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1 Structural and petrological characteristics of a Jurassic detachment
2 fault from the Mont-Blanc massif (Col du Bonhomme area, France)

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24 **domain**

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37 **Abstract**

38 Necking zones, located between the proximal and distal domains of passive margins, are
39 characterized by important crustal thinning and the occurrence of extensional detachment
40 faults. Here we provide a petrographic, microstructural and geochemical study of an
41 extensional detachment fault discovered in the Mont-Blanc massif, which was located in the
42 necking zone of the European margin during Early Jurassic rifting. The footwall of the fault
43 consists of basement gneisses, while its hanging wall presents pre- to syn-rift deposits. A
44 careful observation of the basement / cover contact also reveals the presence of a strain
45 gradient affecting the Variscan gneisses, which are progressively deformed to cataclasite and
46 to black gouges at the top basement, and sealed by undeformed Jurassic syn-rift sandstones,
47 the Grès Singuliers, which rework fault rock pebbles. Microscale observations also indicate
48 fluid-assisted deformation, characterized by the total breakdown of feldspars and the coeval
49 formation of white mica, leading to strain softening along the fault plane. The overall
50 observations point toward a rift-related detachment fault responsible for the exhumation of
51 the internal Mont-Blanc basement, an interpretation which challenges previous models of
52 crustal exhumation in the External massifs of the Alps. Finally, the preservation of an
53 undeformed syn-rift sedimentary cover sealing a tectonized top basement makes this area a
54 fossil analogue for detachment faults in necking domains. It demonstrates that preservation
55 of pre-Alpine, rift related structures is possible even in highly deformed areas, and
56 emphasizes the role played by pre-Alpine inheritance on the building of the Alpine Mountain
57 belt.

58 **Introduction**

59 Recent investigations suggest that orogens resulting from a Wilson cycle (Wilson
60 1966; Wilson et al., 2019) are strongly controlled by rift-related inheritance (e.g., Jammes
61 and Huisman, 2012; Chenin et al., 2017; Jourdon et al., 2019). Such inheritance is largely
62 responsible for the position of major decoupling levels that can be reactivated during
63 convergence, and thus control the final architecture of the orogen, as recently demonstrated
64 for the Pyrenean system (Cadenas et al., 2018; Espurt et al., 2019; Izquierdo-Llavall et al.,
65 2020; Labaume and Teixell, 2020; Lescoutre and Manatschal, 2020; Manatschal et al., 2021)
66 or the eastern Alps (Mohn et al., 2011; Mohn et al., 2014). Paleogeographic reconstructions
67 suggest that the same may apply to the Western Alps (Mohn et al., 2011). Indeed, the
68 present-day External Crystalline Massifs (ECM) are located at the transition between the

69 proximal and the distal domains of the former European margin, which formed during
70 Jurassic rifting and hyperextension (Mohn et al., 2014). However, little work has yet been
71 done on rift-related, pre-Alpine structures in the internal part of the ECM, and their possible
72 impact on the subsequent Alpine reactivation (Butler, 1986, 2017). Jurassic rift-related
73 faults were identified such as the Ornon fault (Belledonne massif; Barfety and Gidon, 1984;
74 Lemoine et al., 1986) or inferred from variations in the thickness of lower Jurassic
75 sedimentary rocks on either side of the ECM (Lemoine and Trümpy, 1987; Bellahsen et al.,
76 2012; Bellahsen et al., 2014), and interpreted as a high-angle normal fault associated with
77 limited crustal thinning. The geometry of these faults contrasts with the large offset normal
78 faults expected at the transition between the proximal and distal domains, as observed at
79 present-day rifted margins (e.g. Osmundsen and Péron-Pinvidic, 2018). Recently, Ribes et
80 al.(2020) proposed the presence of a rift-related, Jurassic detachment fault in the internal
81 Mont-Blanc massif, responsible for the tectonic exhumation of the basement leading to a
82 local emersion and reworking of the Variscan basement in the sediment, as documented in
83 the Lower Jurassic sandstones of the Grès Singulier Formation. The occurrence of such a
84 detachment system predicts important crustal thinning, which is not only in accordance with
85 observations made at present-day rifted margins (Osmundsen and Péron-Pinvidic, 2018;
86 Muñoz-Barrera et al., 2020; Ribes et al., 2020) but is also compatible with the subsidence
87 history observed at the transition between the proximal and distal domains in the Alpine
88 domain (Masini et al., 2014; Ribes et al., 2019b). However, this interpretation is mostly
89 based on field structural and sedimentological observations, calling for a more detailed
90 structural, petrological, and geochemical characterization of the detachment fault previously
91 proposed by Ribes et al. (2020). Corroborating the existence of such a detachment surface is
92 not only important for the understanding of the pre-Alpine evolution and the structure of
93 the European margin, but it also strongly impacts the interpretation of the Alpine orogenic
94 evolution. Its existence would have major implications for the evolution of the Mont Blanc
95 Massif and on a larger scale for the interpretation of how the transition between the internal
96 and external parts of the Western Alps can be interpreted and how far they are
97 linked/controlled by rift inherited structures.

98 In this contribution, we focus on the Col du Bonhomme area, where the proposed
99 detachment surface has been described in most detail by Ribes et al. (2020). Field,
100 microstructural and petrological observations, and elemental geochemical analyses are used

101 to elucidate the kinematics, the mineralogical and geochemical characteristics, and the
102 relative timing of this structure. Our aim is to determine whether this structure was
103 inherited from Jurassic rifting or if it has an Alpine origin.

104

105 **1. Geological setting**

106 The Alpine orogen was formed as a result of the closure of the Alpine Tethys since
107 the Cretaceous. Its evolution is linked with a complex, polyphase kinematic evolution leading
108 to the subduction of the oceanic, proto-oceanic and exhumed subcontinental mantle
109 domains and the subsequent collision of the Jurassic Adriatic and Europe margins (Ford et
110 al., 2004; Rosenbaum and Lister, 2005; Fig.1). Several studies pointed out the link between
111 the main Alpine tectonic units, and the fossil rift domains of the Alpine Tethys (Argand,
112 1916; Debelmas and Lemoine, 1970; Lemoine et al., 1986; Lemoine and Trümpy, 1987;
113 Stampfli and Hochard, 2009; Bellahsen et al., 2014; Mohn et al., 2014). Indeed, wide-spread
114 Jurassic rifting, which initiated with distributed extension during the Early Jurassic (e.g.
115 Bourg d'Oisans basin, Lemoine et al., 1981; Barfety and Gidon, 1984; Chevalier, 2002),
116 resulted in a complex arrangement of hyperextended, exhumed mantle and proto-oceanic
117 domains (e.g., Valais and Liguro-Piemont domains of Dumont et al., 1984; Lemoine and
118 Tricart, 1986; Lagabrielle, 1987; Bousquet et al., 2002; Schmid et al., 2004; Loprieno et al.,
119 2011; Beltrando et al., 2012; Ribes et al., 2019c) during the late Middle Jurassic to
120 Cretaceous. The external units (Jura, Bauges, Bornes, Chartreuse, Vercors) have been
121 commonly assigned to the proximal domain of the former European margin (Fig.1 A,B,C,D;
122 Tricart et al., 2000). They are characterised by the presence of high-angle normal faults of
123 lower Jurassic age associated to half-graben type basins filled with shallow marine sediments
124 (Lemoine et al., 1981). In contrast, the internal units originate from the distal part of the
125 European margin, corresponding to the hyperextended domain, the exhumed mantle and
126 the oceanic domain. Due to important Alpine tectonic overprint, most of the distal parts of
127 the margin have been intensively deformed during subduction (e.g. Pré Piémont domain
128 Pantet et al., 2020), collision and subsequent exhumation (e.g. oceanic domain HP-LT
129 Schistes Lustrés units (Elter, 1971; Lagabrielle et al., 2015). Consequently, rift-related
130 structures are only preserved in a few places (e.g. Petit Saint Bernard (Beltrando et al.,
131 2012); Tasma (Ribes et al., 2019c), Chenaillet (Manatschal et al., 2011), Aiguille Rouge
132 D'Arolla (Decrausaz et al., 2021). A key result of these studies is that the external and

133 internal units show contrasting features not only in their stratigraphic and metamorphic
134 imprint, but also in their inherited structures with major implications for their Alpine
135 tectonic history.

136 The ECM (e.g., Aar, Mont-Blanc, Belledonne, Pelvoux, Argentera Massifs), which
137 belong to the external units of the Alps, are commonly interpreted to derive from the
138 transition between the proximal and distal domains of the European margin (former necking
139 domain; Butler, 1986; Butler, 1989; Mohn et al., 2014; Fig.1C and D). At present, they floor
140 the Helvetic and Penninic nappes (Fig.1B), and reached greenschist P-T conditions during
141 Alpine nappe stacking (Bousquet et al., 2008; Rosenberg et al., 2021). Most of the rift-
142 related structures have been overprinted by Alpine deformation, as for example the
143 reactivation of former normal faults in the Aar massif (Nibourel et al., 2021). Therefore, most
144 of the rift history has been deduced from the facies, thickness and paleo bathymetric
145 evolution of the syn- and post-rift deposits (e.g., Sangle unit; Epard, 1990 and Ultrahelvetic
146 units; Trumpy, 1951) and from the difference of geometry between large fold and thrust
147 nappes (e.g., Vervex unit (Epard, 1990) or Morcles-Doldenhorn nappe (Kiss et al., 2020)).

148 The Mont-Blanc Massif, which is part of the ECM, is a Variscan polymetamorphic
149 basement made up of paragneisses, cordierite-bearing migmatites and granitic intrusions in
150 its western part (external Mont-Blanc massif), and by a large pluton of metaluminous, ferro-
151 potassic granite surrounded by migmatites in its eastern part (internal Mont-Blanc massif;
152 Bussy, 1990). The pluton corresponds to a post-orogenic magmatism related to extensional
153 tectonic (orogenic collapse or back-arc extension), with a mantle-derived source associated
154 to crustal contamination (Bussy et al., 2001). U-Pb dating of the cordierite-bearing
155 leucosomes and of the granites showed that this late Carboniferous magmatic event lasted
156 about 10 Ma in the ECM (migmatization took place at ca. 307 ± 2 Ma and granite
157 emplacement between 307 ± 3 Ma (Montenvers granite) and 303 ± 2 Ma for the Mont-Blanc
158 granite (Bussy et al., 2001)). The absence of a clearly defined metamorphic contact aureole
159 around the Mont-Blanc granite, noted by Bussy (1990), may indicate an emplacement at
160 significant depth, with pressures up to 3.8 kbar at the top of the intrusion, i.e., an assumed
161 intrusion depth at 9 to 12 km for the pluton roof (Marshall et al., 1997). Cooling ages and P-T
162 conditions were obtained from skarns located immediately around the granite (i.e., on the
163 pluton walls), providing temperatures of 400/500°C and pressures of 4/8 kbar (based on Ti
164 thermometry on amphibole and stable isotopes) in Mont-Chemin ore deposits, dated at 303-

165 335 Ma using Ar/Ar on amphiboles (Marshall et al., 1997). Most studies assumed that post-
166 Variscan exhumation to the surface of the external Mont-Blanc Massif was achieved during
167 the Permian (Marshall et al., 1997). It was followed by the development of a Permo-Triassic
168 erosional surface, as observed in the external Mont-Blanc Massif and the Aiguille-Rouge
169 Massif located westward (Epard, 1986). In the western areas, the exhumation was further
170 constrained by the occurrence of Ladinian to Carnian tidal sandstones (Demathieu and
171 Weidmann, 1982) known as the Vieux Emosson Formation (Epard, 1989) directly overlying
172 the basement. Upper Triassic dolostones and evaporites (Arrandellys Fm. ; Epard, 1989)
173 complete the Triassic sequence.

174 The conditions of exhumation in the internal Mont-Blanc Massif are more debated.
175 Most studies considered that similarly to the external part of the massif, the Triassic deposits
176 form a primary, stratigraphic contact with the eroded basement (Eltchaninoff-Lancelot et al.,
177 1982; Epard, 1990). In contrast, Ribes et al. (2020) have proposed that the contact between
178 the pre-rift deposits and the basement corresponds to a Jurassic detachment fault, i.e., that
179 they are allochthonous relative to the underlying exhumed basement. A key observation is
180 the presence of early syn-rift sandstones (Sinemurian/Pliensbachian) corresponding to the
181 Grès Singuliers Formation (Landry, 1976; Eltchaninoff, 1980; Epard, 1989; Ribes et al., 2020)
182 that overlies a deformed and exhumed basement (Fig. 2). These sandstones, up to 50 meters
183 thick, have been interpreted as the result of local uplift and erosion of the Mont Blanc
184 basement associated either with a normal fault (Epard, 1990; Meyer, 2002), or a large offset
185 extensional detachment fault (Ribes et al., 2020). Ribes et al. (2020) proposed that
186 exhumation and related uplift of the internal Mont Blanc basement occurred during rift-
187 related necking at around 185 ± 5 Ma (Pliensbachian), resulting in sub-areal exposure of a
188 core complex type structure. It is interesting to note that the Mont-Blanc core complex of
189 Ribes et al. (2020) can be well compared to the Mont-Blanc island whose presence was
190 proposed and discussed in Trümpy (1971) and Loup (1992) to explain the occurrence of the
191 Grès Singuliers and its primary contact with the basement. During the Aalenian
192 transgression, the core complex/island was drowned and the Triassic to lower Jurassic
193 deposits were covered by Aalenian to Oxfordian black shales of the Dugny Formation
194 (Lemoine et al., 1986; Meyer, 2002, Fig.2). These were followed by Upper Jurassic to
195 Cretaceous limestones.

196 In the Mont-Blanc massif, the Alpine deformation corresponds to: (i) the thrusting of
197 the more internal Helvetic and Penninic nappes over the Mont-Blanc during the Late Eocene
198 to Early Miocene (Leloup et al., 2005; Egli and Mancktelow, 2013); followed by (ii) thick-skin
199 deformation leading to the closure of the Morcles basin (Escher et al., 1993; Herwegh and
200 Pfiffner, 2005) ; and (iii) to thrusting of the Mont-Blanc basement over the Aiguille-Rouges
201 basement (Mont-Blanc shear zone) from middle Miocene to early Pliocene (Leloup et al.,
202 2005). This thrusting was associated with the back thrusting of the basement over sediments
203 along the eastern/internal border of the Massif (Guermani and Pennacchioni, 1998; Leloup
204 et al., 2005; Egli and Mancktelow, 2013). Studies of metamorphic assemblages (muscovite-
205 biotite-chlorite-epidote), U-Th-Pb geochronology on allanite and Ar/Ar geochronology on
206 phengite, biotite and phlogopite indicate that the Alpine metamorphic peak was reached
207 between 28 and 15 Ma, with pressures of 0.5 GPa and temperatures up to 400°C in the
208 basement (Rolland et al., 2003; Rolland et al., 2008; Cenko-Tok et al., 2014). It was followed
209 by Neogene to present day exhumation of the massif (Glotzbach et al., 2008; Boutoux et al.,
210 2016; Herwegh et al., 2020).

211 In this study, we focus on the Col du Bonhomme area located to the NW of Bourg St-
212 Maurice (France). It corresponds to an antiformal fold (Fig.1E) preserving the cover of the
213 southern Mont-Blanc Massif below the Helvetic nappes, thus giving access to the pre-Alpine
214 contacts. In this area (Fig.2B), the syn-rift Grès Singuliers lie unconformably above the pre-
215 rift Triassic cover and the basement. The nature, age, and significance of the contact
216 between the sedimentary cover and the basement is the main goal of our study.

217

218 **2.Methodology**

219 The geological maps of Meyer (2002) and Ribes et al. (2020), complemented by
220 Google Earth satellite images, field observations and drone acquisitions, were used to build a
221 new 11 km² map of the area (Fig.2). Field observations were focused on the facies and
222 thickness distribution of the syn- and pre-rift sedimentary cover, their relationship with the
223 basement, as well as on the recognition and structural characterization of different fault
224 generations and related fault rocks (cataclasites, fault breccias and gouges). In total 71
225 samples were collected (33 fault rocks; 33 samples from the sedimentary cover and 5 from
226 the unfractured basement).

227 In order to quantify the peak metamorphic temperatures reached during Alpine
228 metamorphism in the Col du Bonhomme area, Raman spectroscopy of carbonaceous
229 material (RSCM) analyses were performed on 2 polished thin sections of organic-rich
230 Aalenian shales and one section of pre-rift limestone. A total of 55 Raman spectra were
231 acquired with a Horiba Jobin-Yvon Labram II spectrometer coupled to a microscope, at the
232 Institut des Sciences Moléculaires, Université de Bordeaux, France. Acquisitions parameters
233 include a 514.5 nm argon laser, an extended scanning mode (500-2000 cm^{-1}) and an
234 acquisition time of 30s. Spectra of the first-order region of carbonaceous material were
235 processed using the Fityk software (Wojdyr, 2010). Each spectrum was fitted with 4 bands
236 (D1, D2, D3, and the graphite band G) with Voigt profiles. The calibration of Beyssac et al.
237 (2002), based on the parameter $R2 = D1/(G + D1 + D2)$ was used.

238 Petrographic observations of polished thin sections from sampled rocks ($n = 71$) were
239 performed with a Nikon LV150ND optical microscope, and by cathodoluminescence (CL)
240 using a Cathodyne system (OPEA) with an operating condition of 15 kV–18 kV and a gun
241 current of 300–350 mA under a 60 mTorr vacuum (LFCR, Pau, France). Other techniques
242 include a Zeiss EVO50 scanning electron microscope (SEM) coupled to EDS mapping (20kV
243 beam; PLACAMAT, Pessac, France). Electron Probe Micro-Analyzer (EPMA) analyses were
244 carried out on polished thin sections using a Cameca SXFive instrument at the Centre de
245 Microcaractérisation Raymond Castaing in Toulouse (France), calibrated with natural and
246 synthetic in-house standards: aluminium oxide (Al), wollastonite (Si, Ca), sanidine (K),
247 manganese titanite (Mn), hematite (Fe), albite (Na), barite (Ba), celestine (Sr). For the
248 determination of major element compositions on silicate from the fault rocks and phosphate
249 in the basement minerals, the operating conditions were set at 15 kV, with a probe current
250 of 50 nA, and a 2 μm beam. For elemental mapping, similar conditions, and a step size of 2
251 μm were used.

252 Following microscopic acquisitions, the XMapTools software (Lanari et al., 2014;
253 Lanari et al., 2019) was used to build quantitative X-ray electron microprobe maps, and the
254 AX software (Holland and Powell, 2006) to recalculate the structural formulas of the
255 different mineral phases observed. Optical photomicrographs of the fault rocks were
256 processed with ImageJ (Rasband, 1997) to calculate the surface area, the orientation and the
257 circularity of each grain, and the detection of microfractures. Grain size was calculated using
258 the equal area diameter. These parameters were used to calculate several fault parameters

259 such as the fractal dimension β , the cut-off sharpness α and the characteristic diameter D_c
260 (μm), by fitting the curve obtained with the equation $P \left(\frac{N}{N_t} \right) = \left(1 + \left(\frac{D}{D_c} \right)^\alpha \right)^{-\frac{\beta}{\alpha}}$ (Otsuki, 1998;
261 Monzawa and Otsuki, 2003), with P the number of grains (N) larger than a certain diameter
262 (D in μm) with N_t the total number of grains. Similarly, the microfracture density (μm^{-1}) was
263 calculated using the circle inventory method (Davis et al., 2011; Rossetti et al., 2010),
264 consisting in dividing the total length of microfractures within a circle by the area of this
265 circle.

266

267 **3.Results**

268 **3.1 Alpine metamorphic peak in the Col du Bonhomme area**

269 The peak temperatures calculated from the RSCM analyses of the Aalenian shales
270 and pre-rift limestone samples are 322 ± 18 °C, 336 ± 25 °C and 335 ± 22 °C (Table1 and table
271 S1, S2, S3). Since these shales are rich in organic carbon and are post-tectonic relative to the
272 rifting in the Mont-Blanc/proximal domain, these values likely correspond to the peak
273 temperatures reached during the Alpine metamorphic overprint. These results, which point
274 to lower greenschist facies conditions, are in good agreement with illite crystallinity values
275 measured in the area (1.35; Eltchaninoff and Triboulet, 1980), and with the compilation of
276 RSCM data by Rosenberg et al. (2021), giving a temperature below 350°C for the southern
277 part of the Mont-Blanc area. Petrographic observations on the Grès Singuliers highlight the
278 impact of such metamorphism on the mineralogy of the Jurassic deposits in the internal
279 Mont Blanc Massif where alpine metamorphism resulted in a static thermal overprint
280 marked by overgrowth of muscovite over illite (Fig. S4).

281

282 **3.2 Macroscopic description of the basement / sediment contact**

283 In the Col du Bonhomme area, the contact between the Mont-Blanc basement and
284 the overlying Triassic and Liassic sediments crop out at several places as shown in Figure 2.

285

286 **3.2.1. Basement and tectonized top basement**

287 The basement is composed of a cordierite bearing migmatitic gneiss with a sub-
288 vertical NNE-SSW striking foliation (Fig. 3G). It is intruded by meter-thick magmatic bodies
289 corresponding to aplite.

290 Towards the top of the basement, the gneiss is progressively deformed over about 40
291 meters, showing a gradational transition from the well-preserved gneiss to a cataclasite and
292 a gouge at the basement/cover contact (Fig. 3). Locally, sub-horizontal foliation is visible,
293 possibly corresponding to Variscan mylonites cross-cutting the sub-vertical foliation. They
294 are parallel to the top basement contact (Fig.3 A, F and G) and show a brittle overprint
295 forming tectonic breccia. In more detail, the damaged gneiss is characterized by a network of
296 anastomosed fault zones of cataclastic gneiss with lenses/clasts of preserved gneiss (Fig. 3A),
297 that we define as the damage zone. These anastomosing fault zones form a progressively
298 denser network towards the top, clearly showing a strain gradient toward the top basement.
299 The leucosomes of the migmatitic gneiss are good markers for the strain at the macroscale.
300 Far from the contact, the initial texture of the leucosomes is preserved. Upsection, they are
301 affected by important, centimetre scale fractures without displacement, indicating an
302 increase of the brittle overprint toward the basement/sediment contact (Fig.3E). Closer to
303 the contact (< 10 m Fig.3A), the fracture network becomes denser, the fragments of
304 leucosomes get smaller and gradually develop into cataclasites. The brittle overprint of the
305 former host rock includes the disaggregation of the fabric and the alteration of the feldspars,
306 leading to an increase of the phyllosilicate content upsection. Kinematic markers of fault-
307 related structures are sparse and commonly difficult to interpret. They mainly consist of few
308 lineations giving a N-S direction. In phyllosilicate-rich zones, the kinematics is clearer as the
309 foliation is more marked. *S/C'* structures are observed, where the *C'* plane has an orientation
310 200/36 with a top to the SSW sense of shear.

311 The core of the fault (i.e., the core zone; < 1 meter-thick, Fig. 3A and C) corresponds to a
312 matrix-supported fault rock with a high content of phyllosilicates, here referred to as
313 indurated gouge. It appears as a dark, fine grained discontinuous layer with a well-developed
314 foliation. It forms a sharp contact with the underlying cataclasites forming the damage zone
315 (Fig. 3A). Centimetric injections of material from the gouge are visible in the cataclasite (Fig.
316 3A). Veins parallel to the foliation are also observed (Fig.3C). The gouge has the same
317 appearance at the macro and micro-scale with the alternation of matrix-rich and clast-rich
318 bands, showing N-S lineated clasts (Fig. 3F).

319

320 **3.2.2. Pre-rift, tectonized sedimentary deposits**

321 Pre-rift sediments locally lie above the tectonized basement. Four types can be
322 distinguished, whose spatial distribution is variable along the contact. The first one, which is
323 common, corresponds to discontinuous 20 cm-thick lenses of sandstones composed of
324 micrometric angular lithic fragments and carbonate clasts within a calcite cement (Fig. 3C).
325 The second type consists to a limestone containing millimetric to centimetric, euhedral
326 quartz crystals with 100µm-large rectangular anhydrite inclusions (Fig. S5). The third type
327 corresponds to upper Triassic brownish dolostone (Arandellys Fm.), whose thickness ranges
328 from 0 to approximately 10 meters, and which covers the sandstone lenses and limestones
329 with quartz nodules (Fig. 2,3A and 3C). Although the dolostone origin is sedimentary, it is
330 now significantly deformed up to forming both, metric asymmetric boudins and monomictic
331 breccias. The dolostone layers form an angle of about 10-15° with the tectonic contact
332 separating the basement from the pre-rift sediments suggesting an angular unconformity. In
333 addition, the dolostone boudins can lie directly above the tectonized basement and are
334 associated with tectonic breccias. These breccias are either polymict containing angular to
335 sub-angular pluri-cm to pluri-dm clasts of both dolostones and basement or monomict (Fig.
336 S6 A, B) including dolostone and minor quartz-bearing limestone clasts characterized by a
337 jigsaw geometry (Fig. S6 C, D). Clasts of dolostone embedded in the gouge are frequently
338 observed (Fig. S6 E and F). Finally, the fourth type of pre-rift sediments includes Lower
339 Jurassic (Hettangian) calcschists and limestones preserved in the Rocher du Bonhomme.

340

341 **3.2.3. Syn-rift sediments sealing the contact**

342 At some places, the gouge and cataclasites within the top basement are crosscut by
343 subvertical cracks, which are filled with a sedimentary breccia containing cataclastic
344 material, clasts of gouges and quartz veins floating in a brownish matrix (Fig. 4). The cracks
345 are tens of cm-large and deep and can be best interpreted as neptunian dykes (Fig. 4A). The
346 brown sedimentary infill corresponds to a calcite-rich indurated mud containing quartz
347 grains and bioclasts, mostly crinoids (Fig. 4C, D). The same matrix is also locally observed
348 above the basement gouge and is found within the cracks (Fig. 4B). This observation
349 suggests that the meter-thick brown breccia made mainly of basement clasts, forms a
350 stratigraphic contact with the tectonized basement.

351 The pre-rift dolostones and limestones are truncated by an erosional surface at its top,
352 and are unconformably overlain, along a stratigraphic contact, by Liassic syn-rift sandstones

353 and conglomerates of the Grès Singuliers Fm (Fig. 2,3 and S5 F). These sandstones, 10 to 30
354 meters thick, can also seal directly the tectonized basement at places where the pre-rift
355 deposits are absent. The lower part of the Grès Singuliers corresponds to: (i) monomictic
356 conglomerates (pluri-dm angular carbonate pebbles derived from the pre-rift limestones and
357 dolostones), with a N-S paleocurrent direction as deduced from pebble imbrications
358 different from the NW-SE paleocurrent measurements from the upper part of the
359 formation (Ribes et al., 2020), or (ii) polymictic conglomerates (sub-angular to rounded
360 pebbles of carbonates and basement). Pebbles of cataclasites are also observed
361 ($45^{\circ}43'46.41''\text{N}/6^{\circ}43'47.96''\text{E}$), indicating reworking of the fault rocks from the underlying
362 exhumed basement (Fig.3) into the sediments. The upper part of the Grès Singuliers
363 corresponds to shoreface and foreshore sandstones with strong tidal influence (Ribes et al.,
364 2020). Well-preserved sedimentary figures such as cross beds, small tidal laminations and
365 bioturbations are common.

366 At the top of the Formation, NW-SE stretching lineations related to Alpine deformation
367 are commonly associated with millimetre-thick proto-mylonite shear zones and a pervasive
368 cleavage. Above the sandstones, Aalenian shales belonging to the Dugny Fm. show
369 important deformation with C/S structures giving a top to the NW sense of shear (Fig. 3F).
370 These structures correspond to the decollement level of the Helvetic nappes above the
371 Mont-Blanc massif (Eltchaninoff-Lancelot et al., 1982; Egli and Mancktelow, 2013). It is
372 important to note that the base of the Grès Singuliers is little deformed and that there is a
373 clear strain gradient upsection towards the main Alpine thrust.

374

375 **3.2.4. High-angle faults**

376 Both the basement and the pre-rift deposits are offset by two sets of high-angle normal
377 faults, one striking NE-SW and the second NW-SE. The NE-SW striking normal faults show
378 small offsets (few meters) of the top basement. The NW-SE faults offset the basement and
379 the pre-rift but also the Grès Singuliers and the Aalenian shales. In the Col des Fours area, a
380 NW-SE fault (30/45) offsets the top basement, including the gouges and cataclasites, by
381 about 35 m (Fig. 5). A sedimentary wedge (15°) formed by the Grès Singuliers is visible in the
382 hanging wall of the fault (Fig. 5). Thickening of the sandstones toward the fault plane is
383 compatible with the Grès Singuliers being syn-tectonic to this fault and suggesting a fault
384 activity being Pliensbachian-Toarcian. A roll-over anticline (fold axis 100/14) accommodates

385 the deformation of the pre-rift cover. Slickensides on planes parallel to the fault strike (pitch
386 of 18°) and on secondary planes oblique to the fault (strike: N150) show an important strike-
387 slip component, possibly related to an Alpine reactivation of the fault (coherent with the
388 WSW strike slip direction measured in the area by Egli and Mancktelow, 2013, in the Helvetic
389 cover).

390

391 **3.2.5. Synthesis of macroscopic observations**

392 A strain gradient is observed in the basement toward its top. Variscan foliation is
393 overprinted by brittle deformation, leading to cataclasite and gouge formation. This
394 progressive localisation of the deformation is marked by an increase in the content of
395 phyllosilicates which make most of the gouges. This tectonized basement is overlaid by pre-
396 rift and syn-rift sediments, but the nature of the relationship between basement and its
397 sedimentary cover differs between both types. The pre-rift/basement contact is tectonic
398 while the syn-rift/basement contact is stratigraphic. Therefore, the syn-rift deposits seal an
399 exhumed tectonized top basement, as well as its overlying pre-rift sediments. The syn-rift
400 Grès Singuliers, are affected by a strain gradient as can be seen at its top, which increases
401 toward the overlying Aalenian shales. This alpine deformation is the most intense in the
402 Aalenian shales, which acted as an important decollement during emplacement of the
403 Helvetic nappes (Eltchaninoff-Lancelot et al., 1982; Egli and Mancktelow, 2013).

404

405 **3.3 Petrological and microstructural observations**

406 **3.3.1 Host rock: Variscan gneiss**

407 At distance from the tectonic contact, the subvertical foliation of the migmatitic
408 gneiss basement (quartz (43%), biotite (18%), and feldspar (37%), Fig.6) is highlighted by by
409 the alternation of biotite and quartz-feldspar bands. Accessory minerals are mostly
410 fluorapatite found with biotite. Leucosomes related to melt infiltration are parallel to the
411 gneiss foliation. They are composed of coarse quartz and feldspars. Micro textural analysis
412 shows that quartz grains were affected by grain boundary migration (GBM) and sub-grain
413 rotation with the myrmekitization of some plagioclase feldspars. The migmatites are
414 intruded by aplite veins characterised by a high amount of quartz. The latter are typical of
415 high temperature rock fabrics of migmatitic gneisses.

416 The high temperature (HT) gneiss fabric is also overprinted by microfractures
417 affecting the feldspars, locally forming micro veins showing bulging recrystallisation where
418 the fractures propagate in the quartz. These fractures are mostly transgranular and crosscut
419 the previous HT foliation. They are associated with alteration of feldspars with formation of
420 white micas, albite and microporosity (Fig.7A). Small kinks and crenulation with rutile-rich
421 micro fractures highlight the deformation in the phyllosilicate-rich zones (Fig.7B). This
422 deformation is associated with hydration of the host rock leading to alteration of feldspar
423 and biotite and the formation of white micas. Primary biotite is strongly altered with
424 formation of chlorite and rutile. Alteration is stronger on the border of the crystals and along
425 their cleavage, where rutile is observed, whereas the core of the crystals is better preserved.
426 Chlorite corresponds to an intermediate composition between chlinochore, Fe-sudoite and
427 amnesite (Table 2). The paragenetic association of the hydration of the host rock is $K\text{-Fsp} +$
428 $Ab + Bi \leftrightarrow Chl + Mu + Rt \pm Qtz$ with a total amount of phyllosilicate of around 10%.

429

430 **3.3.2 Damaged zone**

431 In the damage zone, the gneiss protolith is highly altered and affected by brittle
432 deformation. It is composed mainly of quartz, plagioclase and K-feldspars, white micas,
433 sulphides (pyrite and arsenopyrite) and iron oxide minerals. Minerals inherited from the
434 gneiss protolith correspond to quartz, remnants of feldspars and small zircons. Chlorite,
435 fluorapatite and biotite, which were present further away from the tectonic contact, are
436 absent. This fabric is overprinted by fractures filled mostly with iron oxide.

437 In the outer damage zone (>20 m), feldspar constitutes 30% of the rock (10% for
438 plagioclase and 20% for K-feldspar, Fig.6). Feldspars show various amounts of alteration
439 (hydration). Poorly altered K-feldspars show the replacement of orthoclase by albite and
440 small white micas, and formation of secondary porosity (Fig.7A). Highly altered feldspars are
441 mostly replaced by white micas (Fig.7D). Most of the plagioclase has a high proportion of
442 albite. Patches of quartz + micas are common mostly along crystal twins. Minor bulging
443 recrystallisation is related to propagation of transgranular fractures in quartz and undulose
444 extinction of quartz crystals is observed (Fig.7C, D).

445 Close to the top basement / sediment contact (inner damaged zone: < 20 m), the
446 amount of feldspar (Fig.6) decreases (6% of orthoclase and no more albite), correlated with
447 an increase of white micas (20%). Highly altered feldspars show extensive, up to complete

448 white mica replacement (Fig. 6D). The rock is clast-supported, dominated by quartz clasts.
449 Most of the clasts are derived from the gneiss, and the phyllosilicates are a by-product of
450 feldspar alteration leading to the formation of a white mica matrix (M1). White micas are
451 zoned, with an increase of Si and a decrease of Al toward their rims (Table 2). This results a
452 composition departing from the ideal composition of muscovite and being closer to that of
453 the illite-phengite group. Most of the white micas have a grain size below 50 μm ,
454 corresponding to sericite (Eberl et al., 1987), but locally larger micas are observed (> 100
455 μm). Both quartz crystals inherited from the gneiss (where HT dynamic recrystallisation is
456 visible) and from the syn-kinematic veins (coarse grain blocky texture with no dynamic
457 recrystallisation of quartz and locally elongated to fibrous texture) commonly have a lobate
458 crystal border (Fig.7C) similar to a corrosive wear pattern and indicating quartz dissolution.

459 The inherited gneissosity is completely overprinted by brittle deformation (Fig.7C, D,
460 E). In the clast-rich areas, the clasts are rather angular, while in the areas with a higher
461 matrix-to-clast ratios, quartz grains are more rounded (Fig.7E). Most fractures and veins are
462 intergranular, but clusters of intragranular fractures are locally observed. These fractures are
463 branched with each other with an acute angle up to 45° (Fig.7E). Several events of quartz
464 fracturing are also highlighted by different generations of fluid inclusion trails in single quartz
465 grains, which have a fracture density of 28 μm^{-1} . Therefore, deformation is achieved by
466 fracturing associated with significant hydration and mineral neoformation (white micas and
467 quartz precipitation in the veins). These types of “chaotic fault rocks” maintaining cohesion
468 during deformation are typical of cataclasites (Braathen et al., 2004). The grain size shows
469 fractal distribution with a fractal dimension between 1.38 and 1.5.

470 The fabrics described previously are later overprinted by larger transgranular veins
471 (Fig.7E). These veins are filled either with quartz or with iron sulphide or oxide. The quartz
472 veins (> 100 μm -thick) are themselves affected by fracturing and progressively reworked in
473 the cataclastic matrix corresponding to continuous vein formation throughout deformation.
474 The iron sulphide to oxide veins, which crosscut the quartz veins, are thinner (> 10 μm -thick).
475 Far from the contact, only sulphides are observed (6%), while closer to the contact only 1%
476 of pyrite is present, for 2% of rutile and 8% of iron oxide. No specific orientation of the clasts
477 nor the veins is visible.

478 Along the fault (core) zone, calcite-rich fractures are locally visible (Fig.7F). Calcite fills
479 fractures in the cataclasite with a jigsaw monogenic fabric. Cement is brown-coloured, with

480 locally preserved banded texture (2-3 mm). The cement is iron-rich, with a zoned iron
481 composition. The initial texture is composed of large dolomite rhombs with and small
482 ankerite rhombs and formation of calcite filling the space between the former minerals. It is
483 affected by static recrystallization leading to formation of equigranular twinned Mg- calcite.
484 Along the contact between the calcite-cemented fractures and the cataclasite, part of the
485 white micas and feldspars are also replaced by calcite forming patches. The jigsaw fabric
486 with important dilation points toward hydraulic fracturing resulting in breccias. Textural
487 relationships suggest that breccia formation corresponds to one of the latest deformation
488 events along the fault plane.

489

490 **3.3.3 Core zone**

491 The core zone is made of gouges, which are dark, fine-grained matrix-supported
492 rocks, which rework fragments of cataclasites (Fig.8A and B,). The rock shows the same
493 appearance microscopically as macroscopically, with the alternation of clast-rich and matrix-
494 rich layers. The high content in phyllosilicate points toward an indurated black gouge
495 following the definition of Braathen et al. (2004). In the clast-rich layers (at the microscale),
496 the quartz grains range from 1 mm to less than 10 μm in size. The grain size distribution
497 follows a power law with a fractal dimension β between 1.58 and 1.7 (Fig.9C). The shape of
498 the clasts is variable, ranging from angular to well-rounded (up to 0.9 circularity) with a
499 mean circularity of 0.59 (Fig.9F). Deformation of the clasts lead to the formation of σ -clasts
500 (Fig.8A), associated with small folded cataclasite clasts (Fig.8B). Preferential orientation of
501 the clasts (Fig.9E, e.g. Manatschal, 1999) highlights the foliation. The asymmetric clasts are
502 good kinematic markers. Asymmetric clasts are visible in the direction parallel to the N-S
503 stretching lineation indicating a top to the south sense of shear. The gouge matrix (M2)
504 corresponds mostly to illite-rich material less than 5 μm in size, with lower amounts of
505 quartz and pyrite (Fig.8C). Neoformed illite growing perpendicularly from the pyrite walls in
506 pressure shadows are also present. Small zircons (less than 50 μm) inherited from the gneiss
507 are affected by fracturing and minor dissolution at the micro-scale. Oxide-rich bands
508 represent 5% of the whole samples (2% of iron oxide and 2.7% of rutile) (Fig.8E). Most of the
509 pyrite was dissolved, leaving holes in the rock surrounded by quartz pressure shadows. Post-
510 kinematic mineralogical transformations are also observed. Only a small amount of pyrite is
511 preserved or weathered to jarosite. Coarser white micas (up to 100 μm long, $X_{\text{M}} > 0.7$) are

512 visible in the gouge and the cataclasite clasts. They appear post-dating the gouge growing
513 sometimes over the gouge foliation (Fig 8.D). Finally, micrometric hematite-filled fractures
514 can locally be observed in the gouge, where they crosscut and thus postdate the foliation.

515 Dilation and vein formation also occurred in the core zone. Small (2 mm-thick),
516 fractured quartz veins are visible in the matrix. Some veins crosscut the foliation and are
517 affected by fracturing with millimetric offsets, while other veins are parallel to the foliation.
518 Pressure shadows filled with fibrous quartz can have developed around quartz vein
519 fragments. Besides, thin (< 100 μm -thick) iron oxide rich veins parallel to the foliation plane
520 are present. A last vein generation corresponds to centimetre-thick veins visible at the
521 macroscale, parallel to the gouge foliation (Fig.8E, Fig.9A and B). They are not observed in
522 the adjacent cataclasites. The veins are filled by quartz and Ba-rich feldspar (hyalophane)
523 (Fig.8E and F). The Ba content of hyalophane is around 11 to 15% wt, corresponding to a
524 composition of 23% Celsian, 4% Albite and 71.9% orthoclase. The hyalophane crystals, which
525 are euhedral, show no compositional zoning. They are partially replaced by quartz mainly
526 along their cleavage (Fig. S7). On the border of the veins, hyalophane crystals are highly
527 fractured, with reworking of some fragments in the gouge. An important amount of
528 transgranular fractures affecting quartz and hyalophanes are also visible (fracture density of
529 $18 \mu\text{m}^{-1}$) with some corresponding to healed fractures with fluid inclusion trails.

530 More information on the processes occurring during the formation of gouge M2 from
531 the cataclasite is obtained with quantitative compositional mapping by electron probe
532 micro-analyser (EPMA), highlighting the behaviour of major elements and the textural
533 evolution during deformation (Fig.10). Phase maps (Fig.10A) show quartz and feldspars
534 grains in a white mica matrix. For the white micas, a reduction of the grain size is visible for
535 white micas larger than 10 μm in the cataclasite and smaller than 10 μm in the gouge matrix
536 M2 (Fig.10B). The progressive shearing of the cataclasite matrix M1 has an impact not only
537 on the grain size but also on the composition of the micas (Fig.10B and C). Except for the K_2O
538 and SiO_2 contents, which remain stable (10 wt%), and the Al_2O_3 that decreases, other major
539 element contents increase between the cataclasite and the gouge matrix (Fig.10F). The
540 increase is particularly important for TiO_2 as visible in the chemical map (Fig.10C), the
541 chemical profile (Fig.10D) highlighting the absence of TiO_2 in the cataclasite and presence in
542 the gouge matrix. Recalculation of structural formulas show a slight decrease of Al VI and Al
543 IV between the cataclasite matrix and the gouge matrix (1.75 ± 0.07 apfu (atom per

544 formula unit) to 1.67 +/- 0.04 apfu and 0.58 +/- 0.06 apfu to 0.55 +/- 0.040 apfu for Al VI and
545 Al IV, respectively).. Detailed compositional analysis of the phyllosilicates from the gouge
546 matrix (M2 (Fig.10E; Table 2) show departure from the Mu-Ce line (phengite). This indicates
547 that Si-Al substitution is not achieved through a tschermak substitution ($Al^{IV}Al^{VI} \leftrightarrow$
548 $Si^{IV}(Mg,Fe^{2+})^{VI}$) but through a pyrophyllite substitution ($Al^{IV}M^I \leftrightarrow Si^{IV}\square^I$) and illite
549 substitution $Al^{VI}\square^I \leftrightarrow Mg^{VI}M^I$, leading to charge deficit and formation of illite (Meiller (2013))
550 with interfoliar charge between 0.8 and 0.9(Meiller (2013)).

551

552 **4. Discussion**

553 **4.1. Conditions and mechanisms of deformation**

554 **4.1.1. Progressive deformation in retrograde conditions**

555 Detailed petrographical and microstructural observations suggest that the studied rocks
556 forming the top basement of the internal Mont Blanc Massif can be best interpreted as
557 belonging to a fault zone. The fault rocks record the overprint of an initially high-
558 temperature Variscan fabric in the gneissic basement, by a cataclastic deformation at much
559 lower temperatures. In the gneiss, the observation of sub-grain rotation and grain boundary
560 migration during dynamic recrystallisation of quartz grains points to temperatures of around
561 450 to 600°C during the formation of a first, primary foliation (Stipp et al., 2002). Such high
562 temperature fabrics would correspond to depths of 15 to 20 km considering an average
563 gradient of 20°C/km (determined using P-T data on skarns (Marshall et al., 1997). These
564 depths are similar to those of emplacement of the lower part Mont-Blanc pluton at around
565 15 km (6 to 8 Kbar north Mont-Blanc massif; Marshall et al., 1997), suggesting that this
566 Variscan fabric was acquired during the Carboniferous, before any significant exhumation. In
567 the damage zone, this fabric is first overprinted by feldspar fracturing, minor quartz bulging
568 recrystallisation and hydration (Fig.11A). The association of brittle behaviour of feldspars
569 and minor plastic deformation of quartz is indicative of a deformation temperature of
570 around 275°C to 300°C (Stipp et al., 2002), corresponding to a depth of about 7 km assuming
571 a geothermal gradient of 40°C/km (derived from heat flow modeling during necking of the
572 crust (Chenin et al., 2020)). This deformation was associated with significant hydration of the
573 basement, as seen by widespread feldspar sericitization and chloritization of biotite
574 (Fig.11A), still under lower greenschist facies metamorphic conditions. Toward the fault core
575 zone, deformation is more intense and accompanied by an almost complete feldspar

576 replacement by muscovite and illite, and the disappearance of both biotite and chlorite.
577 These metamorphic reactions occurred simultaneous to an increasing brittle overprint
578 during which the host rock lost its initial HT fabric to become a clast supported cataclasite,
579 with an increase of the phyllosilicate content toward the core zone. Cataclastic flow as
580 observed here is found at depth ranging between 12 to 5 km (Sibson, 1986), which is in
581 agreement with a decrease of the overburden during deformation and simultaneous
582 exhumation. The progressive alteration of the rock during brittle deformation resulted in the
583 localisation of deformation in phyllosilicate-rich zones, resulting in the matrix supported core
584 zone (black gouge M2). The observation of rounded clasts of cataclasite in this gouge
585 supports the interpretation that the gouge formed within a previously brittle deformed and
586 altered basement. Petrographical and geochemical observations demonstrate that the
587 gouge matrix was formed by progressive shearing of the cataclasite matrix, associated with a
588 change in composition and grain comminution (Fig.11A). The decrease of Si in the white
589 micas in the cataclasite and corresponding increase in the gouge is consistent with fluid-
590 mediated diffusion of Si and Al across the fault zone and progressive enrichment in Si in the
591 fine-grained gouge matrix, along with the neoformation of illite. The absence of crystal
592 plastic recrystallisation of quartz during gouge formation indicates that frictional
593 deformation was the dominant deformation mode. This restricts the temperature to below
594 275°C, which corresponds to the onset of plastic deformation (Stipp et al., 2002). Brittle
595 behaviour is also visible through the high microfracture density of cataclasite grains, as well
596 as through the significant grain comminution in the gouge. Indeed, the low fractal dimension
597 ($\beta = 1.3-1.7$) calculated in the gouge is an illustration of this friction-dominant, scale
598 independent process leading to grain size reduction. It is in the range of both generic and
599 detachment gouges (Boulton et al., 2009), which have been shown to be commonly formed
600 at levels shallower than 5 km depth (Sibson, 1986), where conditions for extensive clay
601 formation are met (low T and low P) . Finally, transition from plastic/frictional deformation
602 to pure frictional deformation and reworking of cataclasites in the gouge points to
603 exhumation along a retrograde path as deformation proceeded. This exhumation of the
604 basement reached the surface where formation of fissures filled with a crinoid mud affected
605 the already tectonized basement and was sealed by the Grès Singuliers.

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4.1.2 Strain softening due to phyllosilicate neoformation along brittle fault plane

608 Deformation at retrograde, fluid assisted alteration was associated with important
609 phyllosilicate replacement and neoformation in the cataclasite and particularly in the gouge.
610 It is thus likely that the phyllosilicates played a key role in the mechanical evolution and
611 behaviour of the fault (e.g. Wintsch et al., 1995; Jefferies et al., 2006; Wibberley, 2005;
612 Imber et al., 2008; Wallis et al., 2015). Indeed, from the proportion of illite to quartz (80% of
613 clay) in the gouge M2, and using the experimental work on gouge friction of Tembe et al.
614 (2010), the gouge can be estimated to have had a friction coefficient (μ) lower than 0.3
615 (Fig.11A), without considering the possible impact of fluid pressure in facilitating slip along
616 the fault plane. Such a low-friction coefficient must be compared with that of gneissic rocks ,
617 which is expected to be >0.6 (Ikari et al., 2011). This behaviour is best explained by the
618 neoformation of phyllosilicates, which led to significant strain-softening at the transition
619 from a clast to matrix/phyllosilicate supported fabric, promoting, and facilitating slip along
620 the fault plane. In clays, this frictional sliding is enhanced by the presence of interlayer water
621 (when present) and a deficit in interlayer cations, leading to a low bonding of the octahedral
622 and tetrahedral layer facilitating slip (Morrow et al., 2000). In addition to the phyllosilicates,
623 the injection of gouge matrix (M2) into the cataclasites, both at macroscopic and
624 microscopic scales, shows that the gouge locally behaved like a fluid. This is best explained
625 by local fluid overpressure developing in the gouge during transient deformation events (e.g.
626 Manatschal, 1999), likely associated with gouge compaction and pore reduction during
627 shearing (e.g. Faulkner et al., 2018). However, several observations suggest that the
628 mechanical behaviour of the fault gouge evolved from low- to high-friction. The last event
629 affecting the gouge was the formation of quartz/ hyalophane veins emplaced parallel to the
630 gouge foliation. Although the age of the veins is not well constrained, the brecciation of
631 some veins and the presence of clasts of hyalophane in the gouge M2 and the cataclasite
632 tends to indicate a syn-kinematic emplacement of these veins. At present-day the gouge is
633 indurated, which is not compatible with a low friction coefficient. Induration may be
634 explained by syn-kinematic quartz precipitation and presence of Si-rich pervasive fluid or
635 recrystallisation of the illite matrix, possibly during Alpine orogenesis, as discussed later
636 (Fig.11A). The fact that only cataclasites but no gouge clasts were found in the syn-rift
637 sediments might point toward the second hypothesis. Nevertheless, adularia similar to the
638 hyalophane variety from the veins are found commonly reworked in the Grès Singuliers (Fig.
639 S7).

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4.1.3 Fault-related basement hydration and fluid circulation

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The mineral assemblages observed in the damaged basement and the fault rocks highlight the widespread presence of fluids during Jurassic deformation, while their chemical composition suggests the migration of some elements along the fault zone linked to fluid channelling. Indeed, several syn-kinematic metamorphic minerals are hydrated, mostly chlorite and white micas. Chlorite formed in replacement of biotite, which is a common reaction at greenschist facies conditions (Parry and Downey, 1982; Parneix et al., 1985). Chloritization was spatially restricted to the lowly deformed basement. The most important evidence of widespread hydration is the partial transformation of the K-feldspars to sericite and albite, followed by their complete replacement by white micas as deformation continued. The reaction was likely triggered by significant fracturing at both the macro and the microscale in the fault zone, increasing the permeability in the basement, facilitating fluid circulation (e.g., McCaig, 1988; Sibson, 2000; Manatschal et al., 2000; Wibberley and Shimamoto, 2003; Rossetti et al., 2010). Such effect is highlighted by important alteration around microfractures in the damaged zone, which indicates that fracturing at the micro scale increased the reactive surface available for fluid rock interaction (Plümper et al., 2017). In addition to phyllosilicates, several minerals precipitated in veins or locally in the fault gouge during deformation, such as iron rich sulphide/oxides, quartz, hyalophane, or rutile. Some minerals containing elements found in large concentrations in the protolith, such as quartz (Si) or iron sulphides/oxides (Fe), could have easily precipitated in a closed system, triggered by dissolution/precipitation mechanisms (e.g., Jefferies et al., 2006; McAleer et al., 2017). However, the presence of rutile and Ti-rich illite in the gouge matrix, or of hyalophane in veins, which are rich in Ti and Ba respectively, requires element transfer in a more open system. The systematic enrichment of the gouge matrix (M2) in Ti points to mobility of this element during gouge formation. Indeed, the gouge matrix was formed by shearing of the cataclasite matrix (M1), which is devoid of Ti. This indicates at least a minimum amount of Ti transfer in an open system at micro-scale, which is unexpected as this element is usually considered as immobile during fluid-rock interaction (Van Baalen, 1993; Rubin et al., 1993). In the same way, the presence of hyalophane mostly in late veins suggests the input of an external Ba-rich fluid. Such fluid mobility is not surprising, since faults can act as a conduit for fluids both in sedimentary basins and at crustal scales (McCaig,

672 1988; Manatschal et al., 2000; Sibson, 2000; Rossetti et al., 2010). However, fault
673 permeability is extremely heterogenous (Wibberley and Shimamoto, 2003). In the damage
674 zone (cataclasite), permeability reaches values of 10^{-14} - 10^{-16}m^2 , controlled mainly by mineral
675 precipitation and dissolution, and by fracture development (Wibberley and Shimamoto,
676 2003). Dynamic permeability (crack and seal behaviour) related to local fluid overpressure
677 build up along the fault during an interseismic cycle with fluid release during coseismic slip
678 might have contributed to the overall permeability (Sibson, 1990; Sibson, 1992; Poulet et al.,
679 2021). In the core zone (gouges), permeability is generally lower (10^{-17} - 10^{-20}m^2) and
680 anisotropy associated with the planar fabric of the phyllosilicates higher (Wibberley and
681 Shimamoto, 2003; Kopf, 2001). In the studied fault, most of the veins observed are parallel
682 to the foliation, where one could expect the highest permeability. It is thus likely that the
683 gouge fabric had strong control of vein formation and related fluid circulation. Nevertheless,
684 the origin of external fluids, such as a crustal source (dehydration of hydrated rocks) or the
685 infiltration of water from the surface, remains to be determined. Ongoing mass-balance
686 calculations should help to quantify more precisely the conditions of fluid mobility during
687 fault-related deformation in the Col du Bonhomme area.

688

689 **4.2 Is the Col du Bonhomme fault a Jurassic extensional detachment fault?**

690 **4.2.1. Sedimentological evidence for superficial exhumation along a detachment**

691 An extensional detachment fault corresponds to an fault dipping gently ($< 30^\circ$) and
692 which can appear sub-horizontal at the surface (Boulton et al., 2009). Thus, syn-rift
693 sediments can be deposited conformably above an exhumed fault surface, which is not
694 possible for a thrust fault that results in burial. The extensional fault can either initiate at a
695 low-angle, or alternatively initiate at a high-angle and rotate subsequently during
696 exhumation at a low-angle near or at the surface. Isostatic uplift and low elastic strength in
697 the footwall favour such a rotation (Axen and Hartley, 1997) and can lead to an important,
698 kilometre-scale offset (Forsyth, 1992). Such faults are also referred to as rolling hinge faults
699 (Wernicke and Axen, 1988; Lavier and Buck, 2002). They are usually associated with low
700 frictional material along the fault plane. If the fault is exhumed under aerial or marine
701 conditions, it can be submitted to erosion. In such a case, the fault surface can be eroded
702 and becomes difficult to observe. The reworking of the fault rocks in the syn-exhumation
703 sediments is a diagnostic feature of exhumed faults as observed for instance in the Basin and

704 Range (Yin and Dunn, 1992). Epin et al. (2017) defined diagnostic criteria to identify pre-
705 Alpine, rift related detachment faults. A key observation is the presence of fault rocks at the
706 top basement, overlaid by either pre-rift sediments along a tectonic contact, or by syn- to
707 post-tectonic sediments in which case the contact is stratigraphic. Indeed, deposition of
708 sediments above a fault plane is geometrically possible only above an extensional
709 detachment fault, where the footwall is pulled out underneath its hanging wall. Both
710 features are clearly observed in the Col du Bonhomme area.

711 First, remnants of pre-rift Triassic deposits are overlying cataclasites and gouges
712 capping the basement along a tectonic contact. Similar examples have been described in the
713 Pyrenees (Lagabrielle et al., 2010; Jammes et al., 2010) but also in the Alps, where pre-rift
714 remnants are found over the exhumed crust in the necking, the distal and the exhumed
715 mantle domains of the Adriatic margin (Mohn et al., 2010; Epin, 2017). These pre-rift blocks
716 can range from tens of meters to kilometres. Many authors discussed the role of evaporites
717 as potential decoupling levels controlling the preservation of supra-salt pre-rift cover
718 sequence as allochthonous hanging-wall blocks over exhumed footwall (e.g., Jammes et al.,
719 2010; Epin and Manatschal, 2018). In the case of the Col du Bonhomme, the anhydrite-
720 bearing quartz locally found in recrystallized carbonates at the base of the pre-rift deposits
721 suggests the former presence of evaporite levels below the dolomites, suggesting that they
722 may have played a role during exhumation.

723 Second, the syn-rift deposits are sealing the exhumed fault surface, creating a
724 stratigraphic contact. Field observations suggest that the oldest sediments deposited onto
725 the detachment correspond to the crinoid-rich, brown-coloured breccias. The latter fills
726 neptunian dykes that crosscut the basement cataclasites and gouges and is therefore
727 necessarily younger than the fault rocks. A more recent age is also supported by reworking
728 of the cataclasites and gouges in the breccias. A similar breccia, marked by the presence of
729 lamellibranches, echinoderms and *Pentacrines* crinoids was also described by Landry (1976),
730 2 km southeast from the Col du Bonhomme (Crête des Gittes; 45°42'51.6"/ 6°42'37.39").
731 The first occurrence of crinoids in the Helvetic/Dauphinois domain began in the Hettangian,
732 with massive deposition of crinoid-bearing limestones during the Sinemurian/Pliensbachian
733 in the Grande Rouse, Belledonne, and Morcles Massifs (Roux et al., 1988; Loup, 1992;
734 Barfety, 1985; Gillcrist, 1988; Chevalier, 2002; Epard, 1989). A similar, early syn-rift age for
735 the breccia would be consistent with these studies, but also with the age of the overlying

736 Grès Singuliers which is assumed to be Sinemurian to Pliensbachian. Like the breccias, the
737 Grès Singuliers are in stratigraphic contact with the tectonized basement, but also rework
738 the diagnostic cataclasites. This not only indicates that the fault had to be exhumed at the
739 surface, but also supports its pre-Alpine age.

740 Finally, in the Col du Bonhomme area, an unconformity of 10 to 20° is observed
741 between the Grès Singuliers and the top basement surface. Assuming horizontal deposition,
742 which is likely at least in the top Grès Singuliers, where no sedimentary structures indicative
743 of deposition on a major slope were observed, a shallow, < 20° dip-angle between the
744 sediments and the top basement can be reconstructed. All these observations are
745 compatible with an interpretation in which the Mont Blanc massif is capped by a Jurassic
746 low-angle, extensional detachment fault. The shallow dip at the surface may be due to the
747 rotation of the fault due to isostatic uplift in response to fault activity (removal of the
748 hanging wall; Lavier et al., 2000; Lavier and Buck, 2002). This uplift may have lead ultimately
749 to the formation of the Mont-Blanc Island of Trümpy (1971) and Loup (1992), thus providing
750 a source for the detrital syn-rift Grès Singuliers sandstones found around the Mont-Blanc
751 Massif (Fig.11B; Ribes et al., 2020).

752

753 **4.2.2. Interpretation of kinematic markers**

754 Most of the Alpine deformation related to the emplacement of the Helvetic nappes in
755 the Col du Bonhomme area was accommodated in the Aalenian shales overlying the Grès
756 Singuliers, as already noted by Egli and Mancktelow (2013). This may explain why the pre-
757 Alpine, rifted related cover-basement relationships at the Col du Bonhomme were not
758 significantly affected by Alpine deformation. In the damage and core zones of the studied
759 fault, a limited imprint is also supported by the lack of typical Alpine deformation recorded
760 elsewhere in the Mont-Blanc basement, characterized by mylonitic shear zones. In addition,
761 our kinematic measurements obtained in the pre-rift and in the basement (Fig.3G) are
762 consistent with a normal fault having a top to the south sense of shear, which is hardly
763 compatible with the Alpine directions measured in the area (E-W to NW-SE directions with
764 top to the west or to the north-west sense of shear; Meyer, 2002; Egli and Mancktelow,
765 2013).

766 N-S directed syn-rift movement in the Mont-Blanc detachment differs from the NW-
767 SE direction generally assumed for the extensional direction of the Alpine Tethys rift system

768 (Lemoine and Trümpy, 1987; Fig. 1). However, these local measurements cannot be directly
769 compared to the direction of regional extension without considering the specificities of the
770 Alpine Tethys European margin whose three-dimensional characteristics is poorly known.
771 Such scale discrepancies may reflect kinematics complexities related to rift segmentation
772 and inheritance. Many examples both from fossil and present-day margin, show the complex
773 kinematics occurring along faults network, where locally sense of shear can be oblique to the
774 direction of extension. The necking system along the Norwegian margin shows complex fault
775 geometries at large scale (Osmundsen and Péron-Pinvidic, 2018), with fault orientations
776 spreading over 90°, controlled mostly by inheritance (Muñoz-Barrera et al., 2020). More
777 generally segmentation in extensional systems could be associated with the development of
778 transfer zones where local stretching can be highly oblique to the direction of regional
779 extension as demonstrated in metamorphic core complex (Le Pourhiet et al., 2014) and
780 reproduced in rift segment boundaries (Le Pourhiet et al., 2017). Segmentation of the Alpine
781 Tethys rift system by major NW-SE transfer zones have been postulated by many studies in
782 particular between the Pelvoux and Argentera Massif and possibly between the Belledonne
783 and Mont-Blanc Massif (e.g. Lemoine et al., 1989; Decarlis et al., 2017; Ribes et al., 2019a). It
784 is also worth noting that recent studies in the Pyrenean rift have also emphasised that rift
785 segment boundary is better preserved from subsequent underthrusting during orogeny than
786 rift segment centers (Manatschal et al., 2021). For all these reasons, our N-S stretching
787 measurements in the Mont-Blanc preserved detachment may suggest rift segmentation in
788 the southern western border of the internal Mont-Blanc massif.

789

790 **4.2.3. Mineralogical and geochemical evolution: Jurassic exhumation versus Alpine** 791 **metamorphism**

792 In addition to the arguments presented above, the exhumation of the footwall can be
793 deduced from the retrograde path recorded by with the mineralogical assemblages of the
794 fault rocks, which show a progressive decrease of temperature associated with the
795 exhumation from sub-greenschist conditions to the surface. Most mineralogical assemblages
796 identified are diagnostic of lower greenschist metamorphic conditions, with temperatures
797 lower than ~300°C. These temperatures are very close to those reached during the Alpine
798 metamorphic peak in the Col du Bonhomme (~330°C). These temperatures preclude the
799 formation of biotite but are high enough to allow that of muscovite, raising the question of

800 the preservation of the parageneses related to Jurassic fault exhumation. Indeed, our
801 observations show that such syn-kinematic parageneses correspond to extensive
802 phyllosilicate formation, either muscovite in the cataclasite or illite in the gouge, and minor
803 formation of rutile, quartz/hyalophane veins, iron sulphides and calcite. From these
804 minerals, some could clearly have been altered or recrystallized during Alpine prograde
805 metamorphism. This applies first to the white micas. It is not excluded that the present-day
806 illite matrix of the gouge results from the prograde recrystallisation of a lower temperature
807 clay. However, if illite formed during Jurassic extension, its preservation would not be
808 surprising since illite-type white micas are stable until 350°C for the illite end-member, with
809 complete disappearance of smectite above ca. 300°C (Velde, 1985; Merriman, 1999;
810 Rosenberg, 2002; Meunier et al., 2004). In addition, illite whose presence is widespread in
811 the Grès Singuliers and the Aalenian shales (Eltchaninoff and Triboulet, 1980), is known to
812 form at least until 320°C (Eberl et al., 1987; Battaglia, 2004; Meunier et al., 2004). Then, the
813 age of the white micas (sericite-type) observed in the pressure shadows of quartz fragments
814 and sulphides is hard to assess, and an alpine age is possible. Minor muscovite static
815 recrystallisation was also observed locally as overprinting the gouge matrix, possibly as a
816 consequence of prograde metamorphism.

817 For some phases appearing the latest in the sequence (hyalophane-quartz veins, calcite),
818 doubts remain as to a Jurassic or Alpine origin. In the gouge, the hyalophane-quartz veins are
819 deformed and can be broken to clasts, which favors an early, Jurassic formation. The
820 presence of detrital adularia in the Grès Singuliers clearly points in the same direction.
821 However, the composition of the latter needs to be analysed in more detail to confirm a
822 common origin with the hyalophane crystals of the veins. Finally, the Alpine overprint is
823 likely responsible for the static recrystallisation of calcite in the basement, where the initial
824 texture was almost completely lost and replaced by equigranular calcite.

825 To conclude, the contact between the tectonized top basement and the sediments is
826 clearly pre-Alpine, and the Alpine metamorphic overprint seems mostly static with partial
827 recrystallisation of low temperature minerals such as the gouge clay matrix and calcite.

828

829 **4.3 Large scale implication of the Mont-Blanc detachment fault**

830 The presence of a rift-related detachment fault in the Mont-Blanc Massif might have
831 a variety of implications for its Alpine history. First, this study shows that major rift-related

832 structures can be preserved even in deformed parts of an orogen, in accordance with
833 previous observations in the Alps (Barfety and Gidon, 1984; Lemoine and Trümpy, 1987;
834 Manatschal, 1999; Cardello and Mancktelow, 2014; Epin and Manatschal, 2018; Pantet et al.,
835 2020; Nibourel et al., 2021) and the Pyrenees (Izquierdo-Llavall et al., 2020; Espurt et al.,
836 2019; Labaume and Teixell, 2020; Lescoutre and Manatschal, 2020). Indeed, the large-scale
837 architecture of the Mont-Blanc massif proposed by Ribes et al. (2020) put the Mont-Blanc at
838 the transition between the proximal and the distal domains of the former European margin,
839 indicating a specific crustal architecture.

840 The thinning of the crust is highlighted by the increase of the accommodation space
841 and the presence of exhumed mantle (Beltrando et al., 2012). According to the model of the
842 H-Block (Fig. 11C; Lavier and Manatschal, 2006), thinning during the necking of the crust is
843 accommodated by two conjugated faults located in the upper brittle crust and in the lower
844 crust. The Mont-Blanc detachment fault could correspond to the upper crust detachment
845 where the deformation is mostly (if not entirely) brittle. The lower crust conjugate fault
846 might have been preserved as interpreted from the ECORS-CROP profile (Thouvenot et al.,
847 1996; Mohn et al., 2014).

848 Preservation was probably better in the southern edge of the Mont-Blanc Massif,
849 where the rift-related structures are oblique to the main thrust axis, and therefore likely
850 transported passively. At present, the thrust nappes from the external part of the Mont-
851 Blanc (former proximal domain) present polyharmonic folds (e.g., Morcles nappe; Ramsay,
852 1989; Epard, 1990), whereas in the internal part of the Mont-Blanc Massive, the thrust
853 nappes are characterized by disharmonic folding (e.g., Wildhorn and Diableret nappes;
854 Ramsay, 1989). Fold nappes in the external domains are generated by half-graben inversion
855 as shown by field evidence (e.g., Bourg d'Oisan basin; Chevalier, 2002) or by numerical
856 models (Bauville and Schmalholz, 2015; Kiss et al., 2020). The geometry of the disharmonic
857 thrust nappes from the internal Mont-Blanc may thus be explained by the role of structural
858 inheritance from the Mont-Blanc detachment fault, leading to a different Jurassic basin
859 architecture and position of decollement level.

860 In addition, important hydration of the basement around the detachment level
861 during Jurassic rifting, in part linked with an open fluid circulation, could have changed the
862 rheological behaviour of the crust, by diminishing its strength (Wintsch et al., 1995), as
863 postulated for example in the Axial zone of the Pyrenees by Bellahsen et al. (2019) and

864 Airaghi et al.(2020). Indeed, thin-skinned and thick-skinned fold thrust belts are controlled
865 by thermal inheritance but also by structural (Mohn et al., 2014; Tavani et al., 2021) and
866 chemical inheritance (Lacombe and Bellahsen, 2016). A rheologically weakened basement
867 following fault-related hydration promotes thick-skin style of deformation during shortening,
868 and could accommodate large amounts of deformation through a localised or diffused
869 deformation zone, which is observed in the external crystalline massifs of the Alps (Bellahsen
870 et al., 2014; Bellanger et al., 2015; Nibourel et al., 2021; Rosenberg et al., 2021). Finally,
871 because of its potential role in both collision-related deformation and fluid flow during
872 Alpine metamorphism, our study shows that the extent of Jurassic deformation and
873 associated basement hydration clearly deserves further characterisation.

874

875

876 **Conclusion**

877 This work was focused on the field, petrographic, microstructural, and geochemical
878 characterization of a tectonic surface recently discovered in the Col du Bonhomme area as
879 part of the Mont-Blanc Massif, and previously described as a Jurassic detachment fault. The
880 tectonic surface affects mostly the gneissic basement of the massif, which is heavily
881 deformed into cataclasite above which Sinemurian to Pliensbachian sediments were
882 deposited.

883 In detail, the cataclastic basement is increasingly deformed toward its top.
884 Deformation was exclusively the brittle field and associated with hydration of the basement,
885 visible through the increase of phyllosilicate neoformation, which is likely responsible for the
886 observed strain localisation. Overall, the transition from sub-greenschist brittle deformation
887 to gouge formation with low-T clays points towards a retrograde path during fault activity,
888 leading ultimately to exhumation to the surface, which is also supported by the reworking of
889 clasts of fault-rocks in the overlying syn-rift sandstones. All these observations are fully
890 consistent with an interpretation of the top basement as a Jurassic extensional detachment
891 fault. They also suggest that important hydration of the basement and vein formation could
892 be related to fluid circulation along the fault plane, although additional work is needed to
893 confirm this interpretation.

894 Finally, the description of a pre-Alpine detachment fault in the Mont-Blanc Massif
895 shows that preservation of such rift-related structures is possible in the External Crystalline

896 massifs despite significant Alpine deformation. It also raises the question of the role played
897 by this structure during Alpine shortening. Nevertheless, this detachment also provides a
898 field analogue for the necking domain of rifted margins where both the sedimentary cover
899 and the tectonic structure of the basement are preserved.

900

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908

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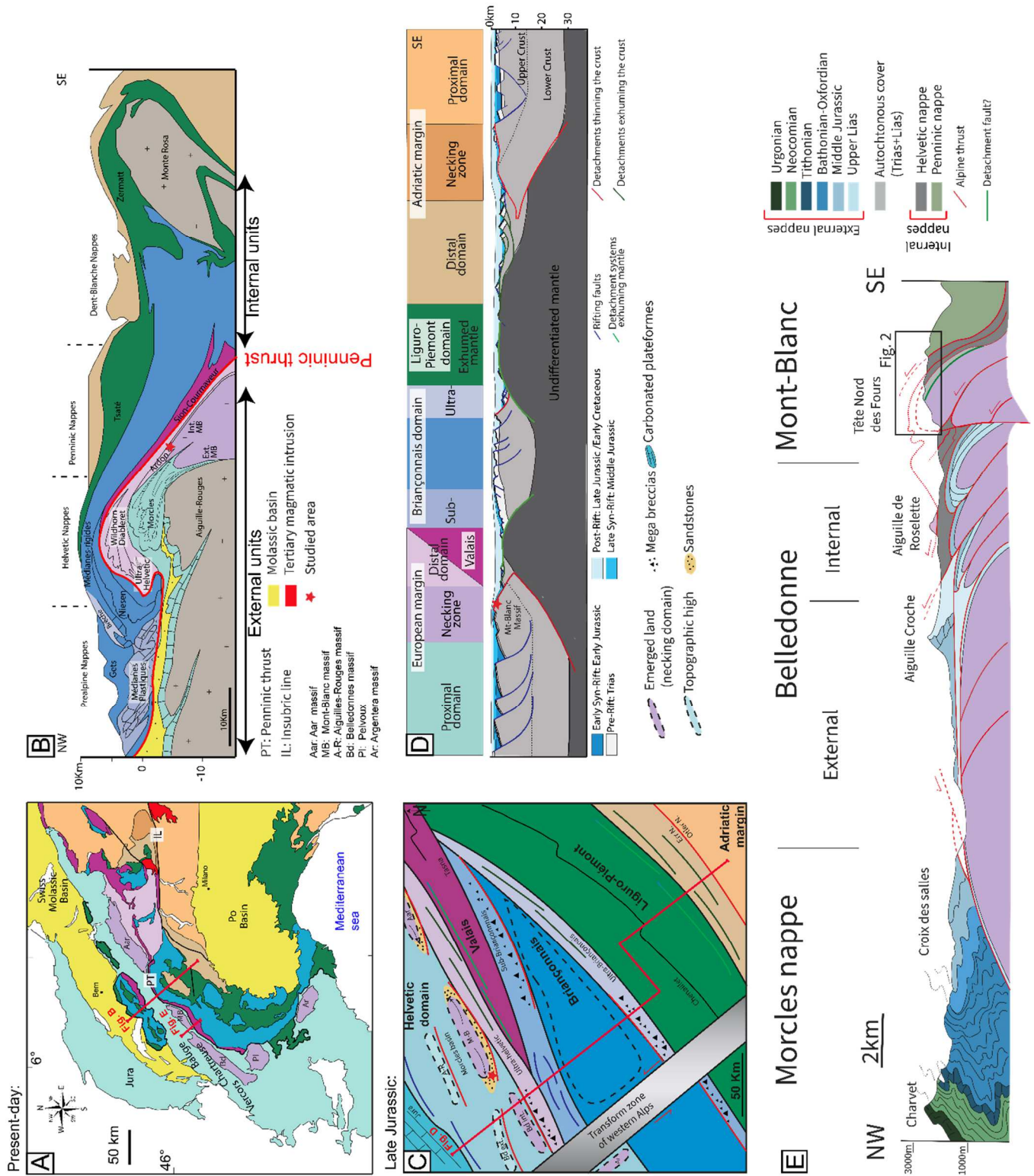


Fig1: A- Tectonic map of the Alps showing the different Alpine tectonic units. Note the presence of the External Crystalline Massifs (ECM) at the transition between the internal and external domains (modified from Mohn et al. (2010); Schmid et al. (2004)). The colour code similar to D (except molassic basins in yellow and Tertiary intrusions in red) highlights the paleogeographic position of the different Alpine units. B- Section across the Western Alps (for location see Fig. 1A; modified from Escher et al. (1993a)). Colour code similar to D. C- Paleogeographic map of the Alpine realm during the late Jurassic (modified from Ribes et al., 2019a). Sandstones (Ribes et al., 2020) and breccias (Ribes et al., 2019b) related to tectonic activity are also represented. Faults in blue correspond to normal faults associated with stretching, the fault in red

corresponds to necking fault thinning the crust, and faults in green correspond to faults exhuming the crust and the mantle. D-Restored section of the Alpine Tethys during the Late Jurassic (Mohn et al., 2010). E- NW-SE directed cross section across the transition zone between the Belledonne and Mont-Blanc Massif (Southern Mont-Blanc massif, modified from Eltchaninoff-Lancelot et al. (1982)). The Penninic and Helvetic nappes thrust above the ECM were later deformed by basement deformation leading to the formation of a large antiformal fold. The area of interest corresponds to the basement anticline pinched between the Helvetic nappes to the south and the Roselette nappe to the north.

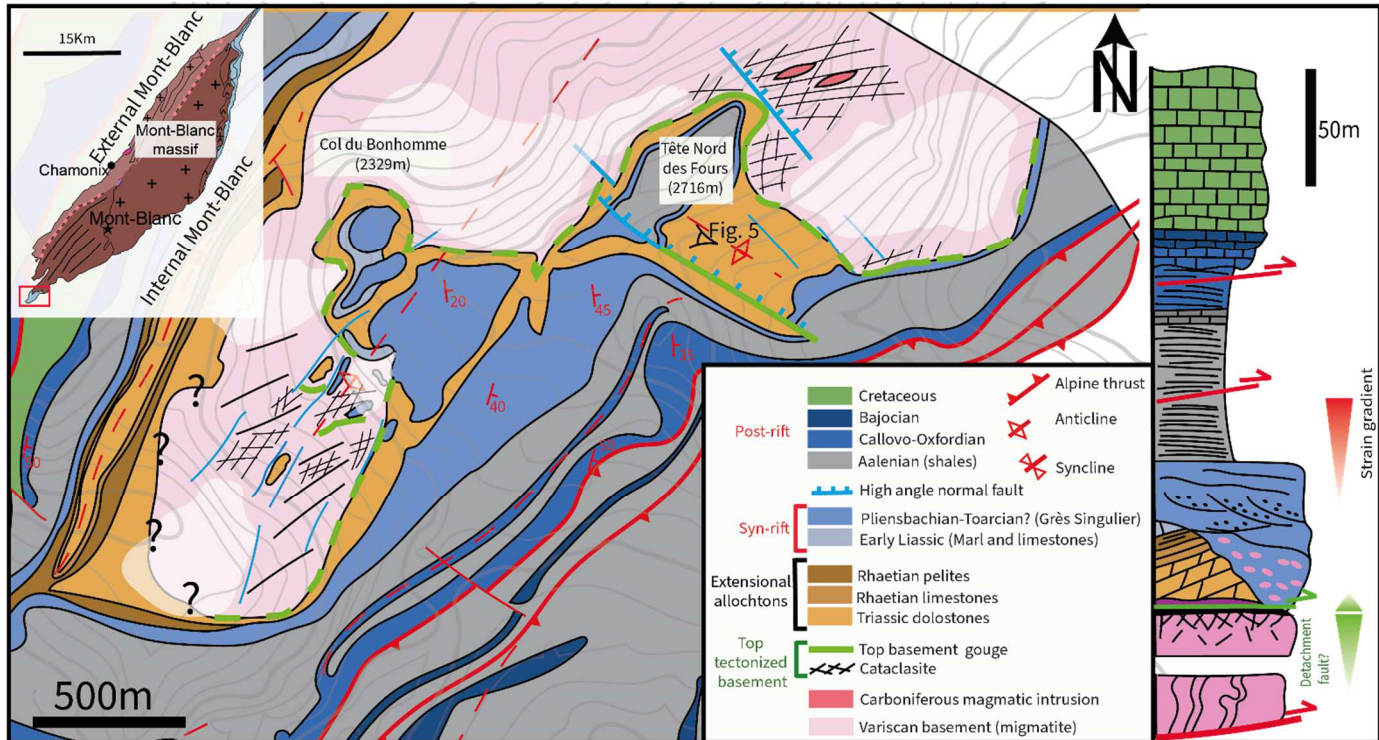


Fig2:-Map of the Col du Bonhomme area (see location on map in the upper left corner, modified from Meyer, (2002)) showing the southern edge of the Mont-Blanc Massif. The tectonic contact described by Ribes et al.(2020) is highlighted by the presence of cataclasite and gouge at the transition between the basement and the sediment. Numerous NE-SW and NW-SE Jurassic faults also offset the top basement tectonic contact. The section on the right shows the simplified stratigraphy from the top basement to the Cretaceous with the location of the main decollement levels in the area in the Aalenian and Oxfordian schists.

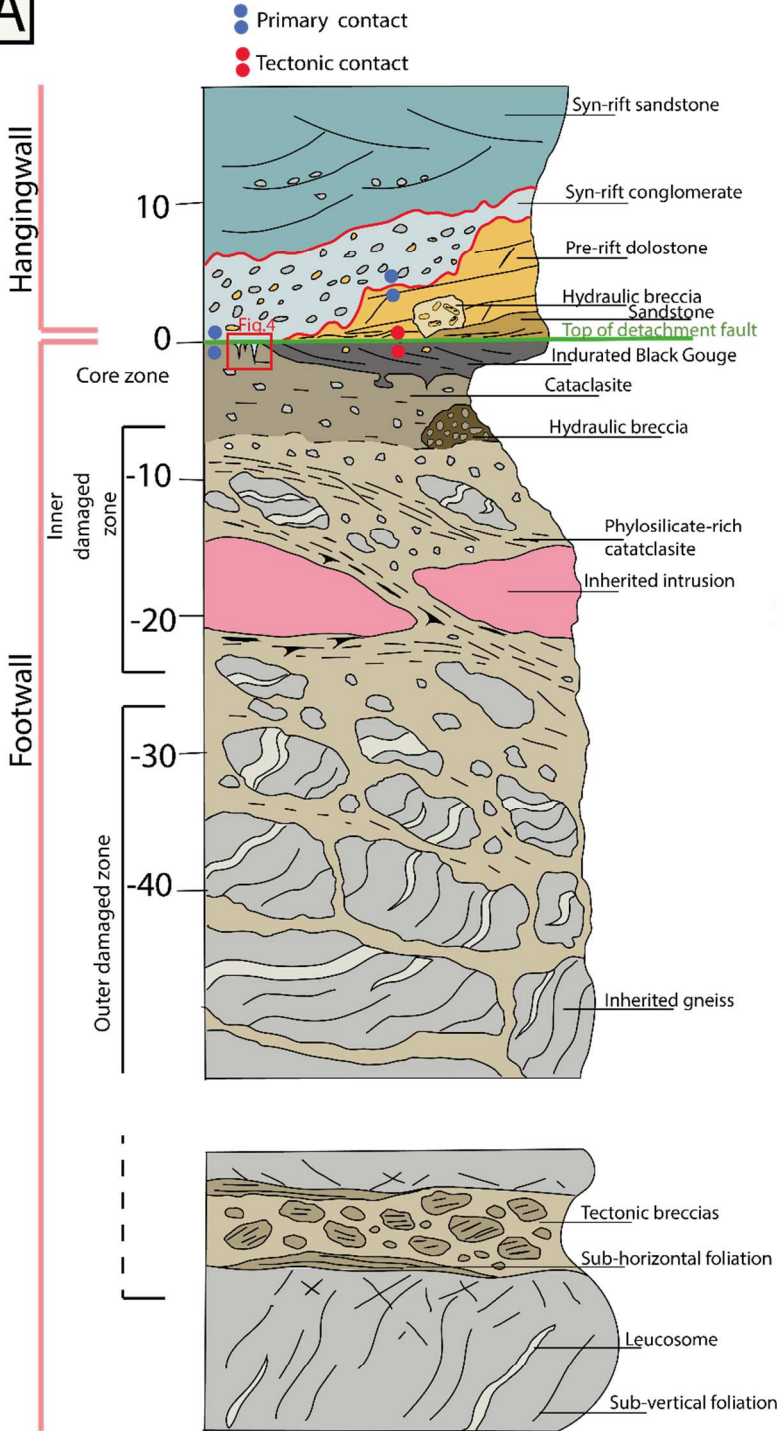
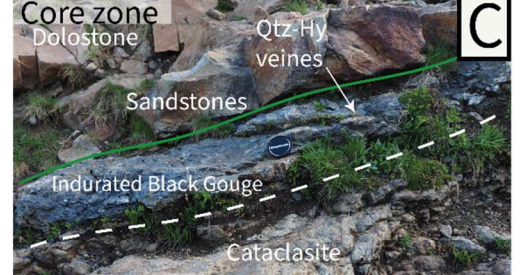
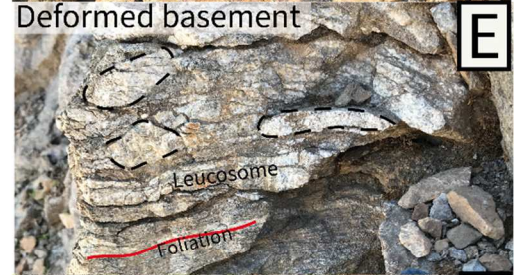
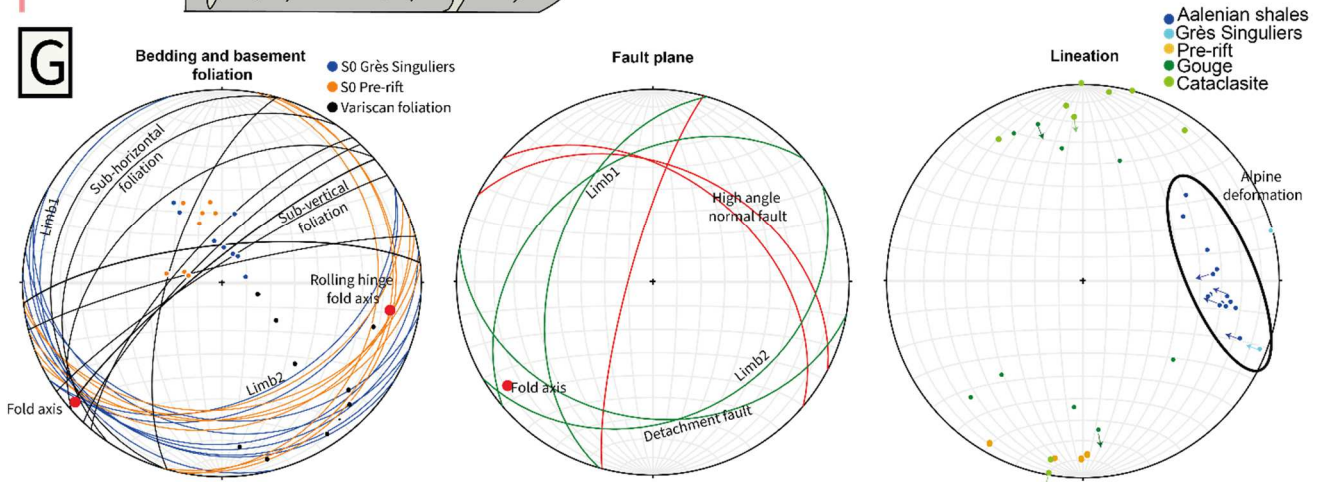
A**B****C****D****E****F****G**

Fig3: A-Schematic section of the top basement/sediment contact showing a strain gradient toward the top of the basement. A local loss of the initial rock texture is visible associated with an increase of alteration and brittle overprint. A sharp contact separates the basement and the sedimentary cover. Pre-rift sediments are separated from the basement by a tectonic contact with deformation visible at the base. The Grès Singuliers are lying above both the pre-rift and the basement in a stratigraphic contact. B-Clast of cataclasite reworked in the Grès Singuliers (45°43'46.41"N/6°43'47.96"E). C- The top basement is composed of cataclasite and locally black indurated gouge (45°43'39.51"N/6°42'40.25"E). D- Phyllosilicate-rich cataclasite giving an NNE-SSSW top to the south sense of shear in the damaged zone (45°46'13.36"N/6°43'46.17"E). E- Cataclased basement with sub-horizontal Variscan foliation is still discernible in the damaged zone (45°44'11.21N/6°43'46.85"E). F- Sub-horizontal mylonites affected by brittle deformation. G-Measures of bedding, fault plane orientations (high angle-normal faults and orientations of the top basement tectonic contact) and stretching lineations from the basement and the cover (projected on Schmidt stereograms, lower hemisphere). Note the NE-SW orientation of the basement fold affecting the basement and the cover (limb 1 and 2).

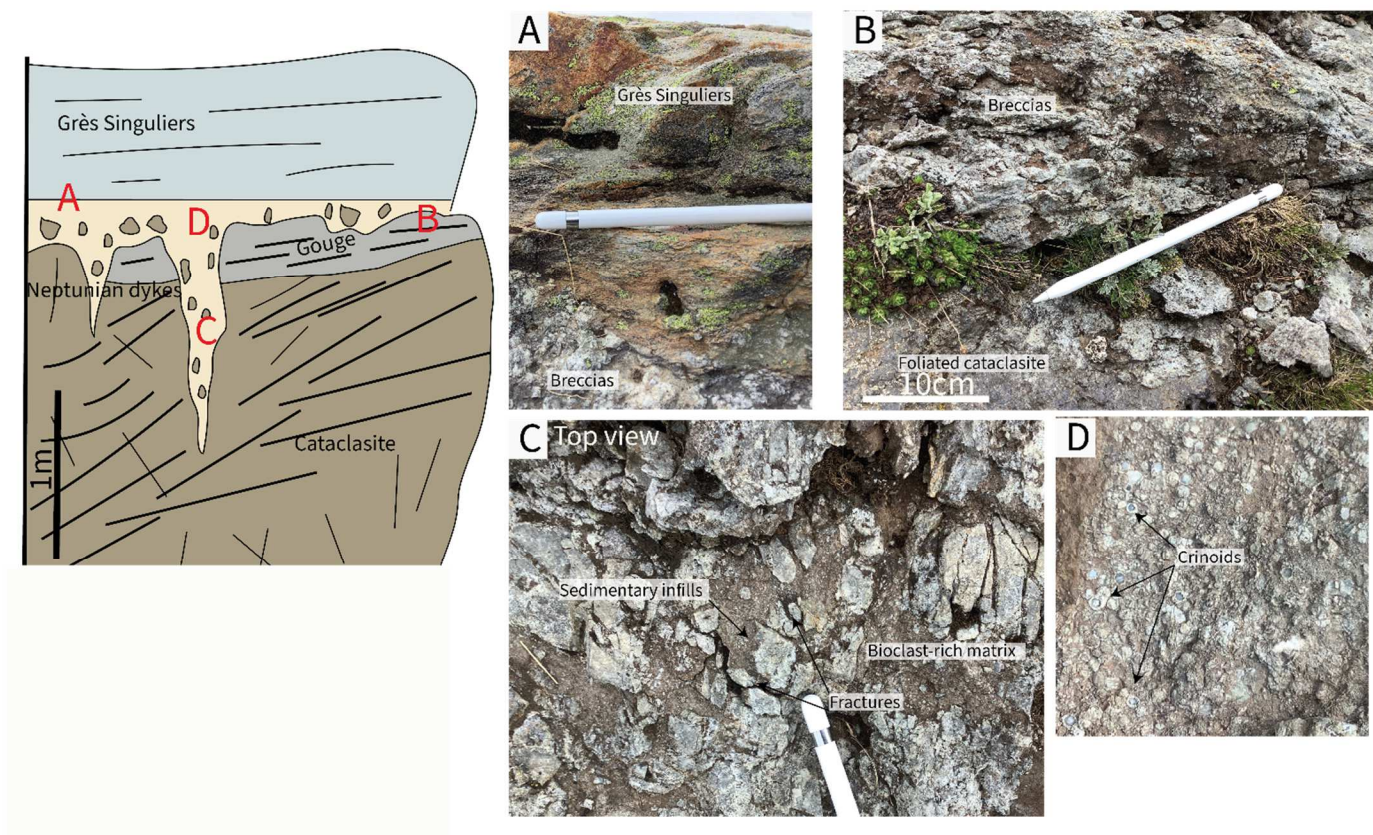


Fig4: Section of the top basement with sediment infill interpreted as neptunian dykes (45°43'39.6"N/6°42'46"E). The position of the pictures is indicated in the section. A- Contact between the top basement and the Grès Singuliers. B-Profile view of foliated cataclasite overlaid by fractured basement. The fractures are filled with a brownish matrix. C- Upper view of the fractured basement with sedimentary infill. The infill is composed of quartz grains and bioclasts fragments in a brownish matrix. D-View of the bioclasts from the matrix corresponding mainly to crinoids.

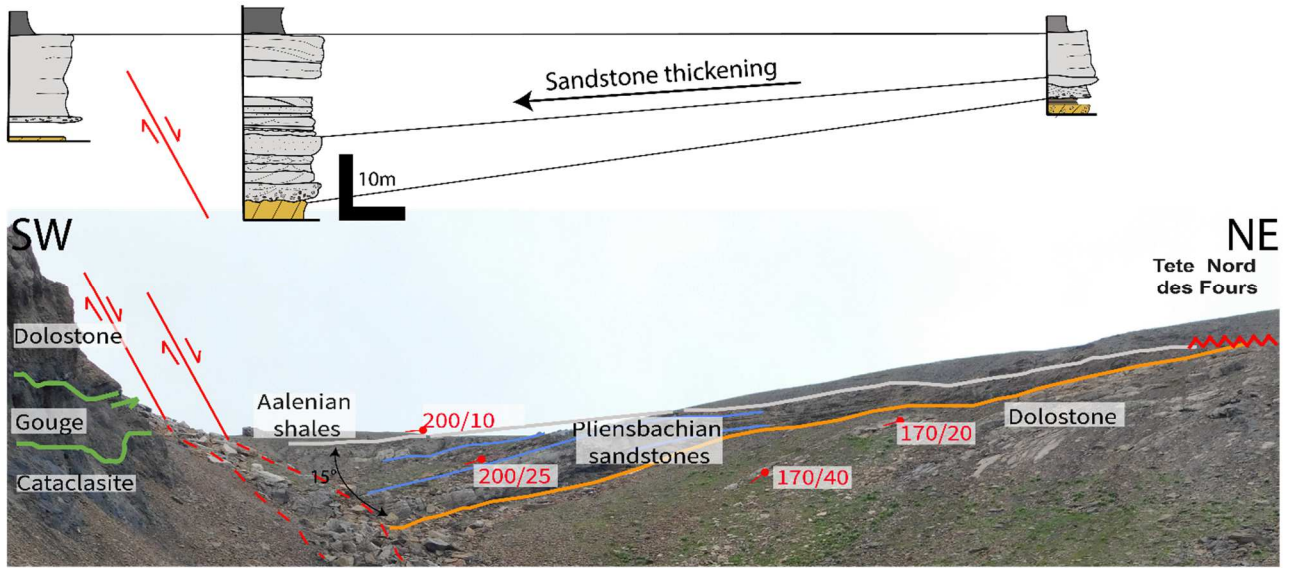


Fig5: High-angle normal fault offsetting the top basement tectonic contact about 30 m ($45^{\circ}43'53.45''/6^{\circ}43'51.70''$). A sedimentary wedge with thickening of sandstones toward the fault is observed in the Grès Singuliers (15°), indicating a likely Pliensbachian/Toarcian age for the fault.

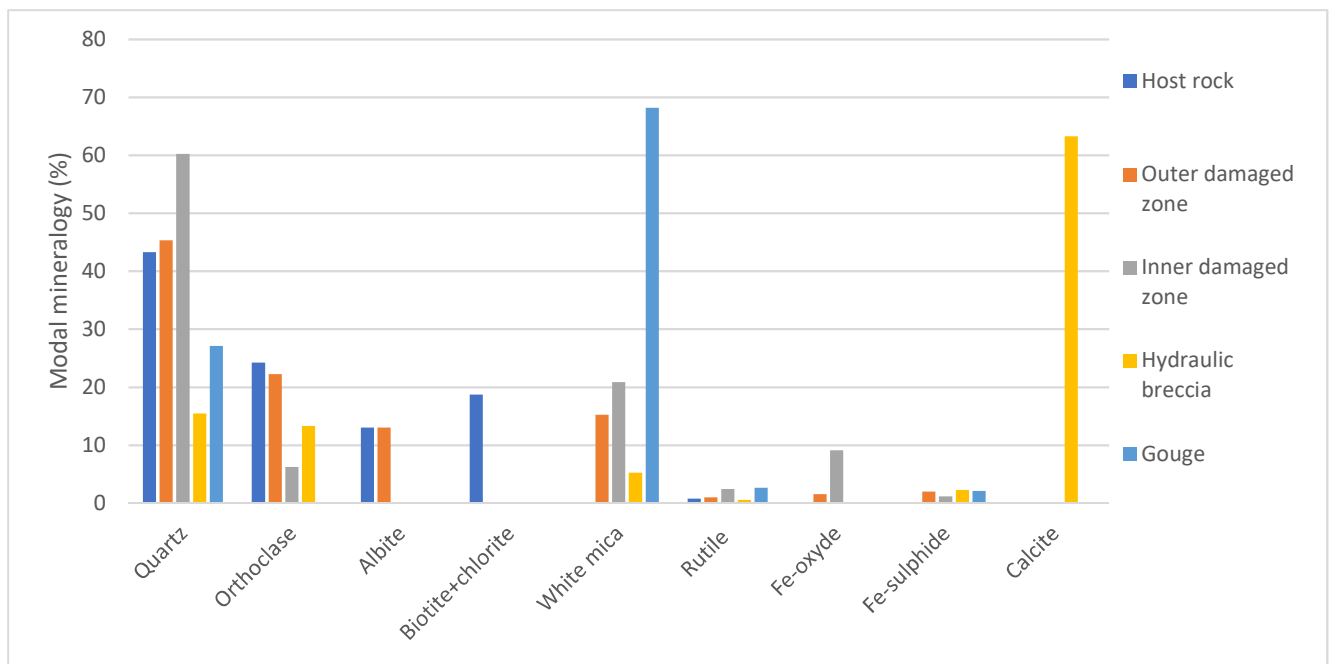


Fig6: Modal composition for some of the different fault rocks found in the Col du Bonhomme area.

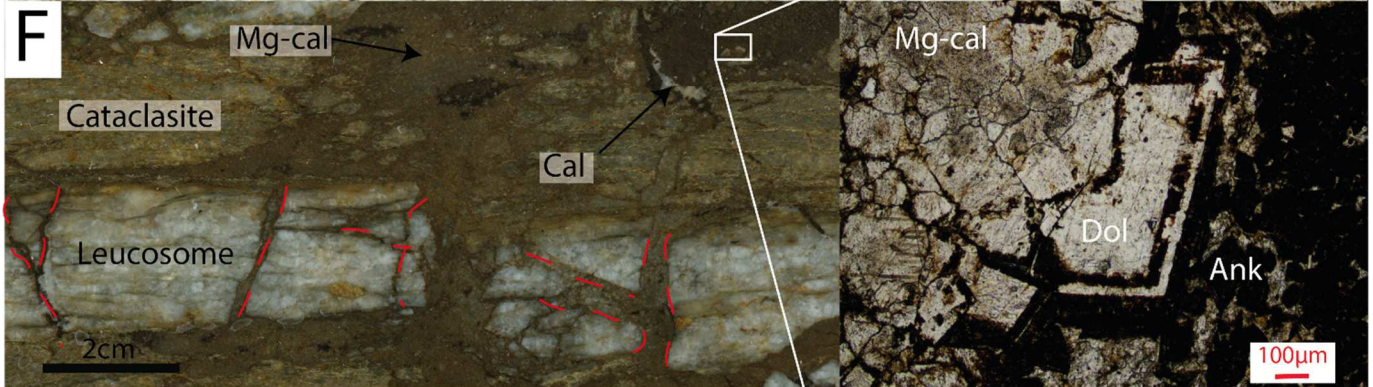
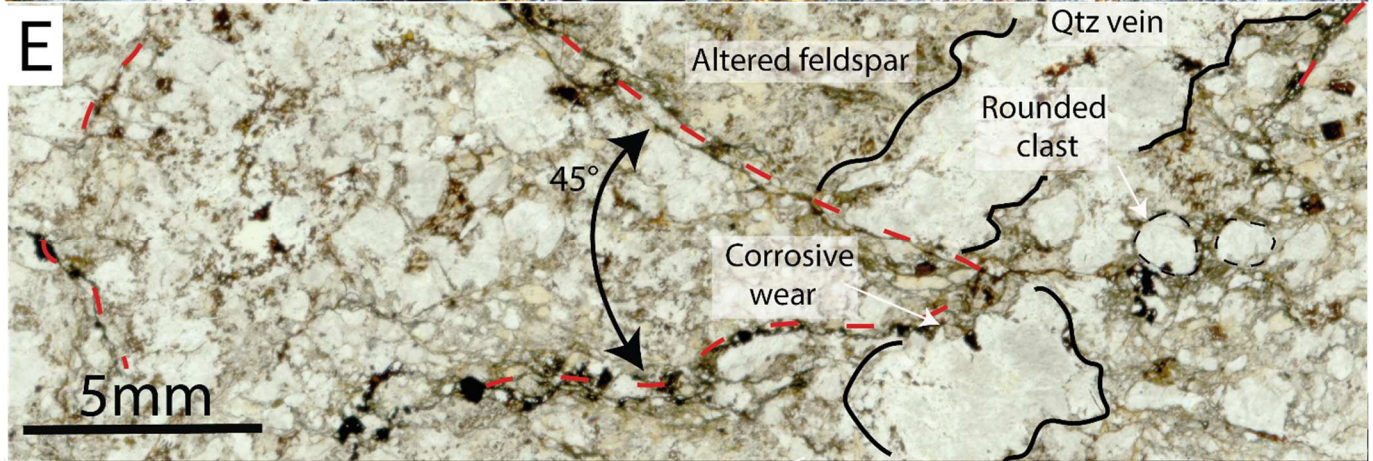
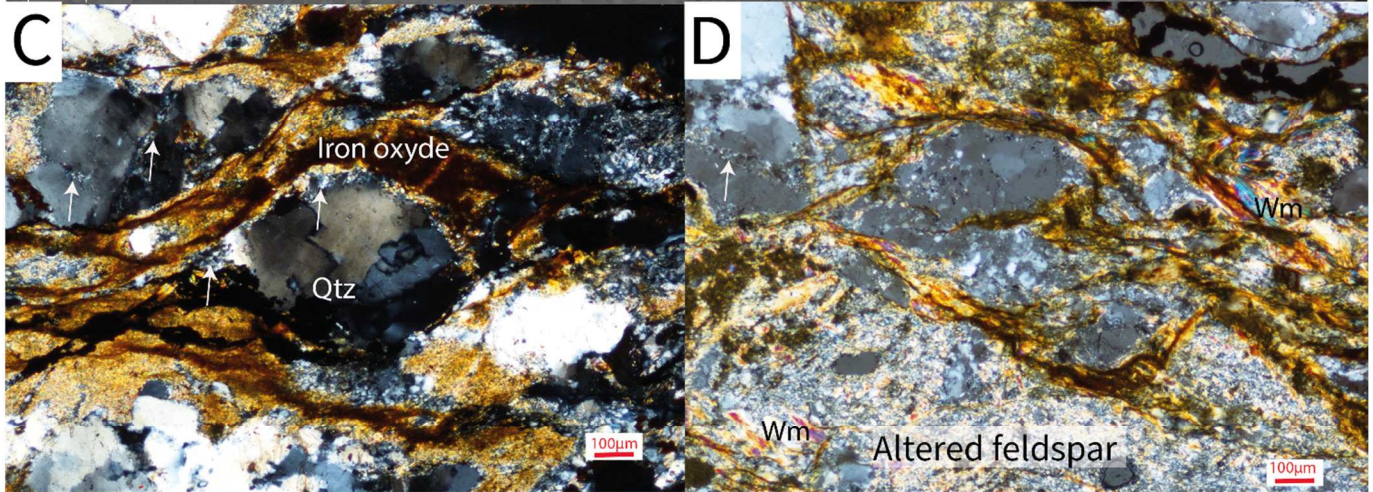
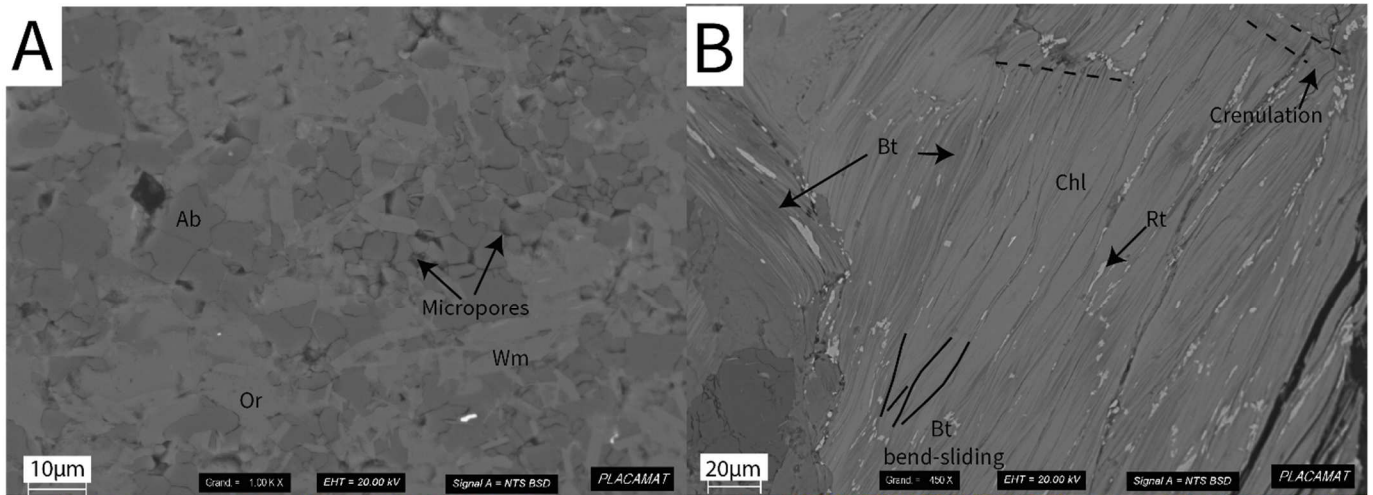


Fig7: A-Lowly altered orthoclase (Or) showing albite (Ab) and white mica (Wm) replacement with creation of microporosity (Scanning Electron microscope- Back Scattered Electron SEM-BSE). B- SEM-BSE image of the alteration of biotite (Bt) to chlorite (Chl) with formation of rutile (Rt). Note the alteration along the cleavage and on the borders of the crystals. Minor deformation such as crenulation is visible in the biotite and chlorite assemblages. C- Closer view of the cataclasite where numerous fractures filled with oxides are observed. Quartz (Qtz) exhibits fractures and minor plastic recrystallisation (bulging, white arrow) associated with undulose extinction (optical photomicrograph, crossed polarized light (XPL)) D- Cataclasite showing important feldspar alteration, with the formation of white micas associated with fracturing and oxides. Minor plastic recrystallization by bulging is visible in the prolongation of fractures (white arrow) (optical photomicrograph, XPL). E- Cataclasite showing fragmentation of the protolith, along with important fracturing of quartz and important hydration of feldspar (optical photomicrograph, plane polarized light (PPL)). Only few remnants of K-feldspars are observed. Partial dissolution of quartz with corrosive wear is clearly visible. Highly deformed quartz veins are observed in the cataclastic matrix, and oxide-rich fractures overprint the cataclasite. F- Hydraulic breccia with calcite cement (left: optical photographs of polished slab; right: XPL photomicrographs). The leucosomes are only affected by perpendicular fractures. The gneiss is affected by stronger alteration and deformation. The rock is overprinted by brecciation and subsequent dolomite (Dol) and ankerite (Ank) formation replaced by later by Mg-calcite (Mg-cal).

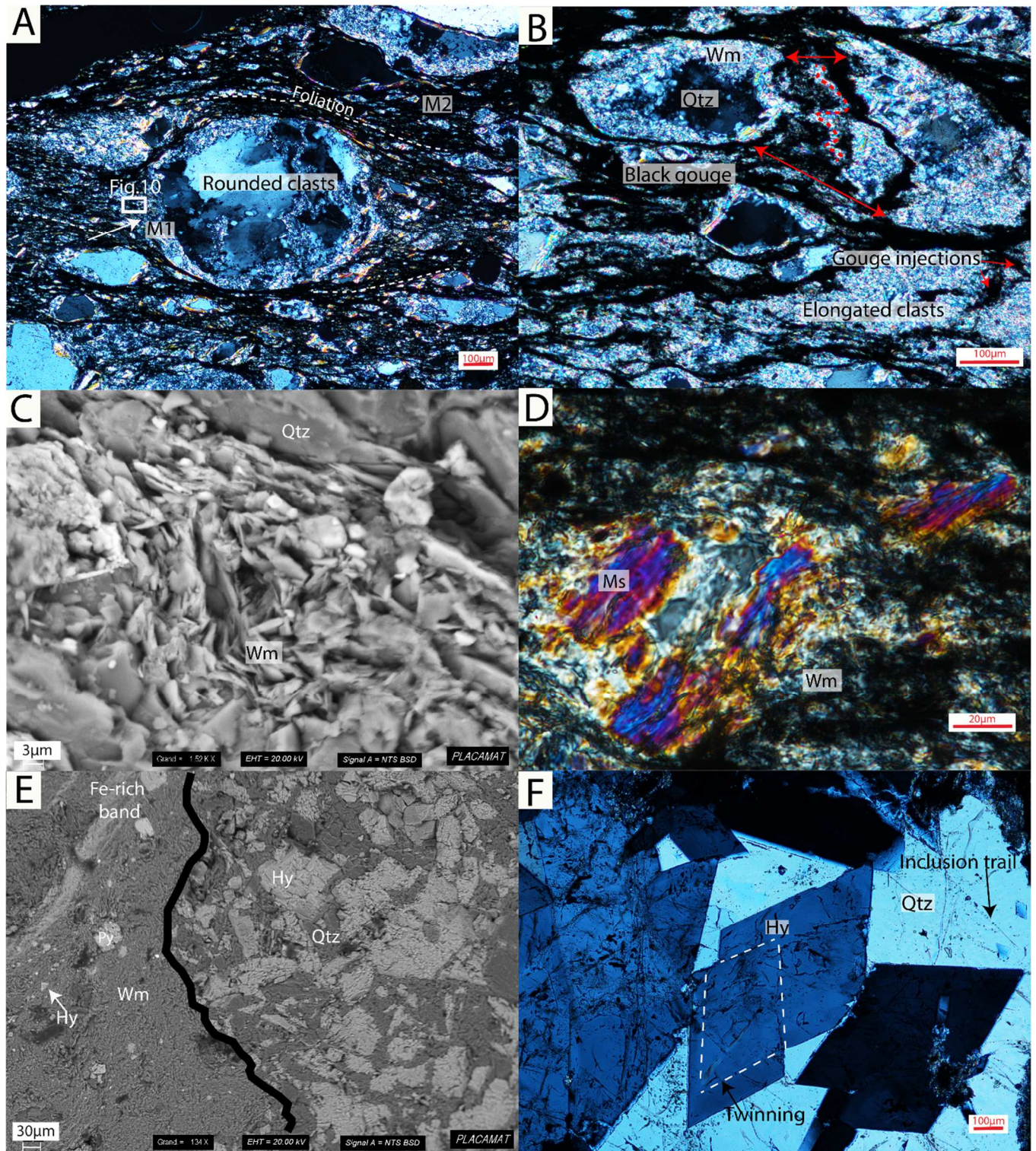


Fig 8: A-Microphotograph of the gouge showing a rounded clast embedded in the black matrix. The foliation is highlighted by the elongated clasts. The white arrow points at the cataclastic matrix (M1) of the clast progressively sheared and incorporated into the gouge matrix M2 (XPL). B- Microphotograph of the gouge matrix M1 with clasts of cataclasite

showing the progressive shearing (red arrows) and transformation into ultra-fine-grained black matrix M2 (XPL). C- Closer view of the black matrix of the gouge (M2) showing small illite minerals (around 5 μm), small quartz fragments (less than 10 μm) and pyrite (SEM-BSE, unpolished sample) D-Muscovite overgrowth over illite in the gouge (XPL). E- Irregular border between a hyalophane-bearing vein and the gouge. The latter is composed of small Fe-rich bands surrounded by illite-rich matrix. Clasts of hyalophane (Hy) are also observed in the gouge (SEM-BSE). F-Microphotograph of hyalophane and quartz in veins affected by important fracturing. The feldspars show important twinning of the crystals (XPL).

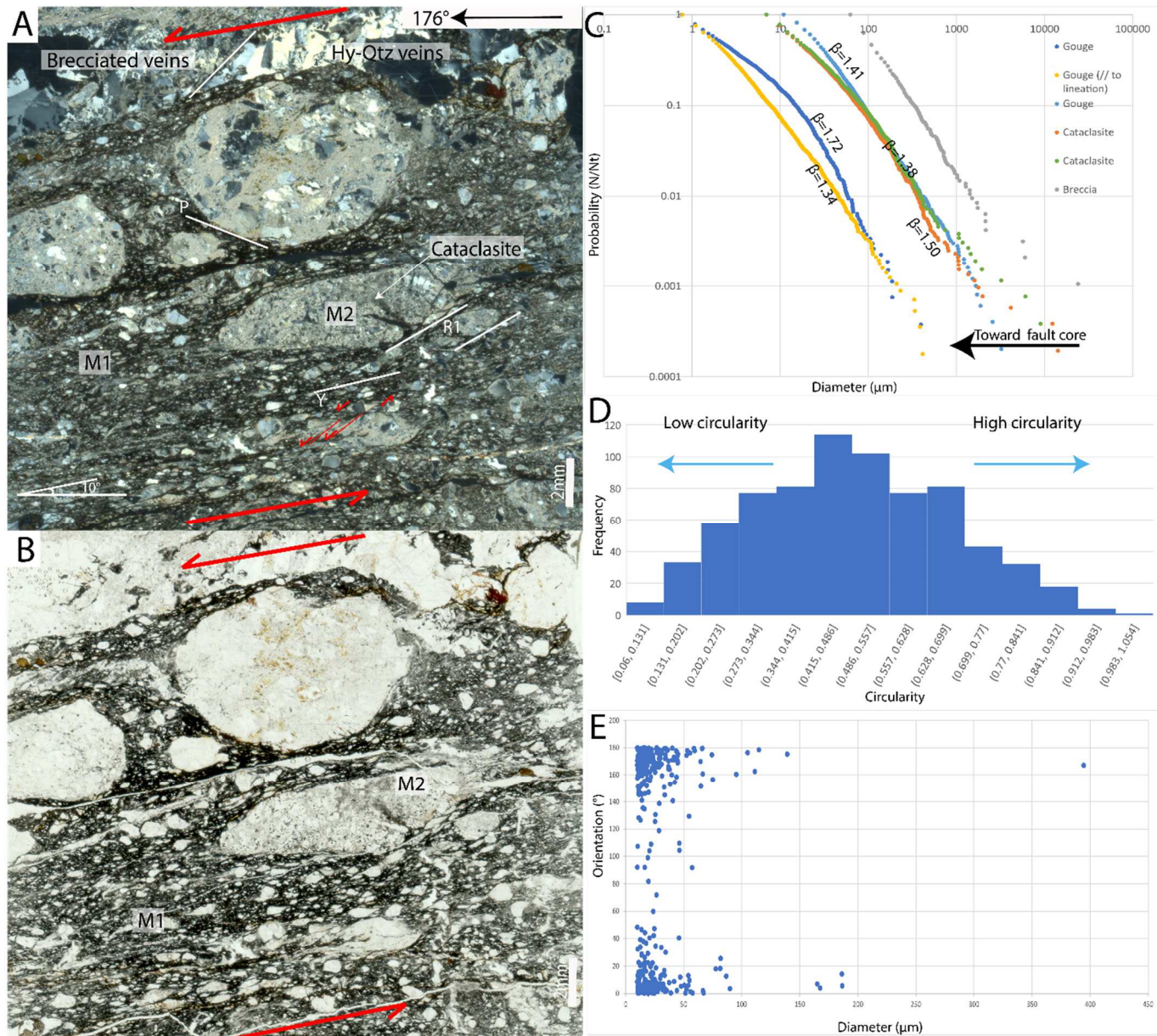


Fig 9: A and B- Microphotograph (XPL and PPL) of the black gouge reworking cataclasite fragments. The elongation of the clasts marks the foliation. Note the reworked clasts of cataclasite in the gouge matrix. C-Example of grain size repartition in gouge and cataclasite samples as obtained from image treatment. All the samples show a fractal dimension between 1.3 and 1.7. Note the progressive grain size diminution toward fault core. D-Grain size circularity of the gouge from image treatment showing a wide range of shapes, from angular to well rounded. E- Grain orientation versus grain size, showing a preferential orientation defining the foliation.

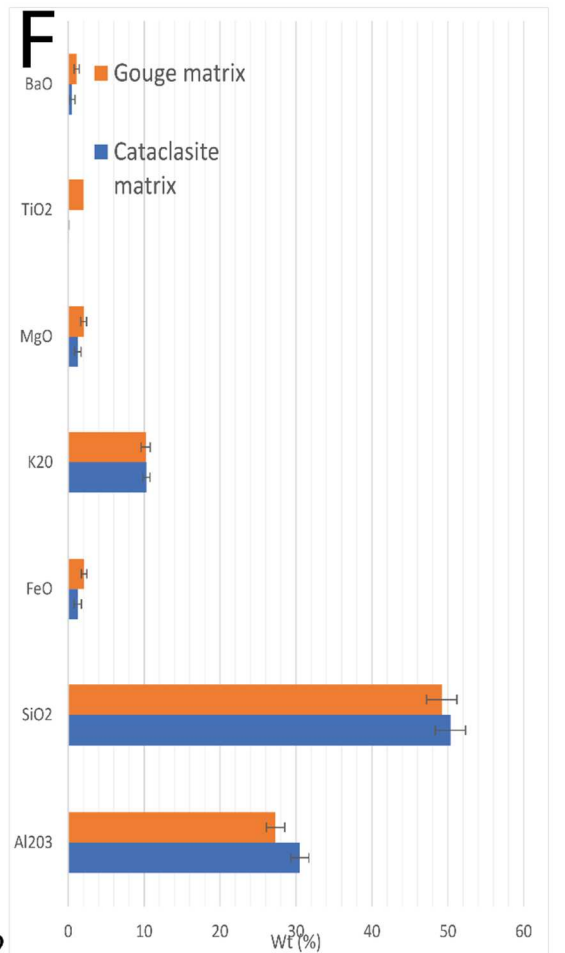
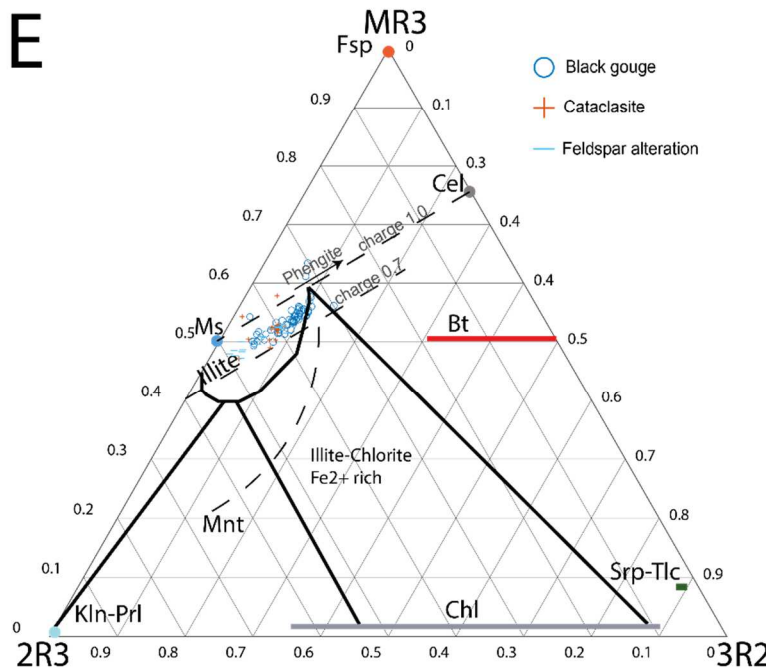
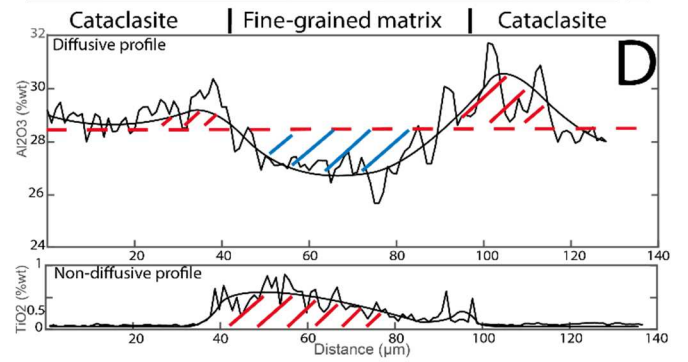
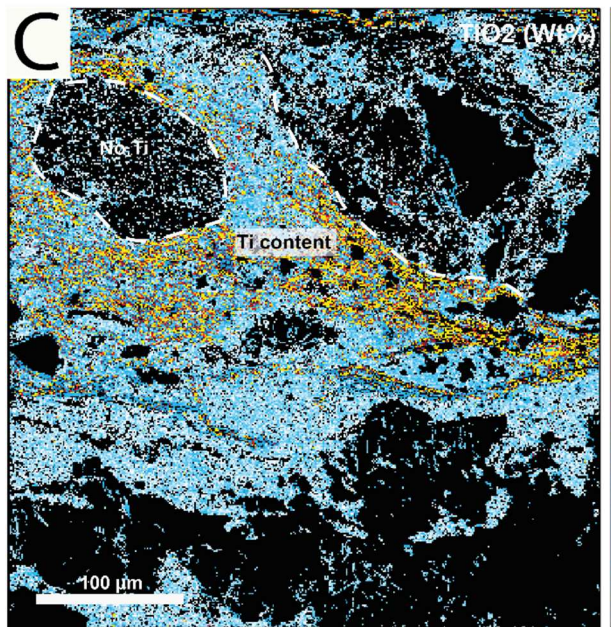
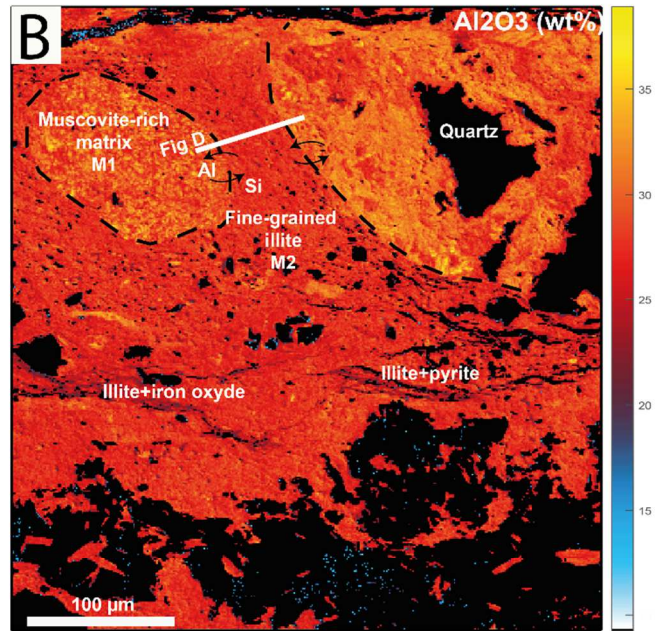
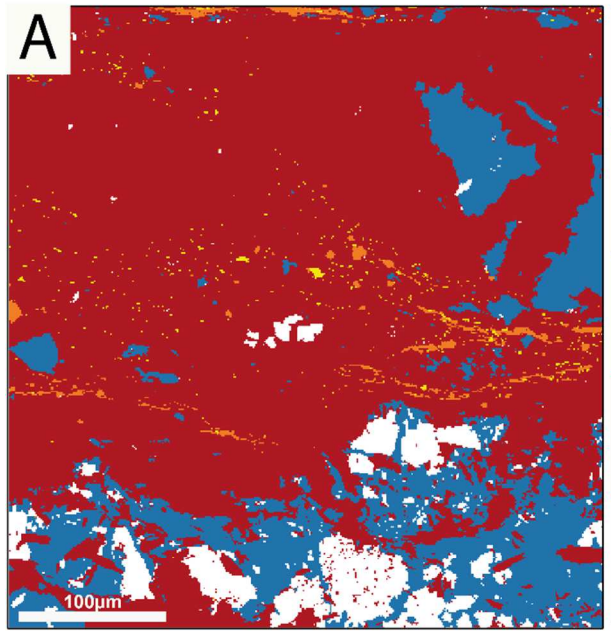


Fig10: A- Phase map from gouge microprobe mapping (Rt=rutile, Py=Pyrite, Wm=White micas, Qtz=Quartz, Kfs=K-feldspars). B and C - Microprobe maps of Al_2O_3 (B) and TiO_2 (C) showing distinct composition for the Wm phase. This highlights the different composition of the clasts of cataclasite (M1) and fine-grained gouge matrix (M2). The transition from the cataclasite matrix M1 to the gouge matrix M2 shows a loss of Al. Ti is absent from the cataclasite but hosted by rutile in the gouge. D- Profile across the micrometric fault zone for Al_2O_3 and TiO_2 . The Al_2O_3 content has a diffusive profile, contrary to TiO_2 (in red, the zone of gain and in blue the zone of loss). E- Ternary plot of the composition of phyllosilicates from the gouge. $MR3 = Na+K+2Ca$, $2R3 = (Al+Fe^{III}-MR3)/2$, $3R2 = (Mg+Fe^{II})/3$ (Velde, 1985). Ms = Muscovite, Cel = Celadonite, Fsp = Feldspars, Kln = Kaolinite, Prl = Pyrophyllite, Chl = Chlorite, Bt = Biotite, Srp = Serpentine, Tlc = Talc, Mnt = Montmorillonite. F- Average microprobe composition of the cataclasite matrix and gouge matrix (muscovite to illite) for the major elements.

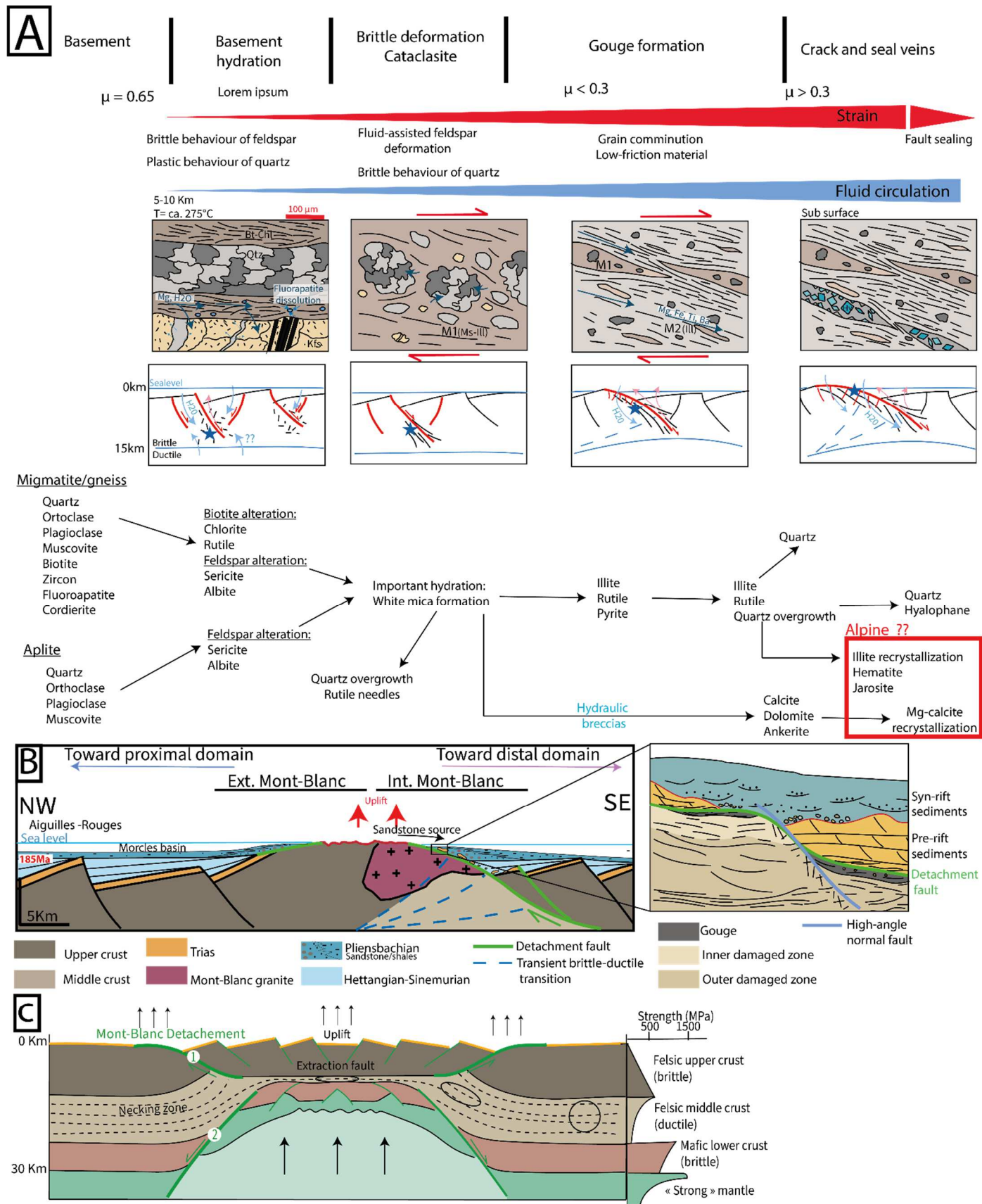


Fig11: A- Sketch summarizing the evolution of the fault rocks during exhumation following a retrograde path and a progressive localisation of the deformation in phyllosilicate rich layers (gouge) toward the surface. This is associated with the petrological evolution observed in the Col du Bonhomme from the gneiss to the matrix-supported gouge. The evolution

of frictional coefficient is indicated. B- Pliensbachian cross section of Mont-Blanc showing the crustal architecture and sedimentary infill associated to Mont-Blanc detachment fault (modified from Ribes et al. (2020)). Uplift related to the detachment fault led to local erosion and sandstones formation around the massif about the exhumed surface. C- The H-block model for the necking of the crust showing to conjugated fault system in the upper brittle crust and in the lower crust leading to crustal thinning (modified from Mohn et al. (2012)).

Sample	Location	Lithologies	Coordinates	Nb of spectra	R1	R2	T (°C)	1σ SD (°C)	95% SD (°C)	95% SEM (°C)
CB20-40	Col des Fours	Aalenian shales	45°45'52.06"/6°44'9.68"	21	2.47	0.71	322	6.52	18	3.96
CB20-41	Col des Fours	Aalenian shales	45°45'51.37"/6°44'7.22"	18	1.73	0.68	336	7.2	25.6	6.02
MB1c	Col des Fours	Liassic Carbonate	45°45'55.10"/6°44'16.06"	16	2.53	0.71	325	7.4	22.5	5.62

Table 1: Results of RSCM thermometry in the Col du Bonhomme area. SD=Standard deviation, SEM= Standard error of the mean, 95% confidence interval.

Samples	CB20-10										5a		5c				
	Protolith							Cataclasite				Gouge					
	Chl	Chl	Bi	Fsp	Fsp	Fsp	Fap	Wm	Wm	Wm	Wm	Wm	Wm	Hy	Hy		
SiO2	24.89	24.88	35.28	SiO2	65.13	63.98	68.63	CaO	52.18	SiO2	51.82	49.14	48.25	51.09	SiO2	54.49	53.50
TiO2	0.00	0.00	3.35	TiO2	0.00	0.00	0.23	MnO	0.98	TiO2	0.10	0.47	2.87	0.27	TiO2	0.00	0.00
Al2O3	20.81	20.68	19.15	Al2O3	19.09	18.46	18.88	FeO	0.57	Al2O3	25.95	26.66	29.73	27.92	Al2O3	20.17	20.17
Cr2O3	0.00	0.00	0.00	Cr2O3	0.00	0.00	0	F	2.67	Cr2O3	0.00	0.00	0.00	0.00	Cr2O3	0.00	0.00
Fe2O3	0.00	0.00	0.00	Fe2O3	0.00	0.00	0.17	MgO	0.03	Fe2O3	0.00	0.25	0.00	0.00	Fe2O3	0.00	0.00
FeO	28.30	27.51	17.96	FeO	0.00	0.00	0	Cl	0.07	FeO	2.56	1.76	1.47	1.99	FeO	0.04	0.04
MnO	0.44	0.44	0.35	MnO	0.00	0.00	0	P2O5	43.02	MnO	0.03	0.09	0.00	0.00	MnO	0.04	0.01
MgO	11.56	11.66	8.45	MgO	0.00	0.01	0.02	SrO	0.01	MgO	3.71	3.64	1.95	3.16	MgO	0.00	0.00
CaO	0.04	0.00	0.00	CaO	0.17	0.00	0.05	SO3	0	CaO	0.00	0.01	0.02	0.07	CaO	0.00	0.06
Na2O	0.00	0.02	0.38	Na2O	2.18	0.36	11.07	Na2O	0.09	Na2O	0.04	0.01	0.08	0.08	BaO	11.75	13.22
K2O	0.04	0.07	9.00	K2O	13.03	15.87	0.46	SiO2	0.06	K2O	10.78	10.58	9.61	10.33	Na2O	0.43	0.46
Total	86.08	85.26	93.92		99.60	98.68	99.51		99.67		94.99	92.62	93.98	94.91	K2O	10.94	10.43
															Total	97.86	97.89
Oxygens:	14.00	14.00	11.00		8.00	8.00	8.00				11.00	11.00	11.00	11.00		8.00	8.00
Si	2.72	2.74	2.71	Si	2.98	2.99	3.012		Si	3.47	3.38	3.24	3.40	Si	2.79	2.76	
Ti	0.00	0.00	0.19	Ti	0.00	0.00	0.008		Ti	0.01	0.02	0.15	0.01	Ti	0.00	0.00	
Al IV	1.28	1.26	1.73	Al	1.03	1.02	0.977		Al IV	0.53	0.63	0.76	0.60	Al	1.22	1.23	
Al VI	1.41	1.42		Cr	0.00	0.00	0		Al VI	1.52	1.53	-0.76	-0.60	Cr	0.00	0.00	
Cr	0.00	0.00	0.00	Fe3	0.00	0.00	0.006		Cr	0.00	0.00	0.00	0.00	Fe3	0.00	0.00	
Fe3	0.00	0.00	0.00	Fe2	0.00	0.00	0		Fe3	0.00	0.01	0.00	0.00	Fe2	0.00	0.00	
Fe2	2.59	2.53	1.15	Mn	0.00	0.00	0		Fe2	0.14	0.10	0.08	0.11	Mn	0.00	0.00	
Mn	0.04	0.04	0.02	Mg	0.00	0.00	0.001		Mn	0.00	0.01	0.00	0.00	Mg	0.00	0.00	
Mg	1.89	1.91	0.97	Ca	0.01	0.00	0.002		Mg	0.37	0.37	0.20	0.31	Ca	0.00	0.00	
Ca	0.01	0.00	0.00	Na	0.19	0.03	0.942		Ca	0.00	0.00	0.00	0.01	Ba	0.24	0.27	
Na	0.00	0.00	0.06	K	0.76	0.95	0.026		Na	0.01	0.00	0.01	0.01	Na	0.04	0.05	
K	0.01	0.01	0.88						K	0.92	0.93	0.82	0.88	K	0.71	0.69	
Sum	9.94	9.93	7.71	Sum	4.98	4.99	4.97		Sum	6.97	6.98	6.85	6.93	Sum	5.00	5.00	
XMg	0.42	0.43	0.46	XAn	0.83	0.00	0.21		XPg	0.48	0.09	0.96	0.98	XClS	23.74	26.73	
				XAb	20.12	3.37	97.11		XMs	50.34	56.12	72.23	57.39	XAb	4.30	4.60	
				XOr	79.05	96.63	2.68		XCel	49.19	43.79	26.81	41.63	XOr	71.96	68.66	
									Charge deficit	0.93	0.93	0.84	0.90				

Table 2: Microprobe analysis of minerals from the protolith, the cataclasite and the gouges. The structural formula were calculated with the software AX from (Holland and Powell, 2006). Ferric iron was recalculated according to the method of (Holland and Powell, 1998). Wm = White micas, Chl = Chlorite, Bi = Biotite, Fsp = Feldspars, Fap = Fluorapatite, Pg = Paragonite, Ms = Muscovite, Cel = Celadonite, An = Anorthite, Ab = Albite, Or = Orthoclase, Hy = Hyalophane, Cls = Celsian, XMg=Mg/(Mg+Fe).