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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 Complex (MTFC), an ENE-WSW oriented regional tectonic structure. We propose a time-14 constrained evolutionary scheme for the MNPM brittle history. By means of remote sensing 15 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 multiscalar 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 oriented and diachronic subsidiary rifting pulses that pre- and postdate the main rifting phase 38 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 evolution recorded by passive margins (e.g., Redfield and Osmundsen, 2009; Ksienzyk et al., 74 75 2016; Gabrielsen et al., 2018; Scheiber and Viola, 2018) and better exploration predictive tools.

While decades of petroleum exploration have unveiled the fine details of the regional structure of 76 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.

The intricated network of brittle fractures and faults affecting the offshore MNPM and the lack of well dated sedimentary markers makes retrieving an absolute time sequence for its deformation history a challenging task. All these difficulties notwithstanding, some studies have demonstrated that it is possible to unravel complex brittle histories by studying the details of geometric, kinematic, and geochronological brittle deformation features (e.g., Wilson et al., 2006; Viola et al., 2009; Saintot et al., 2011; Lacombe et al., 2013; Mattila and Viola, 2014; Scheiber and Viola, 2018; Nordbäck et al., 2022). A significant contribution in this respect stems from the possibility to radiometrically date fault rocks associated with key structural features of the margins so as to add absolute time constraints to brittle structural reconstructions (van der Pluijm et al., 2001; Zwingmann and Mancktelow, 2004; Davids et al., 2013; Torgersen et al., 2015; Ksienzyk et al., 2016; Vrolijk et al., 2018; Tartaglia et al., 2020; Fossen et al., 2021).

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

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## 2. GEOLOGICAL SETTING

The MNPM forms a 350-500 km wide sector of rifted and hyperextended continental crust facing the Norwegian-Greenland Sea (Northeastern Atlantic Ocean; Fig. 1). The studied offshore part of the MNPM is composed of two large segments, which, from SW to NE, include the Møre and the Vøring margins (Fig. 1). Discrete, NW-SE-trending tectonic structures, such as the Jan Mayen Corridor, separate these segments (Faleide et al., 2008; Gernigon et al., 2020), each of which contains an inner platform, the Møre, and Vøring Platforms, and a system of outer basins, 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) muscovite <sup>40</sup>Ar-<sup>39</sup>Ar cooling ages from the area (Kendrick et al., 2004; Walsh et al., 2013). 121

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

The investigated areas include the islands of Hitra, Frøya, and Smøla and the Stad area onshore Norway, located at the transition between the NE-SW oriented MNPM and the N-S oriented North Sea margin (Fig. 1). In the northern MNPM, the islands of Hitra and Smøla consist of the 450-428 Ma granitic to granodioritic Smøla-Hitra batholith (Gautneb and Roberts, 1989), attributed to the Upper Allochthon of the Norwegian Caledonides (Fig. 1). The batholith is overlain by Late Silurian-Devonian sandstone and conglomerate exposed along the southeastern coast of Hitra. On the northern part of Hitra and on Frøya, micaschist and foliated migmatite of the Uppermost Allochthon are exposed. In the Stad area, the tip of Kråkenes peninsula is 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).

The MTFC is a composite fault complex consisting of different subvertical fault strands (e.g., Hitra–Snåsa Fault, Verran Fault), and an enveloping volume of rock deformed by second-order fault zones (Séranne, 1992; Braathen et al., 1999, 2002; Redfield et al., 2004, 2005; Osmundsen et al., 2005, 2006; Redfield and Osmundsen, 2009). The main fault strands are defined by a several meter thick mylonitic core, superimposed by brittle fault rocks (Kendrick et al., 2004). The MTFC has been repeatedly reactivated in response to stress regimes varying through time, 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 km thick mylonitic shear zones, as documented by <sup>40</sup>Ar/<sup>39</sup>Ar dating of synkinematic white micas 181 182 (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 al., 2004, 2005). Pseudotachylytes from the Hitra-Snåsa Fault have been dated by <sup>40</sup>Ar/<sup>39</sup>Ar 186 187 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** 

We carried out a multidisciplinary and multiscalar study to better constrain the brittle evolution of the MNPM in space and through time. Our approach includes the remote sensing of bedrock lineaments, field structural and microstructural analysis, paleostress inversion, and radiometric K-Ar dating coupled with the detailed mineralogical characterization of selected fault rocks. In 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**

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

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

### 221 **3.3 Paleostress Inversion**

Paleostress inversion analysis is useful to reconstruct the orientation of the stress field that 222 223 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.

On the other hand, all these uncertainties notwithstanding, several studies have shown that it is possible to obtain consistent paleostress reconstructions even in multiply deformed metamorphic terrains by acquiring a large dataset, carefully sorting statistically fault populations and by validating the results against independent constraints (e.g., Viola et al., 2009; Saintot et al., 2011; 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  $\alpha$ , 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  $\alpha$  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  $\alpha$  value for a given fault is greater than 30°, then that fault is incompatible with the computed reduced tensor, suggesting that it belongs to a different set and that it formed in a different stress field and, thus, 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 ( $\sigma_1$ ,  $\sigma_2$ , and  $\sigma_3$ , with  $\sigma_1 \ge \sigma_2 \ge \sigma_3$ ) 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  $\sigma_1$  is vertical (extensional stress regime), R' = 2 - R, when  $\sigma_2$  is vertical (strike-slip stress regime) and R' = 2274 275 + R, when  $\sigma_3$  is vertical (compressional stress regime; Delvaux and Sperner, 2003). R' ranges from 0 to 3. R' = 0.5 describes a pure extensional regime, R' = 1.5 describes a pure strike slip 276 277 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.

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

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

K-Ar radiometric dating and X-ray diffraction (XRD) were performed on twelve fault gouges and one non-cohesive altered rock at the laboratories of the Geological Survey of Norway (Trondheim, Norway) following the procedure described by Viola et al. (2018). Collected fault rock samples are representative of the main trends of structural lineaments defined by remote 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  $\mu$ m) 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).

## **4. RESULTS**

## 300 4.1 Bedrock Lineament Pattern

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

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.

#### **4.2 Mesostructural Brittle Fault Characterization**

The structural stations studied along the MNPN are shown in Figure 4a. Faults and fractures at the studied sites cut across different types of rock. The host rocks along the MNPM are mainly represented by granitic and granodioritic gneiss of the WGR, Devonian sandstone and 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 Jøsnøya, 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 Jøsnøya are, in turn, cut by younger calcite veins.

Brittle fault zones along the margin are associated with a broad range of fault rock assemblages. In highly fractured fault cores, fine-grained cataclasite is associated with or crosscut by distinct layers of clay-rich fault gouges (Fig. 5c, d). Due to the lack of marker horizons in the study area, the amount of displacement along individual faults remains largely unknown. Locally, along some brittle structures, the host rock is dismembered, and appears as unconsolidated coarsegrained material. In these fault and fracture zones, the mostly granitic or gabbroic host rock is altered. Brittle structures along the MNPM exhibit different mineralogical coatings, including epidote, chlorite, calcite, and quartz, sometimes decorating the same plane but also crosscutting and overprinting each other, forming a complex multilayered arrangement of fault rocks. Epidote coatings on faults and fractures are generally reworked and/or cut by younger, few mm-thick, milky white prehnite veins (Fig. 5e). Calcite occurs as mm-thick striated synkinematic coating on the fault planes (Fig. 5f), as variably thick veins, or secondary, coarse-grained, post-kinematic 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

Twelve fault gouges and one altered rock were sampled for radiometric K-Ar dating. The details of the sampled fault zones are reported in Table 1 and Figure 6. The sampled fault gouges belong to three main sets of faults: i) low-angle (W)NW-(E)SE striking fault planes (samples 19.006B, 19.078, and 19.016; Fig. 6a-c); ii) variably dipping, ENE-WSW oriented faults (samples 19.007A, 19.011, 19.042A, GT18\_01/2; Fig. 6d-i); iii) medium to high-angle, oblique and dipslip, NNE-SSW oriented faults (samples 19.030A, 19.049, 19.070, 19.076; Fig. 6j-m). Samples
are from different rock types (Table 1), mainly granitic and gabbroic rocks (19.006B, 19.007A,
19.016, 19.030A, 19.042A/E, 19.078), migmatite (19.049) and amphibolite or gneiss (19.011,
19.070, 19.076, GT18\_01, GT18\_02).

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

Samples 19.042A and E are from an ENE-WSW striking strike-slip fault and are from two different structural levels (Fig. 6f, g). In the lower part of the outcrop, a grey, very plastic fault gouge is exposed (sample 19.042A), whereas the upper part of the same brittle structure is filled 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 the397 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 younger399 gouge layer (GT18\_01) has a steeper dip (Fig. 6i).

Finally, some brittle fault zones exploit inherited ductile planar fabrics. For example, NNE-SSW
striking mylonitic shear zones were reworked by brittle faulting (e.g., samples 19.070 and
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., Ep <sub>1</sub> , Ep <sub>2</sub> and Ep <sub>3</sub> 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., $Qtz_1$ , $Qtz_2$ 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).

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

## 432 **4.4. Paleostress Inversion**

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

A pure compressional stress regime with the maximum horizontal compressional stress axis ( $\sigma_1$ ) oriented NW-SE was obtained by the inversion of low angle N-dipping faults (R': 2.54), that invariably containing brittle fault rocks in their cores (Fig. 8a). Some of these faults are reverse, while other faults recorded a later extensional reactivation.

453 Pure compression was followed by a transpressional stress regime (R':1.98) with  $\sigma_1$  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).

A well recorded stress regime along the entire MNPM is an extensional stress field with  $\sigma_3$ oriented E-W (R':0.37; Fig. 8c). It generated NNE-SSW and NNW-SSE normal faults and reactivated some NW-SE oblique fault planes. This stress field is recorded by fault zones mainly decorated by epidote and quartz coatings, and subordinately exposing brittle fault rocks in their cores. The last two inverted stress regimes reactivated variably oriented faults and fractures inherited from the previous deformation episodes. An extensional field with  $\sigma_3$  oriented NW-SE reactivated faults and fractures decorated by prehnite, quartz, zeolite and calcite coatings (R': 0.52; Fig. 8d). Finally, prehnite decorated faults and/or associated with brittle fault rocks, oriented NE-SW and NW-SE formed or were reactivated by an extensional stress regime with  $\sigma_3$ 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 (Table 2, Fig. 9), such that the analyzed  $< 0.1 \,\mu m$  grain size fraction of all the samples only 473 474 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 layers and illite-vermiculite (Fig. 9). K-feldspar is not present in all the samples, but when 479 480 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  $848 \pm 11$  to  $71 \pm 1$  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 µm, and samples 19.016 and 19.042E did not contain sufficient 486 material for dating of the finest (<0.1 µm) 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 ( $< 0.1 \mu m$ ) 489 are  $197 \pm 3$ ,  $176 \pm 2$ ,  $138 \pm 3$ ,  $130 \pm 3$ ,  $126 \pm 3$  (twice),  $91 \pm 3$ ,  $86 \pm 1$ ,  $80 \pm 1$  and  $71 \pm 1$  Ma 490 (Table 3).

## 491 **5. DISCUSSION**

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

### 497 **5.1 Fault Rock Mineral Assemblages**

The sequence of mineral coatings established for the analyzed fault cataclasites constrains a progressive evolution of retrograde cooling of the MNPM, presumably from the Carboniferous onward. Indeed, the epidote + chlorite (+ quartz) assemblage observed in the oldest cataclasite generations suggest deformation conditions close to the frictional-viscous transition in the continental crust under sub-greenschist facies conditions (T  $\leq$ 275 °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  $\approx$  200-250 °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 °C (Tartaglia, 2021). These temperature estimates are in line 514 with zeolite formation temperatures, commonly well below 200 °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.

All in all, this sequence of mineralogical assemblages within the studied fault rocks supports a general cooling trend throughout the brittle evolution of the MNPM. However, the lack of a clear systematic relationship correlating fault strike and fault mineralogy (or sequence of mineral assemblages; see Section 5.2 below, Fig. 8) does not allow us to retrieve general conclusions at the scale of the MNPM. Nevertheless, the mineral paragenesis coupled with K-Ar data from fault zones may provide local, but valuable constraints on the thermal and temporal evolution of the 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  $<0.1 \ \mu m$  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 multiscalar structural analysis) and that the obtained

ages are comparable to the timing of already known regional tectonic events (e.g., Fossen et al., 2021; Osmundsen et al., 2021) suggests that the obtained K-Ar fault gouge dates are truly due to synkinematic crystallization and not to post-tectonic authigenesis and or alteration due to later 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  $(0.1-0.4, 0.4-2 \text{ and } 2-6 \mu \text{m})$  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  $\mu$ m 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  $321 \pm 8$  Ma (MSWD=2.74). A Triassic/Jurassic mean age of  $202 \pm 6$  Ma (MSWD=5.9) is 565 calculated from the age cluster constituted by the  $<0.1 \mu m$  finest grain size fraction of sample 566 19.006B, the coarsest 6-10  $\mu$ m 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  $177 \pm 12$  (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  $122 \pm 5$  Ma (MSWD=3.5). Finally, three

<0.1 μm fraction ages are younger than 100 Ma (c. 71, 80, 86, 91 Ma), indicating an episode of</li>
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 ( $201 \pm 3$ ,  $204 \pm 4$  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.

- To summarize, the obtained K-Ar ages can be grouped into four significant clusters, which, fromolder to younger, date to:
- 591 1) Carboniferous, with a mean age of  $321 \pm 8$  Ma (MSWD=2.74);

592 2) Late Triassic-Jurassic with two distinct mean ages at 202 ± 6 Ma (MSWD=5.9) and at 177 ±
593 12 Ma (MSWD=0.1);

594 3) Early Cretaceous with a mean age of  $122 \pm 5$  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. 320 Ma) have been constrained by K-feldspar <sup>39</sup>Ar/<sup>40</sup>Ar dating from a cataclastic granite in the 605 606 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

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

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

626 Our  $122 \pm 5$  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**

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

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

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

Four dated samples belong to this set of lineaments (Fig. 11). Their K-Ar ages of the  $<0.1 \mu m$ grain size fractions are Early and Late Cretaceous. The coarser grain size fractions consisting of illite-smectite mixed layers and illite yield Carboniferous, Triassic, and Early Cretaceous ages (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.

The NE-SW and NNW-SSE striking faults are geometrically and kinematically compatible with a transpressional stress regime with  $\sigma_1$  oriented NW-SE, which formed epidote and chlorite mineralizations. Additionally, they could be reactivated during the NW-SE extensional stress regime characterized by prehnite, quartz and calcite mineralizations. Some of these faults may thus have exploited a preexisting Caledonian fabric, prior to their own repeated reactivation and 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 ( $201 \pm 3$  Ma, GT18 01, Table 3).

The ENE-WSW striking faults have a variable dip and accommodate strike-slip (mainly sinistral) or normal kinematics. They are mainly associated with chlorite and iron oxide mineralizations and locally with epidote and prehnite. These faults are geometrically compatible with the E-W 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

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

Three faults belonging to this lineament set have been dated (Fig. 11). Samples 19.006B and 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  $\sigma_1$ , 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.

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

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

722 **5.4 Regional Implications** 

The mapping, characterization and dating of the brittle structural framework along the MNPM allows us to place new temporal constraints onto the evolution of the margin, as summarized in Figure 12. This figure shows the obtained inverted paleostress fields through time and the strike 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 <sup>39</sup>Ar/<sup>40</sup>Ar ages of ca. 450 and 435 Ma were obtained from synkinematic micas (Scheiber et al., 730 731 2016). In fact, paleostress inversion of our fault sets resulted in a pure compressional stress 732 regime with  $\sigma_1$  oriented NW-SE and a transpressional stress regime (R':1.98) with  $\sigma_1$  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).

In the area, the MTFC was possibly active in pre-Devonian times accommodating dextral transpression (Séranne, 1992) with the formation of pervasive ductile fabrics that were reactivated during the formation of the MNPM. The MNPM accommodated an extensional history that started in the Early Devonian with the collapse of the Scandinavian Caledonides (Doré et al., 1999; Faleide et al., 2008; Gernigon et al., 2020). A few structural observations onshore East Greenland support the idea that the first rifting phase affecting the NE Atlantic was 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).

During the Triassic-Jurassic faulting event recorded by the onshore MNPM, E-W crustal stretching occurred, as well documented by previous studies from the North Sea and the MNPM (Gómez et al., 2004; Scheiber and Viola, 2018; Gernigon et al., 2020). This rifting phase reactivated older structures and formed epidote-rich cataclasites and quartz veins associated with
NE-SW, NW-SE, and ENE-WSW-striking faults (Figs. 11, 12). Coeval alteration/deep
weathering occurred, likely due to fluid circulation along brittle structures dissecting tilted
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).

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

#### 792 **6. CONCLUSIONS**

The MNPM derives from a complex polyphase faulting history. Its evolution was influenced by the effects of the spatial orientation of old, inherited structures, such as the Caledonian orogenparallel structural grain and the MTFC (*cf.* Osmundsen et al., 2006; Phillips et al., 2019; Schiffer et al., 2019). By means of remote sensing lineament analysis, field work, microstructural analysis, paleostress inversion, mineralogical characterization and K-Ar dating of illite separated from selected fault zones, six tectonic events have been identified. From older to younger these 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  $\sigma_1$  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.
- iv) Late Triassic-Early Jurassic E-W extension at c. 202 and 177 Ma forming epidote and
  quartz-coated, N-S striking, generally normal faults, and coeval alteration of the host
  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;

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

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

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

822

#### 823 Table and Figure Captions

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

827

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

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

833

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

835

Figure 1 - Simplified geological map of the on- and offshore mid-Norwegian Passive Margin.
Drawn after Faleide et al. (2008) and Slagstad et al. (2011). (MTFC: Møre-Trøndelag Fault

838 Complex; WGR: Western Gneiss Region).

839

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

845

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

849

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

853

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

861

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

866

Figure 7 - Microphotographs of representative fault rocks from Hitra island. (a) Granitic protocataclasite characterized by pervasive fracturing and epidote mineralization. The host rock is partially altered, as indicated by the occurrence of chlorite (Chl) and saussurritic plagioclase (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 (Ep<sub>3</sub>). The quartz-rich layer contains a younger quartz vein (Qtz<sub>2</sub>) 873 localized at the contact between an epidote-bearing cataclastic layer (top) and the quartz-rich 874 cataclasite (bottom,  $Qtz_1$ ). Note the occurrence of multiple layers of variably sized epidote 875 cataclasites (Ep<sub>2</sub>, Ep<sub>1</sub>). (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

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

888

**Figure 9** - *Pie charts of mineral concentrations* [*wt%*] *derived from XRD of each of the dated* fractions of three representative fault gouges of different ages (i.e., ages of the finest grain size fractions): 19.070 ( $80 \pm 1$  Ma), 19.042A ( $86 \pm 1$  Ma), 19.006B ( $197 \pm 3$  Ma). 892

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

896

**Figure 11 -** *Radial diagram of fault gouge and altered rock K-Ar ages plotted as a function of the fault plane orientation. (G.s.f.: grain size fraction). Only K-Ar ages interpreted as 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  $\sigma_1$ ,  $\sigma_2$  and  $\sigma_3$ , respectively.

905

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	Q		Coord	inates	ĸ		Fracture		Slip	Line			ation
Set	Sample	Localit	N (003	E	Host Ro	Strike	Dip Dir	Dip	Trend	Plunge	Slip	Description of fault core	Mineraliza
d E-W	19.006B	Hitra	63°28.339'	08°40.505'	Granite	103	13	21	344	27	Ν	reddish and grey gouges	Subparallel calcite vein
W-ESE an	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
MM	19.078	Kråkenes	62°02.087'	04°59.667'	Gabbro	129	219	29				grey light blue gouge and reddish gouge layer	none
JE-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	х	1.5-2 m thick fault zone with clay-rich gouge and cataclasite clasts in the	none
Ξ	19.042E	Hitra	63° 35.661'	08° 57.858'	Granodiorite							lower part; altered granite in the upper part	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				high angle second	none
NE-SW	19.049	Frøya	63°41.762'	08°34.944'	Migmatite	196	106	66	59	52	N	cataclasite layer and gouge	quartz vein and calcite vein
	19.070	Runde	62°23.168'	05°36.732'	Amphibolitic gneiss	203	113	72	152	63	Ν	grey gouge layer	cuts across
	19.076	Vestkapp	62°11.487'	05°11.918'	Amphibolitic gneiss	206	296	84	12	36	S	cm-thick core with	hematite
	19.030A	Hitra	63°28.251'	08°20.128'	Granite	212	122	46				cataclastic layers,	grained

Sample	Grain size fraction	Qtz	Kfs	Pig	ç	ĥ	Amp	Px	IIVMs	mica	Palyg	<b>K</b> In	다	Sm	ChI-Sm	Sm and ChI-Sm	III-Sm and Sm	III-Verm	Lepid	Zeo	Stb	Ap	Anl	GOF
	<0.1			4		6			38				35	17										1.63
Sample 19.0068 19.007A 19.011 19.016 19.030A 19.042A 19.049 19.070 19.077 19.078 GT16_L	0.1-0.4		2	5		15			27				30	21										1.81
9.00	0.4-2	12	3	7		28			14				22	14										1.68
6B	2-6	27	3	9		28	1		8		1		15	10		1								1.68
	6-10	31	3	9		26			6				14	11										1.7
	<0.1	too little																						
-	04.04	material												. 05									.5	
9.00	0.1-0.4	-5		E 10			m  m			<5	_													
17A	2.6	<5		10.15					6 10				5 10	>50									<5	
	6.10	~5		10-15					5.10				10.15	>50									~5	
	0-10	too little		10-13					3-10				10-13	200									~5	
	<0.1	material																						
19.	0.1-0.4								trace				<5	>95										
011	0.4-2	<5		<5					<5				5-10	>80										
	2-6	<5		5-10					5-10				5-10	>70										
	6-10	<5		5-10					5-10				5-10	>70										
	<0.1	too little material																						
Image  Image <th< td=""><td>material</td><td></td><td>&lt;5</td><td></td><td></td><td></td><td></td><td></td><td></td><td></td><td></td><td>trace</td><td>&gt;90</td><td></td><td></td><td></td><td></td><td></td><td></td><td></td><td></td><td>&lt;5</td><td></td></th<>	material		<5									trace	>90									<5		
9.01	0.4-2	trace		<5			trace						trace	>85									<5	
6	2-6	<5		5-10			trace		trace				trace	>80									<5	
	6-10	<5		5-10			trace		trace				trace	>80									<5	
	<0.1	too little																						
	0104	material				<10						<10		> 00										
9.03	0.1-0.4			0		24						10		-00										2.04
0A	2.6			25		24						10	6	37										2.94
	6-10	2		35		10						10	6	28										2.40
	<0.1	-		00		10						10		20			100	_	_					2.10
-	0.1-0.4																100							
9.04	0.4-2	5	7	15					37				23				13							2.37
2A	2-6	10	18	13	1	P  P P  P				2.34														
	6-10	14	16	12	2				26				14				16							2.23
19.0	<0.1	too little																						
		material too littlo																						
	0.1-0.4	material																						
049	0.4-2	3		trace?	trace				30					67										1.38
	2-6	5		7					35				trace	53										2.45
	6-10	5		7					35				trace	53										2.46
	<0.1												4	96										4
10	0.1-0.4												7	93										3.21
.070	0.4-2	3											17	78						2				2.48
0	2-6	7											17	70						4		2		2.63
	6-10	8											17	69						4		2		2.52
	<0.1	material																						
19	0.1-0.4	7											23			70								2.32
.077	0.4-2	9		4			6	6					22			53								1.99
	2-6	12		5			11	7	trace				20			45								2.07
	6-10	12		5			12	6	trace				20			45								2.06
	<0.1									x			x	х	х				х					
19	0.1-0.4									x			х	х	х				х					
.078	0.4-2	trace		trace									XXXXX	XXXXX	XXX				ХХ					
	2-6	x		x									XXXX	XXXX	XXX				x					
	6-10	x		x									XXXX	XXXXX	XXX				x					
~	<0.1										XXXXX			XXXX										
311	0.1-0.4	Irees									XXXX	Incard 2		XXXX										
100	2-6	v	YY								YYYY	v v		****										
-	6-10		***											****										
	<0.1	~~~										~~												
G.	0.1-0.4			trace?										XXXXX				x?			x?			
T18	0.4-2	x		xx										XXXX				x			XXX			
02	2-6	ХХ		XX										XXXXX				x			XXXXX			
	6-10	ХХ		XX										XXXX				x			XXX			
											Alte	red rock sa	mple											
	<0.1	too little																						
-	-	material too little																						
9.00	0.1-0.4	material																						
142E	0.4-2	x		xx					XXX				XXX	XXXX										2.36
	2-6	x		ХХ					XXX				XXX	XXXXX										2.58
	6-10	× ×		1 YY		1	1		1 YYY		1	1	×××	YYYY		1								2.62

Table 3.	

		40 <b>Ar</b> *				к		Age Data				
Sample	Fraction	Mass mg	mol/g	σ (%)	40 <b>Ar</b> * %	Mass mg	wt %	σ (%)	Age (Ma)	σ (Ma)		
Name	<0.1	2 272	9 34E-10	0.29	75.2	50.7	2 502	17	196.8	+3.1		
19.006B	0.1-0.4	4 796	9 06E-10 0.24 82 5 50.9		50.9	1 972	1.7	247.2	±0.1 +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		
	<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		
19.007A	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		
	<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		
19.011	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		
	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		
19.016	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		
	0.1-0.4	3,198	1.02E-10	0.61	57	51.8	0.528	2.3	108.4	±2.5		
19.030A	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		
	0.1-0.4	2,472	6.26E-10	0.3	18.9	51.9	3.02	1.6	115.7	±1.8		
19.042A	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		
	0.1-0.4	2,202	3.42E-10	0.36	43.6	52.6	1.41	1.3	134.7	±1.7		
19.07	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		
	0.1-0.4	2.39	3.89E-10	0.33	55.9	50.3	1.32	1.3	162.4	±2.1		
19.076	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		
	0.1-0.4	1,596	9.14E-10	0.35	37.3	50.8	1.88	1.3	260.6	±3.2		
19.078	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		
	0.1-0.4	3,094	3.69E-10	0.3	32.5	49.9	2,183	1.64	94.9	±1.5		
GT18_01	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		
	0.1-0.4	1,792	3.10E-10	0.45	55.2	50	0.917	2.03	185.2	±3.7		
GT18_02	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		
Altered rock sample												
	0.1-0.4	1,822	2.94E-10	0.45	37.4	30.7	0.921	2.3	175.2	±3.9		
10.0425	0.4-2	2.44	3.49E-10	0.34	42.7	50.2	0.98	2	194.4	±3.8		
19.042E	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		