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1 Time-constrained multiphase brittle tectonic evolution of the
2 onshore mid-Norwegian Passive Margin

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11 **ABSTRACT**

12 The mid-Norwegian Passive Margin (MNPM) is a multiphase rifted margin developed since the
13 Devonian. Its geometry is affected by the long-lived activity of the Møre-Trøndelag Fault
14 Complex (MTFC), an ENE-WSW oriented regional tectonic structure. We propose a time-
15 constrained evolutionary scheme for the MNPM brittle history. By means of remote sensing
16 lineament detection, field work, microstructural analysis, paleostress inversion, mineralogical
17 characterization and K-Ar dating of fault rocks, six tectonic events have been identified: i)
18 Paleozoic NE-SW compression forming WNW-ESE striking thrust faults; ii) Paleozoic NW-SE
19 transpression forming conjugate strike-slip faults; iii) Carboniferous proto-rifting forming NW-

20 SE and NE-SW striking faults; iv) Late Triassic-Jurassic (c. 202 and 177 Ma) E-W extension
21 forming ca. N-S striking epidote and quartz-coated normal faults and widespread alteration; v)
22 renewed rifting in the Early Cretaceous (c. 122 Ma) with a NW-SE extension direction; vi) Late
23 Cretaceous extensional pulses (c. 71, 80, 86, 91 Ma ago) reactivating pre-existing faults and
24 crystallizing prehnite and zeolite. Our multidisciplinary and multiscale study sheds light onto
25 the structural evolution of the MNPM and confirms the active role of the MTFC during the
26 rifting stages. Our sixty-two new radiometric K-Ar ages define discrete episodes of faulting
27 along the MNPM. The proposed workflow may assist the interpretation of the structural
28 framework of the MNPM offshore domain and also help to better understand fault patterns of
29 fractured passive margins elsewhere.

30 **1. INTRODUCTION AND AIMS OF THE STUDY**

31 Passive margins are major morpho-tectonic features of the Earth resulting from extension within
32 continents and the eventual tearing of continental lithosphere to form plate margins and oceanic
33 basins. In addition to the first-order general features common to all margins around the globe,
34 some passive margins may also exhibit remarkable local geological complexities and
35 peculiarities that reflect i) their generally complex polyphase faulting history, with strain and
36 deformation variably distributed in space and through time (Reston, 2005; Duffy et al., 2015;
37 Will and Frimmel, 2018; Phillips et al., 2019; Zastrozhnov et al., 2020); ii) the effects of variably
38 oriented and diachronic subsidiary rifting pulses that pre- and postdate the main rifting phase
39 (Doré et al., 1999; Ren et al., 2003); iii) the effects of the spatial orientation of old, inherited
40 structures, which generally represent zones of mechanical weakness within the crust and tend to
41 preferentially localize rifting (Phillips et al., 2019; Schiffer et al., 2019).

42 The mid-Norwegian Passive Margin (MNPM) is an archetypal multiphase rifted margin that
43 underwent a progressive evolution from the late orogenic collapse of the Caledonian belt in the
44 Early Devonian down to the Late Eocene (Faleide et al., 2008; Theissen-Krah et al., 2017;
45 Péron-Pinvidic and Osmundsen, 2020). It trends c. NE-SW and its southwestern termination
46 intersects the N-S North Sea margin. Its distinctive oblique geometry compared to the North Sea
47 margin is mainly due to the inherited structural grain of the Møre-Trøndelag Fault Complex
48 (MTFC), an ENE-WSW trending crustal-scale shear zone in mid-Norway (e.g., Séranne, 1992;
49 Redfield et al., 2004; Osmundsen et al., 2006; Schiffer et al., 2019; Gernigon et al., 2020; Péron-
50 Pinvidic et al., 2020). While the details of the shear zone's complex faulting history are still
51 being studied and discussed, it is widely accepted that the MTFC has been active at least from
52 the Late Silurian-Early Devonian to the present-day (Séranne, 1992; Braathen et al., 1999, 2002;
53 Osmundsen et al., 2005; Redfield et al., 2005; Sherlock et al., 2004; Nasuti et al., 2011).

54 The discovery of significant unconventional hydrocarbon reserves within exhumed offshore
55 fractured and weathered crystalline basement blocks of the MNPM (e.g., Petford and McCaffrey,
56 2003; Riber et al., 2015, 2017; Trice et al., 2019) has caused a surge of interest in the geometry,
57 petrophysical properties and development of intra-basement fracture networks and weathering in
58 the context of passive margin evolution (e.g., Fossen, 2010; Breivik et al., 2011; Riber et al.,
59 2015; Fredin et al., 2017; Scheiber and Viola, 2018; Trice et al., 2019; Fazlikhani et al., 2020;
60 Ceccato et al., 2021a, b). The MNPM is indeed characterized by a dense network of lineaments
61 that are the expression of brittle fault and fracture zones cumulated during the margin's long and
62 multiphase brittle evolution (Gabrielsen et al., 2002). They occur within both on- and offshore
63 basement units (Fredin et al., 2017; Trice et al., 2019). The structural brittle template of the
64 offshore domain has been mainly studied by seismic imaging and drilling. Currently available

65 geophysical techniques permit to only detect fracture and fault zones characterized by minimum
66 detectable throws >5-20 m, the so-called seismic-resolution scale features (Tanner et al., 2019).
67 On the other hand, onshore structural analysis permits to also study smaller brittle features at a
68 sub-seismic resolution scale (Braathen et al., 2004; Redfield and Osmundsen, 2009; Ksienzyk et
69 al., 2016; Scheiber and Viola, 2018; Scheiber et al., 2019; Ceccato et al., 2021a), which are key
70 to the definition of the structural permeability of basement blocks. A detailed study of the brittle
71 fault network exposed on onshore domains and of the related brittle deformation history can,
72 therefore, assist the reconstruction and interpretation of the regional geological framework and
73 evolution of the offshore domain, leading to high-resolution integrated models for the structural
74 evolution recorded by passive margins (e.g., Redfield and Osmundsen, 2009; Ksienzyk et al.,
75 2016; Gabrielsen et al., 2018; Scheiber and Viola, 2018) and better exploration predictive tools.

76 While decades of petroleum exploration have unveiled the fine details of the regional structure of
77 the offshore domain of the MNPM (e.g., Theissen-Krah et al., 2017; Gernigon et al., 2020;
78 Zastrozhnov et al., 2020 and references therein), so far only a few studies have focused on its
79 onshore counterpart, with the notable exception of the MTFC (e.g., Gabrielsen et al., 1999;
80 Kendrick et al., 2004; Redfield et al., 2004; Osmundsen et al., 2006; Nasuti et al., 2011).
81 Furthermore, although the polyphase evolution of the MNPM is quite well known and
82 established, absolute age constraints on its brittle faulting events are still largely missing, thus
83 preventing detailed reconstructions and correlations along and across the margin.

84 The intricate network of brittle fractures and faults affecting the offshore MNPM and the lack
85 of well dated sedimentary markers makes retrieving an absolute time sequence for its
86 deformation history a challenging task. All these difficulties notwithstanding, some studies have
87 demonstrated that it is possible to unravel complex brittle histories by studying the details of

88 geometric, kinematic, and geochronological brittle deformation features (e.g., Wilson et al.,
89 2006; Viola et al., 2009; Saintot et al., 2011; Lacombe et al., 2013; Mattila and Viola, 2014;
90 Scheiber and Viola, 2018; Nordbäck et al., 2022). A significant contribution in this respect stems
91 from the possibility to radiometrically date fault rocks associated with key structural features of
92 the margins so as to add absolute time constraints to brittle structural reconstructions (van der
93 Pluijm et al., 2001; Zwingmann and Mancktelow, 2004; Davids et al., 2013; Torgersen et al.,
94 2015; Ksienzyk et al., 2016; Vrolijk et al., 2018; Tartaglia et al., 2020; Fossen et al., 2021).

95 Aiming at better constraining the evolution of the MNPM, we present the results of a combined
96 multidisciplinary and multiscalar structural-geochronological study focusing on the post-
97 Caledonian evolution of the onshore MNPM and, for the first time, propose a time-constrained
98 evolutionary scheme for its brittle history based on sixty-two new K-Ar ages obtained from some
99 of its representative onshore fault zones. Our conceptual and analytical approach is of general
100 validity and may be useful to unravel complex brittle histories of similar tectonic settings
101 elsewhere.

102 **2. GEOLOGICAL SETTING**

103 The MNPM forms a 350-500 km wide sector of rifted and hyperextended continental crust
104 facing the Norwegian-Greenland Sea (Northeastern Atlantic Ocean; Fig. 1). The studied offshore
105 part of the MNPM is composed of two large segments, which, from SW to NE, include the Møre
106 and the Vøring margins (Fig. 1). Discrete, NW-SE-trending tectonic structures, such as the Jan
107 Mayen Corridor, separate these segments (Faleide et al., 2008; Gernigon et al., 2020), each of
108 which contains an inner platform, the Møre, and Vøring Platforms, and a system of outer basins,
109 the Møre and the Vøring Basins (Fig. 1). The studied onshore portion of the MNPM between 61°

110 54' N and 63° 55' N is mainly located within the Western Gneiss Region (WGR), (Fig. 1), which
111 is, generally composed of pervasively foliated Proterozoic gneisses locally hosting micaschist,
112 amphibolite and eclogite lenses (Krabbendam and Dewey, 1998; Corfu et al., 2014). The
113 formation of the MNPM results from a long extensional history that started with the collapse of
114 the Scandinavian Caledonides in the Late Silurian-Devonian and continued by deformation
115 localized during discrete tectonic pulses down to the breakup of Greenland from Scandinavia in
116 the early Eocene (Doré et al., 1999; Stemmerik et al., 2000; Faleide et al., 2008; Gernigon et al.,
117 2020). Based on sedimentary and structural observations onshore East Greenland, the initiation
118 of rifting is constrained to the mid-Carboniferous, while it is poorly constrained in the MNPM
119 (e.g., Stemmerik et al., 2000; Rotevatn et al., 2018). By this time, the WGR was exhumed to
120 shallow (<10 km) crustal levels, as inferred from the Early-Middle Devonian (380-400 Ma)
121 muscovite ^{40}Ar - ^{39}Ar cooling ages from the area (Kendrick et al., 2004; Walsh et al., 2013).

122 The MNPM accommodated two main phases of rifting, in the Permo-Triassic and Late Jurassic-
123 Cretaceous, respectively (Gernigon et al., 2020), with its early rifting history strongly controlled
124 by the structural grain inherited from the Caledonian orogen (Doré et al., 1997; Schiffer et al.,
125 2019). During the ENE-WSW-oriented Permo-Triassic rifting phase, the inherited brittle
126 Caledonian fault network was repeatedly reactivated (Faleide et al., 2008). This rifting stage
127 controlled the evolution of the North Sea and the proximal parts of the Vøring margin. (Fig. 1).
128 A quiescent post-rift phase ensued and lasted from the Middle Triassic to the Early Jurassic,
129 when the depositional environment in the area changed from marine to coastal (Gernigon et al.,
130 2020). In the Early and Middle Jurassic, intermittent and moderate episodes of E-W crustal
131 stretching occurred (Scheiber and Viola, 2018; Gernigon et al., 2020), leading to the initial
132 development of the Møre and Vøring margins (Fig. 1). The Late Jurassic–earliest Cretaceous

133 rifting event is related to the northward propagation of the Atlantic rifting (Lundin and Doré,
134 2011), when bulk extension shifted to a NW-SE direction and the final opening direction was
135 roughly perpendicular to the present-day coastline (Doré et al., 1999). During this phase, crustal
136 stretching and pervasive faulting created space to accumulate significant Cretaceous syn-rift
137 deposits (Osmundsen and Péron-Pinvidic, 2018; Zastrozhnov et al., 2020; Osmundsen et al.,
138 2021).

139 During the Cretaceous, the large and deep Møre and Vøring basins formed in the distal part of
140 the MNPM (Fig. 1, Blystad, 1995; Zastrozhnov et al., 2020). Both tectonic and thermal
141 subsidence led to the accumulation of up to 8 km thick sedimentary successions in local
142 depocenters. From the Late Cretaceous, the Møre Basin was subjected to thermal subsidence and
143 passive sedimentation without any significant tectonism (Osmundsen and Péron-Pinvidic, 2018;
144 Osmundsen et al., 2021). The Vøring Basin, instead, experienced renewed extension in the Late
145 Cretaceous (Zastrozhnov et al., 2018; 2020; Osmundsen et al., 2021). The final breakup of the
146 NE Atlantic Ocean initiated at 58-57 Ma in the Møre segment of the margin and later propagated
147 northward to the outer Vøring Margin at 56-55 Ma (Ren et al., 2003; Gernigon et al., 2020;
148 Zastrozhnov et al., 2020). Finally, compressional events affected the Vøring Basin from the
149 Eocene to the Pleistocene as indicated by local basin inversion structures and partial reactivation
150 of the faults bordering the many structural highs of the area (Lundin and Doré, 2002; Doré et al.,
151 2008; Zastrozhnov et al., 2020).

152 The investigated areas include the islands of Hitra, Frøya, and Smøla and the Stad area onshore
153 Norway, located at the transition between the NE-SW oriented MNPM and the N-S oriented
154 North Sea margin (Fig. 1). In the northern MNPM, the islands of Hitra and Smøla consist of the
155 450-428 Ma granitic to granodioritic Smøla-Hitra batholith (Gautneb and Roberts, 1989),

156 attributed to the Upper Allochthon of the Norwegian Caledonides (Fig. 1). The batholith is
157 overlain by Late Silurian-Devonian sandstone and conglomerate exposed along the southeastern
158 coast of Hitra. On the northern part of Hitra and on Frøya, micaschist and foliated migmatite of
159 the Uppermost Allochthon are exposed. In the Stad area, the tip of Kråkenes peninsula is
160 composed of a gabbroic enclave within the WGR (*cf.* Krabbendam et al., 2000).

161 **2.1 The Møre-Trøndelag Fault Complex (MTFC)**

162 The regional-scale, ENE-WSW Møre-Trøndelag Fault Complex (MTFC) extends onshore for
163 more than 300 km in mid- and western Norway before it enters the Norwegian Sea (Fig. 1;
164 Gabrielsen et al., 1999; Redfield et al., 2005; Nasuti et al., 2012). This structure has been related
165 to similarly oriented faults in the Scottish Highlands, such as the Highland Boundary Fault (e.g.,
166 Fossen, 2010). In the offshore domain, it separates the Jurassic–Cretaceous North Sea basin
167 system to the south from the wider and deeper Cretaceous basins of the MNPM to the north (Fig.
168 1; Redfield et al., 2005). Onshore, the MTFC crosscuts the Precambrian autochthonous basement
169 and the overlying Caledonian nappes (Braathen et al., 1999, 2002; Osmundsen et al., 2005;
170 Corfu et al., 2014).

171 The MTFC is a composite fault complex consisting of different subvertical fault strands (e.g.,
172 Hitra–Snåsa Fault, Verran Fault), and an enveloping volume of rock deformed by second-order
173 fault zones (Séranne, 1992; Braathen et al., 1999, 2002; Redfield et al., 2004, 2005; Osmundsen
174 et al., 2005, 2006; Redfield and Osmundsen, 2009). The main fault strands are defined by a
175 several meter thick mylonitic core, superimposed by brittle fault rocks (Kendrick et al., 2004).
176 The MTFC has been repeatedly reactivated in response to stress regimes varying through time,
177 resulting in a complex finite deformation pattern (e.g., Séranne, 1992; Gabrielsen et al., 1999;

178 Redfield et al., 2005; Osmundsen et al., 2006; Nasuti et al., 2011). A possible initial dextral
179 transpression has likely been accommodated by the MTFC in pre-Devonian times (Séranne,
180 1992). However, overall sinistral ductile Early to Middle Devonian shear is recorded by up to 2
181 km thick mylonitic shear zones, as documented by $^{40}\text{Ar}/^{39}\text{Ar}$ dating of synkinematic white micas
182 (Kendrick et al., 2004). From the Devonian onward, progressive regional exhumation led to the
183 transition into a shallower, brittle deformation regime. Previous studies propose multiple oblique
184 reactivations of the MTFC during the Late Devonian, Permo-Triassic, and Jurassic (e.g., Grønlie
185 and Roberts, 1989; Grønlie et al., 1991; Séranne, 1992; Sommaruga and Bøe, 2002; Redfield et
186 al., 2004, 2005). Pseudotachylytes from the Hitra-Snåsa Fault have been dated by $^{40}\text{Ar}/^{39}\text{Ar}$
187 laserprobe to the Late Carboniferous-Early Permian by Sherlock et al. (2004). Additionally, the
188 MTFC experienced normal faulting in the Middle to Late Jurassic–Early Cretaceous, as well as
189 possible Quaternary post-glacial reactivation (Sommaruga and Bøe, 2002; Redfield et al., 2005).
190 The area is still seismically active in an oblique-normal mode (Gabrielsen and Færseth, 1988;
191 Bungum et al., 1991).

192 **3. MATERIALS AND METHODS**

193 We carried out a multidisciplinary and multiscale study to better constrain the brittle evolution
194 of the MNPM in space and through time. Our approach includes the remote sensing of bedrock
195 lineaments, field structural and microstructural analysis, paleostress inversion, and radiometric
196 K-Ar dating coupled with the detailed mineralogical characterization of selected fault rocks. In
197 the following we provide details of the used methodologies.

198 **3.1 Remote Sensing Lineament Detection**

199 Lineament mapping was performed manually on a grayscale hillshaded Digital Elevation Model
200 (DEM) of the onshore MNPM obtained from high-resolution (1m/pixel) airborne-LiDAR (Light
201 Detection And Ranging) surveys. Grayscale hillshade DEMs with an illumination direction from
202 the NW were used to map the lineaments. Lineaments were mapped at the 1:10.000 and
203 1:100.000 scales to test the regional significance of local (smaller scale) fracture patterns. The
204 mapped lineaments are regarded as representing the surface expression of fault and fracture
205 zones (*cf.* Gabrielsen and Braathen, 2014; Scheiber et al., 2015). To this purpose, mapping was
206 assisted by and compared with available geological and topographic maps of the region to
207 prevent tracing ductile foliations and shear zones and artificial lineaments (e.g., roads, fences).
208 The strike of the mapped lineaments is displayed in rose diagrams computed with the MARD
209 (Moving Average Rose Diagram) application by Munro and Blenkinsop (2012).

210 **3.2 Structural Analysis**

211 Field structural analysis focused on the easily accessible coastline of the MNPM, from Hitra in
212 the north to the Stad area in the south (Fig. 1). Fieldwork was used to ground truth the remotely
213 detected lineaments. 86 different structural sites from selected key areas of the MNPM were
214 studied by systematically collecting fault-slip data, including information on fracture or fault
215 orientation, slip direction and sense of movement, lithotype, fault rock type, and fracture
216 mineralogy. In some cases, throw and spatial persistence/extension could also be were measured.
217 This fault-slip dataset represents the input data for our paleostress inversion.

218 In addition, oriented thin sections of selected fault rock samples, cut parallel to the transport
219 direction (fault slip lineation) and perpendicular to the fault plane, were prepared for optical
220 microscopy and petrographic analysis.

221 3.3 Paleostress Inversion

222 Paleostress inversion analysis is useful to reconstruct the orientation of the stress field that
223 caused slip along a given homogeneous set of cogenetic faults. When studying a multiply
224 reactivated and complexly deformed area, paleostress inversion analysis can ideally help to
225 distinguish and reconstruct how the stress field orientation varied through time, during different
226 deformation events. In our study, the collection of fault-slip data included the fault plane
227 orientation, the slip direction and the sense of slip (normal, reverse, dextral or sinistral).
228 Kinematic analysis relied on the growth direction of slickenfibers behind slickensides, Riedel
229 shears and small-scale pull-apart structures (e.g., Hancock, 1985; Petit, 1987).

230 Starting from the fault-slip data, paleostress inversion calculates the reduced stress tensor that
231 best accounts for an internally homogeneous set of faults. This approach is based on the
232 following assumptions (e.g., Wallace, 1951; Bott, 1959; Twiss and Unruh, 1998; Pollard, 2000;
233 Lacombe, 2012; Lacombe et al., 2013): 1) the observed slip direction is parallel to the maximum
234 shear stress resolved on the fault plane; 2) the faulted volume of rock is physically homogeneous
235 and isotropic; 3) the faulted rock volume has to be large compared to the dimension of the
236 studied faults and a homogenous stress distribution therein is necessary; 4) the studied medium
237 responds to the applied stresses as a rheologically linear material; 5) faults do not mutually
238 interact, and 6) no block rotation occurred at the time of faulting. These assumptions still
239 represent oversimplifications of the complexity of faulting in natural systems and by accepting
240 them we necessarily introduce uncertainties in any paleostress analysis. Moreover, they are only
241 seldomly met when studying complexly fractured basements, where the finite fault and fracture
242 pattern derives from multiple and thus superimposed faulting events. This may result in reduced
243 stress tensors whose definition and interpretation are dubious.

244 On the other hand, all these uncertainties notwithstanding, several studies have shown that it is
245 possible to obtain consistent paleostress reconstructions even in multiply deformed metamorphic
246 terrains by acquiring a large dataset, carefully sorting statistically fault populations and by
247 validating the results against independent constraints (e.g., Viola et al., 2009; Saintot et al., 2011;
248 Mattila and Viola, 2014).

249 When dealing with multiple faulting events, the identification of internally consistent subsets of
250 faults separated from the total and heterogeneous available dataset is a fundamental step. A
251 homogeneous set of faults is defined as a group of faults and fractures in any given area that can
252 be genetically bundled together because they formed in response to the same stress field and, as
253 such, they are compatible from a geometric, kinematic and dynamic perspective and have similar
254 fault rock assemblages and mineral coatings. In this study, observations from remote sensing,
255 field constraints and microstructural analysis were considered to sort the total fault-slip dataset
256 into internally consistent subsets of faults from along the studied margin.

257 The individual fault sets were analyzed using the WinTensor software (Delvaux and Sperner,
258 2003) to obtain their reduced stress tensors. This software iteratively applies an inversion
259 algorithm to search for the state of stress that best accounts for the input fault-slip dataset (e.g.,
260 Angelier, 1984; Delvaux and Sperner, 2003). The robustness of the optimization process is
261 measured by the fit between the theoretical slip vectors and those measured in the field (Delvaux
262 and Sperner, 2003). This is achieved by analyzing for all faults in a set the misfit angle α , which
263 is defined as the acute angle between the theoretical maximum shear traction vector and the
264 measured slip vector for an individual fault plane. A set of faults can be considered as
265 homogeneous if the slip deviation (the α value) is lower or equal to 30° , a condition that implies
266 that all processed faults are compatible with the calculated stress tensor. If the α value for a given

267 fault is greater than 30° , then that fault is incompatible with the computed reduced tensor,
268 suggesting that it belongs to a different set and that it formed in a different stress field and, thus,
269 during a different deformation episode (Delvaux and Sperner, 2003).

270 The calculated reduced stress tensor is defined by the orientation of the three main compressional
271 stress axes (σ_1 , σ_2 , and σ_3 , with $\sigma_1 \geq \sigma_2 \geq \sigma_3$) and by the stress ratio R ($R = (\sigma_2 - \sigma_3)/(\sigma_1 - \sigma_3)$)
272 (Angelier, 1984; Lacombe, 2012). To express in a clearer way the stress ratio and the resulting
273 stress regime, we rely on the R' index, which is calculated by $R' = R$, when σ_1 is vertical
274 (extensional stress regime), $R' = 2 - R$, when σ_2 is vertical (strike-slip stress regime) and $R' = 2$
275 $+ R$, when σ_3 is vertical (compressional stress regime; Delvaux and Sperner, 2003). R' ranges
276 from 0 to 3. $R' = 0.5$ describes a pure extensional regime, $R' = 1.5$ describes a pure strike slip
277 regime and $R' = 2.5$ indicates a pure compressional regime. All the intermediate values indicate
278 a mixed regime, e.g., $R' = 1.0$ translates to a transtensional regime.

279 In summary, the reduced stress tensors presented in this study describe paleostress regimes
280 inverted from different, manually sorted sets of faults from along the margin. The obtained
281 paleostress regimes are correlated with specific tectonic events in the regional framework
282 experienced by the MNPM.

283 **3.4 K-Ar Radiometric Dating of Fault Rocks Coupled with Mineralogical Analysis**

284 K-Ar radiometric dating and X-ray diffraction (XRD) were performed on twelve fault gouges
285 and one non-cohesive altered rock at the laboratories of the Geological Survey of Norway
286 (Trondheim, Norway) following the procedure described by Viola et al. (2018). Collected fault
287 rock samples are representative of the main trends of structural lineaments defined by remote
288 sensing analysis. Each sample was disintegrated by repeated freezing and thawing cycles to

289 avoid mechanical grain size reduction of coarse-grained minerals and their contamination in the
290 finer fractions. Samples were, then, separated into five grain size fractions (<0.1, 0.1-0.4, 0.4-2,
291 2-6 and 6-10 μm). The <2, 2-6 and 6-10 μm grain size fractions were separated in distilled water
292 using Stokes' law, whereas the finer fractions (<0.1, 0.1-0.4 and 0.4-2 μm) were concentrated by
293 high-speed centrifugation of the <2 μm fraction. The mineralogical composition of each grain
294 size fraction was determined by XRD analysis (with a Bruker D8 Advance diffractometer).
295 Mineral quantification was carried out on randomly prepared specimens using Rietveld
296 modelling with the TOPAS 5 software. The lower limit of quantification and accuracy are
297 mineral-dependent but are generally 1 wt% and 2-3 wt%, respectively. For further details, the
298 reader is referred to Viola et al. (2018) and Tartaglia et al. (2020).

299 **4. RESULTS**

300 **4.1 Bedrock Lineament Pattern**

301 The results of the manual extraction of bedrock lineaments from the coastal area of the MNPM at
302 the 1:100.000 (Fig. 2a) and 1:10.000 scales (Fig. 2b, c) show that lineaments can be grouped into
303 four main sets:

- 304 i) (E)NE-(W)SW lineaments, largely parallel to the coastline and to the MTFC;
- 305 ii) NNW-SSE lineaments;
- 306 iii) WNW-ESE lineaments;
- 307 iv) E-W lineaments.

308 The (E)NE-(W)SW oriented trend is dominant and particularly evident on Hitra and Smøla (Fig.
309 2b), where one of the strands of the MTFC, the Hitra-Snåsa Fault, is well exposed.

310 When comparing the rose diagrams obtained from the two different scales at which the remote
311 sensing was carried out, some differences can be recognized. This reflects the number of picked
312 lineaments, which is generally one or two orders of magnitude greater at the 1:10.000 scale than
313 at the smaller scale (Fig. 2). Despite the different number of mapped lineaments, however, the
314 same main trends emerge from both mapping scales, with some exceptions for the Runde and
315 Kråkenes areas. The 1:100.000 scale lineament trends in the Runde area are oriented NNW-SSE,
316 NNE-SSW and ENE-WSW (Fig. 2a). The ENE-WSW lineaments define the main trend in the
317 1:10.000 scale stereoplot of the Runde area (Fig. 2c), but the NNW-SSE oriented lineaments are
318 less abundant than at the 1:100.000 scale (Fig. 2a). A similar situation is evident in the Kråkenes
319 area, where the distinct WNW-ESE lineament trend derived from the 1:100.000 scale is less
320 abundant at the 1:10.000 scale (Fig. 2a, c). This observation could indicate that NNW-SSE and
321 WNW-ESE lineaments tend to be better expressed at smaller scale because they represent older,
322 regional structural trends. Their abundance at the 1:10.000 scale could also be relatively smaller
323 in comparison to the great number of variably oriented faults and fractures recorded and mapped
324 at this scale.

325 The mapped lineaments exhibit crosscutting relationships that can be summarized as follows
326 (Fig. 3): WNW-ESE trending lineaments show mutual crosscutting relationships with all the
327 detected trends. In the area of Hitra, the ENE-WSW lineaments are persistent and seem to cut all
328 other lineaments, even if they are locally crosscut by NNW-SSE and NE-SW trends. NNW-SSE
329 and NE-SW trending lineaments are generally crosscut by all the detected trends, and locally,
330 NE-SW trending lineaments crosscut the NNW-SSE lineaments. According to these results,
331 ENE-WSW striking lineaments thus appear to be the youngest, or the ones that have been
332 reactivated more recently, and the NNW-SSE and NE-SW the oldest.

333 4.2 Mesostructural Brittle Fault Characterization

334 The structural stations studied along the MNPN are shown in Figure 4a. Faults and fractures at
335 the studied sites cut across different types of rock. The host rocks along the MNPM are mainly
336 represented by granitic and granodioritic gneiss of the WGR, Devonian sandstone and
337 conglomerates, micaschist, foliated migmatite and locally eclogitic lenses.

338 The collected fault-slip data have variable strike and dip. The associated striae are also variably
339 oriented, indicating both dip-slip, predominantly with normal kinematics, and, to a lesser extent,
340 strike-slip faulting. Faults locally rework and exploit inherited ductile planar fabrics, mainly the
341 mylonitic foliation (Fig. 5a). On the island of Jøsnoya, south of Hitra, high-angle conjugate
342 sinistral ENE-WSW and dextral NE-SW oriented fault planes are exposed. The sinistral ENE-
343 WSW high-angle faults rework the mylonitic fabric (Fig. 5a). The strikes of these faults are
344 parallel to the main remotely detected lineament set in the area. These faults reactivate the
345 mylonitic planar fabric and are systematically cut by high angle NNW-SSE trending normal
346 faults (Fig. 5b). These normal faults are locally decorated by fine-grained epidote mineral
347 coatings. Also, the strike slip faults on Jøsnoya are, in turn, cut by younger calcite veins.

348 Brittle fault zones along the margin are associated with a broad range of fault rock assemblages.
349 In highly fractured fault cores, fine-grained cataclasite is associated with or crosscut by distinct
350 layers of clay-rich fault gouges (Fig. 5c, d). Due to the lack of marker horizons in the study area,
351 the amount of displacement along individual faults remains largely unknown. Locally, along
352 some brittle structures, the host rock is dismembered, and appears as unconsolidated coarse-
353 grained material. In these fault and fracture zones, the mostly granitic or gabbroic host rock is
354 altered.

355 Brittle structures along the MNPM exhibit different mineralogical coatings, including epidote,
356 chlorite, calcite, and quartz, sometimes decorating the same plane but also crosscutting and
357 overprinting each other, forming a complex multilayered arrangement of fault rocks. Epidote
358 coatings on faults and fractures are generally reworked and/or cut by younger, few mm-thick,
359 milky white prehnite veins (Fig. 5e). Calcite occurs as mm-thick striated synkinematic coating
360 on the fault planes (Fig. 5f), as variably thick veins, or secondary, coarse-grained, post-kinematic
361 fracture infill.

362 The studied fault and fracture planes have been classified according to their mineralizations and
363 plotted in rose diagrams (Fig. 4b). Green, ultrafine-grained, epidote coatings are mainly
364 associated with c. N-S trending faults but are also locally present on differently oriented faults
365 (Fig. 4b). Fine-grained quartz mineralizations and quartz veins show a similar pattern but are
366 particularly abundant on E-W striking faults. Red iron oxide coating mainly occurs on WNW-
367 ESE striking faults (Fig. 4b). Milky white, extremely fine-grained, mm to cm thick prehnite
368 coating is mainly associated with roughly N-S and NNW-SSE striking faults. Dark blue/green,
369 mm-thick, chlorite coating does not display any preferential orientation. Calcite is present on all
370 sets of fractures and faults, irrespective of their orientation, but shows a predominance along N-S
371 striking brittle structures (Fig. 4b).

372 ***4.2.1 Fault Zone K-Ar Dating***

373 Twelve fault gouges and one altered rock were sampled for radiometric K-Ar dating. The details
374 of the sampled fault zones are reported in Table 1 and Figure 6. The sampled fault gouges belong
375 to three main sets of faults: i) low-angle (W)NW-(E)SE striking fault planes (samples 19.006B,
376 19.078, and 19.016; Fig. 6a-c); ii) variably dipping, ENE-WSW oriented faults (samples

377 19.007A, 19.011, 19.042A, GT18_01/2; Fig. 6d-i); iii) medium to high-angle, oblique and dip-
378 slip, NNE-SSW oriented faults (samples 19.030A, 19.049, 19.070, 19.076; Fig. 6j-m). Samples
379 are from different rock types (Table 1), mainly granitic and gabbroic rocks (19.006B, 19.007A,
380 19.016, 19.030A, 19.042A/E, 19.078), migmatite (19.049) and amphibolite or gneiss (19.011,
381 19.070, 19.076, GT18_01, GT18_02).

382 Some of the sampled fault zones contain a complex assemblage of different fault rocks within
383 the core with different generations of fault gouges, cataclasites and veins. The low-angle N-
384 dipping fault of sample location 19.006B (Fig. 6a), for example, contains a 7 cm thick fault
385 gouge layer associated with a subparallel blocky calcite vein. The NNE-dipping fault at location
386 19.016 contains a layer of gouge associated with subparallel veins of calcite and zeolite (Fig. 6c).
387 Fault zones 19.078 and 19.011 are composed of a few cm thick fault gouge layer (Fig. 6b, e).

388 Samples 19.042A and E are from an ENE-WSW striking strike-slip fault and are from two
389 different structural levels (Fig. 6f, g). In the lower part of the outcrop, a grey, very plastic fault
390 gouge is exposed (sample 19.042A), whereas the upper part of the same brittle structure is filled
391 with altered/weathered sandy granite (sample 19.042E).

392 Samples 19.030A and 19.049 are from SE-dipping fault zones; both their fault cores are
393 composed of a c. 3 cm thick layer of cohesive fine-grained cataclasite, chunks of blocky calcite
394 veins and the sampled fault gouge (Fig. 6j, k). The reddish fault gouge of the ENE-WSW
395 striking fault at locality 19.007A also embeds calcite veins (Fig. 6d).

396 Samples GT18_01 and GT18_02 were collected from the same c. 7 m thick fault zone on the
397 island of Runde (Fig. 6h). They are from two distinct generations of gouge, with GT18_01

398 cutting across GT18_02. The two gouge layers have different angle of dip, where the younger
399 gouge layer (GT18_01) has a steeper dip (Fig. 6i).

400 Finally, some brittle fault zones exploit inherited ductile planar fabrics. For example, NNE-SSW
401 striking mylonitic shear zones were reworked by brittle faulting (e.g., samples 19.070 and
402 19.076; Fig. 6l, m).

403 **4.3 Microstructural Data**

404 Eight samples of cataclasite were collected from outcrops of representative fault zones at the
405 investigated structural sites (Fig. 4a) for microstructural characterization. Although the collected
406 samples cannot be considered representative of the entire population of faults along the margin,
407 their analysis allows us to derive structural and mineralogical inferences as to the conditions of
408 deformation and to define systematic and mineralogically-constrained crosscutting relationships.
409 Microstructural analyses focused on: (i) the identification of the mineralogy and microstructure
410 of different mineral coatings observed in the field; (ii) the identification of overprinting
411 relationships between different coatings and thereby (iii) deducing a temporal sequence of
412 deformation events.

413 Our study revealed a systematic sequence of overprinting relationships between veins and
414 cataclastic layers characterized by specific mineral phases. From the oldest to the youngest, this
415 sequence is defined by the following evidence:

416 (i) Protocataclasites are generally pervasively fractured and characterized by the
417 occurrence of (coarse-grained) epidote mineralizations (Fig. 7a). Multiple events of
418 fracturing and precipitation of epidote-bearing mineralizations are inferred from the

419 occurrence of epidote-bearing cataclasites with remarkably different grain size and
420 crosscutting relationships (e.g., Ep₁, Ep₂ and Ep₃ in Fig. 7a).

421 (ii) Quartz veins and quartz-rich cataclasites of variable grain size dissect the epidote-
422 bearing cataclasites (Fig. 7b). Several generations of quartz veins and quartz-rich
423 cataclasites can be identified (e.g., Qtz₁, Qtz₂ in Fig. 7b).

424 (iii) Prehnite-bearing veins and cataclastic layers cut across and disrupt the quartz- and
425 epidote-bearing cataclasites described above (Fig. 7c-d). Quartz- and epidote-layers
426 are observed as mm-size clasts embedded within prehnite veins and cataclastic layers.

427 (iv) Fine-grained cataclasites of mixed mineralogical composition containing large clasts
428 of epidote-, quartz-, and prehnite-bearing cataclastic layers and veins (Fig. 7e).

429 Fault gouges are commonly associated with calcite mineralization and subordinated red iron
430 oxides and quartz (Table 1). Calcite-bearing mineralizations can be coarse-grained (mm-size)
431 and are locally associated with euhedral zeolite (Fig. 7f).

432 **4.4. Paleostress Inversion**

433 The total analyzed fault-slip dataset was subdivided into heterogeneous subsets according to a
434 geographic position criterion in order to obtain subsets not exceeding 350 fault-slip data each,
435 which are more easily handled. Inversion was then applied to fault-slip data subsets that were
436 defined based on the criteria explained above, aiming at the identification of internally consistent
437 data according to field constraints, fault orientation, kinematics, and fault characteristics, such as
438 specific mineralization (quartz, epidote, calcite, chlorite), or a specific area with a given
439 geological feature (occurrence of foliation, type of protolith, etc.).

440 Each obtained dataset was inverted, and initial results were progressively refined by discarding
441 faults with misfit angles $>30^\circ$. The inversion procedure was reiterated until stable results were
442 obtained (see also Scheiber and Viola, 2018). As this approach was applied iteratively, fault-slip
443 data of potentially cogenetic faults were tested against their geometric and kinematic
444 compatibility and, in case the misfit angle did not exceed 30° , included into the final dataset to
445 be inverted. Faults without any recorded mineralization were also tested for their compatibility
446 with the calculated stress field. By such an approach, we identified similar stress fields for the
447 different subareas, which stresses the regional significance of the obtained stress regimes.
448 Among them, the most recurrent are shown in Figure 8 from older to younger.

449 A pure compressional stress regime with the maximum horizontal compressional stress axis (σ_1)
450 oriented NW-SE was obtained by the inversion of low angle N-dipping faults (R^2 : 2.54), that
451 invariably containing brittle fault rocks in their cores (Fig. 8a). Some of these faults are reverse,
452 while other faults recorded a later extensional reactivation.

453 Pure compression was followed by a transpressional stress regime (R^2 :1.98) with σ_1 oriented
454 NW-SE, which formed conjugate strike-slip faults, oriented NW-SE (sinistral) and E-W
455 (dextral), invariably decorated by epidote and chlorite (Fig. 8b).

456 A well recorded stress regime along the entire MNPM is an extensional stress field with σ_3
457 oriented E-W (R^2 :0.37; Fig. 8c). It generated NNE-SSW and NNW-SSE normal faults and
458 reactivated some NW-SE oblique fault planes. This stress field is recorded by fault zones mainly
459 decorated by epidote and quartz coatings, and subordinately exposing brittle fault rocks in their
460 cores.

461 The last two inverted stress regimes reactivated variably oriented faults and fractures inherited
462 from the previous deformation episodes. An extensional field with σ_3 oriented NW-SE
463 reactivated faults and fractures decorated by prehnite, quartz, zeolite and calcite coatings (R' :
464 0.52; Fig. 8d). Finally, prehnite decorated faults and/or associated with brittle fault rocks,
465 oriented NE-SW and NW-SE formed or were reactivated by an extensional stress regime with σ_3
466 oriented WNW-ESE (R' : 0.24, Fig. 8e).

467 **4.5 K-Ar Data**

468 The twelve sampled fault rocks and the altered rock were characterized mineralogically by XRD
469 analysis and dated by the K-Ar method. The XRD data are not always quantitative, because of
470 the little amount of material recovered from some of the samples (Table 2). The
471 (semi)quantitative XRD data show that in the different samples the percentage of clay minerals
472 (smectite, illite/muscovite, vermiculite, and palygorskite) increases with decreasing grain size
473 (Table 2, Fig. 9), such that the analyzed $< 0.1 \mu\text{m}$ grain size fraction of all the samples only
474 comprise clay minerals (Fig. 9). The amount of quartz, K-feldspar, plagioclase, epidote,
475 amphibole, calcite, zeolite (when present) generally decreases with decreasing grain size.
476 Generally, the chlorite content also decreases with decreasing grain size and is accompanied by
477 an increase of the smectite content. Focusing on the K-bearing phases, the finer grain size
478 fractions are invariably enriched in K-bearing clay minerals, such as illite, illite-smectite mixed
479 layers and illite-vermiculite (Fig. 9). K-feldspar is not present in all the samples, but when
480 present, its amount decreases progressively and is absent in the finest grain size fraction (Fig. 9,
481 Table 2).

482 The sixty-two new K-Ar data obtained from twelve fault gouges and one altered granitic rock
483 sample range from 848 ± 11 to 71 ± 1 Ma (Table 3). The ages are plotted on an “Age vs. grain
484 size fraction” diagram (Fig. 10). However, sample 19.049 did not contain sufficient material for
485 dating of grain sizes $<0.4 \mu\text{m}$, and samples 19.016 and 19.042E did not contain sufficient
486 material for dating of the finest ($<0.1 \mu\text{m}$) grain size fractions. The graph shows inclined curves
487 with a common trend: coarser grain size fractions have older ages while the younger ages are
488 associated with the finest fractions. The K-Ar ages of the finest grain size fractions ($< 0.1 \mu\text{m}$)
489 are 197 ± 3 , 176 ± 2 , 138 ± 3 , 130 ± 3 , 126 ± 3 (twice), 91 ± 3 , 86 ± 1 , 80 ± 1 and 71 ± 1 Ma
490 (Table 3).

491 **5. DISCUSSION**

492 The complex network of lineaments with different orientations along the MNPM, the abundance
493 of brittle structures and the high variability of mineralogical assemblages within the studied
494 faults and fractures document the polyphase brittle evolution experienced by the MNPM. Despite
495 its long and, at times, convoluted history, our systematic approach contributes to unravelling its
496 deformation and adds new temporal constraints to its polyphase evolution.

497 **5.1 Fault Rock Mineral Assemblages**

498 The sequence of mineral coatings established for the analyzed fault cataclasites constrains a
499 progressive evolution of retrograde cooling of the MNPM, presumably from the Carboniferous
500 onward. Indeed, the epidote + chlorite (+ quartz) assemblage observed in the oldest cataclasite
501 generations suggest deformation conditions close to the frictional-viscous transition in the
502 continental crust under sub-greenschist facies conditions ($T \leq 275 \text{ }^\circ\text{C}$; e.g., Wehrens et al., 2016).

503 This mineral assemblage occurs in several generations of coherent (ultra-)cataclasite (Fig. 7a),
504 suggesting recurrent events of brittle fracturing, fluid infiltration and mineralization precipitation
505 and cataclasis. The mixed mineralogical composition and prehnite-bearing cataclasites indicate
506 the continuation of brittle cataclasis and fluid infiltration processes at lower temperatures at
507 which prehnite becomes stable ($T \approx 200\text{-}250\text{ }^{\circ}\text{C}$; e.g., Malatesta et al., 2021). More than the other
508 described mineralization types, quartz- and prehnite-bearing cataclasites exhibit evidence of
509 fluid-overpressure and vein formation, preserving several generations of both comminuted and
510 euhedral quartz and prehnite crystals (Fig. 7c). Furthermore, gouge-bearing faults are commonly
511 associated with calcite and zeolite mineralizations. Unpublished clumped isotope
512 thermochronological data on calcite veins from the studied faults constrain calcite growth
513 temperatures between c. 190 and 30 $^{\circ}\text{C}$ (Tartaglia, 2021). These temperature estimates are in line
514 with zeolite formation temperatures, commonly well below 200 $^{\circ}\text{C}$ (Weisenberger and Bucher,
515 2010). The preservation of coarse euhedral crystals of both calcite and zeolite (Fig. 7f) may
516 indicate that these veins overprint and post-date the latest stages of brittle faulting and gouge
517 formation.

518 All in all, this sequence of mineralogical assemblages within the studied fault rocks supports a
519 general cooling trend throughout the brittle evolution of the MNPM. However, the lack of a clear
520 systematic relationship correlating fault strike and fault mineralogy (or sequence of mineral
521 assemblages; see Section 5.2 below, Fig. 8) does not allow us to retrieve general conclusions at
522 the scale of the MNPM. Nevertheless, the mineral paragenesis coupled with K-Ar data from fault
523 zones may provide local, but valuable constraints on the thermal and temporal evolution of the
524 brittle deformation history.

525 **5.2 Absolute Dating of MNPM Faults**

526 The obtained inclined “age vs. grain size” curves (Fig. 10) are interpreted as recording variable
527 degrees of physical mixing between authigenic and synkinematic mineral phases and inherited
528 protolithic minerals, and/or mixing of different generations of authigenic minerals (van der
529 Pluijm et al., 2001; Verdel et al., 2012; Torgersen et al., 2015; Viola et al., 2016; Vrolijk et al.,
530 2018). This interpretation is in accordance with the “Age Attractor Model” by Torgersen et al.
531 (2015), wherein the amount of authigenic and synkinematic K-bearing phases progressively
532 increases with decreasing grain size. The finer grain size fractions are, therefore, enriched in
533 synkinematic, authigenic minerals (van der Pluijm et al., 2001; Zwingmann and Mancktelow,
534 2004). However, they may still include inherited protolithic minerals or different generations of
535 K-bearing authigenic phases reworked during multiple stages of deformation, such that the ages
536 obtained from the finest grain size fractions are still to be considered as maximum ages. At the
537 same time, the ages yielded by the $<0.1 \mu\text{m}$ grain size fractions represent the best available
538 constraint on the absolute timing of the most recent faulting event recorded by the studied fault
539 rock (van der Pluijm et al., 2001; Zwingmann and Mancktelow, 2004; Torgersen et al., 2015;
540 Viola et al., 2016, Tartaglia et al., 2020). Within fault rocks, authigenic mineral formation occurs
541 because faulting is a dilatational process that enhances fluid ingress and promotes the (re)-
542 crystallization of synkinematic mineral phases. However, also post-tectonic fluid infiltration
543 within a permeable fault core may lead to authigenesis and, therefore, to dates that do not reflect
544 the age of faulting. Careful fault rock characterization is, thus, necessary to ascertain the details
545 of faulting as preserved and constrained by the microstructure of the fault rocks and to be able to
546 produce a meaningful interpretation of the obtained K-Ar ages. The fact that the fault gouges
547 sampled in this study can be generally associated with the last faulting event experienced by the
548 sampled fault zones (as indicated by our multiscale structural analysis) and that the obtained

549 ages are comparable to the timing of already known regional tectonic events (e.g., Fossen et al.,
550 2021; Osmundsen et al., 2021) suggests that the obtained K-Ar fault gouge dates are truly due to
551 synkinematic crystallization and not to post-tectonic authigenesis and or alteration due to later
552 fluid ingress.

553 The coarser grain size fractions can include protolithic, inherited minerals such as, for example,
554 K-feldspar from the host rock lithology. If several of the coarser grain size fractions from one
555 sample yield a statistically identical or similar age, it is reasonable to conclude that this age has a
556 geological significance (e.g., Viola et al., 2016; Vrolijk et al., 2018; Scheiber et al., 2019;
557 Tartaglia et al., 2020). The “Age Attractor Model” considers the intermediate grain size fractions
558 (0.1-0.4, 0.4-2 and 2-6 μm) as representing a mix of different (synkinematic and protolithic)
559 grains with varying isotopic signatures and they are, therefore, commonly devoid of a geological
560 meaning.

561 In the obtained K-Ar age vs. grain size plots, some age clusters are clearly recognizable (Fig.
562 10). The ages of the >2 μm grain size fractions of samples 19.006B, 19.049, 19.070 and of the
563 0.4-2 μm grain size fraction of sample 19.078 form a Carboniferous cluster with a mean age of
564 321 ± 8 Ma (MSWD=2.74). A Triassic/Jurassic mean age of 202 ± 6 Ma (MSWD=5.9) is
565 calculated from the age cluster constituted by the <0.1 μm finest grain size fraction of sample
566 19.006B, the coarsest 6-10 μm fraction of samples 19.007A, GT18_01 and GT18_02 and the >2
567 μm fractions of the samples 19.042E and 19.076. The ages of the intermediate fractions of
568 19.042E and GT18_01 and of the finest fractions of 19.042E and 19.078 yield a Jurassic cluster
569 with a mean age of 177 ± 12 (MSWD=0.1). The ages of the finest <0.1 μm grain size fractions of
570 samples 19.007A, 19.011, GT18_02, 19.076, and of the >0.4 μm fractions of sample 19.030A
571 define an Early Cretaceous cluster with a mean age of 122 ± 5 Ma (MSWD=3.5). Finally, three

572 <0.1 μm fraction ages are younger than 100 Ma (c. 71, 80, 86, 91 Ma), indicating an episode of
573 illite and smectite crystallization in the Late Cretaceous.

574 The finest grain size fractions, in addition to also some coarser fractions that define the age
575 clusters in the Late Triassic-Early Jurassic and Early and Late Cretaceous, are all composed of
576 authigenic K-bearing minerals, mainly k-bearing smectite, smectite-illite and illite-muscovite.
577 Therefore, they allow to constrain discrete faulting events through the synkinematic
578 (re)crystallization of K-bearing clay minerals. Two exceptions are the 6-10 μm grain size
579 fraction ages of samples 19.042A and GT18_01. In the case of sample 19.042A, the coarser
580 fraction ages show an oscillating trend, suggesting a mixing of protolithic and authigenic grains.
581 The age of its coarser grain size fraction thus remains devoid of an apparent geological meaning.
582 Sample GT18_01 was collected from the same fault zone as sample GT18_02. These two gouges
583 represent two different generations of fault rock, with GT18_01 cutting across GT18_02. Their
584 finest fraction ages (<0.1 μm), 71 Ma and 126 Ma for GT18_01 and GT18_02, respectively,
585 confirm the crosscutting relationship identified at the outcrop (Fig. 6i). Their coarser grain size
586 fractions yield the same age range (201 ± 3 , 204 ± 4 Ma), suggesting that the two gouges derive
587 from a common Triassic “protolith”, developed during a faulting or alteration event, and then
588 reactivated at different times during the subsequent brittle deformation history.

589 To summarize, the obtained K-Ar ages can be grouped into four significant clusters, which, from
590 older to younger, date to:

591 1) Carboniferous, with a mean age of 321 ± 8 Ma (MSWD=2.74);

592 2) Late Triassic-Jurassic with two distinct mean ages at 202 ± 6 Ma (MSWD=5.9) and at $177 \pm$

593 12 Ma (MSWD=0.1);

594 3) Early Cretaceous with a mean age of 122 ± 5 Ma (MSWD=3.5);

595 4) Late Cretaceous with ages between c. 91 and 71 Ma.

596 The Carboniferous cluster is constrained by only the coarser grain size fractions of the dated
597 samples. Among the fractions yielding Carboniferous ages, only the coarser grain size fraction of
598 sample 19.006B contains K-feldspar (3%) in addition to 14% illite/muscovite (Table 2, Fig. 9).

599 The K-feldspar is presumably inherited from the granitic protolith and likely records the Upper
600 Devonian-Carboniferous age of cooling during exhumation of the WGR (cf. Walsh et al., 2013).

601 The other Carboniferous ages derive from illite-smectite mixed layers or illite/muscovite, which
602 most probably formed during faulting. This Carboniferous age cluster possibly represents brittle

603 activity along the MNPM during the initial stages of the Norwegian-Greenland Sea rifting
604 (Rotevatn et al., 2018; Gernigon et al., 2020). More specifically, similar Carboniferous ages (c.

605 320 Ma) have been constrained by K-feldspar $^{39}\text{Ar}/^{40}\text{Ar}$ dating from a cataclastic granite in the
606 hanging wall of the Høybakken Detachment, NE of Hitra (Kendrick et al., 2004). This result has

607 been interpreted as recording the early stages of brittle transtension along the MTFC (Kendrick
608 et al., 2004). A very similar K-Ar illite age (312 Ma) has also been documented from the onshore

609 East Greenland rift system and interpreted as evidence of a Paleozoic rifting phase associated
610 with Carboniferous faulting (Rotevatn et al., 2018). Additionally, similar Carboniferous ages

611 have been reported in a study reporting a microstructurally controlled K-Ar dating approach by
612 Scheiber et al. (2019) and, also, by Viola et al. (2016) farther south in coastal western Norway.

613 In summary, the new results highlight early Carboniferous tectonic activity (likely transtensional,
614 e.g., Kendrick et al., 2004; Osmundsen et al., 2021) onshore the MNPM. Also, Osmundsen et al.

615 (2021) describe a “core-complex and basin” architecture for the proximal MNPM, arguing it to
616 be the result of Late Paleozoic-Early Mesozoic tectonic activity (Osmundsen et al., 2021). It is

617 plausible that these structures formed during the Carboniferous. Finally, the Late Carboniferous
618 ages are coeval also with the Late Carboniferous to Early Triassic Oslo rifting (Larsen et al.,
619 2008; Fossen et al., 2021).

620 Although during the Triassic the offshore MNPM is known to have experienced a relatively
621 quiescent period characterized by only moderate extension (Gernigon et al., 2020), our results
622 indicate some Late Triassic activity. Thus, the c. 202 Late Triassic/Early Jurassic and the 177 Ma
623 Jurassic mean ages possibly represent a distinct tectonic phase associated with crustal stretching
624 during and/or slightly after the first rifting phase reported for the NE Atlantic (Gernigon et al.,
625 2020).

626 Our 122 ± 5 Ma Early Cretaceous age may reflect onshore faulting related to the second rifting
627 phase. In fact, in the Early Cretaceous extension is known to have occurred as recorded by
628 movement along major boundary faults causing block rotation in the offshore MNPM, as well as
629 the reactivation of the MTFC (e.g., Fjellanger et al., 2005; Osmundsen and Péron-Pinvidic,
630 2018). Finally, in the Late Cretaceous, a rifting episode is recorded in the outer Møre and Vøring
631 basins. That rifting defined the pre-breakup setting with the structuring of the complex outer
632 ridges and sub-basin system (Ren et al., 2003; Péron-Pinvidic et al., 2013). However, in the Late
633 Cretaceous the proximal offshore domain is generally described as quiescent (Péron-Pinvidic et
634 al., 2013; Gernigon et al., 2020) Thus, the obtained < 100 Ma Late Cretaceous ages could reflect
635 later reactivation of onshore suitably-oriented faults, due to the later stages of rifting prior to the
636 final breakup of the NE Atlantic Ocean (Gernigon et al., 2020). Similar ages have been reported
637 by other authors onshore along the North Sea Margin (Tartaglia et al., 2020; Fossen et al., 2021
638 and references therein).

639 **5.3 Brittle Evolution through Time**

640 Our study has allowed us to define and characterize the main trends of the brittle structural grain
641 along the MNPM by remote sensing at the 1:10.000 and 1:100.000 scales and field work ground
642 truthing. In order to obtain a comprehensive evolutionary model of the brittle deformation
643 history recorded by the MNPM, the obtained structural and mineralogical data coupled with the
644 K-Ar ages interpreted as geologically meaningful, are discussed for each set of brittle structures,
645 subdivided according to their orientation. Moreover, a radial plot of structural data vs. obtained
646 K-Ar ages is shown in Figure 11 and described in the following paragraphs.

647 ***5.3.1 NE-SW and NNW-SSE Striking Faults***

648 The NE-SW and NNW-SSE striking lineaments are very common along the MNPM. By remote
649 sensing, these lineaments appear to be the oldest structural features, as they are generally cut the
650 other lineament trends (Fig. 3). Most of the NE-SW and NNW-SSE faults and fractures
651 accommodate normal and transtensional kinematics. They are decorated by quartz, epidote,
652 chlorite, calcite, prehnite, and commonly contain fault rocks such as fine-grained cataclasite and
653 distinct layers of fault gouge.

654 The complex arrangement of cataclastic layers and mineral coatings (see Fig. 5e) derives mainly
655 from this set of faults. They mainly contain epidote cataclasites dissected by quartz veins and by
656 later thin prehnite cataclastic layers (Fig. 7).

657 Four dated samples belong to this set of lineaments (Fig. 11). Their K-Ar ages of the <0.1 μ m
658 grain size fractions are Early and Late Cretaceous. The coarser grain size fractions consisting of
659 illite-smectite mixed layers and illite yield Carboniferous, Triassic, and Early Cretaceous ages
660 (Fig. 11). These radiometric data, the widespread occurrence in the field of the NE-SW and

661 NNW-SSE striking faults and their variability in mineralogical coatings suggest that this set of
662 faults recorded several brittle deformation events affecting the margin from the Carboniferous
663 down to the last documented reactivation in the Late Cretaceous.

664 The NE-SW and NNW-SSE striking faults are geometrically and kinematically compatible with
665 a transpressional stress regime with σ_1 oriented NW-SE, which formed epidote and chlorite
666 mineralizations. Additionally, they could be reactivated during the NW-SE extensional stress
667 regime characterized by prehnite, quartz and calcite mineralizations. Some of these faults may
668 thus have exploited a preexisting Caledonian fabric, prior to their own repeated reactivation and
669 reworking during rifting due to their geometrical compatibility with the extensional stress fields.

670 *5.3.2 ENE-WSW Striking Faults*

671 The ENE-WSW striking lineaments are parallel to the northern part of the MNPM coastline,
672 specifically the trend of the MTFC. They are particularly abundant on the island of Hitra, where
673 the main strand of the MTFC, the Hitra-Snåsa Fault, is exposed (Redfield et al., 2005, 2009).
674 Remote sensing analysis indicates that this trend generally cut all the other detected trends, in
675 particular in the area of Hitra. The radiometric dating of our selected ENE-WSW striking fault
676 zones (Fig. 11) indicates that these faults have recorded brittle events along the MNPM ranging
677 from the Late Triassic to the Late Cretaceous. The oldest radiometric age of this set of faults is
678 Late Triassic (201 ± 3 Ma, GT18_01, Table 3).

679 The ENE-WSW striking faults have a variable dip and accommodate strike-slip (mainly sinistral)
680 or normal kinematics. They are mainly associated with chlorite and iron oxide mineralizations
681 and locally with epidote and prehnite. These faults are geometrically compatible with the E-W
682 and NE-SW extensional regimes experienced by the MNPM (Fig. 8). The high angle ENE-WSW

683 striking faults are generally localized along the MTFC mylonitic foliation. According to their
684 radiometric ages, these studied faults record only the evidence of the brittle deformation due to
685 the multiphase rifting of the NE Atlantic. However, pre-Triassic activity along the ENE-WSW
686 faults, parallel to the MTFC, cannot be excluded. Potential evidence is to be found in the
687 presented crosscutting relationships at the macroscale, the absolute dating of strands of the
688 MTFC as Devonian and Late Carboniferous-Early Permian (e.g., Kendrick et al., 2004; Sherlock
689 et al. 2004), as well as in the offshore ENE-WSW-trending lineaments bounding pre-middle
690 Triassic strata in the Froan basin (Osmundsen et al. 2021). Indeed, ENE-WSW striking fault
691 patterns in the area have been previously interpreted as Riedel-shears related the main branches
692 of the MTFC (Hitra- Snåsa and Verran Faults), resulting from sinistral strike-slip tectonics
693 during the late Devonian (e.g., Grønlie and Roberts, 1989). Thus, the inception of activity along
694 the ENE-WSW trending faults is likely older than the obtained K-Ar ages of the studied gouges,
695 and may have started already in the Devonian (Doré et al., 1999; Kendrick et al., 2004; Faleide et
696 al., 2008; Gernigon et al., 2020).

697 *5.3.3 WNW-ESE and E-W Striking Faults*

698 As shown by the remote sensing analysis, WNW-ESE and E-W striking lineaments show mutual
699 crosscutting relationship with the other trends (Fig. 3). In the field, WNW-ESE trending faults
700 record dip-slip or strike-slip kinematics. The WNW-ESE and E-W striking faults are mainly
701 associated with iron oxide mineralizations, incohesive brittle fault rocks, quartz, and calcite
702 mineral coatings.

703 Three faults belonging to this lineament set have been dated (Fig. 11). Samples 19.006B and
704 19.078 were taken from two low angle dip-slip faults. These faults are situated in the northern-

705 and southernmost studied segments of the MNPM where they cut across a granite (19.006B) and
706 a gabbro (19.078). From a structural point of view, these faults are compatible with a
707 compressional stress regime (NW-SE oriented σ_1 , Fig. 1a). This compressional stress field may
708 represent the stress field during the Caledonian orogeny. However, E-W and NW-SE thrusts and
709 folds reported by previous authors in the Sunnfjord region and on Hitra and Frøya have been
710 associated with a N-S shortening related to a contractional/transpressional event in the Late
711 Devonian-Early Carboniferous, referred to as the “Solundian phase” (Braathen, 1999;
712 Osmundsen and Andersen, 2001; Sturt and Braathen, 2001). It cannot be excluded that this set of
713 faults originated during the Solundian phase.

714 K-Ar data of the samples 19.006B and 19.078 indicate that these two faults record a
715 Carboniferous thermal/faulting event, c. 320 Ma ago, and later reactivation during Triassic-
716 Jurassic rifting in an overall E-W extensional regime (Fig. 11). Therefore, they probably
717 experienced a compressional event (the Caledonian or the Solundian phase) and were later
718 reactivated in an extensional manner during Carboniferous and Triassic-Jurassic rifting.

719 Fault gouge 19.016 was collected from a moderately dipping normal fault. It records a different,
720 younger brittle history which documents an Early Cretaceous origin and bears testimony of a
721 Late Cretaceous reactivation (<97 Ma, Fig. 11) during a NW-SE extensional phase.

722 **5.4 Regional Implications**

723 The mapping, characterization and dating of the brittle structural framework along the MNPM
724 allows us to place new temporal constraints onto the evolution of the margin, as summarized in
725 Figure 12. This figure shows the obtained inverted paleostress fields through time and the strike
726 of the brittle structures that accommodated faulting during the constrained deformation phases.

727 Out of the many structures preserved along the MNPM, some sets of faults in the Smøla-Hitra
728 batholith may represent the brittle expression of Caledonian deformation. These sets of faults
729 (Fig. 8a-b) show similar orientations and kinematics as along the North Sea Margin, where
730 $^{39}\text{Ar}/^{40}\text{Ar}$ ages of ca. 450 and 435 Ma were obtained from synkinematic micas (Scheiber et al.,
731 2016). In fact, paleostress inversion of our fault sets resulted in a pure compressional stress
732 regime with σ_1 oriented NW-SE and a transpressional stress regime ($R':1.98$) with σ_1 oriented
733 NW-SE, both of which are compatible with the Caledonian tectonic stresses reconstructed by
734 other studies (e.g., Séranne, 1992; Scheiber and Viola, 2018; Fig. 12). Another possible
735 interpretation (not illustrated in Fig. 12) is that the compressional stress regime could be related
736 to the Late Devonian-Early Carboniferous N-S-oriented contractional/transpressional Solundian
737 phase, which generated the E-W-trending folds and thrusts in the Solund Basin (North Sea) and
738 on Hitra and Frøya (Braathen, 1999; Osmundsen and Andersen, 2001; Sturt and Braathen, 2001).
739 The same compressional structures on Hitra and Frøya have been also previously interpreted as
740 due to Early Devonian transtensional activity of the MTFC, which would have caused local NW-
741 SE compression, coeval to slightly younger than the emplacement of the Smøla-Hitra batholith
742 (e.g., Krabbendam and Dewey, 1998).

743 In the area, the MTFC was possibly active in pre-Devonian times accommodating dextral
744 transpression (Séranne, 1992) with the formation of pervasive ductile fabrics that were
745 reactivated during the formation of the MNPM. The MNPM accommodated an extensional
746 history that started in the Early Devonian with the collapse of the Scandinavian Caledonides
747 (Doré et al., 1999; Faleide et al., 2008; Gernigon et al., 2020). A few structural observations
748 onshore East Greenland support the idea that the first rifting phase affecting the NE Atlantic was
749 in the mid-Carboniferous, but this model remains poorly constrained in the MNPM (e.g.,

750 Stemmerik et al., 2000; Rotevatn et al., 2018). However, our K-Ar ages from NE-SW and NW-
751 SE striking faults support rift initiation along the MNPM in the Carboniferous. This is also in
752 agreement with the dating of Late Carboniferous-Early Permian pseudotachylytes from the
753 Hitra-Snåsa Fault (Sherlock et al., 2004) and of Late Carboniferous cataclasites along the
754 Høybakken Detachment (Kendrick et al., 2004). It can be argued that this Carboniferous event is
755 responsible for the formation of sedimentary basins in the proximal offshore domain of the
756 MNPM, now possibly ascribed to the Late Paleozoic-Early Mesozoic (Osmundsen et al., 2021).
757 More evidence of the Late Carboniferous faulting along the MNPM may have been partly or
758 fully obliterated by later slip deformation and the reworking of originally Carboniferous faults
759 (Fossen et al., 2021). The main rifting phases experienced by the MNPM recorded in the
760 offshore domain are reported in the Permo-Triassic and in the Late Jurassic-Cretaceous
761 (Gernigon et al., 2020). Our data, however, clearly indicate that faulting took place already in the
762 Late Triassic-Early Jurassic and in two pulses during the Cretaceous (Figs. 10, 11, 12). These
763 rifting phases are well documented in the North Sea margin by a rapidly growing database of
764 absolute K-Ar deformation ages (Ksienzyk et al. 2016; Viola et al. 2016; Scheiber and Viola,
765 2018; Scheiber et al. 2019; Fossen et al., 2021). According to our new dataset, the onshore
766 MNPM recorded the effects of two slightly younger rifting phases in comparison to the rifting
767 phases established in the North Sea margin which may be a result of the northward propagation
768 of rifting (Ren et al., 2003; Gernigon et al., 2020; Zastrozhnov et al., 2020).

769 During the Triassic-Jurassic faulting event recorded by the onshore MNPM, E-W crustal
770 stretching occurred, as well documented by previous studies from the North Sea and the MNPM
771 (Gómez et al., 2004; Scheiber and Viola, 2018; Gernigon et al., 2020). This rifting phase
772 reactivated older structures and formed epidote-rich cataclasites and quartz veins associated with

773 NE-SW, NW-SE, and ENE-WSW-striking faults (Figs. 11, 12). Coeval alteration/deep
774 weathering occurred, likely due to fluid circulation along brittle structures dissecting tilted
775 blocks during rifting (e.g., Fredin et al., 2017).

776 The Cretaceous events along the MNPM have played a key role in the reactivation of older fault
777 zones, forming extensive low temperature mineralizations along the exploited fault planes, such
778 as calcite, prehnite and zeolite. The Early Cretaceous event recorded onshore correlates well with
779 the formation of the offshore Møre and Vøring sedimentary basins and the activity of offshore
780 major boundary faults (Osmundsen and Péron-Pinvidic, 2018; Zastrozhnov et al., 2018, 2020).
781 The Late Cretaceous faulting event could also reflect the reactivation of suitably oriented faults
782 during later stages of rifting prior to the final breakup of the NE Atlantic Ocean (Gernigon et al.,
783 2020). Although similar ages have been reported from onshore faults along the North Sea
784 Margin (Tartaglia et al., 2020; Fossen et al., 2021 and references therein), the Late Cretaceous
785 faulting activity in the MNPM is only documented in the distal offshore domain (e.g., Péron-
786 Pinvidic et al., 2013; Gernigon et al., 2020).

787 In the offshore domain, the MTFC separates the Jurassic–Cretaceous North Sea basins from the
788 Cretaceous basins of the MNPM (Redfield et al., 2005), suggesting important offshore
789 Cretaceous activity (Péron-Pinvidic et al., 2013). The faults parallel to the onshore MTFC
790 recorded the rifting events from the Early Jurassic to the Late Cretaceous, supporting the active
791 tectonic role of the multiply reactivated MTFC during rifting and the development of the margin.

792

6. CONCLUSIONS

793 The MNPM derives from a complex polyphase faulting history. Its evolution was influenced by
794 the effects of the spatial orientation of old, inherited structures, such as the Caledonian orogen-
795 parallel structural grain and the MTFC (*cf.* Osmundsen et al., 2006; Phillips et al., 2019; Schiffer
796 et al., 2019). By means of remote sensing lineament analysis, field work, microstructural
797 analysis, paleostress inversion, mineralogical characterization and K-Ar dating of illite separated
798 from selected fault zones, six tectonic events have been identified. From older to younger these
799 are (Fig. 12):

- 800 i) Paleozoic NE-SW compression forming WNW-ESE-trending and N(NE)-dipping
801 low-angle thrust faults;
- 802 ii) Paleozoic transpression with σ_1 oriented NW-SE forming conjugate NW-SE sinistral
803 and E-W dextral strike slip faults;
- 804 iii) A Carboniferous faulting event associated with rift initiation forming NW-SE and
805 NE-SW, variably dipping, faults.
- 806 iv) Late Triassic-Early Jurassic E-W extension at c. 202 and 177 Ma forming epidote and
807 quartz-coated, N-S striking, generally normal faults, and coeval alteration of the host
808 rock due to faulting-enhanced fluid circulation;
- 809 v) Early Cretaceous NW-SE extension representing the second rifting stage documented
810 from the offshore domain of the MNPM, leading to the formation of normal,
811 transtensional NE-SW and N-S striking faults;

812 vi) Late Cretaceous (K-Ar ages of c. 71, 80, 86, 91 Ma) extension reactivating suitably
813 oriented pre-existing faults, with extensive synkinematic precipitation of low
814 temperature coatings (prehnite, zeolite).

815 The lack of a preserved sedimentary cover and the long brittle evolution of the MNPM make the
816 reconstruction of the tectonic evolution of the margin challenging. However, our new
817 radiometric ages fill in the gap of absolute dating of fault activity along the MNPM.

818 Fault dating coupled with multiscale structural analysis has been shown to be key to the study of
819 the polyphase history of the margin. Finally, the applied workflow may assist the interpretation
820 of the structural framework of the offshore domain, leading to high-resolution structural models
821 and better exploration predictive tools.

822

823 Table and Figure Captions

824 **Table 1-** *Field data of sampled and dated fault zones grouped into three sets according to their*
825 *strike (WNW-ESE, ENE-WNW, NE-SW). (N: Normal fault, R: Reverse fault; S: Sinistral fault; X:*
826 *Unknown sense of slip)*

827

828 **Table 2-** *XRD data of the dated samples separated in five grain size fractions (<0.1, 0.1-0.4,*
829 *0.4-2, 2-6 and 6-10 μm). (Qtz: quartz, Kfs: K-feldspar, Plg: plagioclase; Cc: calcite, Ep:*
830 *epidote; Amp: amphibole; Px: pyroxene; Ill: illite; Ms: muscovite; Palyg: palygorskite; Kln:*

831 *kaolinite; Chl: chlorite; Sm: smectite; Verm: vermiculite; Lepid: lepidocrocite; Zeo: zeolite; Stb:*
832 *stilbite; Ap: apatite; Anl: analcime; GOF: Goodness of Fit).*

833

834 **Table 3** - *K-Ar data of 12 fault gouge and 1 altered rock (19.042E) samples.*

835

836 **Figure 1** - *Simplified geological map of the on- and offshore mid-Norwegian Passive Margin.*
837 *Drawn after Faleide et al. (2008) and Slagstad et al. (2011). (MTFC: Møre-Trøndelag Fault*
838 *Complex; WGR: Western Gneiss Region).*

839

840 **Figure 2** - *Map showing hillshaded LiDAR DEMs and bedrock lineaments from the coastal*
841 *MNPM color-coded according to their strike (as shown in the associated rose diagrams). (a)*
842 *Lineaments from the entire MNPN, mapped at the 1:100.000 scale. (b) Lineaments from the*
843 *northern MNPM (Hitra, Frøya and Smøla), mapped at the 1:10.000 scale, (c) Lineaments from*
844 *the southern segment of the MNPM, mapped at the 1:10.000 scale.*

845

846 **Figure 3** – *Maps showing examples of manually mapped lineaments at two different scales on*
847 *(a) northeastern Hitra and (b) western Smøla. The black circles highlight mapped crosscutting*
848 *relationships between variably oriented lineaments.*

849

850 **Figure 4** - (a) Map of the investigated structural sites. Dark grey dots indicate the location of the
851 dated fault gouge samples (Table 1); (b) Rose diagrams showing the strike orientation of fault
852 and fracture zones, sorted according to the main decorating mineral phase.

853

854 **Figure 5** - Photographs of representative fault zones. (a) ENE-WSW outcrop of a strand of the
855 MTFC (Jøsnøya, Hitra island); (b) same outcrop as in (a), this strand of the MTFC is cut by
856 high-angle normal faults, oriented NNW-SSE; (c) c. 10 m thick fault zone, with the principal slip
857 surface oriented 322/76 and a (d) fault core containing cataclasite and gouge (Hitra island); (e)
858 E-dipping fault plane with a 10 cm-thick core containing fine-grained quartz and epidote cut by
859 milky white prehnite veins (pointed out by yellow arrows, Hitra island); (f) NE-SW
860 transtensional fault plane with calcite and zeolite coating (Runde island).

861

862 **Figure 6** – Sampled and dated fault zones divided into three sets according to their strike:
863 WNW–ESE and E-W (a - c); ENE – WSW (d – –); –NE-SSW (j - m). Red lines highlight principal
864 slip surfaces, and the yellow squares indicate the sampled area. More details in the text and Table
865 1.

866

867 **Figure 7** - Microphotographs of representative fault rocks from Hitra island. (a) Granitic
868 protocataclasite characterized by pervasive fracturing and epidote mineralization. The host rock
869 is partially altered, as indicated by the occurrence of chlorite (Chl) and saussuritic plagioclase
870 (Plg). Several generations of progressively finer-grained epidote mineralizations can be

871 identified (Ep₁, Ep₂, Ep₃). (b) Quartz (Qtz)-rich cataclasite containing a clast of fine-grained
872 epidote-bearing cataclasite (Ep₃). The quartz-rich layer contains a younger quartz vein (Qtz₂)
873 localized at the contact between an epidote-bearing cataclastic layer (top) and the quartz-rich
874 cataclasite (bottom, Qtz₁). Note the occurrence of multiple layers of variably sized epidote
875 cataclasites (Ep₂, Ep₁). (c) Prehnite (Prh) vein and cataclastic layer (top) cutting across an older
876 quartz-epidote-bearing cataclasite. This cataclasite contains clasts of epidote-bearing layers
877 embedded in a coarse-grained, quartz-rich vein/cataclastic layer. (d) Prehnite vein disrupting the
878 layered structure of older epidote- and quartz-bearing cataclasite. (e) Fine-grained, cataclasite
879 containing coarse grained clasts of epidote-, quartz and prehnite-bearing cataclasite embedded in
880 a fine-grained matrix. (f) Calcite (Cc) vein associated with euhedral zeolite crystals. (a, c-e)
881 Sample 19'009A, N 63°'1.079' E 08°48.838'; (b) Sample 1'.002B, N 63°'8.868' E 08°47.255'; (f)
882 Sample '9.006, N 63°'8.339' E 08°40.505'.

883

884 **Figure 8** - *Paleostress tensors computed from the inversion of fault-slip data from the onshore*
885 *MNPM: stereonets (a) to (e) are tentatively sorted from oldest to youngest (see text for*
886 *explanation). Stereoplots are Schmidt, lower hemisphere projections. (bfr: brittle fault rocks;*
887 *Ep: epidote; Chl: chlorite; Qz: quartz; Prh: Prehnite; Cc: Calcite).*

888

889 **Figure 9** - *Pie charts of mineral concentrations [wt%] derived from XRD of each of the dated*
890 *fractions of three representative fault gouges of different ages (i.e., ages of the finest grain size*
891 *fractions): 19.070 (80 ± 1 Ma), 19.042A (86 ± 1 Ma), 19.006B (197 ± 3 Ma).*

892

893 **Figure 10** - K-Ar age vs. grain size fraction diagram. Each point in the diagram represents the
894 age of a specific grain size fraction. The gray bars indicate deduced tectonic events and are also
895 reported in Figs. 11 and 12.

896

897 **Figure 11** - *Radial diagram of fault gouge and altered rock K-Ar ages plotted as a function of*
898 *the fault plane orientation. (G.s.f.: grain size fraction). Only K-Ar ages interpreted as*
899 *geologically meaningful (cf. text and figure 10) are shown.*

900

901 **Figure 12** - *Schematic summary of the obtained paleostress fields through time and strike of*
902 *active brittle structures. The stereoplots represent the inverted paleostress regimes in present-*
903 *day coordinates. The white, grey, and black arrows in the stereoplots represent the orientation of*
904 *σ_1 , σ_2 and σ_3 , respectively.*

905

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916

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Figure 1

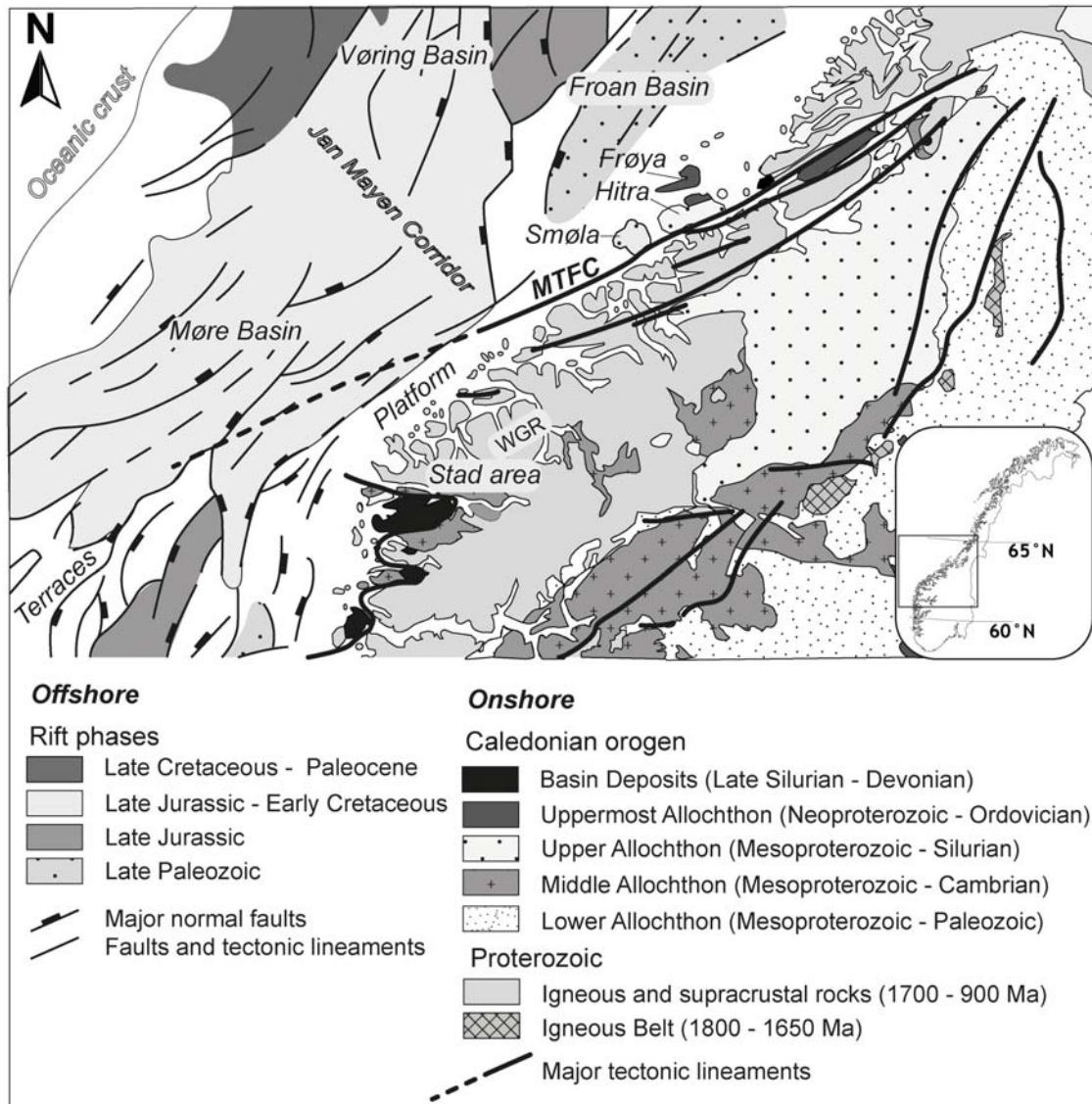


Figure 2

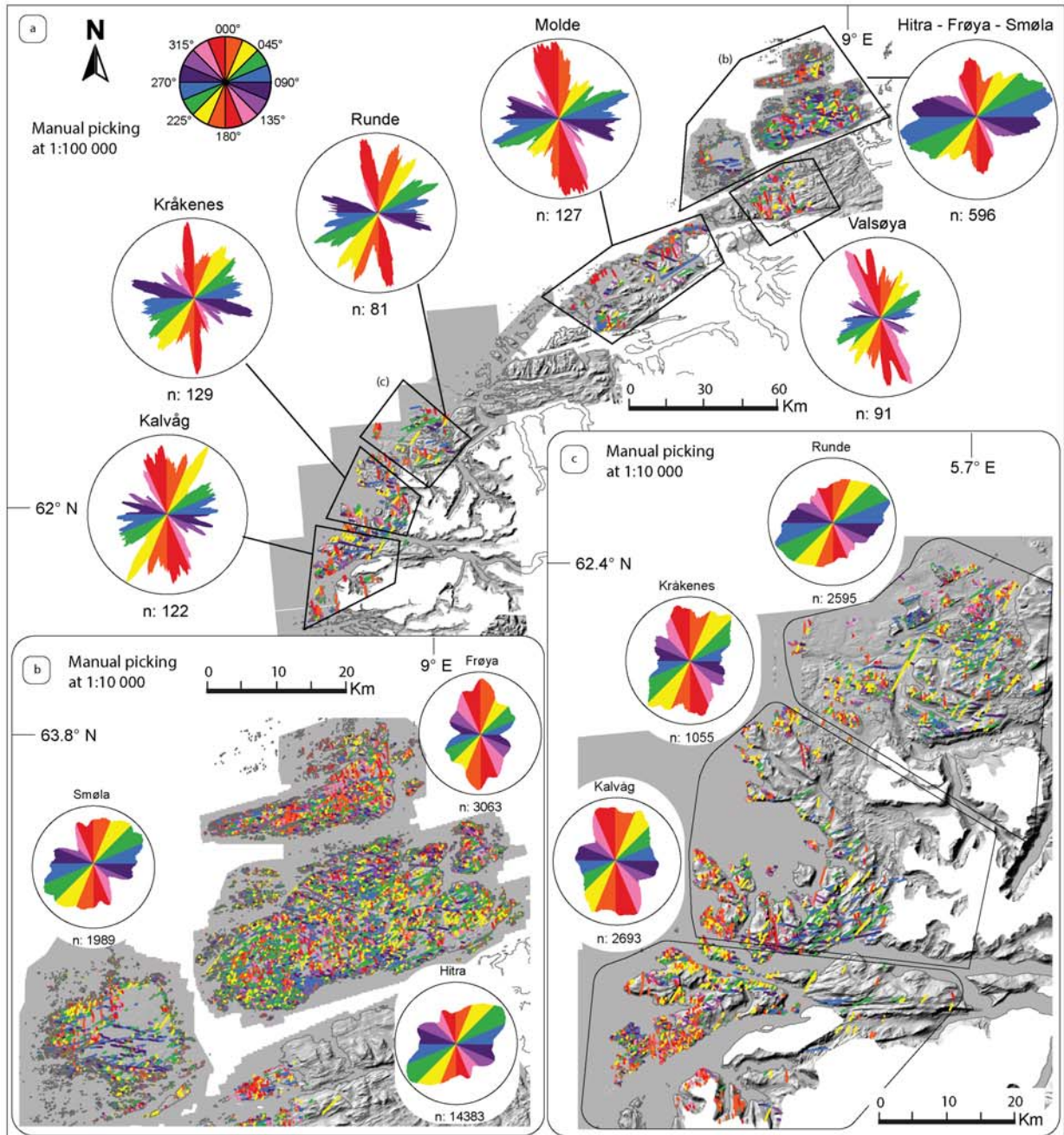


Figure 3

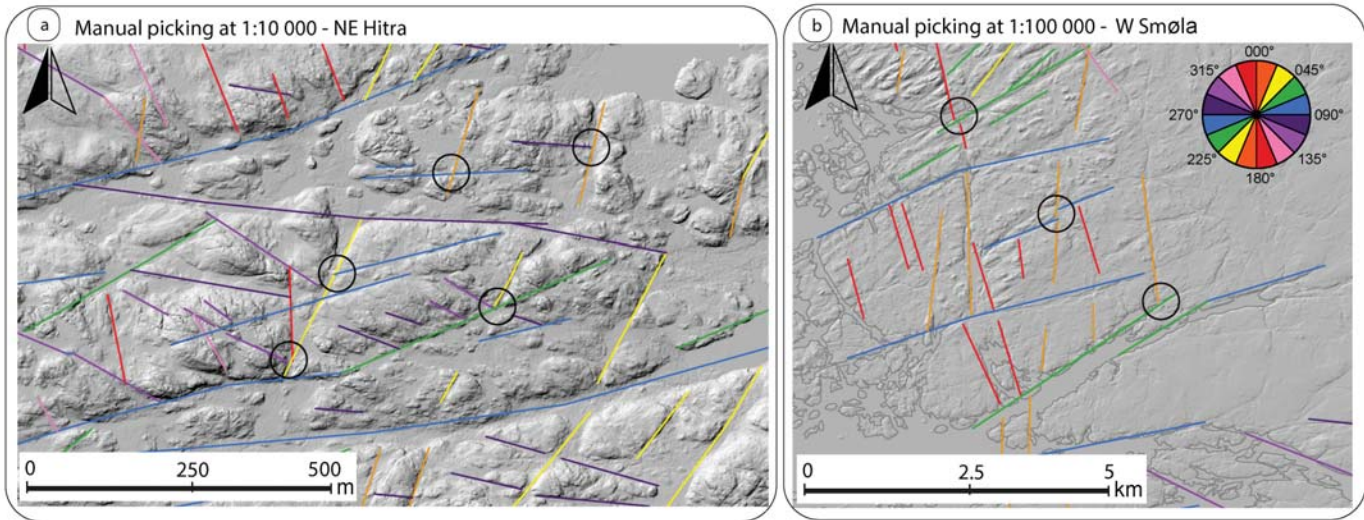


Figure 4

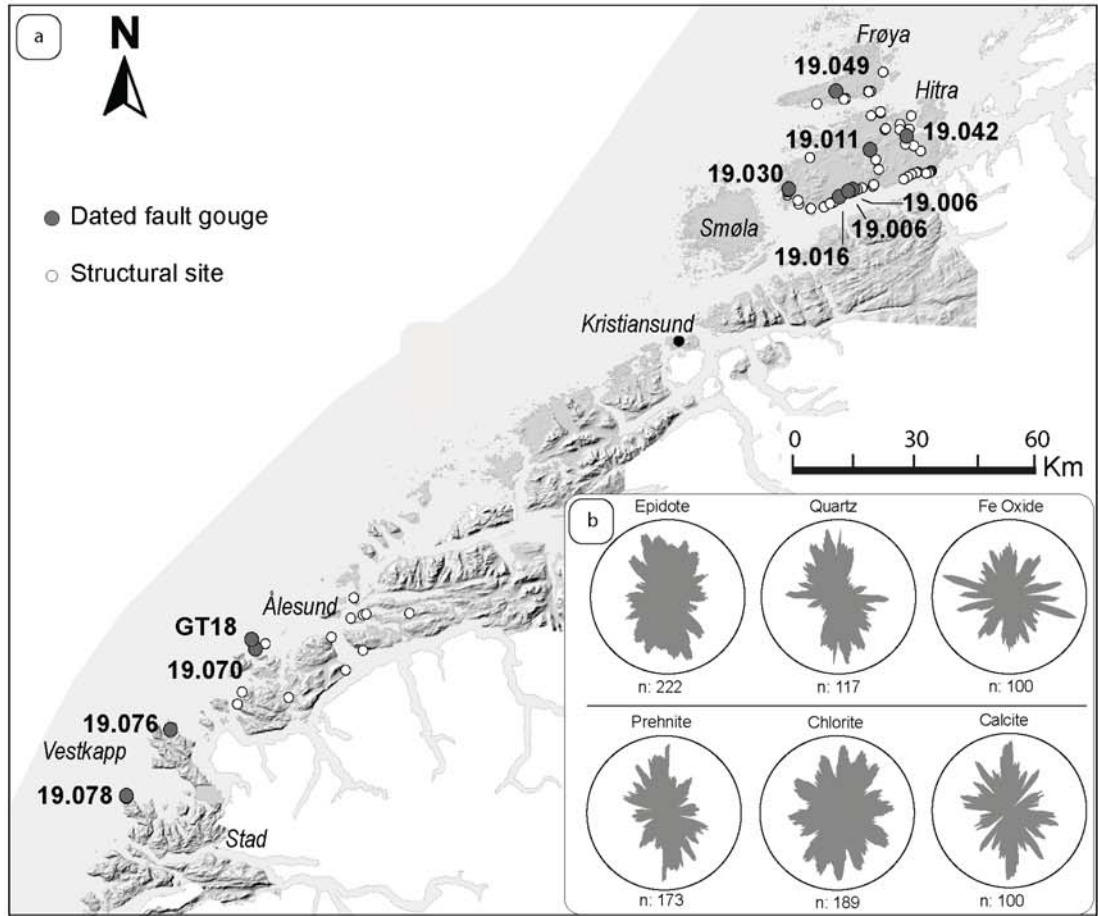


Figure 5

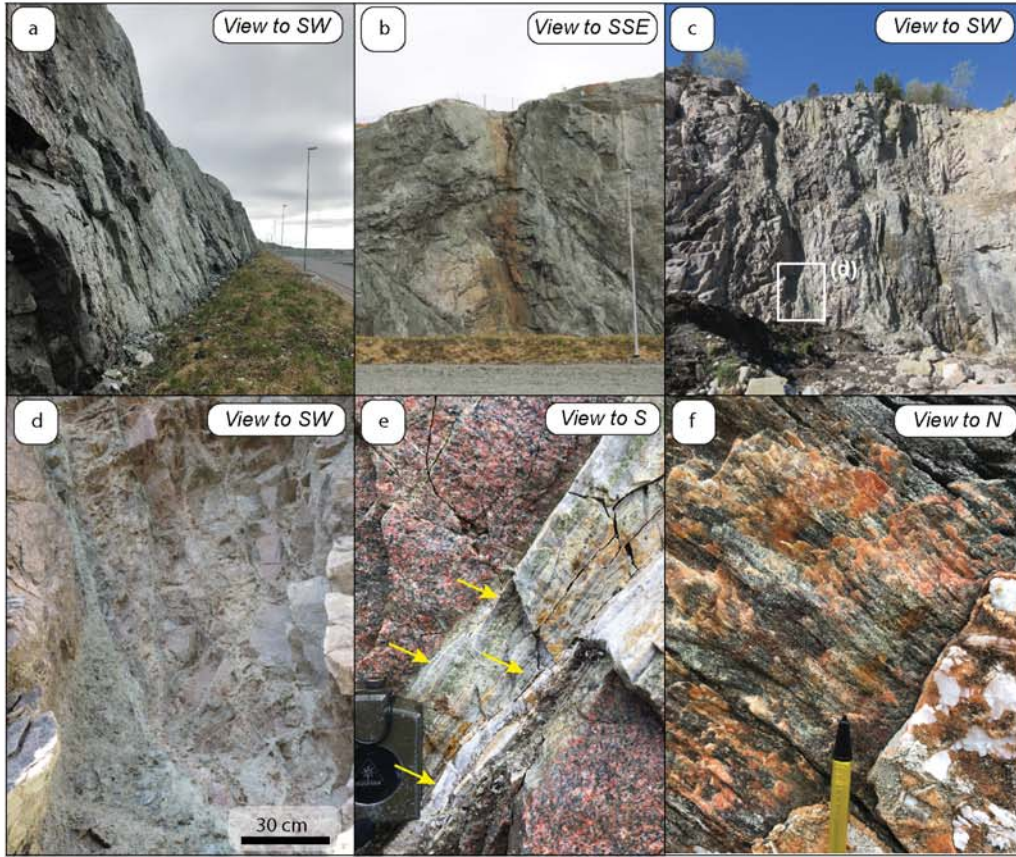
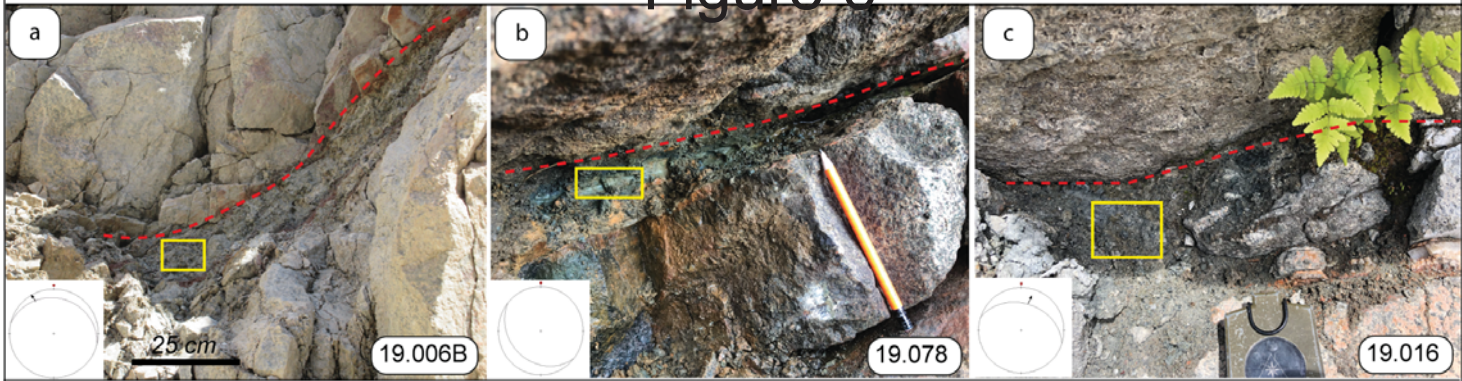
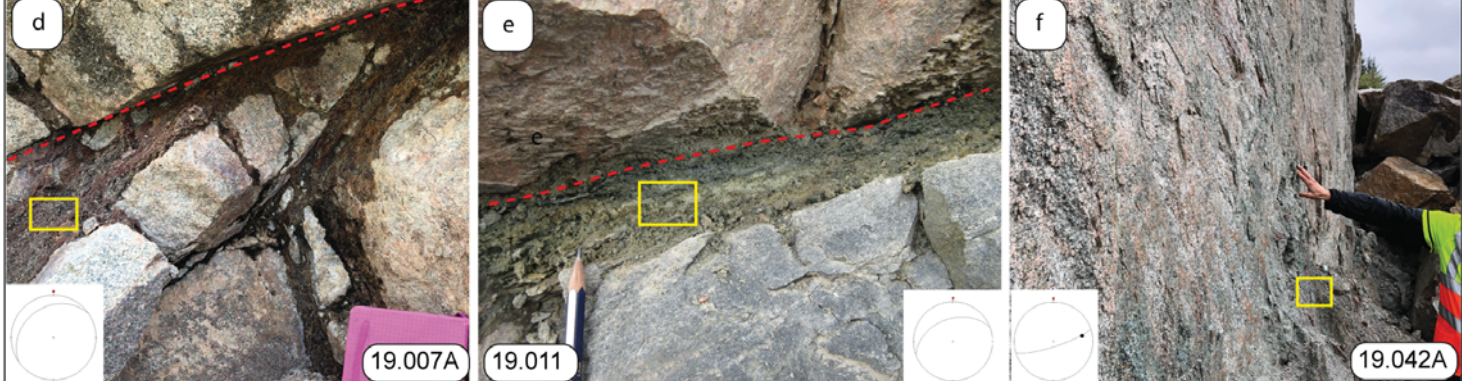


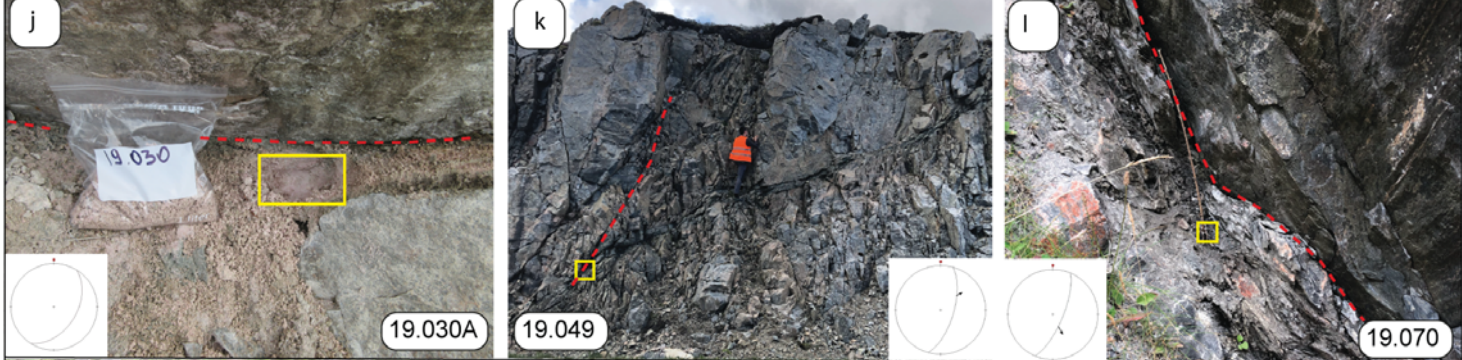
Figure 6



ENE - WSW



NNE - SSW



Key
- - - Principal Slip Surface
□ Sampling area

Figure 7

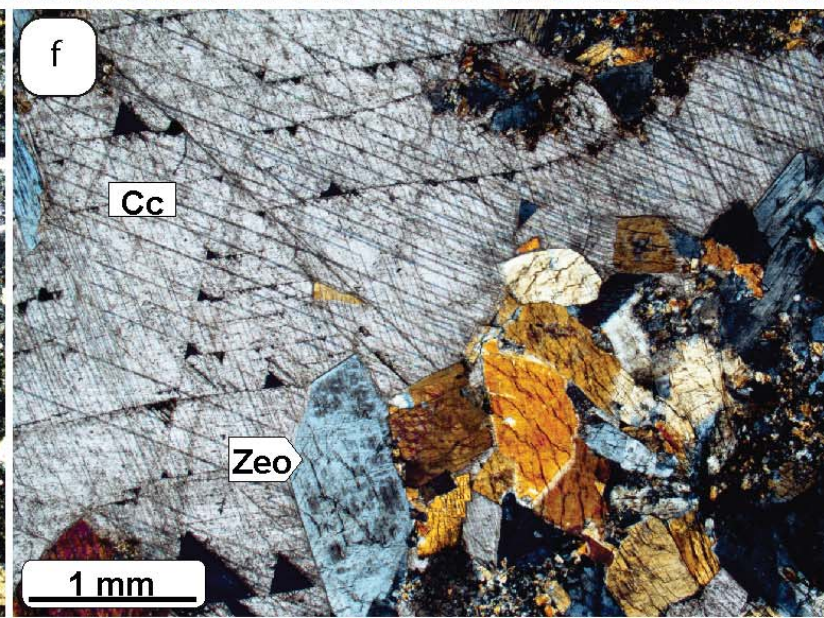
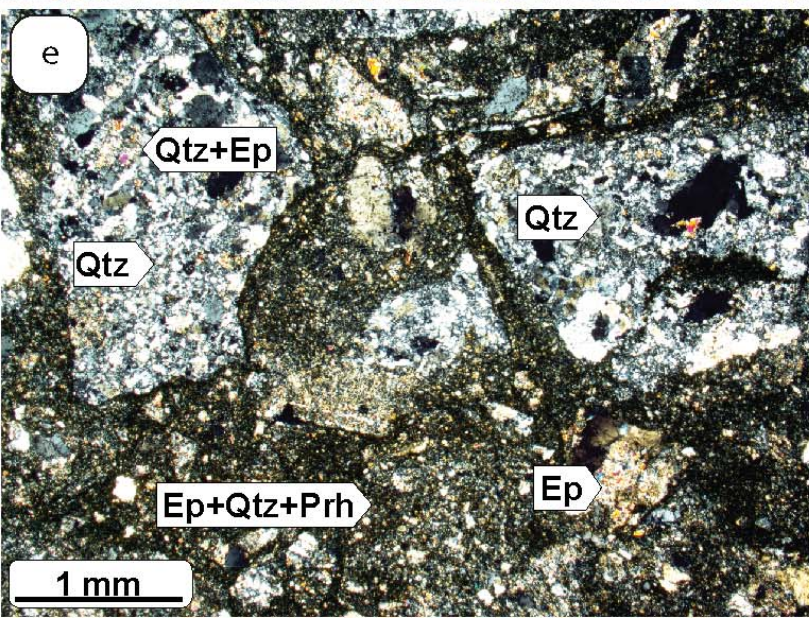
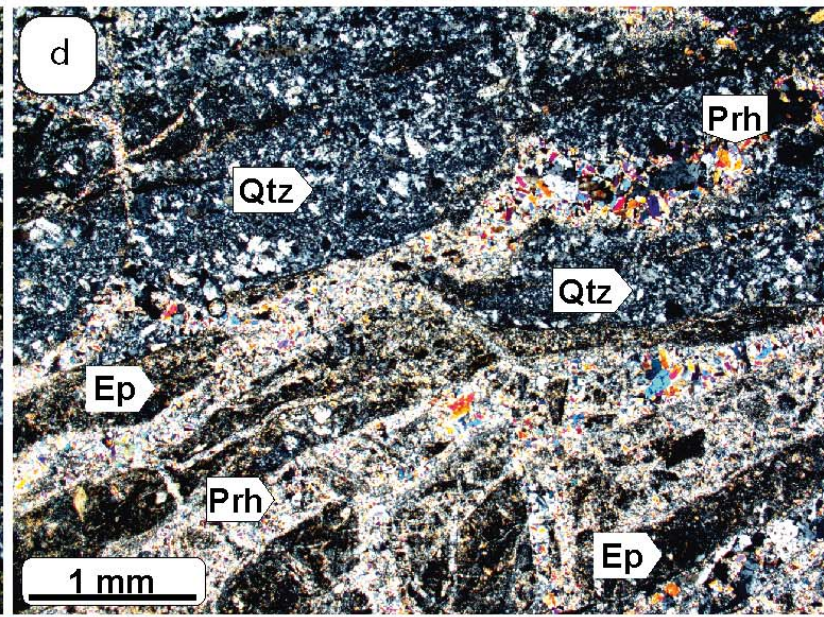
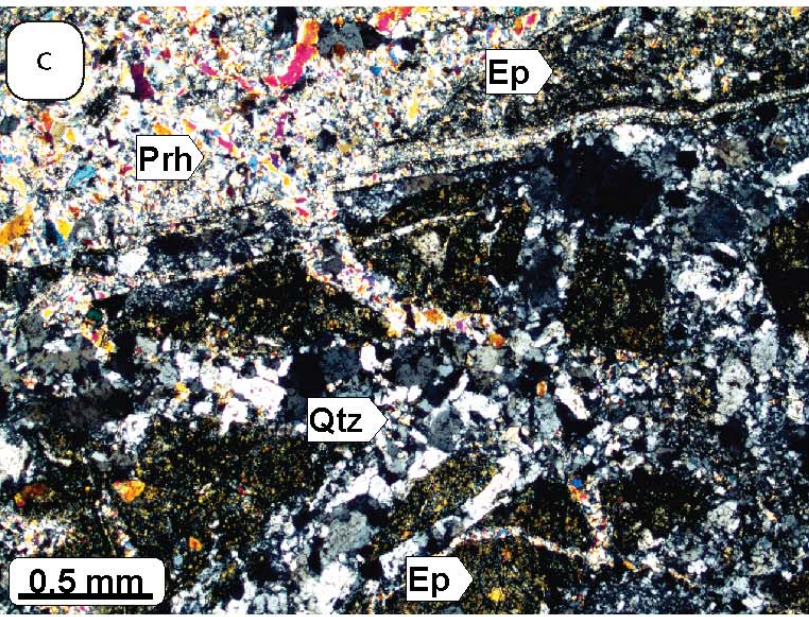
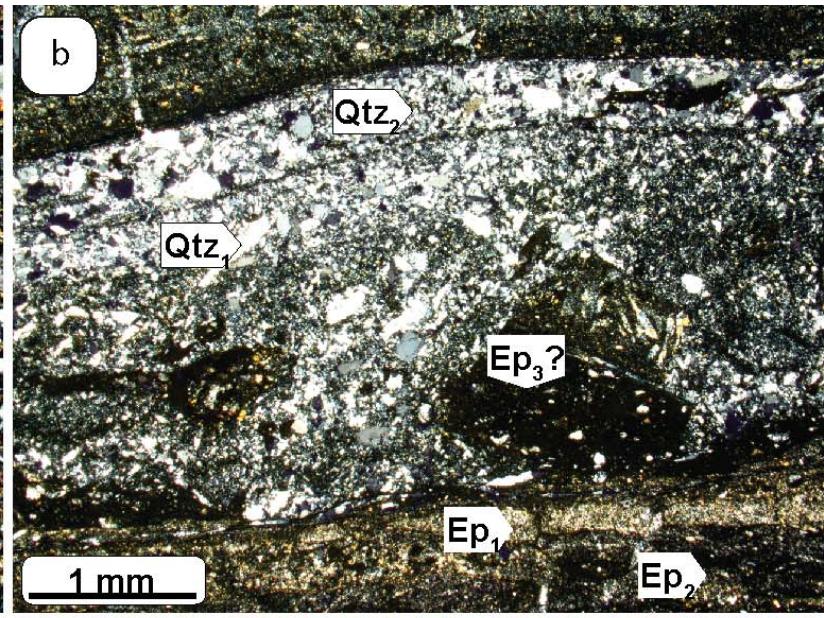
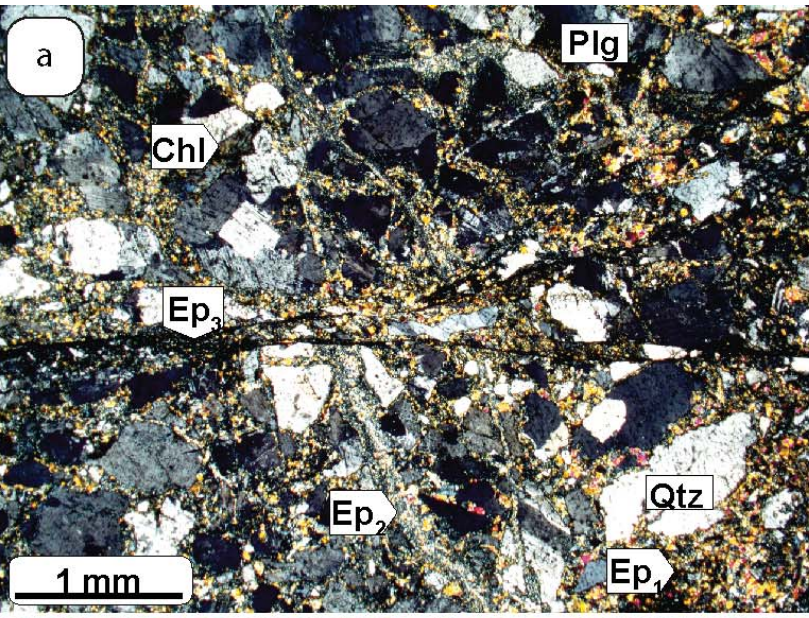


Figure 8

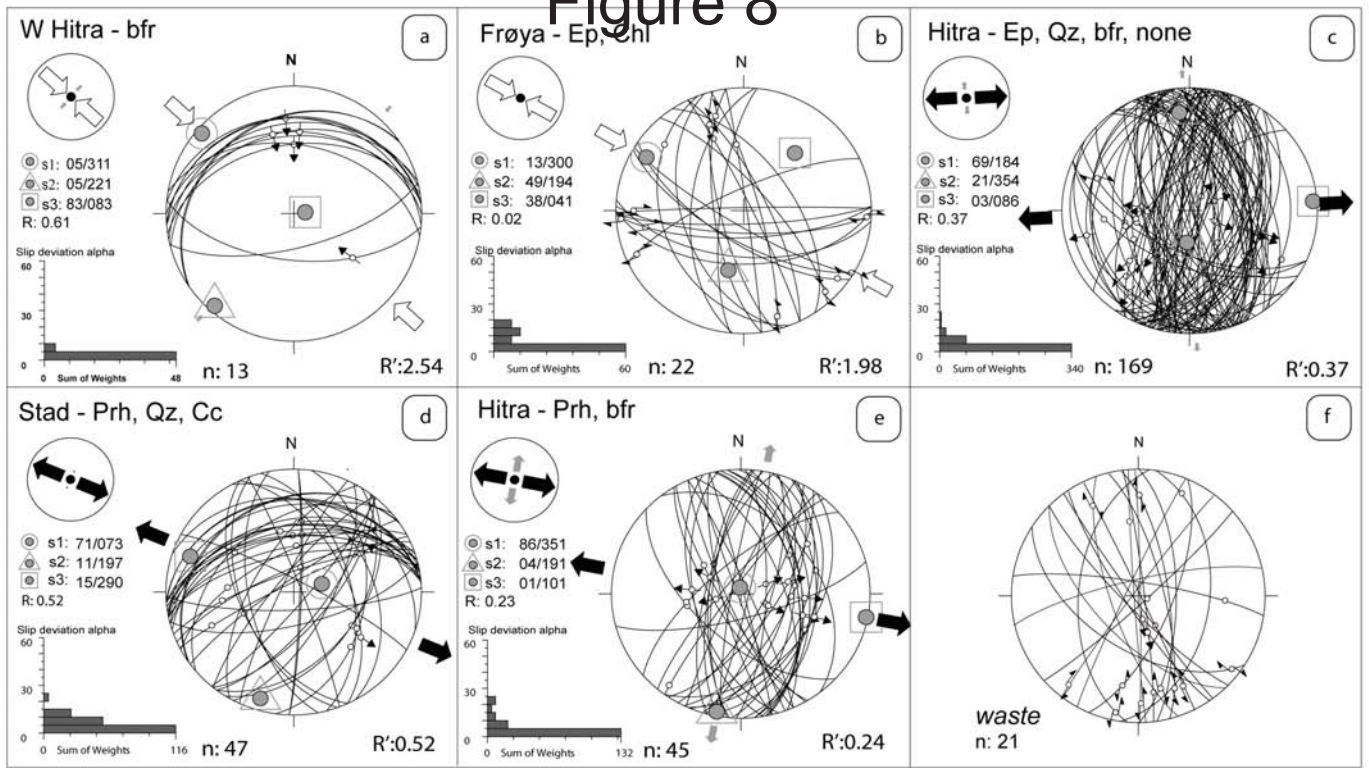


Figure 9

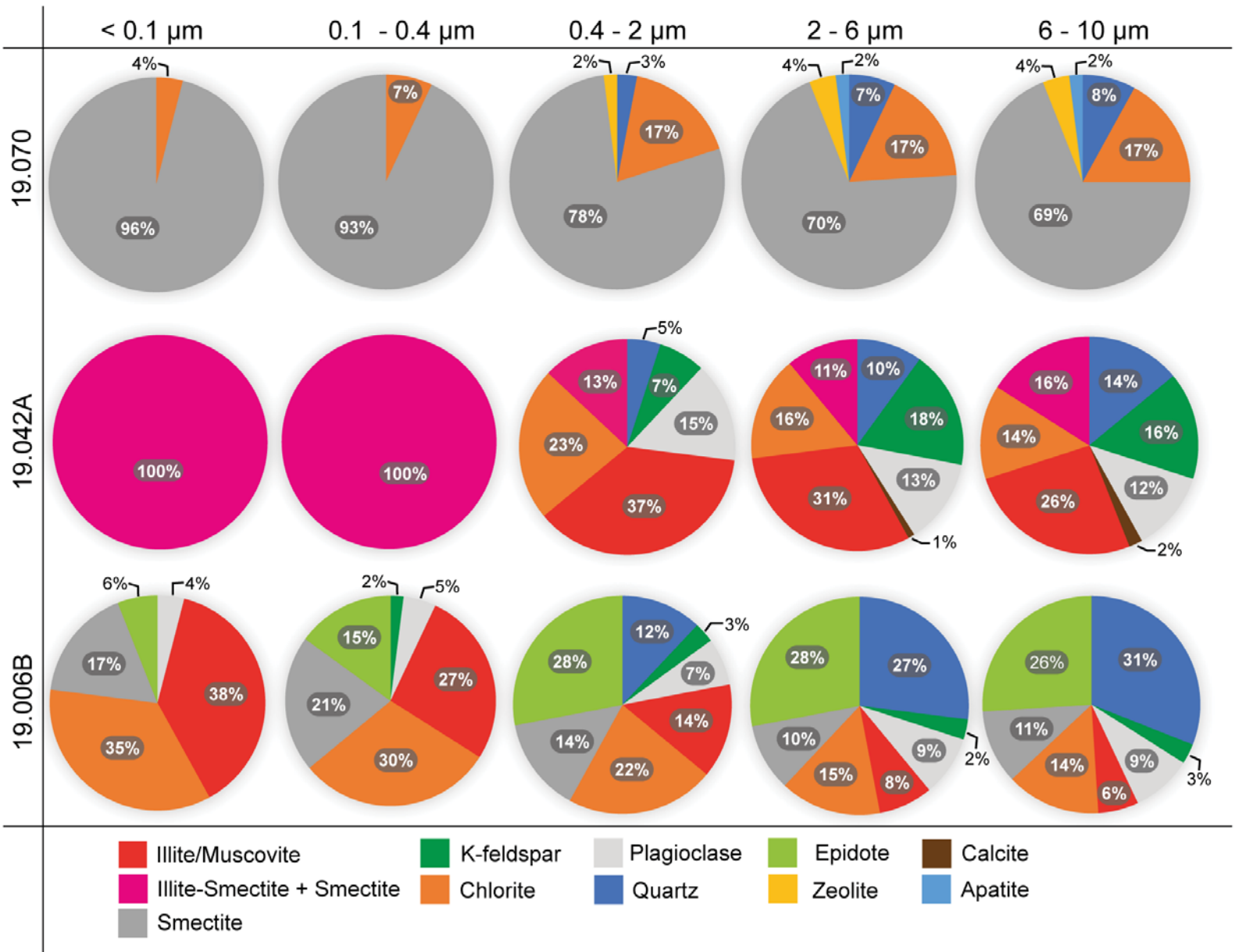


Figure 10

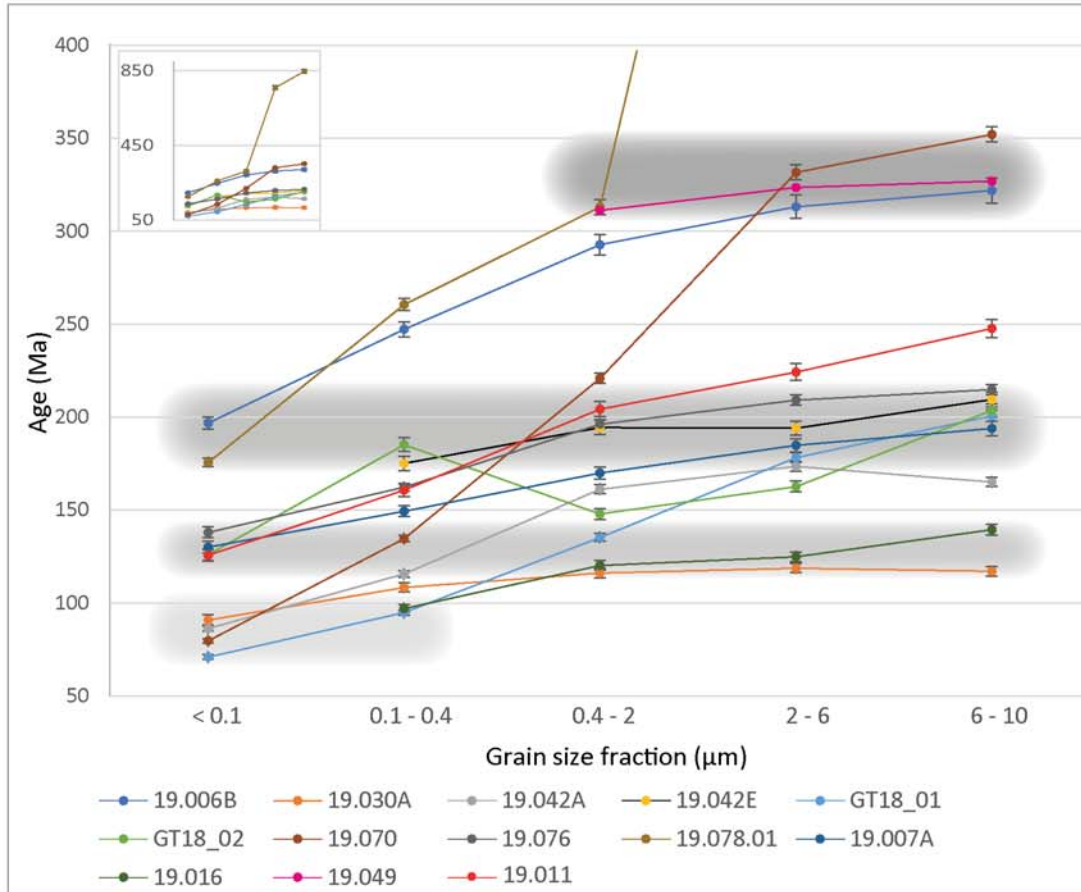


Figure 11

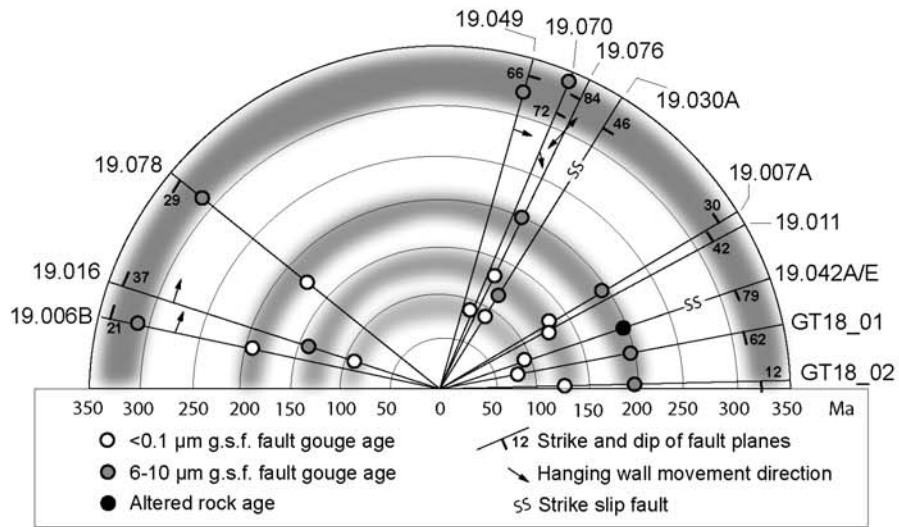


Figure 12

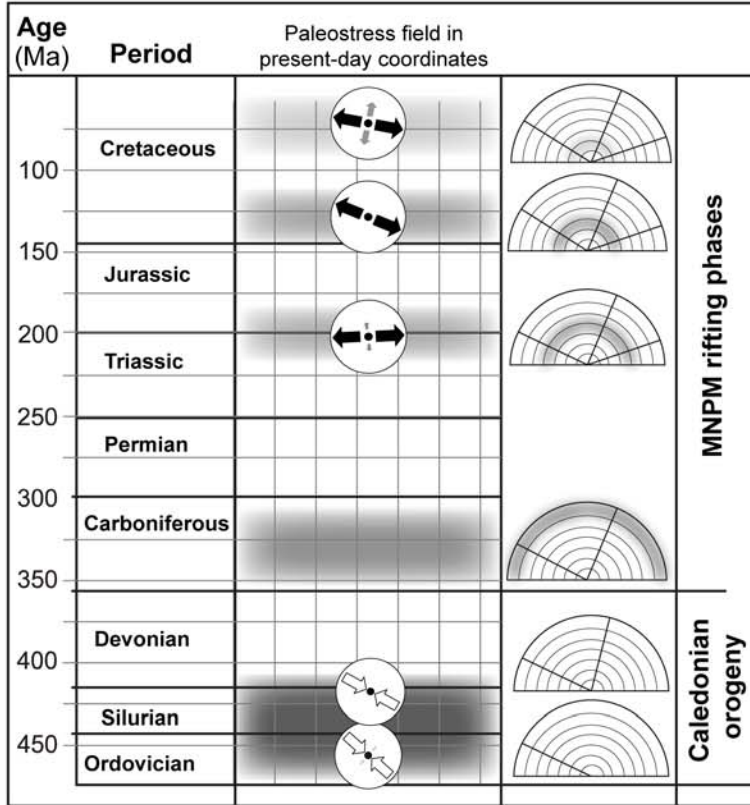


Table 1.

Set	Sample ID	Locality	Coordinates		Host Rock	Strike	Fracture Plane		Slip Line		Slip	Description of fault core	Mineralization
			(WGS84)				Dip Dir	Dip	Trend	Plunge			
			N	E									
WNW-ESE and E-W	19.006B	Hitra	63°28.339'	08°40.505'	Granite	103	13	21	344	27	N	reddish and grey gouges	Subparallel calcite vein
	19.016	Hitra	63°28.116'	08°39.247'	Granodiorite	108	18	37	4	30	I/N	4-cm-thick gouge	calcite vein and zeolite
	19.078	Kråkenes	62°02.087'	04°59.667'	Gabbro	129	219	29				grey light blue gouge and reddish gouge layer	none
ENE-WSW	19.007A	Hitra	63°27.347'	08°36.213'	Granite	235	325	27				cm-thick reddish gouge	Reworked network of thin calcite veins
	19.011	Hitra	63°33.823'	08°46.071'	Amphibolite	250	340	42				foliated gouge	none
	19.042A	Hitra	63° 35.661'	08° 57.858'	Granodiorite	252	162	79	86	10	X	1.5-2 m thick fault zone with clay-rich gouge and cataclastic clasts in the lower part; altered granite in the upper part	none
	19.042E	Hitra	63° 35.661'	08° 57.858'	Granodiorite							foliated gouge	none
	GT18_01	Runde	62° 24.318'	5° 35.563'	Amphibolitic gneiss	260	170	62				foliated gouge	none
	GT18_02	Runde	62° 24.318'	5° 35.563'	Amphibolitic gneiss	270	180	12				foliated gouge cut by a high angle second generation of gouge	none
NE-SW	19.049	Frøya	63°41.762'	08°34.944'	Migmatite	196	106	66	59	52	N	cataclastic layer and gouge	quartz vein and calcite vein
	19.070	Runde	62°23.168'	05°36.732'	Amphibolitic gneiss	203	113	72	152	63	N	reworked shear zone, grey gouge layer embedding deformed	calcite vein cuts across the fault
	19.076	Vestkapp	62°11.487'	05°11.918'	Amphibolitic gneiss	206	296	84	12	36	S	reworked shear zone, 10-cm-thick core with pinkish gouge reworking	hematite
	19.030A	Hitra	63°28.251'	08°20.128'	Granite	212	122	46				reworked cataclastic and cataclastic layers, embedding chunks of	coarse-grained calcite vein

Table 3.

Sample Name	Fraction	⁴⁰ Ar*				K			Age Data	
		Mass mg	mol/g	σ (%)	⁴⁰ Ar* %	Mass mg	wt %	σ (%)	Age (Ma)	σ (Ma)
19.006B	<0.1	2,272	9.34E-10	0.29	75.2	50.7	2,592	1.7	196.8	±3.1
	0.1-0.4	4,796	9.06E-10	0.24	82.5	50.9	1,972	1.8	247.2	±4.1
	0.4-2	2,884	6.02E-10	0.28	86.6	50.9	1,092	2	292.7	±5.4
	2-6	2,678	3.78E-10	0.32	88.5	51	0.637	2.2	313.2	±6.4
	6-10	6,752	3.23E-10	0.24	88.9	50.7	0.528	2.3	321.9	±6.8
19.007A	<0.1	2,508	4.00E-10	0.31	26.6	6.5	1.71	2.7	130	±3.4
	0.1-0.4	1,754	3.70E-10	0.39	26.7	51.6	1.37	2	149.4	±3.0
	0.4-2	4.04	5.10E-10	0.25	39.6	50.8	1.65	2	169.9	±3.3
	2-6	2.02	5.84E-10	0.32	52.1	50.4	1.73	2	184.9	±3.6
	6-10	1.58	5.96E-10	0.37	55.4	50.5	1.68	2	193.9	±3.8
19.011	<0.1	1,854	1.52E-10	0.61	12.6	10.1	0.671	2.4	125.7	±3.0
	0.1-0.4	1,898	1.85E-10	0.53	15.4	50.7	0.635	2.1	160.7	±3.4
	0.4-2	3,556	2.80E-10	0.3	19.2	52.3	0.746	2.1	204.3	±4.1
	2-6	2.56	3.43E-10	0.32	26	51.5	0.827	2.1	224.3	±4.5
	6-10	1,778	4.05E-10	0.38	33.6	53.5	0.88	2.1	247.7	±4.9
19.016	0.1-0.4	1,662	9.50E-11	0.97	13.4	50.6	0.549	2.1	97.1	±2.2
	0.4-2	1,858	1.43E-10	0.64	19.4	52.7	0.664	2.1	120.2	±2.6
	2-6	3.3	1.58E-10	0.39	23.5	50.1	0.705	2.1	125	±2.6
	6-10	2.27	1.91E-10	0.44	27.2	50.7	0.759	2.1	139.5	±2.9
	<0.1	2.31	1.20E-10	0.72	40.2	10.5	0.739	2.9	90.9	±2.7
19.030A	0.1-0.4	3,198	1.02E-10	0.61	57	51.8	0.528	2.3	108.4	±2.5
	0.4-2	2,428	1.50E-10	0.56	66.3	51.8	0.722	2.2	116.1	±2.5
	2-6	4,718	1.72E-10	0.31	79.2	50.5	0.809	2.1	118.8	±2.5
	6-10	1.76	1.81E-10	0.66	79.5	51.2	0.865	2.1	117.1	±2.5
	<0.1	3,394	4.60E-10	0.29	12.4	51.6	3,001	1.6	86.3	±1.4
19.042A	0.1-0.4	2,472	6.26E-10	0.3	18.9	51.9	3.02	1.6	115.7	±1.8
	0.4-2	2,902	9.40E-10	0.27	37.4	50.1	3,216	1.6	161.2	±2.5
	2-6	3,028	1.10E-09	0.26	49.1	50.2	3,473	1.5	173.6	±2.6
	6-10	2,644	1.01E-09	0.27	54.5	50.8	3,373	1.5	165.1	±2.5
	0.4-2	1,104	1.27E-09	0.43	86.5	25.2	2.15	2.1	311.3	±6.2
19.049	2-6	1,646	1.36E-09	0.33	87.4	50.6	2.21	1.9	323.6	±5.9
	6-10	1,744	1.47E-09	0.32	89.8	53.9	2.37	1.9	326.9	±5.8
	<0.1	3,444	1.88E-10	0.37	28.4	50.1	1.33	1.3	79.6	±1.1
19.07	0.1-0.4	2,202	3.42E-10	0.36	43.6	52.6	1.41	1.3	134.7	±1.7
	0.4-2	2,206	7.70E-10	0.3	69.9	54.3	1.89	1.2	220.9	±2.7
	2-6	1,554	1.06E-09	0.35	83.3	51.2	1.68	1.3	331.7	±4.0
	6-10	2,058	1.15E-09	0.3	85	54	1.7	1.3	352.1	±4.1
	<0.1	1,036	3.26E-10	0.68	41.5	7.1	1.31	2.2	138.1	±3.0
19.076	0.1-0.4	2.39	3.89E-10	0.33	55.9	50.3	1.32	1.3	162.4	±2.1
	0.4-2	1,936	4.75E-10	0.36	66.2	52.4	1.32	1.3	196.2	±2.5
	2-6	2,268	4.66E-10	0.32	68.3	52.7	1.21	1.3	209.2	±2.7
	6-10	2.75	4.71E-10	0.29	71	51.7	1.19	1.3	214.8	±2.8
	<0.1	0.736	6.21E-10	0.7	32.5	51.5	1.94	1.2	175.7	±2.4
19.078	0.1-0.4	1,596	9.14E-10	0.35	37.3	50.8	1.88	1.3	260.6	±3.2
	0.4-2	1.13	7.65E-10	0.46	37.6	54.8	1.29	1.3	312.9	±4.0
	2-6	1,766	7.07E-10	0.34	38.4	51.7	0.432	1.5	759.4	±9.6
	6-10	1,456	8.56E-10	0.38	47.6	51.1	0.456	1.5	847.8	±10.5
	<0.1	1,868	1.93E-10	0.58	20.5	50.9	1,533	1.79	71	±1.3
GT18_01	0.1-0.4	3,094	3.69E-10	0.3	32.5	49.9	2,183	1.64	94.9	±1.5
	0.4-2	4,078	6.08E-10	0.26	48.3	50.3	2,493	1.58	135.3	±2.1
	2-6	2.1	9.03E-10	0.31	65.3	50.6	2.78	1.53	178.2	±2.6
	6-10	1,694	1.06E-09	0.34	74.2	49.9	2,879	1.52	200.8	±3.0
	<0.1	2,028	1.74E-10	0.6	40.3	23.2	0.768	2.48	126	±3.1
GT18_02	0.1-0.4	1,792	3.10E-10	0.45	55.2	50	0.917	2.03	185.2	±3.7
	0.4-2	1.91	2.99E-10	0.44	54.8	50.5	1.12	1.93	147.8	±2.8
	2-6	1,604	3.91E-10	0.44	65.7	49.7	1,324	1.87	162.7	±3.0
	6-10	1,592	5.07E-10	0.4	69.1	50.8	1,356	1.85	203.6	±3.6
	<i>Altered rock sample</i>									
19.042E	0.1-0.4	1,822	2.94E-10	0.45	37.4	30.7	0.921	2.3	175.2	±3.9
	0.4-2	2.44	3.49E-10	0.34	42.7	50.2	0.98	2	194.4	±3.8
	2-6	3,112	3.96E-10	0.29	47.9	51	1,115	2	194	±3.7
	6-10	1,988	4.42E-10	0.36	53.4	50.2	1,147	2	209.5	±4.0