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**Camille Dusséaux:** Conceptualization, Investigation, Sampling, Data Curation, Writing - Original Draft, Visualization

**Aude Gébelin:** Conceptualization, Sampling, Investigation, Validation, Writing - Original Draft, Supervision, Project administration

**Philippe Boulvais:** Conceptualization, Sampling, Writing - Review & Editing, Supervision

**Gilles Ruffet:**  $^{40}\text{Ar}/^{39}\text{Ar}$  geochronology - investigation, formal analysis, visualisation, writing

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# Timing and duration of meteoric water infiltration in the Quiberon detachment zone (Armorican Massif, Variscan belt, France)

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## Key Points

- $\delta D$  values of synkinematic hydrous minerals document the presence of 300 Ma-old meteoric fluids in a key Variscan detachment zone.
- Meteoric fluid-rock interactions occurred during high-temperature deformation ( $>500^{\circ}\text{C}$ ) in the Quiberon detachment footwall.
- Coeval detachment activity, leucogranite emplacement and migmatization allowed the hydrothermal system to be maintained for  $\sim 17$  Myr.

## Keywords

Variscan; detachment; shear zone; hydrous silicates; hydrogen isotope;  $^{40}\text{Ar}/^{39}\text{Ar}$ ; U(-Th)/Pb; geochronology; fluid-rock interaction; meteoric fluids

## Abstract

Assessing the geochemical signature and the role of fluids in a key Variscan detachment zone demonstrates the link between crustal deformation, thermo-mechanical events and Variscan mineralization. We document meteoric fluid infiltration into the ductile segment of the Late-Carboniferous Quiberon detachment zone (QDZ), when synkinematic muscovite and tourmaline crystallized and equilibrated with deuterium-depleted surface-derived fluids during high-temperature deformation. Titanium-in-muscovite thermometry supported by microstructures indicate that syntectonic isotope exchange between fluids and hydrous minerals occurred above  $500^{\circ}\text{C}$ .  $^{40}\text{Ar}/^{39}\text{Ar}$  muscovite data ( $\sim 319$  to  $\sim 303$  Ma) and U(-Th)/Pb geochronology on zircon, monazite and apatite ( $\sim 318$  to  $\sim 305$  Ma) from syntectonic leucogranites together with microstructural and geochemical (U and REE contents) data suggest that meteoric fluid-rock-deformation interaction started at  $\sim 320$  Ma and played a major

38 role in leaching uranium at ~305 Ma. U-Th/Pb data (~330 to ~290 Ma) from migmatites located  
39 below the QDZ strengthen the idea that meteoric fluid infiltration, detachment activity,  
40 syntectonic leucogranite emplacement and migmatization were coeval and allowed the  
41 development of a sustained hydrothermal system.

## 42 Plain Language Summary

43 We document the presence of 300-million-year-old rainwater in a normal fault that was in the  
44 internal zones of an eroded mountain range called the Variscan Belt, now outcropping at the  
45 surface in Brittany (France). Water-bearing minerals preserved the hydrogen isotope  
46 composition of the water that penetrated the fault zone at depth. The same minerals can be  
47 dated based on the radioactive decay of Argon. Combined with other dating methods applied  
48 to various minerals, we conclude that the rainwater infiltrated the fault zone 320 to 300 million  
49 years ago. This study allows a better insight of how fluids circulate in the Earth's crust, which  
50 is critical for our understanding of mineralization processes that form ore deposits.

## 51 1 Introduction

52 Extensional shear zones play a major role during lithospheric extension as they control mass  
53 transport through lateral and vertical displacements in mid to lower levels of the continental  
54 crust (e.g. Teyssier & Whitney, 2002; Tirel et al., 2008; Whitney et al., 2013). The zones of  
55 planar weakness that separate the relatively cool upper crust from the hotter middle-lower crust  
56 may act as fluid conduits and, therefore, represent sites of strong fluid-rock interaction where  
57 meteoric, metamorphic and magmatic fluids interact (e.g. Dusséaux et al., 2019; Famin et al.,  
58 2004; Gébelin et al., 2011, 2015, 2017; Gottardi et al., 2011, 2018; Methner et al., 2015;  
59 Morrison & Anderson, 1998; Nesbitt & Muehlenbachs, 1995). Assessing the source of fluids  
60 in such hydrothermal systems has a range of applications including constraining the conditions  
61 for mineralization and ore deposition (Ballouard et al., 2017, 2018a; Beaudoin et al., 1991;  
62 Boiron et al., 2003) or paleoaltimetry reconstructions if surface-derived (meteoric) fluids are  
63 present (e.g. Dusséaux et al., 2021; Gébelin et al., 2012, 2013; Grambling et al., 2022; Mulch  
64 et al., 2004).

65 Based on the low hydrogen ( $\delta D$ ) and oxygen ( $\delta^{18}O$ ) isotope values of synkinematic hydrous  
66 silicates (e.g.  $\delta D_{\text{Muscovite}}$  values  $< -90\text{‰}$ ), meteoric water infiltration has been documented in  
67 the footwall of detachment zones, e.g. in the North American Cordillera (e.g. Gébelin et al.,  
68 2011, 2015; Mulch et al., 2004; Person et al., 2007), in the Himalayas (e.g. Gébelin et al., 2017,  
69 2013), and in the Variscan belt of Western Europe (e.g. Ballouard et al., 2018a, 2017, 2015;  
70 Dusséaux et al., 2021, 2019). Some of these studies (e.g. Dusséaux et al., 2019, 2021; Gébelin  
71 et al., 2012, 2013) highlight three main conditions essential for the downward infiltration of  
72 meteoric fluids: (1) the development of brittle normal faults and associated fracture networks  
73 in the upper crust that facilitate permeability of the crust in the brittle domain, (2) a high  
74 geothermal gradient with advection of partially molten material in the footwall that provides  
75 heat to support an active convection system, and (3) the presence of a hydraulic head provided  
76 by topography.

77 In this study, we focus on the Quiberon detachment shear zone (QDZ; Figure 1; Southern  
78 domain, Armorican Massif, France), an extensional shear zone with excellent exposures, that  
79 was active during the Late Carboniferous and developed as a result of late-orogenic extension  
80 (e.g. Gapais et al., 2015, 1993). The QDZ represents an excellent target to look at fluid-rock-  
81 deformation interactions and understand how, why and when fluids circulated in the crust  
82 during the Late Carboniferous because: (1) hydrogen isotope ratios ( $\delta D$ ) of hydrous minerals  
83 indicate that meteoric fluids infiltrated the mylonitic footwall of the QDZ (Dusséaux et al.,

84 2019), (2) the structural, metamorphic, and geochronological record of the region is relatively  
85 well established (e.g. Gapais et al., 2015, 1993; Turrillot, 2010; Turrillot et al., 2011); (3) brittle  
86 normal faults have been identified in the hanging wall of the QDZ (Figure 2A; Ballouard et al.,  
87 2015); (4) the QDZ is considered to have played a major role in the exhumation of high-grade  
88 metamorphic rocks (Brown & Dallmeyer, 1996; Gapais et al., 2015), and (5) the region was  
89 characterized by a high geothermal gradient in part generated by the emplacement of  
90 peraluminous syntectonic leucogranites and migmatites (e.g. Ballouard et al., 2015; Brown and  
91 Dallmeyer, 1996; Gapais et al., 2015).

92 In addition, different tools (hydrogen and oxygen isotopes of whole rock, minerals and fluid  
93 inclusions) applied on diverse geological materials (e.g undeformed and mylonitic granites,  
94 host rocks and quartz veins) across the strike-slip South Armorican Shear Zone (Ballouard et  
95 al., 2018a; Dusséaux et al., 2019; Lemarchand et al., 2012; Tartèse et al., 2013), within the  
96 footwall of nearby detachment zones (Ballouard et al., 2017, 2015; Dusséaux, 2019; Dusséaux  
97 et al., 2019) and in kilometer-scale quartz veins cutting through the lower crustal units in the  
98 southern Armorican domain (Lemarchand et al., 2012), as well as on shear zones of the western  
99 French Massif Central (Dusséaux et al., 2021) that constitutes the SE prolongation of the  
100 southern Armorican domain, all agree with the idea that cold meteoric fluids infiltrated the  
101 upper plate during Late-Carboniferous extensional processes and reached significant depths  
102 (10-15 km) while mixing with deep crustal fluids. These Earth's surface-derived fluids played  
103 a major role in the coupled grain-scale physical and geochemical processes resulting from  
104 fluid-rock-deformation interactions that influenced the thermomechanical evolution of  
105 Variscan deformation zones and led to the formation of uranium ore deposits (e.g. Ballouard  
106 et al., 2017, 2018a; Boulvais et al., 2019).

107 Here, we present structural, microstructural, electron backscatter diffraction (EBSD)  
108 petrofabrics, Ti-in-muscovite thermometry, hydrogen isotope ( $\delta D$ ), geochemical and  
109 geochronological data ( $^{40}Ar/^{39}Ar$  and U-Th/Pb) from mylonitic leucogranites exposed in the  
110 QDZ footwall and associated high-grade metamorphic rocks. This new dataset supports  
111 previous premises that surface-derived fluids penetrated the ductile segment of the QDZ at  
112 ~320 Ma while at the same time high-grade metamorphic rocks were emplacing in the lower  
113 crust. In addition, this study indicates that meteoric fluid-rock-deformation interaction lasted  
114 for about 17 million years.

## 115 **2 Tectonic context and timing of metamorphism in the southern Armorican** 116 **domain**

117 The Armorican Massif (Western France) is part of the Ibero-Armorican arc, which forms the  
118 internal zone of the Variscan belt of Western Europe (Figure 1A). Three main domains,  
119 characterized by contrasting tectonic, geochronological and metamorphic features, are  
120 delimited by two major dextral strike-slip shear zones, the North Armorican Shear Zone  
121 (NASZ) to the north and the South Armorican Shear Zone (SASZ) to the south. In contrast to  
122 the northern and central domains, the southern domain underwent substantial Carboniferous  
123 crustal thickening and high-pressure metamorphism (e.g. Ballèvre et al., 2013). From top to  
124 bottom three main units can be identified in the southern domain (Figures 1A and 1B): (1) The  
125 upper unit comprising blueschists (1.4-1.8 GPa, 550°C) and metavolcanics (0.8 GPa, 350-  
126 400°C; Bosse et al., 2002; Le Hebel et al., 2002, 2007); (2) The intermediate unit characterized  
127 by micaschists that preserved greenschist and amphibolite-facies metamorphism (Bossière,  
128 1988; Goujou, 1992; Triboulet & Audren, 1988); and (3) The lower unit represented by (S-  
129 type) syntectonic leucogranites and high-grade metamorphic rocks (0.8 GPa, 700-750°C; Jones  
130 and Brown, 1990). Peak high-pressure metamorphism in the upper unit (1.8-2.0 GPa and 450-

131 500°C) and amphibolite-facies in the intermediate unit (0.9 GPa for 650-700°C) have been  
 132 dated between ~370 Ma and ~345 Ma, respectively (U/Pb,  $^{40}\text{Ar}/^{39}\text{Ar}$  and Rb/Sr; e.g. (Bosse et  
 133 al., 2005, 2002; El Korh et al., 2011; Le Hébel et al., 2002; Paquette et al., 2017). Strike-slip  
 134 ductile shear zones at ~350-345 Ma are responsible for the exhumation of these two units (e.g.  
 135 Ballèvre et al., 2013; Ballouard et al., 2018b; Bosse et al., 2002, 2005; El Korh et al., 2011;  
 136 Jegouzo, 1980; Tartèse et al., 2011a, 2011b, 2011c; Tartèse and Boulvais, 2010).

137 In contrast, Late-Carboniferous detachment zones played a major role in the exhumation of  
 138 migmatites and associated syntectonic leucogranites that form the lower unit, and mark the  
 139 interface that separates the ductile lower crust from blueschist and metavolcanics  
 140 systematically located stratigraphically above (e.g. Ballouard et al., 2017, 2015; Brown and  
 141 Dallmeyer, 1996; Cagnard et al., 2004; Gapais et al., 2015, 1993; Goujou, 1992; Turrillot  
 142 et al., 2011). In agreement with a high Variscan geothermal gradient (~35°C/km, e.g.  
 143 Vanderhaeghe et al., 2020), extreme thermal gradients across detachment zones (e.g. Gottardi  
 144 et al., 2013), and gravity modelling (e.g. Gébelin et al., 2006), syntectonic leucogranites,  
 145 including the Quiberon granite (Figure 1), emplaced at shallow depths (~3 to 10 km) in the  
 146 footwall of detachment zones by relatively cold (<850°C) partial melting of metasediments  
 147 (e.g. Ballouard et al., 2017; Capdevila, 2012; Gapais et al., 2015; Le Hébel et al., 2007). This  
 148 is consistent with the emplacement conditions estimated for garnet-cordierite-rich syntectonic  
 149 leucogranites (~0.5-0.6 GPa, 750-800°C; Gébelin et al., 2009) that represent the granitic melt  
 150 fraction of biotite-garnet-sillimanite-cordierite metapelites in the western part of the French  
 151 Massif Central (SE extension of the southern Armorican domain).

152 At the scale of the southern domain, detachment shear zones predominantly trend N-S to NW-  
 153 SE and are associated with top-to-the-west sense of shear (e.g. Gapais et al., 2015) observed  
 154 parallel to the stretching lineations that have orientations that vary locally (see Quiberon: top  
 155 to the WNW, Sarzeau: top to the ESE and Piriac: top to the NNE on Figure 1A). These near-  
 156 horizontal shear zones developed due to WNW-ESE crustal extension and are connected by  
 157 dextral transfer zones parallel to the SASZ (e.g. Turrillot et al., 2011). Both strike-slip and  
 158 detachment shear zones were active during the Late-Carboniferous and served as conduits for  
 159 aqueous fluids and/or melt migration (e.g. Gapais et al., 2015).

## 160 **3 Results**

### 161 **3.1 Structural study**

#### 162 **3.1.1 Outcrop description**

163 The Quiberon peninsula provides exceptional exposures of the Quiberon Variscan detachment  
 164 footwall represented by a ~2 km-thick high-strain zone (Figure 1; Gapais et al., 1993). The  
 165 contact between the footwall and hanging wall of the QDZ is currently located below sea level  
 166 at  $\geq 500\text{m}$  from the coast (Figures 1 and 2C). Therefore, the location of the samples in the  
 167 footwall was estimated based on their structural position from the hanging wall/footwall  
 168 contact as it is marked on the geological map of the continental margin (Thinon et al., 2008),  
 169 and taking into consideration a dip of ~38° to the WNW for the foliation (see description below  
 170 and Figures 1 and 2, Table 1). Our study is based on two representative outcrops of coarse-  
 171 grained mylonitic leucogranite in the Quiberon detachment footwall: 1) QUIB01, located at  
 172 ~300m beneath the QDZ, and 2) QUIB03 at ~322m beneath the QDZ. Both outcrops are  
 173 characterized by a shallow to moderate (10-20°) WSW-dipping foliation (S plane) with a ESE-  
 174 WNW trending stretching lineation highlighted by quartz and muscovite grains (~N280)  
 175 (Figures 1 and 2). S-planes, emphasized by sheared feldspar and muscovite grains and quartz  
 176 ribbons, are affected by C-planes dipping with an angle of ~30° to the WNW (Figures 1D and

177 2D). The 2 – 5 cm spaced C-planes are emphasized by large deformed muscovite (> 0.5 cm)  
 178 oriented parallel to the shear zone boundary and together with S-planes form shear bands  
 179 indicating a top-to-the-WNW sense of shear (Figures 2C and 2D). Shear bands form a  
 180 heterogeneous and anastomosing network that isolates sigmoidal quartz veins and micaschist  
 181 lenses (Figure 2B).

182 Pegmatites intrude the mylonitic leucogranites in different places and especially between  
 183 QUIB01 and QUIB03 that form their upper and lower boundaries (Figure 2F). Here, pegmatites  
 184 are intensively deformed and form corridors from a few centimeters up to 1- to 2-meter thick  
 185 which are near-parallel to the mylonitic leucogranite foliation (Figure 2F). As observed in the  
 186 leucogranites, pegmatite and quartz veins display a ESE-WNW stretching lineation marked by  
 187 tourmaline and quartz grains (Figures 2E and 3A). Based on their spatial geometric  
 188 relationships with the mylonitic leucogranites and their fabrics indicative of recrystallisation  
 189 during deformation, we interpret their syntectonic emplacement during the Quiberon  
 190 detachment activity (samples QUIB03-06; Figure 3A). In the same area, undeformed  
 191 pegmatites and quartz veins crosscut the mylonitic leucogranite foliation with an angle of 30-  
 192 50° (QUIB09; figure 2E). Structurally lower in the section, C-S structures progressively  
 193 disappear due to an increased heat-flow leading to a high granitic melt fraction with biotite  
 194 schlieren highlighting the primary near-horizontal foliation (Figure 3B).

195 In addition to those observed in the lower part of the Quiberon section, migmatitic rocks  
 196 outcrop further east at Port Navalo (Gapais et al., 1993, 2015). Here, two main rock types  
 197 characterize these high-grade rocks: biotite-garnet-sillimanite-cordierite metasedimentary  
 198 gneiss and garnet-cordierite leucogranite. The metasediments, representing the migmatite  
 199 paleosome/mesosome, display a NNW-SSE striking foliation that dips steeply to the ENE  
 200 (>65°) and contains a near horizontal stretching lineation defined by cordierite and garnet  
 201 (Figure 3C). Pressure shadows formed by cordierite-sillimanite-biotite assemblages surround  
 202 large garnet grains (> 2 cm) (Figure 3C). In some places, the garnet-cordierite leucogranite  
 203 forms dykes near-parallel to the foliation, illustrating the increasing granitic melt fraction from  
 204 the paleosome/melanosome (Figure 3D). These alternating quartzo-feldspathic-rich  
 205 (leucosome) and ferromagnesian-rich (paleosome/melanosome) layers are folded and based on  
 206 the fold axial surface indicate a syn-to-post migmatitic NNE-SSW shortening event (Figure  
 207 3D).

### 208 3.1.2 Microstructures

209 At the microscopic scale, muscovite mica fish from mylonitic leucogranites (i.e. QUIB01 and  
 210 QUIB03) develop along both shear (C) and schistosity (S) planes (Figures 4A and 4D).  
 211 Phyllosilicate mineral cleavages are parallel to S planes and are relatively displaced by C planes  
 212 made of finely recrystallized material (e.g. Figures 4B, 4E and 4H; Berthé et al., 1979; Lister  
 213 and Snoke, 1984). Together, they form C-S structures that support the idea that leucogranites  
 214 were emplaced while the Quiberon detachment was active (Figures 4A and 4D). Overall, C-S  
 215 structures indicate a top-to-the-WNW sense of shear (Figure 4A), but opposite shear sense can  
 216 be observed in some leucogranite samples (Figure 4D). Following the mica fish morphological  
 217 grouping proposed by ten Grotenhuis et al. (2003), the truncated upper and lower parts of  
 218 lenticular muscovite grains (group 1) indicate that they underwent rotation and solution-  
 219 precipitation during deformation (Figure 4F, 4G and 4B). Some of these group 1 muscovite  
 220 fish evolve into group 2 by drag along micro shear planes that developed along the upper and  
 221 lower sides of the grain, resulting in bent tips and cleavage planes (Figures 4G). While  
 222 muscovite grains in sample QUIB01 display clear boundaries (Figures 4F and 4G), mica fish  
 223 in sample QUIB03 shows recrystallization of tiny muscovite grains (<50 µm) along the rims  
 224 and in pressure shadows (Figure 4B). In mylonitic leucogranite, tourmaline grains are oriented



225 with their long axis parallel to the foliation and perpendicular to the lineation, indicating that  
226 they rotated along the shear plane (Figure 4H). Sub-solidus deformation microstructures such  
227 as rectangular and castellated quartz grain boundaries (Figures 4C and 4I) indicate that grain  
228 boundary migration (regime 3; Hirth and Tullis, 1992) was the dominant dynamic  
229 recrystallization process that affected the mylonitic syntectonic leucogranite, reflecting a high  
230 recovery: strain accumulation ratio resulting from high-temperature or low strain rate  
231 deformation (e.g. Kilian & Heilbronner, 2017). Feldspar microstructures such as perthite and  
232 abundant micro-fracturing indicate both ductile and brittle deformation close to the brittle-  
233 ductile transition for feldspar ( $\sim 450 \pm 50^\circ\text{C}$ ) and are consistent with a top to the WNW sense  
234 of shear (Figures 4A, 4D and 4E). In addition, microfractures, brecciation and sericitization of  
235 feldspar and grain boundaries, as well as minor chloritic alteration along biotite grain  
236 boundaries and shear planes overprinted the previous deformation fabrics in both mylonitic  
237 leucogranite samples and provide evidence for hydrothermal alteration (Figures 4A and 4D).

238 The ultramylonitic pegmatite displays a foliation (S) supported by fine grains of feldspar and  
239 quartz ( $< 100 \mu\text{m}$ ) that form ribbons wrapping some remains of K-feldspar porphyroclasts  
240 (QUIB04; Figures 5A, 5B and 5C). C-planes marked by mica fish layers make an angle of  $\sim 30^\circ$   
241 with S-planes. In agreement with microstructures from mylonitic leucogranites, pegmatites  
242 display C-S structures depicting a top-to-the-WNW sense of shear that confirm their  
243 syntectonic emplacement (e.g. Gapais & Boundi, 2014). This kinematic inference is supported  
244 by drag or quarter folds made by quartz and feldspar-rich layers (Figure 5B) and discontinuous  
245 C' shear planes (Figure 5C). This top-to-the WNW kinematic is also strengthened by asymmetric  
246 sigmoidal feldspars (Figures 5A and 5D). Biotite in ultramylonitic pegmatite is more frequent  
247 than in leucogranite and marks the 0.25 cm-spaced shear planes (Figures 5A, 5B and 5D). The  
248 mylonitic foliation is overprinted by lower-grade microstructures. Narrow zones of cataclasite  
249 and ultracataclasite developed parallel to the shear plane in mica-poor areas (Figure 5A).  
250 Feldspars are commonly affected by fractures that are sometimes filled by muscovite or quartz  
251 (Figure 5B, 5D and 5E). Recrystallized quartz surrounds feldspar grains in pressure shadows  
252 (Figures 5D and 5E). Also, sericitization of feldspar and muscovite is frequent including in  
253 pressure shadows and along shear planes (Figures 5C and 5D).

254 The observation of high-grade metamorphic rocks at the microscopic scale confirms the  
255 presence of biotite, garnet, sillimanite, and cordierite in addition to K-feldspar, plagioclase and  
256 quartz in the melanosome/mesosome part (Figure 5F). At the macroscopic scale (Figure 3C),  
257 sillimanite and cordierite are observed in pressure shadows around garnet. The leucosome part  
258 exhibits a medium-grained fabric ( $>1 \text{ mm}$ ) made of K-feldspar, plagioclase, quartz, biotite,  
259 garnet, and large cordierite minerals that form  $\sim 50\%$  of the matrix/sample (Figure 5G).

### 260 3.1.3 EBSD on quartz grains

261 The crystallographic-preferred orientation (CPO) of quartz grains from sample QUIB01 is used  
262 here to obtain information about the quartz slip system operated in mylonitic granite from the  
263 QDZ footwall (Figure 6; Text 2 of the supplementary material). As observed in thin section,  
264 the quartz grain maps obtained by EBSD reveal sigmoidal quartz ribbons made of quartz grains  
265 (re)crystallizing by grain boundary migration (Figures 6A and 6B). Quartz c-axis pole figures  
266 exhibit four maxima symmetrically distributed at  $\sim 25^\circ$  around the shear plane (C) and the  
267 lineation direction (Figure 6C), indicating crystal plastic deformation dominated by prismatic  
268  $\langle c \rangle$  glide during coaxial strain with either constriction or plane strain (see framed model on  
269 Figure 6 from Barth et al., 2010). The c-axes data also show two maxima located at  $\sim 70^\circ$  with  
270 a monoclinic symmetry to the direction of the lineation which reflect activation of basal [a] slip  
271 (e.g. Stipp et al., 2002). The activation of the prismatic  $\langle c \rangle$  slip system is also supported by  
272 quartz a-axes that form a single asymmetric girdle indicating a non-coaxial kinematic to the

273 WNW (see framed model on Figure 6 from Barth et al., 2010). Based on the a-axes defining a  
 274 girdle rather than point maxima we propose that deformation occurred more likely under  
 275 constriction than under plane strain (e.g. Barth et al., 2010). Both quartz c- and a-axes point to  
 276 the activation of prism  $\langle c \rangle$  glide (e.g. Blumenfeld et al., 1986).

### 277 3.2 Hydrogen isotope geochemistry

278 The hydrogen isotope ratios ( $\delta D$ ) of muscovite and tourmaline from fractions  $> 250 \mu m$  were  
 279 measured in 11 samples of sheared leucogranite, pegmatite, micaschist and quartz veins  
 280 collected across 300m of structural section into the underlying mylonitic footwall of the QDZ  
 281 (Figure 7 and Table 1). Analyses were performed at the Goethe University-Senckenberg BiK-  
 282 F Stable Isotope Facility, Frankfurt (analytical procedure detailed in Text 1 of the  
 283 supplementary material). Samples presented here include  $\delta D$  values for muscovite previously  
 284 reported in Dusséaux et al. (2019).

285 Muscovite from leucogranite samples displays  $\delta D_{\text{Muscovite}}$  values of -85 to -79‰ at distances of  
 286 250-322m beneath the QDZ. These  $\delta D_{\text{Muscovite}}$  values are lower than those from the  
 287 ultramylonitic pegmatites that show  $\delta D_{\text{Muscovite}}$  values ranging between -75 and -64‰.  
 288 Muscovite from quartz veins and micaschist yields intermediate  $\delta D_{\text{Muscovite}}$  values that range  
 289 from -80 to -76‰. Tourmaline in leucogranite samples yields lower  $\delta D_{\text{Tourmaline}}$  values (-87 to  
 290 -86‰) than those from pegmatites (-80‰). Tourmaline from quartz veins has intermediate  
 291  $\delta D_{\text{Tourmaline}}$  values that vary between -85 and -81‰.

### 292 3.3 Muscovite geochemistry and Ti-in-Ms geothermometry

293 We measured the chemical composition of muscovite from QUIB01 and QUIB03 to determine  
 294 (1) the hydrothermal or magmatic origin of muscovite and, (2) the temperature of muscovite  
 295 crystallization using the Titanium-in-muscovite thermometer (Wu & Chen, 2015; Text 3 and  
 296 Table 1 of the supplementary material).

297 We first focused on muscovite from QUIB01 from which the lowest  $\delta D_{\text{Muscovite}}$  values have  
 298 been obtained (-85‰) that were previously interpreted to reflect interaction with surface-  
 299 derived fluids (Dusséaux et al., 2019). The titanium, magnesium and sodium contents attain  
 300 values of  $0.02 < \text{Ti} < 0.05$  apfu,  $0.08 < \text{Mg} < 0.10$  apfu and  $0.02 < \text{Na} < 0.03$  apfu, respectively,  
 301 indicating a magmatic to hydrothermal origin as also highlighted in a Ti-Na-Mg ternary  
 302 diagram (Figure 8) (Miller et al., 1981). We then focused on muscovite grains from QUIB03  
 303 that yielded slightly higher  $\delta D_{\text{Muscovite}}$  values (-82‰). Their titanium ( $0.02 < \text{Ti} < 0.03$  apfu)  
 304 and sodium ( $0.02 < \text{Na} < 0.03$  apfu) contents are similar to those obtained for the QUIB01  
 305 muscovite, but a higher Mg content ( $0.09 < \text{Mg} < 0.11$  apfu) makes them plot towards the  
 306 hydrothermal muscovite field (Figure 8).

307 We then applied the Titanium-in-muscovite thermometer (Ti-in-Ms; Wu and Chen, 2015) to  
 308 determine the metamorphic temperature conditions using a pressure estimate of  $0.4 \pm 0.1$  GPa  
 309 (Text 4 and Table 1 of the supplementary material; e.g. Gapais et al., 1993; Turrillot, 2010).  
 310 Results from the two samples are consistent with a temperature of  $569 \pm 42^\circ\text{C}$  for QUIB01 and  
 311  $546 \pm 41^\circ\text{C}$  for QUIB03.

### 312 3.4 $\delta D_{\text{Water}}$ values of syntectonic fluids

313 To calculate the hydrogen isotopic composition of the fluid present during the synkinematic  
 314 crystallization of muscovite ( $\delta D_{\text{Water}}$ ), we used the temperatures of  $569 \pm 42^\circ\text{C}$  and  $546 \pm 41^\circ\text{C}$   
 315 (average =  $558 \pm 42^\circ\text{C}$ ) determined using the Titanium-in-muscovite thermometer applied for  
 316 QUIB01 and QUIB03, respectively, which are consistent with temperature estimates deduced  
 317 from quartz microstructures and quartz c axes fabrics ( $\sim 550$ - $600^\circ\text{C}$ ; see above).

318 Using a temperature of  $569 \pm 42^\circ\text{C}$ , a  $\delta\text{D}_{\text{Muscovite}}$  value of  $-85\text{‰}$  and the muscovite-water  
 319 hydrogen isotope fractionation factor of Suzuoki and Epstein (1976), we calculated a  $\delta\text{D}_{\text{Water}}$   
 320 value of  $-73 \pm 4\text{‰}$  for QUIB01. This value is consistent with the  $\delta\text{D}_{\text{Water}}$  value of  $-68 \pm 4\text{‰}$   
 321 calculated for QUIB03, using a  $\delta\text{D}_{\text{Muscovite}}$  value of  $-82\text{‰}$ , a temperature of isotopic exchange  
 322 of  $546 \pm 41^\circ\text{C}$  and the hydrogen isotope muscovite-water fractionation of Suzuoki and Epstein  
 323 (1976).

324 Using the average temperature of  $558 \pm 42^\circ\text{C}$  obtained from QUIB01 and QUIB03, the  
 325 equation of Suzuoki and Epstein (1976) and the  $\delta\text{D}_{\text{Muscovite}}$  values of each individual sample  
 326 from the Quiberon section, we obtained  $\delta\text{D}_{\text{Water}}$  values ranging from  $-73$  to  $-66 \pm 4\text{‰}$  in  
 327 mylonitic leucogranite, from  $-67$  to  $-63 \pm 4\text{‰}$  in quartz veins and micaschist, and from  $-62$  to  
 328  $-51 \pm 4\text{‰}$  in ultramylonitic pegmatite (Table 1). Using the hydrogen isotope composition of  
 329 tourmaline in equilibrium with muscovite, the hydrogen water-tourmaline isotope fractionation  
 330 factor of Kotzer et al. (1993) and the average temperature of  $558 \pm 42^\circ\text{C}$ , we calculated  $\delta\text{D}_{\text{Water}}$   
 331 values ranging from  $-77$  to  $-74 \pm 5\text{‰}$  in mylonitic leucogranite, from  $-74$  to  $-70 \pm 4\text{‰}$  in quartz  
 332 veins, and  $-69 \pm 4\text{‰}$  in ultramylonitic pegmatite (Table 1). The  $\delta\text{D}_{\text{Water}}$  values calculated from  
 333 the  $\delta\text{D}_{\text{Tourmaline}}$  values ( $-77 \leq \delta\text{D}_{\text{Water}} \leq -69\text{‰}$ ) are 4 to 9‰ lower than the ones calculated from  
 334 the  $\delta\text{D}_{\text{Muscovite}}$  values ( $-73 \leq \delta\text{D}_{\text{Water}} \leq -60\text{‰}$ ), but both are comparable within uncertainty ( $< \pm$   
 335  $5\text{‰}$ ).

### 336 **3.5 Geochronology**

#### 337 **3.5.1 Muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ dating of mylonitic leucogranite**

338 Four single muscovite grains from QUIB01 and QUIB03 were step-heated for  $^{40}\text{Ar}/^{39}\text{Ar}$   
 339 analyses using a  $\text{CO}_2$  laser probe coupled with a MAP215 mass spectrometer (Figure 8; Text  
 340 5 and Tables 2 and 3 of the supplementary material). The experiments were carried out in  
 341 Geosciences Rennes, France. The analytical procedure described by Ruffet et al. (1995, 1991)  
 342 is provided as supplementary material (Text 5) with parameters used for calculations (Table  
 343 2). Analytical data are available in Table 3 of the supplementary material.

344 Two muscovite grains from QUIB03 provide consistent plateau ages in the 305 - 303.5 Ma  
 345 range (mean at  $304.1 \pm 1.4$  Ma), calculated over more than 90% of the total degassed  $^{39}\text{Ar}_K$ .

346 In contrast, two muscovite grains from QUIB01 yield older ages. One of the grains provides a  
 347 perfectly flat age spectrum allowing a plateau age to be calculated at  $319.5 \pm 1.9$  Ma ( $2\sigma$ ). The  
 348 age spectrum obtained on the other grain indicates younger apparent ages, in the range of 306-  
 349 309 Ma, and a characteristic saddle shape which will be discussed below.

#### 350 **3.5.2 Monazite, zircon and apatite U-Th/Pb dating of leucogranite**

351 Two samples of weakly deformed (QUIB21) and mylonitic (QUIB20) leucogranite from the  
 352 QDZ footwall were dated in the GeOHeLiS analytical platform (Univ Rennes 1), France, using  
 353 zircon, monazite and apatite LA-Q-ICP-MS U(-Th)/Pb dating (See location on the simplified  
 354 stratigraphic column of Figure 2, Figure 9; Text 6 and Tables 4, 5, 6 and 7 of the supplementary  
 355 material).

356 Monazite grains extracted from the weakly deformed leucogranite sample QUIB21 plot in a  
 357 concordant position in the  $^{206}\text{Pb}/^{238}\text{U}$  versus  $^{208}\text{Pb}/^{232}\text{Th}$  concordia diagram (Figure 9), with  
 358 apparent  $^{206}\text{Pb}/^{238}\text{U}$  ages ranging from 335 Ma down to 285 Ma (Table 5 of the supplementary  
 359 material). A group of twelve consistent and concordant analyses yield a date of  $318 \pm 2.2$  Ma  
 360 (MSDW = 0.5; Figure 9A). Analyses 13 and 22 are within error with this concordant group,  
 361 although they yield older apparent ages. The remaining data plot along the concordia curve  
 362 yielding apparent ages between 300 and 285 Ma. Apatite grains from the same sample yield a

363  $^{207}\text{Pb}$  corrected date of  $312.9 \pm 2.2$  Ma (MSDW = 0.5) equivalent within error with both the  
 364 free- ( $323 \pm 12$  Ma) and anchored- ( $313.2 \pm 3.3$  Ma) isochron dates (Figure 9B; Table 7 of the  
 365 supplementary material). Note that these apatite crystals are characterized by very consistent  
 366 REE spectra and a U content of  $\sim 70$  ppm (Figure 9C).

367 For the more deformed sample QUIB20, 23 measurements out of 34 zircon grains were  
 368 performed. Plotted in a Tera-Wasserburg diagram (Figure 9D), data are concordant to  
 369 discordant with  $^{206}\text{Pb}/^{238}\text{U}$  apparent ages ranging from 875 down to 315 Ma (Table 5 of the  
 370 supplementary material). Three groups of concordant data can be distinguished. The first and  
 371 oldest group yields a concordant date of  $676 \pm 2.3$  Ma ( $n=4$ , Zr 18 & 19), the second a  
 372 concordant date of  $335 \pm 2.3$  Ma ( $n=4$ , Zr 19 & 26) and the youngest one a concordant date of  
 373  $317 \pm 4$  Ma ( $n=5$ , MSDW = 1.2; Zr 3, 8, 16 & 19). In this sample, the REE spectra for the  
 374 apatite grains are different from one to the other and associated with a U content of  $\sim 50$  ppm  
 375 (Figure 9F). The QUIB20 apatite yield a  $^{207}\text{Pb}$  corrected date of  $304.9 \pm 3.2$  Ma (MSDW = 0.9)  
 376 that is comparable with the free ( $310 \pm 14$  Ma) and anchored ( $302.9 \pm 4.3$  Ma) isochron dates  
 377 (Figure 9E; Table 7 of the supplementary material).

### 378 3.5.3 Monazite U-Th/Pb REE petrochronology of migmatite

379 The melanosome/mesosome part of the Port Navalo migmatite located in the QDZ footwall  
 380 was dated and characterized using in-situ laser ablation petrochronology techniques at the  
 381 University of Portsmouth, UK (Full methodology and results can be found in Text 7 and Tables  
 382 8 and 9 of the supplementary material). Sample NAV04 yields a spread of concordant U-Pb  
 383 analyses (Figures 10A-10D) that range in  $^{206}\text{Pb}/^{238}\text{U}$  and  $^{208}\text{Pb}/^{232}\text{Th}$  apparent age from  $\sim 330$   
 384 to 290 Ma ( $n = 35$ ; Figure 10A). This spread is not statistically resolvable into separate  
 385 populations. All monazite analyses yield consistent trace element signatures, with relative  
 386 enrichments in LREE, significant Eu anomalies and a slight range in HREE (Figure 10F).  
 387 Monazite grains show faint zoning in BSE (Figure 10G) and some variation in their Y  
 388 concentrations (Table 8 of the Supplementary material). However, there is no systematic  
 389 variation in either REE or Y concentration with age, and in grains where multiple spots were  
 390 analyzed, they were within uncertainty and therefore non-resolvable.

## 391 4 Discussion

### 392 4.1 Interpretation of geochronological data

#### 393 4.1.1 Monazite petrochronology

394 Monazite grains from both the leucogranite and migmatite analyzed in this study yielded a  
 395 spread of apparent  $^{206}\text{Pb}/^{238}\text{U}$  ages that almost overlap within uncertainty e.g.  $319 \pm 8$  Ma –  
 396  $283 \pm 8$  Ma (leucogranite) and  $329 \pm 9$  Ma –  $288 \pm 8$  Ma (migmatite). This spread can be  
 397 interpreted either as Pb loss, incomplete resetting during fluid-related dissolution precipitation  
 398 (e.g. Grand'Homme et al., 2016), or due to a genuine spread in geological ages (e.g. Foster et  
 399 al., 2004, and commonly seen in Himalayan monazite ages). Monazite is resistant to Pb-loss  
 400 by volume diffusion alone (e.g. Cherniak et al., 2004; Gardés et al., 2007, 2006; Parrish, 1990),  
 401 however is prone to dissolution-precipitation (recrystallization; e.g. Hetherington et al., 2017).  
 402 This may result in the formation of patchy zoning at grain scale and discordance in the U-Th-  
 403 Pb system (e.g. Krohe and Wawrzenitz., 2000; Seydoux-Guillaume et al., 2012; Wawrzenitz  
 404 et al., 2015) causing disruption of the isotopic age due to incomplete resetting, and a trend of  
 405 data towards the right of a Tera Wasserburg Concordia diagram (Grand'Homme et al., 2016).  
 406 Monazites in our studied migmatite sample contain patchy zoning (Figure 10G) that could be  
 407 interpreted as evidence for dissolution-precipitation or growth zoning. The spread seen in the

408 data could therefore be interpreted as incomplete resetting or could represent a spread of  
 409 analyses between two populations. It is not possible within the precision of our dataset to  
 410 distinguish between these scenarios.

411 Monazite trace elements in the migmatite yield consistent Eu anomalies that can be interpreted  
 412 as peritectic monazite crystallization with feldspar from (partial) melt (Figure 10F). This is  
 413 consistent with observations of selvages of quartz and feldspar and lobate grain boundaries  
 414 (Figure 10E) as well as the presence of cordierite (Figure 5G) in thin section, microstructures  
 415 indicative of partial melting within the migmatitic host rock. Together with the U-Th-Pb data,  
 416 it can be interpreted that monazite grew during melting between ~330–290 Ma, either during  
 417 several stages (e.g. Foster et al., 2004; Mottram et al., 2014), or due to the older peritectic  
 418 monazite being partially reset during later fluid flow event(s) (e.g. Grand'Homme et al., 2016).

## 419 **4.1.2 Interpretation of Ar/Ar and U(-Th)/Pb ages**

### 420 **4.1.2.1 Syntectonic leucogranite emplacement in the QDZ footwall**

421 To determine the timing of meteoric fluid infiltration and subsequent meteoric fluid-rock  
 422 interactions that occurred in the QDZ footwall, we applied different geochronology methods  
 423 (U(-Th)/Pb on zircon, monazite, apatite and  $^{40}\text{Ar}/^{39}\text{Ar}$  on muscovite) to date both highly and  
 424 weakly deformed leucogranites (see summary in Figure 11).

425 We interpret the U-Th/Pb age of  $318 \pm 3$  Ma obtained on monazite from a weakly deformed  
 426 leucogranite (QUIB21) as its age of emplacement. This age is comparable to the U/Pb age of  
 427  $317 \pm 4$  Ma obtained on magmatic zircon from a similar but intensively deformed leucogranite  
 428 (QUIB20) collected at the top of the section that we also interpret to reflect the period of  
 429 syntectonic emplacement. Both ages are in agreement with the  $^{40}\text{Ar}/^{39}\text{Ar}$  age of  $319.7 \pm 1.9$  Ma  
 430 ( $2\sigma$ ) obtained on muscovite from the QUIB01 mylonitic leucogranite sample (this study),  
 431 which is similar to the muscovite  $^{40}\text{Ar}/^{39}\text{Ar}$  age of  $319.2 \pm 1.8$  Ma ( $2\sigma$ ) obtained in the  
 432 Questembert granite emplaced along the SASZ (Tartèse et al., 2011b, 2011c).

433 The similarity of the U(-Th)/Pb and  $^{40}\text{Ar}/^{39}\text{Ar}$  ages at ~318 Ma on the weakly and more  
 434 deformed granite types indicate that following their emplacement, leucogranites have  
 435 experienced a rapid cooling through the isotopic closure temperature ( $T_c$ ) of argon in  
 436 muscovite. Based on Harrison et al. (2009)'s experimental determinations of muscovite  
 437 diffusion coefficients, (Pitra et al., 2010) calculated a  $T_c$  of ~540°C for a diffusion radius of 1  
 438 mm and a cooling rate of 100°C/Ma. Therefore, our data agree with a rapid cooling and suggest  
 439 that, as proposed by Gapais et al. (1993), the QDZ may have played a major role in exhuming  
 440 rocks from the lower unit.

441 The presence of C-S structures in mylonitic leucogranites within the top ~500 m of the section  
 442 supports a syntectonic crystallization age of monazite at  $318 \pm 3$  Ma, and although such fabrics  
 443 are not clearly observed within the weakly deformed leucogranite from the bottom of the  
 444 section (~700 m) and dated at  $317 \pm 4$  Ma on zircon, the presence of biotite schlieren and their  
 445 orientations related to the magmatic lineation suggest that they emplaced while the detachment  
 446 zone was already active. Based on the similarity of the U(-Th)/Pb and  $^{40}\text{Ar}/^{39}\text{Ar}$  ages at ~318  
 447 Ma on the weakly and more deformed granite types, we conclude that 1) the QDZ was already  
 448 active at that time, 2) based on the structural and microstructural features, the mylonitic and  
 449 weakly deformed leucogranites emplaced during the QDZ activity, and that the deformation  
 450 was heterogenous over a thickness of ~700 m with highly deformed rocks at the top and weakly  
 451 deformed rocks at the bottom.

452 To help reconstructing the cooling history, we acquired additional  $^{40}\text{Ar}/^{39}\text{Ar}$  data from sample  
 453 QUIB03 from which two muscovite grains yielded two concordant ages of  $304.8 \pm 2.1$  Ma and

454  $303.4 \pm 2.1$  Ma ( $2\sigma$ ) (Figure 8). A similar age of  $305 \pm 3$  Ma has been obtained on apatite using  
 455 U/Pb geochronology from the mylonitic leucogranite QUIB20. The spread REE spectrum of  
 456 apatite supports the idea that these minerals experienced a partial leaching of REE during  
 457 hydrothermal alteration (see Ballouard et al., 2018a).

458 Although showing a saddle shape spectrum, a single muscovite grain from QUIB01 provides  
 459 a plateau age of  $307.7 \pm 1.7$  Ma with apparent ages ranging between  $309.0 \pm 1.5$  Ma and  $305.9$   
 460  $\pm 1.5$  Ma. When comparing with the plateau age of  $\sim 319.5$  Ma provided by a single muscovite  
 461 grain from the same sample (see above), these ages are younger and close to those obtained  
 462 from the QUIB03 muscovite and QUIB20 apatite grains (Figure 11). It is known that the partial  
 463 re/neo-crystallization of white mica can generate saddle-shaped age spectra that result from  
 464 distinctive degassing patterns of the initial/inherited and re/neo-crystallized domains for a  
 465 given crystal (see Alexandrov et al., 2002; Castonguay et al., 2007; Cheilletz et al., 1999;  
 466 Tartèse et al., 2011c; Tremblay et al., 2011). The observed saddle shape can result from two  
 467 distinct phenomena: (1) the mixing of two distinct radiogenic components that are imperfectly  
 468 separated, an initial/inherited one which could be as old as  $\sim 319.5$  Ma and a re/neo-crystallized  
 469 one which could be as young as  $\sim 304$  Ma; (2) the isotopic record of period of long protracted  
 470 recrystallization history from  $\sim 309$  to  $306$  Ma linked to deformation and/or fluids. However,  
 471 the relatively young ages obtained for mylonitic leucogranite (QUIB03 muscovite  $^{40}\text{Ar}/^{39}\text{Ar}$   
 472 ages of  $\sim 305$  and  $\sim 303.5$  Ma and QUIB20 apatite U/Pb age of  $\sim 305$  Ma) and the associated  
 473 apatite REE and muscovite  $\delta\text{D}$  results seem to be in favor of the second option, involving the  
 474 syntectonic fluid-assisted (re)crystallization of mica through time.

475 Apatite from weakly deformed leucogranites (QUIB21) yielded a U/Pb date of  $313 \pm 2$  Ma. In  
 476 agreement with the magmatic structures observed at the mesoscopic scale on the weakly  
 477 deformed leucogranite (Figure 3B), apatite display a narrow REE distribution pattern. Based  
 478 on the theoretical closure temperature of apatite ranging between  $350$  and  $550^\circ\text{C}$  (e.g. Chew  
 479 and Spikings, 2015; Pochon et al., 2016), we interpret this U/Pb date of  $313 \pm 2$  Ma as a cooling  
 480 age.

481 To summarize, two main events can be defined from these geochronological results:

482 1) An early phase at  $\sim 318$  Ma (U-Th/Pb on monazite and U/Pb on magmatic zircon) marked  
 483 by leucogranite emplacement in the Quiberon detachment footwall while the shear zone was  
 484 active. Following their emplacement, these leucogranites ensued a rapid cooling as highlighted  
 485 by  $^{40}\text{Ar}/^{39}\text{Ar}$  ages from the QUIB01 muscovite grains.

486 2) A late phase at  $\sim 305$  Ma characterized by a main magmatic-hydrothermal event highlighted  
 487 by apatite U/Pb and muscovite  $^{40}\text{Ar}/^{39}\text{Ar}$  ages, and apatite REE spectra and uranium content  
 488 from mylonitic leucogranites. Both apatite U/Pb and muscovite  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of  $\sim 305$  Ma are  
 489 in agreement with previous published muscovite  $^{40}\text{Ar}/^{39}\text{Ar}$  ages between  $\sim 304$  and  $301$  Ma  
 490 measured in similar mylonitic leucogranite from Quiberon (Gapais et al., 2015; Turrillot, 2010)  
 491 and U/Pb ages of  $\sim 303$  Ma acquired on zircon and monazite from an aplitic dyke of the  
 492 neighboring Guérande leucogranite (Ballouard et al., 2015).

#### 493 **4.1.2.2 Migmatization in the QDZ footwall**

494 U-Th/Pb monazite from the Port Navalo migmatite provided a spread of apparent ages  
 495 between  $\sim 330$ – $290$  Ma. A more precise age would be useful to explore the exact tectono-  
 496 magmatic and metamorphic link between the migmatites and leucogranites. However, our  
 497 relatively imprecise age range demonstrates that the lower-crustal migmatites were partially  
 498 melting while monazite crystallized coevally in the syntectonic leucogranite. Our dates also  
 499 suggest that there was several (tens) of millions of years of metamorphism and fluid-flow that  
 500 potentially (partially) reset the ages. This is consistent with a lower-crustal setting where rocks

501 are deformed, metamorphosed and experience partial melting in the presence of fluids over  
502 prolonged periods of time (such as in the Himalaya; Mottram et al., 2014).

## 503 **4.2 Syntectonic hydrothermal alteration during high-temperature deformation**

### 504 **4.2.1 Source of fluids**

505 Hydrogen isotopes measured on syntectonic muscovite from mylonitic leucogranite,  
506 ultramylonitic pegmatite and quartz veins allow to calculate  $\delta D_{\text{Water}}$  values from -73 to  $-51 \pm$   
507 5‰ that correspond to the hydrogen isotopic composition of water that equilibrated with  
508 muscovite in the detachment footwall (Table 1). Muscovite from mylonitic leucogranite  
509 samples provides the lowest calculated  $\delta D_{\text{Water}}$  values (-73 to -66‰) whereas the highest  
510  $\delta D_{\text{Water}}$  values (-62 to -51‰) have been obtained from muscovite from ultramylonitic  
511 pegmatite. Quartz veins and micaschist indicate intermediate  $\delta D_{\text{Water}}$  values (-67 to -63‰).

512 The difference of up to 22‰ between the  $\delta D_{\text{Water}}$  values extracted from the Quiberon mylonitic  
513 leucogranites and ultramylonitic pegmatites allow to identify two different reservoirs of fluids  
514 that have triggered intense fluid-rock interactions in the Quiberon detachment footwall. The  
515 new hydrogen dataset presented here extracted from various lithologies from the QDZ footwall  
516 (Figure 7) is in good agreement with the results published in Dusséaux et al. (2019) showing a  
517 regional-scale mixing relationship between deep crustal fluids ( $\delta D_{\text{Water}}$  values up to -33‰) and  
518 surface-derived fluids ( $\delta D_{\text{Water}}$  values down to -74‰) in syntectonic leucogranite emplaced at  
519 different crustal levels in the southern Armorican domain. Fluid inclusions aligned along  
520 synkinematic structural planes in quartz grains from detachment footwalls (Quiberon and  
521 Piriac; Dusséaux, 2019) contain very low to medium salinity water (0 to 7 wt% eq. NaCl) and  
522 yield  $\delta D$  and  $\delta^{18}\text{O}$  values plotting between the meteoric water line and the  
523 metamorphic/magmatic ranges, further strengthening this interpretation. This mixing  
524 relationship has also been demonstrated based on  $\delta D_{\text{Water}}$  values from syntectonic leucogranites,  
525 pegmatites, quartz veins and episyenites in the western part of the French Massif Central  
526 (merging to the northwest with the southern Armorican domain) where high  $\delta D_{\text{Water}}$  values of  
527 up to  $\sim -30$ ‰ reveal a signature of deep crustal fluids and  $\delta D_{\text{Water}}$  values as low as -104 ‰  
528 indicate a contribution of meteoric fluids sourced at high elevation (Dusséaux et al., 2021;  
529 Turpin et al., 1990).

530 In the Iberian Massif, other studies involving measurements of hydrogen and oxygen isotope  
531 ratios of hydrous minerals (phengite, chlorite) from altered granites and host metasediments  
532 (Rodríguez-Terente et al., 2018), or fluid inclusions from granite, mineralized veins and skarns  
533 (Tornos et al., 2000) systematically point to the mixing of fluids from different sources  
534 (magmatic/metamorphic and meteoric). For instance,  $\delta D_{\text{Water}}$  values from barren quartz veins  
535 ranging from -70 to -35‰ are interpreted to reflect strong meteoric water/rock interactions in  
536 the Iberian Massif (Martín Crespo et al., 2002). As a consequence, based on previous studies  
537 conducted in similar geological objects, the hydrogen isotope values obtained from syntectonic  
538 muscovite from the Quiberon granite are consistent with a mixing relationship between  
539 deuterium-depleted fluids and metamorphic/magmatic water with high  $\delta D$  values.

540 Due to the presence of graphite in the overlying micaschist, metavolcanics and quartz veins of  
541 the upper crust (e.g. Ballouard et al., 2017; Caroff et al., 2016), we cannot rule out the  
542 possibility that graphite crystallization from a methane-bearing fluid could have played a role  
543 in lowering the  $\delta D_{\text{Muscovite}}$  values from the Quiberon granitic samples (see Craw, 2002).  
544 However, the significant number of previous geochemical studies conducted in the southern  
545 Armorican domain and the western French Massif Central (See introduction, Ballouard et al.,  
546 2015, 2017, 2018a; Dusséaux, 2019; Dusséaux et al., 2019, 2021; Lemarchand et al., 2012;  
547 Tartèse et al., 2012, 2013), constitute a tangible proof that meteoric fluids penetrated the

548 extending and fractured upper crust to significant depths in the active footwall of detachment  
 549 zones during the Late Carboniferous, sometimes leading to the formation of uranium ore  
 550 deposits (Ballouard et al., 2017; Boulvais et al., 2019).

551 The  $\delta D_{\text{Water}}$  values found in pegmatites ( $-62 \leq \delta D_{\text{Water}} \leq -51\text{‰}$ ) are typical of deep  
 552 crustal fluids ( $-70\text{‰} < \delta D_{\text{metamorphic fluids}} < -20\text{‰}$  and/or  $-80\text{‰} < \delta D_{\text{magmatic fluids}} < -40\text{‰}$ ; e.g.  
 553 Field and Fifarek, 1985). However, the 11 ‰ difference ( $-62$  to  $-51\text{‰}$ ) within the  $\delta D_{\text{Water}}$  values  
 554 can be explained by variable meteoric fluid-rock ratios (e.g. Dusséaux et al., 2019). In a rock-  
 555 buffered system, pegmatites that intruded mylonitic leucogranites during the latest stages of  
 556 deformation may not have had enough time to equilibrate with meteoric fluids. In accordance  
 557 with pegmatite successive emplacement during syntectonic crystallization of late magmatic  
 558 fluids (Gapais & Boundi, 2014) and the regional-scale mixing relationship of Dusséaux et al.  
 559 (2019), we consider that the highest  $\delta D_{\text{Water}}$  value of  $-51\text{‰}$  most closely approximate the  
 560 hydrogen isotope composition of a magmatic fluid.

561 In contrast, the lowest  $\delta D_{\text{Water}}$  values ( $-73$  to  $-66\text{‰}$ ) obtained from mylonitic leucogranite  
 562 suggest that muscovite interacted with surface-derived deuterium-depleted fluids during  
 563 deformation (Dusséaux et al., 2019). As mentioned above, lozenge-shaped muscovite grains  
 564 (mica fish) located along shear bands and their associated low  $\delta D_{\text{Water}}$  values suggest that the  
 565 growth of micas occurred through solution-precipitation mechanisms during the QDZ activity  
 566 which was the site of intense meteoric water-rock interaction. These low  $\delta D_{\text{Water}}$  values are  
 567 strengthened by those down to  $-77\text{‰}$  obtained from tourmaline in mylonitic leucogranite  
 568 (Figure 7). In agreement with a  $\sim 5\text{‰}$  difference in hydrogen isotope fractionation between the  
 569 muscovite and the tourmaline at  $550^\circ\text{C}$  and  $0.3$  GPa (Blamart et al., 1989), the  $\delta D_{\text{Tourmaline}}$   
 570 values are systematically  $3$  to  $7 \pm 2\text{‰}$  lower than the  $\delta D_{\text{Muscovite}}$  values (Figure 7 and Table 1)  
 571 indicating that these two minerals have reached hydrogen isotope equilibrium during high-  
 572 temperature deformation. These  $\delta D_{\text{Water}}$  values down to  $-77\text{‰}$  corroborate previous studies that  
 573 interpreted  $\delta D_{\text{Water}}$  values down to  $-74\text{‰}$  provided by syntectonic leucogranites to reflect an  
 574 interaction with meteoric fluids during the activity of the Piriac and Quiberon detachment zones  
 575 (Dusséaux et al., 2019). However, the  $\delta D_{\text{Water}}$  value of Late Carboniferous meteoric fluids at  
 576 the surface may have been more negative than  $-74\text{‰}$  due to deuterium enrichment during the  
 577 downward penetration of meteoric and consequent meteoric fluid-rock interactions at low  
 578 meteoric-fluid rock ratios (e.g. Dusséaux et al., 2021; Gébelin et al., 2012).

#### 579 4.2.2 Timing and duration of fluid-rock interactions

580 Four lines of evidences lead us to conclude that the  $\sim 319$  Ma muscovite samples (QUIB01;  
 581 Figure 11) crystallized in presence of meteoric fluids during high temperature deformation in  
 582 the QDZ footwall: (1) muscovite from this sample provides the lowest  $\delta D_{\text{Muscovite}}$  value ( $-85\text{‰}$ ;  
 583 Figure 7), which is  $22\text{‰}$  lower than the ones in pegmatite, and has been interpreted to reflect  
 584 the highest meteoric fluid-rock ratio found in the QDZ footwall (Dusséaux et al., 2019); (2)  
 585 recrystallization involving solution-precipitation is indicated by primary lenticular mica fish  
 586 (Group 1; ten Grotenhuis et al., 2003) that evolved through time into a more bent secondary  
 587 mica fish with deflected tips due to continuous shearing (Group 2; ten Grotenhuis et al., 2003)  
 588 (Figure 4); (3) the chemical composition of these mica fish plot in both the primary and  
 589 secondary muscovite fields ( $0.02 < \text{Ti} < 0.05$  apfu; Figure 8); (4) the Ti-in-muscovite  
 590 thermometry results obtained from muscovite fish associated with low  $\delta D_{\text{Muscovite}}$  values  
 591 (Figure 7) indicated that they formed during high temperature deformation ( $> 500^\circ\text{C}$ ) which is  
 592 consistent with quartz grain boundary migration. Therefore, we conclude that  $319.5 \pm 1.9$  Ma  
 593 represents a minimum argon age for the infiltration of meteoric fluids which interacted with  
 594 synkinematic muscovite during high temperature deformation in the QDZ footwall.



595 Lenticular muscovite grains from the QUIB03 sample provide younger  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of ~305  
 596 - 303 Ma (Figure 11) that could reflect a period of intense hydrothermal activity because (1)  
 597 muscovite grains provide low  $\delta\text{D}_{\text{Muscovite}}$  values (<-80‰; Figure 7), (2) the chemical  
 598 composition of these mica fish have a tendency to plot into the field of hydrothermal muscovite  
 599 ( $0.02 < \text{Ti} < 0.03$  apfu), (3) group 1 mica fish (ten Grotenhuis et al., 2003) show evidence of  
 600 secondary newly recrystallized grains on the grains boundaries, (4) apatite from the same  
 601 mylonitic leucogranites that yield similar ages ( $305 \pm 4$  Ma; Figure 9) are associated with varied  
 602 REE spectra reflecting hydrothermal alteration.

603 We propose that meteoric fluid-rock interactions occurred in the footwall of the QDZ during  
 604 high-temperature deformation at least between  $319.5 \pm 1.9$  Ma and  $303.4 \pm 2.1$  Ma, supported  
 605 by syntectonic mica growth (see section 4 above; e.g. G ebelin et al., 2011; Kelley, 2002;  
 606 Tart ese et al., 2011b).

607 Field observations show that the walls of planar ultramylonitic pegmatite veins form an angle  
 608 of  $\sim 10^\circ$  with mylonitic leucogranite foliation suggesting syntectonic pegmatite emplacement  
 609 within the high strain zone while the ductile deformation was still active (Figure 2F). The high  
 610  $\delta\text{D}_{\text{Water}}$  values calculated from syntectonic muscovite from deformed pegmatites ranging from  
 611 -62‰ to -51‰ do not reflect a meteoric fluid signature but rather a magmatic imprint. By  
 612 comparison with mylonitic leucogranites from which  $\delta\text{D}_{\text{Water}}$  values between -77 and -66‰  
 613 have been deduced, we interpret the difference in reconstructed hydrogen isotopic composition  
 614 of meteoric water to reflect a difference in microstructure, chemical composition and  
 615 permeability of the rock and/or a short period of meteoric fluid-rock interactions.

616 The geochronological results presented in this study do not preclude the possibility that  
 617 meteoric fluid infiltration and detachment activity may have occurred before ~320 Ma and/or  
 618 after ~303 Ma. The  $\delta\text{D}$  values measured in leucogranite can be interpreted as the presence of  
 619 surface-derived fluids in the footwall during this time period (see paragraph 4.2.1). In addition,  
 620 sericitization along muscovite rims and in pressure shadows in mylonitic leucogranite (e.g.  
 621 Figure 4B) support fluid-assisted dissolution-precipitation during the final stages of the QSZ  
 622 activity as indicated by  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of ~305 and ~303.5 Ma. Moreover, both mylonitic and  
 623 cataclastic fabrics revealed in mylonitic leucogranite and pegmatite reveal a normal sense of  
 624 movement suggesting that brittle fabrics formed during exhumation of metamorphic footwall  
 625 rocks to the hanging wall along the Quiberon detachment zone while syntectonic granites were  
 626 crossing the brittle-ductile transition. A late, intense fluid circulation event sustained by high  
 627 (paleo-) geothermal gradient likely occurred in the Quiberon detachment footwall until ~298  
 628 Ma as indicated by monazite U/Pb dates of  $297 \pm 8$  Ma and  $298 \pm 8$  Ma from the Port Navalo  
 629 migmatites (Figure 10) and shown by ~298 Ma muscovite  $^{40}\text{Ar}/^{39}\text{Ar}$  ages obtained on late shear  
 630 bands and deformed pegmatite from the Quiberon peninsula (Gapais et al., 2015; Turrillot,  
 631 2010).

### 632 4.2.3 Implications for elemental mobility

633 REE spectra from apatite grains show a difference in the uranium content from high (~70 ppm)  
 634 in the weakly deformed granite to low (~50 ppm) in the mylonitic leucogranite. Based on the  
 635 ages obtained for these two granites, we can say that the uranium content decreased through  
 636 time from at least ~313 Ma (QUIB21) to ~305 Ma (QUIB20). These observations agree with  
 637 the deuterium depletion in muscovite and tourmaline identified in the younger granite,  
 638 associated to a spread REE spectrum reflecting REE partial leaching from apatite. Therefore,  
 639 we propose that the depletion in uranium and rare earth elements in apatite and in deuterium  
 640 detected in muscovite and tourmaline are related to a same fluid-rock interaction event that  
 641 occurred during the final stage of the Quiberon magmatic-hydrothermal system activity. This

642 agrees with studies of uranium deposits in the southern Armorican domain (e.g. Ballouard et  
643 al., 2017, 2018a, 2018b; Boulvais et al., 2019; Tartèse et al., 2013) showing that meteoric fluids  
644 leached uranium from the magmatic uraninite in syntectonic peraluminous granites emplaced  
645 in the footwall of detachments (Piriatic, and in this study, Quiberon) or along the SASZ (e.g.  
646 Pontivy, Questembert). On their way back to the surface, the oxidized uranium-bearing fluids  
647 interacted with reducing black schists and triggered the precipitation of hydrothermal uraninite.  
648 At the broader scale of the Variscan belt of Western Europe, coeval late-orogenic extensional  
649 shear zones, hydrothermal fluid flow, leucogranite syntectonic emplacement and granulite-  
650 facies metamorphism in the lower crust led to U, W, Au, Sb, P and rare earth element  
651 mineralization (e.g. Ballouard et al., 2017; Boiron et al., 2003; Bouchot et al., 2005; Cuney et  
652 al., 2002; Cuney, 2014; Harlaux et al., 2018).

### 653 **4.3 Mechanisms and depth of meteoric fluid infiltration**

654 The infiltration of surface-derived fluids down to the detachment footwall could be explained  
655 through fractures and steep normal faults (Figure 2A) that developed in the extending brittle  
656 upper crust and soled down on to the QDZ (e.g. Gébelin et al., 2017, 2015, 2013, 2011; Mulch  
657 et al., 2006, 2004; Person et al., 2007). In addition to brittle structures that could have served  
658 as conduits for fluids and generated the necessary porosity for fluid percolation from the  
659 surface down to the middle crust, anastomosing C-S structures (Figures 2, 4 and 5) localize the  
660 strain, weaken the rock and promote permeability (e.g. Bauer et al., 2000; Hunter et al., 2016;  
661 McCaig, 1988; Tartèse et al., 2013). Meteoric fluid infiltration to substantial depths involves  
662 the presence of both hydraulic (e.g. Person et al., 2007) and hydrostatic heads (e.g. Sutherland  
663 et al., 2017; Upton & Sutherland, 2014) and can be achieved through dip slip faulting-related  
664 dilatancy (e.g. Nüchter and Ellis, 2011) that episodically injects small amount of meteoric  
665 fluids and leads to protracted periods of fluid-rock interactions (e.g. Sibson, 1981; Upton et al.,  
666 1995) in agreement with our rock- buffered meteoric fluid signatures.

667 In addition, driving forces are essential to maintain fluid-rock interactions for more than 15 Ma  
668 in the active ductile segment of the QDZ. As suggested for other fossil hydrothermal systems  
669 (e.g. Gébelin et al., 2011, 2015, 2017; Methner et al., 2015; Person et al., 2007), advection of  
670 partially molten materials in the lower crust can sustain buoyancy-driven fluid convection  
671 while the detachment is active and continuously exhume rocks from the footwall to the hanging  
672 wall. Together with radioactive heat production (e.g. Jolivet et al., 1989; Vigneresse et al.,  
673 1989), partial melting and leucogranite emplacement in the southern Armorican domain from  
674 ~320 to ~305 Ma (Figure 11; e.g. Augier et al., 2015; Ballouard, 2016; Ballouard et al., 2018a;  
675 Lemarchand et al., 2012; Peucat, 1983; Tartèse et al., 2012, 2011a, 2011c; Turrillot et al., 2011,  
676 2009; Turrillot, 2010), the new U-Th/Pb ages acquired on the Port Navalo migmatites ranging  
677 from ~330 to ~ 290 Ma (Figure 10) suggest that high heat flow occurred at the same time as  
678 the mylonitization across the QDZ ( $^{40}\text{Ar}/^{39}\text{Ar}$  on muscovite and U/Pb on apatite and monazite  
679 from the leucogranite; Figures 8 and 9), helping to sustain the hydrothermal system.

680 Coeval lower crustal migmatization and syntectonic leucogranite emplacement has been dated  
681 in similar rocks at ~315 Ma in the western part of the French Massif Central that constitutes  
682 the eastern extent of the southern Armorican domain (e.g. Gébelin et al., 2009). In the broader  
683 framework of the Variscan belt, the results from this study fit well with coeval syntectonic  
684 leucogranite emplacement along ductile shear zones and lower crustal partial melting  
685 recognized widely in the European Variscan hinterlands at the end of the Carboniferous (e.g.  
686 Ballouard et al., 2015; Gébelin et al., 2009; López-Moro et al., 2012; Padovano et al., 2014;  
687 Rolland et al., 2009; Tartèse et al., 2012). Age-equivalent migmatization and leucogranite  
688 emplacement was also recognized in the Montagne Noire (southern part of the French Massif

689 Central), considered as the foreland of the belt (e.g. Pitra et al., 2021; Poujol et al., 2017;  
690 Whitney et al., 2015).

691 A hydraulic head generated in high-relief areas is also essential for the penetration of surface-  
692 derived fluids at depth. As demonstrated by stable isotope paleoaltimetry estimates for the  
693 western French Massif Central (minimum mean elevation of  $3400 \pm 700\text{m}$ ; Dusséaux et al.,  
694 2021), the Variscan Belt of Western Europe was standing at high elevation during the late  
695 Carboniferous. These paleoaltimetry results cannot necessarily be applied to our study area that  
696 represents the north-western extension of the French Massif Central. However, based on the  
697 internally thickened crust (e.g. Ballèvre et al., 2013) and preliminary stable isotope  
698 paleoaltimetry results indicating a mean paleoelevation of  $\sim 2500\text{ m}$  for the southern Armorican  
699 domain (Dusséaux, 2019), this region probably imposed a high regional hydraulic head that  
700 allowed fluids to migrate down to the brittle-ductile transition and reach the QDZ footwall (e.g.  
701 Gébelin et al., 2013, 2012; Raphaël Gottardi et al., 2013; Person et al., 2007).

702 Although it is difficult to define accurately the depth of meteoric water incursion, we propose  
703 that meteoric fluids reached the brittle-ductile transition zone at a depth of  $\sim 8\text{ km}$  based on a  
704  $300^\circ\text{C}$  isotherm approximately corresponding to the start of brittle-ductile deformation in  
705 quartzo-feldspathic rocks (Stöckhert et al., 1999) and a warm Variscan geothermal gradient of  
706  $\sim 35\text{-}40^\circ\text{C/km}$  (e.g. Vanderhaeghe et al., 2020). However, the local depth of the brittle-ductile  
707 transition may be shallower as the active QDZ together with hot material advection may have  
708 led to extreme geothermal gradient and tight isotherms across the detachment zone (e.g.  
709 Gottardi et al., 2011; Gébelin et al., 2011).

## 710 **5 Conclusion**

711 This study documents the infiltration of meteoric fluids in the footwall of one of the best  
712 exposed Variscan extensional shear zones, the Quiberon detachment (QDZ) in the southern  
713 domain of the Armorican Massif (France). Combined structural, hydrogen isotope,  
714 geochemical and geochronological data from syntectonic leucogranites emplaced in the QDZ  
715 footwall allow meteoric fluid-rock interactions at depth to be bracketed between  $\sim 320$  and  $303$   
716 Ma. Synkinematic hydrous minerals recrystallized and equilibrated in the QDZ footwall with  
717 deuterium-depleted water during high temperature deformation ( $\sim 500^\circ\text{C}$ ) deduced from  
718 microstructural observations and titanium-in-muscovite thermometry. Mica fish provide  
719  $^{40}\text{Ar}/^{39}\text{Ar}$  ages between  $\sim 320$  and  $\sim 303\text{ Ma}$  that, together with microstructural observations,  
720 suggest that fluid-rock interactions occurred for at least  $17\text{ Ma}$ . This interpretation is in good  
721 agreement with U(-Th)/Pb data obtained on magmatic zircon and monazite from the same  
722 leucogranites that indicate an age of emplacement of  $\sim 318\text{ Ma}$ . In addition, U/Pb ages, REE  
723 spectra and uranium content of apatite in leucogranites strengthen the idea that meteoric fluid-  
724 rock interactions occurred for several million years and played a major role in leaching uranium  
725 at  $305 \pm 3\text{ Ma}$ . U-Th/Pb ages from  $\sim 330$  to  $290\text{ Ma}$  acquired on migmatites that form the lower  
726 crust suggest that, as demonstrated in Metamorphic Core Complexes of the Western part of the  
727 US (e.g. Gébelin et al., 2015, 2011; Mulch et al., 2004), high heat flow below the detachment  
728 is essential to sustain convection of fluids at depth while the upper crust undergoes extension  
729 as demonstrated by brittle fractures and late shear bands dated at  $\sim 302 - 298\text{ Ma}$  (Gapais et al.,  
730 2015; Turrillot, 2010).

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## Timing and duration of meteoric water infiltration in the Quiberon detachment zone (Armorican Massif, Variscan belt, France)

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**Table 1.** GPS locations, hydrogen isotope composition ( $\delta D$ ) of muscovite (Ms) and tourmaline (To) from leucogranite, micaschist, quartz vein and pegmatite found in the mylonitic footwall of Quiberon detachment zone.  $\delta D_{\text{Water}}$  values have been calculated by using the hydrogen isotope muscovite-water and tourmaline-water fractionation factors ( $\alpha$ ) of Suzuoki and Epstein [1976] and Kotzer et al. [1993], respectively, and using temperatures indicated by the Ti-in-Ms thermometer ( $546$  and  $569 \pm 42^\circ\text{C}$  for QUIB03 and QUIB01 respectively, and the average temperature of  $558 \pm 42^\circ\text{C}$  for the other samples). Calculated  $\delta D_{\text{Water}}$  values have propagated uncertainties of  $\pm 5.2\text{‰}$ , considering the precision of isotopic analyses ( $\delta D_{\text{hydrous silicate}} \pm 2\text{‰}$ ) and the uncertainties linked to the temperature of recrystallization ( $T \pm 42^\circ\text{C}$  results in  $\delta D_{\text{Water}}$  uncertainties of  $\pm 5\text{‰}$ ). Structural distances of samples below the estimated detachment interface are indicated in m (see text for explanation).



Name	Rock type	Distance (m)	$\delta D_{\text{Muscovite}}$ (‰) $\pm 2\text{‰}$ 250 $\mu\text{m} < f$	$\delta D_{\text{Tourmaline}}$ (‰) $\pm 2\text{‰}$ 250 $\mu\text{m} < f$	$\delta D_{\text{Ms}}$ - $\delta D_{\text{Tro}}$ (‰)	Temperature (°C) based on Ti-in-Ms thermometry	$\pm$ (°C)	$\delta D_{\text{Water}}$ (‰) based on the $\delta D_{\text{Muscovite}}$ values and the Ti-in-Ms thermometry			$\delta D_{\text{Water}}$ (‰) based on the $\delta D_{\text{Tourmaline}}$ values and the Ti-in-Ms thermometry			Location	
								$\delta D_{\text{Water}}$ (‰)	-	+	$\delta D_{\text{Water}}$ (‰)	-	+	Latitude (DD)	Longitude (DD)
QUIB 07	Micaschist	200	-76			558	42	-63	-3	4				47°30'21.85"N	03°09'00.54"W
QUIB 08	Quarz vein	250	-76	-81	5	558	42	-63	-3	4	-70	-4	4	47°30'26.2"N	03°08'58.6"W
QUIB 09	Quarz vein	255	-76			558	42	-63	-3	4				47°30'26.1"N	03°08'58.6"W
QUIB 01	Mylonitic leucogranite	300	-85	-87	2	<b>569</b>	<b>42</b>	-73	-2	4	-77	-3	5	47°29'0.85"N	3° 8'42.68"W
QUIB 02	Mylonitic leucogranite	305	-79			558	42	-66	-3	4				47°29'0.85"N	3° 8'42.68"W
QUIB 06	Ultramylonitic pegmatite	320	-75			558	42	-62	-3	4				47°29'0.85"N	3° 8'42.68"W
QUIB 04	Ultramylonitic pegmatite	321	-64			558	42	-51	-3	4				47°29'0.85"N	3° 8'42.68"W
QUIB 05	Ultramylonitic pegmatite	321.5	-73	-80	7	558	42	-60	-3	4	-69	-4	4	47°29'0.85"N	3° 8'42.68"W
QUIB 03	Mylonitic leucogranite	322	-82	-86	4	<b>546</b>	<b>41</b>	-68	-4	3	-74	-5	3	47°29'0.85"N	3° 8'42.68"W
QUIB 16	Quarz vein	465	-79			558	42	-66	-3	4				47°28'54.5"N	03°08'11.0"W
QUIB 17	Quarz vein	500	-80	-85	5	558	42	-67	-3	4	-74	-4	4	47°28'54.3"N	03°08'07.8"W

Propagated uncertainties	Error	Error (‰)	Propagated error in $\delta D_{\text{Water}}$
Isotopic analysis ( $\delta D$ )	$\pm 2\text{‰}$	$\pm 2\text{‰}$	$\pm 5.2\text{‰}$
Temperature deduced from Ti-in-Ms geothermometry	$\pm 42^\circ\text{C}$	$\pm 5\text{‰}$	

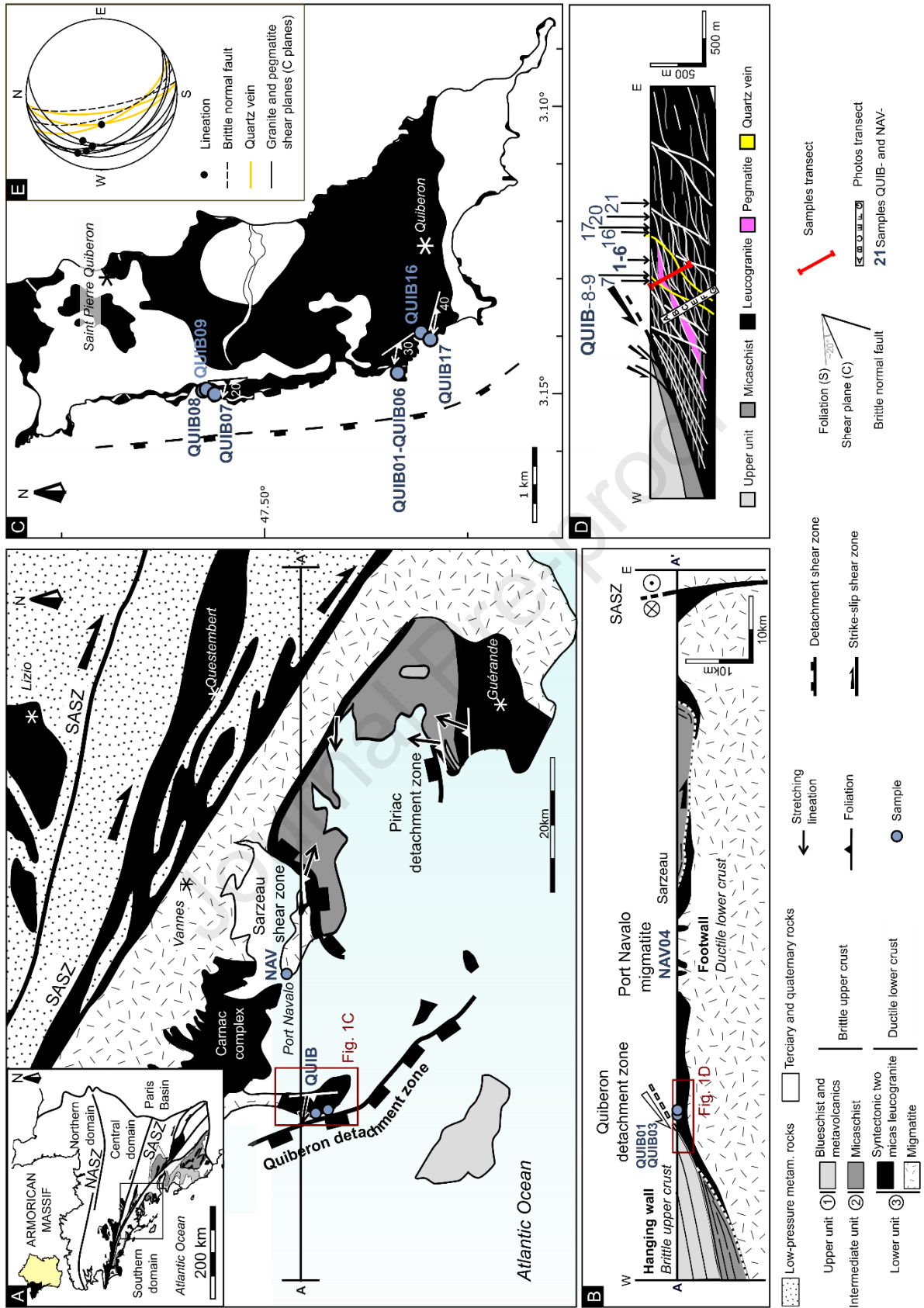
## Timing and duration of meteoric water infiltration in the Quiberon detachment zone (Armorican Massif, Variscan belt, France)

Camille Dusséaux<sup>1\*</sup>, Aude Gébelin<sup>1</sup>, Philippe Boulvais<sup>2</sup>, Gilles Ruffet<sup>2</sup>, Marc Poujol<sup>2</sup>, Nathan Cogné<sup>2</sup>, Yannick Branquet<sup>2,3</sup>, Catherine Mottram<sup>4</sup>, Fabrice Barou<sup>5</sup>, Andreas Mulch<sup>6,7</sup>

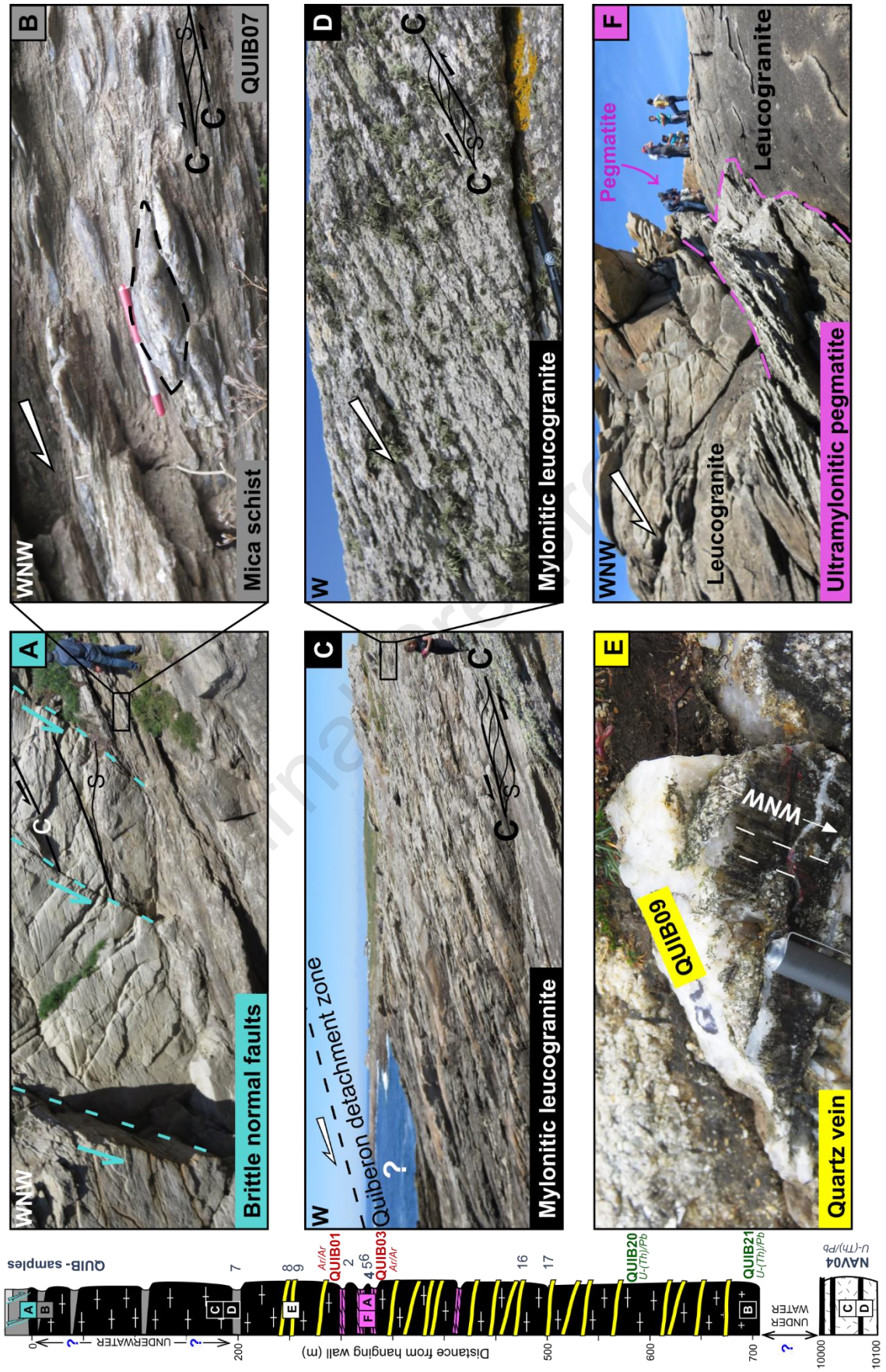
### Figure 1.

(A) Map of the southern Armorican domain. SASZ: South Armorican Shear zone; (B) W-E cross-section across the Quiberon detachment and the SASZ; (C) Map of the Quiberon island; (D) W-E cross-section and (E) stereographic projections showing the main structures and rock types found in the Quiberon leucogranite footwall. Modified after Gapais et al (1993, 2015), Thinon et al., (2008) and Turillot (2010). Sample coordinates are indicated in Table 1.





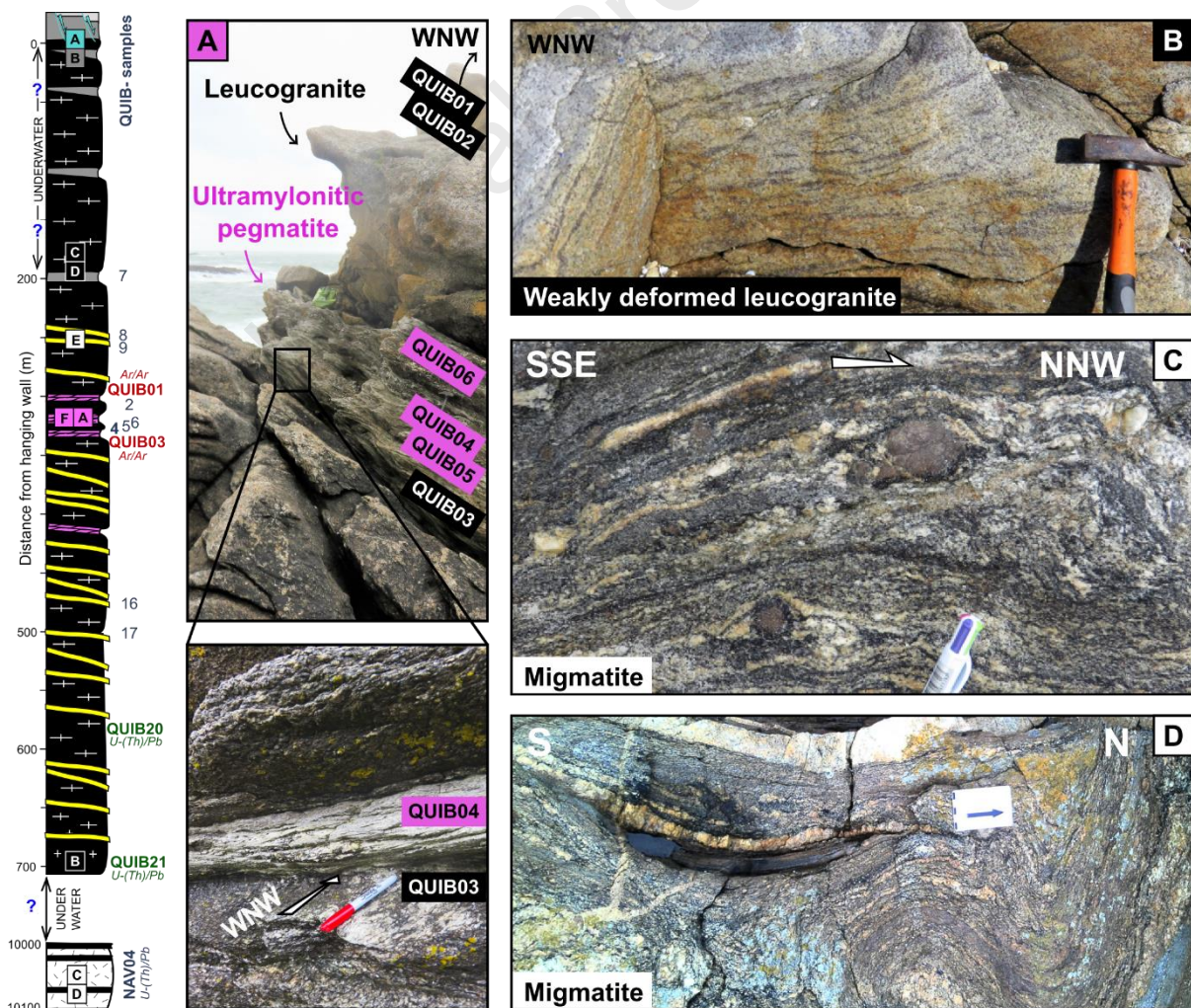






**Figure 2.** Lithologic section in the Quiberon detachment footwall and associated field pictures; (A) Brittle normal faults in micaschist, (B) Sigmoidal quartz in micaschist (sample QUIB07), (C) Syntectonic leucogranite emplaced in the footwall of the Quiberon detachment zone, (D) Mylonitic leucogranite with C-S structures highlighting a top-to-the-WNW sense of shear, (E) West-dipping quartz vein with WNW-ESE trending lineation supported by tourmaline; (F) High-strain zone made of ultramylonitic pegmatite (see text for explanation and Table 10 of the supplementary material for GPS coordinates).

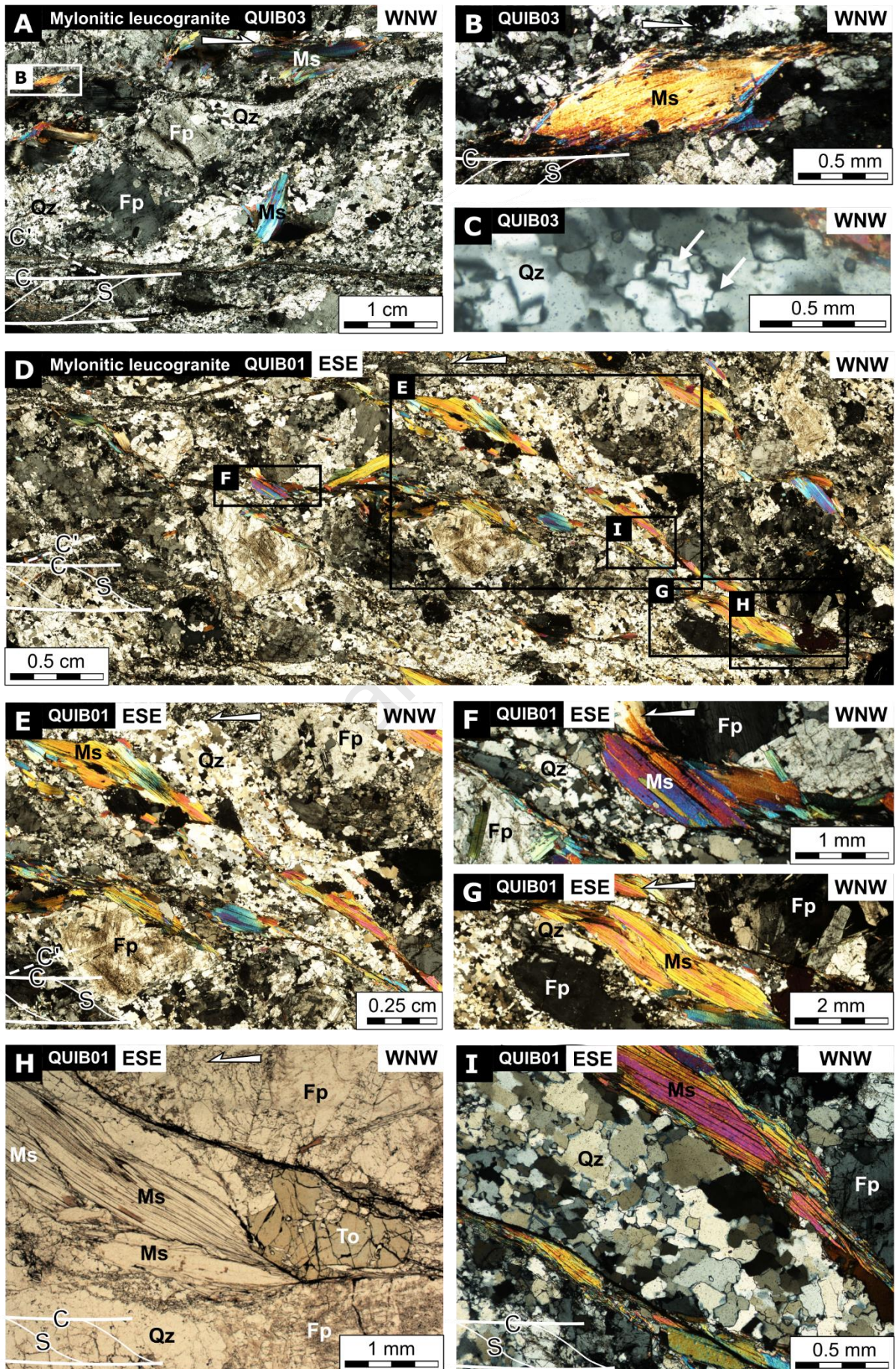
**Figure 3.** Lithologic section in the Quiberon detachment footwall and associated field pictures; (A) Ultramylonitic pegmatite (QUIB04, 05, 06) intruding coarser-grained mylonitic leucogranite (QUIB01, 02, 03), (B) Magmatic foliation in weakly deformed leucogranite, (C) and (D) Folded paleosome/mesosome and leucosome layers characterizing the migmatites at Port-Navalo (see text for explanation and Table 10 of the supplementary material for GPS coordinates).

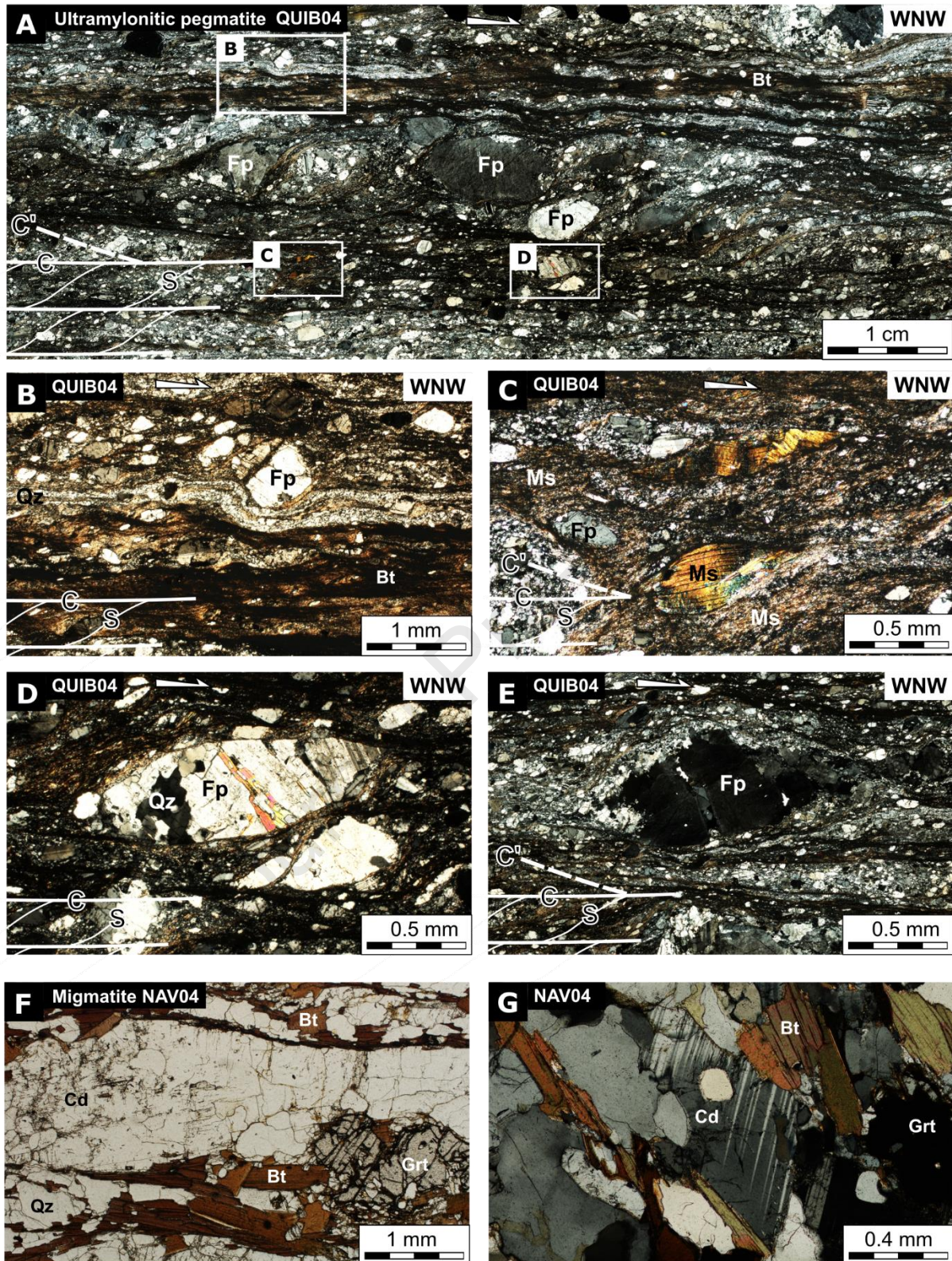


**Figure 4.** Microstructures from the Quiberon detachment footwall. Sections are cut perpendicular to foliation and parallel to lineation. Ms: muscovite; Qz: quartz; Fp: felspar; To: Tourmaline. (A-E) Mylonitic leucogranite (QUIB01); (B) C-S structures; (C) Lenticular muscovite fish (group 1 of ten Grotenhuis et al., 2003); (D) Group 2 mica fish that form C-S structures indicating a top-to-the-WNW sense of shear; (E) muscovite fish in equilibrium with tourmaline form shear planes in mylonitic leucogranite; (F) Sub-solidus deformation microstructures such as rectangular and castellate quartz grain boundaries suggest that grain boundary migration (regime 3, Hirth & Tullis, 1992) was the dominant dynamic recrystallization process; (G-I) Mylonitic leucogranite (QUIB03); (H) Group 1 micafish; (I) castellated quartz grain boundary. See text for explanation and Table 10 of the supplementary material for GPS coordinates.



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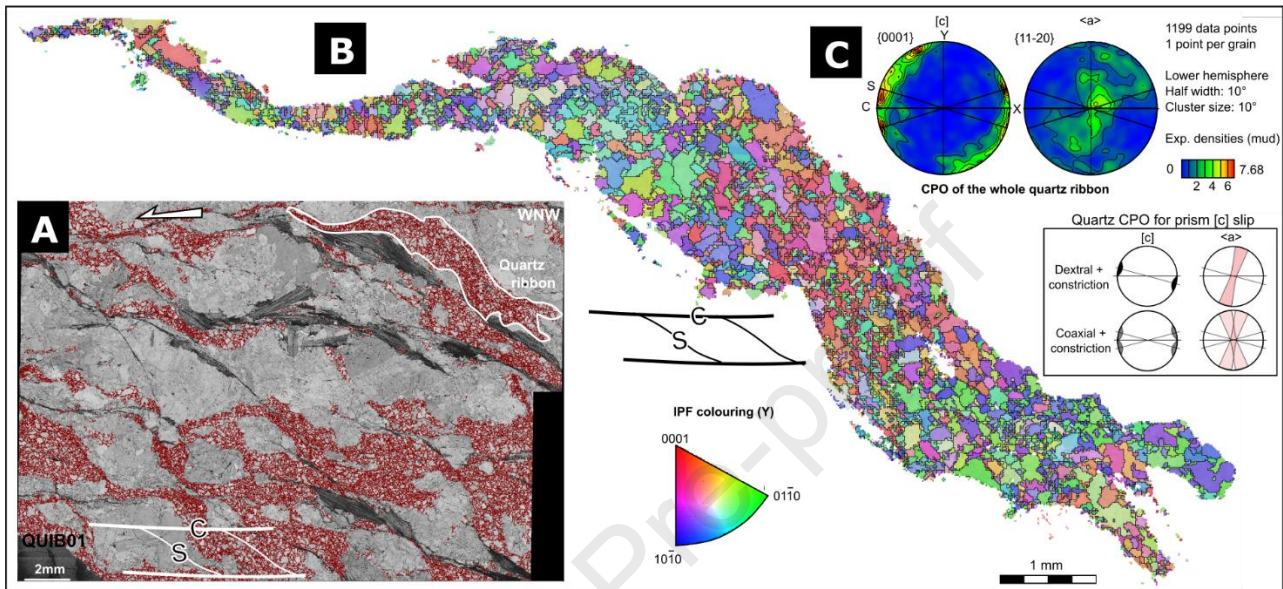




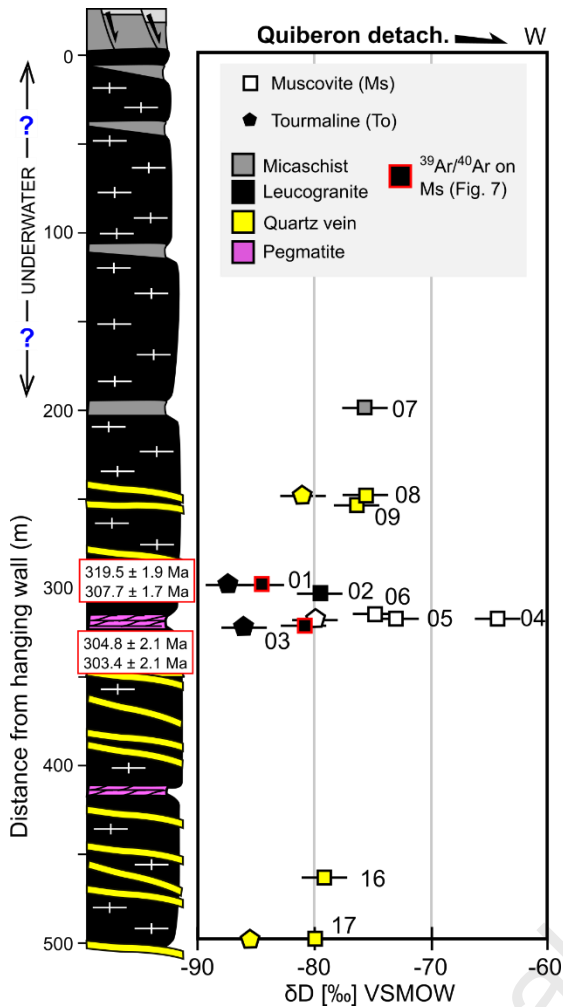
**Figure 5.** Microstructures from the Quiberon detachment footwall. Sections are cut perpendicular to foliation and parallel to lineation. Ms: muscovite; Qz: quartz; Fp: feldspar; Bt: biotite; Cd: cordierite; Grt: garnet. (A-C) Ultramylonitic pegmatite (QUIB04); (D-E) Garnet-cordierite-bearing migmatite from Port Navalo (NAV04). See text for explanation and Table 10 of the supplementary material for GPS coordinates.

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**Figure 6.** Microstructure and Crystallographic Preferred Orientation (CPO) of quartz grains from the QUIB01 mylonitic leucogranite sample measured using EBSD. Equal-area projection, Lower hemisphere. Foliation (XY plane) is vertical, and lineation (X) is horizontal in this plane. (A) Band contrast map with quartz grain boundary highlighted in red, (B) Map of a quartz ribbon quartz grain boundary with Inverse Pole Figure (IPF) coloring (Y represents the lineation direction) and (C) corresponding CPO. Framed quartz CPO for prism  $\langle c \rangle$  slip come from *Barth et al.* [2010].



**Figure 7.** Hydrogen isotope analysis ( $\delta D$  [‰] VSMOW  $\pm 2$ ‰) of hydrous silicates (muscovite and tourmaline) from micaschist, leucogranite, quartz veins and pegmatite located in the footwall of the Quiberon detachment zone with respect to the distance from the hanging wall ( $\sim 200$  to  $500$  m).

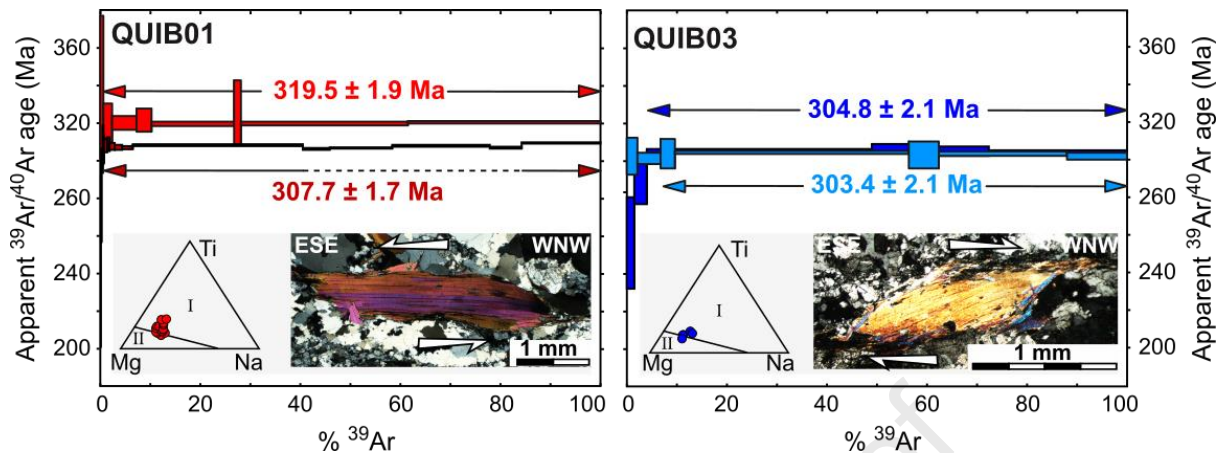
**Table 1.** GPS locations, hydrogen isotope composition ( $\delta D$ ) of muscovite (Ms) and tourmaline (To) from leucogranite, micaschist, quartz vein and pegmatite found in the mylonitic footwall of Quiberon detachment zone.  $\delta D_{\text{water}}$  values have been calculated by using the hydrogen isotope muscovite-water and tourmaline-water fractionation factors ( $\alpha$ ) of Suzuoki and Epstein [1976] and Kotzer et al. [1993], respectively, and using temperatures indicated by the Ti-in-Ms thermometer ( $546$  and  $569 \pm 42^\circ\text{C}$  for QUIB03 and QUIB01 respectively, and the average temperature of  $558 \pm 42^\circ\text{C}$  for the other samples). Calculated  $\delta D_{\text{water}}$  values have propagated uncertainties of  $\pm 5.2$ ‰, considering the precision of isotopic analyses ( $\delta D_{\text{hydrous silicate}} \pm 2$ ‰) and the uncertainties linked to the temperature of recrystallization ( $T \pm 42^\circ\text{C}$  results in  $\delta D_{\text{water}}$  uncertainties of  $\pm 5$ ‰). Structural distances of samples below the estimated detachment interface are indicated in m (see text for explanation).



Name	Rock type	Distance (m)	$\delta D_{\text{Muscovite}}$ (‰) $\pm 2\text{‰}$ 250 $\mu\text{m} < f$	$\delta D_{\text{Tourmaline}}$ (‰) $\pm 2\text{‰}$ 250 $\mu\text{m} < f$	$\delta D_{\text{Ms}} - \delta D_{\text{To}}$ (‰)	Temperature (°C) based on Ti-in-Ms thermometry	$\pm$ (°C)	$\delta D_{\text{Water}}$ (‰) based on the $\delta D_{\text{Muscovite}}$ values and the Ti-in-Ms thermometry			$\delta D_{\text{Water}}$ (‰) based on the $\delta D_{\text{Tourmaline}}$ values and the Ti-in-Ms thermometry			Location	
								$\delta D_{\text{Water}}$ (‰)	-	+	$\delta D_{\text{Water}}$ (‰)	-	+	Latitude (DD)	Longitude (DD)
QUIB 07	Micaschist	200	-76			558	42	-63	-3	4				47°30'21.85"N	03°09'00.54"W
QUIB 08	Quarz vein	250	-76	-81	5	558	42	-63	-3	4	-70	-4	4	47°30'26.2"N	03°08'58.6"W
QUIB 09	Quarz vein	255	-76			558	42	-63	-3	4				47°30'26.1"N	03°08'58.6"W
QUIB 01	Mylonitic leucogranite	300	-85	-87	2	<b>569</b>	<b>42</b>	-73	-2	4	-77	-3	5	47°29'0.85"N	3° 8'42.68"W
QUIB 02	Mylonitic leucogranite	305	-79			558	42	-66	-3	4				47°29'0.85"N	3° 8'42.68"W
QUIB 06	Ultramylonitic pegmatite	320	-75			558	42	-62	-3	4				47°29'0.85"N	3° 8'42.68"W
QUIB 04	Ultramylonitic pegmatite	321	-64			558	42	-51	-3	4				47°29'0.85"N	3° 8'42.68"W
QUIB 05	Ultramylonitic pegmatite	321.5	-73	-80	7	558	42	-60	-3	4	-69	-4	4	47°29'0.85"N	3° 8'42.68"W
QUIB 03	Mylonitic leucogranite	322	-82	-86	4	<b>546</b>	<b>41</b>	-68	-4	3	-74	-5	3	47°29'0.85"N	3° 8'42.68"W
QUIB 16	Quarz vein	465	-79			558	42	-66	-3	4				47°28'54.5"N	03°08'11.0"W
QUIB 17	Quarz vein	500	-80	-85	5	558	42	-67	-3	4	-74	-4	4	47°28'54.3"N	03°08'07.8"W

Propagated uncertainties	Error	Error (‰)	Propagated error in $\delta D_{\text{Water}}$
Isotopic analysis ( $\delta D$ )	$\pm 2\text{‰}$	$\pm 2\text{‰}$	$\pm 5.2\text{‰}$
Temperature deduced from Ti-in-Ms geothermometry	$\pm 42^\circ\text{C}$	$\pm 5\text{‰}$	



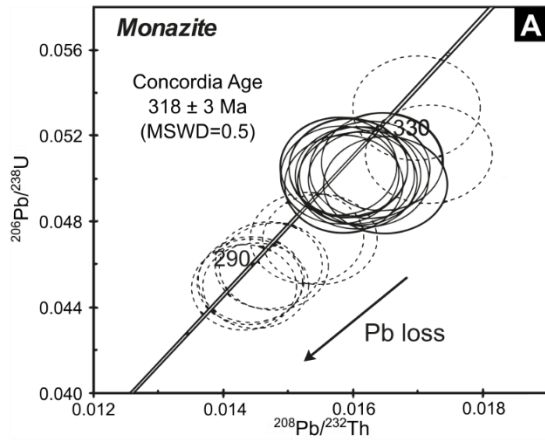


**Figure 8.**  $^{40}\text{Ar}/^{39}\text{Ar}$  step-heating spectra of muscovite from mylonitic leucogranite samples QUIB01 and QUIB03, with associated ternary Mg–Ti–Na diagram for each analyzed muscovite grain that allow to decipher between (I) the primary and (II) the secondary muscovite fields (Miller et al., 1981) and aspect of muscovite fish (see text). Note that micafish of QUIB01 show clear grain boundaries whereas QUIB03 micafish display secondary recrystallization of small mica grains on rims. Apparent age errors are plotted at the  $1\sigma$  level and calculated ages are indicated at  $2\sigma$ .

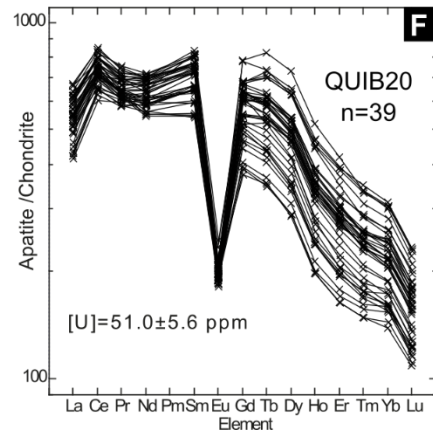
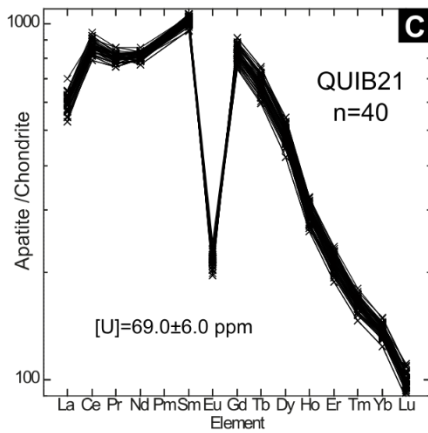
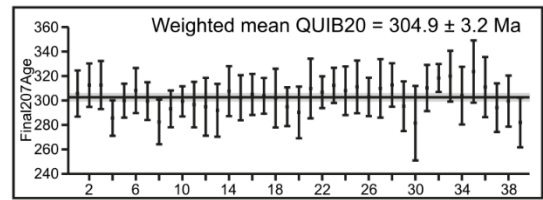
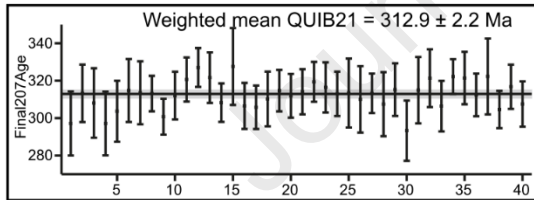
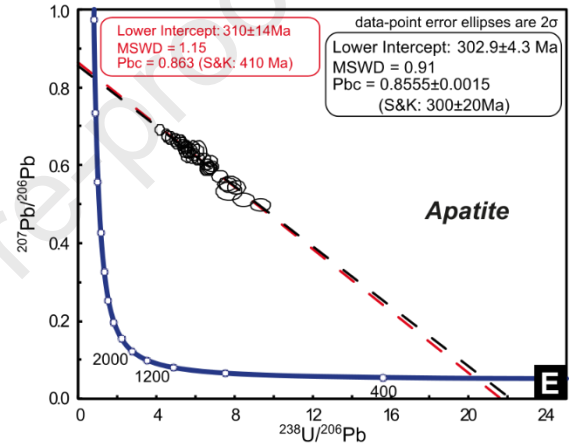
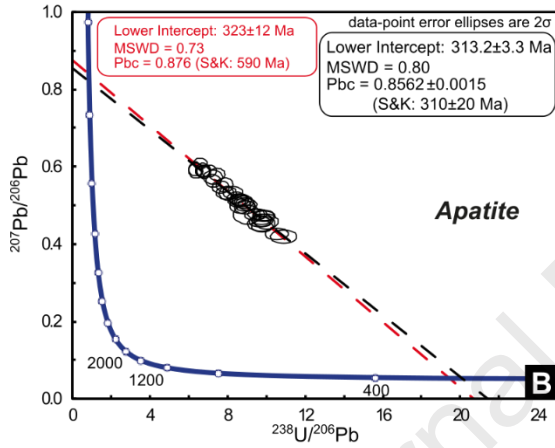
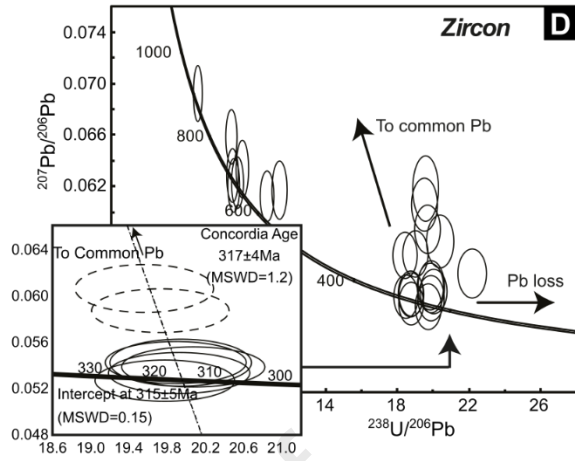
**Figure 9.** Monazite, zircon and apatite U–Th/Pb dating of weakly deformed (QUIB21) and mylonitic (QUIB20) Quiberon leucogranite samples. Results from the weakly deformed leucogranite sample: (A)  $^{206}\text{Pb}/^{238}\text{U}$  versus  $^{208}\text{Pb}/^{232}\text{Th}$  concordia diagram for monazite; (B)  $^{238}\text{U}/^{206}\text{Pb}$  versus  $^{207}\text{Pb}/^{206}\text{Pb}$  diagram with free- and anchored- isochron dates and weighted mean of the  $^{207}\text{Pb}$  corrected dates obtained on apatite (unforced age in red and forced age in black); (C) apatite REE spectra. Results from the mylonitic leucogranite: (D) Tera–Wasserburg diagram for zircon; (E)  $^{238}\text{U}/^{206}\text{Pb}$  versus  $^{207}\text{Pb}/^{206}\text{Pb}$  diagram with free- and anchored- isochron dates and weighted mean of the  $^{207}\text{Pb}$  corrected dates obtained on apatite; (F) apatite REE spectra. Note that apatite REE spectra is homogeneous in the weakly deformed sample (QUIB20) compared to the variable REE spectra in the mylonitic leucogranite sample (QUIB21). S&K: Stacey and Kramers [1975] lead evolution model.

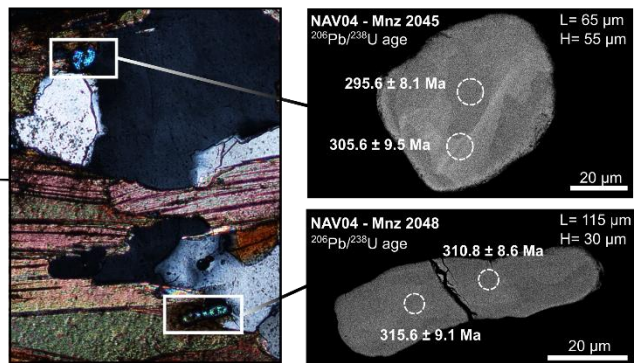
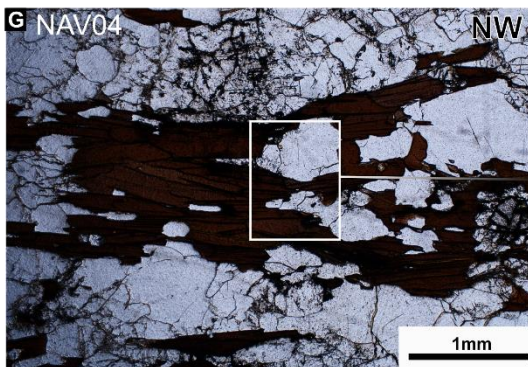
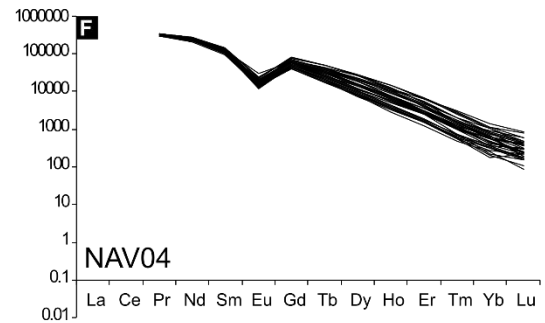
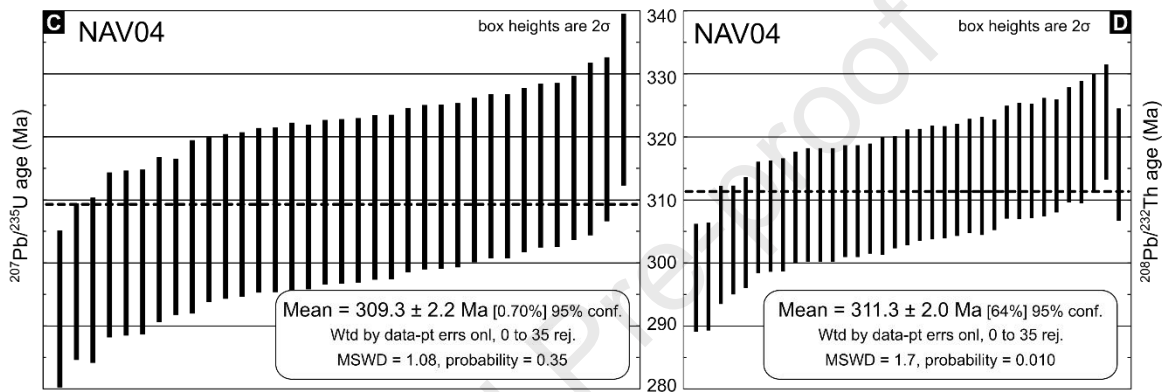
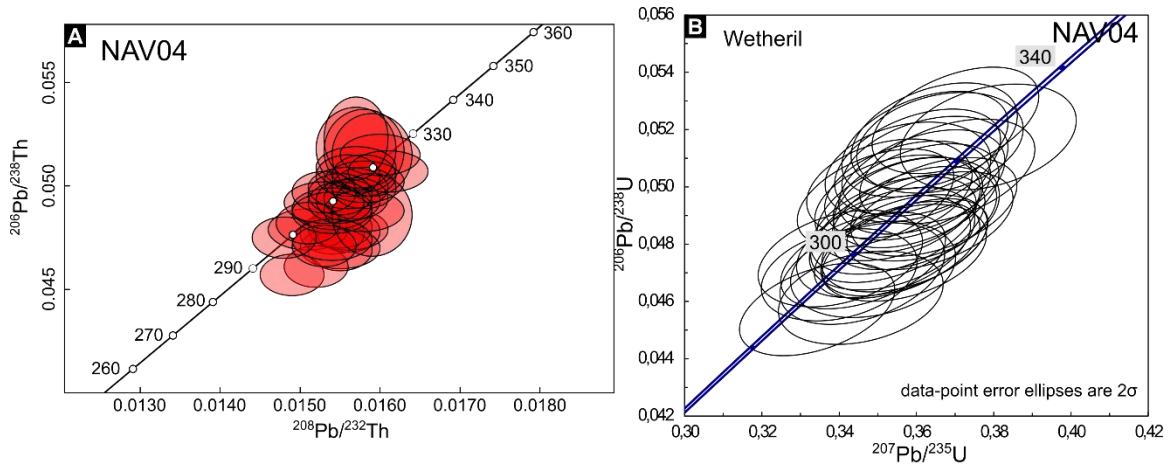


**Weakly deformed granite QUIB21**



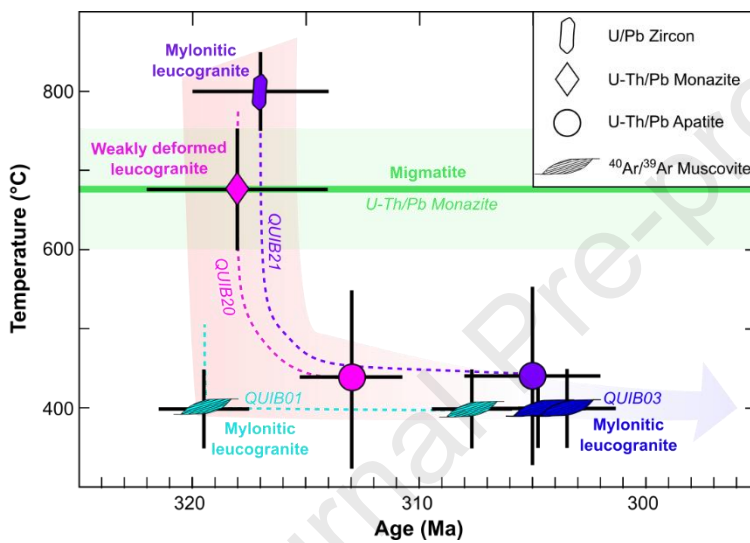
**Deformed granite QUIB20**







**Figure 10.** Monazite U-Th/Pb ages obtained on Port-Navalo migmatite (sample NAV04) located in the footwall of the Quiberon detachment zone. (A)  $^{206}\text{Pb}/^{238}\text{U}$  vs  $^{208}\text{Pb}/^{238}\text{Th}$  diagram; (B) Wetherill diagram; (C) Weighted average of the  $^{207}\text{Pb}/^{235}\text{U}$  ages; (D) Weighted average of the  $^{208}\text{Pb}/^{232}\text{Th}$  ages; (E) Quartz grain boundary migration; (F) Monazite REE spectrum highlighting a magmatic signature; (G) Microstructural aspect of monazite grains and associated  $^{206}\text{Pb}/^{286}\text{U}$  spot ages.



**Figure 11.** Temperature (°C) versus time (Ma) plot summarizing the geochronology results obtained for weakly deformed to mylonitic leucogranite in the Quiberon footwall.

1                   **Timing and duration of meteoric water infiltration in the**  
2 **Quiberon detachment zone (Armorican Massif, Variscan belt, France)**

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16 **Highlights**

- 17       •  $\delta D$  values of synkinematic hydrous minerals document the presence of 300 Ma-old  
18 meteoric fluids in a key Variscan detachment zone.  
19       • Meteoric fluid-rock interactions occurred during high-temperature deformation  
20 ( $>500^{\circ}\text{C}$ ) in the Quiberon detachment footwall.  
21       • Coeval detachment activity, leucogranite emplacement and migmatization allowed the  
22 hydrothermal system to be maintained for  $\sim 17$  Myr.

**Declaration of interests**

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

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