

## Beyond Byerlee friction, weak faults and implications for slip behavior

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# 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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34	<b>Highlights</b>	
35	Fault friction drops from 0.6 to 0.2 when interconnected networks of phyllosilicates are	
36	present.	
37	Fluid-assisted reaction softening is a general weakening mechanism within the seismogenic	
38	crust.	
39	The integration of geological data and results from laboratory experiments depicts structural	
40	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.  $0.01 \mu\text{m/s} < v < 100 \mu\text{m/s}$ , 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



102 homogeneous and isotropic; b) one principal stress is vertical since the Earth's surface is a  
103 free surface; and c) brittle faults form in accordance with the Coulomb criterion for shear  
104 failure:

105

$$106 \quad \tau = C + \mu_i \sigma'_n = C + \mu_i (\sigma_n - P_f) \quad (1)$$

107

108 where  $\tau$  and  $\sigma_n$  are, respectively, resolved shear and normal stresses on the failure plane,  $C$  is  
109 the cohesive strength,  $\mu_i$  is the coefficient of internal friction, ( $C$  and  $\mu_i$  are rock material  
110 properties) and  $P_f$  is the pore fluid pressure. However with increasing displacement the  
111 cohesive strength of a fault is very small compared to the shear and normal stresses to be  
112 neglected and the internal friction coefficient is replaced by a sliding friction coefficient,  $\mu_s$ ,  
113 resulting in the Amontons' law:

114

$$115 \quad \tau = \mu_s (\sigma_n - P_f) \quad (2)$$

116

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

123

## 124 *2.2 Supporting evidence for Byerlee friction*

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

130 The first evidence is produced by the dip distribution of moderate-to-large ruptures in  
131 extensional and compressional environments that seem to be controlled by frictional fault  
132 reactivation theory under Byerlee friction (Sibson, 1985; Collettini and Sibson, 2001). For the  
133 two-dimensional case in which an existing fault containing the  $\sigma_2$  axis lies at a reactivation  
134 angle,  $\theta_r$ , to  $\sigma_1$ , equation 2 may be written in term of principal stresses as:

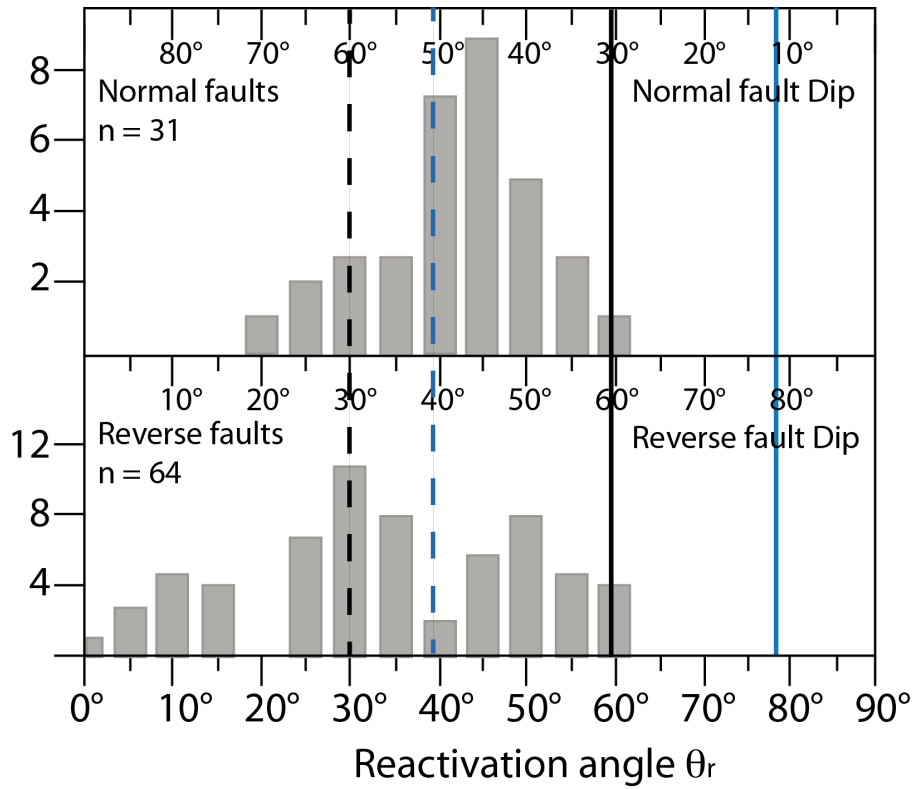
135

$$136 \quad R = \frac{(\sigma_1 - P_f)}{(\sigma_3 - P_f)} = \frac{(1 + \mu_s \cot \theta_r)}{(1 - \mu_s \tan \theta_r)} \quad (3)$$

137

138 defining the relative ease of reactivation for faults at varying angles to  $\sigma_1$  (Sibson, 1985),  
 139 where  $\sigma_1$ ,  $\sigma_2$  and  $\sigma_3$  are the maximum, intermediate and minimum principal stresses  
 140 respectively. The optimal orientation for frictional reactivation is given by  $\theta_r^* = 0.5 \tan^{-1}$   
 141  $(1/\mu_s)$ . As  $\theta_r$  increases or decreases away from this optimal position, the stress ratio required  
 142 for reactivation increases. Frictional lock-up ( $R \rightarrow \infty$ ) occurs when  $\theta_r^* = 2 \theta_r \tan^{-1} (1/\mu_s)$ . For  
 143 Byerlee's friction range, optimal reactivation occurs when  $\theta_r = 25^\circ - 30^\circ$  and frictional lockup is  
 144 expected at  $\theta_r = 50^\circ - 59^\circ$ .

145 Figure 1 shows the dip distribution of reverse and normal fault ruptures obtained from focal  
 146 mechanisms of shallow, intracontinental earthquakes ( $M > 5.5$ ; slip vector raking  $90^\circ \pm 30^\circ$  in  
 147 the fault plane) where the rupture plane is unambiguously discriminated. On the same figure,  
 148 under the assumption of vertical and horizontal  $\sigma_1$  trajectories for extensional and  
 149 compressional regimes, respectively, the dip distributions,  $\delta$ , of ruptures are plotted also as  
 150 functions of the reactivation angle,  $\theta_r$ . The cut-off at  $\theta_r \approx 60^\circ$  for both varieties of dip-slip faults  
 151 is consistent with frictional lockup for  $\mu_s = 0.6$ , at the bottom of the Byerlee range. Lower  
 152 coefficients are possible, but  $\mu_s = 0.6$  is also consistent with the dominant peak at  $\theta_r \approx 30^\circ$  in  
 153 the reverse-slip distribution representing the optimal orientation for reactivation. The  
 154 absence of this peak for normal faults is explicable by: a) the different reactivation curves for  
 155 the two faulting modes with a more acutely defined minimum at optimal orientation for  
 156 reverse faults than for normal faults (Collettini and Sibson, 2001); and b) the significant  
 157 number of normal faults ruptures obtained from earthquakes occurring in central Italy and  
 158 nucleating on faults forming with dip angles of  $45^\circ$  as ductile shear zones following planes of  
 159 maximum shear stress (Collettini et al., 2009a). The dataset and frictional analysis presented  
 160 in figure 1 provide strong evidence for seismogenic faults possessing sliding friction  
 161 coefficient similar to those measured in the laboratory by Byerlee, but see also contradictory  
 162 observations for earthquakes occurring in the oceanic lithosphere (Craig et al., 2014; Tesei et  
 163 al., 2018) or microseismicity along low-angle normal faults (Collettini, 2011). The fact that  
 164 some faults are reactivated close to frictional lock-up also implies that localized fluid  
 165 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,  $\theta_r = 90^\circ - \delta$ , and for reverse faults,  $\theta_r = \delta$ , where  $\delta$  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).

200

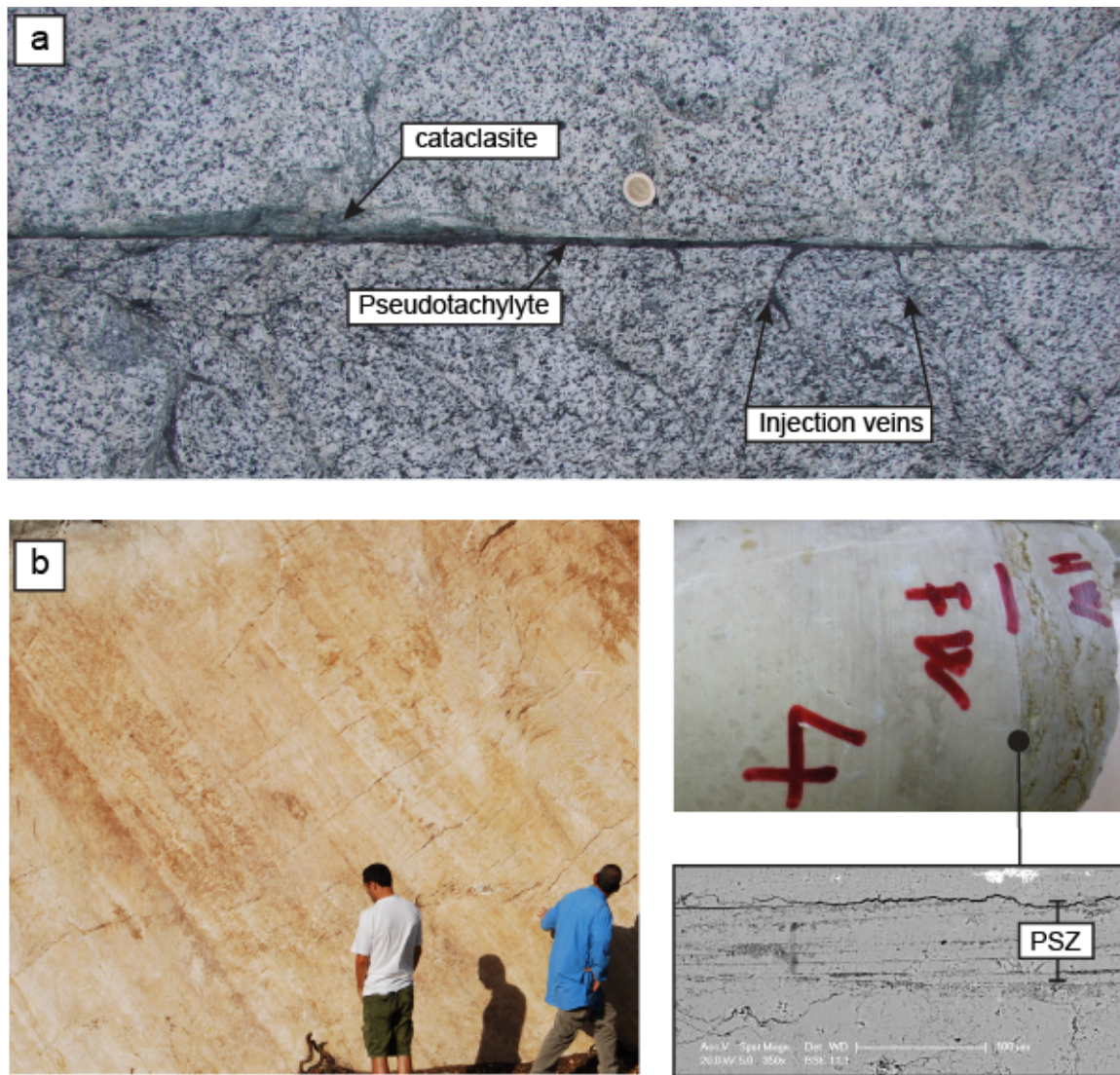
### 201 *2.3 Structural and mechanical characteristics of strong faults*

202 The localization of strain along crustal faults produces a fault structure that generally consists  
203 of a fault core, where most of the deformation is localized, surrounded by a damage zone  
204 formed by distributed fractures and subsidiary, small displacement faults (e.g. Chester et al.,  
205 1993). Cataclastic deformation (brittle fragmentation by macroscopic fracturing and grain  
206 comminution) intensifies toward the fault core that consists of one or more tabular zones of  
207 ultracataclasite, within which bands of intense grain-size reduction define principal slip zones  
208 (e.g. Sibson, 1977; Chester and Chester, 1998). A comprehensive characterization of fault zone  
209 structure is beyond the scope of the present manuscript and it is well described in several  
210 excellent review papers such as Caine et al., (1996), Ben-Zion and Sammis, (2003), Sibson  
211 (2003), Wibberley et al., (2008) and Faulkner et al., (2010). Here we try to link structural and  
212 mechanical data in order to summarize the main mechanical characteristics of strong faults.

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

224 Chester, 1998) to explain the weakness of the San Andreas: in this view the fault would be  
225 statically strong yet dynamically weak (Rice et al., 2009). Some faults that cross-cut silica-rich  
226 sediments, like the Corona Heights fault, in San Francisco, show a mirror-like finish due to the  
227 presence of 1-3 mm thick zone of vitreous silica formed during earthquake slip (Kirkpatrick et  
228 al., 2013). Faulting within sandstones results in one or more through-going slip surfaces  
229 where the great part of the displacement is accommodated by quartz grain-size reduction and  
230 localization along principal slip surfaces (Shipton and Cowie, 2001). The Pretorius fault in Tau  
231 Tona mine, South Africa, during its Archaean activity, experienced multiple slip events along a  
232 quartz-rich principal slip surface and in 2004 it was reactivated by a  $M = 2.2$  earthquake. The  
233 mapped earthquake rupture at 3.6 km of depth reveals that the seismic slip produced 1-5 mm  
234 thick fault gouge along four quasi-planar segments of the ancient fault-zone (Heesakkers et al.,  
235 2011). Faulting within massive carbonates (Fig. 2b) is characterized by localization along sub-  
236 parallel slipping zones where the deformation is localized along very thin ( $< 500 \mu\text{m}$ ) zones  
237 bounded by mirror-like slipping surfaces (De Paola et al., 2008; Fondriest et al., 2013; Siman-  
238 Tov et al., 2013; Collettini et al., 2014). In some of these carbonate-bearing faults, calcite  
239 crystals exhibiting localized disaggregation together with a high concentration of vesicles  
240 indicate thermal decomposition during past earthquakes (e.g. Rowe et al., 2012; Collettini et  
241 al., 2013). In other structures nanograins texture with polygonal grain boundaries suggest  
242 superplastic deformation of carbonates during earthquake-slip (De Paola et al., 2015).

243



**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  $\mu_s = 0.6 - 0.85$  and hydrostatic fluid pressure. These structures represent strong faults since the differential stress,  $(\sigma_1 - \sigma_3)$ , or the shear stress required for their reactivation is quite high. At 10 km of depth  $(\sigma_1 - \sigma_3) > 100$  MPa and it increases significantly from extensional to compressional environments (Sibson, 1974); for example in

260 the KTB borehole the differential stress at 9 km of depth is about 170 MPa (Townend and  
261 Zoback, 2000).

262 However, several observations cast doubt on the fact that deformation within the crust is  
263 exclusively controlled by strong faults and amongst these, we report the two that we consider  
264 the most relevant. First, there is an important number of faults that experience reactivation  
265 although they are severely misoriented within the regional stress field. Examples of these  
266 structures are given by the San Andreas fault in a strike slip regime (Zoback et al., 1987;  
267 Carpenter et al., 2011; Lockner et al., 2011), low-angle normal faults in extensional  
268 environments (Wernicke, 1981; Collettini 2011) and sub-horizontal thrusts in compressional  
269 regimes (Price, 1988; Davis et al., 1983; Suppe 2007; Tesei et al., 2015). Reactivation along  
270 these structures is possible only following significant fault weakening that can be achieved by  
271 either an increase in fluid pressure, or a reduction in friction coefficient or a combined effect  
272 (e.g. Hubbert and Rubey, 1959; Rice, 1992; Faulkner et al., 2006; Suppe 2007; Tesei et al.,  
273 2015). Second, in the last 15 years, geophysical observations combined with numerical  
274 models have shown that aseismic creep is common and sometimes prevalent within the  
275 seismogenic layer (e.g. Avouac, 2015), and a continuum spectrum of fault slip behaviour,  
276 including slow slip phenomena, can be present at all depths of crustal faults (e.g. Bürgmann,  
277 2018). This richness in fault slip behaviour for crustal faults is difficult to be captured only by  
278 strong faults with Byerlee friction.

279

### 280 *3.1 Weak-fault structure*

281 In marked contrast to strong faults, there is a significant number of crustal structures showing  
282 distributed deformation along interconnected, anastomosing shear zones rich in  
283 phyllosilicates. The geometry and internal fabric of these shear zones strongly resembles, and  
284 is sometimes derived from, ductile shear zones of high metamorphic grade (e.g. Berthé et al.,  
285 1979; Platt and Vissers, 1980). In this paragraph we will present some well-documented field  
286 examples of phyllosilicate-rich crustal faults (Fig. 3 and Table 1).

287

288 For strike-slip faulting, the Carboneras fault in Spain (Fig. 3a) is characterized by 1 km thick  
289 fault core consisting of foliated and interconnected networks of phyllosilicate-rich zones  
290 (Faulkner et al., 2003; Rutter et al., 2012). These anastomosing networks are up to 50 m thick  
291 and are rich in chlorite and illite derived from a mica schist protolith (Solum and van der  
292 Pluijm, 2009). The creeping section of the San Andreas fault at SAFOD consists of multiple  
293 fault strands, made of foliated serpentinite and smectite clays (Holdsworth et al., 2011), that



294 are several meters wide and creep simultaneously (Zoback et al., 2010). Some segments of the  
295 Median Tectonic Line in Japan consist of several meters thick, foliated fault rocks rich in  
296 chlorite (Wibberley and Shimamoto, 2003; Jeffereis et al., 2006). During the final activity of  
297 the fault at shallow crustal levels the deformation is concentrated along clay rich shear zones  
298 (Wibberley and Shimamoto, 2003). Exhumed shallow portions of the North Anatolian fault  
299 are characterized by hundreds of meter thick faults where the deformation is mainly  
300 accommodated within sub-parallel shear-zones rich in talc, kaolinite and chlorite (Kaduri et  
301 al., 2017). The Livingstone fault zone in New Zealand (Tarling et al., 2018) is dominated by a  
302 serpentinite *mélange* tens to several hundreds of metres wide, in which a pervasive  
303 anastomosing fabric surrounds pods of more competent material (e.g., metasediments,  
304 rodingite, massive serpentinite) ranging from tens to hundreds of metres in size. The foliated  
305 serpentinite shear zone is mostly made of fibrous serpentine and lizardite, consistent with the  
306 estimated ambient temperature during shearing of 300–350 °C.

307

308 For thrusts faults there are plenty of examples of foliated phyllosilicate-rich shear zones.  
309 Notable examples may include the classic phyllonites associated to the Moine thrust zone,  
310 possibly formed by retrogression and shearing of mylonitic gneissose protoliths under lower  
311 greenschist facies conditions (McClay and Coward, 1981, Wibberley, 2005). Other examples of  
312 phyllonitic fault rocks have been documented in the Karakoram fault zone in the Himalayas  
313 (Wallis et al., 2015) or along the Red River shear zone in the Yunnan Province in China (e.g.  
314 Wintsch and Yeh, 2013). The Perdido thrust in the Pyrenees consists of a several meters thick  
315 and foliated shear zone, rich in illite and chlorite (Lacroix et al., 2011), separating more  
316 competent lithologies in the hanging-wall (limestones) and footwall (turbidites sandstone).  
317 The Monte Fico thrust in the Elba Island in Italy (Fig. 3b) is a c. 200 m thick shear zone formed  
318 by competent lenses of retrograde pseudomorphic serpentinite surrounded by an  
319 anastomosing network of foliated serpentinites (Viti et al., 2018). Several tectonic *mélanges*  
320 around the world, thought to constitute exhumed analogues of subduction channels, are  
321 constituted by anastomosing shear zones enriched in phyllosilicates enveloping lenses of  
322 more competent lithologies (e.g. Cowan, 1974; Byrne et al., 1988; Cloos and Shreve, 1988,  
323 Kimura et al., 2012; Morley et al., 2017; Rowe et al., 2013). For instance, the Rodeo Cove  
324 thrust, near San Francisco, consists of a 200 m thick shear zone with a foliated fabric rich in  
325 chlorite that formed at about 8-10 km of depth (Meneghini and Moore, 2008). The Chrystalls  
326 Beach Complex accretionary *mélange* of New Zealand (Fagereng and Sibson, 2010) consists of  
327 competent lenses of chert, sandstone and metabasalts surrounded by an interconnected



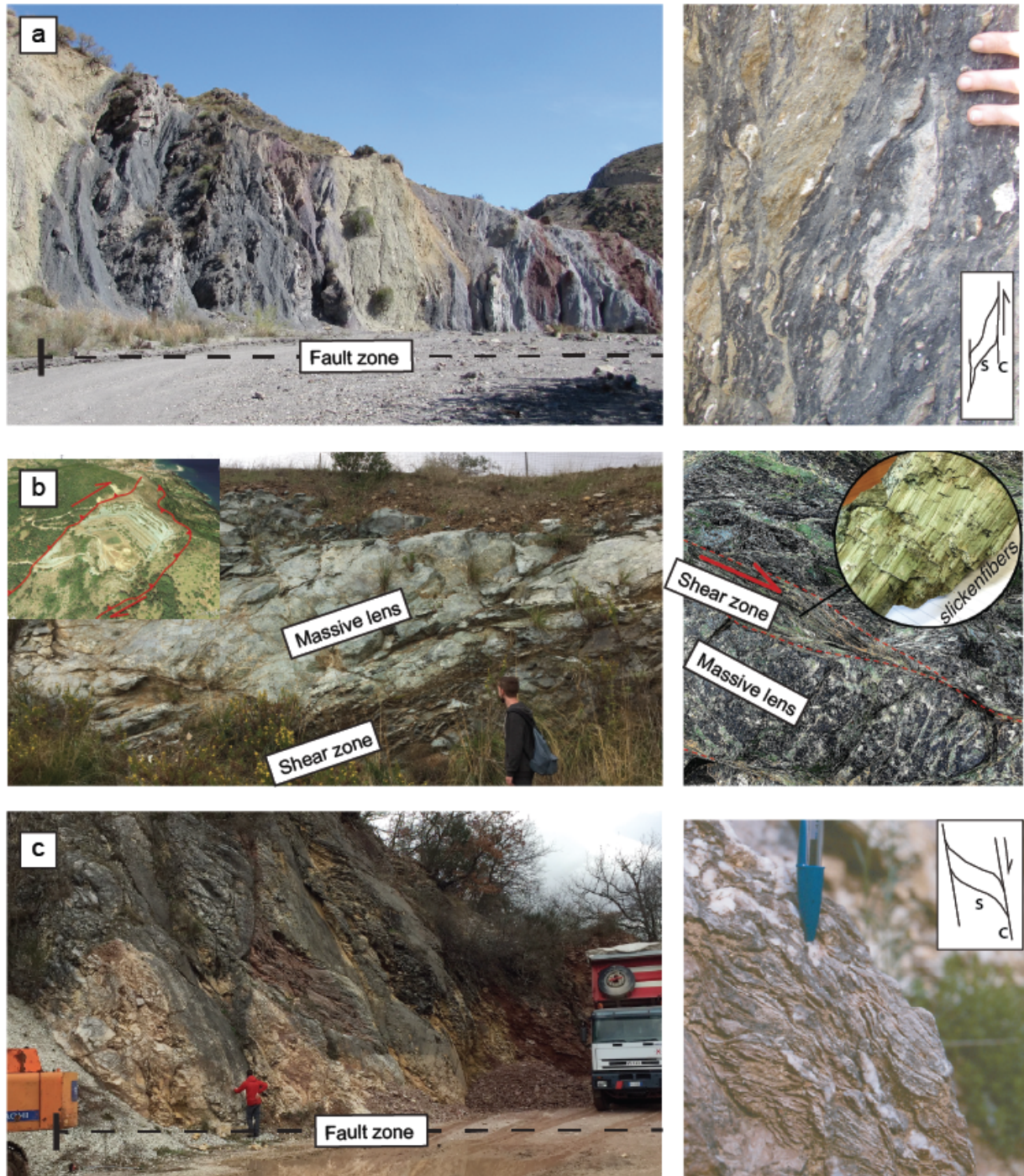
328 network of phyllosilicates, i.e. illite-muscovite, developed at depth < 300°C (Fagereng and  
329 Cooper, 2010). At shallower crustal levels other examples of tectonic *mélange* are reported in  
330 Vannucchi et al., (2012 and references therein). In the Apennines of Italy, in marly limestones,  
331 shear zones formed at crustal depths < 4 km show a thickness ranging from several to  
332 hundreds of meters with concentration of illite and smectite within the interconnected and  
333 foliated network of the fault core (e.g. Koopman, 1983; Tesei et al., 2013).

334

335 Some large displacement extensional faults formed in the US within quartz-feldspathic rocks  
336 reveal a heterogeneous structural zonation with fault zone thickness ranging from < 10 m for  
337 the Mineral Mountains fault to locally about 20 m for segments of the Wasatch and Dixie  
338 Valley faults (Bruhn et al., 1994). The fault zone is made of fault breccia, fine-grained  
339 cataclasites and foliated zones rich in muscovite, biotite, chlorite and clays. Some portions of  
340 the Zuccale low-angle normal fault in Italy, consist of an up to 8 m thick fault core rich in talc  
341 and smectite (Collettini et al., 2011). The Black Mountains detachment in California shows  
342 deformation that is asymmetrically distributed, increasing upward from the footwall (Cowan  
343 et al., 2003). A well-defined slip zone separates hangingwall Quaternary conglomerates from  
344 fault rocks consisting of foliated fault breccia and fault gouge where weak mineral phases  
345 (illite, chlorite, smectite and saponite) are concentrated (Hayman, 2006). The Gubbio normal  
346 fault in the Apennines of Italy shows some segments characterized by foliated SC fabric from  
347 the metric (Fig. 3c) to the microscale. These segments form predominantly within marly  
348 carbonates and show the concentration of clays within the shear zone (Bullock et al., 2014).  
349 The Err Nappe detachment in Switzerland shows a fault core made of a continuous layer of  
350 black gouge, with thickness ranging from a few centimetres to some metres (Manatschal,  
351 1999). In the footwall the granitic host rock is affected by brittle fracturing associated with a  
352 complex vein system and towards the fault core fluid-assisted diffusion mass transfer  
353 processes occurring under lowermost greenschist facies conditions promoted the  
354 development of an SC fabric rich in chlorite and illite (Manatschal, 1999).

355

356



357

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

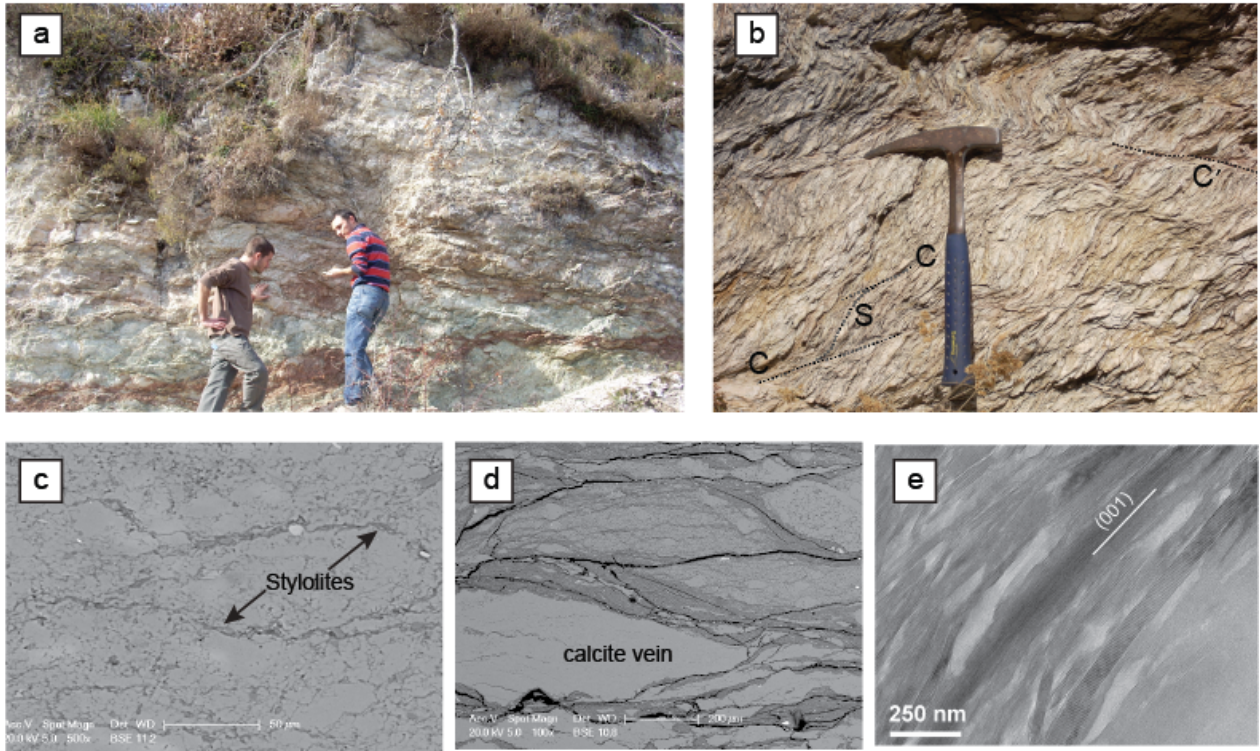
368

### 369 3.2 Reaction softening

370 Fault zone architecture exerts a primary control on fluid flow in crustal shear zones (Sibson,  
371 1992; Caine et al., 1996; Wibberley et al., 2008; Faulkner et al., 2010). The influx of fluids into  
372 fault zones can trigger two main types of weakening process that operate over different  
373 timescales. In the short term of the seismic cycle, crustal fluids can be trapped within low-  
374 permeability fault zones promoting the development of fluid overpressure, e.g. the  
375 mechanical weakening that reduces the effective normal stress (equation 2 and Hubbert and  
376 Rubey, 1959). During the entire fault history fluid circulation within shear zones might exert a  
377 chemical role facilitating the replacement of strong mineral phases with weak mineral phases,  
378 hence promoting reaction softening (e.g. Janecke and Evans, 1988; Bruhn et al., 1994; Evans  
379 and Chester, 1995; Wintsch et al., 1995; Manatschal, 1999; Imber et al., 1997; Wibberley,  
380 1999; Collettini and Holdsworth 2004; Schleicher et al., 2010; Warr et al., 2014). In the  
381 following we will review some examples of fluid assisted reaction softening generated in  
382 different protoliths and occurred at different crustal levels. Then we will integrate these  
383 observations in a general mechanism for fluid assisted fault weakening.

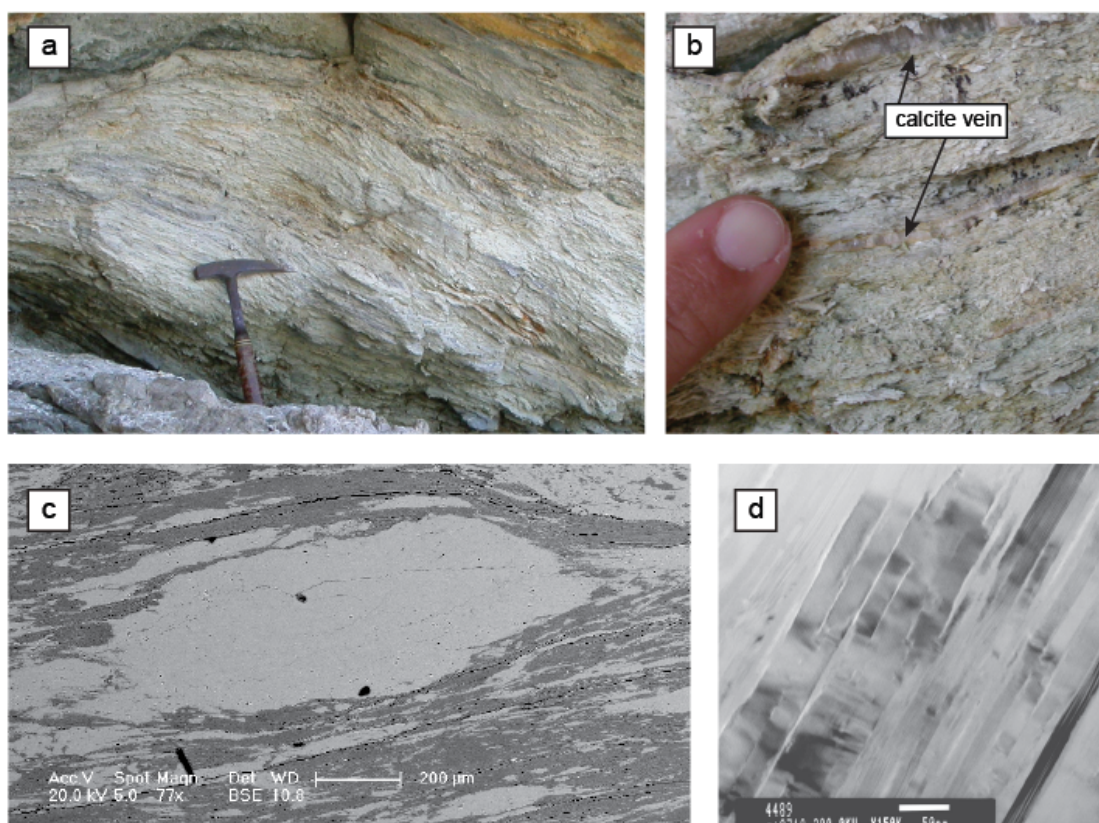
384 In the thrusts of the Apennines, when shear zones develop within marly limestones, faults  
385 with kilometric displacement show distributed deformation along thick (up to 200 m) shear  
386 zones (Fig. 4a and Tesei et al., 2013) characterized by SCC' fabric (e.g. Ramsay and Graham,  
387 1970; Berthé et al., 1979; Bos and Spiers, 2001 and Fig. 4b). The evolution of the fault zone  
388 structure, inferred from the analysis and comparison of small, intermediate and kilometric  
389 displacement faults, indicates that during the early stages of deformation dissolution of the  
390 carbonates favours the concentration of insoluble clay minerals within stylolitic surfaces (Fig.  
391 4c, Tesei et al., 2013; Gratier et al., 2013; Lacroix et al., 2015). With increasing deformation  
392 the stylolites evolve into an interconnected foliation (Fig. 4d, see also Gratier and Gamond,  
393 1990), formed by smectitic clays in nanosized (001) lamellae (Fig. 4e) with preferred  
394 orientation parallel to the local slipping surface (Viti et al., 2014). Foliation parallel veins (Fig.  
395 4c) with crack-and-seal texture, suggest that the low permeability and the anisotropy of clay  
396 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  $\approx 20$  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 are 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.

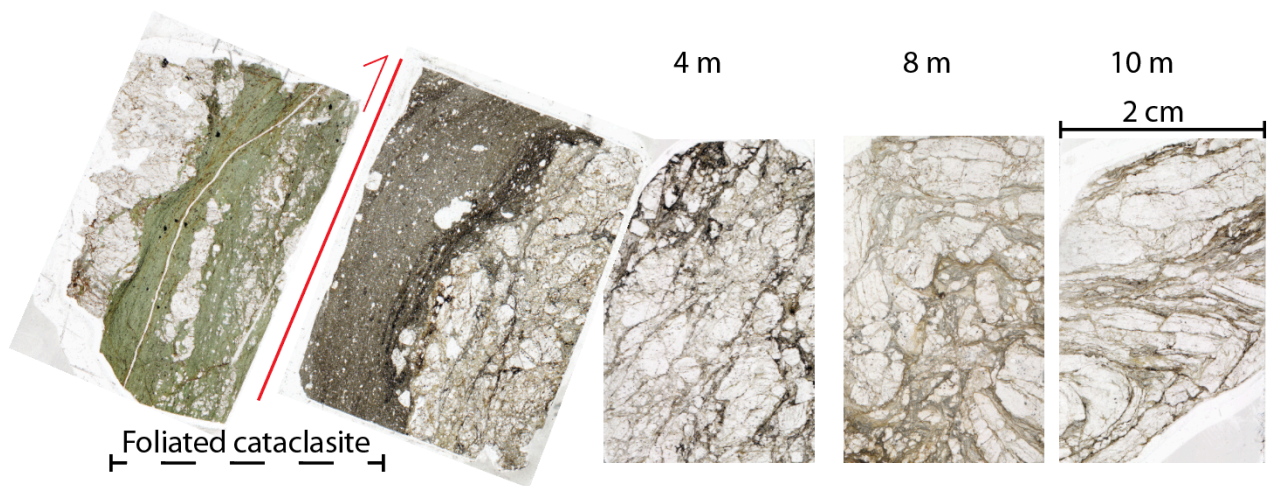
Along the Zuccale low-angle normal fault (Elba island, Italy) during the initial phase of deformation, fracturing of a dolomitic protolith favoured the influx of silica-rich fluids into the fault zone. In the low-strain domains fluids interacted with the fine-grained cataclasite promoting dissolution 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).

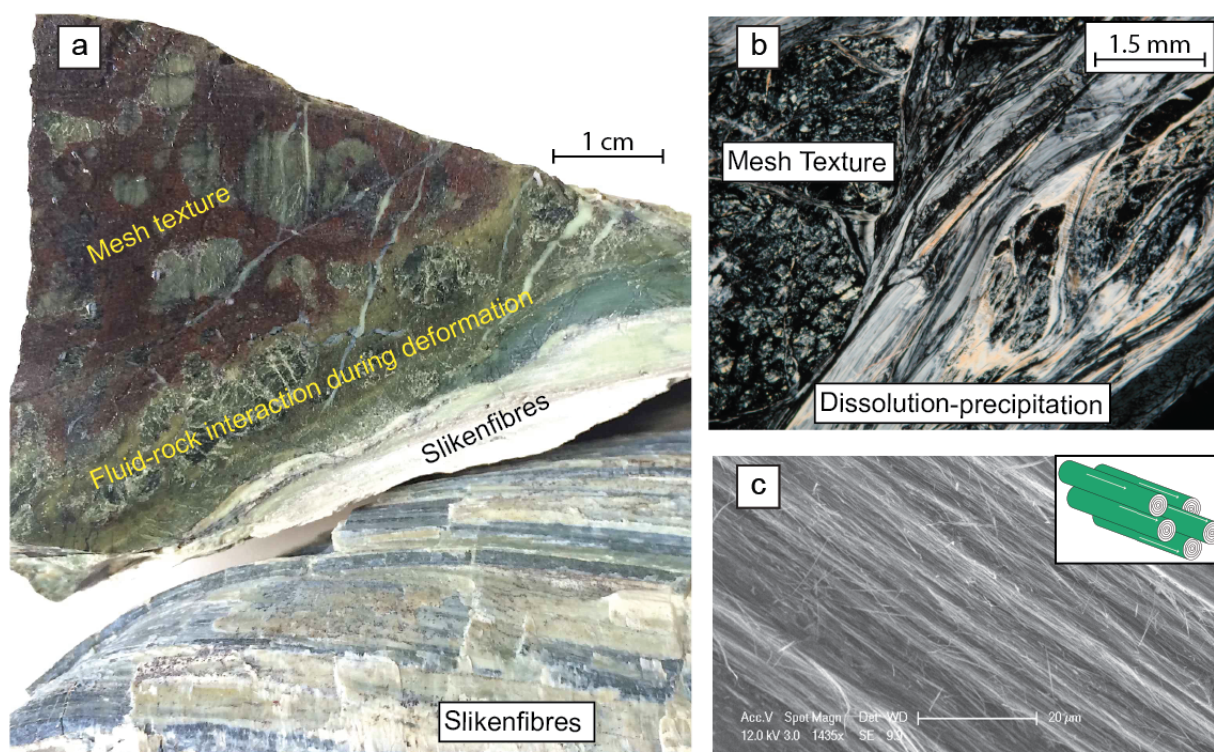
The Moonlight Fault in southern New Zealand consists of a 20 m thick shear zone formed within grey and green schists. Microstructural studies suggest that in the early stages of the fault activity brecciation promoted grain-scale dilatancy favouring the influx of hydrous fluids into the fault zone (Fig. 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-feldspathic 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 (Plummer 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).

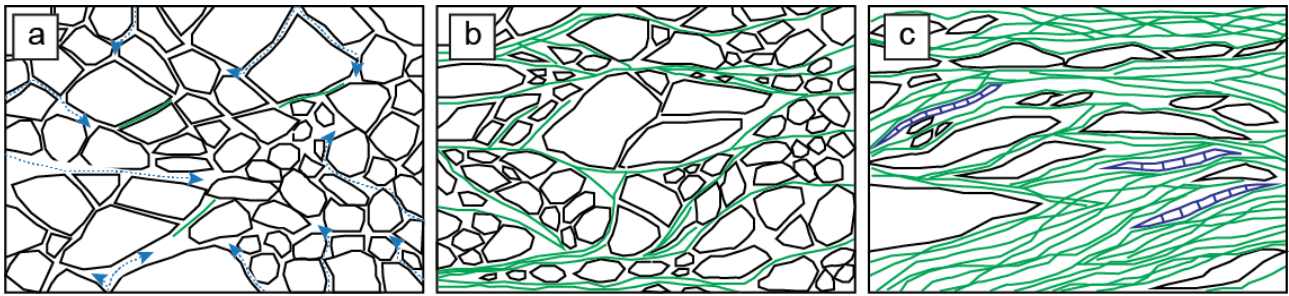


464

465 **Figure 7.** Monte Fico fault in Elba island, Italy: fault weakening in ultramafic rocks. a) Hand  
 466 sample showing the transition from the mesh texture to the slikenfibres coating the shear  
 467 zones. b) Mesh texture and slikenfibres at the optical microscope. c) Parallel fiber axis of  
 468 fibrous serpentines, SEM image (details in Tesei et al., 2018; Viti et al., 2018).  
 469

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

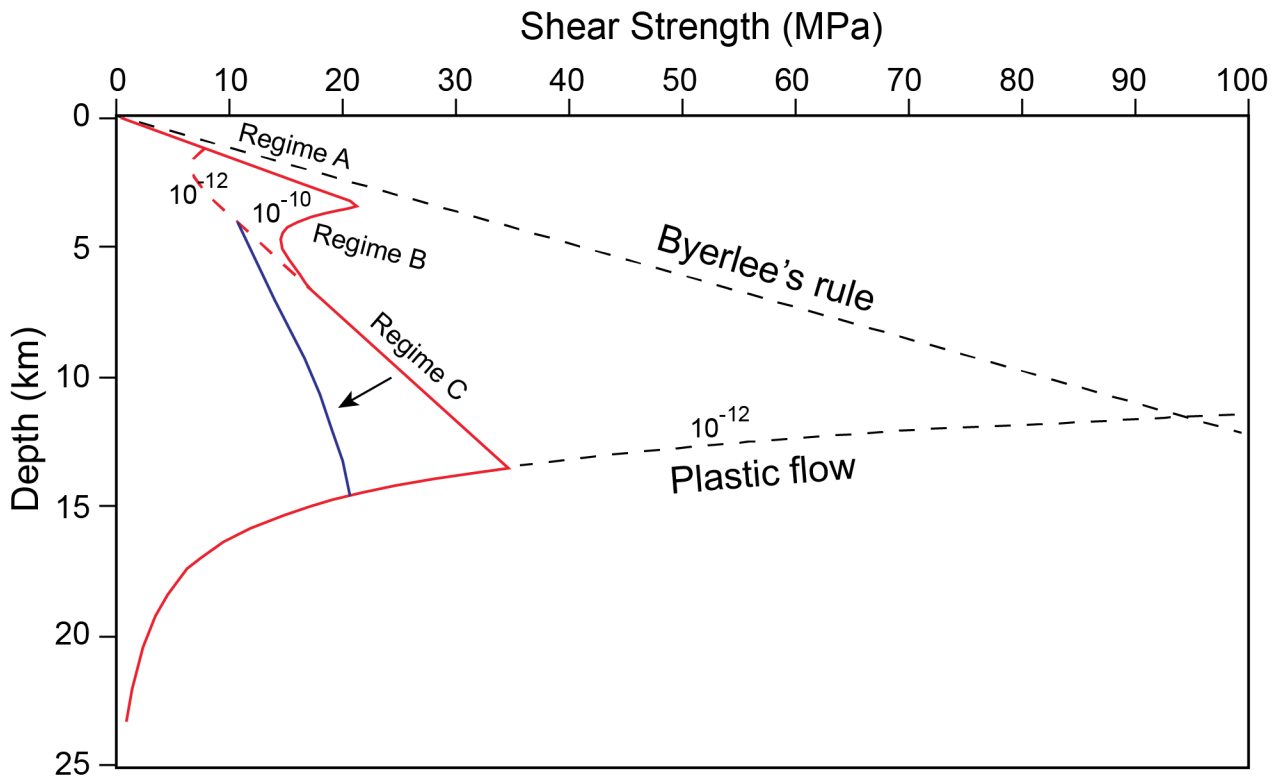
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**Figure 8.** Schematic representation of reaction softening with increasing strain. a) At the onset of deformation fracturing associated to cataclasis increases permeability favouring the influx of fluids (blue arrows) into the fault zone. b) Fluids 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 ( $\mu = 0.6-0.85$ ) 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 ( $\mu = 0.3$ ) 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).

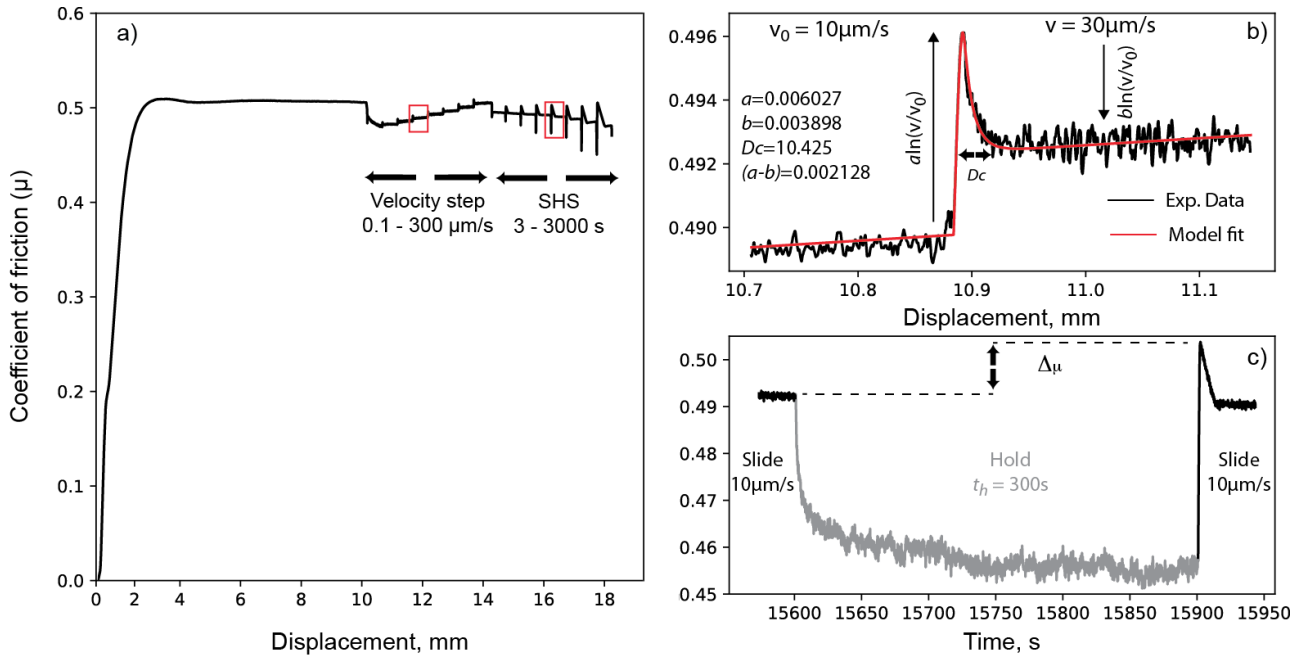
### 3.3 Frictional properties of phyllosilicate-rich faults

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  $\mu\text{m/s}$ . 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,  $\Delta\mu$ , 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:

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

565

566 where  $\Delta\mu_s$  is the change in steady state friction. Positive values of  $(a - b)$ , indicate velocity-  
 567 strengthening behavior, that favours stable sliding and fault creep. Negative values of  $(a - b)$ ,  
 568 represent a velocity-weakening behavior, that is a requirement for the nucleation of slip  
 569 instability (e.g. Dieterich and Kilgore, 1994; Marone, 1998a; Scholz, 2002).

570 In slide-hold-slide tests (e.g., Dieterich and Kilgore, 1994; Marone, 1998b; Carpenter et al.,  
 571 2016) slip at constant velocity is followed by a hold period,  $t_h$ , usually ranging from 1 to 3000  
 572 s, during which sliding is halted and subsequently resumed. The amount of frictional healing,  
 573  $\Delta\mu$ , is measured as the difference between the peak friction measured upon re-shear after  
 574 each hold and the pre-hold steady state friction,  $\mu_{ss}$  (Fig. 10c). Frictional healing rate  $\beta$  is  
 575 calculated as:

576

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

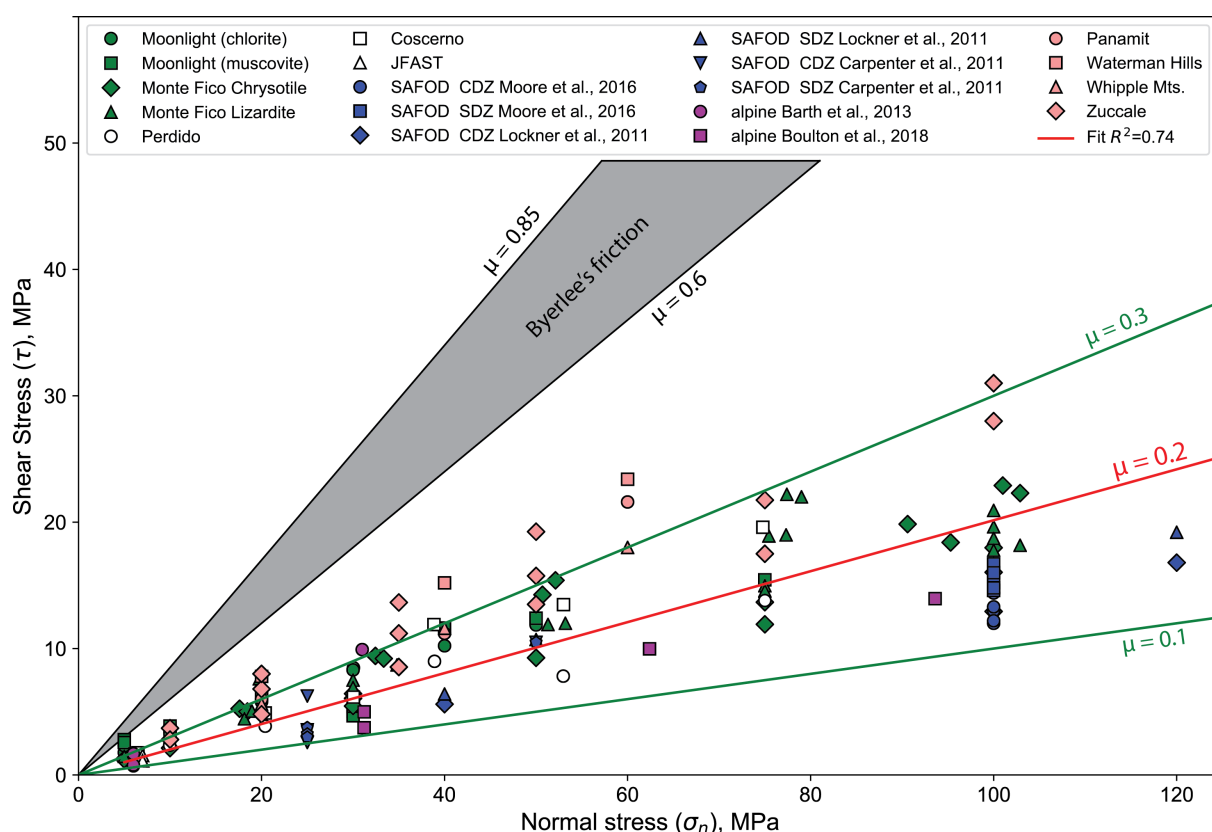
578

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

583

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

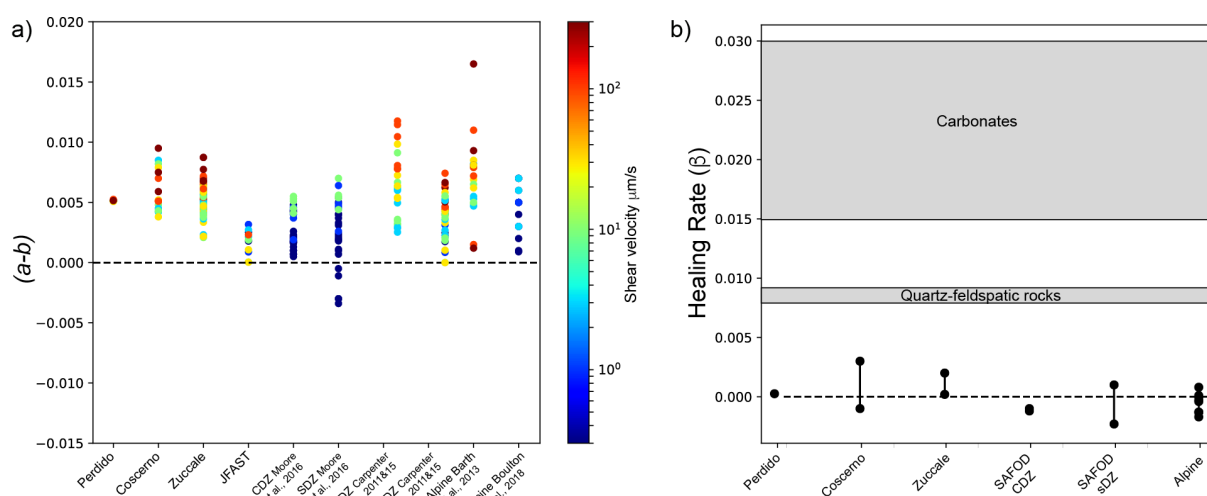
597 a) the original foliated microstructure is preserved and 2) most of the deformation occurs by  
 598 frictional sliding along the phyllosilicate-rich network with very limited cataclasis and grain-  
 599 size reduction (Collettini et al., 2009b; Tesei et al., 2014; 2015; Smith et al., 2017). Other  
 600 laboratory experiments on powdered fault rocks mixing different percentages of weak and  
 601 strong mineral phases have documented a decrease in frictional strength with the addition of  
 602 weak phases (Logan and Rauenzahn, 1987; Saffer and Marone, 2003; Takahashi et al., 2007;  
 603 Crawford et al., 2008; Giorgetti et al., 2015). In particular an amount of weak phases greater  
 604 than 40-50% of the rock volume results in significant fault weakness because it is sufficient to  
 605 promote interconnectivity of the weak mineral phase through the entire experimental fault.  
 606 This implies that when the weak mineral phases are abundant, laboratory experiments on  
 607 powdered materials are similar to those conducted on wafers.  
 608



609  
 610 **Figure 11.** Frictional properties of phyllosilicate-rich faults. The dataset together with key  
 611 references is summarized in Table 2.  
 612

613 Figure 11 shows normal stress vs. steady state shear stress from friction tests conducted on a  
 614 large number of natural fault rocks rich in phyllosilicates, both wafers or powdered material  
 615 with phyllosilicates percentages > 40%, under a wide range of experimental conditions (the  
 616 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

644 kilometres long, can be characterized by weak fault patches in zones where crustal fluids  
645 exerted a chemical role, facilitating the replacement of strong with weak mineral phases, and  
646 strong fault patches, where fluid-assisted reaction softening were not efficient, and crustal  
647 deformation was achieved predominantly by fragmentation, grain-size reduction and  
648 localization (Fig. 13). Similar heterogeneities have been documented on the grounds of  
649 frequency content of seismic waveforms of large and great earthquakes on subduction zone  
650 megathrusts (e.g. Lay et al., 2012), from high-resolution spatio-temporal behaviour of  
651 seismicity (e.g. Rubin et al., 1999; Waldhauser et al., 2004; Chiaraluce et al., 2007), and from  
652 geodetic imaging of active faults (e.g. Avouac, 2015 and references therein).

653

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

strengthening 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

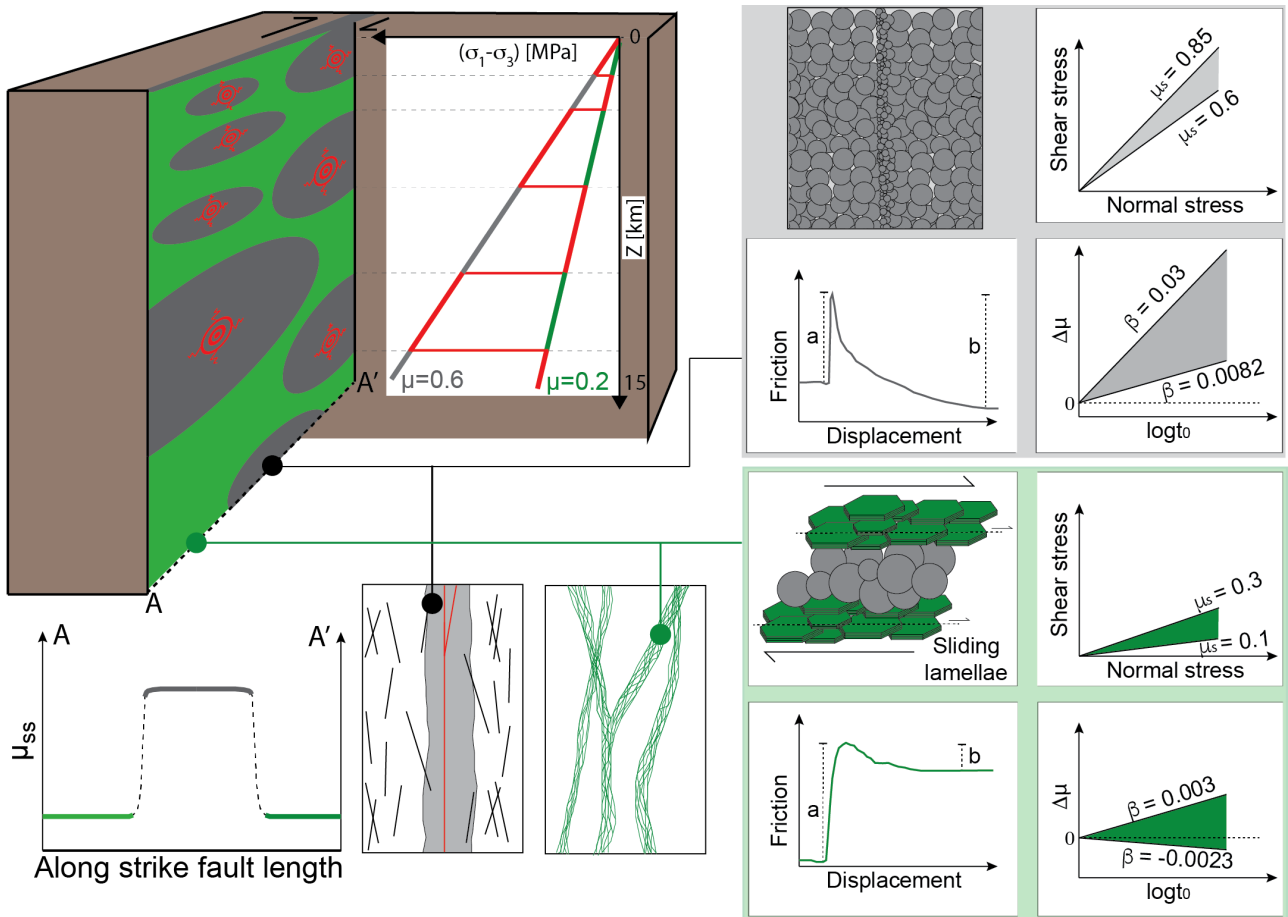
712 periods resulting in null healing rates (Fig. 12b and Carpenter et al., 2011; Tesei et al., 2012).  
713 Moreover, phyllosilicate coating strongly hinders rock cementation (Dewers and Ortoleva,  
714 1991; Worden and Morad, 2000), which together with null frictional healing suggests the  
715 inability of these fault zones to store elevated elastic stress. Collectively, these frictional  
716 properties indicate that foliated, phyllosilicate-rich faults are weak, they accommodate  
717 deformation predominantly by aseismic creep, or to be more precise (see discussion below) a  
718 seismic rupture nucleation is inhibited within these lithologies, and remain weak over long  
719 time scales.

720

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

734





**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,  $\beta$ . 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

756 observation that more strain is accommodated in the phyllosilicate-rich networks, support  
757 the idea of stress concentration in the strong lenses (e.g. Fagereng and Sibson, 2010; Fagereng  
758 and Den Hartog, 2016). When the shear stress overcomes the frictional strength of the strong,  
759 velocity-weakening patch an earthquake rupture might nucleate. In particular, frictional stick-  
760 slip instabilities occur when the fault-weakening rate with slip exceeds the maximum rate of  
761 elastic unloading, resulting in a force imbalance and fault acceleration (e.g., Scholz, 2002). If  
762 the condition to nucleate an earthquake instability is satisfied, the extent of the rupture will  
763 be controlled by the distribution and dimensions of velocity weakening vs. velocity  
764 strengthening fault patches, their state of stress, and the energy dissipated as seismic slip  
765 propagates (e.g. Boatwright and Cocco, 1996; Kaneko et al., 2010; Faulkner et al., 2011; Noda  
766 and Lapusta 2013; Avouac, 2015). Numerical models, reproducing the rich earthquake slip  
767 behaviours similar to that of natural faults, show that velocity strengthening fault patches  
768 tend to inhibit rupture propagation and the probability for a seismic rupture to propagate  
769 through a velocity-strengthening patch is related to the dimension and the frictional  
770 properties of the patch itself (Kaneko et al., 2010). However high-velocity friction  
771 experiments have shown that wet clay at low slip velocity is velocity strengthening, but at  
772 high slip velocity it weakens immediately or remains weak resulting in negligible fracture  
773 energy, making rupture propagation through clay-rich fault portions energetically very  
774 favourable (Faulkner et al., 2011). These experimental findings have been incorporated in  
775 numerical models and it has been shown that stable velocity strengthening segments may  
776 host seismic ruptures as result of mechanisms such as dynamic weakening (Noda and  
777 Lapusta, 2013). The transition from velocity strengthening at low sliding velocities (0.1-300  
778  $\mu\text{m/s}$ ) to slip dominated by dynamic weakening approaching seismic slip rates (tens of cm  
779 per seconds) have been invoked (Wibberley et al., 2008; Faulkner et al., 2011; Noda and  
780 Lapusta, 2013) to explain the exceptionally large seismic slip, of as much as 50 m (Ide et al.,  
781 2011), in the shallower area of the Mw 9.0 Tohoku-Oki earthquake, where clay rich materials  
782 are present (Kameda et al., 2015).

783 Our conceptual model for the slip behaviour of heterogeneous faults (Fig. 13) agrees with the  
784 increasing number of geodetic observations showing that in the interseismic period some  
785 fault areas remain locked whereas others creep aseismically (e.g. Avouac 2015 and references  
786 therein). The model is also consistent with the analysis of the stress orientations in  
787 subduction zones suggesting that creeping subduction zones are weaker than locked ones  
788 (Hardebeck and Loveless, 2018). However, another source of structural heterogeneities, that  
789 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).

800

#### 801 *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

824 remains a great challenge (e.g. Gombert et al., 2018) and some earthquakes show no evidence  
825 of aseismic slip in the earthquake preparatory phase, at least in the absence of measurements  
826 in the near field of the hypocentre (Ellsworth and Bulut, 2018), other evidences suggest the  
827 presence of slow-slip phenomena in or nearby the fault patch hosting the future mainshock.  
828 The foreshock activity of the Mw 9.0 Tohoku-Oki earthquake indicates that two sequences of  
829 slow slip transients propagated toward the initial rupture point of the earthquake (Kato et al.,  
830 2012). For the Mw 8.1 Iquique earthquake, more than 1 m of aseismic slip has been  
831 documented in the 15 days preceeding the event, in the same area where the mainshock  
832 occurred (Ruiz et al., 2014). High-resolution seismological data have shown that an Mw 3.7  
833 earthquake in Alaska was preceded by a slow slip phase that accelerates into a fast rupture  
834 (Tape et al., 2018). Fluid injection experiments on laboratory and natural faults reveal a  
835 similar phase of sustained aseismic creep that increases shear stress beyond the pressure  
836 front and promote earthquake triggering (Cappa et al., 2019). Collectively, these observations  
837 significantly renew the interest in precursory signals and although the detection of reliable  
838 earthquake precursors remains an open issue, now there is some cause for optimism. The  
839 structural and frictional heterogeneous nature of crustal faults presented in this manuscript  
840 suggests that creep and slow slip play a fundamental role in the deformation of crustal faults  
841 including the earthquake preparatory phase. Preparatory processes such as slow slip and  
842 seismicity migration before large earthquakes can be monitored with a combination of  
843 seismic and geodetic observations, and the physics of these processes can be constrained by  
844 structural geology and laboratory experiments, and implemented in numerical models.  
845 Innovative and original data sets will be needed to establish whether and if the observation of  
846 such signals before the mainshocks repeats in time, then leading to a reliable contribution to  
847 the forecast.

848

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1542 Table 1. Examples of fault zone structures characterized by interconnected networks of  
 1543 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

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1546 Table 2. Frictional properties of phyllosilicate-rich faults; w = wafer experiments, p powder  
 1547 experiments.  
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Fault & weak minerals	Normal stress, Temperature	Friction	(a-b) 0.01-300 $\mu\text{m/s}$	Healing rate $\beta$	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