbonates (Fig. 2b) is characterized by localization along sub-parallel slipping zones where the deformation is localized along very thin (< 500 μm) zones bounded by mirror-like slipping surfaces (De Paola et al., 2008; Fondriest et al., 2013; Siman-Tov et al., 2013; Collettini et al., 2014). In some of these carbonate-bearing faults, calcite crystals exhibiting localized disaggregation together with a high concentration of vesicles indicate thermal decomposition during past earthquakes (e.g. Rowe et al., 2012; Collettini et al., 2013). In other structures nanograins texture with polygonal grain boundaries suggest superplastic deformation of carbonates during earthquake-slip (De Paola et al., 2015).

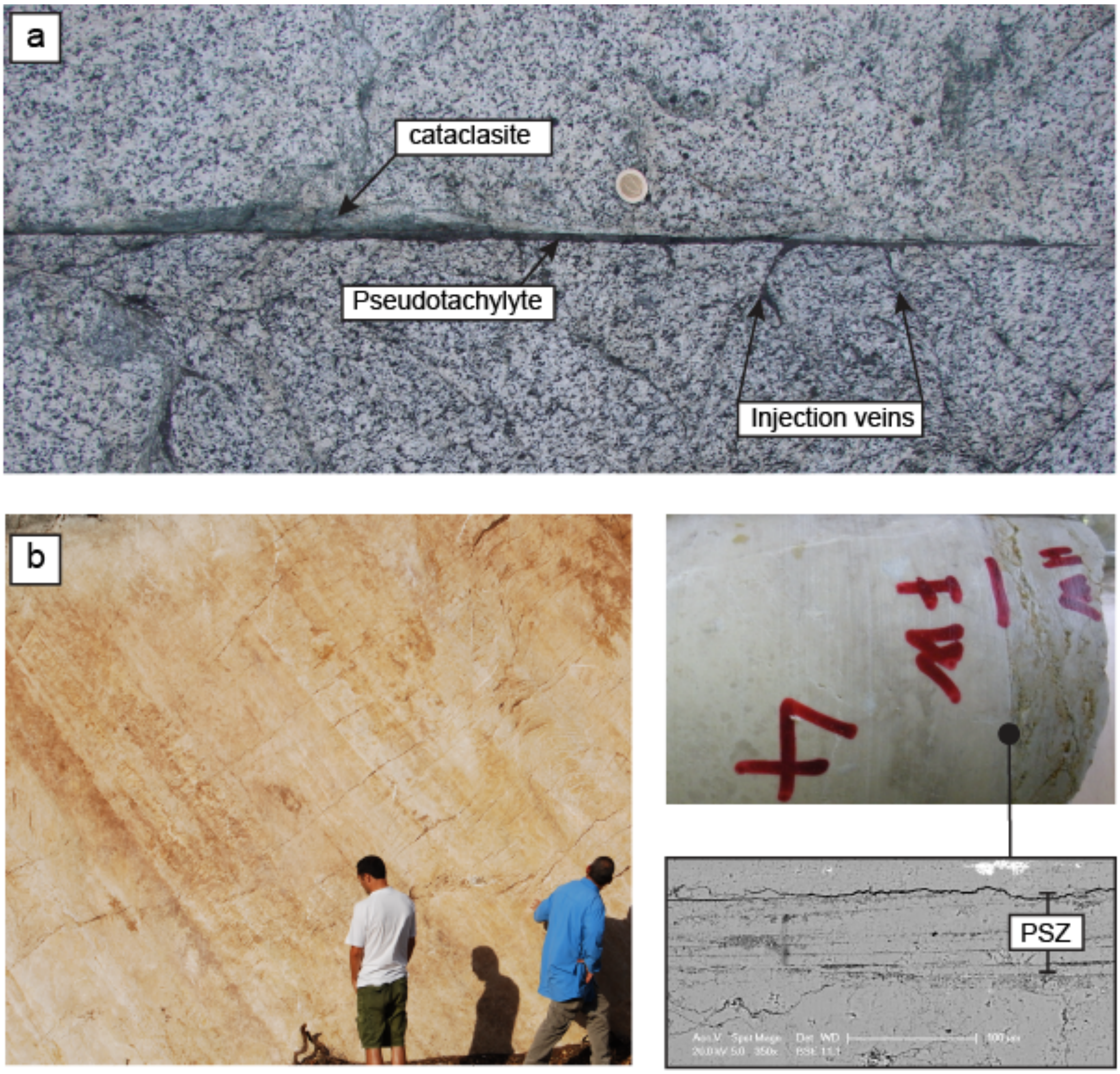

Figure 2. Strong faults. a) Pseudotachylyte-bearing fault from the Gole Larghe outcrop in the Italian Alps (Di Toro and Pennacchioni, 2005; Smith et al., 2013). Cataclasite- and pseudotachylyte-bearing faults surround relatively intact blocks of tonalite, and pseudotachylyte overprints cataclasites. b) Mirror-like slip surface from the Monte Maggio fault in the Apennines of Italy (Collettini et al., 2014). The panels on the right show localized deformation along a sharp slipping zone, affected by grain-size reduction and thermal decomposition processes (localized disaggregation with a high concentration of holes) along the Principal Slipping Zone, PSZ; Scanning Electron Microscope, SEM, image.

3. From strong to weak faults

The datasets presented above support the interpretation that crustal faults in general fail according to equation (2) with μ_s = 0.6 – 0.85 and hydrostatic fluid pressure. These structures represent strong faults since the differential stress, (σ_1 - σ_3), or the shear stress required for their reactivation is quite high. At 10 km of depth (σ_1 - σ_3) > 100 MPa and it increases significantly from extensional to compressional environments (Sibson, 1974); for example in the KTB borehole the differential stress at 9 km of depth is about 170 MPa (Townend and Zoback, 2000).

However, several observations cast doubt on the fact that deformation within the crust is exclusively controlled by strong faults and amongst these, we report the two that we consider the most relevant. First, there is an important number of faults that experience reactivation although they are severely misoriented within the regional stress field. Examples of these structures are given by the San Andreas fault in a strike slip regime (Zoback et al., 1987; Carpenter et al., 2011; Lockner et al., 2011), low-angle normal faults in extensional environments (Wernicke, 1981; Collettini 2011) and sub-horizontal thrusts in compressional regimes (Price, 1988; Davis et al., 1983; Suppe 2007; Tesei et al., 2015). Reactivation along these structures is possible only following significant fault weakening that can be achieved by either an increase in fluid pressure, or a reduction in friction coefficient or a combined effect (e.g. Hubbert and Rubey, 1959; Rice, 1992; Faulkner et al., 2006; Suppe 2007; Tesei et al., 2015). Second, in the last 15 years, geophysical observations combined with numerical models have shown that aseismic creep is common and sometimes prevalent within the seismogenic layer (e.g. Avouac, 2015), and a continuum spectrum of fault slip behaviour, including slow slip phenomena, can be present at all depths of crustal faults (e.g. Bürgmann, 2018). This richness in fault slip behaviour for crustal faults is difficult to be captured only by strong faults with Byerlee friction.

3.1 Weak-fault structure

In marked contrast to strong faults, there is a significant number of crustal structures showing distributed deformation along interconnected, anastomosing shear zones rich in phyllosilicates. The geometry and internal fabric of these shear zones strongly resembles, and is sometimes derived from, ductile shear zones of high metamorphic grade (e.g. Berthé et al., 1979; Platt and Vissers, 1980). In this paragraph we will present some well-documented field examples of phyllosilicate-rich crustal faults (Fig. 3 and Table 1).

For strike-slip faulting, the Carboneras fault in Spain (Fig. 3a) is characterized by 1 km thick fault core consisting of foliated and interconnected networks of phyllosilicate-rich zones (Faulkner et al., 2003; Rutter et al., 2012). These anastomosing networks are up to 50 m thick and are rich in chlorite and illite derived from a mica schist protolith (Solum and van der Pluijm, 2009). The creeping section of the San Andreas fault at SAFOD consists of multiple fault strands, made of foliated serpentinite and smectite clays (Holdsworth et al., 2011), that are several meters wide and creep simultaneously (Zoback et al., 2010). Some segments of the Median Tectonic Line in Japan consist of several meters thick, foliated fault rocks rich in chlorite (Wibberley and Shimamoto, 2003; Jeffereis et al., 2006). During the final activity of the fault at shallow crustal levels the deformation is concentrated along clay rich shear zones (Wibberley and Shimamoto, 2003). Exhumed shallow portions of the North Anatolian fault are characterized by hundreds of meter thick faults where the deformation is mainly accommodated within sub-parallel shear-zones rich in talc, kaolinite and chlorite (Kaduri et al., 2017). The Livingstone fault zone in New Zealand (Tarling et al., 2018) is dominated by a serpentinite mélange tens to several hundreds of metres wide, in which a pervasive anastomosing fabric surrounds pods of more competent material (e.g., metasediments, rodingite, massive serpentinite) ranging from tens to hundreds of metres in size. The foliated serpentinite shear zone is mostly made of fibrous serpentine and lizardite, consistent with the estimated ambient temperature during shearing of 300–350 °C.

For thrusts faults there are plenty of examples of foliated phyllosilicate-rich shear zones. Notable examples may include the classic phyllonites associated to the Moine thrust zone, possibly formed by retrogression and shearing of mylonitic gneissose protoliths under lower greenschist facies conditions (McClay and Coward, 1981, Wibberley, 2005). Other examples of phyllonitic fault rocks have been documented in the Karakoram fault zone in the Himalayas (Wallis et al., 2015) or along the Red River shear zone in the Yunnan Province in China (e.g. Wintsch and Yeh, 2013). The Perdido thrust in the Pyrenees consists of a several meters thick and foliated shear zone, rich in illite and chlorite (Lacroix et al., 2011), separating more competent lithologies in the hanging-wall (limestones) and footwall (turbidites sandstone). The Monte Fico thrust in the Elba Island in Italy (Fig. 3b) is a c. 200 m thick shear zone formed by competent lenses of retrograde pseudomorphic serpentinite surrounded by an anastomosing network of foliated serpentinites (Viti et al., 2018). Several tectonic mélanges around the world, thought to constitute exhumed analogues of subduction channels, are constituted by anastomosing shear zones enriched in phyllosilicates enveloping lenses of more competent lithologies (e.g. Cowan, 1974; Byrne et al., 1988; Cloos and Shreve, 1988, Kimura et al., 2012; Morley et al., 2017; Rowe et al., 2013). For instance, the Rodeo Cove thrust, near San Francisco, consists of a 200 m thick shear zone with a foliated fabric rich in chlorite that formed at about 8-10 km of depth (Meneghini and Moore, 2008). The Chrystalls Beach Complex accretionary mélange of New Zealand (Fagereng and Sibson, 2010) consists of competent lenses of chert, sandstone and metabasalts surrounded by an interconnected network of phyllosilicates, i.e. illite-muscovite, developed at depth < 300°T (Fagereng and Cooper, 2010). At shallower crustal levels other examples of tectonic mélange are reported in Vannucchi et al., (2012 and references therein). In the Apennines of Italy, in marly limestones, shear zones formed at crustal depths < 4 km show a thickness ranging from several to hundreds of meters with concentration of illite and smectite within the interconnected and foliated network of the fault core (e.g. Koopman, 1983; Tesei et al., 2013).

Some large displacement extensional faults formed in the US within quartz-feldspatic rocks reveal a heterogeneous structural zonation with fault zone thickness ranging from < 10 m for the Mineral Mountains fault to locally about 20 m for segments of the Wasatch and Dixie Valley faults (Bruhn et al., 1994). The fault zone is made of fault breccia, fine-grained cataclasites and foliated zones rich in muscovite, biotite, chlorite and clays. Some portions of the Zuccale low-angle normal fault in Italy, consist of an up to 8 m thick fault core rich in talc and smectite (Collettini et al., 2011). The Black Mountains detachment in California shows deformation that is asymmetrically distributed, increasing upward from the footwall (Cowan et al., 2003). A well-defined slip zone separates hangingwall Quaternary fanglomerates from fault rocks consisting of foliated fault breccia and fault gouge where weak mineral phases (illite, chlorite, smectite and saponite) are concentrated (Hayman, 2006). The Gubbio normal fault in the Apennines of Italy shows some segments characterized by foliated SC fabric from the metric (Fig. 3c) to the microscale. These segments form predominantly within marly carbonates and show the concentration of clays within the shear zone (Bullock et al., 2014). The Err Nappe detachment in Switzerland shows a fault core made of a continuous layer of black gouge, with thickness ranging from a few centimetres to some metres (Manatschal, 1999). In the footwall the granitic host rock is affected by brittle fracturing associated with a complex vein system and towards the fault core fluid-assisted diffusion mass transfer processes occurring under lowermost greenschist facies conditions promoted the development of an SC fabric rich in chlorite and illite (Manatschal, 1999).

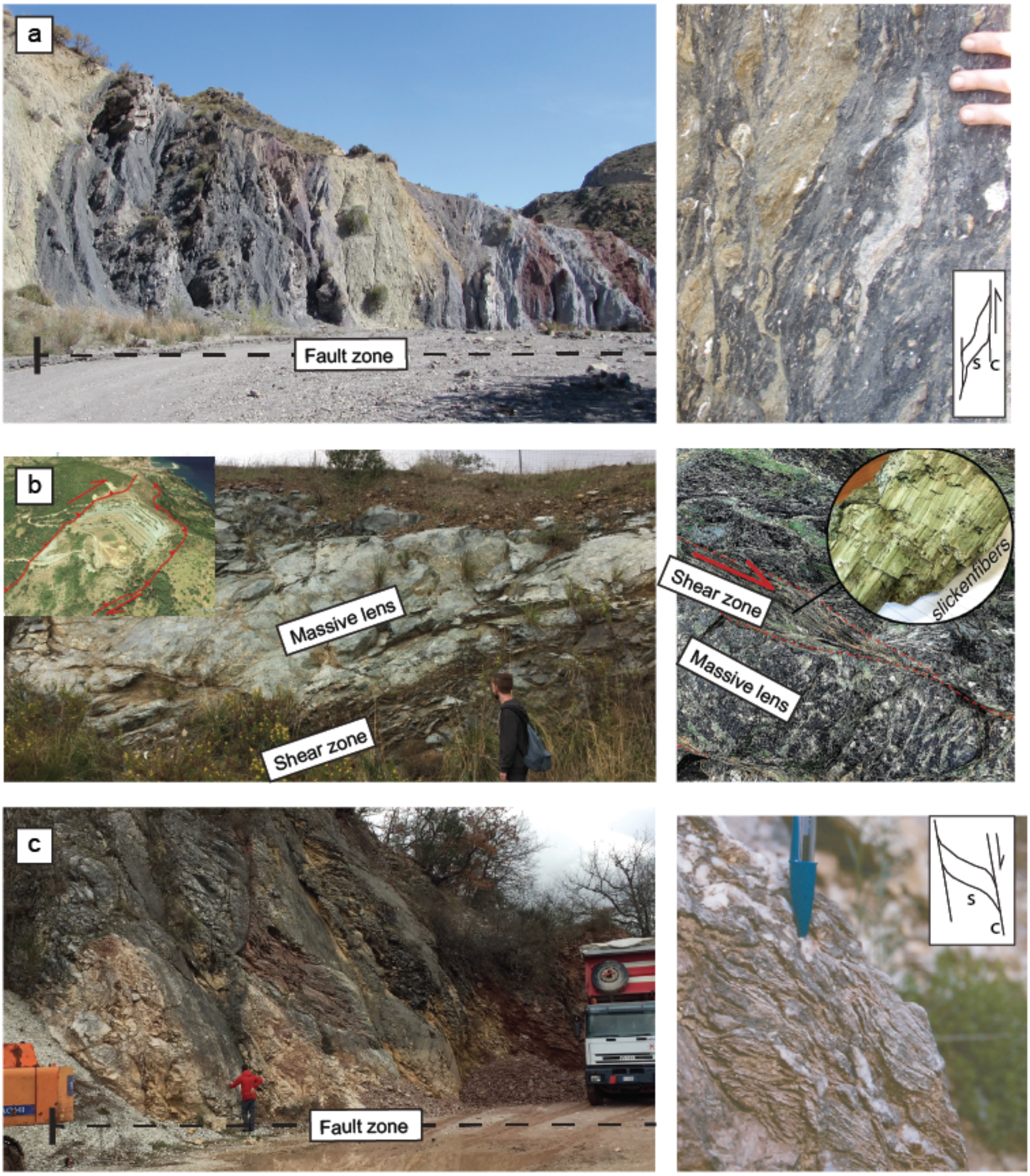

Figure 3. Weak faults. a) Outcrop view of the Carboneras fault SE Spain (Faulkner et al., 2003; Rutter et al., 2012). Each of the coloured bands represents different types of fault gouge that have been juxtaposed due to movements on the fault. The image on the right shows details, in map-view, of the SC fabric with left-lateral kinematics. The fault rock consists of a clay-bearing gouge derived from graphitic mica schist (Rutter et al., 2012). b) The Monte Fico thrust (Elba Island, Italy) is a 200m thick structure (panoramic view in the inset) characterized by SCC' fabric. Sigmoidal competent lenses of serpentinites are surrounded by shear-zones coated with fibrous serpentine and lizardite (Tesei et al., 2018; Viti et al., 2018). c) Gubbio normal fault in the Apennines of Italy. In this outcrop the fault is up-to 30 m wide and is characterized by an SCC' fabric at the metric and centimetric scale (details in Bullock et al., 2014).

3.2 Reaction softening

Fault zone architecture exerts a primary control on fluid flow in crustal shear zones (Sibson, 1992; Caine et al., 1996; Wibberley et al., 2008; Faulkner et al., 2010). The influx of fluids into fault zones can trigger two main types of weakening process that operate over different timescales. In the short term of the seismic cycle, crustal fluids can be trapped within low-permeability fault zones promoting the development of fluid overpressure, e.g. the mechanical weakening that reduces the effective normal stress (equation 2 and Hubbert and Rubey, 1959). During the entire fault history fluid circulation within shear zones might exert a chemical role facilitating the replacement of strong mineral phases with weak mineral phases, hence promoting reaction softening (e.g. Janecke and Evans, 1988; Bruhn et al., 1994; Evans and Chester, 1995; Wintsch et al., 1995; Manatschal, 1999; Imber et al., 1997; Wibberley, 1999; Collettini and Holdsworth 2004; Schleicher et al., 2010; Warr et al., 2014). In the following we will review some examples of fluid assisted reaction softening generated in different protoliths and occurred at different crustal levels. Then we will integrate these observations in a general mechanism for fluid assisted fault weakening.

In the thrusts of the Apennines, when shear zones develop within marly limestones, faults with kilometric displacement show distributed deformation along thick (up to 200 m) shear zones (Fig. 4a and Tesei et al., 2013) characterized by SCC' fabric (e.g. Ramsay and Graham, 1970; Berthé et al., 1979; Bos and Spiers, 2001 and Fig. 4b). The evolution of the fault zone structure, inferred from the analysis and comparison of small, intermediate and kilometric displacement faults, indicates that during the early stages of deformation dissolution of the carbonates favours the concentration of insoluble clay minerals within stylolitic surfaces (Fig. 4c, Tesei et al., 2013; Gratier et al., 2013; Lacroix et al., 2015). With increasing deformation the stylolites evolve into an interconnected foliation (Fig. 4d, see also Gratier and Gamond, 1990), formed by smectitic clays in nanosized (001) lamellae (Fig. 4e) with preferred orientation parallel to the local slipping surface (Viti et al., 2014). Foliation parallel veins (Fig. 4c) with crack-and-seal texture, suggest that the low permeability and the anisotropy of clay fabric favoured cyclic fluid overpressure development during the fault activity.

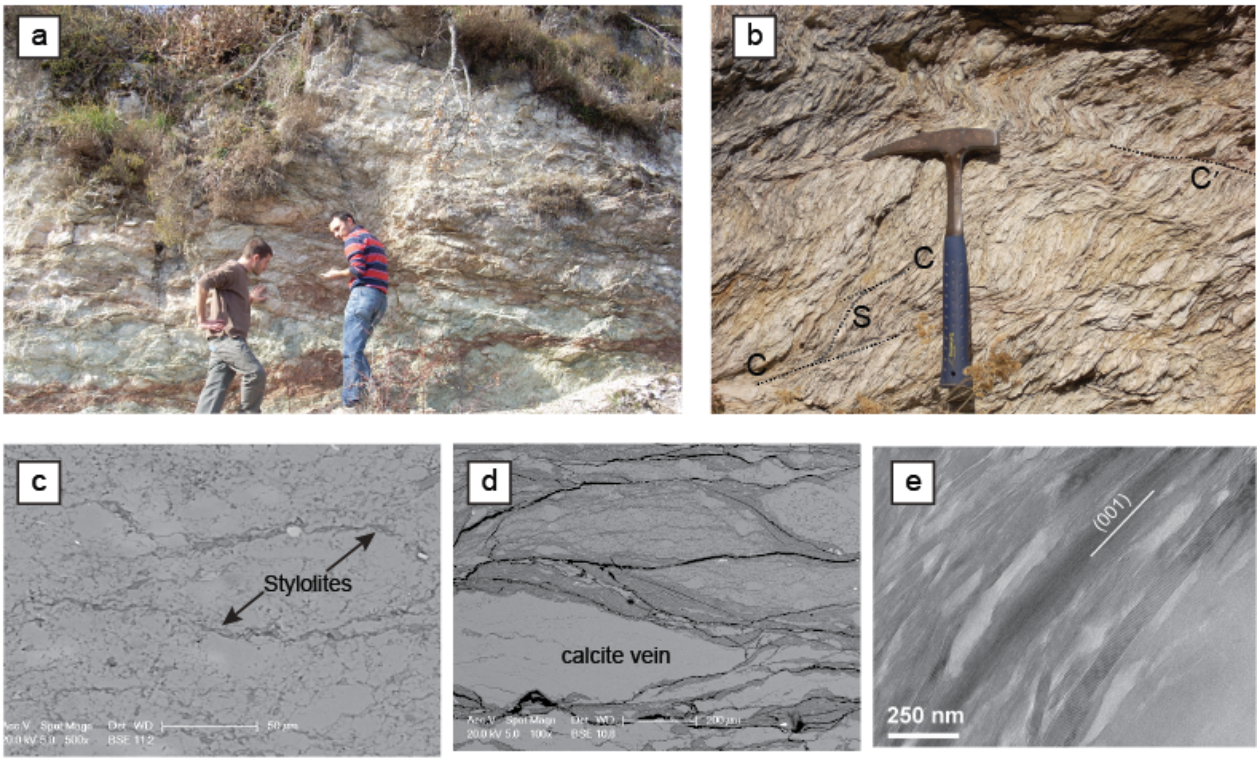

Figure 4. Fault weakening in marly limestones. a) and b) In the exposed outcrop, the Coscerno thrust (Northern Apennines of Italy) is ≈20 m thick and is characterized by a fault rock showing a pervasive SCC' fabric (e.g. Tesei et al., 2013). c) From a marly protolith, characterized by calcite (light grey) with a small amount of clay (heavy grey), dissolution of calcite favours the concentration of insoluble clays initially along stylolites. d) In mature fault rocks the clay minerals form interconnected networks, with calcite veins parallel to the foliation. In some cases these veins a re-worked by dissolution processes. e) Within the interconnected clay-rich networks, the deformation is accommodated by frictional sliding along smectite lamellae nearly parallel to the sense of shear. c) and d) are Scanning Electron Microscope, SEM, images and e) is a Transmission Electron Microscope, TEM, image.

Along the Zuccale low-angle normal fault (Elba island, Italy) during the initial phase of deformation, fracturing of a dolomitic protolith favoured the influx of silica-rich fluids into the fault zone. In the low-strain domains fluids interacted with the fine-grained cataclasite promoting dissolution of the dolomite and precipitation of talc (Collettini, et al., 2009c). In the high-strain domains the mature fault zone structure consists of an interconnected foliated network (Fig. 5a-c) that deforms by frictional sliding along 50–200 nm-thick talc and smectite lamellae (Fig. 5d and Viti and Collettini, 2009). Here again foliation-parallel veins (Fig. 5b) indicate fluid involvement and fluid overpressure development during the fault activity.

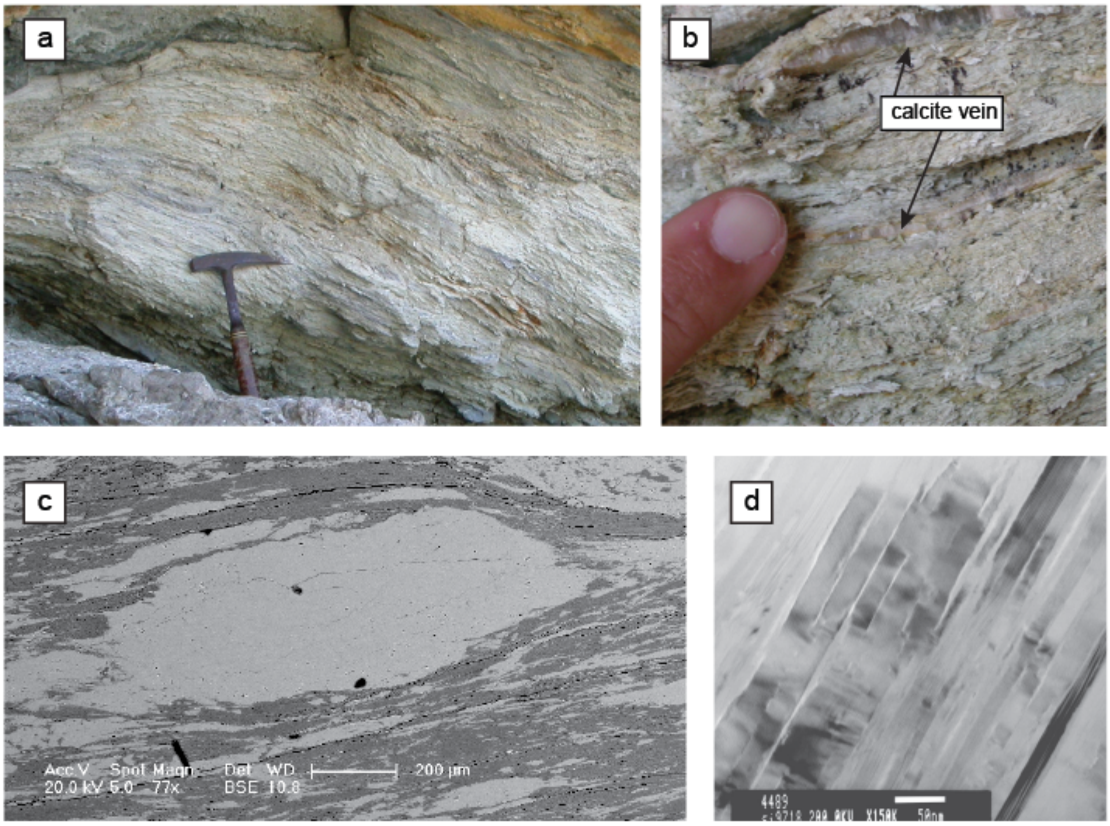

Figure 5. Fault weakening in dolostones. a) Talc-rich foliated structure of the Zuccale low-angle normal fault in the Elba Island, Italy (Collettini et al., 2009c). b) Detail of the foliated fault rock with foliation parallel calcite veins. c) SEM image showing sigmoids of calcite, light grey, within interconnected talc-rich foliated microstructure, heavy grey. d) TEM image showing interlayer delamination and frictional sliding along talc (001) lamellae (Viti and Collettini, 2009).

The Moonlight Fault in southern New Zealand consists of a 20 m thick shear zone formed within grey and green schists. Microstructural studies suggest that in the early stages of the fault activity brecciation promoted grain-scale dilatancy favouring the influx of hydrous fluids into the fault zone (Fig. 6 and Alder et al., 2016). Seams of fine-grained insoluble phyllosilicates that anastomose clasts of quartz and albite, and overgrowth of chlorite indicate that the shear zone deformed by dissolution - precipitation mechanisms that accompanied frictional sliding along the phyllosilicate foliae. In the high strain domains of the shear zone it is evident a high concentration of chlorite and muscovite derived from the hangingwall green-schists and footwall grey-schists respectively (Fig. 6 and Alder et al., 2016). Similar weakening processes in quartz-feldspatic rocks have been extensively documented in other fault zones worldwide in Janecke and Evans, (1988), Bruhn et al., (1994), Evans and Chester, (1995), Wintsch et al., (1995), Wibberley, (1999), Imber et al., (1997), Wintsch and Yeh, 2013 and Wallis et al., 2015.

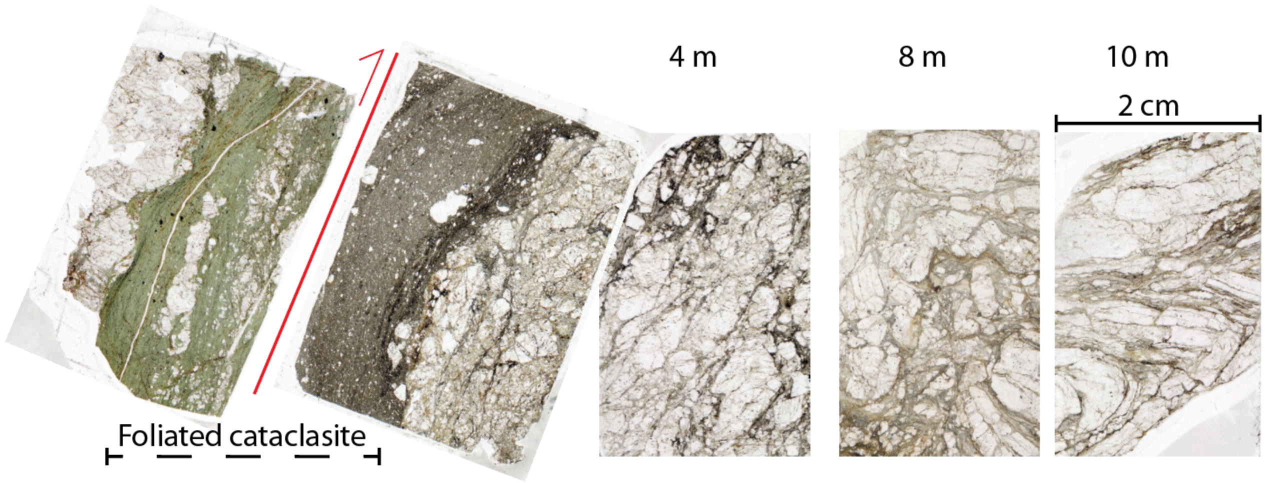

Figure 6. Fault weakening in quartz-feldspatic schists. The Moonlight fault in New Zealand shows a fault core containing an up to 20 m thick sequence of breccias, cataclasites and foliated cataclasites. Series of scanned thin sections from samples collected at distances of <10 m from the shear zone illustrating the progression of deformation towards the fault core via a: 1) decrease in grain-size/increase in matrix proportion; 2) modal increase in abundance of phyllosilicates; 3) increasing alignment of phyllosilicate lamellae, chlorite in green and muscovite in black. Modified from Alder et al., 2016.

In ultramafic rocks different stages of weakening have been documented. Brittle fracturing and grain-size reduction is a fundamental process in the early stages of peridotite serpentinisation, because it promotes the formation of fluid pathways and allows for efficient interface hydration reactions of primary peridotitic minerals (Plumper et al., 2012). Serpentinisation results in pseudomorphic textures (mesh cores/rims and bastites, from olivine and pyroxene, respectively), consisting of a mixture of lizardite, chrysotile and polygonal serpentine (e.g. Escartin et al., 2001). In retrograde serpentinites, during shear deformation, two main processes favour further progressive weakening (Fig. 7 and Viti et al., 2018): 1) preferential dissolution of mesh cores favours the development of an interconnected network of sub-parallel lizardite lamellae with (001) planes parallel to the shear direction; and 2) subsequent precipitation promotes the development along shear zones of fibrous serpentines (chrysotile and polygonal serpentine) with the fibre axis oriented parallel to the shear direction (Fig. 7 c). Frictional sliding along (001) lizardite lamellae or along fiber axis of fibrous serpentinites results in a friction of $0.15 < \mu_s < 0.19$ (Tesei et al., 2018).

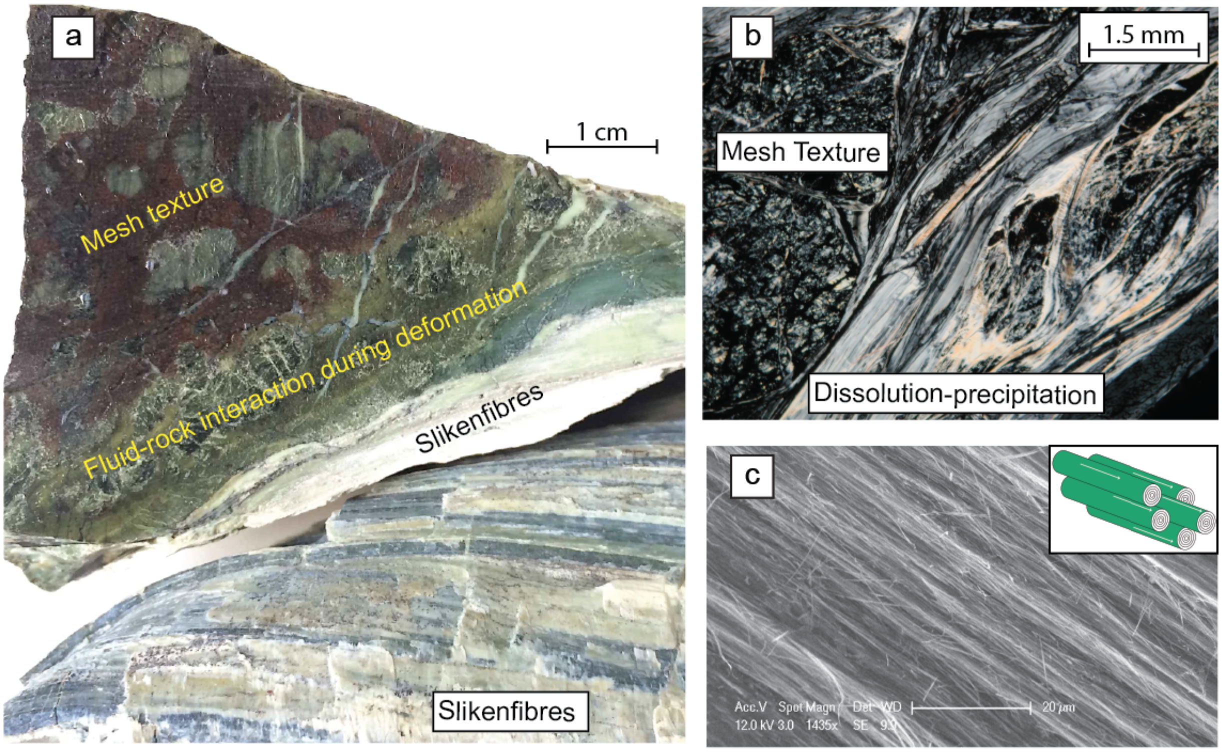

Figure 7. Monte Fico fault in Elba island, Italy: fault weakening in ultramafic rocks. a) Hand sample showing the transition from the mesh texture to the slikenfibers coating the shear zones. b) Mesh texture and slikenfibers at the optical microscope. c) Parallel fiber axis of fibrous serpentines, SEM image (details in Tesei et al., 2018; Viti et al., 2018).

Collectively the examples of fault evolution reported above indicate that in the early stages of deformation, brittle fracturing favours the increase of fault zone permeability promoting the influx of fluids into the fault zone (Fig. 8a). Fluids interact with the fine-grained portions of the cataclasites (Fig. 8b) and promote dissolution and precipitation processes that favour the replacement of strong mineral phases (quartz, feldspar, olivine, pyroxene, calcite, dolomite) with weak mineral phases (clays, talc, chlorite, muscovite, lizardite, fibrous serpentine). With increasing deformation, this fluid assisted reaction softening, allows the development of an interconnected and phyllosilicate-rich microstructure where a significant amount of slip is accommodated by frictional sliding along the phyllosilicate foliae (Fig. 8c). The development of foliated networks rich in platy minerals makes the fault a low permeability barrier that can trap crustal fluids and generate fluid overpressures as suggested by the numerous veins documented parallel to the foliated networks (e.g. Fig. 4d, 5b, 7b).

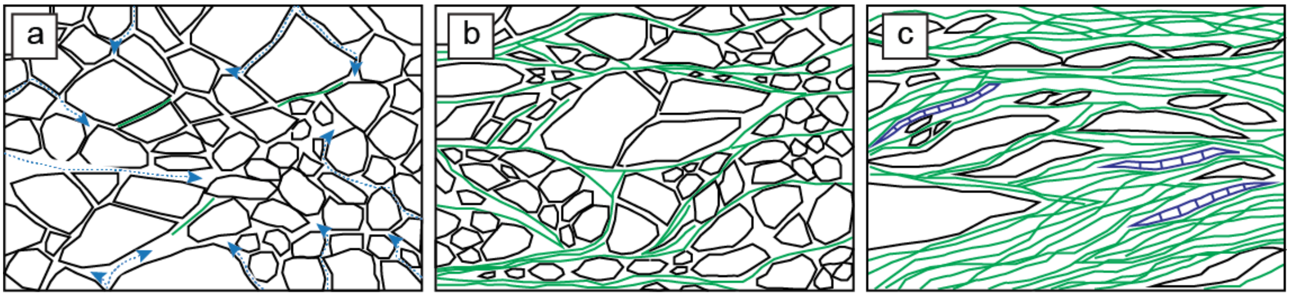

Figure 8. Schematic representation of reaction softening with increasing strain. a) At the onset of deformation fracturing associated to cataclasis increases permeability favouring the influx of fluids (blue arrows) into the fault zone. b) Fluids react with the fine-grained cataclasite promoting dissolution of the strong granular phases and precipitation of phyllosilicates (green lines). c) At high strains the microstructure consists of an interconnected phyllosilicate-rich network where the deformation is predominantly accommodated by frictional sliding along the (001) phyllosilicate lamellae. The phyllosilicate network is also a low-permeability horizon for transversal fluid flow favouring the development of fluid overpressure testified by foliation parallel veins with crack-and-seal texture (dark-blue). Key-references on the processes highlighted in this picture are reported on the main text.

Within the fault zone, from early to mature stages of deformation, there is an evolution from a granular load-bearing network to a weak and interconnected foliated microstructure (e.g. Handy, 1990; Holdsworth, 2004). A similar microstructural evolution involving the combined effect of pressure solution of granular materials and frictional sliding on phyllosilicates have been reproduced in the laboratory, mainly at Utrecht University (e.g. Bos & Spiers 2001). The associated mechanical data have been used to characterize a frictional-viscous behavior (i.e., both normal stress and strain rate dependent) active within the seismogenic crust and describe its implications for crustal strength profiles (Bos & Spiers 2001; Niemeijer & Spiers, 2005; Den Hartog and Spiers, 2014). The rheological strength profiles for a foliated and phyllosilicate-rich faults contained within the seismogenic crust is composed of three main deformation regimes (Fig. 9): A) at shallow crustal level, the deformation is mainly accommodated by cataclastic processes involving dilation: this behaviour closely resembles the Byerlee's rule with a linear trend controlled by high (μ = 0.6-0.85) friction; C) at greater depth, following the development of interconnected phyllosilicate-rich networks via pressure-solution processes, the slip behaviour is mainly controlled by frictional sliding along the phyllosilicate foliae, i.e. linear trend with low friction (cf. paragraph 3.3 for details on friction); B) the transition region represents the pressure solution controlled regime (regime B), where mechanical behavior is strongly rate-sensitive as well as normal stress sensitive. At larger crustal depth, frictional sliding along the phyllosilicates (regime C) is replaced by plastic flow.

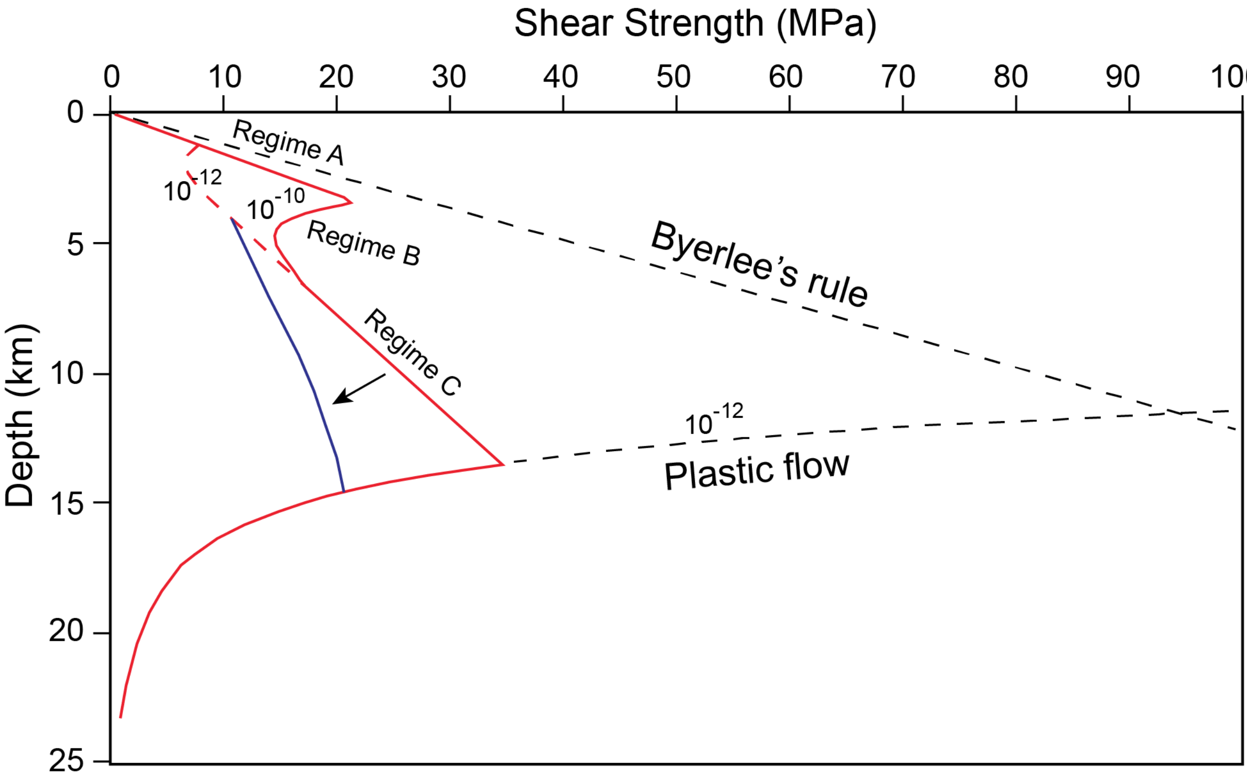

Figure 9. Crustal strength profile for a quartz (Byerlee's friction) muscovite (μ = 0.3) assemblage within a strike-slip fault (from Niemeijer and Spiers, 2005). The model of Niemeijer and Spiers (red curve) defines three main deformation regimes in which the strength is dominated by: A) cataclastic deformation; B) pressure solution; C) frictional sliding on phyllosilicate foliae. With increasing crustal depth frictional sliding along the phyllosilicates is replaced by crystal-plastic flow of quartz. The strength profile is constructed for a geothermal gradient of 25°C/km and the influence of strain rate on the depth of pressure solution accommodated deformation (regime B) is shown for strain-rates of 10^{-10} and 10^{-12} s^{-1}. Reduction in phyllosilicate-friction or the onset of phyllosilicate plasticity for high geothermal gradients (e.g. Wallis et al., 2015) can promote further strength reduction (blue curve).

3.3 Frictional properties of phyllosilicate-rich faults

The simple conceptual model we can reconstruct by merging the observations presented in paragraphs 3.1 and 3.2 suggests that fluid-rock interaction within fault zones might produce frictional-viscous flow with the development of interconnected networks of phyllosilicates resulting in fault weakness. In regime C, the low-strength of the fault is mainly controlled by frictional sliding along phyllosilicate foliae. To provide mechanical evidence for this weakness, here we present a vast compilation of frictional measurements on phyllosilicate-rich faults.

Data reported here have been collected during friction experiments at low sliding velocities, in general 0.01- 100 μm/s. These experiments are used as a proxy to evaluate the steady state frictional strength of a fault during the interseismic or pre-seismic phase of the seismic cycle.

To characterize the frictional properties of fault rocks during laboratory experiments a constant normal stress is applied on the rock sample and then a shear stress is induced by shearing the fault at constant sliding velocity. The experimental fault generally shows an initial phase, where the shear stress increases rapidly during elastic loading, before a yield point, followed by shear at a steady-state friction value (Fig. 10a). During these friction experiments velocity steps and slide-hold-slide sequences are usually performed to characterize the frictional stability together with the healing properties of the tested material (e.g. Dieterich, 1979; Ruina, 1983; Marone 1998a and references therein).

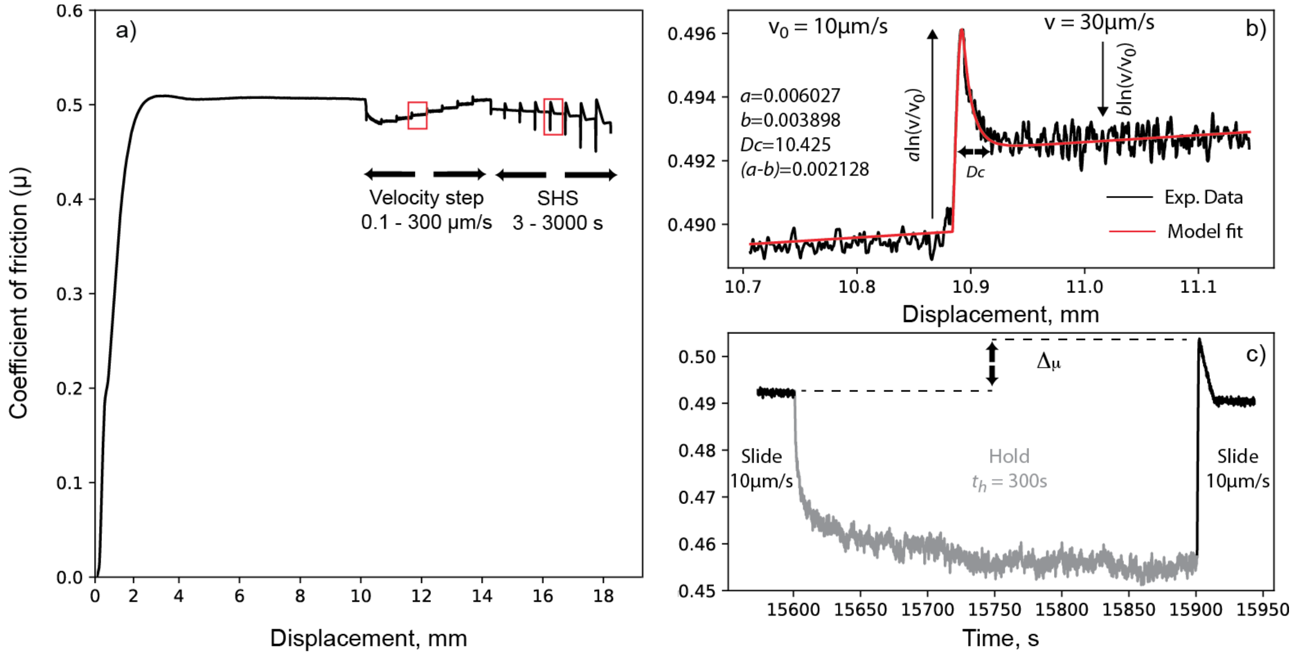

Figure 10. Laboratory experiment to characterize the rock frictional properties. The tested material is powdered illite at room humidity and room temperature. a) A typical experiment to characterize the frictional properties of a fault consists of a run-in phase to achieve steady state shear strength and measure steady state friction. Then velocity steps b) and/or c) slide-hold slide, SHS, tests are usually performed to constrain the velocity dependence of friction, *a*-*b*, the critical slip distance D_c, and the healing properties, Δ_μ, of the experimental fault.

During velocity-stepping tests (Fig. 10b) a near-instantaneous step change in sliding velocity from V_0 to V is imposed and the new sliding velocity is held constant until a new steady state shear stress level is attained. The instantaneous change in friction scales as the friction parameter $a\ln(V/V_0)$, where a is an empirical constant defined as the direct effect (e.g., Ruina, 1983). The subsequent drop to a new steady state value of friction scales as the friction parameter $b\ln(V/V_0)$, where b is an empirical constant defined as the evolution effect (e.g., Ruina, 1983). D_c is the critical slip distance that is the displacement over which the population of asperity contacts that control friction are renewed. The velocity dependence of steady state friction (a - b) is defined as:

$$(a-b) = \frac{\Delta\mu_s}{ln\left({}^{V}/_{V_0}\right)} \quad (4)$$

where $\Delta\mu_s$ is the change in steady state friction. Positive values of (a - b), indicate velocity-strengthening behavior, that favours stable sliding and fault creep. Negative values of (a - b), represent a velocity-weakening behavior, that is a requirement for the nucleation of slip instability (e.g. Dieterich and Kilgore, 1994; Marone, 1998a; Scholz, 2002).
In slide-hold-slide tests (e.g., Dieterich and Kilgore, 1994; Marone, 1998b; Carpenter et al., 2016) slip at constant velocity is followed by a hold period, t_h, usually ranging from 1 to 3000 s, during which sliding is halted and subsequently resumed. The amount of frictional healing, $\Delta\mu$, is measured as the difference between the peak friction measured upon re-shear after each hold and the pre-hold steady state friction, μ_{ss} (Fig. 10c). Frictional healing rate β is calculated as:

$$\beta = {}^{\Delta\mu}/_{\Delta\, log_{10}(t_h)} \quad (5)$$

It is important to note that in the Byerlee's (1978) dataset, friction measurements on pure clays show values well below the range he proposed. However, since fault zones are not composed of a single mineral phase, it was not clear how the mixing of strong and weak phyllosilicates would have affected fault strength.

Fault weakness resulting from fluid-assisted reaction softening originates at the scale of individual mineral grains and the ability to transmit this local effect to crustal scale faults is due to the interconnectivity of the phyllosilicate-rich networks. In other words, even a low percentage of weak mineral phases can induce significant fault weakening if the interconnectivity of the weak minerals is very high (Niemeijer et al., 2010). In order to capture the role of phyllosilicate interconnectivity in frictional properties of weak faults, instead of running traditional friction experiments on powdered fault rocks or on bare rock surfaces, we collected large blocks of foliated natural fault rocks and we cut them to form wafers 0.8–1.2 cm thick and 5 cm x 5 cm in area. This approach is akin to the one used to constrain the role of anisotropy in the strength of foliated metamorphic rocks (e.g. Shea and Kronenberg, 1993). The wafers were oriented so that they could be sheared in their in situ orientation, with foliation parallel to shear direction (e.g. Collettini et al., 2009b). We refer these as wafer experiments and microstructural analyses of these sheared samples show that:

a) the original foliated microstructure is preserved and 2) most of the deformation occurs by frictional sliding along the phyllosilicate-rich network with very limited cataclasis and grain-size reduction (Collettini et al., 2009b; Tesei et al., 2014; 2015; Smith et al., 2017). Other laboratory experiments on powdered fault rocks mixing different percentages of weak and strong mineral phases have documented a decrease in frictional strength with the addition of weak phases (Logan and Rauenzahn, 1987; Saffer and Marone, 2003; Takahashi et al., 2007; Crawford et al., 2008; Giorgetti et al., 2015). In particular an amount of weak phases greater than 40-50% of the rock volume results in significant fault weakness because it is sufficient to promote interconnectivity of the weak mineral phase through the entire experimental fault. This implies that when the weak mineral phases are abundant, laboratory experiments on powdered materials are similar to those conducted on wafers.

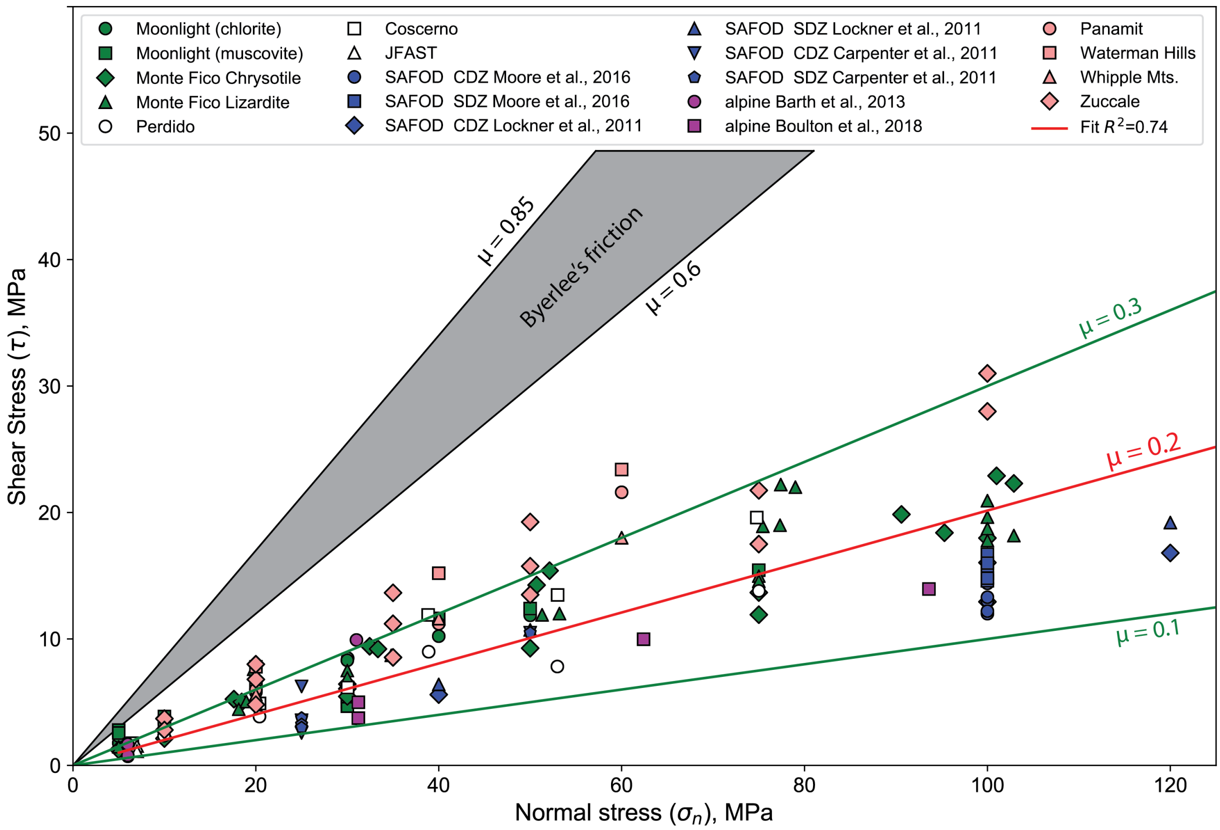

Figure 11. Frictional properties of phyllosilicate-rich faults. The dataset together with key references is summarized in Table 2.

Figure 11 shows normal stress vs. steady state shear stress from friction tests conducted on a large number of natural fault rocks rich in phyllosilicates, both wafers or powdered material with phyllosilicates percentages > 40%, under a wide range of experimental conditions (the dataset together with key references is summarized in Table 2). Each rock type plots along a line consistent with a brittle failure envelope and the frictional strength of all the tested materials is significantly below the Byerlee range. The average friction for all the tested material is 0.2 and at first approximation it is not influenced by the applied normal stress and temperature. The highest friction values, $0.25 < \mu_s < 0.43$ (pink colour in figure 11), are recorded for dry experiments on fault rocks from US detachment and from the Zuccale fault. These fault rocks are rich in clays and further weakening for these minerals is expected in the presence of water (Moore and Lockner, 2004, 2007). The tested phyllosilicate-rich fault rocks are characterized by a velocity strengthening behaviour, i.e. by positive *(a-b)*, which in general becomes more pronounced with increasing sliding velocity (Fig. 12a). Limited velocity weakening is reported systematically for the Southern Deforming Zone, SDZ, of the San Andreas at SAFOD for temperatures above 200 °C (Moore et al., 2016). The phyllosilicate-rich fault rocks also show near zero or limited healing rates, $-0.0023 < \beta < 0.003$ (Fig. 12b), that is significantly lower in comparison to granular materials such as quartz, $0.0082 < \beta < 0.0086$ (Marone, 1998b), quartz-feldspatic rocks $0.007 < \beta < 0.008$ (Carpenter et al., 2016) and calcite, $0.015 < \beta < 0.03$ (Carpenter et al., 2014).

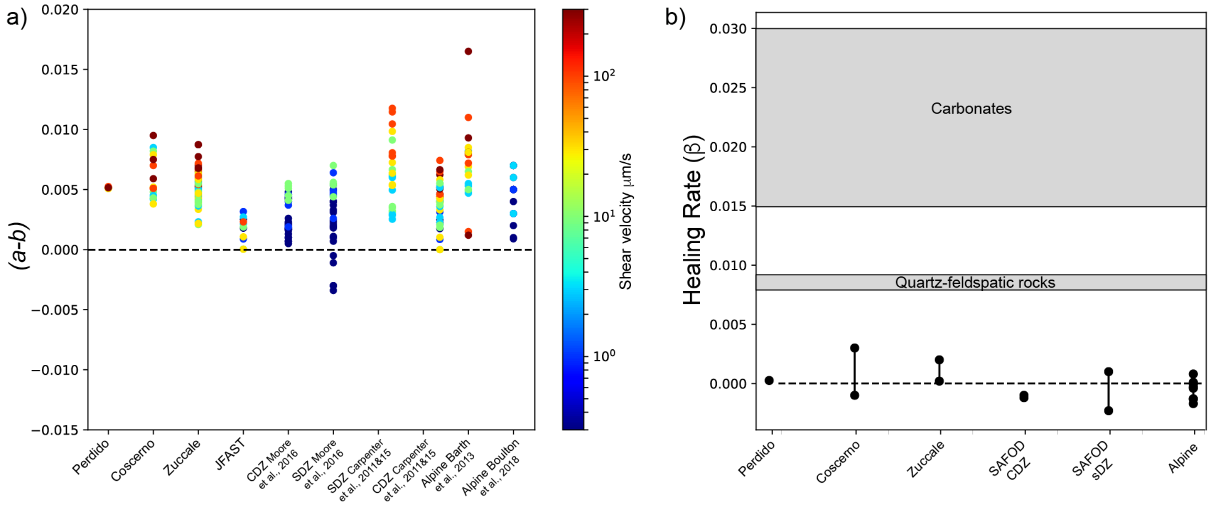

Figure 12. a) Velocity dependence of friction of phyllosilicate-rich faults. b) Healing properties of phyllosilicate-rich faults and comparison with quartz-feldspatic (Marone, 1998b; Carpenter et al., 2016) and calcite-rich (Carpenter et al., 2014) rocks. The dataset together with key references is summarized in Table 2.

4. Discussion

4.1 Structural and frictional heterogeneous crustal faults

In this review we have documented (via outcrops, microstructural and laboratory data) several examples of crustal faults characterized by either high (paragraph 2) or low (paragraph 3) strength. When considered all together these data point to the heterogeneous structural nature of crustal scale faults. In particular, a single crustal scale fault, tens of kilometres long, can be characterized by weak fault patches in zones where crustal fluids exerted a chemical role, facilitating the replacement of strong with weak mineral phases, and strong fault patches, where fluid-assisted reaction softening were not efficient, and crustal deformation was achieved predominantly by fragmentation, grain-size reduction and localization (Fig. 13). Similar heterogeneities have been documented on the grounds of frequency content of seismic waveforms of large and great earthquakes on subduction zone megathrusts (e.g. Lay et al., 2012), from high-resolution spatio-temporal behaviour of seismicity (e.g. Rubin et al., 1999; Waldhauser et al., 2004; Chiaraluce et al., 2007), and from geodetic imaging of active faults (e.g. Avouac, 2015 and references therein).

Strong faults or strong fault portions like those described in paragraph 2.3 (Fig. 13) form predominantly when granular mineral phases like quartz, feldspar, pyroxene, olivine, calcite and dolomite are dominant and repeated fault reactivation during the geologic fault history allows the development of fault rocks like gouge or cataclasite. The similarities in the internal structure of experimental and natural faults point to similarities in the deformation mechanism (Tchalenko, 1970). Large-displacement natural faults in granular materials show the evidence of localization along fault-parallel principal slip zones that are present at all depths through the seismogenic crust (Chester and Chester 1998; Sibson, 2003). Similarly, experimental faults show that at high-strains the deformation is accommodated along distinct fault-parallel shear zones, showing intense grain-size reduction (Logan et al., 1979; Beeler et al., 1996; Scuderi et al., 2017). Collectively these faults are characterized by a granular load-bearing microstructure with a frictional strength controlled by Byerlee's rule (e.g. Weeks and Tullis, 1985; Biegel et al., 1989; Marone, 1998a; Beeler et al., 1996; Verberne et al., 2010; Scuderi et al., 2013; Carpenter et al., 2014). With increasing strain, localization along a principal slip zone promotes the passage from rate strengthening to rate weakening behaviour (Beeler et al., 1996; Ikari et al., 2011; Scuderi et al., 2017) favouring the occurrence of a frictional instability. Frictional instabilities and associated earthquake slip along strong faults is documented by extreme localization, < 1 cm (e.g. Chester and Chester, 1998), and by the presence within the principal slip zone of fault rocks such as pseudotachylites, amorphous silica, polygonal nano-grains and decomposed minerals, produced by intense frictional heating during earthquake slip (paragraph 2.3 and Rowe and Griffith, 2014 for a comprehensive review). Following the earthquake, the fault regains strength during the interseismic period because in the fault zone the increase in grain contact quantity and quality promotes significant fault healing (Marone 1998b; Carpenter et al., 2014). Further restrengthening is also achieved via sealing and cementation processes (e.g. Sibson, 1992; Tenthorey et al., 2003).

Weak faults or weak fault patches result from fluid-assisted fault weakening (e.g. paragraph 3.2) that, during the long-term evolution of the fault, might promote the replacement of strong minerals (quartz, feldspar, olivine, pyroxene, calcite, dolomite) with weak phyllosilicates. Fluid flow into the fault zone is controlled by fracturing and therefore mineral replacement is a pervasive process within the fault core and the damage zone (Caine et al., 1996), and results in interconnected networks of weak phyllosilicates (Fig. 13).

The development of phyllosilicate networks formed via fluid-assisted reaction softening has been documented at different crustal depths, from shallow to deep crust, in different tectonic regimes and from different protoliths such as carbonates, dolostones, marly-carbonates, sandstones, ultramafic rocks and quartz-feldspatic dominated lithologies (paragraph 3.2 and table 1). Therefore, fluid-assisted reaction softening and associated frictional-viscous behavior can be considered as a general weakening mechanism within the seismogenic crust. The weakness resulting from frictional-viscous behaviour originates at the scale of individual mineral grains and the ability to transmit this local effect to crustal scale faults is due to the interconnectivity of the phyllosilicate-rich networks (e.g. Handy, 1990; Holdsworth, 2004; Collettini et al., 2009b). In laboratory experiments the low frictional strength of phyllosilicate-rich faults (Fig. 11), despite the fact that in some cases phyllosilicates are present in relatively small quantities, is mainly due to the interconnectivity of phyllosilicates and frictional sliding along the phyllosilicate foliae with very limited cataclastic processes (Fig. 13 green parts and details in Collettini et al., 2009b, Tesei et al., 2015). Friction measurements on these phyllosilicate-rich natural fault rocks show that friction is significantly below Byerlee's range and extends from $0.1 < \mu_s < 0.3$ (Fig. 11 and table 2). Frictional sliding along phyllosilicates-rich faults acts in concert with dissolution and precipitation processes (Niemeijer and Spiers, 2005; Gratier et al., 2011; Fagereng and den Hartog 2016), and in some cases dissolution accommodated deformation is so fast that it does not contribute to strength (e.g. regime 3 in Fig. 9). Frictional sliding along the phyllosilicate foliae (Fig. 13) is a sliding mechanism by which mineral surfaces tend to remain in complete contact and thus the real contact area does not evolve during the velocity steps or during the hold periods. The concept of contact saturation has been proposed as reasonable explanation for: a) the low *b* values during the velocity steps, promoting the velocity strengthening behaviour (Fig. 12a and Saffer and Marone, 2003; Ikari et al., 2009); and b) the absence of contact growth during the hold periods resulting in null healing rates (Fig. 12b and Carpenter et al., 2011; Tesei et al., 2012). Moreover, phyllosilicate coating strongly hinders rock cementation (Dewers and Ortoleva, 1991; Worden and Morad, 2000), which together with null frictional healing suggests the inability of these fault zones to store elevated elastic stress. Collectively, these frictional properties indicate that foliated, phyllosilicate-rich faults are weak, they accommodate deformation predominantly by aseismic creep, or to be more precise (see discussion below) a seismic rupture nucleation is inhibited within these lithologies, and remain weak over long time scales.

Our conceptual model of heterogeneous fault zone structures builds on the documented fluid-assisted reaction softening (paragraph 3) and on measured frictional properties of the resulting natural fault rocks. The model indicates that single faults might be characterized by strong and weak fault patches (Fig. 13). Further weakening, not considered in the conceptual model, is due to fluid pressure development during the fault activity related to the spatial and temporal variability of fault zone permeability (e.g. Cox, 1995; Miller et al., 1996; Faulkner et al., 2010; Sibson, 2017). For example, the development of continuous and phyllosilicate-rich networks strongly reduces fault permeability (Faulkner and Rutter, 2001; Ikari et al., 2009) facilitating the entrapment of crustal fluids deriving from various sources, e.g. meteoric fluids, mantle degassing, metamorphic reactions. Fluid pressure build-up and release during the deformation of phyllosilicate-rich faults is testified by the numerous hydrofracture systems contained within these shear zones (e.g. figure 4c, 5b, 7b and details in Chester et al., 1993; Imber et al., 1997; Collettini et al., 2006; Fagereng and Sibson, 2010; Sibson, 2017).

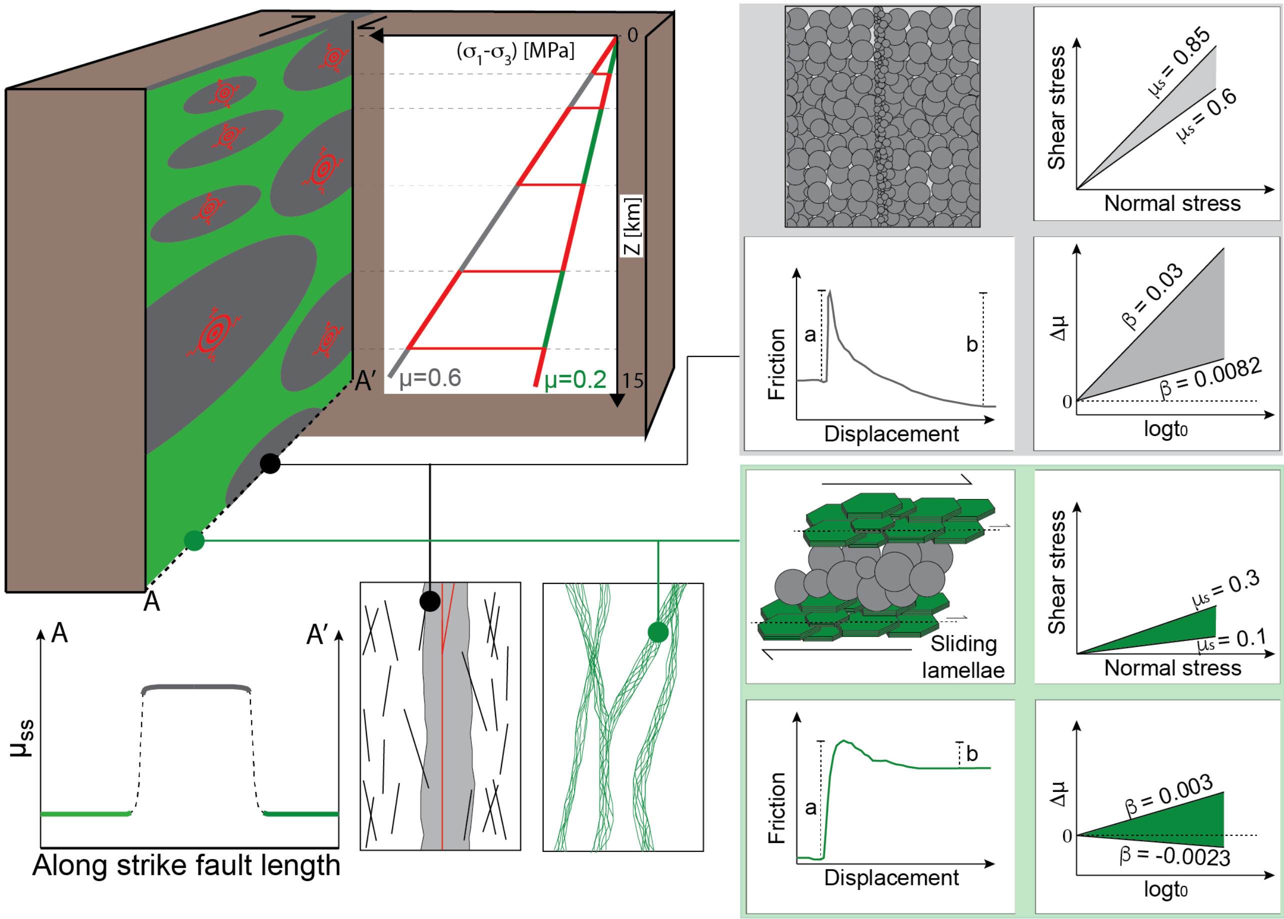

Figure 13. An integrated view of the internal structure and mechanical properties of heterogeneous faults. Schematic cross-section with strong (grey) and weak (green) fault portions (top-left). Steady state frictional strength profile along a vertical (red path) and sub-horizontal transect (A-A' bottom left). The strong portion (grey drawings and data) is mainly characterized by granular mineral phases affected by cataclasis and grain-size reduction with localization along a principal slipping zone (red-line), usually < 1 cm in thickness. This is associated to Byerlee friction, a velocity weakening behaviour, i.e. negative *(a-b)*, and high healing rates, β. In the weak fault portions (green drawings and data), distributed deformation occurs along interconnected phyllosilicate-rich networks and frictional sliding along phyllosilicate lamellae favours low friction, $0.1 < \mu_s < 0.3$, velocity strengthening, i.e. positive *(a-b)*, and very low healing rates.

4.2 Fault patches interaction during tectonic loading

During tectonic loading, weak and velocity-strengthening phyllosilicate-rich patches slip aseismically (Fig. 13 green parts). Creep along weak patches, facilitated also by pressure solution processes (e.g. Gratier et al., 2011; 2013; Fagereng and Den Hartog, 2016), allows for stress build up in the surrounding strong patches that during the interseismic phase remain locked due to high strength and healing rates (Fig. 13 grey parts). Analyses based on the assumption of stress and strain continuity across the weak-strong interface, and the observation that more strain is accommodated in the phyllosilicate-rich networks, support the idea of stress concentration in the strong lenses (e.g. Fagereng and Sibson, 2010; Fagereng and Den Hartog, 2016). When the shear stress overcomes the frictional strength of the strong, velocity-weakening patch an earthquake rupture might nucleate. In particular, frictional stick-slip instabilities occur when the fault-weakening rate with slip exceeds the maximum rate of elastic unloading, resulting in a force imbalance and fault acceleration (e.g., Scholz, 2002). If the condition to nucleate an earthquake instability is satisfied, the extent of the rupture will be controlled by the distribution and dimensions of velocity weakening vs. velocity strengthening fault patches, their state of stress, and the energy dissipated as seismic slip propagates (e.g. Boatwright and Cocco, 1996; Kaneko et al., 2010; Faulkner et al., 2011; Noda and Lapusta 2013; Avouac, 2015). Numerical models, reproducing the rich earthquake slip behaviours similar to that of natural faults, show that velocity strengthening fault patches tend to inhibit rupture propagation and the probability for a seismic rupture to propagate through a velocity-strengthening patch is related to the dimension and the frictional properties of the patch itself (Kaneko et al., 2010). However high-velocity friction experiments have shown that wet clay at low slip velocity is velocity strengthening, but at high slip velocity it weakens immediately or remains weak resulting in negligible fracture energy, making rupture propagation through clay-rich fault portions energetically very favourable (Faulkner et al., 2011). These experimental findings have been incorporated in numerical models and it has been shown that stable velocity strengthening segments may host seismic ruptures as result of mechanisms such as dynamic weakening (Noda and Lapusta, 2013). The transition from velocity strengthening at low sliding velocities (0.1-300 μm/s) to slip dominated by dynamic weakening approaching seismic slip rates (tens of cm per seconds) have been invoked (Wibberley et al., 2008; Faulkner et al., 2011; Noda and Lapusta, 2013) to explain the exceptionally large seismic slip, of as much as 50 m (Ide et al., 2011), in the shallower area of the Mw 9.0 Tohoku-Oki earthquake, where clay rich materials are present (Kameda et al., 2015).

Our conceptual model for the slip behaviour of heterogeneous faults (Fig. 13) agrees with the increasing number of geodetic observations showing that in the interseismic period some fault areas remain locked whereas others creep aseismically (e.g. Avouac 2015 and references therein). The model is also consistent with the analysis of the stress orientations in subduction zones suggesting that creeping subduction zones are weaker than locked ones (Hardebeck and Loveless, 2018). However, another source of structural heterogeneities, that is likely to play a key-role on fault slip behaviour and not considered in our simplified conceptual model, is due to fault roughness. For example in areas of extremely rugged subducting seafloor, creeping is the predominant mode of subduction (e.g. Wang and Bilek, 2014, but see also Scholz and Small, 1997 for a different interpretation) whereas mega-earthquakes rupture seems to be more likely along flat portions of the megathrust because the shear strength is more homogeneous, and hence more likely to be exceeded simultaneously over large areas (Bletery et al., 2016). Laser-based methods to map exposed fault surfaces have shown that large-slip faults are polished at small scales but contain elongated quasi-elliptical bumps and depressions at scales of a few to several meters (Renard et al., 2006; Sagy et al., 2007; Brodsky et al., 2011). This difference in geometry play an important role in the nucleation, growth, and termination of earthquakes (Sagy et al., 2007).

4.3 Some challenging topics

While there is an increasing amount of geological evidence that relates the occurrence of seismic ruptures within strong fault patches (some details and references in paragraph 2.3 and Rowe and Griffith, 2014 for an extensive review), clear geological evidence of rupture propagation through weak and velocity strengthening phyllosilicate fault patches at depth is rare. Recently Tarling et al., (2018) showed that within a tens to several hundreds of metres wide serpentinite shear zone, where the bulk deformation is accommodated by dissolution and precipitation processes plus frictional sliding along lizardite and fibrous chrysotile (i.e. processes indicative of aseismic creep) polished and localized fault surfaces in magnetite-rich patches contain high temperature reaction products, in the form of nanocrystalline olivine and enstatite, that likely formed during earthquake rupture propagation through the creeping serpentinites. Similar examples can be represented by the narrow and clay-rich slip zone within the Median Tectonic Line in Japan (Wibberley and Shimamoto, 2003) or by the pseudotachylyte-bearing fault rocks contained within the foliated argillaceous matrix of the Pasagshak Point thrust in Alaska (Rowe et al., 2011). On this research line, more field studies are required to better document the occurrence of seismic rupture propagation within rate-strengthening fault rocks. Furthermore, future laboratory tests should aim at reproducing experimental faults composed of both rate-strengthening and rate-weakening fault patches (e.g. Corbi et al., 2017 in rock analogues) to better illuminate the interaction between weak and strong fault patches.

The observations on fault zone structure and frictional processes presented here suggest that fault creep and slow slip might play an important role in the earthquake preparatory phase. Although distinguishing the processes operating during the earthquake nucleation still remains a great challenge (e.g. Gomberg et al., 2018) and some earthquakes show no evidence of aseismic slip in the earthquake preparatory phase, at least in the absence of measurements in the near field of the hypocentre (Ellsworth and Bulut, 2018), other evidences suggest the presence of slow-slip phenomena in or nearby the fault patch hosting the future mainshock. The foreshock activity of the Mw 9.0 Tohoku-Oki earthquake indicates that two sequences of slow slip transients propagated toward the initial rupture point of the earthquake (Kato et al., 2012). For the Mw 8.1 Iquique earthquake, more than 1 m of aseismic slip has been documented in the 15 days preceeding the event, in the same area where the mainshock occurred (Ruiz et al., 2014). High-resolution seismological data have shown that an Mw 3.7 earthquake in Alaska was preceded by a slow slip phase that accelerates into a fast rupture (Tape et al., 2018). Fluid injection experiments on laboratory and natural faults reveal a similar phase of sustained aseismic creep that increases shear stress beyond the pressure front and promote earthquake triggering (Cappa et al., 2019). Collectively, these observations significantly renew the interest in precursory signals and although the detection of reliable earthquake precursors remains an open issue, now there is some cause for optimism. The structural and frictional heterogeneous nature of crustal faults presented in this manuscript suggests that creep and slow slip play a fundamental role in the deformation of crustal faults including the earthquake preparatory phase. Preparatory processes such as slow slip and seismicity migration before large earthquakes can be monitored with a combination of seismic and geodetic observations, and the physics of these processes can be constrained by structural geology and laboratory experiments, and implemented in numerical models. Innovative and original data sets will be needed to establish whether and if the observation of such signals before the mainshocks repeats in time, then leading to a reliable contribution to the forecast.

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Table 1. Examples of fault zone structures characterized by interconnected networks of phyllosilicates formed during the fault activity.

Fault	Depth	Protolith	Phyllosilicates	References
Carboneras, Spain	2-4 km	Mica schists	Chlorite, illite	Faulkner et al., 2003; Solum and van der Pluijm, 2009
San Andreas at SAFOD, US	3 km	Serpentinites	Smectite clays (saponite)	Schleicher et al., 2010; Holdsworth et al., 2011
Midian Tectonic Line, Japan	5-10 km	Granitoids	Chlorite, muscovite	Jefferies et al., 2006
North Anatolian Fault, Turkey	< 5 km	Dolomite, quartz & calcite rich rocks	Talc, kaolinite & chlorite.	Kaduri et al., 2017
Livingstone fault, New Zealand	300–350 °C	Ultramafic rocks	Lizardite, chrysotile	Tarling et al., 2018
Rodeo Cove, California	8-10 km	Basalts	Chlorite	Meneghini & Moore, 2008
Thrusts in the Apennines	1-4 km	Marly limestone	Illite, smectite	Tesei et al., 2013
Perdido thrust Pyrenees	6-7 km	Limestones and sandstone	Illite, chlorite	Lacroix et al., 2011
Chrystalls Beach mélange New Zealand	T < 300°C	Sandstone and metabasalts	Illite, muscovite	Fagereng and Cooper, 2010; Fagereng and Sibson, 2010
M. Fico thrust Italy	T < 300°	Ultramafic rocks	Lizardite, chrysotile & polygonal serpentine	Viti et al., 2018; Tesei et al., 2018
Wasatch and Dixie valley faults, US	< 10 km	Quartz-feldspatic rocks	Muscovite, chlorite & clays	Bruhn et al., 1994
Zuccale fault Italy	4-6 km	Dolostone	Talc & smectite	Viti and Collettini., 2009
Black Mountains Detachment, California	From > 3 km to shallow	Carbonate, siliceous gneisses quartz-feldspatic basement, volcanic rocks	Illite, chlorite, smectite & saponite	Hayman, 2006
Gubbio Fault, Apennines	< 4 km	Limestone and	Illite & smectite	Bullock et al., 2014

		marly limestones		
Err Nappe, Switzerland	< 300°C	Quartz-feldspatic rocks	60% chlorite illite in the fault core	Manatschal, 1999

Table 2. Frictional properties of phyllosilicate-rich faults; w = wafer experiments, p powder experiments.

Fault & weak minerals	Normal stress, Temperature	Friction	(a-b) 0.01-300 μm/s	Healing rate β	Reference
Moonlight (NZ), chlorite	5-75 MPa T = 25° C, wet	0.24 w			Smith et al., 2017
Moonlight (NZ), muscovite	5-50 MPa T = 25° C, wet	0.19 w			Smith et al., 2017
M. Fico Thrust (ITA) chrysotile & Pol. serpentine	20-100 MPa T = 170° C, wet	0.19 p			Tesei et al., 2018
M. Fico Thrust (ITA) chrysotile & Pol. serpentine	5-100 MPa T = 25° C, wet	0.15 p			Tesei et al., 2018
M. Fico Thrust (ITA) lizardite	20-100 MPa T = 170° C, wet	0.18 p			Tesei et al., 2018
M. Fico Thrust (ITA) lizardite	5-100 MPa T = 25° C, wet	0.18 p			Tesei et al., 2018
Perdido thrust (SPA) illite, chlorite	10-75 MPa T = 25° C, wet	0.17 w	0.0050-0.0052	$\beta \approx 0.0008$	Tesei et al., 2015
Coscerno thrust (ITA) smectite	10-100 MPa T = 25° C, wet	0.27 w	0.0038-0.0095	$-0.001 < \beta < 0.003$	Tesei et al., 2014
Zuccale normal fault (ITA), talc & smectite	10-150 MPa T = 25° C, room humidity	0.25-0.31 w	0.0021-0.0087	$0.0002 < \beta < 0.002$	Collettini et al., 2009b; Collettini et al., 2011; Healing Tesei et al., 2012
J-FAST (JPN), smectite	5-7 MPa T = 25° C, wet	0.2-0.26 p	0-0.003		Ikari et al., 2015
SAFOD CDZ (USA) saponite	40-200 MPa, 25-250°C, fluid pressure	0.1-0.17 p	0.0007- 0.0067		Lockner, 2011 Moore et al., 2016
SAFOD SDZ (USA) saponite	40-200 MPa, 25-250°C, fluid pressure	0.15-1-19 p	0.0011-0.007 v. weakening (a few) at T>200°C		Lockner, 2011 Moore et al., 2016
SAFOD CDZ (USA) smectite	25-50 MPa, 25°C, fluid pressure	0.1-0.25 p	0.002–0.008	$-0.0012<\beta<-0.001$	Carpenter et al., 2011 Carpenter et al., 2015
SAFOD SDZ (USA) smectite	25-50 MPa, 25°C, fluid pressure	0.12-0.15 w	-0.0015–0.011	$-0.0023<\beta<0.001$	Carpenter et al., 2011 Carpenter et al., 2015
Alpine Fault (NZ) Gaunt Creek, Illite, chlorite	6 MPa, 25°C, fluid pressure	0.28 p	0.0076–0.0153	$-0.0013<\beta<-0.0004$	Barth et al., 2013
South Alpine Fault (NZ) Martyr River, illite, chlorite.	31 MPa, 25°C, fluid pressure	0.32 p	0.0035–0.0089	$-0.0013<\beta<0.0008$	Barth et al., 2013
South Alpine Fault (NZ) McKenzie Creek, smectite, chlorite, liz.,	6 MPa, 25°C, fluid pressure	0.13 p	0.0051–0.0098	$-0.0017<\beta<0.0001$	Barth et al., 2013
South Alpine Fault (NZ) Hokuri Creek,	6 MPa, 25°C, fluid pressure	0.12 p	0.0049–0.0085	$-0.0003<\beta<-0.00004$	Barth et al., 2013

smectite, chlorite.					
South Alpine Fault (NZ) Hokuri Creek Smectite, lizardite.	31-94 MPa, 25-210 C°, fluid pressure	0.12-0.16 p	0.001-0.007		Boulton et al., 2018
Whipple Mts. Detachment (USA), clay	20-60 MPa, 25 °C, room humidity	0.29-0.30 p			Haines et al., 2014
Panamint Detachment (USA), clay	20-60 MPa, 25 °C, room humidity	0.28-0.38 p			Haines et al., 2014
Waterman Hills detachment (USA), clay	20-60 MPa, 25 °C, room humidity	0.38-0.43 p			Haines et al., 2014