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In-situ quantification of mechanical and permeability properties on outcrop analogues of offshore fractured and weathered crystalline basement: Examples from the Rolvsnes granodiorite, Bømlo, Norway

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1 **In-situ quantification of mechanical and permeability properties on outcrop analogues of**  
2 **offshore fractured and weathered crystalline basement: Examples from the Rolvsnes**  
3 **granodiorite, Bømlo, Norway**

4  
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9 **Abstract**

10 Fractured and weathered crystalline basement units below erosional unconformities potentially  
11 represent unconventional reservoirs for georesources (oil, mineral and water). The reservoir  
12 properties and characteristics strongly depend on secondary processes connected to the local  
13 structural and alteration/weathering history. Here we present the results of in-situ field quantification  
14 of mechanical (uniaxial compressive strength) and petrophysical (permeability) properties of a  
15 fractured and weathered crystalline basement at selected outcrops on the island of Bømlo (western  
16 Norway). The Bømlo outcrops are believed to represent an onshore analogue of the unconventional  
17 oil reservoir hosted in the offshore Utsira High granodioritic fractured basement (northern North Sea).  
18 The off- and onshore crystalline basements have both undergone surficial weathering during the  
19 Mesozoic, as shown by the occurrence of a dated, variably thick saprolitic profile on top of fresh  
20 fractured basement blocks. The Bømlo crystalline basement is characterized by a complex and highly  
21 permeable fracture network. Fault rocks within its fault zones are characterised by an anisotropic  
22 mechanical strength and by an average permeability that is two orders of magnitude larger than that  
23 of the host rock. The matrix permeability and mechanical strength are significantly affected by  
24 alteration/weathering products. Analysis of the textural and mineralogical characteristics of the

25 weathered outcrops allowed us to constrain the variation of permeability and mechanical strength as  
26 a function of increasing alteration and to infer their distribution in the, now eroded, top–basement  
27 weathering profile on Bømlo. Weathering enhances permeability and drastically decreases the  
28 mechanical strength. Nevertheless, evolved saprolitic horizons may act as low–permeability top–seal  
29 units to the fractured and weathered crystalline basement reservoir. The obtained permeability and  
30 mechanical data are finally used to better constrain the potential reservoir rocks, the fluid migration  
31 pathways, and to discuss their role in the geomechanics of a conceptualised fractured and weathered  
32 crystalline basement unconventional reservoir.

### 33 **Keywords**

34 Fractured basement reservoir; Saprolite; Utsira High; Fault rock permeability.

35

36

## 37        **1. Introduction**

38    Faults, fractures, and related alteration zones control fluid transport within low-permeability  
39    crystalline basement rocks at different crustal levels within all tectonic settings and geological  
40    conditions (Stober and Bucher, 2015). Indeed, brittle deformation and alteration processes may  
41    drastically affect the mechanical strength and permeability of inherently stiff, low-permeability  
42    crystalline basement rocks through the development of secondary porosity (Brace, 1984; Bruhn et al.,  
43    1994; Caine et al., 1996; Stober and Bucher, 2007; Place et al., 2016; Staněk and Géraud, 2019). A  
44    better understanding of how these processes affect the rock mechanical strength and the properties  
45    controlling fluid flow is, therefore, of interest to geoscientific research and many applied fields.  
46    Examples are the geological characterisation necessary to the construction of deep nuclear waste and  
47    CO<sub>2</sub> repositories (Armitage et al., 2013), the better definition of crustal rheology and rock mechanics  
48    (Caine et al., 1996; Sibson and Rowland, 2003) and the exploration and production of georesources  
49    (Dewandel et al., 2006; Stober and Bucher, 2015; Kitchka et al., 2017). Specific to this last point, in  
50    the recent past, economically viable unconventional oil plays have been defined in structural highs  
51    within fractured and weathered crystalline basement blocks (Koning, 2003; Luthi, 2005; Kitchka et  
52    al., 2017; Braathen et al., 2018; Bonter and Trice, 2019; Holdsworth et al., 2019, 2020; Trice et al.,  
53    2019; McCaffrey et al., 2020). There, the constituent elements of a reservoir include (Riber et al.,  
54    2015, 2017; Fredin et al., 2017b; Braathen et al., 2018; Lothe et al., 2018; Walter et al., 2018;  
55    McCaffrey et al., 2020): (1) the host crystalline rock, characterised by primary low-  
56    porosity/permeability; (2) fractures, fault zones, mineral veins and open fissures characterised by  
57    enhanced micro- and mesoscopic fracture porosity; (3) fluid-rock interaction zones related to either  
58    alteration along structural discontinuities or to top-basement paleo-weathering profiles; (4) the  
59    buffering/sealing fault zones bounding the structural highs; (5) the sedimentary cover overlying the  
60    weathered crystalline basement acting as either reservoir rock, top-seal and/or source rock. The  
61    reservoir rock consists of the fractured and altered/weathered crystalline basement, in which fluid  
62    storage and transport are dependent mainly on the spatial arrangement, geometry and distribution of

63 “matrix” and “structural” permeability (Braathen et al., 2018; Trice et al., 2019). The “matrix”  
64 permeability is directly related to the textural, compositional, and alteration characteristics of the host  
65 crystalline basement (Bruhn et al., 1994; Géraud et al., 2010; Braathen et al., 2018). The “structural”  
66 permeability, on the other hand, reflects and is related to the presence of structural discontinuities at  
67 all scales, from microfractures to crustal-scale fracture and fault zones (Bruhn et al., 1994; Sibson  
68 and Rowland, 2003; Holdsworth et al., 2020). Brittle faulting and fluid percolation within structural  
69 discontinuities may either increase and/or decrease the matrix permeability of the host crystalline  
70 basement (Bruhn et al., 1994; Caine et al., 1996; Holdsworth et al., 2019, 2020; McCaffrey et al.,  
71 2020). The geological conditions (depth, pressure, temperature) and the origin (metamorphic,  
72 hydrothermal, or meteoric) of fluids may lead to different brittle fault rock development and different  
73 types of alteration from textural and compositional points of view (Steefel and Mäher, 2009).

74 The nature, spatial arrangement, and geometry of structural and textural features at the reservoir scale  
75 are commonly defined through seismic geophysical investigations. Textural variations, fracture, and  
76 fault zones displaying sufficient seismic impedance contrast as well as fault thickness or throws  $> 4$ –  
77 10 m are usually detected by reservoir-scale seismic investigations and are thus classified as seismic-  
78 resolution-scale (SRS) features (Tanner et al., 2019). However, the site- (well-, borehole- or tunnel-  
79 ) scale fluid transport and storage capabilities are effectively controlled by sub-seismic-resolution  
80 scale (SSRS) features (thickness/throw  $< 4$ –10 m; Damsleth et al., 1998; Walsh et al., 1998).

81 In-situ analysis of outcrop analogues of onshore fractured and weathered basement blocks may  
82 provide the unique opportunity to characterise the details of the petrophysical and mechanical  
83 properties of SSRS features and to understand their relationships with larger, reservoir-scale SRS  
84 features (Jones et al., 2008; Howell et al., 2014; McCaffrey et al., 2020; Primaleon et al., 2020). We  
85 present here the results of the in-situ structural, geomechanical (uniaxial compressive strength, UCS),  
86 and petrophysical (permeability,  $k$ ) characterisation of selected outcrops of fractured and weathered  
87 crystalline basement rocks on the island of Bømlo, western Norway (Fig. 1a–b). The crystalline

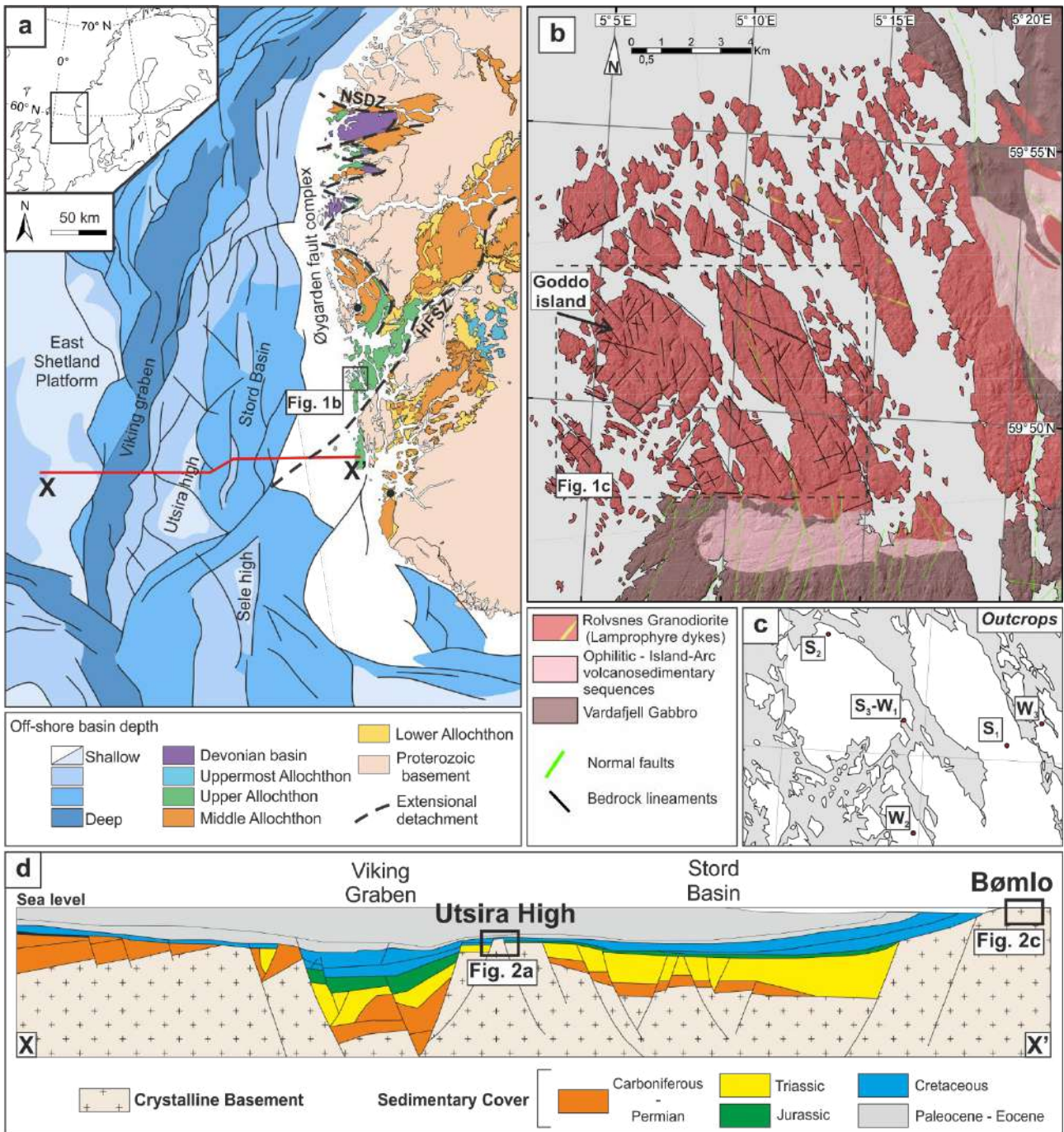
88 basement on Bømlo is being currently studied as the potential onshore analogue of fractured and  
89 weathered basement reservoirs of the Utsira High, an economically viable unconventional reservoir  
90 located offshore western Norway in the northern North Sea (Figs. 1a–2a; Banks et al., 2019; Trice et  
91 al., 2019). The crystalline basement on Bømlo consists of igneous (granodiorite, granite and gabbro)  
92 and volcano–sedimentary units of Ordovician–Silurian age belonging to the Upper Allochthon of the  
93 Caledonian orogen (Slagstad et al., 2011). From the Permian onward, the crystalline basement on  
94 Bømlo underwent a series of brittle deformation events (Scheiber and Viola, 2018) and experienced  
95 pervasive alteration/weathering in response to subaerial exposure in tropical humid climate  
96 conditions during the Triassic, forming a thick weathering profile (Fredin et al., 2017b). Similarly,  
97 the offshore Utsira High reservoir is characterised by a pervasively fractured crystalline basement,  
98 mainly composed of felsic intrusive and gabbroic rocks of Ordovician–Silurian age (Slagstad et al.,  
99 2011; Riber et al., 2015), on top of which rests a paleo–weathering profile formed during the late  
100 Triassic (Fredin et al., 2017b; Riber et al., 2017; Lothe et al., 2018) (Fig. 2b).

101 On Bømlo, we have worked on a few selected outcrops (Fig. 1c) that are representative of (1)  
102 structural discontinuities affecting the crystalline basement “matrix” properties at different scales,  
103 from SSRS fracture corridors to SRS fault zones (Fig. 2c) and (2) textural heterogeneities related to  
104 the progressive development of alteration/weathering products (Fig. 2b). Our analyses aimed  
105 specifically at the characterisation of the different structural and alteration products controlling the  
106 current “matrix” mechanical and permeability properties of the crystalline basement at the sub–  
107 seismic–resolution scale. The role of the different structural and alteration elements in controlling  
108 fluid transport and storage, and the characterisation of potential reservoir rocks in fractured and  
109 weathered crystalline basement structural highs, are then discussed to the benefit of broader, more  
110 general conceptual scenarios.

111 Quantitative, high–resolution geomechanical and petrophysical datasets are quite rare, yet  
112 fundamental for interpreting increasingly higher–resolution geophysical datasets, for reducing the

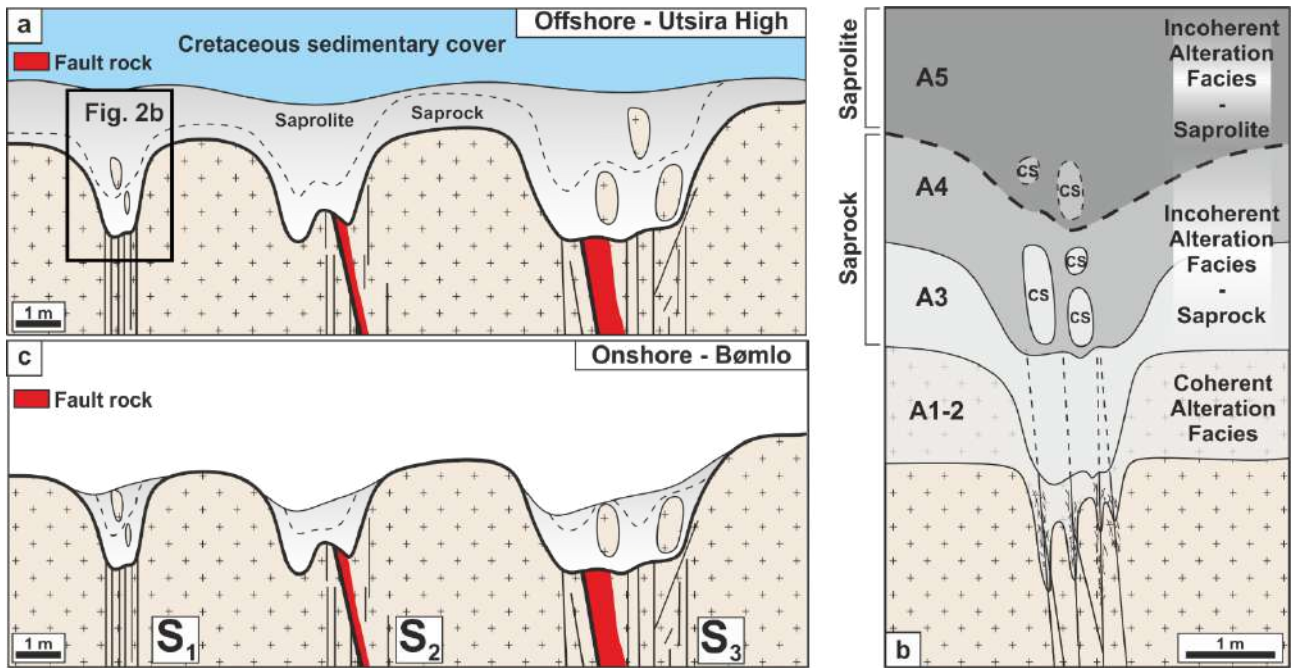
113 uncertainty in reservoir modelling, for better constraining fault zone mechanics and for assessing  
114 borehole stability during exploration and drilling (Rutqvist and Stephansson, 2003; Wibberley et al.,  
115 2008; Steer et al., 2011; Jeanne et al., 2017). Our results are thus useful to help bridge the gap between  
116 field observations and large-scale geological contexts, where the lack of direct access to the  
117 geological objects of interest and the processes determining their spatial and temporal evolution  
118 commonly preclude straightforward across-scale correlations and quantifications. In addition, the  
119 geological/structural contextualisation offered by the quantitative, high-resolution dataset presented  
120 here may help to better read and interpret the at times overwhelming wealth of global and regional  
121 geomechanical/petrophysical datasets (Lothe et al., 2018; Walter et al., 2018; Scibek, 2020).





122

123 **Figure 1:** (a) Geological map of southwestern Norway and the northern North Sea (modified after  
 124 Slagstad et al., 2011; Scheiber and Viola, 2018). HFSZ: Hardangerfjord Shear Zone; NSDZ:  
 125 Nordfjord–Sogn Detachment Zone. The red line X–X' represents the trace of the geological cross  
 126 section presented in Fig. 1d. (b) Geological map of the northern Bømlo area, including the island of  
 127 Goddo (redrawn from the NGU geological map of Norway 1:50000). (c) Simplified topographic map  
 128 of Bømlo showing the locations of the studied outcrops. (d) Schematic geological cross section  
 129 through the central North Sea (X–X' in (a)) (modified from Riber et al., 2015).



131

132 **Figure 2:** (a) Simplified schematic representation of the fractured and weathered basement top section  
 133 preserved offshore in the Utsira High beneath the Mesozoic sedimentary cover. This schematic  
 134 representation focuses on the fractured crystalline basement, which is characterised by different  
 135 typologies of fractures and fault zones, the weathering profile (saprock–saprolite in grey) and the  
 136 overlying sedimentary cover (light blue). The dashed line represents the boundary between saprock  
 137 and saprolite within the weathering profile. (b) Simplified sketch of the top–basement weathering  
 138 profile showing the gradual transition from unaltered and fractured host rock at the base of the profile,  
 139 toward saprock and saprolite at the top. The weathering profile locally reaches greater depths where  
 140 in correspondence with structural discontinuities (in this case, a SSRS fracture corridor, S<sub>1</sub> of Fig. 2a  
 141 and 2c). The progression of weathering is illustrated by the continuous darkening of the grey shades  
 142 and the increasing number in the alteration/weathering grade on the left (A1–A5). (CS: core stones).  
 143 (c) Simplified schematic representation of the fractured and weathered basement top section now  
 144 exposed onshore on the island of Goddo. The erosion of the weathering profile has led to the exposure  
 145 of an “etched” basement top surface (Fredin et al., 2017). S<sub>1</sub>, S<sub>2</sub> and S<sub>3</sub> represent the three typologies

146 of deformation structures studied in this work. S<sub>1</sub>: SSRS fracture corridors; S<sub>2</sub>: SSRS fault zone; S<sub>3</sub>:  
147 SRS fault zone.

148

149

150        **2. Geological setting**

151        The crystalline basement rocks exposed on the island of Bømlo in western Norway belong to the  
152        Upper Allochthon units of the Caledonian orogen (Gee et al., 2008) (Fig. 1a). They mainly consist of  
153        variably metamorphosed igneous, volcanic and sedimentary units derived from the Iapetus oceanic  
154        domain (Roberts, 2003). The study area is located within the Rolvsnes granodiorite (RGD,  $466 \pm 3$   
155        Ma, U/Pb on zircon) (Scheiber et al., 2016), a plutonic body related to pre–Scandian magmatism that  
156        intrudes the ophiolitic, gabbroic, and volcano–sedimentary orogenic units (Slagstad et al., 2011) (Fig.  
157        1b). After its emplacement in the Ordovician, the RGD remained at shallow crustal levels ( $< 10\text{--}15$   
158        km depth), thus escaping pervasive ductile deformation and metamorphic re–equilibration during the  
159        Scandian tectonometamorphic event (Scheiber et al., 2016; Scheiber and Viola, 2018). During the  
160        post–Caledonian orogenic collapse and extension, the Upper Allochthon units in the Bømlo region  
161        belonged to the hanging wall of the Hardangerfjord Shear Zone, a regional NW–dipping extensional  
162        detachment reactivating earlier Caledonian thrusts and nappe contacts (Fig. 1a) (Fossen and Hurich,  
163        2005). The RGD recorded a prolonged and multi–phased brittle deformation history, which has been  
164        characterised in detail by means of structural and geochronological analysis of the different fracture  
165        patterns exposed on Bømlo (Scheiber et al., 2016; Scheiber and Viola, 2018).

166        *2.1 Brittle deformation and alteration history of the Bømlo basement*

167        The latest stages of Caledonian compression during the Mid Ordovician–Silurian are recorded by  
168        ENE–WSW–striking reverse and kinematically consistent conjugate sets of strike–slip shear fractures  
169        and minor faults accompanied by the emplacement of greisens, dykes, and mineralised veins  
170        (Scheiber et al., 2016). A first phase of NW–SE extension occurred during the Devonian in response  
171        to Caledonian orogenic collapse, as recorded by the development of minor, NE–SW–striking normal  
172        faults. Ordovician to Devonian mineralised veins, joints, shear fractures, and minor faults are  
173        pervasive and widely distributed, and they are commonly filled with– and sealed by a variety of  
174        minerals. In the Permian–Early Cretaceous, the area underwent renewed regional extension related

175 to the long-lived series of E–W–directed North Sea rifting events. The first main North Sea rifting  
176 stage is recorded by the development of widespread NE–dipping normal faults active during Permian  
177 to Triassic times (Viola et al., 2016; Scheiber and Viola, 2018). This rifting event led to the formation  
178 of the most prominent structural highs and basins of the North Sea, namely the Stord Basin, Utsira  
179 High, and the Viking Graben (Bell et al., 2014; Scheiber and Viola, 2018). Rifting and crustal  
180 stretching continued during the Late Triassic and Jurassic leading to the formation of ENE–dipping  
181 normal faults associated with the deepening and development of the offshore Viking Graben  
182 (Scheiber and Viola, 2018) and the progressive reactivation of the previously formed NE–dipping  
183 normal faults (Viola et al., 2016). This rifting phase enhanced the rapid exhumation of the Bømlo  
184 crystalline basement from 6–8 km to shallow (<2 km) depths during Permian–to–Jurassic times  
185 (Scheiber and Viola, 2018). From the Late Jurassic onward, the crystalline basement has resided at  
186 very shallow depth (<2 km; Fredin et al., 2017b; Scheiber and Viola, 2018). Far–field stresses related  
187 to the Early Cretaceous rifting stage of the northern North Sea and the Mid–Norwegian margin  
188 subsequently led to the development of N–S–trending fracture corridors in the Bømlo region (Viola  
189 et al., 2016; Scheiber and Viola, 2018).

190 A distinctive characteristic of the Bømlo fractured crystalline basement is the occurrence of alteration  
191 and weathering products spatially related to fracture and fault zones (Viola et al., 2016; Fredin et al.,  
192 2017b; Scheiber and Viola, 2018). Altered/weathered granodiorite is invariably found as lenses along  
193 fractures and variably altered volumes within high–fracture density deformation zones. The dense  
194 network of fractures and fault zones likely promoted fluid infiltration at depth (Place et al., 2016). K–  
195 Ar dating of alteration–related authigenic illite has revealed the occurrence of multiple shallow–  
196 crustal levels alteration/weathering events during the prolonged and multi–phased brittle deformation  
197 history discussed above (Viola et al., 2016; Fredin et al., 2017b; Scheiber and Viola, 2018). A  
198 significant event occurred during the Late Triassic (ca. 220–200 Ma), when the fractured crystalline  
199 basement on Bømlo was exposed to sub–aerial alteration and formation of deep–weathering horizons

200 (saprock – saprolite) related to the Mesozoic tropical–humid climate of the region (Fredin et al.,  
201 2017b). Other alteration events likely occurred in the Permian and Jurassic–Cretaceous (A. Margreth  
202 personal communication; Viola et al., 2016; Scheiber and Viola, 2018). The detailed genetic  
203 processes and geological conditions under which these latter alteration events occurred are still not  
204 well constrained (A. Margreth personal communication). Despite the possible diachronism and  
205 superposition of different alteration/weathering events, all the observed alteration products of the  
206 crystalline basement on Bømlo present similar field textural and mineralogical characteristics.  
207 Progressive disaggregation of the host RGD into sandy “grus–type” aggregates (*sensu* Migoń and  
208 Thomas, 2002) is commonly accompanied by mineral (feldspar) alteration and clay–mineral  
209 authigenesis. Such characteristics are ascribable to a process of arenitisation, kaolinitisation, and/or  
210 saprolitisation of the host RGD occurring at shallow crustal depths (< 6 km) up to surficial conditions  
211 (Viola et al., 2016; Fredin et al., 2017b, 2017a; Scheiber and Viola, 2018). It is indeed quite well  
212 known that hydrothermal kaolinitisation and saprolite formation under tropical–humid climate  
213 conditions may lead to alteration products with similar textural/mineralogical/mechanical  
214 characteristics (Coggan et al., 2013).

## 215 *2.2 Onshore–offshore correlation – Bømlo basement and Utsira High*

216 Alteration characteristics similar to those documented onshore have been detected in offshore  
217 exploratory wells from the Utsira High fractured crystalline basement at variable depth beneath the  
218 sedimentary cover (Riber et al., 2015; Trice et al., 2019). Amidst the central North Sea, the Utsira  
219 High represents a N–S trending basement structural high bounded by the Viking Graben to the west  
220 and the Stord Basin to the east (Fig. 1a, d) (Gabrielsen et al., 2001). The basement unit in its core is  
221 mainly composed of Ordovician–Silurian granitoid and gabbroid rocks (Riber et al., 2015).  
222 Petrological, geochronological, and geophysical data suggest that, prior to the opening of the North  
223 Sea, the crystalline basement of Bømlo and the basement units of the Utsira High likely belonged to  
224 the same unit of the Upper Allochthon (Slagstad et al., 2011; Lundmark et al., 2014; Riber et al.,

225 2015; Fredin et al., 2017b; Lothe et al., 2018; Trice et al., 2019). The seismically imaged top-  
226 basement topography of the Utsira High shows the occurrence of widespread fracture lineaments and  
227 fault zones, resembling the onshore joint-aligned valley morphology of the Bømlo region (Fredin et  
228 al., 2017b). The pervasive and variably oriented fracture lineaments and fault zones thus suggest that,  
229 similarly to the onshore margin, the Utsira High also underwent a prolonged and multi-phased brittle  
230 deformation history (Fredin et al., 2017b). Brittle deformation probably started in the Permian-  
231 Triassic during the first rifting phase of the North Sea, and ended in the Cretaceous-Paleogene with  
232 the northward migration of the rifting activity toward the Mid-Norwegian margin (Bell et al., 2014;  
233 Scheiber and Viola, 2018). Offshore exploration wells have identified the occurrence of a variably  
234 preserved 1–10 m thick weathering profile atop of the fractured granitoid basement (Fig. 2a) (Riber  
235 et al., 2015; Trice et al., 2019). The weathering profile, from top to bottom, consists of (Fig. 2b)  
236 (Riber et al., 2015, 2016, 2017, 2019a): (i) clay-rich, saprolite – incoherent alteration facies A5 of  
237 Riber et al., (2016); (ii) grus-type saprock, in which the granitoid magmatic texture is still preserved  
238 – incoherent alteration facies A3–A4 of Riber et al., (2016); (iii) coherent, partially weathered  
239 granitoid, still preserving the magmatic texture and mechanical properties of the host rock – alteration  
240 facies A1–A2 of Riber et al., (2016). K–Ar dating on authigenic illite from core samples of the  
241 saprolitic level in the weathering profile has shown that the Utsira High crystalline basement was  
242 exposed to sub-aerial conditions and weathering during the Triassic ( $206 \pm 4$  Ma) (Fredin et al.,  
243 2017b).

244 After the Late Triassic, burial and sedimentation affected the whole passive margin (including the  
245 Utsira High and Bømlo region), causing the deposition of a several km thick Mesozoic sedimentary  
246 succession (Sclater and Christie, 1980; Gabrielsen et al., 2001; Bell et al., 2014). The buried Utsira  
247 High fractured-and-weathered basement acted as a system reservoir-trap for oil and currently, thus,  
248 represents an economically viable unconventional oil play (Trice et al., 2019). Offshore exploration  
249 wells have identified exploitable oil resources hosted in the basement-top weathering profile beneath

250 the sedimentary cover (Fig. 2a–b) (Riber et al., 2015, 2017). Pre–Cretaceous SRS faults bound  
251 laterally the reservoir basement blocks, which are overlain by the Jurassic–Cretaceous sedimentary  
252 cover that acted as both lateral source rock and top–seal unit for oil–migration within the structural  
253 high (Trice et al., 2019). Jurassic–Cretaceous sedimentation likely affected also the Bømlo region, as  
254 inferred from the occurrence of Late–Mesozoic sedimentary basins close to the present–day onshore  
255 area (Fredin et al., 2017b, 2017a). Onshore, the Mesozoic sedimentary cover and the underlying  
256 Triassic top–basement weathering profile have been almost completely eroded by geomorphic and  
257 glaciogenic processes during the Cenozoic (Fredin et al., 2017b; Riber et al., 2017). As a  
258 consequence, only the deepest portions of the top–basement weathering profile are locally preserved,  
259 particularly where the effects of weathering penetrated for a few tens of m into the crystalline  
260 basement along discrete structural discontinuities (i.e. fracture and fault zones; Fig. 2c) (Fredin et al.,  
261 2017b). The deepest portion of the weathering profile, preserved in these structurally–controlled  
262 depressions may, therefore, have escaped from significant textural, mechanical, and mineralogical  
263 modifications related to the subsequent Cenozoic erosion, weathering or alteration (Fredin et al.,  
264 2017b, 2017a; Lothe et al., 2018)

265 For these reasons, an increasing number of multidisciplinary studies in the region considers the  
266 fractured and weathered crystalline basement exposed on Bømlo as a reliable onshore analogue of  
267 the offshore crystalline basement of the Utsira High (Fredin et al., 2017b; Lothe et al., 2018; Banks  
268 et al., 2019; Trice et al., 2019).

269

### 270 **3. Methods**

271 Our structural analysis focussed on the identification and characterisation of meso–scale fracture  
272 patterns and the definition of the main structural domains related to faulting. The investigation of  
273 petrophysical (permeability) and geomechanical (uniaxial compressive strength) properties has been  
274 carried out directly in the field on selected representative outcrops. In–situ permeability



275 measurements have been carried out with a New England Research TinyPerm-3 air-  
276 minipermeameter on both fault rocks and alteration products related to basement weathering and  
277 saprolitisation. The minipermeameter is calibrated by the manufacturer with known standards. The  
278 instrument allows a reliable field investigation of rock permeability within small volumes ( $1-1.5 \text{ cm}^3$ )  
279 in the  $10^{-3}$ - $10 \text{ D}$  range, even though controlled laboratory tests have demonstrated its capability to  
280 measure permeability values as low as  $10^{-5} \text{ D}$  (Filomena et al., 2014). In the bulk-rock permeability  
281 mode, the instrument directly yields an estimate of the permeability based on the outgoing air flow  
282 rate from the built-in compression vessel. Permeability values obtained from air-minipermeametry  
283 need to be corrected and standardised in order to be comparable with permeability values obtained  
284 from laboratory tests on rock plugs or image analysis (Fossen et al., 2011; Filomena et al., 2014;  
285 Torabi et al., 2018). Air-minipermeameters yield either larger permeability values (by a factor of  
286 1.7-1.8) when compared with results derived from image analysis quantifications (Fossen et al.,  
287 2011), or lower permeabilities (-37%) than those obtained from small (<10 cm) rock plugs used in  
288 laboratory tests (Filomena et al., 2014). In addition, permeability estimates from permeametry tests  
289 adopting a gas as permeating fluid (as in air-minipermeametry) can be significantly larger than the  
290 estimates from permeability tests relying on a liquid as permeating fluid as a consequence of possible  
291 slip-flow effects (Klinkenberg effect) (van Noort and Yarushina, 2019). This has to be considered  
292 when using absolute gas permeability values to analyse the flow of a liquid in the same porous  
293 medium.

294 The geomechanical characterisation has been carried out with a L-type Schmidt hammer (DRC  
295 GeoHammer, 0.735 N·m impact energy) calibrated according to international standards (ASTM  
296 D5873-00; EN 12 504-2; ASTM C 805-02). The reliability range of the instrument extends between  
297 10 MPa and 300 MPa of UCS (Aydin and Basu, 2005).

298 We ran petrophysical and geomechanical analyses both on compositionally and texturally  
299 homogeneous volumes of the outcrops, thought to be representative of the bulk rock properties, and

300 along transects. Where rock textural anisotropies were present, we performed permeability and  
301 Schmidt hammer analysis both perpendicular and parallel to the dominant planar rock fabric, aiming  
302 at evaluating the potential anisotropy of petrophysical and geomechanical properties. Transects  
303 allowed us to constrain the variation of the analysed properties across weathering fronts and high-  
304 density fracture zones. Surface spot analyses for both permeability and Schmidt hammer rebound  
305 analyses have been done by probing representative outcrop surfaces with at least 10 measurements.  
306 Selected measuring spots were at least 10–15 cm away from mesoscopic fractures wherever possible  
307 (Aydin and Basu, 2005). Both permeability and Schmidt hammer rebound analyses were repeated  
308 twice in the immediate surroundings of the selected spot along transects, for a total of three  
309 measurements. The replicates were not exactly on the same point to avoid results biased by subtle  
310 rock modifications (e.g., compaction and microfracturing) due to the previous measurements. Each  
311 measure spot was kept at c. 5–10 cm from the nearest measuring point both along transects and within  
312 selected areas, and taken to be representative of a 5 cm–radius hemispherical volume below it (ASRM  
313 standards; Aydin and Basu, 2005; Aydin, 2009; Demirdag et al., 2009). UCS values were retrieved  
314 from the analysis of Schmidt hammer rebound values (Aydin and Basu, 2005). Schmidt hammer  
315 measurements were carried out perpendicularly to clear, fresh rock surfaces (wherever possible),  
316 parallel to the horizontal direction. Schmidt hammer rebound values obtained from the in-situ  
317 measurement were converted into UCS by following the hammer orientation-dependent calibration  
318 curves provided by the manufacturer. Horizontal rebound values were converted into UCS by the  
319 following equation:

$$320 \quad \text{UCS} = 0.0232 * R^{2.2637} \text{ (MPa)}$$

321 Presented data are not corrected or standardised, so as to fully capture the heterogeneity and potential  
322 spread of rock mechanical properties (Aydin, 2009).

323 At one outcrop, fracture pattern characteristics have been quantified by using circular scan windows  
324 along a transect previously adopted for petrophysical and geomechanical analysis (Watkins et al.,

2015). The results of fracture intensity ( $P_{21}$ ) quantification from circular scan windows 10-cm-diameter have been compared with the along-transect variation of mechanical strength (Vignaroli et al., 2019). The same outcrop has been analysed also with a 1.4 m radius circular scan window to capture the outcrop scale characteristics of the fracture pattern. A schematic representation of the tectonic fractures of the outcrop has been obtained by line drawing on an outcrop field photo. The resulting line drawing has been analysed with FracPaQ (Healy et al., 2017) in order to quantify fracture intensity and 2D permeability at the outcrop scale in the direction of fluid flow (Long et al., 1982; Watkins et al., 2018).

333

## 334 **4. Results**

### 335 *4.1 Outcrop description*

336 On Bømlø, we have investigated five outcrops representative of the different deformation and weathering product end members, to constrain their mesoscopic structural characteristics as well as their petrophysical and geomechanical properties. Some of the analysed outcrops contain both structural and weathering features. For the clarity of presentation and discussion, outcrops are grouped into Group S (representative of *structural features*, S<sub>1</sub> to S<sub>3</sub>) and Group W (*weathering features*, W<sub>1</sub> to W<sub>3</sub>) (Figs. 1c, 2c–d). Magmatic dykes, greisens and mineralised veins are common in the RGD. The detailed characterisation of their mechanical strength and permeability properties is, however, beyond the scope of this paper.

344 Group S includes outcrops where we measured the geomechanical and petrophysical characteristics of discrete mesoscopic deformation zones (Fig. 2c). Our field investigations focussed mainly on the analysis of fracture corridors and major (either SSRS or SRS) fault zones formed during the Permian to Cretaceous interval, which significantly contributed to the development and modification of the crystalline basement permeability (Scheiber and Viola, 2018; McCaffrey et al., 2020). The pervasive and widely distributed Ordovician to Devonian mineralised joints, shear fractures and minor faults

350 are commonly filled and sealed by different minerals (Scheiber and Viola, 2018), such that their  
351 contribution to the overall current permeability of the crystalline basement is assumed to be very  
352 limited (Watkins et al., 2018; McCaffrey et al., 2020). The field criteria adopted for the classification  
353 of different structures include: (i) the width of the deformation zone, as indicated by the across-strike  
354 thickness of the fractured domain; (ii) the occurrence/lack of fault rocks within the deformation zone;  
355 (iii) the width of the fault rock-bearing deformation zone (fault core). The width of the fault core may  
356 help us to quantify the minimum fault throw, and thus discriminate between SSRS and SRS  
357 deformation zones. The structural features included in Group S are: SSRS fracture corridors ( $S_1$ ), a  
358 SSRS fault zone ( $S_2$ ), and a SRS fault zone ( $S_3$ ). Group S is numbered according to the increasing  
359 size of the studied structural features (fracture/fault zone width) as constrained by outcrop  
360 observations. The analysis of these outcrops allowed us to constrain the mechanical strength and  
361 permeability variation related to: (1) high-intensity fracture zones within (SSRS) fracture corridors;  
362 (2) fault rock development and distribution within SSRS and SRS fault cores.

363 Group W includes outcrops used to characterise the weathering features affecting the crystalline  
364 basement. Each Group W outcrop has already been previously analysed and sampled for illite K–Ar  
365 dating aiming at constraining the age of weathering (Viola et al., 2016; Fredin et al., 2017b; Scheiber  
366 and Viola, 2018). Results from XRD analysis (Table 1) reported in the cited references are here  
367 adopted to qualitatively characterise the mineral composition of weathering products. Group W  
368 includes: outcrop  $W_1$  formed by partially altered RGD with incipient weathering and discolouring  
369 mainly localised along fractures; outcrop  $W_2$  formed by cohesionless, altered RGD, still preserving  
370 the RGD magmatic structure but completely transformed into a sandy aggregate (saprock – grus);  
371 outcrop  $W_3$  formed by cohesionless, altered RGD with increasing clay content (saprolite). The W  
372 outcrops are presented herein from  $W_1$  to  $W_3$  according to the increasing degree of weathering (from  
373 incipient to evolved weathering). The degree of weathering has been qualitatively deduced directly  
374 on the field comparing the mechanical/textural characteristics of the weathering products with those

375 described by Riber et al. (2016) for the alteration facies of the Utsira High core samples. In all the  
376 analysed outcrops, alteration/weathering products invariably postdate the formation of brittle  
377 deformation zones.

378

#### 379 *4.1.1 Group S – Fractures and fault zones within the crystalline basement*

380 **Outcrop S<sub>1</sub>.** This outcrop contains a deformation zone defined by a 10–15 m thick fracture corridor.  
381 The central portion of the fracture corridor is characterised by a 2–5 m thick zone of high fracture–  
382 intensity (average fracture spacing = 0.05–0.1 m), composed of NNE–SSW–trending subvertical  
383 open fractures (Fig. 3a). Mesoscopic fractures are usually longer than 0.5–1 m along the dip direction,  
384 generally subparallel, but still locally forming a well–interconnected fracture network. Their typical  
385 aperture (*sensu* Ortega et al., 2006) is in the order of <1 mm. This deformation zone likely  
386 accommodated minor (<1m) lateral displacement, as inferred by the only local and rare development  
387 of fault striae (mainly strike–slip) on fracture surfaces and the limited offset (c. 6 cm) of crosscut  
388 magmatic markers (Scheiber and Viola, 2018). The outcrop is generally fresh, although fractures are  
389 in some cases filled with granular material probably related to either alteration or presence of fault  
390 rocks along the major fracture planes. The presence of now–eroded saprolite horizons has been  
391 inferred above the outcrop (Fig. 6g–h in Scheiber and Viola, 2018). The finest grain–size fractions of  
392 the alteration rock observed within the fractures are enriched in smectite–like clays of Jurassic age  
393 ( $187.5 \pm 16.7$  Ma; Table 1), as inferred from K–Ar dating on authigenic illite (sample TSC–1 of  
394 Scheiber and Viola, 2018). Alteration of the rock postdates the formation of the fracture corridor,  
395 which is thus older than Jurassic in age. The only limited throw (< 1 m) and width of the fracture  
396 corridor core (< 10 m) suggest that outcrop S<sub>1</sub> represents a SSRS fracture corridor.

397

398 **Outcrop S<sub>2</sub>.** Outcrop S<sub>2</sub> contains two subparallel, well-localised SSRS fault zones characterised by  
399 a 10 cm thick cataclastic fault core (Fig. 6i–j in Scheiber and Viola, 2018). Each fault zone is  
400 characterised by a polished principal slip surface (PSS) oriented 70°/140° (dip/dip direction) (Fig.  
401 3b). The PSS is decorated by transtensive–kinematics slickenlines plunging 30°/070° (plunge/trend).  
402 Several scattered, sub-vertical open fractures surround the fault core, defining a poorly developed 1–  
403 m-thick damage zone. Speculatively, the thickness of the fault core (10 cm) would be consistent with  
404 a fault throw in the order of 1–10 m (Torabi and Berg, 2011). The limited fault core width and  
405 estimated throw suggest that these fault zones may represent SSRS fault zones. The host RGD and  
406 the cataclastic core are locally altered, as documented by the occurrence of an irregular volume of  
407 sandy material (Fig. 3b). The alteration product is characterised by an increasing kaolinite content  
408 over smectite (sample TSC–36 of Scheiber and Viola, 2018). K–Ar dating on authigenic clays  
409 constrains alteration to the Cretaceous ( $127.4 \pm 16.5$  Ma, Table 1). Therefore, the age of faulting must  
410 be older than Cretaceous in age.

411

412 **Outcrop S<sub>3</sub>.** The Goddo Fault Zone (GFZ) crops out along a road cut on the island of Goddo  
413 (northwestern Bømlo, Figs. 1b–4). The GFZ represents a E-dipping normal fault zone formed and  
414 reactivated during Permian-to-Cretaceous extensional tectonics (Viola et al., 2016; Fredin et al.,  
415 2017b; Scheiber and Viola, 2018). It is characterised by two separate and distinct fault cores (marked  
416 as S<sub>3A</sub> and S<sub>3B</sub> in Fig. 4a–b), separated and surrounded by highly fractured damage zones (Figs. 4a–  
417 b, 5a–b). The two fault cores consist of several brittle structural facies (*sensu* Tartaglia et al., 2020),  
418 i.e. different fault rocks characterised by different texture, mineralogical assemblages, and ages of  
419 formation (Viola et al., 2016). Brittle structural facies observed within the fault core include: (a) a  
420 polished PSS overlain by (b) a massive, well-sorted clay-rich gouge, (c) a phyllonitic gouge, and (d)  
421 a cohesive cataclasite (Fig. 5a–d). The PSS of the GFZ consists of a 1–2 cm-thick fine-grained  
422 quartz-coated polished surface, dipping 50°/070° and bearing dip-slip slickenlines (50°/070°)

423 associated with top-to-the-east normal fault steps (Fig. 5c–d). The overlying gouge core is 20–30  
424 cm thick and it is composed of two main gouge layers showing different textures and age of formation  
425 (Viola et al., 2016). The clay-rich gouge consists of a 5–10 cm thick homogeneous and isotropic  
426 plastic gouge (Fig. 5c–d). The phyllonitic-gouge consists of a scaly, phyllosilicate-rich gouge  
427 containing a pervasive S–C' composite fabric (Fig. 5d). K–Ar illite dating showed that the phyllonitic  
428 gouge of the fault core S<sub>3B</sub> developed during Permian extension ( $264.1 \pm 5.4$  Ma; Table 1); Triassic–  
429 Early Jurassic fault reactivation led to the formation of the clay-rich gouge and reworking of the  
430 phyllonitic gouge in the S<sub>3B</sub> fault core ( $200.2 \pm 4.1$  Ma; Table 1) (Viola et al., 2016). The gouges are  
431 overlain by a 20–40 cm thick layer of indurated cataclasite, containing internal discrete shear planes  
432 decorated by purple–reddish–orange coatings of Fe-oxides (Fig. 5d). The occurrence of fault rocks  
433 is limited to the two described fault cores. The phyllonitic gouge is only observed in the S<sub>3B</sub> fault core  
434 (Fig. 5d). No fault breccia bounding the fault core is observed.

435 The thickness of each fault core is in the order of 0.6–1 m (Fig. 5c–d), suggesting that each fault core  
436 may have accommodated a cumulative normal throw of 10–100 m (Torabi and Berg, 2011), thus  
437 defining a valuable example of a SRS fault zone (Viola et al., 2016). Accordingly, the damage zone  
438 width would be in the order of 10 to 100 m (Wilson et al., 2003; Faulkner et al., 2006, 2010). The  
439 damage zone is characterised by densely spaced (average spacing = 0.1–0.5 m) mesoscopic fractures  
440 related to both the background basement fracturing and to the GFZ itself (Scheiber and Viola, 2018).  
441 The GFZ damage zone is well exposed in the footwall, and between fault cores, whereas the geometry  
442 and exposure of the outcrop did not allow us to quantify the exact thickness of the hanging wall  
443 damage zone and the entire fault zone in the field (Fig. 4).

#### 444 *4.1.2 Group W – Weathered outcrops*

445 **Outcrop W<sub>1</sub>**. The outcrop is in the same locality of outcrop S<sub>3</sub>–GFZ. The whole GFZ is crosscut by  
446 a series of poorly exposed alteration/weathering zones (outcrops marked with W<sub>1A</sub>–W<sub>1B</sub> in Fig. 4a;  
447 Fig. 6a–b). These weathering zones are spatially related to NNE–SSW, 1–2 m thick fracture corridors

448 cutting across both the undeformed RGD host (Fig. 6a) and fault rocks (Fig. 6b) (Viola et al., 2016;  
449 Scheiber and Viola, 2018). The granodiorite within the fracture corridor mostly preserves its primary  
450 structure and cohesion at the hand specimen scale. Partially altered host rock lithons within the  
451 fracture corridor are embedded in a cohesionless granular matrix, mainly composed of quartz and  
452 clay minerals, which forms much less than 50 vol% of the altered outcrop (Fig. 6a–b). The  
453 mineralogical composition of the finest fraction of this matrix is dominated by smectite, as inferred  
454 from the sample BO–OFR–1 presented in Viola et al. (2016) and Fredin et al. (2017b). The  
455 weathering material included therein has been dated to the Early Cretaceous ( $125.2 \pm 4.2$  Ma, Table  
456 1). The mineralogical composition and textural characteristics of the weathering products suggest that  
457 this outcrop is equivalent to the alteration facies A1 described by Riber et al. (2016).

458 **Outcrop W<sub>2</sub>.** Outcrop W<sub>2</sub> is characterised by the occurrence of a pervasive NNW–SSE–trending  
459 fracture set (Fig. 6c). Fractures are preferentially organised in 1-m-thick clusters, separated by 2–5 m  
460 of non–fractured RGD host. Commonly, rock volumes within fracture clusters are heavily altered and  
461 transformed in a variably cohesive granular material preserving the magmatic fabric. Measurements  
462 were performed along a transect perpendicular to the main fracture trend, on a vertical surface and  
463 crossing through an alteration zone exhibiting a variable degree of weathering (from left to right in  
464 Fig. 6c). Qualitatively, the outcrop was subdivided into three main domains: (i) partially altered host  
465 granodiorite with some localised mesoscopic fractures (domain *d1* in Fig. 6c), (ii) granular  
466 cohesionless aggregate preserving the granodiorite fabric but no mesoscopic fractures (domain *d2* in  
467 Fig. 6c), (iii) granular aggregate with increasing clay content (domain *d3* in Fig. 6c), as inferred from  
468 plastic deformation of hand samples. Unpublished K–Ar illite dating data suggest that this alteration  
469 is older than Triassic and its mineralogical composition dominated by smectite (A. Margreth personal  
470 communication). As reported earlier, similar weathering characteristics have been observed in  
471 outcrop S<sub>2</sub>.



472 **Outcrop W<sub>3</sub>**. The outcrop is characterised by a 2 m thick altered rock volume within granodiorite,  
473 bounded by N–S trending fractures. The altered volume consists of a cohesionless granular aggregate,  
474 grading from saprock to mature, fine–grained clay–rich saprolite, enveloping variably altered “core  
475 stones” (i.e., remnant blocks of granodiorite less weathered than the embedding material and resulting  
476 from spheroidal weathering processes; cfr. Ryan et al., 2005; Fig. 6d) (Fredin et al., 2017b, 2017a).  
477 Weathering products in outcrop W<sub>3</sub> are well represented by the samples (Bomlo2–3–4) analysed by  
478 Fredin et al. (2017a,b), with a predominantly kaolinitic composition and likely developed during sub–  
479 aerial exposure of the crystalline basement in the Triassic (220–200 Ma, Table 1; Fredin et al., 2017b).

480

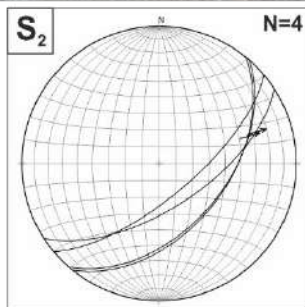
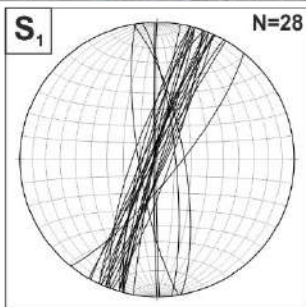
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Outcrop	Sample ID	Grain Size (µm)	XRD Data								K-Ar Dating		Ref.	
			Quartz	Kaolin	Illite/Mica 2M1	Illite/Mica 1M	Diocahedral smectite	Interstratified illite/smectite	Albite/Anorthite	Anatase	Lepidocrocite	Age (Ma)		Error (Ma)
S1	TSC-1	0.1 - 0.4		5			95		<1			187.5	16.7	[3]
		0.4 - 2	<1	9			87		4			192.6	13.7	
		2 - 6	1	18			75		6			260.4	8.1	
S2	TSC-36	<0.1		7			93					127.4	16.5	[3]
		0.1 - 0.4		25			75		<1	<1		177.6	16.2	
		0.4 - 2	<1	29	2		67		2	<1		261.7	8.4	
		2 - 6	<1	22	4		71		3	<1		324.0	9.0	
S3	BO-GVI-1	<0.1		1	20			79				200.2	4.1	[1]
		0.1 - 0.4		5	21		6	68				218.9	4.4	
		0.4 - 2	3	16	28		24	29		<1		240.7	5.0	
		2 - 6	34	8	14		13	30	1			272.1	5.5	
		6 - 10	35	8	17		12	27	1			282.9	5.7	
	BO-GVI-2	<0.1	<1	12	14	4	20	49	1			264.1	5.4	[1]
		0.1 - 0.4										262.0	5.3	
		0.4 - 2										303.8	6.3	
		2 - 6	2	9	32	2	16	35	4			349.6	7.1	
		6 - 10	2	9	33	4	17	31	4			354.4	7.1	
W1	BO-OFR-1	<0.1		7			93		<1			125.2	4.2	[1]
		0.1 - 0.4		7			93		<1			121.4	5.3	
		0.4 - 2		16			82		2			167.7	6.5	
		2 - 6	<1	22	3		71		3	<1		272.1	5.8	
		6 - 10	<1	25	2		68		4	1		287.5	6.2	
W3	Bomlo 2	<0.1					89				11	210.0	13.1	[2]
		0.1 - 0.4		4			78		2		16	290.2	10.6	
		0.4 - 2	<1	7			83		3		7	307.5	38.3	
		2 - 6	<1	8	6		79		3	<1	4	406.9	20.3	
	Bomlo 3	<2	<1	76	1		23					217.3	4.8	[2]
		2 - 6	<1	72	3		24							
	Bomlo 4	<2	<1	53			46		1			233.0	5.0	[2]
		2 - 6	<1	53	2		43		2					

482

483 **Table 1.** XRD data and K–Ar dates from synkinematic/authigenic illite samples of the outcrops  
484 studied in this paper. References: [1] Viola et al. (2016); [2] Fredin et al. (2017b); [3] Scheiber and  
485 Viola (2018).

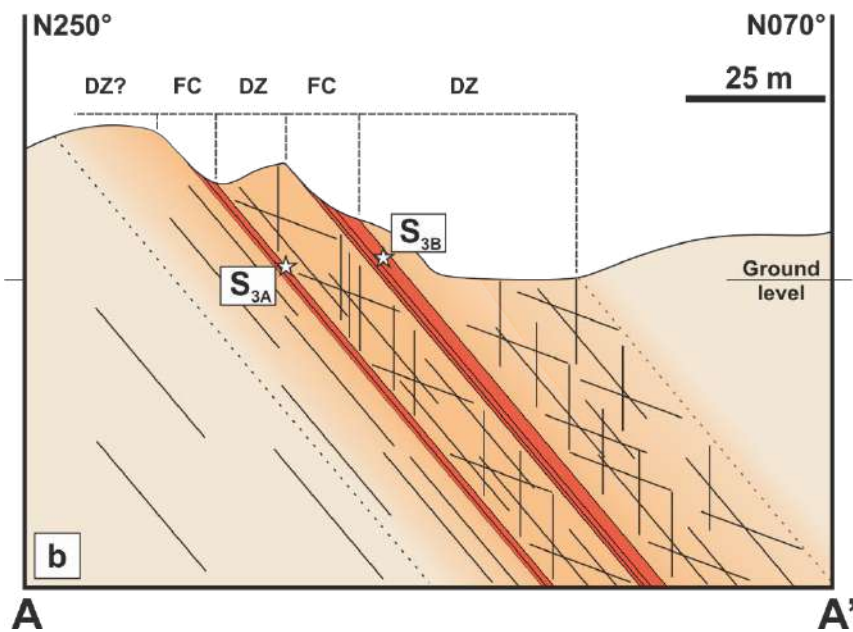
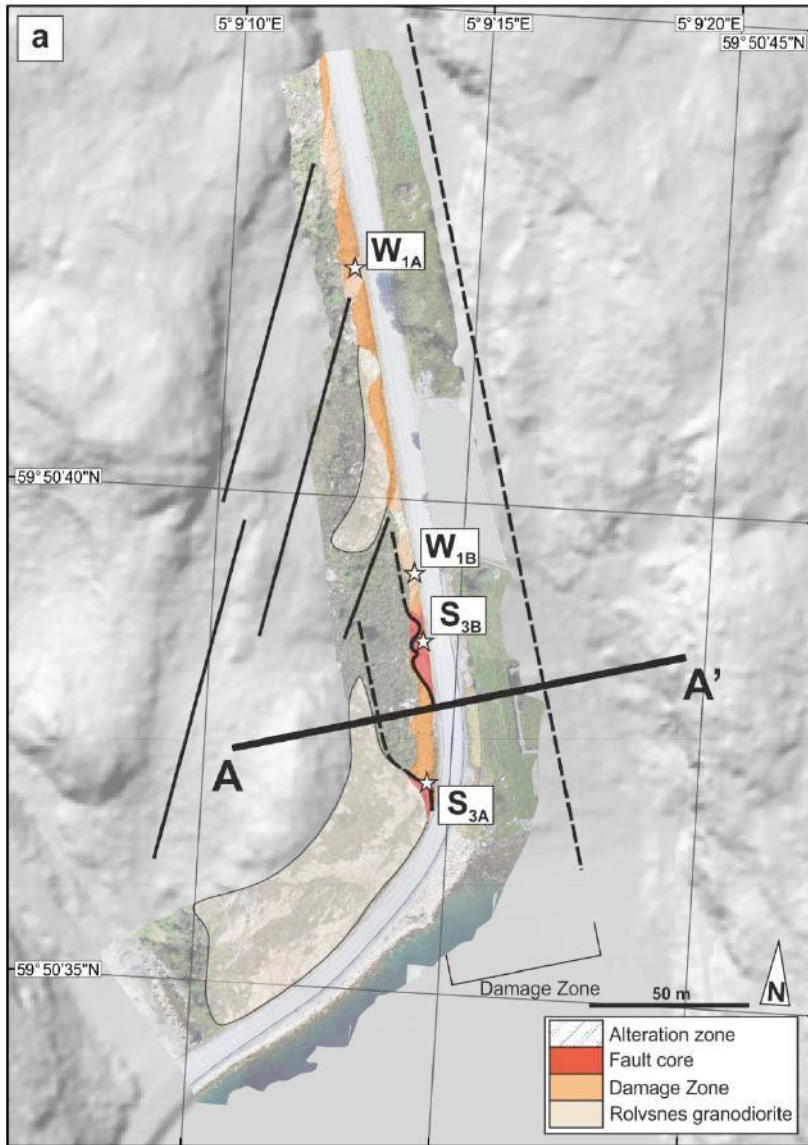
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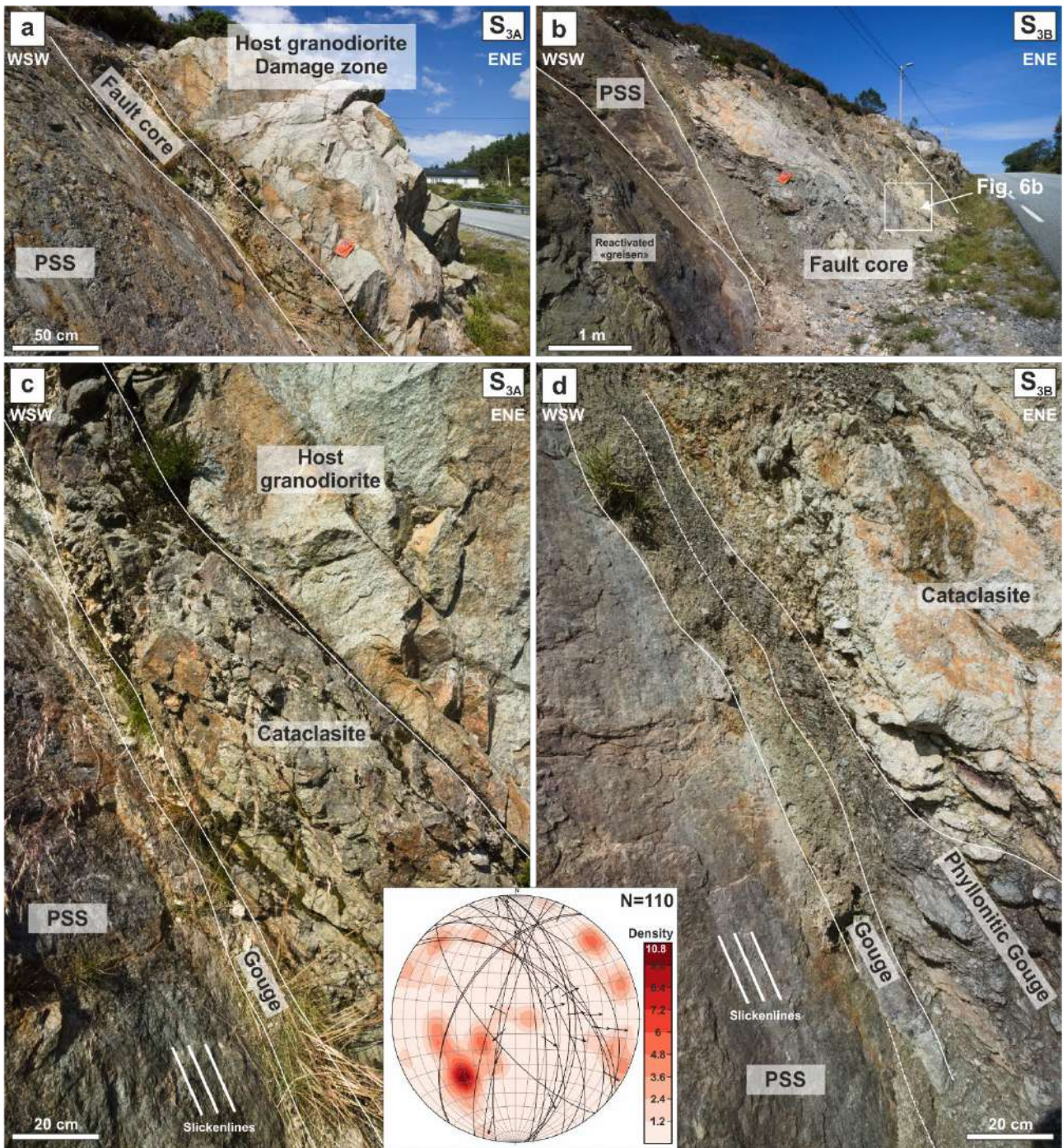
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489 **Figure 3:** (a) Fracture corridor at outcrop S<sub>1</sub>. (b) The fault zone of outcrop S<sub>2</sub> is characterised by a  
490 thin, well-localised, gouge- and cataclasite-rich fault core enveloped by a 1 m thick, poorly-  
491 developed damage zone. Fluids percolating through the fault core led to the alteration of cataclasite  
492 and the damage zone observed in the central portion of the image. PSS: Principal Slip Surface.  
493 Stereonets: Lower hemisphere, equal area projection of fractures and fault planes. The black arrows  
494 represent the direction of movement of the hanging wall along the fault plane.



496 **Figure 4:** (a) Schematic map of the Goddo Fault Zone (outcrop S<sub>3</sub>) overlain on a composite UAV  
497 orthophoto and LiDAR digital elevation model image. The location of the studied outcrops along the  
498 fault zone (S<sub>3A</sub>, S<sub>3B</sub>, W<sub>1A</sub>, W<sub>1B</sub>) is also reported. Black thick lines represent fracture lineaments  
499 spatially related to discrete alteration zones. The black dashed lines bracket the supposed maximum  
500 width of the damage zone. (b) Geological cross section along the A–A' profile of the Goddo Fault  
501 Zone. FC: Fault core; DZ: Damage zone.

502

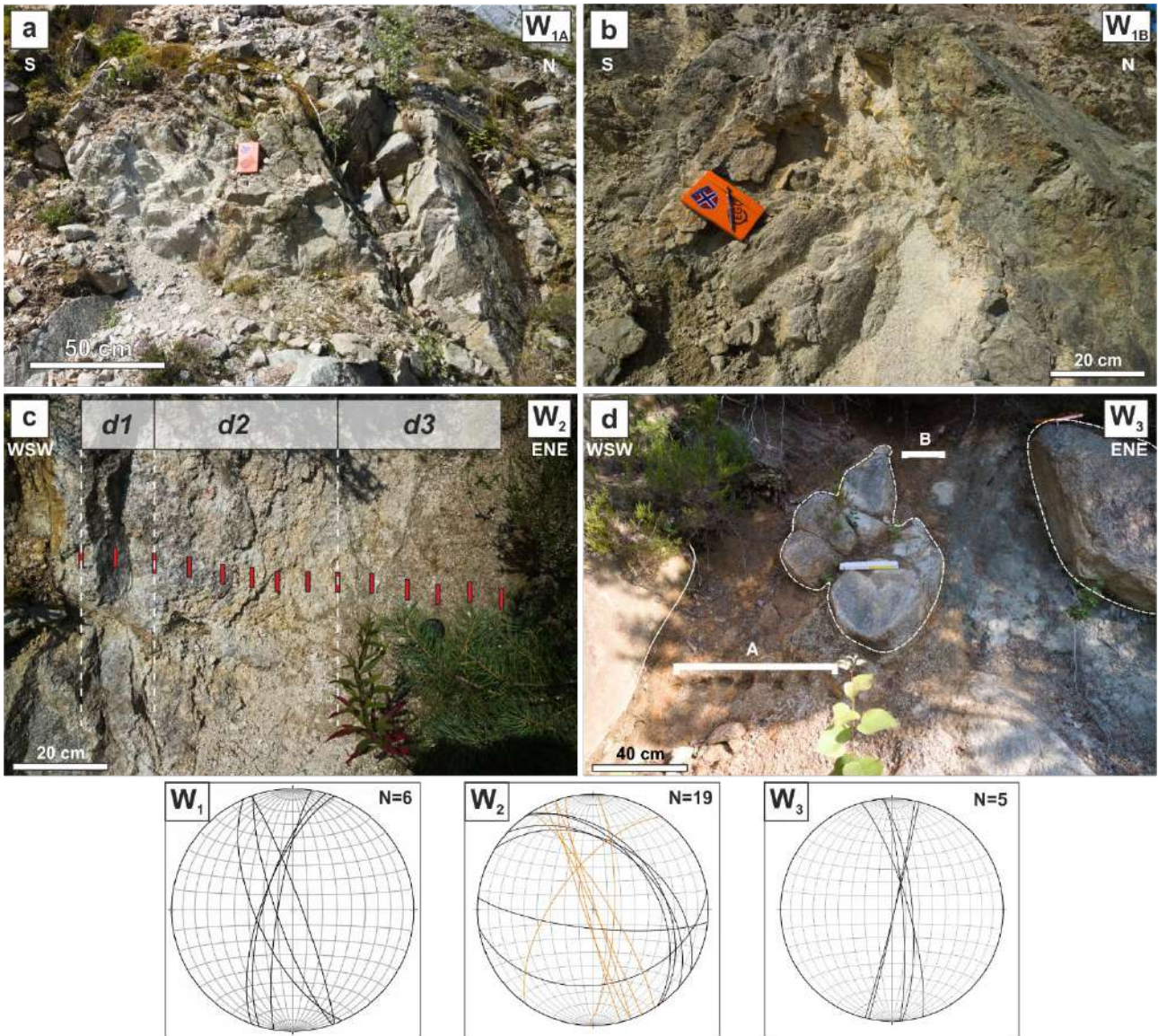


503  
 504 **Figure 5:** Representative outcrops of the Goddo Fault Zone. (a) Fault core at outcrop S<sub>3A</sub>, showing  
 505 the composite fault core. (b) Composite fault core and the reactivated slip surface of a “greisen” pre-  
 506 faulting fracture exposed at S<sub>3B</sub>. (c) Composite fault core at S<sub>3A</sub>, showing the juxtaposition of  
 507 cataclasite, gouges, and PSS. (d) Composite fault core at S<sub>3B</sub>, showing the juxtaposition of cataclasite,  
 508 gouge, and PSS. Note the alteration of the cataclasite layers. Stereonets: Lower hemisphere, equal  
 509 area projection of fractures and fault planes at S<sub>3</sub>. The whole dataset of fracture and fault planes

510 collected on the field is reported as contoured distribution of poles to fracture and fault planes (in  
511 red). Black great circle represent only those fracture and fault planes characterised by slip lines  
512 (lineations, slickenlines). The black arrows represent the direction of movement of the hanging wall  
513 along the fault plane.

514





515  
 516 **Figure 6:** Representative outcrops of weathered granodiorite. (a) Alteration zone  $W_{1A}$  related to a  
 517 NNE-SSW-trending fracture lineament. (b) Alteration of the host granodiorite within the damage  
 518 zone of the GFZ at  $W_{1B}$  (detail of Fig. 5b). (c) Outcrop  $W_2$  showing an increasing alteration profile  
 519 along which we performed Schmidt hammer analysis and permeability measurements. (d) Outcrop  
 520  $W_3$  (saprolite) showing relatively unaltered “core stones” (enclosed by dashed white lines). The thick  
 521 white lines (A, B) represent the profile along which permeability and UCS measurements have been  
 522 performed. Stereonets: Lower hemisphere, equal area projection of fracture planes at outcrops  $W_1$ ,  
 523  $W_2$  and  $W_3$ . Orientation of fractures within fracture clusters related to weathering zones at outcrop  
 524  $W_2$  are highlighted in orange in the stereoplot.



526 *4.2 UCS and permeability data*

527 Data are presented in the following according to rock type (host rock, fault rocks and weathering  
528 products) as observed from many different outcrops, to derive an average value and constrain the  
529 natural variability range for each lithology. Average values of UCS and related standard deviations  
530 are reported in the text and Fig. 7a. Permeability ranges are presented in the text and data distributions  
531 are reported as boxplots in Fig. 7b.

532 *4.2.1 Host rock*

533 The RGD on the island of Bømlo still well preserves its magmatic texture and mineralogical  
534 composition. No pervasive ductile fabrics have been observed at the analysed outcrops, confirming  
535 the observations of Scheiber and Viola (2018). UCS values for the RGD host as observed at three  
536 different localities (outcrops S<sub>1</sub>–S<sub>2</sub>–W<sub>3</sub>). Fresh, non–altered and non–mineralised granodiorite  
537 surfaces yield an average UCS value of  $169 \pm 36$  MPa. Granodiorite surfaces bounding  
538 alteration/weathering zones (measured at outcrops W<sub>2</sub>–W<sub>3</sub>) exhibit a lower range of UCS variability,  
539 with an average value of  $107 \pm 29$  MPa (“Weathered Granodiorite surface” in Fig. 7a). Results from  
540 permeability measurements on the pristine, non–fractured host granodiorite at different outcrops  
541 range between  $<10^{-4}$  D (below the actual reliability limit of the air–minipermeameter) and 0.1 D (Fig.  
542 7b).

543

544 *4.2.2 Outcrop S<sub>1</sub> – SSRS Fracture corridor*

545 UCS measurements are from along a transect perpendicular to fracture strike. In order to evaluate the  
546 influence of fracture intensity on the UCS, we have quantified fracture intensity with circular scan  
547 windows (10 cm in diameter) centred on each measurement spot along the transect (Fig. 8a) (Watkins  
548 et al., 2015). UCS values comparable to fresh and slightly altered host RGD are measured in the  
549 outermost portion of the fracture corridor (Fig. 8b–c). Mechanical strength drastically decreases to  
550 15–20 MPa toward the centre of the fracture corridor, where fracture intensity is higher ( $P_{21} = 74 \text{ m}^{-1}$

551 <sup>1</sup>; Fig. 8b–c, e). The permeability is inferred to be related to only meso–scale fracture aperture and  
552 interconnectivity. In order to have a relative comparison with the permeability data presented in the  
553 next sections, 2D structural permeability at the scale of the outcrop has been evaluated by analysing  
554 the digitised fracture network obtained from line drawing on a field photo of the core zone with the  
555 software FracPaQ (Healy et al., 2017). During line drawing, only the subvertical fractures were  
556 considered for the definition of the fracture network pattern (Fig. 8d). The subhorizontal fractures are  
557 likely related to post–tectonic outcrop modifications related to either post–glacial unloading or man–  
558 made excavation and blasting operations, given that such fracture orientation is not observed in any  
559 other fracture set related to the brittle deformation history of the Bømlo crystalline basement  
560 (Scheiber and Viola, 2018). The 2D permeability of the fracture pattern in the direction of flow was  
561 calculated considering a range of different values of fracture aperture, ranging from 0.01 to 1 mm and  
562 perfect connectivity among fractures. The outcrop–scale maximum permeability is always oriented  
563 parallel to the dominant subvertical fractures, and maximum values range between  $10^3 D$  to  $10^{-3} D$ ,  
564 for a fracture aperture of 1 mm and 0.01 mm, respectively (Fig. 8e).

565

#### 566 *4.2.3 Outcrops S<sub>2</sub>–S<sub>3</sub> –Fault rocks*

567 UCS bulk rock data on fault rocks have been measured at the S<sub>2</sub> and S<sub>3</sub> outcrops. The PSS is  
568 characterised by high UCS values when analysed normal to the surface ( $274 \pm 32$  MPa, Fig. 7a).  
569 Permeability measured normal to the slip surface is in the 0.05 to 0.06 D range (Fig. 7b). UCS of the  
570 clay–rich gouge layer is  $12 \pm 2$  MPa, measured parallel to fault strike. Gouge layers present a lower  
571 range of permeability, between  $10^{-3} D$  and 0.4 D measured parallel to fault strike (Fig. 7b). The rough  
572 exposure surface of the phyllonitic gouge facies did not allow us to perform reliable measurements  
573 (Fig. 5d). The cataclasite shows strongly anisotropic UCS, as inferred comparing the results obtained  
574 from measurements performed parallel ( $36 \pm 22$  MPa) and perpendicular ( $195 \pm 37$  MPa) to fault  
575 strike (Fig. 7a). The cataclastic core presents a variable permeability ranging between  $10^{-3} D$  and 3 D

576 when measured parallel to fault strike (Fig. 7b). The rough exposure surface of the cataclasite did not  
577 allow us to perform reliable measurements of permeability perpendicular to fault strike.

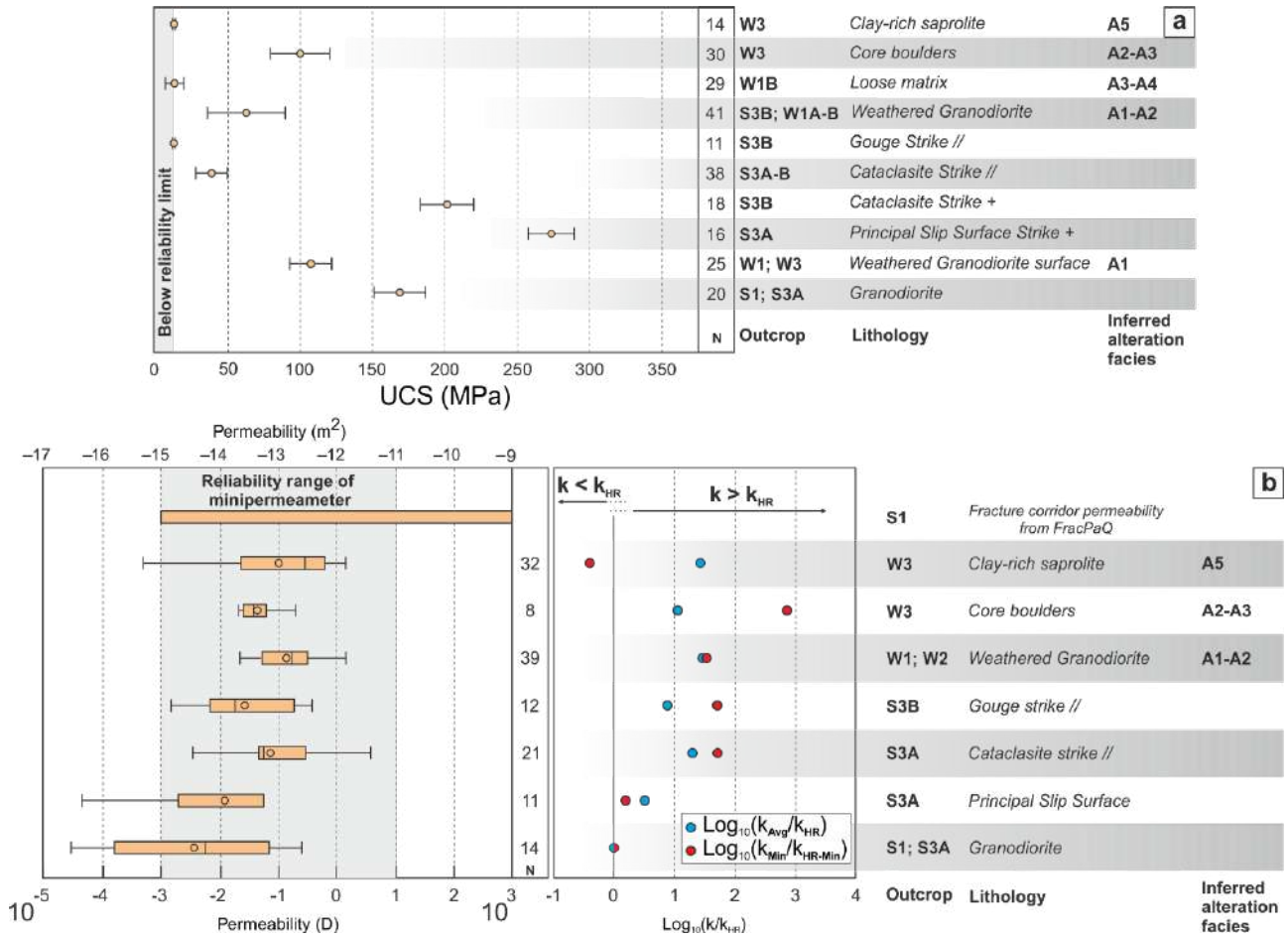
#### 578 4.2.4 Group $W_1$ – $W_3$ . Weathering along fractures and fault zones

579 **Outcrop  $W_1$ .** UCS for altered granodiorite ranges between 30 and 90 MPa (“Weathered  
580 Granodiorite” in Fig. 7a). Values larger than 100 MPa are found in the weathered host granodiorite  
581 where the alteration is locally less developed. In the fracture corridor, altered granodiorite lithons are  
582 surrounded by a fine-grained sandy, rather non-cohesive matrix that exhibits low UCS values (<15  
583 MPa) (“loose matrix” in Fig. 7a). The host granodiorite altered lithons exhibit a variable permeability,  
584 ranging between  $10^{-3}$  D and 0.84 D (Fig. 7b).

585 **Outcrop  $W_2$ .** Outcrop  $W_2$  displays a clear variation of mechanical strength and permeability with  
586 increasing weathering grade (Figs. 6c, 9a). Incipient alteration in domain  $d1$  exhibits variable UCS  
587 between 20 and 50 MPa and variable permeability ranging between  $10^{-3}$  D and 1 D. Similar values  
588 were measured from the altered RGD observed at outcrop  $S_2$ . The cohesionless alteration products in  
589 domain  $d2$ , consisting of a grus-type sandy aggregate, yields UCS values <20 MPa. Permeability  
590 increases (from 0.1 D up to several Darcy’s) and then decreases (from >1 D down to 0.01–0.1 D)  
591 moving from the left-hand side boundary to domain  $d1$  toward the right-hand side boundary to  
592 domain  $d3$  (Figs. 6c, 9a). Alteration domain  $d3$  is characterised by variable UCS < 20 MPa, and low  
593 permeability (from 0.01–0.1 D down to few mD).

594 **Outcrop  $W_3$ .** Partially weathered “core stones” exhibit a broad range of UCS values, ranging between  
595 75 MPa and 125 MPa (Fig. 7a). They have a permeability included between 0.01 D and 0.1 D (Fig.  
596 7b). The clay-rich saprolite volume has been analysed along two transects (Fig. 6d). Both transects  
597 yield almost constant UCS and permeability values within the clay-rich saprolite (UCS = 10–15 MPa,  
598  $k = 10^{-2}$ – $10^{-3}$  D) (Fig. 9b).

599

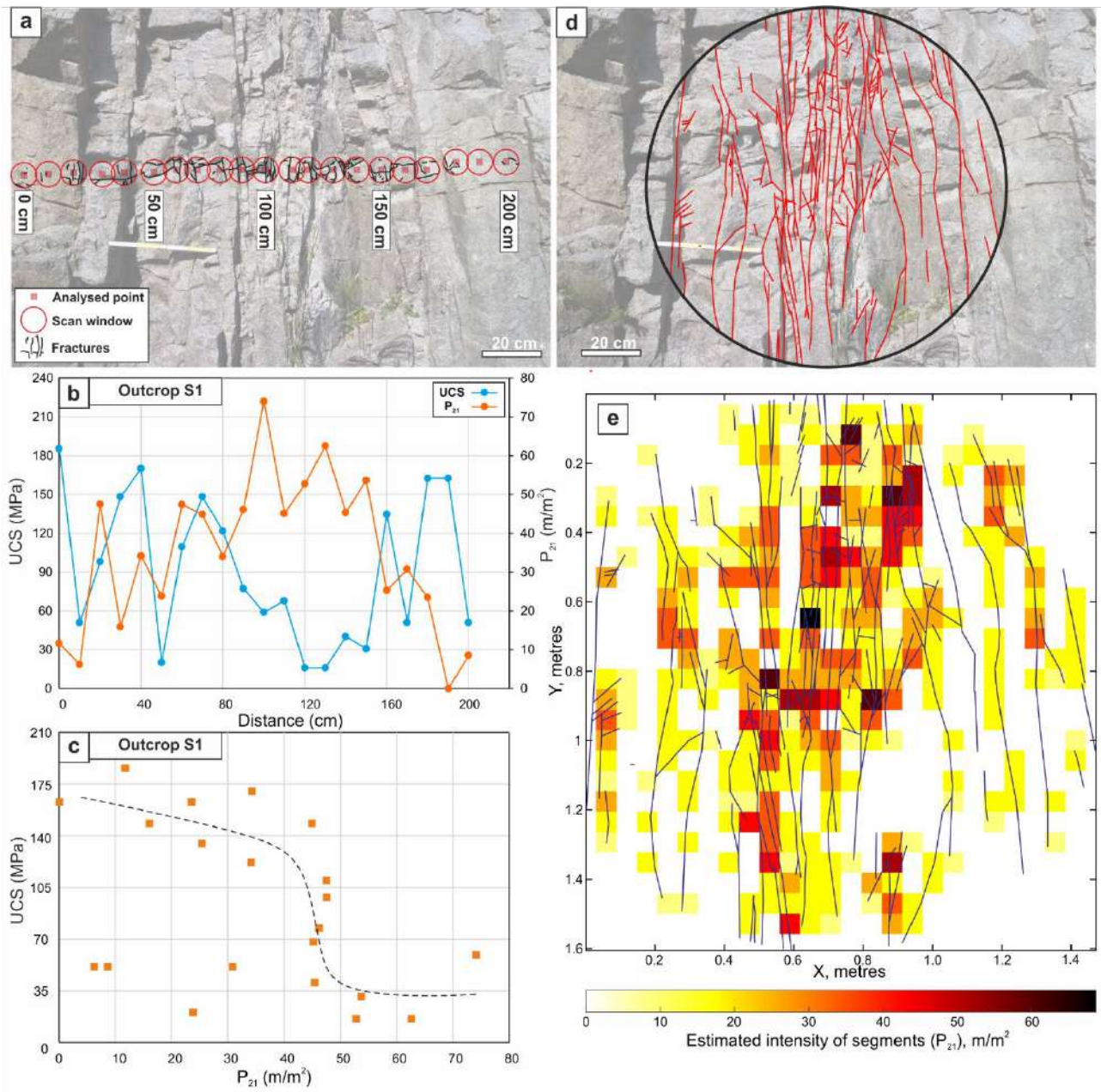


601

602 **Figure 7:** Mechanical strength and permeability box-plot diagrams. “Strike //” stands for data  
 603 measured parallel to the fault plane. “Strike + ” stands for data measured perpendicular to the fault  
 604 plane. (a) Uniaxial Compressive Strength (UCS, MPa). For each analysed lithology we report the  
 605 mean (orange dot), the standard deviation ( $1\sigma$ , black bars), the number of measurements (N), the  
 606 outcrop(s) name on which the measurements were performed, and the corresponding alteration facies.  
 607 (b) Permeability ( $m^2$ , D). The outcrop name and the corresponding alteration facies are reported. N:  
 608 number of measurements for each lithology. Each box of the box-and-whiskers plot represents the  
 609 range between the 1<sup>st</sup> and 3<sup>rd</sup> quartile of the distribution. The whole data range is represented by the  
 610 extension of the whiskers. In (b) we also plot the logarithm of the permeability ratio ( $\text{Log}_{10}(k/k_{HR})$ )  
 611 between the host rock and the selected lithology (Scibek, 2020). The ratio between mean values of  
 612 permeability (blue dots) and the ratio between the minimum reported values (red dots) for each

613 lithology are also reported. It is worth noting that air–minipermeametry retrieved permeability values  
614 for saprolites are comparable to those obtained from laboratory measurements on saprolite samples  
615 from Bømlo and the Utsira High exploration wells (Lothe et al., 2018).

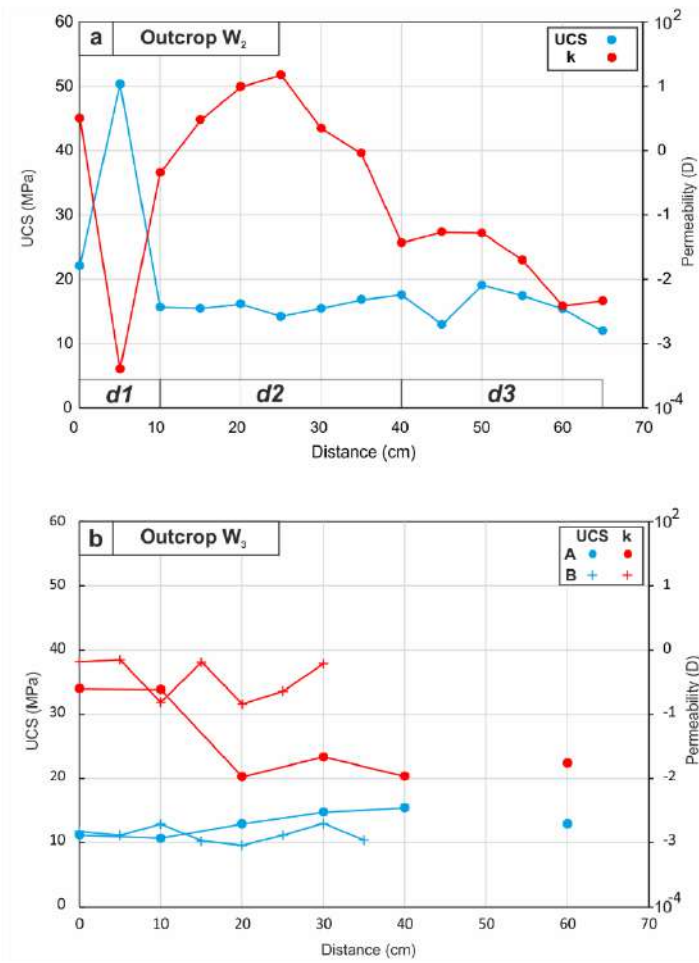
616



617  
 618 **Figure 8:** Results of the geomechanical and permeability measurements at outcrop S1. (a) Location  
 619 of the measurements for the Schmidt hammer analysis, and circular scan windows for fracture  
 620 intensity  $P_{21}$  quantification; (b) Variation of mechanical strength (UCS) and fracture intensity ( $P_{21}$ )  
 621 along the analysed transect. (c) UCS against fracture intensity  $P_{21}$ . Note the drastic decrease of  
 622 mechanical strength after a threshold value of  $P_{21}$  at about  $45 m^{-1}$ . (d) Line drawing of the subvertical  
 623 fracture network analysed with FracPaQ (Healy et al., 2017); (e) Results of 2D permeability and  
 624 fracture intensity quantification from FracPaQ..

625





626

627 **Figure 9:** Results of mechanical strength and permeability quantification along the transect at  
 628 outcrops (a) W<sub>2</sub> and (b) W<sub>3</sub>. Letters A and B in (b) refer to the two transects shown in Fig. 6d.

629

630

631       **5. Discussion**

632       We have characterised the structural and weathering features controlling the matrix permeability and  
633       mechanical properties of the fractured and altered crystalline basement at a number of representative  
634       outcrops. In the following paragraphs we discuss (i) the host rock properties, (ii) the matrix  
635       permeability related to SSRS and SRS deformation zones, (iii) the mechanical strength and  
636       permeability related to weathering products, and (iv) their bearings on the definition of reservoir rock  
637       in crystalline basement structural highs.

638       The data presented in this paper represent the present-day, actual mechanical strength and  
639       permeability at surficial, ambient conditions. The extrapolation of these values to depth (at deeper  
640       structural levels or buried below a sedimentary cover) needs to take into consideration the effect of  
641       increasing confining pressure and temperature and the compaction processes (Rutqvist and  
642       Stephansson, 2003; Faulkner, 2004; Faulkner et al., 2006; Lothe et al., 2018). A further complication  
643       arises from the fact that the analysed onshore outcrops may have been affected by renewed alteration  
644       related to Cenozoic geomorphic and glaciogenic processes, which the fractured and weathered  
645       crystalline basement buried in the North Sea by the Mesozoic cover has, instead, probably escaped.  
646       The petrophysical and mechanical data presented here, however, are comparable (on average) with  
647       other results in the literature from standardised laboratory tests on fresh samples from onshore  
648       outcrops and offshore core samples of the same rock types analysed here (Lothe et al., 2018; Walter  
649       et al., 2018; Høien et al., 2019). This suggests that possible Cenozoic geomorphic and glaciogenic  
650       processes have only slightly affected the textural, mechanical and petrophysical characteristics of the  
651       studied onshore outcrops.

652       *5.1 Host rock properties*

653       The observed host granodiorite UCS is comparable to the reference values reported in the database  
654       by SINTEF for Norwegian rock types (Høien et al., 2019). Permeability data from the RGD are quite  
655       different from available laboratory permeability measurements on intact rocks (Sibson and Rowland,

656 2003). Most of the host rock permeability values are comparable with permeability values obtained  
657 from fractured crystalline and igneous rock in fault damage zones (Forster and Evans, 1991; Evans  
658 et al., 1997; Cappa and Rutqvist, 2011; Gomila et al., 2016). This discrepancy might be related to  
659 either: (i) an analytical error during measurement (i.e. air slippage from the non-perfectly sealed  
660 contact between the probe tip and the rough surface); (ii) enhanced permeability related to the  
661 increased microfracture density in the proximity of fracture and fault zones (Mitchell and Faulkner,  
662 2012; Belaidi et al., 2018; Torabi et al., 2018; Staněk and Géraud, 2019); (iii) enhanced permeability  
663 related to recent, very local surficial weathering processes. Therefore, further investigations are  
664 needed to properly constrain the permeability of the host RGD.

## 665 *5.2 Group S – Mechanical strength and permeability of deformation zones*

666 The host crystalline basement (RGD) is overprinted by several sets of fractures, including fracture  
667 corridors and SSRS–SRS fault zones (Gabrielsen and Braathen, 2014; Scheiber and Viola, 2018).  
668 These fracture lineaments affect differently the mechanical and permeability properties of the host  
669 RGD and thus require a separate discussion.

### 670 *5.2.1 SSRS fracture corridors*

671 Fracture corridors are characterised by high–fracture–intensity deformation zones, accommodating  
672 limited ( $< 1\text{m}$ ) throw, mostly barren of fault rock. The outcrop mechanical strength and bulk  
673 permeability are mainly controlled by the spatial distribution of meso–scale fractures (Fig. 8c, e). The  
674 increase in local fracture intensity leads to a drastic decrease of the host rock mechanical strength for  
675  $P_{21} > 40\text{--}50 \text{ m/m}^2$  (Fig. 8b–c). High fracture intensity leads to a significant increase of fracture  
676 permeability up to  $1\text{--}10^3 \text{ D}$  (Fig. 7b). The largest 2D permeability in the direction of flow is observed  
677 parallel to the longer, subvertical and well interconnected fractures (Fig. 8e). Such a fracture network  
678 represents a preferential pathway for fluid flow (Souque et al., 2019). Thus, the permeability of  
679 limited–throw SSRS fracture corridors is controlled by the mesoscopic fracture porosity. On the other  
680 hand, microscopic fracture porosity may contribute to the matrix mechanical strength and

681 permeability to only a very limited extent (Faulkner et al., 2006; Mitchell and Faulkner, 2012; Rempe  
682 et al., 2018). A local increase in permeability related to microscopic fracture porosity is reported by  
683 Torabi et al. (2018) for fracture corridors developed within the Øygarden Gneiss Complex in the  
684 Bergen area. Therefore, such SSRS, limited–throw fractures remarkably affect both the overall  
685 mechanical and permeability properties of the crystalline basement.

### 686 5.2.2 SSRS and SRS fault cores

687 Strain accommodation and development of fault rocks along SSRS–SRS fault zones resulted in a  
688 drastic modification of textural, mechanical and petrophysical properties of the host RGD. Fault  
689 zones, either SSRS or SRS, are characterised by the development of a variably thick fault core,  
690 defined by the occurrence of fault rocks, and a variably developed damage zone (Caine et al., 1996).  
691 SSRS fault cores are likely to be characterised by thin, single fault cores (< 1m thickness) surrounded  
692 by poorly developed damage zones (e.g., outcrop S<sub>2</sub>). SRS fault zones, instead, may be characterised  
693 by the occurrence of thicker ( $\geq 1$  m), multi–strand fault cores surrounded by complex, thick and high–  
694 fracture–intensity damage zones (Walsh et al., 1999; Gabrielsen and Braathen, 2014).

695 In both SSRS and SRS fault zones, the fault core is characterised by a composite fault plane–parallel  
696 sequence of fault rocks including a PSS, cataclasite, and potentially (several) gouge layers. Extreme  
697 strain localisation and rock comminution resulted in the formation of a thin, yet mechanically strong,  
698 PSS with a low permeability (Fig. 7a,b). Cataclasite mechanical strength is strongly anisotropic (Fig.  
699 7a). Cataclasites are stronger (UCS  $\gg 100$  MPa) where measured normal to the fault plane and  
700 weaker (UCS <50 MPa) where measured parallel to the fault plane. Cataclasite permeability is up to  
701 two orders of magnitude larger than the average permeability of the host rock (Fig. 7b). Gouge layers  
702 yield a very–low mechanical strength (UCS < 20 MPa), and low permeability parallel to the fault  
703 plane ( $k < 10^{-1}$  D). A strong anisotropy in permeability is expected for both cataclasite and gouge  
704 layers (up to three orders of magnitude) (Faulkner and Rutter, 1998). The lower strength and increased  
705 permeability parallel to the fault plane is probably also related to a higher density of fault–parallel

706 microfractures and micro–shear planes within cataclasite and gouge layers (Zhang and Tullis, 1998).  
707 A pervasive phyllosilicate–bearing foliation characterises the phyllonitic gouge of the GFZ. Thus,  
708 even though we lack direct measurements, a strong permeability and strength anisotropy should be  
709 expected for the phyllonitic gouge (Shea and Kronenberg, 1993; Niemeijer and Spiers, 2005; Leclère  
710 et al., 2015). Cataclasites and gouge layers are in any case two orders of magnitude more permeable  
711 than the host RGD (Fig. 7b). Grain–size reduction, cataclasis and microfracturing processes likely led  
712 to the increased micro–fracture–related porosity of fault rocks with respect to that of the host  
713 granodiorite (Staněk and Géraud, 2019).

### 714 *5.2.3 Damage zones*

715 The discrete and multi–strand fault cores are surrounded by variably developed fracture–dominated  
716 damage zones, whose width is one order of magnitude larger than fault core thickness (Faulkner et  
717 al., 2010). The mechanical strength and bulk permeability of the damage zone is strictly dependent  
718 on the intensity, mechanical properties and geometry of multi–scale fractures related to both the fault  
719 zone and the basement background fracturing (Bruhn et al., 1994; Faulkner et al., 2006;  
720 Gudmundsson et al., 2010; Mitchell and Faulkner, 2012; Rempe et al., 2018). Local permeability  
721 within the damage zone might be as large as the permeability of similarly high–fracture–intensity  
722 fracture corridors (Gabrielsen and Braathen, 2014). An accurate characterisation of the fracture  
723 pattern geometry is needed to properly quantify the effect of mesoscopic fractures on the evolution  
724 of mechanical properties and bulk permeability in a damage zone (Healy, 2008; Gudmundsson et al.,  
725 2010).

726 In conclusion, SSRS and SRS deformation zones have distinctive structural characteristics controlling  
727 their permeability structure. Low–strain, low–displacement SSRS fracture corridors characterised by  
728 open, non–mineralised fractures greatly enhance the permeability of the crystalline basement at the  
729 sub–seismic–resolution scale. They form very localised, narrow (1–10 m in width) highly–efficient  
730 conduits for fluid flow. The largest permeability is observed parallel to the average fracture plane

731 orientation. SSRS and SRS fault zones are characterised by two domains with characteristic  
732 permeability (Caine et al., 1996; Evans et al., 1997): (i) the fault core, including low-permeability  
733 fault rocks; (ii) the damage zone, defined by a high-permeability multiscale fracture network. The  
734 permeability of fault rocks and damage zones is in any case several orders of magnitude larger than  
735 the host rock. Therefore, despite their effective (SSRS or SRS) or relative (fault core/damage zone  
736 width ratio) size characteristics, fault zones developed within the granodioritic crystalline basement  
737 of Bømlo are likely to act invariably as preferential conduits for fluid flow. Fluid flow will be  
738 promoted in any directions parallel to the fault zone, along the highly fractured damage zones. Fluid  
739 flow from footwall to hanging wall of SRS fault zones, or vice versa, might be buffered/limited by  
740 the occurrence of relatively low-permeability fault cores.

### 741 *5.3 Group W outcrops and a comparison with the Utsira High weathering profile*

742 Several authors described in detail the mineralogical and textural changes of the granitoid crystalline  
743 basement of the Utsira High related to the development of a saprolitic weathering profile identified  
744 in the core samples of offshore exploratory wells (Riber et al., 2015, 2016, 2017, 2019a; Lothe et al.,  
745 2018). They describe five different alteration facies accounting for the observed gradual  
746 disaggregation of the igneous host rock, the alteration of feldspar, and clay mineral formation. In their  
747 model, progressive downward penetration of meteoric fluids from the top-basement surface led to  
748 the development of a thick weathering profile containing both saprock and more mature saprolite  
749 formed at the expense of the host granitoid (Fig. 2a–b) (Riber et al., 2015, 2019b; Walter et al., 2018;  
750 Zauyah et al., 2018). Weathering extends from the top-basement surface (that was exposed to a sub-  
751 aerial tropical-humid climate) into depth in the rock column: the progressive degradation of the host  
752 rock primary structure and the presence of a pervasive network of fractures and fault zones promoted  
753 the penetration at depth of fluids enhancing alteration (Braathen et al., 2018; Walter et al., 2018;  
754 Zauyah et al., 2018). The weathering process induces specific textural and mineralogical modification  
755 of the host rock. Incipient weathering along fractures leads to discolouring and limited mineral

756 alteration while still preserving the igneous/metamorphic rock texture (“altered coherent facies A1–  
757 A2” in Riber et al., 2016; Fig. 2b). The alteration of biotite and plagioclase leads to progressive grain  
758 disaggregation and grain fracturing, clay mineral and Al–Fe oxide formation (saprock – saprolite,  
759 “altered incoherent facies A3–A5” in Riber et al., 2016; Fig. 2b) (Goodfellow et al., 2016; Hayes et  
760 al., 2019). Progressive weathering is also reflected in a varying mineralogy of authigenic clay  
761 minerals, with smectite being progressively substituted by kaolinite as the degree of weathering  
762 increases (Coggan et al., 2013; Riber et al., 2016).

763 A similar transition in mineralogical and textural characteristics of the RGD with increasing alteration  
764 can be deduced by the comparison of the preserved weathering remnants at the analysed Group W  
765 outcrops. Each outcrop effectively shows different textural, mineralogical (clay–mineral amounts and  
766 species), and mechanical characteristics, which indicate progressive weathering and alteration.  
767 Outcrops W<sub>1</sub>–S<sub>1</sub> only display partial alteration of the wall rock along fractures while preserving the  
768 textural and mechanical properties of the host rock. The weathering products are represented by  
769 samples BO–OFR–1 (Viola et al., 2016) and TSC–1 (Scheiber and Viola, 2018). The finest–fractions  
770 of these samples are enriched in smectite–like phases, suggesting only incipient alteration (Coggan  
771 et al., 2013; Riber et al., 2016). The characteristics of this outcrop, thus, resemble those of the  
772 “alteration facies A1–A2” described by Riber et al. (2016). The alteration products observed at  
773 outcrop W<sub>2</sub> document, instead, a transition from a coherent host rock (domain *d1* in Fig. 6c) toward  
774 a cohesionless, sandy, grus–type aggregate preserving the magmatic texture but not the mechanical  
775 cohesion of the host rock and increasing clay content (domain *d3* in Fig. 6c). The intermediate  
776 alteration domain *d2* at outcrop W<sub>2</sub> presents textural and mechanical characteristics similar to the  
777 alteration products observed at outcrop S<sub>2</sub>. Sample TSC–36 from outcrop S<sub>2</sub> shows increasing  
778 kaolinite content over smectite (Scheiber and Viola, 2018). The clay content and the textural  
779 characteristics of domain *d2* at outcrop W<sub>2</sub> and outcrop S<sub>2</sub> are similar to those described for “alteration  
780 facies A3–A4” by Riber et al. (2016). The clay–rich alteration product in outcrop W<sub>3</sub> is present in the

781 set of samples analysed by Fredin et al. (2017b) (Bomlo2–3–4). Kaolinite is predominant over  
782 smectite-like phases, suggesting an advanced stage of alteration (Coggan et al., 2013; Riber et al.,  
783 2016). Outcrop W<sub>3</sub> mineralogical and textural characteristics are similar to the “alteration facies A5”  
784 described by Riber et al. (2016).

785 Even though the alteration products analysed in the Group W outcrops may result from diachronic  
786 events, they all present textural and mineralogical characteristics which are ascribable to  
787 kaolinitisation and/or sapolitisation of the RGD host (Coggan et al., 2013; Riber et al., 2016; Fredin  
788 et al., 2017b). Indeed, alteration processes took place at either shallow crustal levels (<6km? depth,  
789 Scheiber and Viola, 2018) or surficial conditions (weathering – sapolitisation) (Fredin et al., 2017b).  
790 It is worth noting that, in each of the analysed outcrops, the weathering products postdate the  
791 formation of brittle structures and are not overprinted by any subsequent brittle deformation. This  
792 may suggest that weathering likely developed during periods of quiescent tectonic activity, as it is  
793 postulated for the development of thick saprolitic horizons (Fredin et al., 2017b)

794 Therefore, we speculate that the studied Group W outcrops may be adopted as representative of the  
795 different stages of progressive weathering. The partially altered outcrop W<sub>1</sub> represents the deepest  
796 portions of the weathering profile; the outcrop W<sub>2</sub> represents the intermediate alteration stage and the  
797 transition toward more evolved alteration products, which are well represented by outcrop W<sub>3</sub> (Fig.  
798 10) (Lothe et al., 2018).

799 Such a defined weathering profile displays a progressive degradation of mechanical strength and an  
800 overall permeability enhancement with the increasing weathering grade (from A1 to A5, Fig. 10b).  
801 Incipient alteration A1–A2 leads, on average, to a progressive decrease in mechanical strength of  
802 about 50% of the intact RGD. The permeability at this stage of weathering is increased up to 0.01–1  
803 D. The transition from coherent (A2) to incoherent alteration facies (A3–A5) is characterised by a  
804 drastic decrease in mechanical strength (UCS down to < 20 MPa, Fig. 10b). Conversely, the



805 permeability increases up to 10 D within alteration facies A3–A4, and then progressively decreases  
806 toward  $10^{-4}$  D when approaching alteration facies A5 (Fig. 10b).

807 The observed variability of the studied properties through the weathering profile is similar to that  
808 commonly reported from saprolitic profiles and alteration profiles related to granite kaolinitisation  
809 (Coggan et al., 2013; Lothe et al., 2018; Walter et al., 2018). In such contexts, the variation of  
810 mechanical strength and permeability are commonly related to the development and evolution of  
811 micro-porosity with increasing weathering (Goodfellow et al., 2016; Lothe et al., 2018; Walter et al.,  
812 2018; Hayes et al., 2019). Microstructural investigations suggest that the petrophysical properties of  
813 incipient alteration stages (coherent alteration facies A1–A2) are controlled by microfracture-related  
814 porosity and biotite alteration (Goodfellow et al., 2016; Walter et al., 2018). The increasing alteration  
815 of biotite and plagioclase into clay-minerals (alteration facies A3–A4) promotes the development of  
816 vacuole-shaped porosity and triggers micro-fracturing as a consequence of positive volume changes  
817 caused by the oxidation reactions in the altering rock (Goodfellow et al., 2016; Walter et al., 2018).  
818 At this stage, highly interconnected pores and micro-fractures enhance the effective permeability and  
819 storage capacity of the weathered rock (Walter et al., 2018). Accordingly, increased micro-porosity  
820 and progressive clay-mineral formation lead to the observed drastic decrease in mechanical strength  
821 between cohesive and cohesionless alteration facies, similarly to what reported by Coggan et al.  
822 (2013) during progressive granite kaolinitisation. The most advanced stages of weathering (alteration  
823 facies A5) are characterised by the complete alteration of biotite and plagioclase into clay-minerals  
824 and Fe-oxides. At this stage, mineral alteration and neoblastesis lead to obstruction of the previously  
825 developed micro-porosity and to the observed reduction of permeability (Walter et al., 2018). As a  
826 result, weathering leads to immediate and drastic decrease of mechanical strength, whereas  
827 permeability firstly increases and then decreases during weathering process progression (Fig. 10).

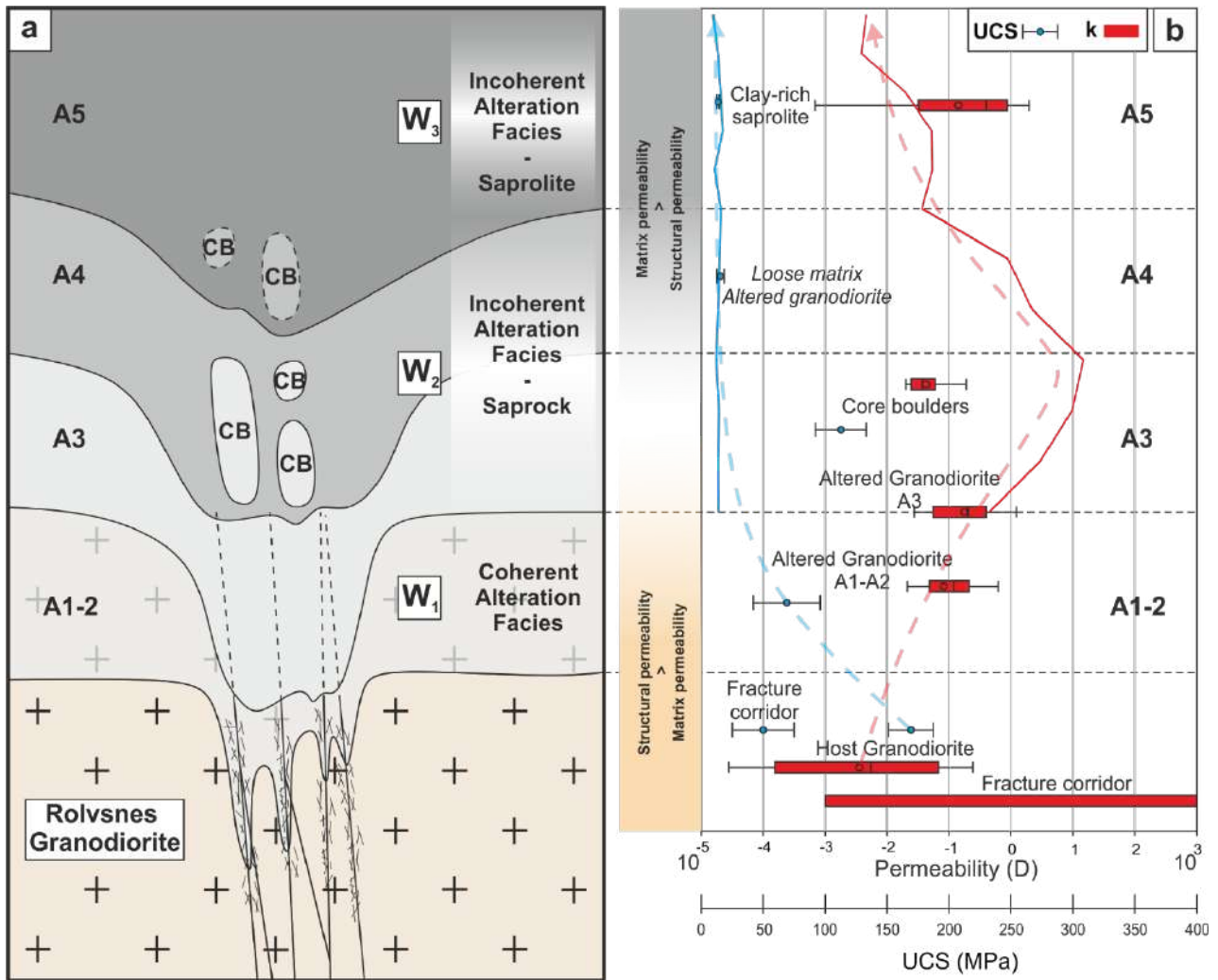
828 The characteristics of the weathering profile studied here have three main important implications for  
829 the characterization of reservoir rocks buried offshore beneath the sedimentary cover in the North  
830 Sea structural highs.

831 (i) Weathering around fracture zones, progressive matrix disaggregation, homogenization,  
832 and the resulting blurring of the structural discontinuities drive the transition from  
833 structurally controlled permeability of the intact host rock toward the matrix-controlled  
834 permeability of intermediate alteration facies (Fig. 10). Fracture zones are characterised  
835 by high permeability, which is, however, spatially confined to the mesoscopic fracture  
836 porosity and fracture aperture. The highly permeable alteration facies A3 is two orders of  
837 magnitude less permeable than the maximum permeability inferred for fracture corridors  
838 (with a fracture aperture of 1 mm). However, it involves larger volumes of the host rock,  
839 and thus dramatically increases the storage capacity of the reservoir rocks.

840 (ii) The low permeability of alteration facies A5 (saprolite) suggests that the top layers of the  
841 weathering profile, when preserved, may act as a partial seal to fluid flow at the base of  
842 the sedimentary cover atop the basement structural highs (Dewandel et al., 2006; Walter  
843 et al., 2018).

844 (iii) The analysed paleo-weathering profile developed during the Mesozoic, and was then  
845 buried below the Jurassic-Cretaceous sedimentary succession of the central North Sea.  
846 The observed differential mechanical strength may have led to different compaction  
847 patterns and behaviours, and thus a differential modification of the permeability of  
848 weathering products during the burial history of the paleo-weathering profile (Lothe et  
849 al., 2018). The weaker clay-rich saprolites may have undergone to a significant change of  
850 the original secondary porosity and permeability when compared to the more stiff,  
851 granular saprock and incipient alteration facies (Lothe et al., 2018; Walter et al., 2018).

852



853

854 **Figure 10:** (a) Schematic representation of the top-basement weathering profile. The penetration of  
 855 weathering processes into the crystalline basement is promoted by the structural discontinuities, such  
 856 as SSRS fractures and fault zones. Alteration intensity decreases with depth. With increasing  
 857 alteration, structural permeability is progressively replaced by matrix permeability related to the  
 858 alteration products. (b) Mechanical strength and permeability variations through the alteration profile.  
 859 The continuous red and light blue curves represent the mechanical strength and permeability observed  
 860 at W<sub>2</sub>. The dashed red and blue curves represent the overall general trend of variation of k and UCS,  
 861 respectively, through the weathering profile. An initial permeability increment in the deepest portion  
 862 of the alteration profile is superseded by a drastic permeability decrement toward the more evolved  
 863 weathering stages.

864

865 *5.4 Reservoir rocks in fractured and weathered basement*

866 The reservoir rocks in fractured and weathered crystalline basement are made up of (Braathen et al.,  
867 2018): (1) the fractures and fault (damage) zones within the non-weathered crystalline basement, in  
868 which permeability is structurally controlled (Figs. 11, 12), and (2) the products of weathering and  
869 alteration, such as the overlying top-basement paleo-weathering profile where the permeability is  
870 dominated by the matrix properties of the weathered products (Figs. 10, 12).

871 The crystalline basement is characterised by the occurrence of a complex network of variably oriented  
872 brittle deformation zones (Gabrielsen and Braathen, 2014; Scheiber and Viola, 2018). Fracture  
873 corridors and damage zones related to both SSRS and SRS fault zones represent high-permeability  
874 conduits for fluid flow (Fig. 11a). The volume of the fractured reservoir suited for fluid storage is  
875 therefore limited to the overall deformation zone volume. However, the occurrence of fault rocks  
876 within SSRS and SRS fault cores may limit the migration of fluids between highly permeable footwall  
877 and hanging wall damage zones. In addition, the intersection of variably oriented fault zones may  
878 limit the overall lateral fluid flow through the large-scale network of SSRS-SRS fracture lineaments.  
879 This would lead to the formation of fault bounded polyhedral domains (i.e. “reservoir compartments”)  
880 of the fractured crystalline basement (Fig. 11b) (Watkins et al., 2018). Within each domain, fluid flow  
881 is limited to the high permeability fracture corridors and footwall/hanging wall damage zone  
882 connected to the domain bounding faults (Fig. 11b).

883 The reservoir storage volume is likely increased within the top-basement weathering profile and  
884 fracture-related alteration zones. Weathering may lead to the formation of very thick (up to 100 meter  
885 thick) profiles characterised by high porosity and permeability (Braathen et al., 2018; Walter et al.,  
886 2018) (Fig. 12). In addition, fluid percolation through fracture zones may lead to the deep penetration  
887 of weathering processes, down to even 300 m below the top-basement paleo surface (Place et al.,  
888 2016; Walter et al., 2018). Furthermore, the most altered A5 facies may represent a top-seal layer to

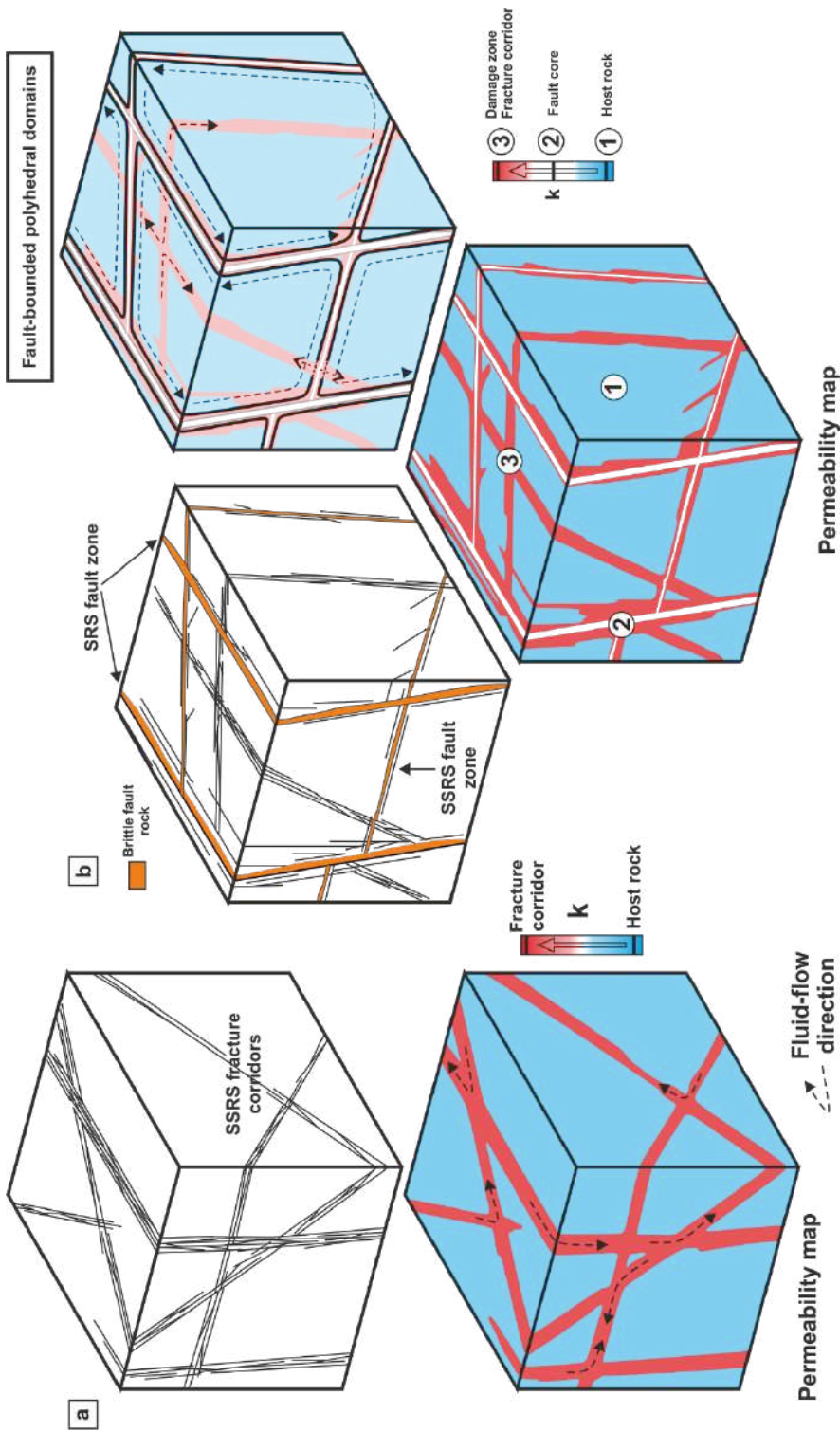
889 the underlying weathered and fractured basement high, buffering the upward migration of fluids  
890 toward the overlying sedimentary succession (Figs. 10, 12) (Walter et al., 2018).

### 891 *5.5 Mechanical strength of fractured and weathered crystalline basement*

892 The fractured and weathered unconventional reservoir under consideration is thus composed of  
893 several structural and textural elements with strikingly different mechanical strength. The fracture  
894 network and the spatial distribution of weathering/alteration products lead to a significant mechanical  
895 heterogeneity/anisotropy within the crystalline basement. In turn, this mechanical heterogeneity and  
896 anisotropy may in places significantly perturb the regional stress field (Gudmundsson et al., 2010).  
897 This could lead to the relevant modification of the overall fracture network geometry, fracture  
898 orientation and spacing within the mechanically heterogeneous rock volume during both natural  
899 deformation and induced fracturing of the reservoir (Jeanne et al., 2013; Smart et al., 2014). The  
900 occurrence of mechanically weak layers, such as the top–basement weathering profile, the gouge–  
901 bearing fault cores, and the partially altered fracture corridors within the crystalline basement may  
902 control the orientation and propagation of natural or induced fractures during either tectonic  
903 deformation or reservoir production (Douma et al., 2019; Forbes Inskip et al., 2020). In particular,  
904 the occurrence of these low–mechanical strength layers and localised heterogeneities may arrest the  
905 propagation of fractures artificially induced during the exploitation phase of unconventional  
906 reservoirs to enhance the local structural permeability of the host low–permeability crystalline rocks  
907 and stimulate production (Forbes Inskip et al., 2020).

908

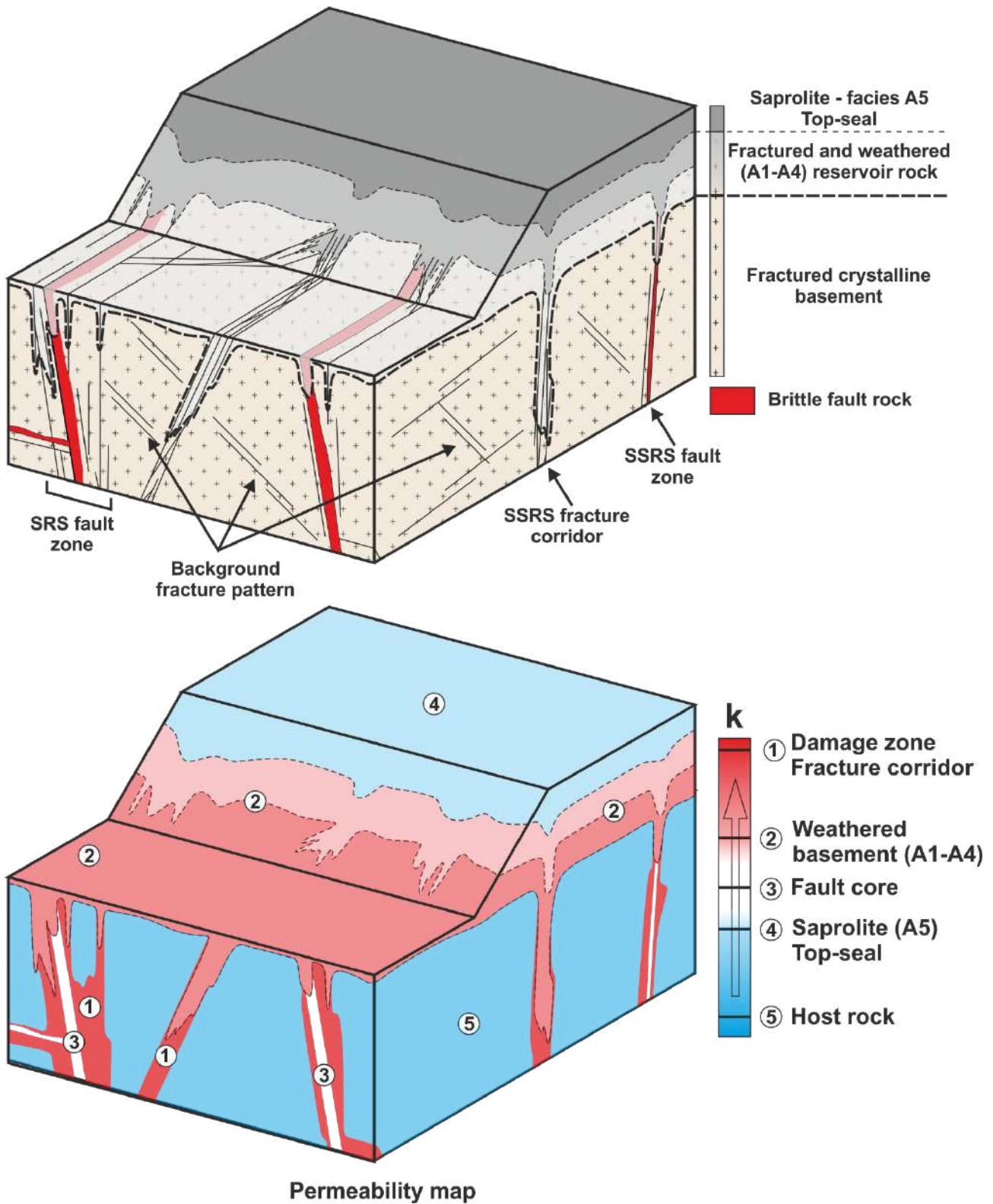
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910

911 **Figure 11:** Schematic representation of the conceptualised fractured basement reservoir. (a) Low-  
 912 strain fracture corridors lead to the development of a high-permeability network of fracture  
 913 lineaments, dominated by the structural permeability. The randomly oriented dashed arrows in the  
 914 coloured permeability map indicate the lack of reservoir compartmentalization. (b) The development

915 of sub-seismic-resolution scale (SSRS) and seismic-resolution scale (SRS) fault zones, and related  
916 fault rocks within fault cores lead to the development of local low-permeability barriers to fluid flow.  
917 The large-scale fracture network is then “compartmentalised” into polyhedral fault-bounded  
918 domains, whose dimension depend on the geometry of the fault zone network. Dashed arrows  
919 describe a closed path for fluid flow within each fault-bounded domain.



920

921 **Figure 12.** Block diagram summarising a conceptualised model of reservoir within fractured and  
 922 weathered crystalline basement (not to scale). (a) The top-basement weathering profile overprints the  
 923 fractured crystalline basement, exploiting the network of fractures and fault zones to penetrate to even  
 924 great depth (down to to 300 m, Walter et al., 2018). (b) Schematic permeability map of the



925 conceptualised model described in (a). The reservoir is deformed by highly permeable fractures and  
926 fault zones (1) and contains the deepest portion of the weathering profile (2). Fault cores (3) are  
927 inherently more permeable than the unaltered host rock, thus acting as a conduit for fluid flow parallel  
928 to fault strike. The potential reservoir rocks (1–2) and potential fluid conduits (1 and 3) are overlain  
929 by the low permeability saprolitic horizon (4), which may act as top–seal layer. Weathering processes  
930 greatly increase the fluid storage and transport potential of the reservoir close to the top–basement  
931 surface, which is otherwise limited to the network of fractures and fault zones.

932

933

934

935 **6. Conclusion**

936 The matrix mechanical properties and permeability constrained by this study within the crystalline  
937 basement rocks exposed on Bømlø result from the prolonged and complex interplay between tectonics  
938 and lithosphere–atmosphere interactions (Fredin et al., 2017b). Brittle fracturing and weathering led  
939 to the development of secondary porosity at different scales: (a) meso–scale fracture porosity; (b)  
940 micro–porosity related to cataclastic processes and micro–fracturing; (c) micro–porosity related to  
941 mineral alteration and reaction–induced micro–fracturing. Thus, fault rock formation and weathering  
942 processes related to subaerial exposure and fluid percolation through deformation zones deeply  
943 affected both the mechanical strength and the matrix permeability of the RGD.

944 The fractured and weathered crystalline basement reservoir rocks are, therefore, composed of two  
945 main constituent elements: (1) the fractured crystalline basement and (2) the weathered/altered rock  
946 volume. The multiscale fracture network is composed of different types of fracture zones (SSRS  
947 fracture corridor, SSRS and SRS fault zones). Fault zones, either SSRS or SRS, represent zones of  
948 enhanced permeability with respect to the host crystalline basement and conduits for fluid flow  
949 parallel to the fault planes at all stages of evolution and at all scales (Caine et al. 1996). Meso–scale  
950 fractures spatially related to low strain and limited throw fracture corridors drastically enhance the  
951 overall permeability of the basement. Similarly, the formation of fault rocks, which are on average  
952 two orders of magnitude more permeable than the host rock, significantly increases the crystalline  
953 basement matrix permeability. They are characterised by anisotropic mechanical and permeability  
954 properties. In addition, they may act as partial seal/buffer layers to fluid flow within the high  
955 permeability structural network defined by fracture corridors and damage zones (Bruhn et al., 1994;  
956 Caine et al., 1996).

957 The comparison of weathering characteristics between different outcrops has allowed us to quantify  
958 the mechanical strength and permeability as a function of the actual position within the reconstructed

959 weathering profile (Dewandel et al., 2006; Lothe et al., 2018). The weathering profile shows a  
960 progressive degradation of mechanical strength coupled to an initial increase in permeability during  
961 the incipient stages of weathering (up to 10 D, corresponding to its deepest portion), followed toward  
962 shallower depth by a significant decrease in permeability toward the more evolved weathering stages  
963 (down to  $10^{-3}$  D). The matrix secondary permeability created during weathering greatly increases  
964 fluid storage and migration in a potential reservoir, which would otherwise be limited to volumes  
965 with sufficient structural permeability. Mature saprolitic horizons, on the other hand, may act as top–  
966 seal layer to the underlying permeable weathered products. As a consequence, weathering profiles  
967 may act both as reservoir and top–seal unit in sub–unconformity unconventional reservoirs (Braathen  
968 et al., 2018; Holdsworth et al., 2020; McCaffrey et al., 2020).

969 Fractured and weathered crystalline basement reservoirs are characterized by a significant mechanical  
970 heterogeneity, as inferred by the strikingly different mechanical properties of deformation zones,  
971 brittle fault rocks and weathering products. The mechanical heterogeneity is particularly developed  
972 toward the top–basement unconformity, where brittle deformation zones may be decorated by  
973 alteration products and abut the overlying weathering profile. The detailed quantification of the  
974 mechanical heterogeneity and its spatial distribution is therefore fundamental to optimize the planning  
975 strategies for the exploration and the operative exploitation of sub–unconformity unconventional  
976 reservoirs.

977

978

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988



990

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