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1 **Hybrid compactive faults formed during burial in micritic limestone (Montpellier area,**  
2 **France)**

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8

9 **ABSTRACT:** Detailed fieldwork, analyses including SEM and EBSD investigations, and  
10 dynamic inversion method are used to describe and propose a conceptual model for the formation  
11 of compactive faults related to burial in micritic lacustrine-palustrine limestones (Eocene,  
12 Pyrenean foreland basin, South of France). These faults are characterized by sinuous shape, no  
13 calcite precipitation and no macroscopic fault core evidencing their development in poorly  
14 consolidated sediments. The fault pattern is composed of two sub-synchronous systems of  
15 conjugate meter-scale faults (a normal-sense system and a strike-slip system), each one with high  
16 dihedral angle (130°) and sometimes showing continuity between faults of both sets. Their  
17 pervasive distribution, angle relative to the maximum principal stress direction, incipient shear-  
18 offset, porosity reduction, their crystallographic preferred orientation, and the “ridge and groove”  
19 morphology of their slickenside, all argue for a hybrid compactive and shear deformation process  
20 combining disaggregation, comminution, and pressure-solution-recrystallization processes. A  
21 conceptual model using estimated stress path and theoretical yield envelope is proposed to  
22 explain their formation along the end cap part during early burial and compaction of the

23 limestone beds. The slope to the South associated to the syn-sedimentary basin flexure is  
24 proposed to trigger their specific pattern.

25 .

26 **Key-words:** compactive faults, shear localization, pressure-solution, crystallographic preferred  
27 orientation, micritic limestone

28

## 29 **1. Introduction**

30 In micritic carbonate rocks, it is generally considered that fault initiate by a combination of  
31 various deformation structures such as veins, shearing planes, and/or compactive pressure-  
32 solution seams (Peacock and Sanderson, 1995; Willemse et al., 1997; Mazzoli and Di Bucci,  
33 2003; Crider and Peacock, 2004; Micarelli and Benedicto, 2008; Delle Piane et al., 2017 and  
34 references therein). Faults mainly grow by propagation and segment linkage (Kelly et al., 1998;  
35 Ferrill and Morris, 2003; Soliva and Benedicto, 2004; Agosta et al., 2012; Roche et al., 2012;  
36 Ferrill et al., 2017; Mayolle et al., 2019 among others). Because fine-grained carbonates are  
37 generally of low-porosity, fracturing and brecciation are dominant in fault zones, expressing  
38 dilation and shear behaviour (Micarelli et al., 2006; Bussolotto et al., 2015; Balsamo et al., 2016).  
39 This deformation can progressively evolve into cataclastic processes within the main deformation  
40 area called the “fault core” (Billi, 2005; Hausegger et al., 2010; Schröckenfuchs et al., 2015;  
41 Ferraro et al., 2018). It can be accompanied by other various processes generally related to  
42 seismic events, i.e. clast-cortex grains, calcite crystallographic preferred orientation (CPO), or  
43 fluidized layers (Delle Piane et al., 2017; Demurtas et al., 2019). In such model, compactional  
44 processes are generally restricted to specific areas, such as fault dip changes (Ferrill et al., 2020)

45 or fault tips (Rispoli, 1981; Petit and Mattauer, 1995; Agosta et al., 2010). Compaction by  
46 pervasive pressure-solution can also occur within the cataclastic zone, showing scaly foliation  
47 (Bullock et al., 2014; Clemenzi et al., 2014). However, fault initiation by compaction processes  
48 (mainly pressure-solution pressure-solution but also disaggregation and cataclasis) remains  
49 unclear, especially in micritic carbonates (Billi, 2003; Graham et al., 2003; Agosta and Aydin,  
50 2006).

51 Conversely, fault initiation by shear and compaction failures is well known in porous and  
52 granular rocks such as sandstones (Antonellini and Aydin, 1994; Ballas et al., 2014; Fossen et al.,  
53 2018), grainstones (Tondi et al., 2006; Cilona et al., 2012; Antonellini et al., 2014; Rotevatn et  
54 al., 2016) or chalks (Wennberg et al., 2013; Kaminskaite et al., 2019). In such lithologies, faults  
55 initiate by deformation bands showing different micro-mechanisms (Fisher and Knipe, 2001;  
56 Aydin, 2006; Fossen et al., 2007) and various ratios of shear to compaction; i.e. pure compaction  
57 bands (Mollema and Antonellini, 1996; Agosta et al., 2010; Rotevatn et al., 2016), shear-  
58 enhanced compaction bands (Eichhubl et al., 2010; Ballas et al., 2013; Fossen et al., 2015),  
59 compactive-shear bands (Saillet and Wibberley, 2013), and shear bands (Rotevatn et al., 2008).  
60 Cataclastic processes are dominant and the related grain crushing induces pore collapse and  
61 porosity reduction (Antonellini and Aydin, 1995; Fossen and Bale, 2007; Ballas et al., 2015)  
62 leading to material hardening and shear localization (Aydin and Johnson, 1978; Schultz and  
63 Fossen, 2002; Shipton and Cowie, 2003; Philit et al., 2018). Other micro-mechanisms; such as  
64 disaggregation (grain rotation, torsion or displacement), or pressure-solution-recrystallization; are  
65 described in deformation bands and faults (Cilona et al., 2012; Ballas et al., 2013; Philit et al.,  
66 2015; Tavani et al., 2018a; Kaminskaite et al., 2019 among others). However, these processes  
67 remain difficult to integrate in mechanical model such as the one developed for cataclastic

68 structures (Saillet and Wibberley, 2010; Soliva et al., 2013, 2016; Robert et al., 2018; Fossen et  
69 al., 2018) because of the lack of combined descriptive-mechanical approach on such deformation  
70 features.

71 Mechanical concepts suggest a potential formation of compactive-shear structures under  
72 progressive burial, i.e. out of regional tectonic influence (Soliva et al., 2013). However, only few  
73 natural examples propose compactive shear localization under gravitational loading (Guiraud and  
74 Séguret, 1987; Wennberg et al., 2013), whereas almost all described compactive-shear structures  
75 are related to tectonic stresses. During progressive burial, compaction follows general porosity  
76 reduction curves (Ehrenberg and Nadeau, 2005) but can also localize bed parallel structures, such  
77 as compaction bands (Aydin and Ahmadov, 2009; Rustichelli et al., 2012) or stylolites (Laronne  
78 Ben-Itzhak et al., 2014). The part of these localization structures within the whole compaction of  
79 sediment remains unclear. A better understanding of compactive-shear structures initiation  
80 appears necessary, especially within fine-grained materials and under burial “gravitational”  
81 loading.

82 In such context, we investigate a network of meter-scale compactive-shear structures exposed  
83 within micritic continental limestone in the Aumelas - Hérault basin in the South of France  
84 (Montpellier area). Detailed fieldwork, natural light, SEM and EBSD (scanning electron  
85 microscopy, electron backscatter diffraction) microstructural investigations, and kinematic  
86 inversion method have been used to describe the fault systems and discuss the mechanisms and  
87 triggers. Theoretical mechanical models are used to discuss the origin of such structures.

## 88 **2. Methods**

### 89 *2.1. Field observations and measurements*

90 Sedimentological log of the main studied section was done to identify the relationship between  
91 the occurrence of fault and bed characteristics. Beds are analyzed following the Dunham  
92 classification (*M*: Mudstone, *W*: Wackstone, *P*: Packestone, *G*: Grainstone, *B*: Boundstone + *C* to  
93 identify clays). Evidences of subaerial alterations (desiccation, karstification, brecciation)  
94 observed on bed surfaces and the beds affected by fault sets (i.e. the location of fault networks in  
95 the sedimentary succession) are also reported.

96 The geometry of the fault network is analyzed from stereographic data collected using compass  
97 and digit tablet (FieldMOVE software). These data are reported on stereoplots and histograms of  
98 dip and rake, i.e. the angle between the slickenline and the strike direction of individual fault  
99 planes. The data are presented after restoring the bedding to horizontal.

100

## 101 *2.2. Microstructure and rock properties*

102 Nine polished thin-sections were cut from blocks sampled below fault slickensides previously  
103 impregnated by epoxy resin. These thin-sections were cut in the dip-slip direction of faults (6  
104 thin-sections) or the strike direction of faults (3 thin-sections) to analyze fault micro-mechanisms  
105 and kinematic criteria (evidence and sense of shear). They were analyzed by backscattered  
106 electron imagery using a FEI Quanta 200 environmental scanning-electron microscope (SEM).  
107 Mosaics of SEM backscattered electron images, with a gray-scale intensity range from 0 to 255,  
108 were obtained and used to observe deformation features and to estimate the porosity of the  
109 deformed zones and the host rocks (2D porosity obtained from black areas on SEM images).  
110 Several SEM images were also acquired on 3D blocks to observe grains and slickensides micro-  
111 morphology. Analyses with optical microscope in natural or cross-polarized light, and  
112 cathodoluminescence mode were also used to identify potentially different cement phases.

113 Electron BackScatter Diffraction (EBSD) analyses were done in the host rocks and along the  
114 fault zones, including steps and fault core. This analysis edits a map of the particles from their  
115 crystallographic assemblage and allows analyzing their size, shape and crystallographic  
116 orientation. Data were collected with the scanning electron microscope CamScan Crystal Probe  
117 X500FE and were processed using AZtecHKL software (Oxford Instruments). The acquisition  
118 was done with a 0.3-0.4  $\mu\text{m}$  step size at a working distance of 25 mm. Grains were reconstructed  
119 from orientation data and their boundaries were defined as having a misorientation angle between  
120 neighboring pixels of  $> 5^\circ$ . The grain slope (i.e. the angle of the grain longer axis with respect to  
121 vertical) was measured on grains showing surface area larger than 5  $\mu\text{m}$  in order to avoid artefact  
122 due to the image resolution. These data are represented on rose diagram. Crystallographic  
123 orientations are represented on stereographic diagram (equal area, lower hemisphere) and the  
124 multiple of the uniform distribution (*m.u.d.* value) is used to estimate the crystallographic  
125 preferred orientation of grains (CPO). Calcite twinning is automatically mapped using the  
126 AZtecHKL software.

127

### 128 2.3. *Fault slip and stress analysis*

129 The state of stress during fault slip is obtained by inversion of the fault and slickenline data  
130 (Angelier, 1984) with the FSA software that was developed to analyse various aspects of the  
131 relationship between fault slips and stresses (Tajima and Célérier, 1989; Burg et al., 2005;  
132 Célérier, 2020). The inversion seeks the stress tensor that best align shear stress with observed  
133 slickenlines (Wallace, 1951; Bott, 1959) and uses the random search proposed by Etchecopar et  
134 al. (1981). This method allows to invert either the whole dataset in the case of monophasic data or  
135 only a fraction of the dataset in the case of polyphasic data. It yields the principal stress

136 orientations  $s_1$ ,  $s_2$ ,  $s_3$ , and stress tensor aspect ratio  $r_0 = (\sigma_1 - \sigma_2) / (\sigma_1 - \sigma_3)$ , where  $\sigma_1 \geq \sigma_2 \geq \sigma_3$  are the  
137 principal stress magnitudes corresponding to the orientations  $s_1$ ,  $s_2$ ,  $s_3$ . We use the quality  
138 criterion proposed by Heuberger *et al.* (2010) to distinguish robust solutions from statistical  
139 artifacts. It is based on a comparison between the total number of data ( $N_{total}$ ), the number of data  
140 with a rake misfit lower than  $15^\circ$  ( $N_{exp15}$ ), and the number of data ( $NO_{exp15}$ ) that could be obtained  
141 below the same misfit by inverting a random dataset of the same size  $N_{total}$ . The result of this  
142 comparison is summarized by an integer quality index, *InvQual*, which varies from -1 for barely  
143 acceptable solutions to 2 for robust estimates.

144

### 145 **3. Geological Setting**

#### 146 *3.1. Regional tectonic framework*

147 The Aumelas - Hérault basin is located 20 km to the west of Montpellier city in the South of  
148 France (Fig. 1a). This basin form a part of the continental foreland basin developed at the front of  
149 the Montpellier thrust (also called North Pyrenean Frontal Thrust) (Arthaud & Séguret 1981).  
150 This thrust is related to the Pyrenean orogeny occurring in this area during two main shortening  
151 events, during Early Paleocene and Late Eocene. A relatively quiescent phase of tectonic activity  
152 occurred during Lower Eocene (Debrand-Passard *et al.*, 1984) when carbonate lacustrine-  
153 palustrine sequence were deposits (Bonnet *et al.*, 1971). However, the accommodation of a 300-  
154 500 m thick sedimentary sequence showing locally growth strata (syndimentary folding) in  
155 lacustrine deposits, indicates that some moderate tectonic activity remained during this period.

156 The studied outcrops are located along the A750 highway, at the “La Taillade” site (Lat.  
157  $43^\circ37'36''N$ ; Long.  $3^\circ35'54''E$ , Fig. 1a). This area is located in front of the WNW-ESE St Bauzille



158 Thrust (reactivated Mesozoic normal-fault) and close to the change from NE-SW to EW trend of  
159 the main Montpellier Thrust. Lacustrine-palustrine limestones were mainly folded and faulted  
160 during Late Eocene second shortening phase of the Pyrenean orogeny (Fig. 1b). These strata are  
161 now tilted 15-30° to the S and SW (up to 65° in the central part of the section, Fig. 2a), according  
162 to their location in the fore-bulge of the Pyrenean thrusts and to the tilting of the NE-trending  
163 Hérault half-graben during Oligocene rifting (Maerten & Seranne 1995). We note that the  
164 maximum burial reached by the studied section is difficult to evaluate due to an uplift (>100 m)  
165 and erosion during late Miocene (Séranne et al., 2002). However, we consider a maximum burial  
166 in an order of magnitude of 100 m since Oligocene sediments are not deposited above the study  
167 site (only formed in half-grabens) and the studied limestone section corresponds to the upper part  
168 of the Eocene sequence.

169

### 170 3.2. Host rock characteristics

171 The main outcrop lies at the top of the 300 m thick “Lutetian” limestones strata related to a  
172 lacustrine-palustrine system (Hartenberger et al., 1969) (Fig. 2). The studied section is limestone  
173 dominated and shows a succession of mostly 25-50 cm thick, stacked, tabular carbonate beds  
174 (Figs. 2 b, c). The beds are mainly composed of a massive white mudstone including whole or  
175 fragments of fossils of continental gastropods (*limnea* and *planorbis*) and ostracods bioclasts. We  
176 note the presence of a 7 m thick wackestone to packstone interval (particle sizes well above 30  
177 microns meters and often rather millimeters to centimeters). In some places, the mudstone beds  
178 are intercalated with few centimeters thick brownish silty marls (Fig. 2d). Such silty marls can  
179 also reached 1 m thick at the base of the studied section. The top of the section ends with a 3 m

180 thick marly interval including sandy-carbonate beds. The complete sedimentary succession does  
181 not display any particular depositional architecture nor sedimentary structures. The main study  
182 section is located down West of the 3 km long exposure along the highway, where the carbonate  
183 beds only dip of 25-35° which permits to make measurements (Fig. 2a). A second section is  
184 located to the East, at the close vicinity of the Montpellier thrust and the Oligocene La Boissière  
185 normal fault (Fig. 2a). This second outcrop of “Lutetian” limestone strata is separated by map-  
186 scale strike-slip sinistral faults and cannot be precisely correlated to the whole outcrop as no  
187 benchmark beds is observed from both sections.

188 In thin section, the matrix of mudstones is composed of an unclear micrite sometimes including  
189 initially mud-size crushed bioclastic debris and peloids. The SEM image from 3D samples show  
190 micritic euhedral crystals of calcite less than 5 µm in size and few crystals of dolomite (less than  
191 5%) (Fig. 3a). The present-day porosity of the studied mudstones, measured in laboratory on 11  
192 mini cores using the triple weighting method, ranges from 4.4% to 22.3%. These mudstones are  
193 therefore qualified as fine-grained and porous carbonate rocks. The dark and homogeneous  
194 images from cathodoluminescence microscopy analysis reveal no difference between calcite  
195 grains and interstitial cements. We note that the imbricated contacts between calcite particles are  
196 observed at microscale disseminated in the material (Fig. 3b). The bed surfaces are affected by  
197 pedogenic modifications (i.e. desiccation cracks, micro-karstification, brecciation, iron oxide  
198 deposits...) related to shallow palustrine environments (wetlands), and therefore to common  
199 exposition to drying conditions and evaporative/groundwater recharge phases (Freytet and  
200 Plaziat, 1982; Alonso-Zarza and Wright, 2010). The mudstone beds of the second (eastern)  
201 outcrop show less evidences of emersion and contain several brecciated beds with higher quartz  
202 grains fraction. These specificities suggest a formation of this sedimentary section in a lacustrine

203 domain rather than palustrine environment, or at least a deeper water “lake” setting that records  
204 less the evaporative phases compared to the wetland environment of carbonate section of the  
205 western outcrop.

206

#### 207 **4. Description of faults**

208 Different fault sets are observed along the study outcrop. We focus on meter-scale faults showing  
209 specific slickensides (see description section 4.2.) whereas Late Pyrenean or Oligocene fault  
210 networks (i.e. faults of several meters to tens meters length, cutting across the whole observable  
211 limestone section, with a brecciated fault core, a damage zone with mode I joint systems and  
212 calcite fiber along their surface) are not considered in the present study. Studied faults are  
213 organized in two different sets: a set of dip-slip normal faults at low-angle with respect to  
214 bedding, and a set of strike-slip faults at high-angle to bedding.

215

##### 216 *4.1. Location of faults within the limestone succession*

217 Faults are heterogeneously distributed in the limestone succession. They affect almost all beds  
218 within the first 30 m and the last third of the mudstone sequence from 82 to 88 m (Fig. 4). They  
219 are mainly absent from 55 m to 75 m, especially in the wackestone-packstone interval. These  
220 faults are also observed restricted to specific mudstone beds that are intercalated within beds  
221 unaffected by faults from 30 to 55 m and from 75 to 82 m. The faults are only observed in  
222 limestone beds thicker than 20 cm (mainly absent in thinner ones). They can be observed  
223 whatever the secondary alteration occurring in affected limestone beds during syn-sedimentary  
224 subaerial exposure (desiccation, oxidation...) (Fig. 4). However, we note that faults are more

225 abundant in limestone beds showing few secondary alteration (for example the 2 m to 17 m  
226 section) whereas they are mainly absent in highly altered limestone beds (for example the 55 m to  
227 75 m section).

228

#### 229 *4.2. Specificity of fault slickensides*

230 The studied faults differ from other fault systems affecting the limestone sections in the area by  
231 their morphology and slickenside characteristics (Fig. 5). At the first order, these faults are  
232 characterized by the lack of a macroscopic fault core (i.e. fault rock visible at the naked eyes) and  
233 the lack of calcite or iron oxide precipitation along their slickensides (Fig. 5a). The slickensides  
234 show grooves as frequent as ridges along the fault planes. These grooves and ridges features are  
235 of several centimeters length and show U-shape with flat bottom (or flat top) and larger wave  
236 length than amplitude (Figs. 5b, c). These grooves are observed at two imbricated scales of  
237 width: 1 cm and 100s  $\mu\text{m}$  (Figs. 5b, e). The ridges of one wall of the fault plane perfectly fit into  
238 the grooves of the other wall, i.e. no fracture porosity or cementation is observed along the fault  
239 surface. These grooves show no specific asperity at the lower tip whereas two different  
240 morphologies are observed at the upper tip of the groove: (1) V “crescentic” markings and (2)  
241 step markings formed by sharp mm to cm-high plane, at high angle to the slickenside (following  
242 the classification of Doblas, 1998) (Figs. 5b, c, d and f). These step markings are irregular planes  
243 of centimeter to several centimeters length (Fig. 5d). They are uncommonly related to joint below  
244 the fault plane but are more generally observed without associated deformation structures. No  
245 stylolitic teeth are observed on these step markings.

246

247 *4.3. Normal fault system*

248 The low-angle to bedding faults are generally composed of meter-scale length structures (Fig.  
249 6a). These faults are mainly restricted within single carbonate beds, and rarely affecting 2 or 3  
250 beds. They show a sigmoidal shape (or flat S-shape) in cross-section with a maximum dip in the  
251 central part of the beds, progressively decreasing at the top and base of the beds (listric to bed  
252 interface) (Figs. 6b, c). Laterally, they form curved and irregular bended surfaces and show  
253 connections with surrounding faults. The steps at the upper tip of the grooves along the  
254 slickensides suggest a normal-sense kinematic (Petit & Laville 1987) (Fig. 5), but no offset  
255 markers or tip folding are observable. These faults form a conjugate system with opposite dip.  
256 After restoration to horizontal, stereographic plots reveal East-West oriented conjugate fault  
257 system, with mainly North-South slickenlines (grooves) (Fig. 7a). Each dipping set is  
258 crosscutting the other with a dominant observation of South-dipping set cutting the North-dipping  
259 one (Figs. 6d, e). These crosscutting zones uncommonly show millimeter-scale normal-sense  
260 shear-displacement but no measurable offset is more generally detected along these faults.  
261 Moreover, numerous abutting contacts are observed between both sets (Fig. 6d, e). Few NW or  
262 SW-striking slickenlines are also measured. The dip of the faults, measured systematically in the  
263 central part of beds, is on average  $25^\circ \pm 7^\circ$ , suggesting a high dihedral angle of about  $130^\circ$   
264 between conjugate fault sets (Fig. 7b). The rake data show purely dip-slip motion (Fig. 7c).

265

266 *4.4. Strike-slip fault system*

267 The high-angle to bedding faults show large grooves and ridges similarly to low-angle normal  
268 faults (Figs. 8a, b). These faults generally cross-cut several limestone layers (2 to 5 beds). They

269 show numerous centimeter-scale dilational jogs between meter-scale planar fault segments (Fig.  
270 8c). These dilational jogs observed along the fault revealed a shear-displacement of about 1 cm  
271 (Fig. 8d). These faults form then a strike-slip system with a dextral set of strike ranging of strike  
272 ranging from N°20E to N°50E and a sinistral set striking from N°150E to N°180E (Fig. 9a). The  
273 strike-slip faults form a conjugate system with a general dihedral angle between 120° and 140°  
274 (similarly to the normal-sense set). These faults are almost vertical after restoration of bedding to  
275 horizontal, on average  $79^\circ \pm 9^\circ$ , and show mostly horizontal slickenlines (grooves) with a plunge  
276 of  $11^\circ \pm 13^\circ$  on average (both North or South) (Fig. 9b) and a rake around the 0° and the 180°  
277 values (i.e. horizontal) (Fig. 9c).

278 The strike-slip fault systems generally cross-cut the normal-sense fault system (Fig. 8a), but, a  
279 same surface is sometimes observed as laterally evolving from strike-slip to dip-slip slickensides,  
280 with a progressive decrease of dip and showing parallel grooves of similar trend (Figs. 10a, b).  
281 We note that both dip-slip and strike-slip systems are commonly observed together (Fig. 10d).  
282 However, a bed-by-bed analysis of fault system within the first 47 beds of the limestone  
283 succession reveals 6 beds where only dip-slip normal faults are observed and 3 beds with only  
284 strike-slip faults (Fig. 10c). The fault spacing of only few tens centimeters revealed a pervasive  
285 distribution of both sets in bed affected by faulting (Fig. 10d). We note that several examples of  
286 strike-slip slickensides are observed along the plane of late strike-slip brittle faults (i.e. faults of  
287 tens meters length, with a brecciated fault core and calcite fiber along the fault mirror).

288

289 *4.5. Microstructure of faults*

290 The microstructure of the limestone in the wall along the fault plane changes from areas where  
291 the grain framework is similar to the host rock (grain size, shape, composition, porosity) into  
292 zones where the microporosity is significantly reduced close to the main fault surface (Fig. 11).  
293 The thickness of this low-porosity zone is variable and can reach in the best-case 0.3 mm below  
294 the fault plane. These compacted zones are preferentially observed at fault plane irregularities,  
295 i.e. where small dip changes occurs and forms compressive jogs along the slickenside (Figs. 11a,  
296 c, d). This specific location of compaction is consistent with the kinematics of the faults deduced  
297 from field observations. Smooth pressure-solution seams are observed particularly in these zones  
298 as sinuous cleavages following grain contacts (Figs. 11c, d). These cleavages are parallel or  
299 slightly oblique to the faults and are sometimes open (probably related to thin-section  
300 preparation). Concerning step markings located at the upper tip of the groove, these areas show  
301 no evidence of compaction. The presence of dilation joint is sometimes observed (Fig. 11e), but  
302 several examples show no deformation feature below the step. A sharp plane penetrating the  
303 material is often observed in the continuity of the fault plane and could be interpreted as P plane  
304 structure (Fig. 11e). No significant grain orientation is observed in the host material (Fig. 12a).  
305 However, compacted areas show a preferential orientation of grains parallel to the fault plane  
306 (Figs. 12b, c).

307 A low-porosity and compacted zone of few tens of micrometers is observed above the main fault  
308 plane (Fig. 11b). In this zone, the particles appear smaller than the ones of host material (Fig.  
309 13a). We note that only few grains with microcracks and no hertzian contacts were identified in  
310 these structures. This zone is bordered by the main fault slickenside. Several tens  $\mu\text{m}$  long planes  
311 are also observed within the comminuted zone (Fig. 11b). This comminution is accompanied by a  
312 crystallographic preferred orientation (CPO) of grains as revealed by the dominant reddish color

313 of grains within the compacted fault zone (EBSD acquisition showing colors depending on the  
314 crystallographic organization of each grains) (Fig. 13a). This CPO shows a maximum *m.u.d.*  
315 value of 4.85 whereas this value nether exceeds 1.70 in the host rock (Fig. 13b). This CPO is  
316 observed in the c-axis of calcite grains and is oriented about 70° to the fault plane. We note that  
317 twining process is revealed by μm-thick bands showing a different color than the grain they  
318 affect (Fig. 14a). These twins are only observed within the compacted area of the fault core or the  
319 compactive jogs (Figs. 14b, c) whereas no twins are observed on step marking at the upper tip of  
320 grooves (Fig. 14d). This specific location of calcite twins is consistent with the normal-sense  
321 kinematic proposed for the fault system at low-angle dip relative to the bedding.

322

#### 323 4.6. Fault slip and stress analysis

324 Strike directions of faults are clustered East-West and rakes have 3 modes: a major mode that is  
325 pure normal (rake around -90, Fig. 7c) and two minor modes, around pure sinistral (left lateral)  
326 and pure dextral (right lateral) (rake around 0 and 180, Fig. 9c). Inverting the restored dataset and  
327 taking into account all of the 264 data yields a solution with only three outliers with a rake misfit  
328 larger than 160°. This can be due either to discordant movements or to erroneous slip sense  
329 determination. It was thus decided to invert for 99% of the data to disable any influence of the  
330 outliers on the result. The resulting solution is at the transition between extensional and wrench  
331 regime with  $s_3$  horizontal and about N°10E (Fig. 15a). In this solution,  $s_1$  is vertical and show a  
332 very close value with  $s_2$  suggested by the  $r_0 = 0.0$ . This solution also yields a low average angular  
333 error (<10°, computed on 99% of the data), a large number of faults with rake misfit below 15°  
334 ( $N_{exp15} = 216$ ), and the highest value  $InvQual = 2$  (Fig. 15b). The normal faults are at high angle



335 ( $\geq 60^\circ$ ) with  $s_1$  because of their low (15-30°) dip with respect to bedding. The strike-slip faults  
336 are also at high angle with  $s_2$ , which plays the same role as  $s_1$  when  $r_0 = 0.0$ . As a consequence  
337 both types of fault data are located on the right hand side of Mohr's circle (Fig. 15c). The quality  
338 of the solution supports the fact that all faults are dynamically consistent.

339

## 340 **5. Multi-beds normal faults**

341 Several multi-beds normal faults (i.e. non-stratabound faults) are observed at the second studied  
342 site, eastward of the main study outcrop (Fig. 2a). Because of the specificity of these structures  
343 and their absence in the main 2.5 km long outcrop, we decided to describe them separately.

344 Slickensides of the multi-beds normal-faults show closely similar morphology than the bed-  
345 restricted faults, i.e. grooves and ridges, marking steps, no calcite precipitation, and curved planes  
346 (Fig. 16a). The specificity of such slickensides is the presence of visible indentations at lower tip  
347 of grooves interpreted as slickolites (i.e. sheared pressure-solution zone, Arthaud and Mattauer,  
348 1969) (Fig. 16b). The height of these faults is at least 10 m and their horizontal length is about  
349 several tens of meters. They are mainly trending EW and dip exclusively to the South (Fig. 16c).  
350 Their average dip is  $23^\circ \pm 7^\circ$  (after bedding restored to horizontal). The fault throw can reach few  
351 centimeters (Fig. 16d, see the evidence of footwall bed marks on the fault slickenside). We note  
352 evidences of reactivation in reverse-sense motion on one of these faults, i.e. fine and regular  
353 striation oblique to the main normal slickenlines, consistently with a reactivation during late  
354 Eocene Pyrenean shortening. No multi-bed normal fault dipping to the North was observed at this  
355 site. Brittle-like (late) strike-slip structures are observed with brecciated fault core, calcite fiber

356 along the fault plane, and mode I fracture damage zone (joints), but no strike-slip slickensides  
357 were clearly identified.

358

## 359 **6. Discussion**

### 360 *6.1. Model of fault formation*

361 The low intergranular porosity zones having wavy cleavage observed close to the fault planes  
362 (Fig. 11), the comminution, CPO and twinning observed in the fault core, as well as the high  
363 dihedral angle between conjugate faults ( $130^\circ$  to the shortening axis) reveals that the deformation  
364 mechanism is hybrid between compaction and shear, and not controlled by a standard Coulomb  
365 type frictional strength law. These characteristics expressing failure on the right hand side of the  
366 Mohr's circle (Fig. 15c) are consistent with shear-enhanced compaction and pore collapse modes  
367 observed in porous granular material having yield envelope with elliptical end cap (Wong and  
368 Baud, 2012; Cilona et al., 2014) (Fig. 17a). Indeed, studied limestones are composed of micritic  
369 grains of calcite with relatively high interstitial porosity (up to 22%; we note that these faults are  
370 very rarely or never observed in wackestone or packstone carbonate beds). They probably  
371 behaved similarly to other porous and fine granular carbonates such as chalk (Risnes, 2001;  
372 Risnes et al., 2005) or marine limestones (Vajdova *et al.* 2004; Wong & Baud 2012 and  
373 references therein). The size of such yield envelopes is mainly controlled by the porosity and the  
374 grain size of the host material (if grain fragmentation is involved) as well as particle cementation  
375 (e.g. Estrada and Taboada 2013; Soliva *et al.* 2013). This mechanical model implies that the  
376 faults were formed along the cap under low ratios between applied shear and normal stress (Fig.  
377 17a).

378 The stress state leading to fault formation can be obtained under a progressive increase of burial  
379 with a high coefficient of Earth pressure at rest ( $K_0$  the ratio of vertical to horizontal stresses in  
380 oedometric conditions, i.e. no horizontal remote stress). This ratio ranges from 0.5 to 1 in very  
381 fine-grained materials (Goultly 2003; Mitchell & Soga 2005) but is not constrained for continental  
382 unlithified micritic carbonate muds to your knowledge. The occurrence of both sub-synchronous  
383 normal-sense and strike-slip systems suggests that vertical ( $\sigma_v$ ) and horizontal ( $\sigma_{H_{EW}}$ ) stresses are  
384 very close. This fault pattern and the occurrence of faults in an unconsolidated material argue for  
385 a high  $K_0$  context since no evidence of EW regional (tectonic) contraction is observed in this area.  
386 We here consider a range of  $K_0$  from 0.7 to 0.9 to estimate stress paths leading to faults formation  
387 (Fig. 17a). All these observations and mechanical considerations allow considering these faults as  
388 formed by burial driven compaction during the increase of burial depth. The presence of faults  
389 within the top of the sequence suggests that structures can form under a burial of only few tens  
390 meters depth (Fig. 4). Their curved and irregular geometry is consistent with subsurface  
391 conditions of formation in poorly lithified rocks (Guiraud and Séguret, 1987). We note that faults  
392 are generally observed in the limestone succession showing less evidences of subaerial exposure  
393 (karstification, desiccation...) (Fig. 4). Subaerial exposure, by its impact on the host material  
394 rheology, would probably prevent the formation of such faults.

395 However, it is not common to localize shear plane in such mechanical context favourable to  
396 compaction banding with more or less shear component (Schultz and Siddharthan, 2005; Baud et  
397 al., 2006; Soliva et al., 2013). In these cited examples, the compaction and the pore collapse  
398 mainly occurred by cataclasis, leading to pervasive network of compactive cataclastic bands  
399 (Solum et al., 2010; Sallet and Wibberley, 2010; Ballas et al., 2015; Soliva et al., 2016;  
400 Cavailhes and Rotevatn, 2018; Robert et al., 2018) and uncommonly conduct to shear plane

401 localization following the strain hardening model of Aydin and Johnson (1978) (e.g. Rotevatn et  
402 al., 2008; Tueckmantel et al., 2010). The fault mechanism described in the present study differs  
403 from this model of localization as a combination of processes is involved, i.e. grain rotation,  
404 pressure-solution, comminution, and a dynamic recrystallization as revealed by the  
405 crystallographic preferred orientation (CPO) of grains (Smith et al., 2013; Delle Piane et al.,  
406 2017). By this way, it is interesting to compare the characteristics of these planes to the  
407 slickensides described in C-S structures in crystalline rocks at the deep brittle-ductile transition  
408 (Berthé et al., 1979; Lin et al., 2007). In such contexts, slickensides share similar characteristics  
409 than the presently studied structures: grooves and ridges morphology without noticeable down-  
410 slip ends, and slickenlines (grooves) of several centimeters length for a fault of cm-scale shear-  
411 displacement (e.g. Means, 1987). To explain this strain localization in a ductile context, a plastic  
412 behaviour of material which is not related to a frictional sliding process is proposed by Lin and  
413 Williams (1992). As expressed by Will and Wilson (1989) from analog laboratory experiments:  
414 “the excess length and the nesting between footwall and hanging wall ridges and grooves is  
415 therefore a product of this plastic deformation”.

416 We propose that quite similar hybrid and evolving processes occurred under sub-surface  
417 conditions in response of plastic processes followed by shear localization and an early blocking  
418 of the fault movement. The fault initiation starts by disaggregation and grain comminution within  
419 10s  $\mu\text{m}$  thick zones (Fig. 17b, stage 1). The process triggering the grain comminution remains  
420 unclear. No evidence of significant cataclasis was observed even if this process cannot be totally  
421 excluded. However, the grain comminution is accompanied by a dynamic recrystallization of  
422 calcite as revealed by the CPO within the comminuted zone. The potential hardening induced by  
423 the grain comminution (Schultz and Siddharthan, 2005; Wibberley et al., 2007) evolves

424 contemporaneously with a softening behaviour induced by the dynamic recrystallization (Smith  
425 et al., 2013), which conducts to shear localization. This shear localization seems initiated by  
426 segments of hundreds  $\mu\text{m}$  to several mm long, and is developed within or, more frequently, along  
427 the comminuted-recrystallized zone (Fig. 17b, stage 2). We note that grain rotation and pressure-  
428 solution evidences are generally observed within compressive jogs and are not observed all along  
429 the fault plane. This specific location suggests that these processes acts as secondary mechanisms  
430 related to the fault plane heterogeneities and then mainly occurred after the fault localization (Fig.  
431 17b, stage 3). However, these processes reveal a compactive behaviour during the last stage of  
432 the fault evolution which certainly triggers an early blocking of the fault movement as expressed  
433 by the very low shear displacement measured along the fault slickensides.

434 The formation of two sub-synchronous conjugate systems rather than radial or random patterns is  
435 quite unexpected considering that this mechanism of faulting is triggered by compaction and  
436 burial context. This particularity in the fault pattern can be explained by a decrease of the NS ( $\sigma_3$ )  
437 stress induced by the syn-burial tilting of sedimentary layers since they are located in the extradors  
438 of a very smooth antiform associated with the flexure of the foreland basin (foreland bulge, Fig.  
439 17c) (e.g. Ferrill et al., 2021). In this particular setting, a conjugate system of E-W oriented  
440 normal-sense faults formed consistently with the polarization of horizontal stresses (NS  $s_3$ ),  
441 impeding more random pattern formation. The strike-slip faults generally crosscut the low-angle  
442 normal faults whereas their connection may be a local reactivation of the normal system. The  
443 strike-slip system could then initiate slightly later, consistently with a close value of  $\sigma_1$  and  $\sigma_2$   
444 (Morris and Ferrill, 2009), and could be linked to local stress release induced by the formation of  
445 the normal-sense system. More precisely, the softening occurring during the shear plane  
446 localization of normal-sense faults could be enough to imply simultaneously the permutation

447 between a vertical  $s_1$  to a horizontal and East-West  $s_1$  (local stress perturbation due to fault slip),  
448 especially if we considered a high  $K_0$  value. We underline that the similarity between dihedral  
449 angles for both the strike-slip and normal fault sets ( $130^\circ$ ), suggests that the material had the  
450 same behaviour during normal faulting and strike-slip faulting, and also that no significant  
451 vertical compaction occurred after fault formation (i.e. vertical compaction would have only  
452 increased the dihedral angle between normal faults). This observation suggests that the material  
453 reaches its maximum mechanical compaction during the development of the studied faults.

454 The occurrence of fault sets within specific beds or beds section of the limestone succession (Fig.  
455 4) suggests that additional factors affect the fault formation. Low-porosity and/or early  
456 cementation as well as the rock texture (distribution of grain size and nature of grains) in beds  
457 may certainly impede the initiation of such hybrid compactive faults, similarly to the well-known  
458 conditioning parameters described for deformation bands occurrence in porous sandstones  
459 (Fossen et al., 2018a and references therein). Moreover, external larger-scale factors such the  
460 sedimentation rate controlling the overloading stress or the tectonic features (tilting and slope,  
461 regional stresses) can also periodically create favourable or unfavourable conditions for fault  
462 initiation.

463

#### 464 *6.2. The specific case of multi-beds faults*

465 The specificity of multi-beds faults remains elusive. Most of their characteristics argue for a  
466 similar mechanical context of formation by compaction processes, under the influence of a NS  
467 trending  $s_3$ . These multi-beds faults cannot be considered as the evolution of bed-restricted ones  
468 as their slickensides show any evidences of segment connection and cut across several limestone

469 beds without marked plan bending. We propose that these multi-beds faults formed in an area of  
470 higher sedimentation rate consistently with the lack of significant emersion and the presence of  
471 several sedimentary breccias (debris flow from the thrust relief). This higher sedimentation rate  
472 coupled with less marked mechanical stratigraphy (fewer differences between rheology of  
473 limestone beds compared to more fluctuating palustrine environment) is proposed to explain the  
474 occurrence of such large-scale slickensides. The apparent absence of the conjugate set is also  
475 interpreted to be linked with a larger influence of the slope during fault formation. We note that  
476 these faults are not significantly reactivated during late Eocene Pyrenean shortening despite of  
477 their EW orientation and their location in front of the main thrust system. This suggests that these  
478 faults, from their deformation mechanisms, morphology, and diagenetic origin, are not suitable to  
479 form preferential plane of weakness.

480

### 481 *6.3. Comparison with other fault systems*

482 Other fault systems share several geometrical attributes with the fault system described in the  
483 present study. For example, distributed patterns of compactive faults have already been described  
484 as cataclastic shear-enhanced compaction bands in porous sandstones (Eichhubl et al., 2010;  
485 Ballas et al., 2013; Fossen et al., 2015) and porous limestones (Rustichelli et al., 2012; Rotevatn  
486 et al., 2016; Kaminskaite et al., 2019), or as deformation bands with particles reorientation-  
487 torsion in clayey materials (Ujjiie et al., 2004; Milliken and Reed, 2010). Furthermore, meter-  
488 scale shear structures showing both cataclasis and pressure-solution were described in porous  
489 limestones (Tondi et al., 2006; Rotevatn et al., 2016; Tavani et al., 2018b), chalk (Kaminskaite et  
490 al., 2019) or undetermined carbonate (Marshak et al., 1982). These examples and the fault system

491 described in the present study complete the range of localized deformation structures observed  
492 along end cap yield envelop of porous materials (Fossen et al., 2018b; Ferrill et al., 2021).  
493 However, these cited examples of pervasive compactional structures differ from our fault system  
494 as there are mainly deformation bands without localization of slickensides and they are mainly  
495 related to regional contractional stress fields.

496 Compaction of sediment during burial conditions is mainly diffuse across the entire material  
497 (Ehrenberg and Nadeau, 2005). The compaction can also take place in discontinuity such as bed-  
498 parallel pressure-solution seams (Toussaint et al., 2018; Meng et al., 2021) or compaction bands  
499 (Tondi et al., 2006; Aydin and Ahmadov, 2009). However, the participation of shear  
500 discontinuities in the sediment compaction remains mainly not considered although polygonal  
501 fault patterns observed worldwide in offshore sediments (Cartwright et al., 2003; Gay et al.,  
502 2004; Cartwright, 2011; Gay, 2017) are discussed to be triggered during sediment burial and  
503 diagenesis. In chalks, Hibschi et al., (2003) described polygonal faulting where shearing is  
504 associated with decrease of porosity by pressure solution and cementation. Conversely,  
505 Wennberg et al., (2013) described cataclastic compaction bands triggered by vertical loading  
506 during burial (no information of the fault pattern). In our study, faults appear to be a hybrid  
507 system between these two previously cited examples as cataclasis, pressure-solution and dynamic  
508 recrystallization is observed.

509 We note that the polygonal faults described by Tewksbury et al. (2014) show calcite  
510 mineralization along the fault plane which implies a dominant dilatant behaviour. The meter-  
511 scale compactive normal faults described by Guiraud and Séguret, (1987) in clayey sediments are  
512 also related to water-escape structures during sediment compaction (their low-angle to bedding  
513 and their slickenside morphology are quite similar to our study faults). These examples reveal



514 that different micro-mechanisms can occur within the shear discontinuities initiated during  
515 sediment compaction. These structures participate to sediment compaction by compactive  
516 mechanisms within the shear structures (cataclasis, pressure-solution) or by dilatant behaviour  
517 which permits fluid expulsion and diffuse compaction of the host material.

518

## 519 **7. Conclusion**

520 Compactive faults showing hybrid micromechanisms of deformation are described in micritic  
521 porous limestones from the South of France. These faults are organized in distributed patterns of  
522 meter-scale structures and form two sub-synchronous conjugate sets: a set composed of low-  
523 angle normal faults and a set composed of subvertical strike-slip faults. Their curve plane,  
524 incipient shear offset, high angle relative to  $s_1$ , and porosity reduction argues for a dominant  
525 compaction behaviour during faulting. Comminution, dynamic recrystallization (revealed by the  
526 crystallographic preferred orientation from EBSD acquisitions), pressure-solution and grain  
527 rotation are observed in these structures initiated during early sediment burial and basin flexure.  
528 These faults complete the range of compactive structures formed along the end cap yield envelop  
529 of porous materials. This example underlines the complexity of deformation processes occurring  
530 under compaction and early burial context, especially in such poorly consolidated carbonate  
531 material.

532

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## 902 **Caption List**

903 **Fig. 1. (a)** Schematic geological map of the study area modified from the 1/1 000 000 geological  
904 map of France. **(b)** SSW- NE Structural cross-section within the Aumelas - La Boissière basin  
905 using field cartographic measurements and the St Bauzille Well data. La Taillade is the study site.

906 **Fig. 2. (a)** Aerial pictures (GoogleEarth) showing the location of the studied outcrop along the  
907 A750 highway. The reported geological map is modified form the 1/50 000 maps of Montpellier  
908 (Bonnet et al., 1971) and Lodève (Alabouvette et al., 1982). Main bed dip values are reported for  
909 the three main parts of the outcrop. The main studied section is located westward. A second  
910 studied outcrop is located eastward but is potentially disconnected form the other studied  
911 exposures by map-scale sinistral faults. **(b)** Field view of the main studied limestone-dominated  
912 section. **(c)** Details of the studied outcrop showing mudstone beds. **(d)** Example of brownish silty  
913 marl deposits intercalated in mudstone beds.

914 **Fig. 3.** Host rock microstructure observed from SEM images. **(a)** 3D mudstone sample showing  
915 micritic euhedral crystals of calcite less than 5  $\mu\text{m}$  in size. **(b)** Image from thin-section showing  
916 imbricated contact between calcite particles.

917 **Fig. 4.** Sedimentological log of the main studied section showing location of studied faults along  
918 the succession. Evidence of pedogenetic alterations (desiccation, karstification, brecciation)  
919 observed on bed surfaces are also reported. Beds are draw following the Dunham classification  
920 (*M*: Mudstone, *W*: Wackstone, *P*: Packestone, *G*: Grainstone, *B*: Boundstone + *C* to identify  
921 clays). Studied fault symbol represents indifferently both normal-sense and strike-slip fault sets.

922 **Fig. 5.** Slickenside morphology. **(a)** example of slickenside from a normal-sense fault. **(b), (c),**  
923 **and (d)** Detailed observations of grooves and ridges morphology and tips. **(e)** SEM image  
924 showing slickenlines morphology at the microscale. **(f)** SEM image showing the 3D morphology  
925 of a step marking on a slickenside. Thick arrows show the sense of displacement of the absent  
926 wall.

927 **Fig. 6.** Normal fault system. **(a)** Limestone layers containing constrained normal-sense faults. **(b)**  
928 and **(c)** S-Shape of faults. **(d)** and **(e)** Cross-section of limestone beds showing the conjugate  
929 system of faults. White arrows show the fault kinematic. *S0* is the bedding (base of the succession  
930 at the bottom of the photography).

931 **Fig. 7.** Stereographic analysis of normal-sense fault set. **(a)** Stereograms showing projection of  
932 fault planes and slickenlines. **(b)** Histogram of fault dip. **(c)** Histogram of fault rake. (Restored  
933 data)

934 **Fig. 8.** Strike-slip fault system. **(a)** and **(b)** Examples of dextral and sinistral fault slickensides,  
935 respectively. **(c)** Bed surface showing the conjugate sets of strike-slip faults. **(d)** Detailed view of  
936 dilation jogs showing the fault kinematic and the centimeter-scale shear-displacement. White  
937 arrows show the fault kinematic.

938 **Fig. 9.** Stereographic analysis of strike-slip fault set. **(a)** Stereograms showing projection of fault  
939 planes and slickenlines (grooves). **(b)** Histogram of fault dip. **(c)** Histogram of fault rake.  
940 (Restored data)

941 **Fig. 10.** Interaction between both conjugate fault sets. **(a)** and **(b)** Continuity between normal-  
942 sense and strike-slip faults. **(c)** Schematic bloc diagram showing faults organization. **(d)**  
943 Photograph showing the pervasive distribution of both normal-sense and strike-slip fault systems.  
944 White and red arrows show the fault kinematic. *S0* is the bedding (base of the succession at the  
945 bottom of the photography). See the evidence of shear offset of the NNW-SSE trending sinistral  
946 fault fig. 10a.

947 **Fig. 11.** Microstructure of faults. **(a)** Mosaic of SEM images showing the rock microstructure  
948 below the fault slickensides and photograph of the block sample (see the orientation and dip of  
949 the fault). **(b)** Focus on the compacted zone above the main fault plane. **(c)** Example of the  
950 compressive jog showing compaction and evidences of enhanced pressure-solution. **(d)** Other  
951 example of the compacted zone below the fault plane. **(e)** Example of fault step marking at the  
952 upper tip of groove. This example illustrates the presence of dilation joint and the lack of  
953 compaction.

954 **Fig. 12.** Grain framework reconstruction maps from EBSD acquisition and rose diagram of grain  
955 slope (angle from the image top) in **(a)** the host rock, **(b)** and **(c)** the compressive jogs along the  
956 fault plane. The color of grain indicates the angle of their crystallographic preferred orientation  
957 with the different calcite crystallographic axis (*0001*: c-axis; *0110* and *1010*: a-axes).

958 **Fig. 13.** Crystallographic preferred orientation (CPO) in the fault zone, EBSD acquisitions. **(a)**  
959 Grain framework reconstruction maps from EBSD acquisition showing deformation zone above

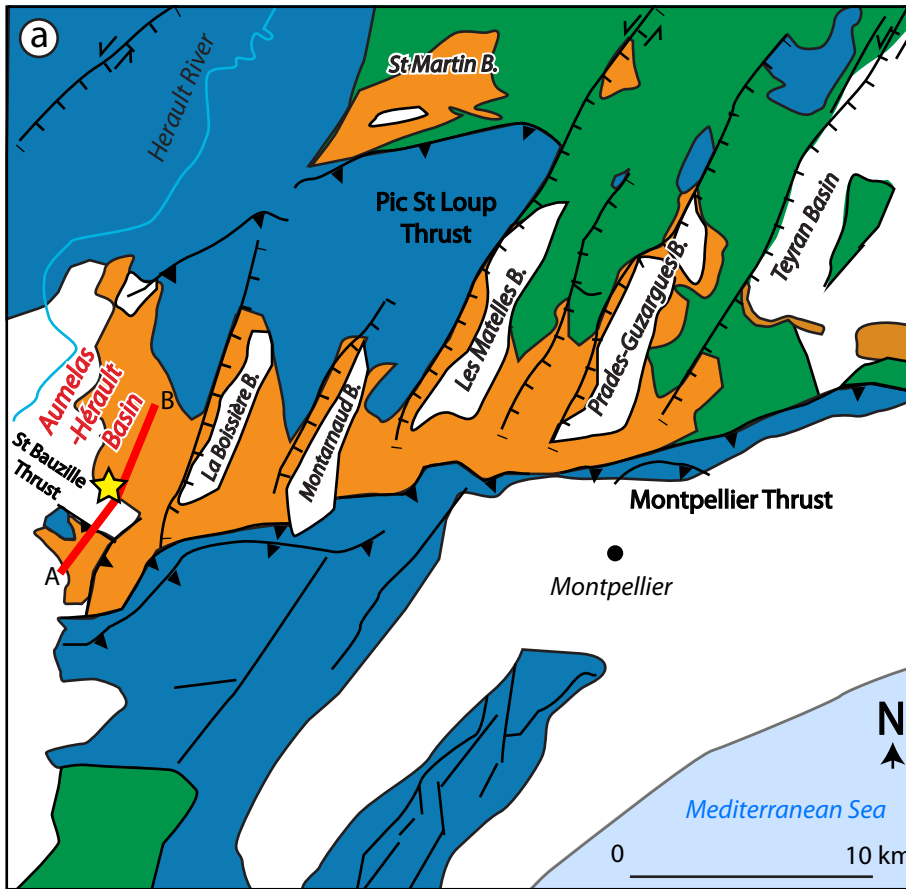
960 the main fault surface. **(b)** Calcite orientation data for grains of the fault zone (within dashed line  
961 in (a)). (*0001*: c-axis; *0110*, *1010* and *1120*: a-axes).

962 **Fig. 14.** **(a)** EBSD acquisitions showing example of calcite twinning within the fault core. Maps  
963 of calcite twins detected at the fault core **(b)**, a compressive jog **(c)**, and a step marking **(d)**. The  
964 calcite twins are draw by red lines. High concentration of calcite twins is observed in the fault  
965 core and along the compressive jog whereas no twins are observed in the host rocke and at  
966 marking dilatant step.

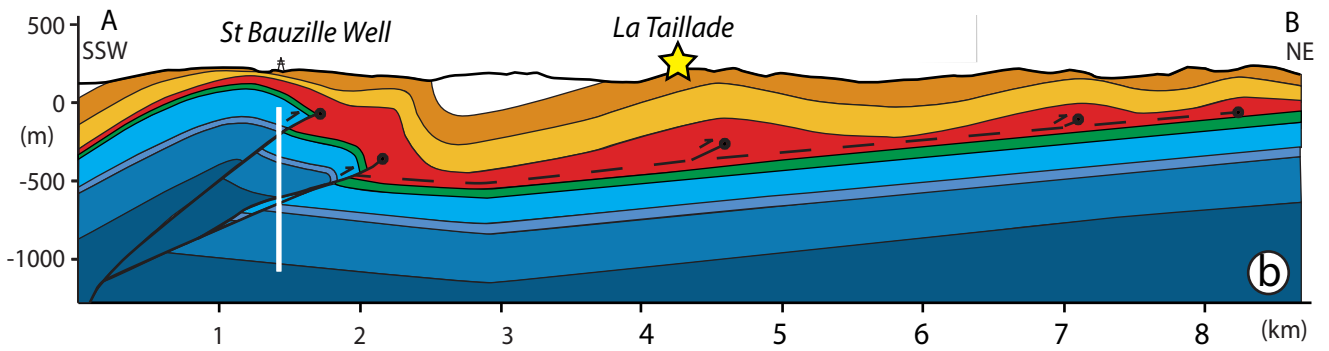
967 **Fig. 15.** Stress state during fault formation. **(a)** Stereogram projection showing state of stress  
968 determination by inversion method. (*r<sub>0</sub>*: stress tensor aspect ratio; *s<sub>1</sub>*, *s<sub>2</sub>*, *s<sub>3</sub>*, principal stress  
969 orientations). **(b)** Table showing stress inversion parameters. **(c)** Reduced Mohr circle. The faults  
970 are color coded as a function of their angular misfit range: 0-5° blue, 5-15° magenta, 15-30°  
971 green, 30-180° grey. The procedure is detailed in Appendix 1 of Burg *et al.* (1995). *N<sub>total</sub>*: Total  
972 number of fault slip data. *P<sub>inv</sub>*: Proportion of slip data used in the inversion.  $\Delta\lambda_{av}$ : Average  
973 angular misfit of the rake; this average is computed on the *P<sub>inv</sub>* lowest misfits only (Etchecopar  
974 *et al.*, 1981). *N<sub>exp15</sub>*: Number of fault slip data with rake misfit < 15°. *N<sub>0exp15</sub>*: Number of fault slip  
975 data with rake misfit < 15° that could be obtained from a random sample. *InvQual*: Quality index  
976 as in Heuberger *et al.* (2010).

977 **Fig. 16.** Multi-beds normal faults. **(a)** Exposure of such fault. **(b)** Detail of slickenside  
978 morphology showing slickolites (teeth evidencing pressure-solution at the down-slip ends of  
979 grooves). **(c)** Stereographic data showing faults organization (bedding restored to horizontal). **(d)**  
980 Example of such faults with their estimated shear-displacement. White arrows show the fault  
981 kinematic. *S<sub>0</sub>* is the bedding. *ds* is the apparent shear displacement.

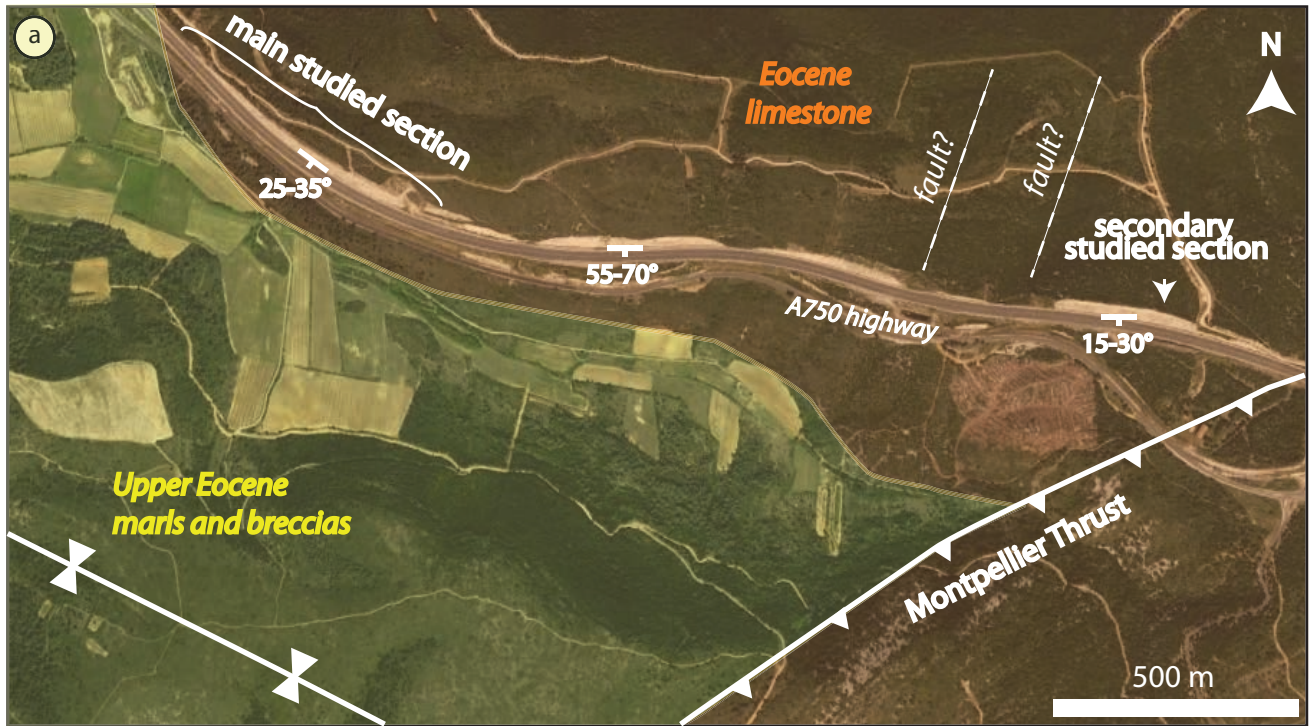
982 **Fig. 17.** Conceptual model of faults formation. **(a)** Differential Stress (Q) – Mean Effective Stress  
983 (P) space diagram showing location of faults formation using yield envelopes of porous carbonates  
984 (Risnes *et al.* 2005) and estimated stress path ( $K_0$  : coefficient of Earth Pressure at Rest; blue  
985 lines: estimated stress paths). **(b)** The successive stages of faults formation: comminuted zone  
986 with dynamic recrystallization (stage 1), shear localization (stage 2), and early blocking of fault  
987 movement by grain rotation and pressure-solution processes (stage 3). **(c)** 3D schematic bloc  
988 diagram showing the location of fault in the basin and the influence of the slope.  
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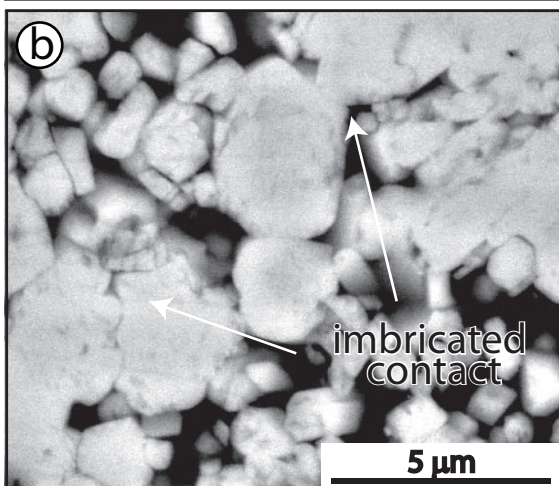
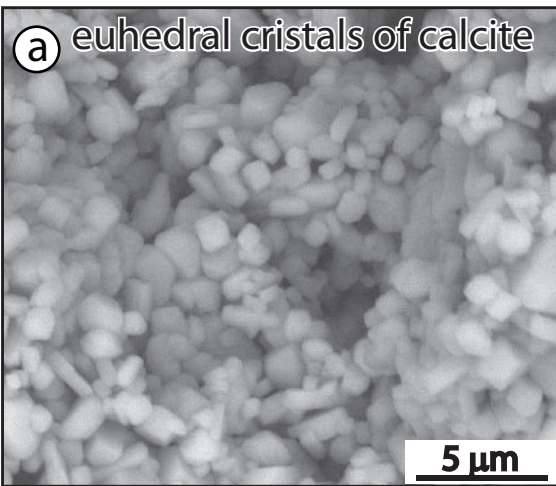
- ★ Study site
  - ▲ Thrust Fault
  - ↔ Strike-slip Fault
  - ⚡ Other Faults
  - Upper Eocene-Quaternary
  - carbonate
  - marl - sand
  - breccia - clays
  - Cretaceous
  - Jurassic
- Paleocene - Eocene Foreland Basin**

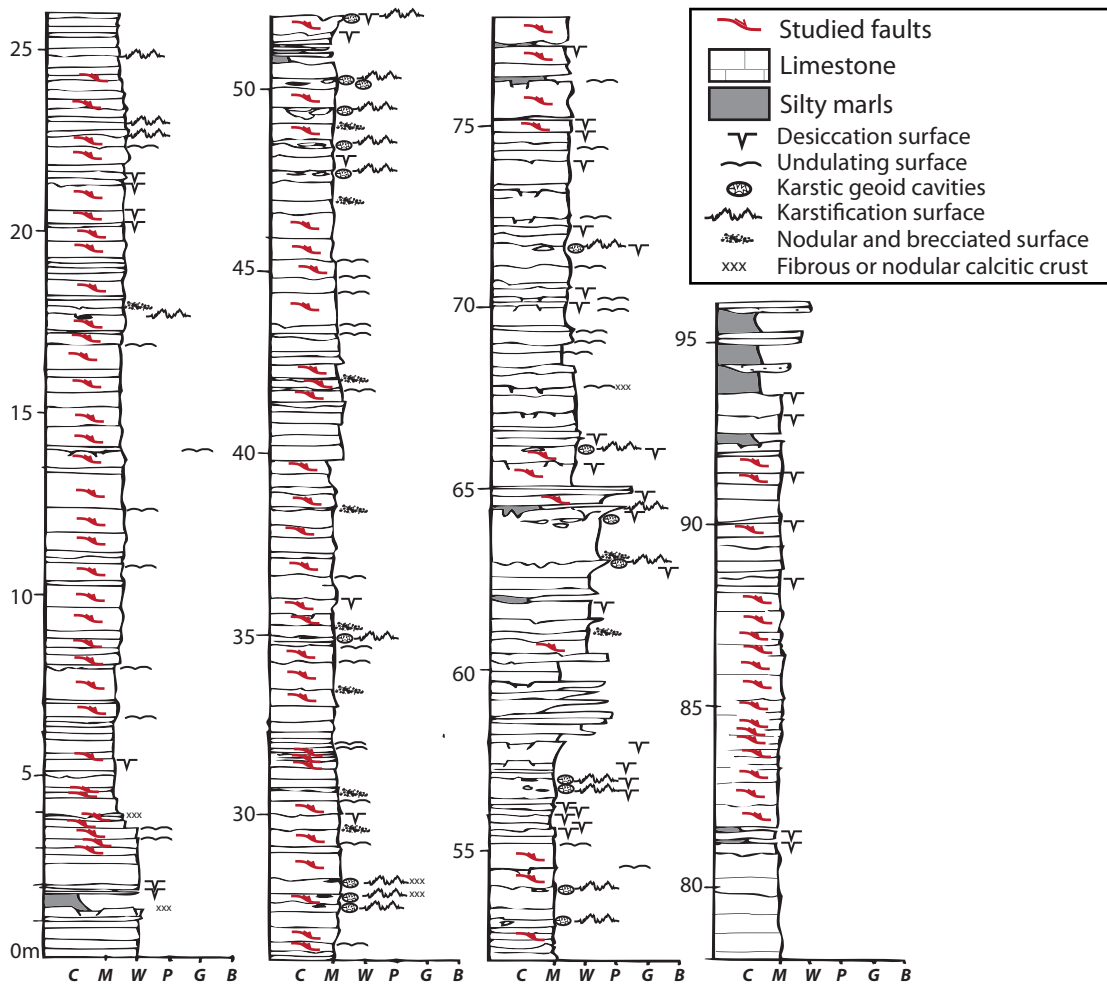


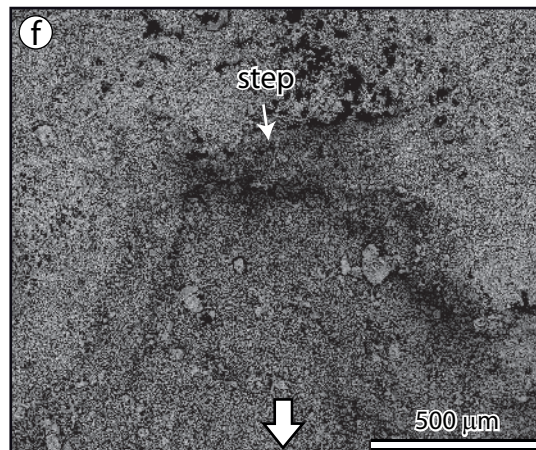
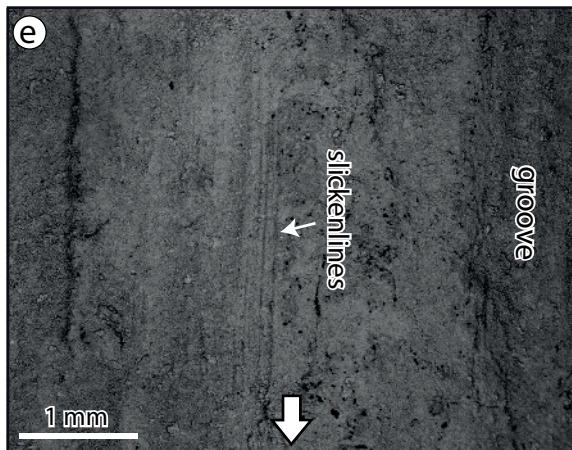
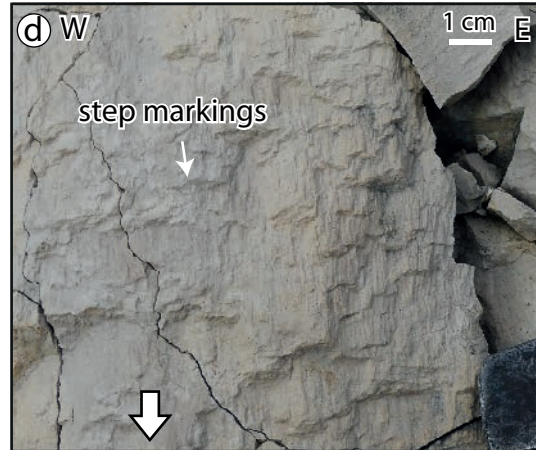
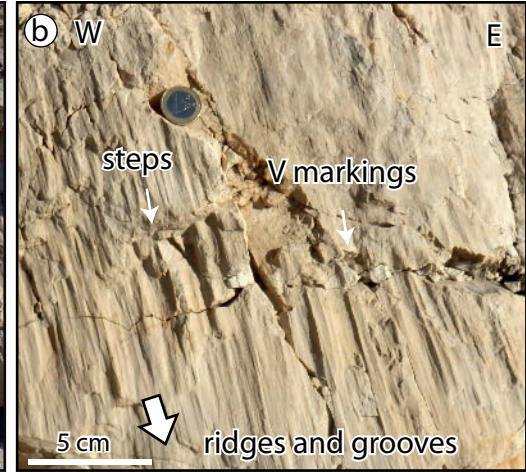




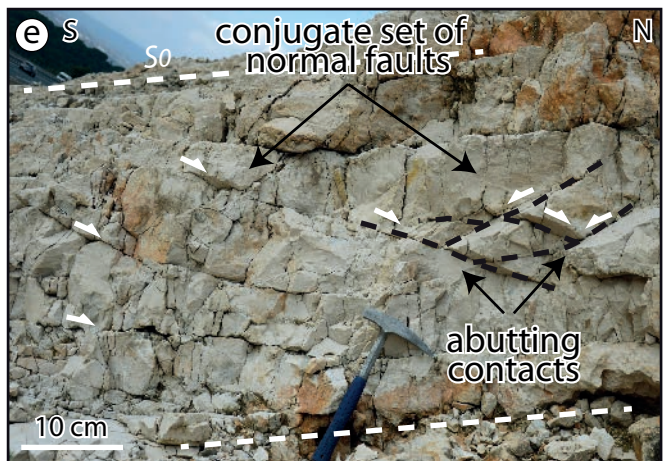
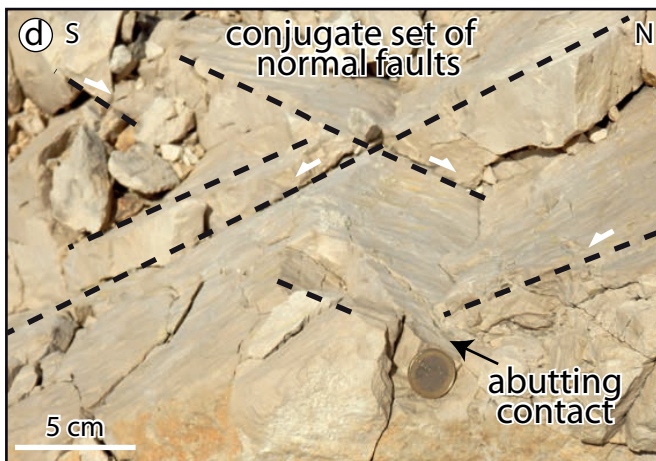


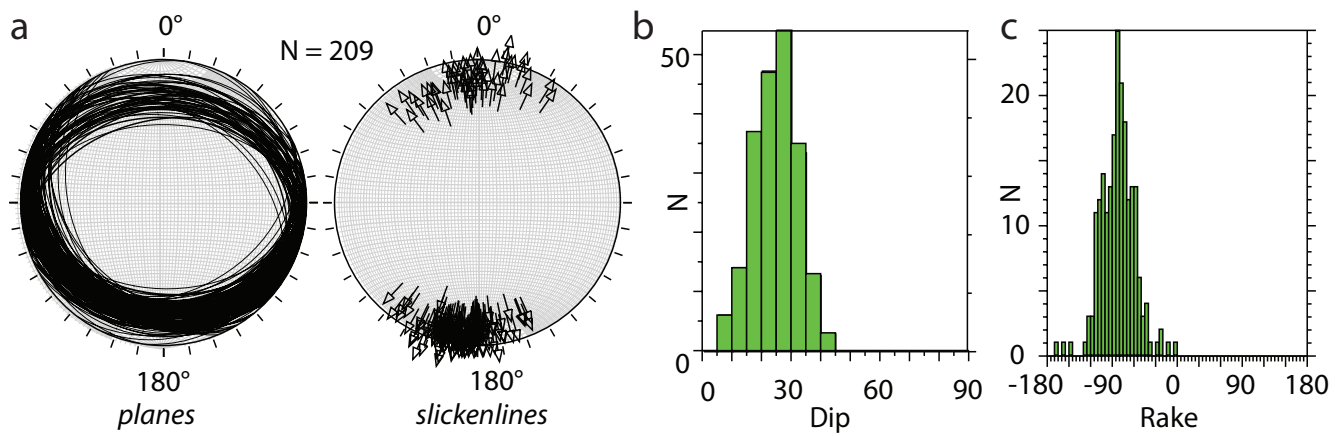




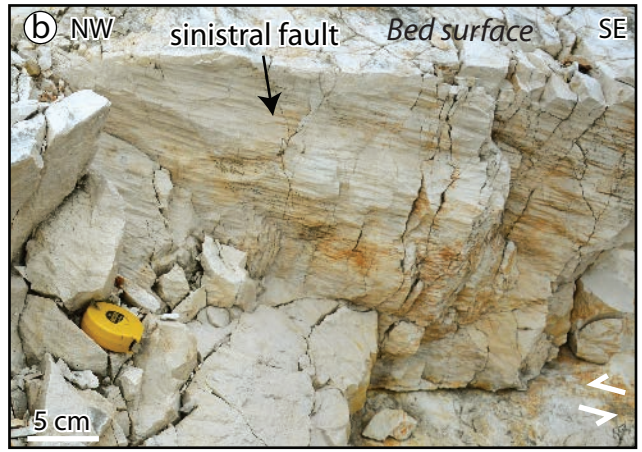
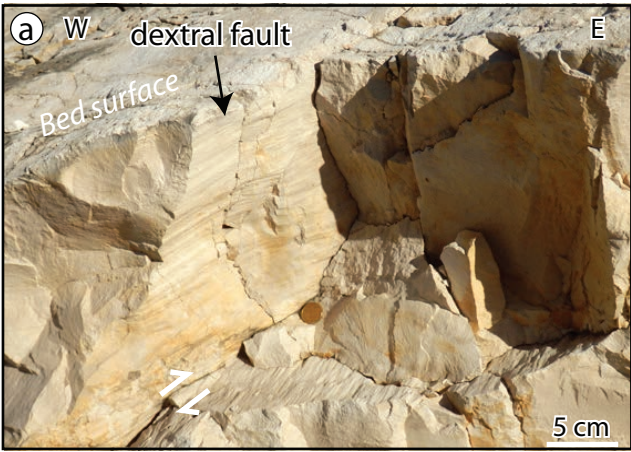


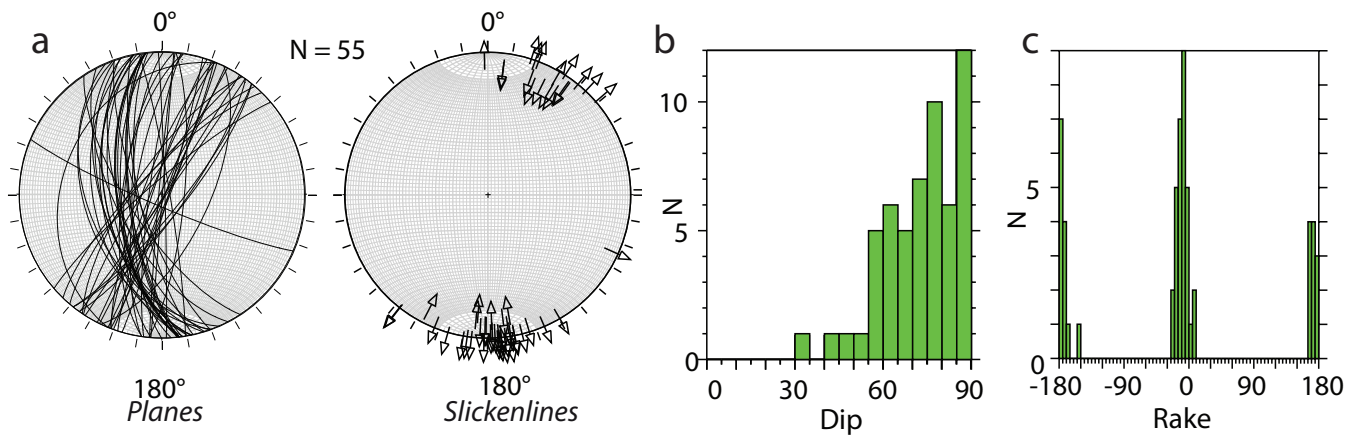




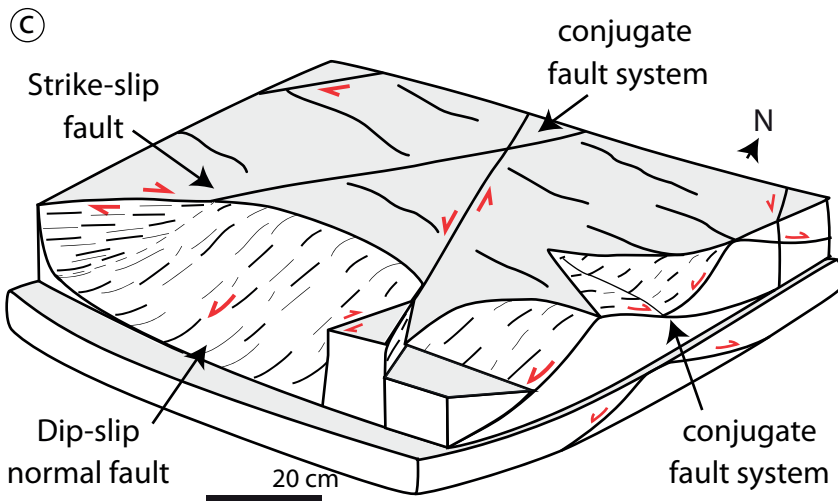
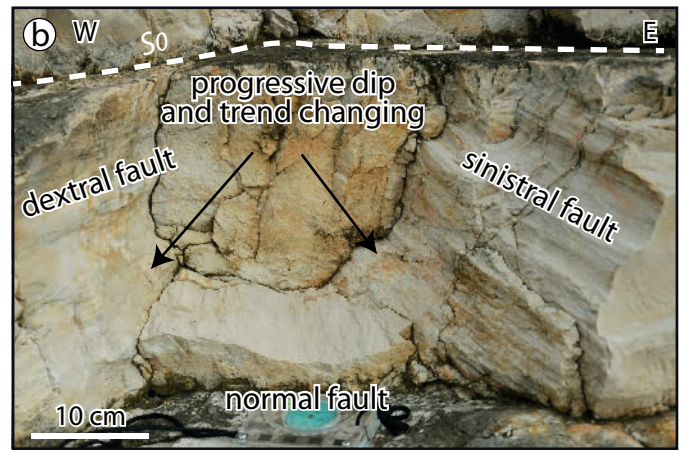
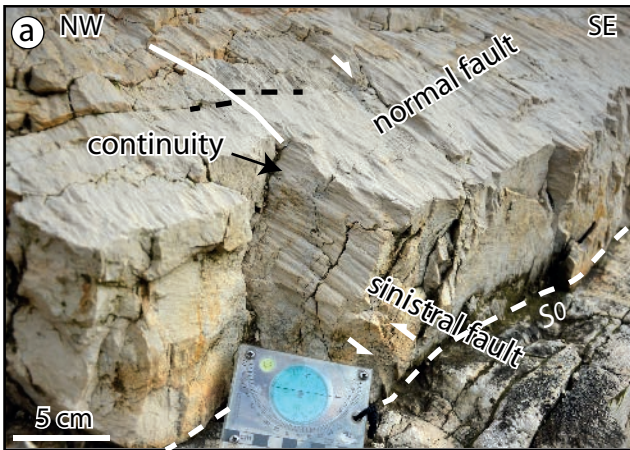




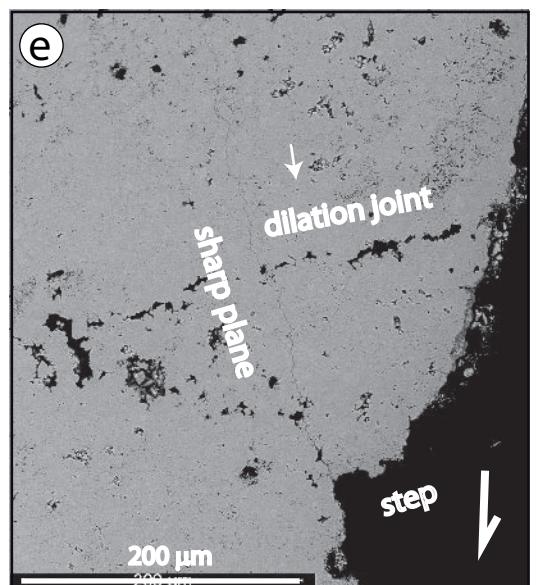
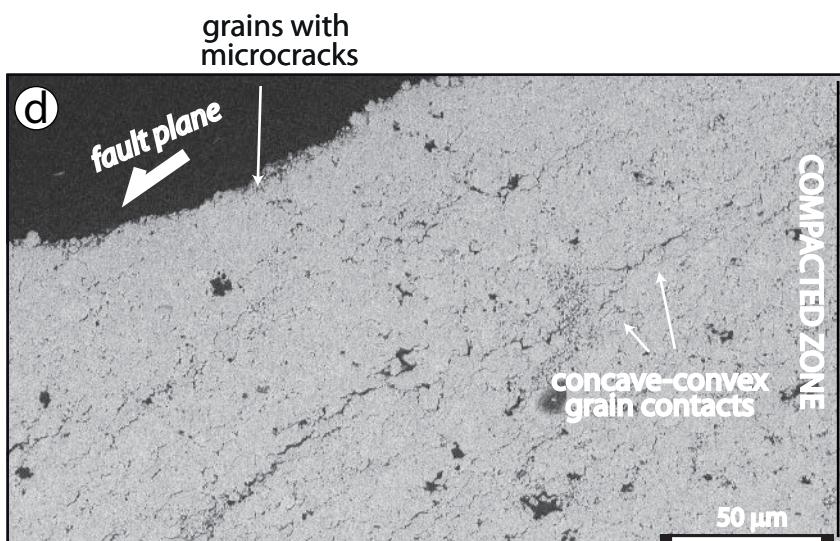
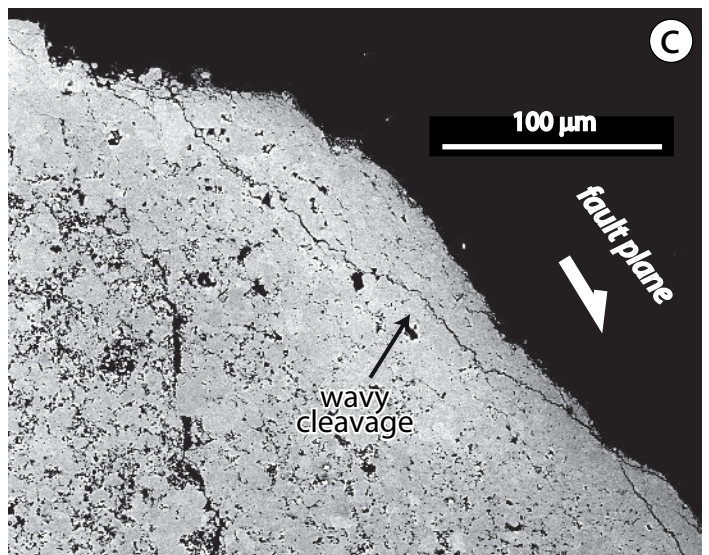
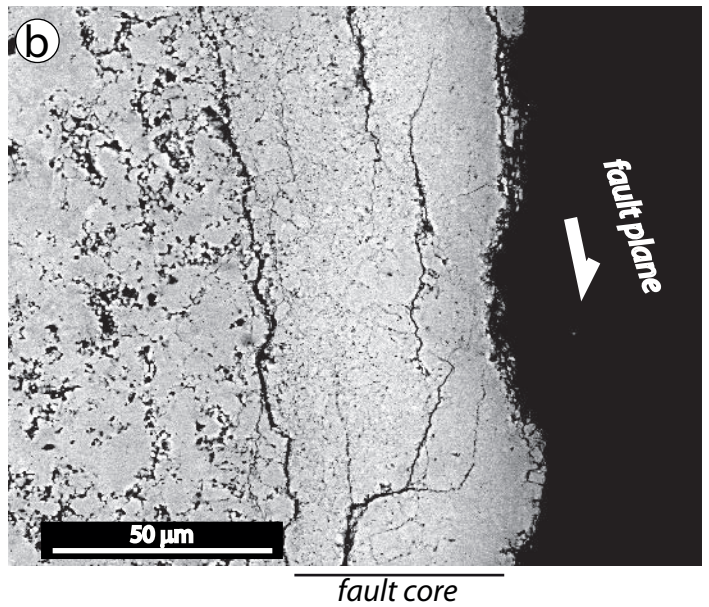
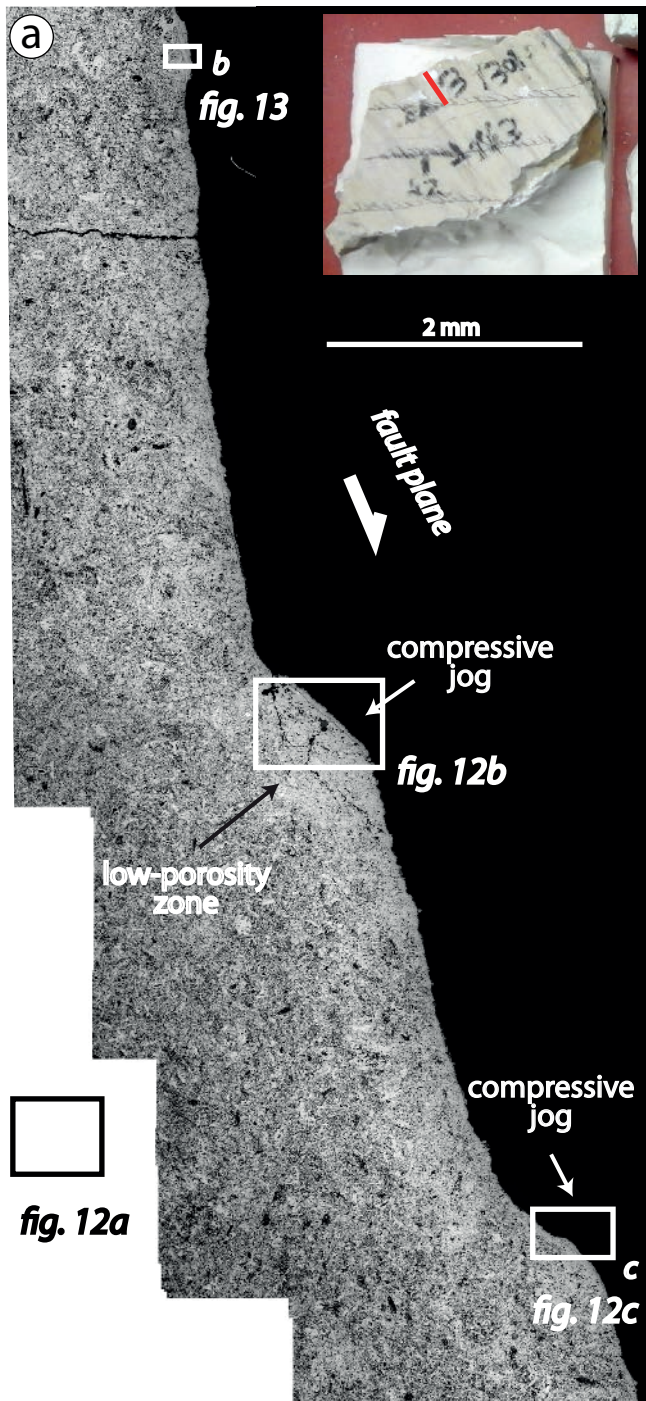




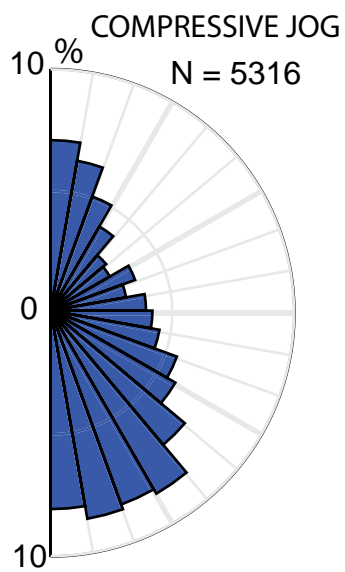
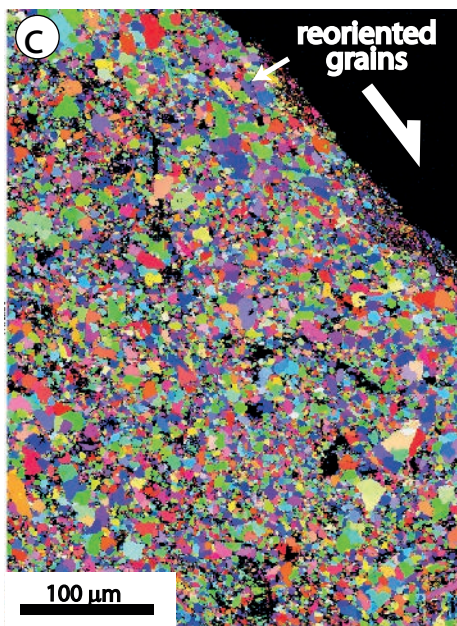
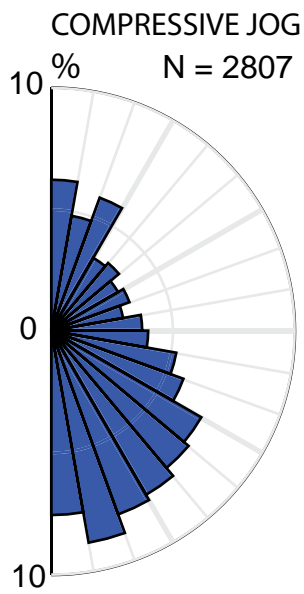
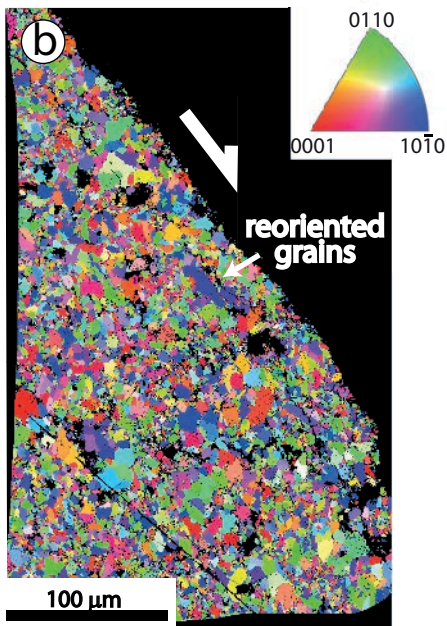
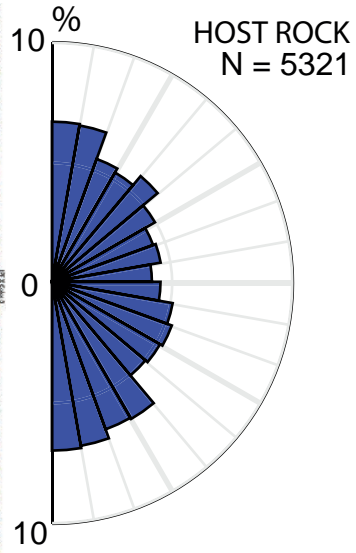
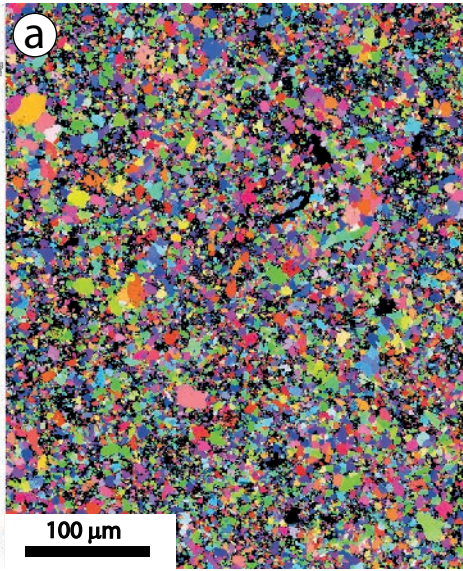


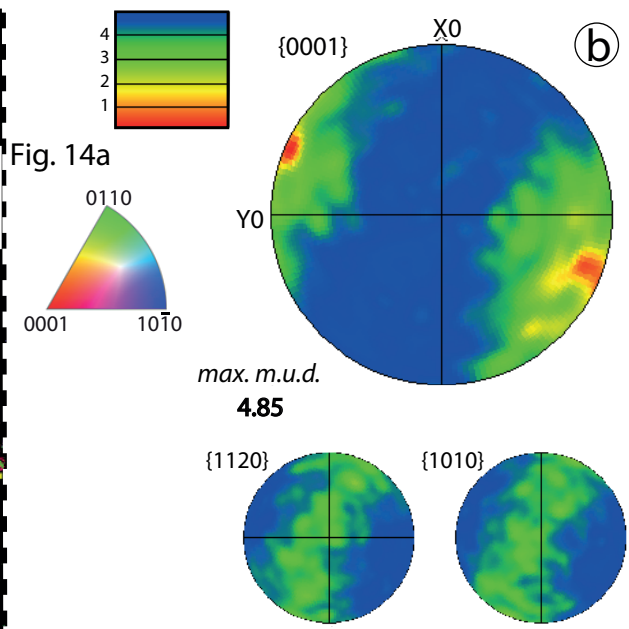
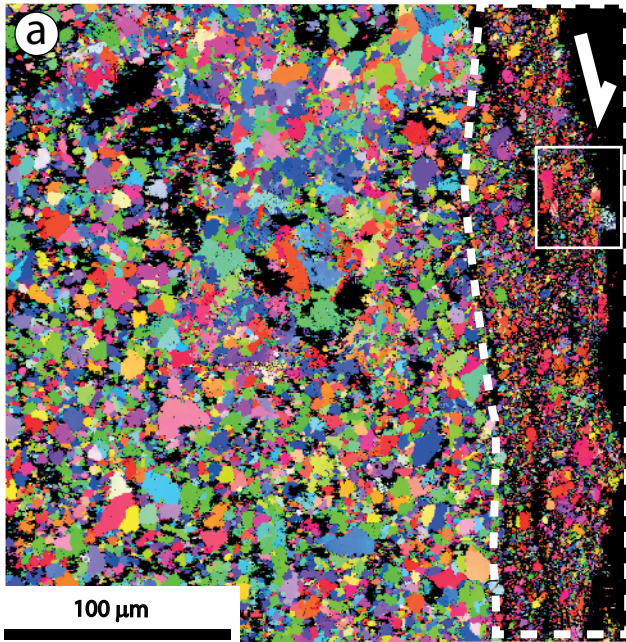




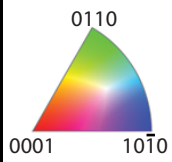
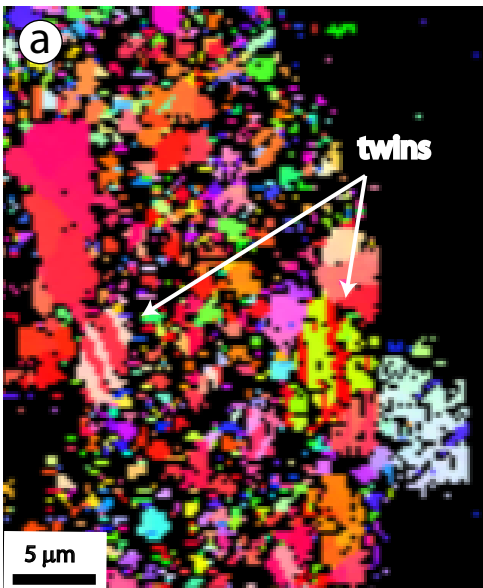




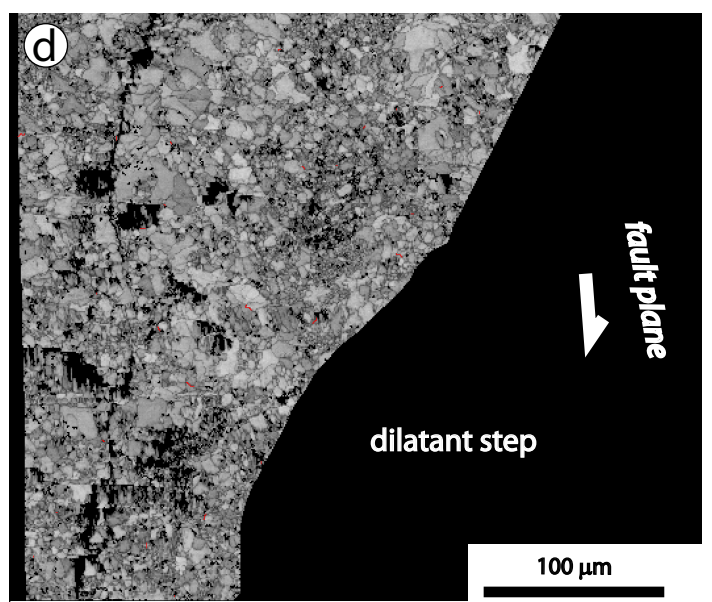
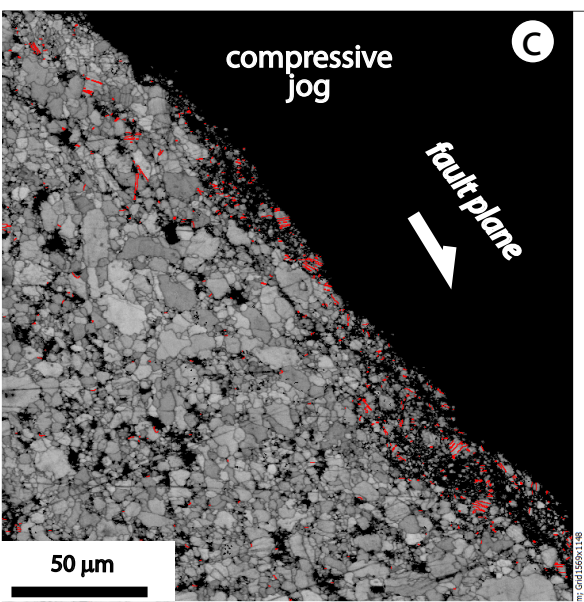
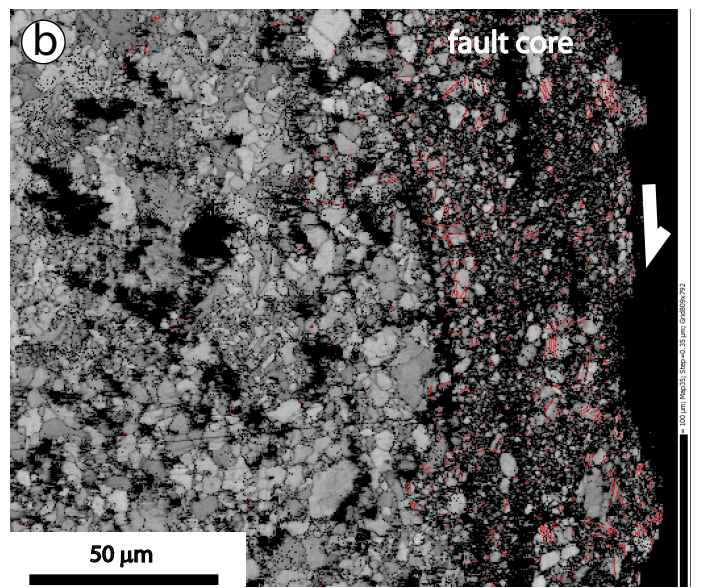


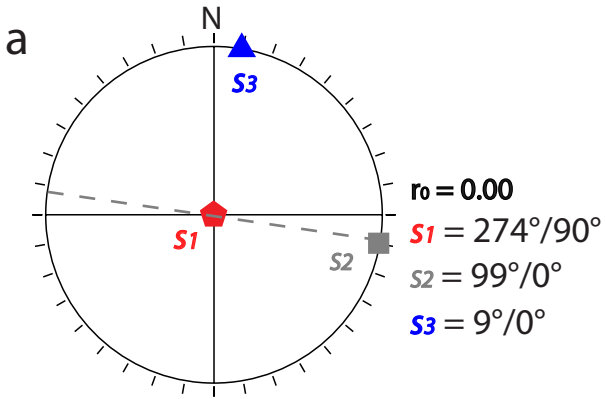






twins  
location





b

Stress tensor	$N_{total}$	$P_{inv}$ (%)	$\Delta\lambda_{av}$ ( $^\circ$ )	$N_{exp15}$	$NO_{exp15}$	$InvQual$
	264	99	9.6	216	25.7	2

c

