

1 **Earthquakes as precursors of ductile shear zones in the dry and strong lower crust**

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15
16 **Key Points:**

- 17 • Multiple generations of pseudotachylytes and mylonites formed at 650-750 °C and 0.7-
18 0.8 GPa in dry anorthosites from Lofoten (Norway)
- 19 • Strain localized in recrystallized pseudotachylytes deforming by diffusion creep
- 20 • Earthquakes are agents of weakening in the dry and strong lower crust
- 21

22 Abstract

23 The rheology and the conditions for viscous flow of the dry granulite facies lower crust are still
 24 poorly understood. Viscous shearing in the dry and strong lower crust commonly localizes in
 25 pseudotachylyte veins, but the deformation mechanisms responsible for the weakening and
 26 viscous shear localization in pseudotachylytes are yet to be explored. We investigated examples
 27 of pristine and mylonitized pseudotachylytes in anorthosites from Nusfjord (Lofoten, Norway).
 28 Mutual overprinting relationships indicate that pristine- and mylonitized pseudotachylytes are
 29 coeval and resulted from the cyclical interplay between brittle and viscous deformation. The
 30 stable mineral assemblage in the mylonitized pseudotachylytes consists of plagioclase,
 31 amphibole, clinopyroxene, quartz, biotite, \pm garnet \pm K-feldspar. Amphibole-plagioclase
 32 geothermobarometry and thermodynamic modelling indicate that pristine- and mylonitized
 33 pseudotachylytes formed at 650-750°C and 0.7-0.8 GPa. Thermodynamic modelling indicates
 34 that a limited amount of H₂O infiltration (0.20-0.40 wt%) was necessary to stabilize the mineral
 35 assemblage in the mylonite. Diffusion creep is identified as the main deformation mechanisms in
 36 the mylonitized pseudotachylytes based on the lack of crystallographic preferred orientation in
 37 plagioclase, the high degree of phase mixing, and the synkinematic nucleation of amphiboles in
 38 dilatant sites. Extrapolation of flow laws to natural conditions indicates that mylonitized
 39 pseudotachylytes are up to 3 orders of magnitude weaker than anorthosites deforming by
 40 dislocation creep, thus highlighting the fundamental role of lower crustal earthquakes as agents
 41 of weakening in strong granulites.

42 1 Introduction

43 Most of crustal seismicity is confined towards the base of the brittle upper crust at
 44 temperatures $< 300^\circ\text{C}$ [e.g. *Chen and Molnar*, 1983]. Interaction between lower crustal
 45 deformation and the earthquake cycle in the overlying seismogenic zone is observed as
 46 postseismic afterslip and interseismic creep in the lower crust [*Bodin and Horton*, 2004; *Wright*
 47 *et al.*, 2013]. However, the lower crust is also involved in earthquake nucleation [e.g., *Maggi et*
 48 *al.*, 2000; *Fagereng*, 2013], and this indicates a mechanically strong lower crust [*Maggi et al.*,
 49 2000; *Jackson et al.*, 2004], at least transiently [*Handy and Brun*, 2004].

50 Important volumes of pseudotachylytes (generally interpreted as quenched melts caused
 51 by the frictional heat released during seismic slip) in exhumed lower crustal rocks, and formed at
 52 lower crustal conditions, have been taken as direct geological evidence for deep fracturing
 53 [*Austrheim and Blundy*, 1994; *Jackson et al.*, 2004; *Austrheim*, 2013]. Examples of deep
 54 fracturing is particularly common in dry rocks, and it is generally accepted that a strong,
 55 seismogenic lower crust requires anhydrous granulite facies material [*Jackson et al.*, 2004].
 56 Anhydrous conditions in granulites inhibit crystal plastic flow and result in high strength
 57 [*Jackson et al.*, 2004; *Menegon et al.*, 2011]. Furthermore, anhydrous granulites can survive
 58 metastably in the course of Wilson cycles, unless they are infiltrated by fluids that trigger
 59 metamorphic reactions [*Austrheim*, 2013]. A picture is emerging in which a seismically active
 60 lower crust provides an environment for enhanced metamorphic and rheological transitions, due
 61 to fluid-rock interaction and associated reaction weakening in the fractured domains [*Austrheim*,
 62 2013; *Austrheim et al.*, 2017]. The observation that viscous shearing in a dry and strong lower
 63 crust often localizes in pseudotachylyte veins [*Pennacchioni and Cesare*, 1997; *White*, 1996;
 64 2012; *Steltenpohl et al.*, 2006; 2011; *Austrheim*, 2013; *Pittarello et al.*, 2013] is consistent with
 65 this picture. However, the deformation mechanisms responsible for the associated weakening

66 and viscous shear localization in pseudotachylyte veins are yet to be fully explored. Furthermore,
67 whether lower crustal pseudotachylytes develop by fault-related frictional melting [McKenzie
68 and Brune, 1972], by thermal runaway in ductile shear zones developing extreme localization
69 and accelerated creep rates [John *et al.*, 2009], or by in-situ amorphization of crystalline material
70 at high stresses and strain rates [Pec *et al.*, 2012] is still controversial.

71 Lofoten is one of the rare localities where the interplay between fracturing and viscous
72 flow, and the processes leading to strain localization in dry, granulitic lower crust can be
73 accurately studied, because (1) Lofoten exposes a relatively "intact" lower crust largely
74 consisting of anhydrous granulites [Corfu, 2004], and (2) lower crustal shear zones in Lofoten
75 are commonly associated with large volumes of pseudotachylytes [Steltenpohl *et al.*, 2006;
76 2011]. The study presented here investigates an exceptionally well preserved network of coeval
77 lower crustal pseudotachylytes and mylonites from Lofoten, and discusses the conditions for
78 pseudotachylytes development and the deformation mechanisms responsible of the ductile shear
79 localization in pseudotachylyte veins.

80

81 **2 Geological Setting**

82 Lofoten islands in northern Norway consist of a NNE-SSW trending basement horst that
83 is part of the Baltic Shield (Fig. 1a). Lofoten represents a tectonic window beneath the
84 Caledonian orogen and is composed of an Archean to Paleoproterozoic metamorphic complex of
85 para- and orthogneisses intruded, between 1870 and 1770 Myr, by a large Anorthosite-
86 Mangerite-Charnockite-Granite (AMCG) suite [Griffin *et al.*, 1978; Corfu, 2004]. The AMCG
87 suite consists of several plutons of anhydrous composition [Markl *et al.*, 1998; Corfu, 2004]
88 emplaced into the granulitic crust at ambient conditions estimated at 750-800°C, and in the range
89 between 0.4 and 1.0-1.2 GPa [Markl *et al.*, 1998, and references therein]. The primary igneous
90 texture and the granulite facies mineral assemblage are generally well preserved.

91 Lofoten largely escaped the Caledonian tectono-metamorphic overprint, as result of
92 water-deficient conditions that inhibited crystal plastic deformation and metamorphic re-
93 equilibration of granulites [Steltenpohl *et al.*, 2004; Leib *et al.*, 2016; Okudaira *et al.*, 2017].
94 Despite the lack of a regional post-intrusive fabric, eclogite and upper amphibolite facies,
95 localized shear zones are common in Lofoten [Kullerud *et al.*, 2001; Steltenpohl *et al.*, 2006,
96 2011; Menegon *et al.*, 2013]. Fracturing and pseudotachylyte formation have been reported to be
97 key processes for shear zone initiation in Lofoten [Steltenpohl *et al.*, 2006; Menegon *et al.*, 2013,
98 2015], and available evidence indicates that the pseudotachylytes developed under lower crustal
99 conditions [Moecher and Steltenpohl, 2009].

100 The Nusfjord eastern ridge in Flakstadøy (Fig. 1) exposes one of the largest bodies of
101 anorthosites in Lofoten [Markl *et al.*, 1998]. The Nusfjord anorthosites are massive, coarse-
102 grained rocks (1-5 cm average grain size of plagioclase) without any evident solid-state fabric.
103 These rocks host a network of shear zones containing both pseudotachylytes and ultramylonites
104 that are the main subject of this paper.

105 3 Methods

106 3.1 Microstructural observations and Electron Backscatter Diffraction (EBSD) analysis

107 Deformation microstructures and petrography were investigated using light and scanning
108 electron microscopy (SEM). Observations were made on polished thin sections cut perpendicular
109 to foliation and parallel to the stretching lineation (mylonites) and perpendicular to the
110 pseudotachylyte vein boundary. SEM analysis was performed at the Plymouth University
111 Electron Microscopy Centre using a JEOL LV6610 SEM and a JEOL 7001 FEG-SEM.

112 EBSD data were acquired on the JEOL LV6610 SEM equipped with a NordlysNano
113 EBSD detector (Oxford Instruments). Working conditions during acquisition of the EBSD
114 patterns were 20 kV, 70° sample tilt and high vacuum. EBSD patterns were acquired and
115 indexed with the AZtec software on rectangular grids with step size of 0.8 µm and 2.6 µm, and
116 processed with the Channel 5 software (Oxford Instruments).

117 3.2 Mineral chemistry

118 Major element mineral chemistry was measured with the JEOL 8200 Super Probe
119 electron micro-probe analyser (EMPA) at the Department of Earth Sciences, University of
120 Milan, Italy. Working conditions were 15 kV accelerating voltage, 5 nA current, 1 µm beam
121 diameter, using wavelength dispersive spectrometry (WDS) and natural silicates as standards.
122 All standards were calibrated within 0.5 % at one standard deviation and a PhiRhoZ routine was
123 used for matrix correction.

124 Mineral chemistry of amphibole-plagioclase pairs, used for the of P-T estimates, was
125 measured with the Cameca SX100 microprobe at the School of Earth Sciences, University of
126 Bristol, UK. Working conditions were 20 kV accelerating voltage, 5nA current, 1 µm beam
127 diameter, using wavelength dispersive spectrometry (WDS) and natural silicates as
128 standards. Fe^{3+} of amphiboles and garnets has been calculated from stoichiometry.

129 3.3 XRF analysis of bulk-rock chemical composition

130 Whole rock chemical analysis of major elements was performed by Wavelength
131 Dispersive X-Ray Fluorescence (WD-XRF) analysis with a Philips PW2400 equipped with a
132 rhodium tube at the Department of Geosciences, University of Padova, Italy. Powder samples
133 were mixed and diluted at 1:10 with $\text{Li}_2\text{B}_4\text{O}_7$ and LiBO_2 flux and melted into glass beads. Loss
134 on Ignition (LOI) was determined from weight lost after ignition at 860°C for 20 minutes and at
135 980°C for 2 hours. FeO was determined with permanganometry using a rhodium tube.

136 3.4 Thermodynamic modelling

137 H_2O -saturated P-T and T- H_2O isochemical sections were calculated with *Perple_X*
138 software [Connolly, 2005], in the system $\text{SiO}_2\text{-Al}_2\text{O}_3\text{-MgO-CaO-Na}_2\text{O-K}_2\text{O-TiO}_2\text{-FeO-(Fe}_2\text{O}_3)$,
139 using the bulk composition of the mylonitized pseudotachylyte sample N13-10D and of the host
140 anorthosite sample N1310B (Table 1). We used the thermodynamic database and equation of
141 state for H_2O of *Holland and Powell* [1998, upgrade 2002] and no phase was excluded from the
142 calculation. We used the solution models of *Holland and Powell* [1998, 2003] for olivine,

143 orthopyroxene, clinopyroxene, garnet, spinel, phengite and feldspar, of *Tajcmanova et al.* [2009]
144 for Ti-biotite, and of *Dale et al.* [2005] for calcic amphibole.

145 3.5 SIMS analysis

146 Intracrystalline hydrogen content in plagioclase was measured on polished thin sections
147 using the Cameca ims-4f Secondary Ion Mass Spectrometry (SIMS) at the NERC Ion
148 Microprobe Facility in Edinburgh, UK. Prior to analysis the sample was placed in an oven at
149 105°C for 30 minutes to remove surface humidity. The sample was then coated with a gold film
150 (<0.02µm) and kept in the SIMS chamber under high vacuum (5×10^{-9} Torr) for > 48 hours.
151 Analysis was performed with a 5nA primary beam of $^{16}\text{O}^-$ ions accelerated to 14.5 kV. To avoid
152 surface contamination, few nm of surface material were removed by sputtering the spot for 4
153 minutes while rastering the beam (25 µm grid) before each measurement. Measurements were
154 acquired using a beam spot size of 20 µm (± 5 µm). Each analysis consisted of 20 cycles of the
155 isotopes ^1H , ^{23}Na , ^{30}Si , ^{39}K and ^{49}Ti . For hydrogen only the last 15 cycles were averaged as ^1H
156 signals stabilized only after the first cycles. Previously determined SiO_2 concentration was used
157 as internal standard. Probing locations were identified beforehand and attention was paid to
158 avoid cracks, grain-boundaries and alterations, which could contain free water. Basaltic glass
159 standard St81A9 [*Lesne et al.*, 2011] was used to calibrate the water content in plagioclase.
160 Anhydrous olivine standard (Kilbourne Hole) was used to correct for background ^1H signals
161 (estimated at 12 wt ppm H_2O).

162

163 4 Field observations

164 The eastern ridge of Nufjord peninsula is dominated by coarse-grained anorthosites with
165 minor troctolites and gabbros. Satellite images show a network of 3 main sets of lineaments
166 trending ENE-WSW to NE-SW (*Set 1*), NW-SE (*Set 2*), and NNE-SSW (*Set 3*) (Figs. 1b-c). In
167 the field, the lineaments correspond to major (of as much as a few meters wide), steeply dipping
168 mylonitic shear zones associated with sub-parallel 2nd order (cm-dm thick) discrete shear zones
169 in between (Fig. 2). All three sets contain associations of pseudotachylytes and mylonites. The
170 study presented here focuses on the *Set 1* shear zones, because it contains the largest volumes of
171 pseudotachylytes in the area.

172 Set 1 shear zones dip steeply mostly towards the SE and contain a moderately to steeply
173 plunging stretching lineation (Fig. 1c). The shear zones show normal-oblique kinematics, with a
174 top-to-S component. The simplest type of 2nd order shear zone consists of a thin sharply
175 bounded, homogeneously fine-grained (Figs. 2a-c), dark green sheared layer. This layer is
176 typically surrounded by a < 10 cm wide bleached halo (contrasting in colour with the dark host
177 anorthosite), which was partly involved in shearing (Figs. 2a-c). The bleached halo results from a
178 pervasive network of white-coloured layers forming transgranular micro-fractures or exploiting
179 the cleavage of the coarse anorthosite plagioclase (Fig. 2c).

180 The dark layers are homogeneously fine grained, regardless of the accommodated strain
181 (Fig. 2d), indicating that the fine grain size did not result from progressive grain size refinement,
182 but was a pristine feature of the shear zone precursor. These layers are identified as sheared

183 pseudotachylyte fault veins as indicated by the local preservation of pseudotachylyte breccia
 184 pockets (Fig. 2e) and injection veins (Fig. 2f-g), and by the local occurrence of undeformed
 185 pseudotachylytes trending parallel to the shear zones (Fig. 2h).

186 Thick shear zones have commonly a domainal foliation defined by elongated, alternating
 187 domains of whitish and dark-green colour (Fig. 3a-b). These latter domains are fine grained and
 188 similar to the deformed pseudotachylytes of the 2nd order shear zones; locally they are discordant
 189 to the main foliation (Fig. 3a). There is no compositional heterogeneity in the pristine anorthosite
 190 to explain the origin of such compositional banding of mylonitic rocks. The dark domains tend to
 191 anastomose around the white domains that represent stretched portions of the bleached/damaged
 192 host anorthosite. With increasing strain both types of domains tend to form a more regular,
 193 planar layering and macroscopic plagioclase porphyroclasts disappear. Large, low strain domains
 194 associated with some major shear zones locally show thick (20-50 cm) pseudotachylyte breccias
 195 with cm-sized lithic clasts of bleached anorthosite and clast-free layers of pseudotachylyte (Fig.
 196 3c), parallel to the shear zone boundary. The origin of the sheared dark domains from
 197 pseudotachylytes is indicated by the common occurrence of undeformed pseudotachylyte
 198 injection veins in the anorthosite wall rock of shear zones.

199 In some cases, the pseudotachylyte veins crosscut the mylonitic foliation. In Fig. 3d, the
 200 foliation of a shear zone, that includes sheared pseudotachylytes, is involved in a breccia
 201 cemented by a pseudotachylyte that exploited the shear zone boundary. Fig. 3e shows a
 202 pseudotachylyte-bearing fault sharply displacing a localized ductile shear zone and being in turn
 203 dragged into a ductile shear zone that exploits older pseudotachylyte veins (Fig. 3f).

204

205 **5 Microstructure**

206 **5.1 Anorthosite host rock**

207 Anorthosite consists of large (as much as 20 cm in size) plagioclase crystals and
 208 subordinate (< 10%) clinopyroxene, orthopyroxene and olivine. (Fig 4a). Clinopyroxene is the
 209 most common mafic phase. Hydrous phases are very rare (< 1 %) and consist of local biotite
 210 rims around orthopyroxene and of discontinuous amphibole coronas around clinopyroxene .
 211 Corundum locally occurs as idiomorphic single crystals or as aggregates included in mm-sized
 212 magnetite. A detailed description of the Nusfjord anorthosites is reported in *Markl et al.* [1998].

213 **5.2 Pseudotachylyte and associate damage zone**

214 Pseudotachylytes locally preserve the pristine microstructure that includes microlites of
 215 plagioclase, clinopyroxene, amphibole and orthopyroxene (Figs. 4a-c), flow structures, and
 216 chilled margins (Fig. 4d). Pristine pseudotachylytes consist of single-generation individual veins
 217 with no evidence of multiple veins along the same fault.

218 The pseudotachylyte matrix is microcrystalline and dominantly consists of very fine-
 219 grained (< 5 μm) plagioclase. Plagioclase, and rarer clinopyroxene and orthopyroxene, occur as
 220 clasts in the pseudotachylyte matrix together with lithic clasts. The shape of clasts ranges from
 221 angular to smoother shapes showing resorption structures. In the damage zone flanking

222 pseudotachylyte, plagioclase and clinopyroxene show undulatory extinction, bent twins and bent
223 lamellae (in pyroxenes). The grains show fractures filled with new small polygonal grains (20-30
224 μm in size) forming bands up to 100 μm wide (Fig. 4a). In clinopyroxene the intracrystalline
225 bands of new grains locally contain minor amounts of amphibole.

226 Some pseudotachylytes have completely lost the pristine microlitic microstructure and
227 have recrystallized into a fine-grained (average grain size $\sim 10 \mu\text{m}$) mixture of plagioclase,
228 amphibole, clinopyroxene, biotite, quartz \pm K-feldspar \pm orthopyroxene \pm ilmenite (Fig. 4e).
229 Since these pseudotachylytes do not show a mylonitic overprint, we refer to them as to
230 *recrystallized pseudotachylytes*.

231 Garnet is common in pristine and recrystallized pseudotachylytes as well as in the wall
232 rock anorthosite. In pristine pseudotachylytes, garnet occurs as small (50-100 μm) euhedral
233 grains arranged as: (i) overgrowths on plagioclase clasts (Fig. 4b); (ii) framboidal rims around
234 opaque grains; (iii) trails along thin healed microcracks confined within the pseudotachylyte
235 vein; and (iv) trails along the pseudotachylyte boundary.

236 5.3 Ductile shear zones

237 Mylonitized pseudotachylytes consist of a fine-grained (average grain size: 5-30 μm)
238 mixture of plagioclase, amphibole and clinopyroxene, with less abundant garnet, biotite, quartz,
239 K-feldspar and ilmenite (Fig. 5b). Amphibole is typically found as isolated grains and small
240 clusters ($< 50 \mu\text{m}$ in length) at triple junctions between plagioclase grains (Fig. 5b), as well as
241 along C' shear bands. The amount of amphibole in the phase mixture, estimated with image
242 analysis on SEM backscatter electron images, is in the range of 5-10 vol%. Garnet is present as
243 isolated grains (typically $< 20 \mu\text{m}$ in size) dispersed in the phase mixture, and as polycrystalline
244 clusters where individual grains are up to 200 μm in size (Fig. S1). Most of the garnets contain
245 inclusions of plagioclase, clinopyroxene and amphibole. Isolated grains of garnet are typically
246 inclusion-free.

247 Mylonitic pseudotachylytes wrap around lenses deriving from the grain size reduction of
248 original plagioclase and clinopyroxene of the anorthosites. In these lenses, plagioclase is variably
249 fractured and recrystallized to fine-grained (average grain size of 25 μm) mono-mineralic
250 aggregates elongated parallel to the shear zone foliation. The relict porphyroclasts show
251 undulatory extinction and bent twins (Figs. 5a, 5c). Clinopyroxene porphyroclasts are partially to
252 completely replaced by fine-grained ($< 30 \mu\text{m}$) clinopyroxene, amphibole and quartz neoblasts
253 along intracrystalline bands and along the boundaries with plagioclase (Fig. 5d). The
254 clinopyroxene-derived aggregates are also stretched parallel to the mylonitic foliation (Fig. 5c).

255 Plagioclase and clinopyroxene in the damage zone flanking the ductile shear zones are
256 pervasively fractured and show undulatory extinction. Fractures are filled with new small
257 polygonal grains of plagioclase (within plagioclase) and clinopyroxene + amphibole \pm quartz
258 (within clinopyroxene). The new grains are 5-30 μm in size and form up to 200 μm wide

259 intracrystalline bands (Fig. 5e). These microstructures are identical to those of the damage zone
260 flanking pristine pseudotachylyte veins.

261

262 **6 EBSD analysis**

263 The crystallographic preferred orientation (CPO) of plagioclase determined by EBSD in
264 the polyphase matrix of the mylonitic pseudotachylyte (Fig. 6a) does not show an obvious
265 alignment of lattice planes and axes with the foliation and with the stretching lineation,
266 respectively (Fig. 6b). On the contrary, amphibole has a CPO of the (100) planes distributed
267 along a discontinuous girdle subparallel to the YZ plane of the finite strain ellipsoid, with a
268 maximum near the pole to the foliation (Fig. 6b). The [001] axis forms a maximum on the
269 foliation plane at a low angle to the stretching lineation.

270 The misorientation axis distribution (MAD) of random pairs of both plagioclase and
271 amphibole is remarkably close to the theoretical curve for a random distribution (Fig. 6c).
272 Correlated pairs of both plagioclase and amphibole show that misorientations $< 50^\circ$ occur with a
273 higher frequency than for a random distribution (Fig. 6c).

274 The long axes of amphibole grains are preferentially oriented at an angle of $5\text{-}20^\circ$
275 measured counter-clockwise from the trace of the ultramylonite foliation (Fig. 6d). For the
276 sinistral sense of shear of the ultramylonite in Fig. 7a, the preferred orientation of the long axes
277 of amphibole corresponds to a C' -type shear band orientation.

278 Recrystallized plagioclase in an elongated mono-mineralic aggregate embedded in
279 mylonitized pseudotachylytes (Fig. 7a) shows a grain size range from $4\ \mu\text{m}$ to about $120\ \mu\text{m}$
280 (Fig. 7b). Grains larger than $30\ \mu\text{m}$ are preferentially oriented with the (001) planes subparallel
281 to the foliation and with the [110] direction subparallel to the stretching lineation. Grains smaller
282 than $30\ \mu\text{m}$ show a remarkable dispersion of the crystallographic planes and axes, but many
283 small grains overlap in orientation with the larger grains (Fig. 7c). Misorientations $< 70^\circ$ occur
284 with a higher frequency than for a theoretical random distribution for both correlated and
285 uncorrelated pairs (Fig. 7d). The misorientation axis for low misorientations ($3\text{-}10^\circ$) is
286 preferentially parallel to the [-100] axis (Fig. 7d).

287

288 **7 Mineral chemistry and P-T-H₂O conditions of deformation**

289 The major element compositions of representative minerals in pristine, recrystallized, and
290 mylonitized pseudotachylytes are reported in Tables S1-S3 in the supplementary material.
291 Plagioclase of the host rock has ~ 77 mol% of anorthite component (bytownite). Plagioclase
292 microlites and recrystallized plagioclase (both in recrystallized and in mylonitized
293 pseudotachylytes) have lower anorthite contents, mostly in the range between 46 and 52 mol%

294 (andesine), but with a few recrystallized grains containing up to 58 mol% of anorthite (Table
295 S1).

296 Clinopyroxene is diopside, with similar Al contents of 0.15-0.20 atoms per formula unit
297 (a.p.f.u.) in both the host rock and in the recrystallized grains in the mylonite (Table S2). X_{Mg}
298 values ($Mg/(Mg+Fe^{2+})$) range from 0.76 to 0.84, with the highest values found in clinopyroxene
299 inclusions in garnet (table S2). Orthopyroxenes occur in the host rock, as survivor clasts and as
300 microlites in the pseudotachylytes, and as porphyroclasts in the mylonites. They have similar
301 compositions with X_{Mg} of ~0.57–0.68 and Al contents from 0.10 to 0.14 a.p.f.u., and up to 0.20
302 a.p.f.u. in the microlites (Table S2).

303 Amphiboles are pargasites and have Ca ~1.90, $(Na+K)_A$ from 0.71 to 0.85 and Si ranging
304 from 5.98 to 6.35 a.p.f.u. (Table S3). The values of X_{Mg} range from 0.55 to 0.63, up to 0.68 for
305 those included in garnet in the mylonites. Amphiboles do not show any variation of $(Na+K)_A$ and
306 $Al^{(VI)}$ indicative of edenite substitution. The almost homogeneous composition of amphiboles in
307 the various microstructural sites of pristine, recrystallized and mylonitized pseudotachylytes
308 suggests similar equilibrium conditions during the formation, recrystallization, and shearing of
309 pseudotachylytes.

310 Table S3 also shows the compositions of garnets overgrowing plagioclase survivor clasts
311 in pseudotachylytes, in coronas around corundum in the host rock, and in the mylonitized
312 pseudotachylytes. All garnets have the same almandine content (~50 mol%), except those around
313 corundum (59 mol%). They instead record variable grossular and pyrope contents with values
314 ranging from 28 to 39 mol% of grossular, corresponding to a variation from 24 to 11 mol% of
315 pyrope (Table S3).

316 Tables S1 and S3 also report representative compositions of plagioclase and amphibole
317 pairs used for conventional geothermobarometry calculations. Temperature was estimated with
318 the thermometer of *Holland and Blundy* [1994], which considers the exchange reaction edenite +
319 albite = richterite + anorthite and was calibrated in the P, T range of 0.1-1.5 GPa and 400-1000
320 °C. Pressure was estimated with the barometer of *Anderson and Smith* [1995], which is based on
321 the increase of Al content in hornblende with increasing P and was calibrated in the T range
322 between 675 and 760 °C. The resulting P, T conditions of sin-kinematic recrystallization in the
323 mylonitized pseudotachylytes are estimated at 700-750°C, 0.7-0.9 GPa.

324 These results are in agreement with the P-T stability field modelled in Figure 8a for H₂O-
325 saturated conditions, which were chosen because we interpret the syn-kinematic nucleation of
326 amphibole at triple junctions as the evidence of the presence of H₂O at grain boundaries. The
327 mineral assemblage consisting of biotite – amphibole – plagioclase – clinopyroxene – quartz –
328 ilmenite found along the recrystallized and mylonitized pseudotachylytes (Figs. 4e, 5b) is stable
329 in the red P-T area of Fig. 8a, at around 650 °C and 0.7 GPa. Syn- to post-kinematic garnet
330 extends the stability field up to 750 °C and 0.85 GPa. These maximum temperatures correspond
331 to the stability limit of amphibole for the studied composition. Below 750 °C, the stability of
332 amphibole is also governed by the amount of water assisting the deformation, as portrayed in the
333 T-X(H₂O wt%) section of Figure 8b, calculated at 0.7 GPa. At least 0.25 wt% of H₂O is required
334 to crystallise amphibole and 0.5 wt% to saturate the system in the temperature range from 575 to
335 750 °C. For higher H₂O contents, the modal amount of amphibole is only dependent on the

336 equilibration temperature, as shown by the grey lines that indicate the locus of points of equal
337 amount (vol%) of amphibole for the various modelled mineral assemblages.

338 To evaluate if fluid infiltration was required in order to stabilize the mineral assemblage of
339 the mylonitized pseudotachylytes, we modelled the variation of modal abundances of mineral
340 phases in the pristine anorthosites as a function of the H₂O content. The calculations have been
341 performed at the estimated P, T conditions of deformation (700° C and 0.7 GPa) and the results
342 are shown in Fig. 9. The diagram shows that < 1 vol% of biotite requires approximately 0.04
343 wt% of H₂O, and that such a low amount of H₂O would not result in the presence of free
344 intergranular H₂O in the anorthosite protolith at the estimated P, T conditions of deformation.
345 Thus, the amount of fluid present in the anorthosite is about six to ten times less than that
346 required to stabilize the mineral assemblage of the mylonitized pseudotachylytes.

347

348 **8 Intracrystalline water content in plagioclase**

349 Brittle fracturing can introduce fluids into grain interiors even at high P-T conditions and
350 this typically favours hydrolytic weakening [e.g. *Stünitz et al.*, 2017]. The measurement of the
351 intracrystalline water content in plagioclase was carried out to test if fracturing and
352 pseudotachylyte formation was accompanied by fluid infiltration in the interior of the grains. We
353 measured the intracrystalline H content in plagioclase grains along 1.5 cm long transects from
354 the undeformed anorthosite to the pristine pseudotachylyte, where the measurements were
355 restricted to survivor clasts.

356 Most of the measurements indicate intracrystalline water contents < 40 ppm both in the
357 anorthosites and in survivor clasts in the pseudotachylyte (Fig. 10). There is no progressive
358 increase of water concentration in plagioclase towards the pseudotachylyte vein, and many clasts
359 inside the veins and at the pseudotachylyte vein boundary have water concentration < 20 ppm.
360 There are few grains with intracrystalline water content > 70 ppm both in the host rock and in the
361 pseudotachylyte. SEM inspection of the probe pits indicates that these higher water
362 concentrations occur in close proximity to lamellae of K-feldspar, presumably resulting from
363 exsolution of original ternary feldspars (Fig. S2).

364

365 **9 Discussion**

366 **9.1 Cyclic interplay between brittle and viscous deformation at lower crustal conditions**

367 Field observations in Nusfjord indicate that ductile shear zones exploited a set of
368 pseudotachylyte-bearing faults. This is consistent with the general observation that nucleation of
369 ductile shear zones requires the presence of a planar compositional or structural precursor [e.g.
370 *Pennacchioni and Mancktelow*, 2007; *Pennacchioni and Zucchi*, 2013]. In Nusfjord,
371 pseudotachylytes provided both the structural anisotropy of long, sub-parallel fault planes and
372 the compositional precursor of weak, fine-grained pseudotachylyte fault veins decorating the slip
373 surfaces. The alternation of thinner and thicker shear zones apparently reflects the original
374 hierarchical organization of the precursor pseudotachylyte-bearing faults [*Pennacchioni*, 2005;

375 *Di Toro and Pennacchioni, 2005*]. Thin shear zones derive from single mm-cm thick
376 pseudotachylyte fault veins that only locally include pockets of pseudotachylyte breccias. Major
377 shear zones overprinted complex bands of more pervasive pseudotachylyte veining and
378 brecciation.

379 The mutual overprinting relationships between pristine and mylonitized pseudotachylytes
380 (Figs. 3d-f) are interpreted to result from the cyclic interplay between brittle and viscous
381 deformation. Thus, the P-T conditions of deformation estimated for the mylonitized
382 pseudotachylytes are considered representative also of the conditions of formation of
383 pseudotachylytes. This is supported by the similarity in microstructure of the damage zone
384 flanking both pristine and mylonitized pseudotachylytes showing neoblasts of clinopyroxene and
385 amphibole, and the lack of low-grade minerals (Figs. 4a, 5e).

386 The P-T conditions of deformation are constrained at 650-750°C and 0.7-0.8 GPa
387 (depending on the stability of garnet), using both amphibole-plagioclase conventional
388 geothermobarometry and thermodynamic forward modelling. Assuming an average crustal rock
389 density of 2700-3000 kg/m³, these conditions correspond to a depth range of 24-30 Km and
390 indicate a lower crustal setting for the brittle-viscous deformation cycles.

391 9.2 Origin of pseudotachylytes at lower crustal conditions

392 Generation of pseudotachylytes at depths below the classic (10-15 km depth) brittle-ductile
393 transition in the continental crust has been interpreted to result from different processes: (1)
394 frictional melting during seismic slip, either due to the downward propagation of seismic
395 ruptures nucleated in the upper brittle crust [e.g. *Ellis and Stöckhert, 2004; Lin et al., 2005;*
396 *Moecher and Steltenpohl, 2011*] or to the nucleation of lower crustal earthquakes [*Jackson et al.,*
397 *2004; Austrheim, 2013; Austrheim et al., 2017*], (2) plastic instabilities [*Hobbs et al., 1986;*
398 *White, 1996, 2012*], (3) self-localizing thermal runaway [e.g. *John et al., 2009; Andersen et al.,*
399 *2014*], (4) in-situ amorphization during high stress deformation at slow displacement rates [*Pec*
400 *et al., 2012*]. Plastic instabilities and self-localizing thermal runaway require a ductile precursor
401 of a pseudotachylyte-bearing fault. In these models, seismic slip localization and melting result
402 from the feedback between shear heating and thermal softening within a ductile shear zone.
403 However, pseudotachylytes by themselves cannot be taken as unequivocal field evidence for
404 thermal runaway processes, unless they contain clasts of mylonites that could suggest the
405 presence of a ductile precursor. The plastic instability and thermal runaway models do not seem
406 appropriate for the Nufjord case, because a large amount of pseudotachylytes clearly predates
407 the mylonitic deformation (e.g. Figs. 2f, 3f). The pristine pseudotachylytes contain in most of the
408 cases abundant survivor clasts and lithic clasts of the host shattered anorthosite, but not of
409 mylonites (Fig. 4a). In-situ amorphization is difficult to reconcile with the extensive occurrence
410 of microlites (Figs. 4b, 4c), with the formation of chilled margins (Figs. 2g, 4d) and with the
411 large volumes of pseudotachylytes. Thus, we interpret the Nufjord pseudotachylytes as the
412 result of frictional melting on fault planes during seismic slip in the lower crust.

413 Generation of pseudotachylytes by brittle deformation at lower crustal conditions has been
414 explained by local high fluid pressure promoting fracturing [e.g. *Lund and Austrheim, 2003;*
415 *Steltenpohl et al., 2006*]. However, the Nufjord pseudotachylytes are not spatially associated
416 with any vein system, and there is no evidence of hydration of the pseudotachylytes with respect

417 to their host rock (Table 1, Fig. 8, and see section 9.3). This is consistent with the development
418 of the Nufjord pseudotachylytes as the product of co-seismic brittle deformation of effectively
419 dry rocks devoid of an intergranular pore fluid. The absence of free aqueous fluid in
420 interconnected intergranular pore space inhibits viscous deformation and results in a high
421 strength of lower crustal rocks [Fitz Gerald *et al.*, 2006; Menegon *et al.*, 2011]. In the absence of
422 elevated pore fluid pressure, high differential stresses are necessary to fracture dry lower crustal
423 rocks. Thus, we interpret the widespread occurrence of pseudotachylytes generated by co-
424 seismic brittle deformation in Nufjord as the evidence of a strong, seismogenic lower crust with
425 a rheology equivalent to that of anhydrous granulite facies material [Maggi *et al.*, 2000; Jackson
426 *et al.*, 2004]. Crustal strength calculations suggest that deep seismicity at depths greater than 30
427 km in the Rukwa-Malawi rift zone is best explained by the rheology of a mafic granulitic lower
428 crust [Fagereng, 2013]. Likewise, an analysis of the aftershocks following the 2001 M_w 7.6 Bhuj
429 earthquake revealed that they were indeed located mostly in the lower crust at a depth of about
430 26 km, and this was explained invoking a mafic granulite rheology of the deeper crustal levels
431 [Bodin and Horton, 2004]. Thus, we conclude that the lower crustal Nufjord earthquakes are
432 consistent with the nucleation of seismic ruptures at depth > 25 km, or with a downward
433 propagation of seismic ruptures from the deep brittle-ductile transition in dry anorthosites, and
434 not from the typical shallower transition in the upper continental crust.

435 9.3 Deformation mechanisms in the mylonitized pseudotachylytes and the role of fluids

436 Monomineralic plagioclase layers embedded in mylonitized pseudotachylytes show a
437 moderate CPO consistent with dislocation creep by $\frac{1}{2}$ (001) [110] slip system (Fig. 7c) [e.g.
438 Kruse *et al.*, 2001]. Deformation by dislocation creep in plagioclase monomineralic layers is
439 suggested by the overlap in crystallographic orientation between porphyroclasts and
440 recrystallized grains (Fig. 7c), by the excess of low angle boundaries ($< 20^\circ$, Fig. 7d), and by the
441 clusters of misorientation axes parallel to crystallographic axes for low misorientations (3-10°,
442 Fig. 7d). Many recrystallized grains show a significant dispersion of crystallographic orientations
443 from the one of the host porphyroclasts (Fig. 7c). This is consistent with a contribution of grain
444 boundary sliding after the recrystallization of plagioclase. However, monomineralic layers are
445 typically wrapped by recrystallized pseudotachylytes, and this suggests that the polyphase
446 mixture deriving from recrystallization of pseudotachylytes is rheologically weaker than
447 monomineralic plagioclase domains deforming by dislocation creep.

448 The fine grain size of the phase mixture (5-30 μm) and the high degree of phase mixing
449 suggest that mylonitized pseudotachylyte deformed by diffusion creep and grain boundary
450 sliding. The plagioclase weak CPO is not consistent with dislocation creep along common slip
451 systems (Fig. 6b). Amphibole has a clear CPO of the (100) planes and [001] direction, but the
452 same CPO has been observed in amphibolites deformed experimentally in the diffusion creep
453 regime [Getsinger and Hirth, 2014]. We interpret the amphibole CPO as the result of oriented
454 grain growth and rigid body rotation during diffusion creep, as concluded for other natural
455 samples deformed at lower crustal condition [e.g. Berger and Stünitz, 1996; Getsinger *et al.*,
456 2013; Menegon *et al.*, 2015; Okudaira *et al.*, 2015]. Thus, we conclude that the brittle grain size
457 reduction due to pseudotachylyte formation, the fine grain size, and the polymineralic
458 composition (inhibiting grain growth) of pristine- and recrystallized pseudotachylytes (Figs. 4c,
459 4e) predisposed pseudotachylytes veins and the surrounding damage zone to subsequent strain
460 localization by grain size sensitive creep. A similar origin of ultramylonites deforming by grain

461 size sensitive creep was proposed by *White* [1996], who identified the recrystallization of
462 precursor pseudotachylyte veins as the key grain-size reduction mechanism necessary to produce
463 ultramylonites.

464 Strain localization triggered by lower crustal seismicity has typically been associated with
465 fluid infiltration and associated reaction weakening in the fractured and hydrated domains
466 [*Austrheim*, 2013, and references therein]. Our results indicate that the formation of mylonitized
467 pseudotachylytes required a six- to ten-fold increase in H₂O content with respect to the
468 anorthosite protolith (Figs. 8b and 9).

469 In the mylonitized pseudotachylytes, amphibole crystallisation in equilibrium with biotite,
470 plagioclase, clinopyroxene, quartz, ilmenite, ± garnet occurred for H₂O contents above 0.25 wt%
471 at P = 0.7 GPa and T up to 750 °C. At H₂O-undersaturated conditions, amphibole content is
472 strongly dependent on the water content of the system. It is worth noting that at T = 700 °C, 0.45
473 wt% of water enabling the crystallisation of 8 vol% of amphibole (which represents the average
474 amount of amphibole in the mylonitized pseudotachylytes as estimated with image analysis) is
475 coincident with the limit of H₂O saturation. At higher temperatures, the limit is reached at even
476 lower water contents. This means that very low H₂O is needed to stabilise the observed syn-
477 deformation paragenesis and, more important, that 0.25-0.45 wt% of H₂O is sufficient to saturate
478 the entire rock system at the P, T conditions of deformation. Thus, although H₂O infiltration is
479 required to stabilize the mineral assemblage in the mylonitized pseudotachylytes, the amount of
480 H₂O infiltration that resulted in rheological weakening is minimal. *Okudaira et al.* [2015]
481 reported similar low values of water contents during diffusion creep deformation in gabbro
482 ultramylonites following a precursor fracturing stage at granulite facies conditions.

483 This is consistent with experimental results of *Milke et al.* [2012], which showed that very
484 little amounts of water present as a pore fluid are effective in facilitating mineral reactions if rock
485 deformation maintains a sufficient porosity. Moreover, relatively low water contents (~0.07
486 wt%) are commonly referred to as wet conditions during experimental rock deformation of
487 feldspar rocks in the viscous regime, and are sufficient to activate diffusion creep [*Rybacki and*
488 *Dresen*, 2000; *Dimanov et al.*, 2005]. We argue that brittle grain size reduction followed by
489 diffusion-accommodated viscous grain boundary sliding enhanced the porosity of the
490 mylonitized pseudotachylytes, and resulted in the distribution of the infiltrated fluids along the
491 grain boundaries. Diffusion-accommodated viscous grain boundary sliding can maintain a
492 dynamic porosity in ultramylonites [*Fusseis et al.*, 2009], which can evolve into the development
493 of creep cavitation bands decorated by new grains that precipitate from the grain boundary fluid
494 [*Menegon et al.*, 2015]. In the Nufjord mylonitized pseudotachylytes, the possible signature of
495 creep cavitation bands is the preferred orientation of amphibole long axis parallel to a C'-type
496 shear band orientation (Fig. 6d), similar to the microstructures described in *Menegon et al.*
497 [2015]. The water redistributed along the grain boundaries made the system wet at the shear zone
498 scale, and facilitated heterogeneous phase nucleation, as demonstrated by the syn-kinematic
499 growth of amphiboles at triple junctions and dilatant sites (Figs. 5b, 6a). This stabilized strain
500 localization in the polymineralic, fine-grained mylonitized pseudotachylytes. Therefore, it is the

501 water at grain boundaries that has a major rheological effect on lower crustal shear zones, by
 502 facilitating diffusion creep deformation and phase nucleation.

503 Measurements of intracrystalline water content of plagioclase do not show a systematic
 504 increase from the host rock to the pseudotachylytes. This lends further support to our
 505 interpretation that the infiltrated aqueous fluid was redistributed along the grain boundaries
 506 rather than penetrating the interior of nominally anhydrous minerals via (micro)cracking. The
 507 low intracrystalline water contents in plagioclase (Fig. 10) were apparently sufficient for limited
 508 dislocation creep to occur (Fig. 7). However, this did not result in the nucleation of ductile shear
 509 zones in intact anorthosites, which instead required a precursor stage of co-seismic brittle
 510 deformation with associated fluid redistribution at the grain boundaries.

511

512

513

514 9.4 Rheology of mylonitized pseudotachylytes

515 To evaluate the extent of weakening resulting from diffusion creep in the mylonitized
 516 pseudotachylytes, we compared the rheology of plagioclase deforming by dislocation creep with
 517 the rheology of a plagioclase + clinopyroxene aggregate deforming by diffusion creep. The
 518 general flow law for dislocation creep is:

$$519 \quad \dot{\epsilon} = A\sigma^n \exp(-Q/RT) \quad (1)$$

520 where $\dot{\epsilon}$ is the strain rate, A an empirical constant, σ the differential stress, n the stress
 521 exponent, Q the activation energy, R the gas constant, and T the temperature. We used the flow
 522 laws for dislocation creep of wet and dry anorthite of *Rybacki and Dresen* [2000]. Dry conditions
 523 were chosen to predict the rheology of the anorthosite host rock deforming by dislocation creep
 524 and devoid of an intergranular pore fluid phase, whereas wet conditions were chosen to simulate
 525 the rheology of monomineralic plagioclase layers embedded in mylonitized pseudotachylytes,
 526 which deformed in the presence of H₂O (Figs. 4b, 8).

527 Deformation accommodated by diffusion creep is described by the flow law:

$$528 \quad \dot{\epsilon} = A\sigma^n d^m \exp(-Q/RT) \quad (2)$$

529 where d is the grain size and m the grain size exponent. We use the flow law for diffusion
 530 creep of anorthite + diopside aggregates of *Dimanov and Dresen* [2005]. This flow law was
 531 chosen because (1) it approximates the composition of the mylonitized pseudotachylytes, and (2)
 532 it was derived for grain boundary diffusion-controlled creep, which is consistent with the
 533 deformation mechanisms identified in the mylonitized pseudotachylytes. We considered the flow

534 law for wet diffusion creep of an aggregate consisting of 75% anorthite (average grain size of ca.
535 3.5 μm) and 25% diopside (grain size < 35 μm) [*Dimanov and Dresen, 2005*].

536 Rheological calculations show that at $T = 700^\circ\text{C}$ the differential stress required to sustain
537 geologically realistic strain rates of 10^{-10} – 10^{-13} s^{-1} in the anorthite + diopside aggregate
538 deforming by diffusion creep would be 1–3 orders of magnitude smaller than the stress required
539 to deform plagioclase by wet dislocation creep, and up to 5 orders of magnitude smaller than
540 then stress required to deform plagioclase by dry dislocation creep (Fig. 11). Furthermore, in the
541 differential stress range between 10 and 0.1 MPa, deformation by diffusion creep in the anorthite
542 – diopside aggregate would occur 3-7 orders of magnitude faster than deformation by dislocation
543 creep in wet plagioclase. Thus, the weakening from activation of diffusion creep in the fine-
544 grained polyphase mixture derived from recrystallization of pseudotachylytes is predicted to be
545 significant. Similar effects of grain size reduction, marked weakening, and enhanced creep rates
546 through activation of diffusion creep in recrystallized pseudotachylytes have been proposed by
547 *White [1996]*.

548

549 **10 Conclusions**

550 In the dry anorthosites from Nusfjord (Lofoten, northern Norway) different generations of
551 pseudotachylytes and mylonites formed at lower crustal conditions of 650-750 $^\circ\text{C}$ and 0.7-0.9
552 GPa, as estimated with conventional geothermobarometry and with thermodynamic forward
553 modelling. Mylonites exploited pristine pseudotachylytes veins, which provided both a structural
554 and a compositional precursor for the nucleation of ductile shear zones.

555 Pristine and mylonitized pseudotachylytes show mutual overprinting relationships that are
556 interpreted as the result of the cyclic interplay between brittle and viscous deformation at lower
557 crustal condition. Pseudotachylytes formed by brittle co-seismic deformation in relatively dry
558 and strong anorthosites, as the result of seismic ruptures nucleated in the deep crust or of the
559 downward propagation of seismic ruptures from the deep brittle-ductile transition in dry
560 anorthosites.

561 Strain localization in mylonitized pseudotachylytes was facilitated by the grain size
562 reduction due to the formation and subsequent recrystallization of pseudotachylytes to a fine-
563 grained (< 30 μm) polyphase aggregate. Microstructural observations and EBSD analysis are
564 consistent with diffusion creep as the dominant deformation mechanism in the mylonitized
565 pseudotachylytes. Thermodynamic modelling combined with microstructural observations
566 indicate that the fluid infiltration necessary to stabilize the mineral assemblage of the
567 mylonitized pseudotachylytes was minimal, on the order of 0.20-0.40 wt% H_2O . The infiltrated
568 aqueous fluid was redistributed along the grain boundaries in the fractured domains, and made
569 the system wet and weak at the shear zone scale. Diffusion-accommodated viscous grain
570 boundary sliding maintained a dynamic porosity in the mylonitized pseudotachylytes, and

571 produced creep cavitation and precipitation of material from grain boundary fluids collected in
572 creep cavities and in other dilatant sites.

573 Extrapolation of experimentally derived flow laws to geologically realistic strain rates at a
574 temperature of 700°C indicates up to 3 orders of magnitude weakening in the mylonitized
575 pseudotachylytes deforming by diffusion creep with respect to plagioclase deforming by
576 dislocation creep in the anorthosite host rock. This highlights that pseudotachylytes caused by
577 brittle faulting can be precursors of viscous, weak shear zones in the dry lower crust, indicating
578 lower crustal earthquakes as agents of rheological change from strong, brittle crust, to strong
579 crust with embedded fine grained, weak viscous shear zones.

580

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593

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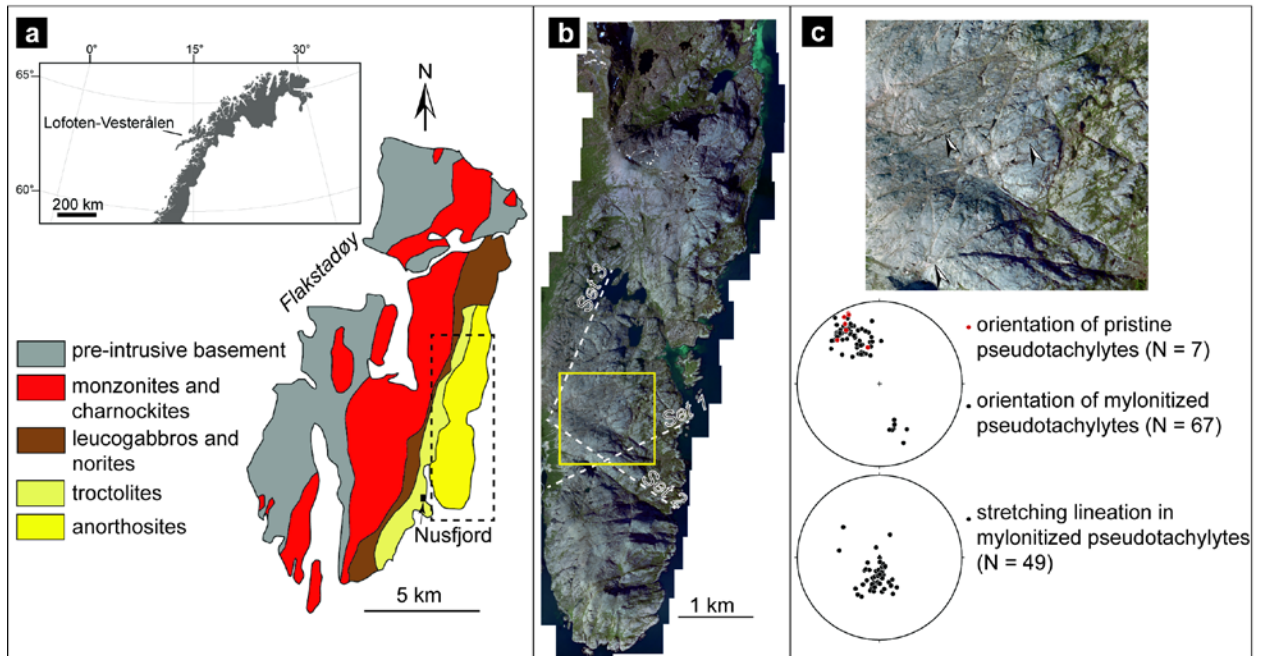
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Lithology	Anorthosite host rock	Pristine pseudotachylyte	Mylonitized pseudotachylyte
Sample	N13-10B	N22	N13-10D
Sample location	0432042 E 7549520 N	0432009 E 7550172 N	0432042 E 7549520 N
SiO ₂	53.83	51.37	53.54
TiO ₂	0.14	0.32	0.30
Al ₂ O ₃	27.22	25.67	25.11
Fe ₂ O _{3tot}	1.37	3.98	3.80
MnO	0.02	0.04	0.04
MgO	0.41	1.87	1.86
CaO	10.75	10.85	9.89
Na ₂ O	4.75	4.25	4.24
K ₂ O	0.61	0.76	0.92
P ₂ O ₅	0.04	0.04	0.04
Tot	99.14	99.15	99.74
FeO	<i>n.d.</i>	<i>n.d.</i>	3.32
L.O.I.	0.32	0.30	0.32

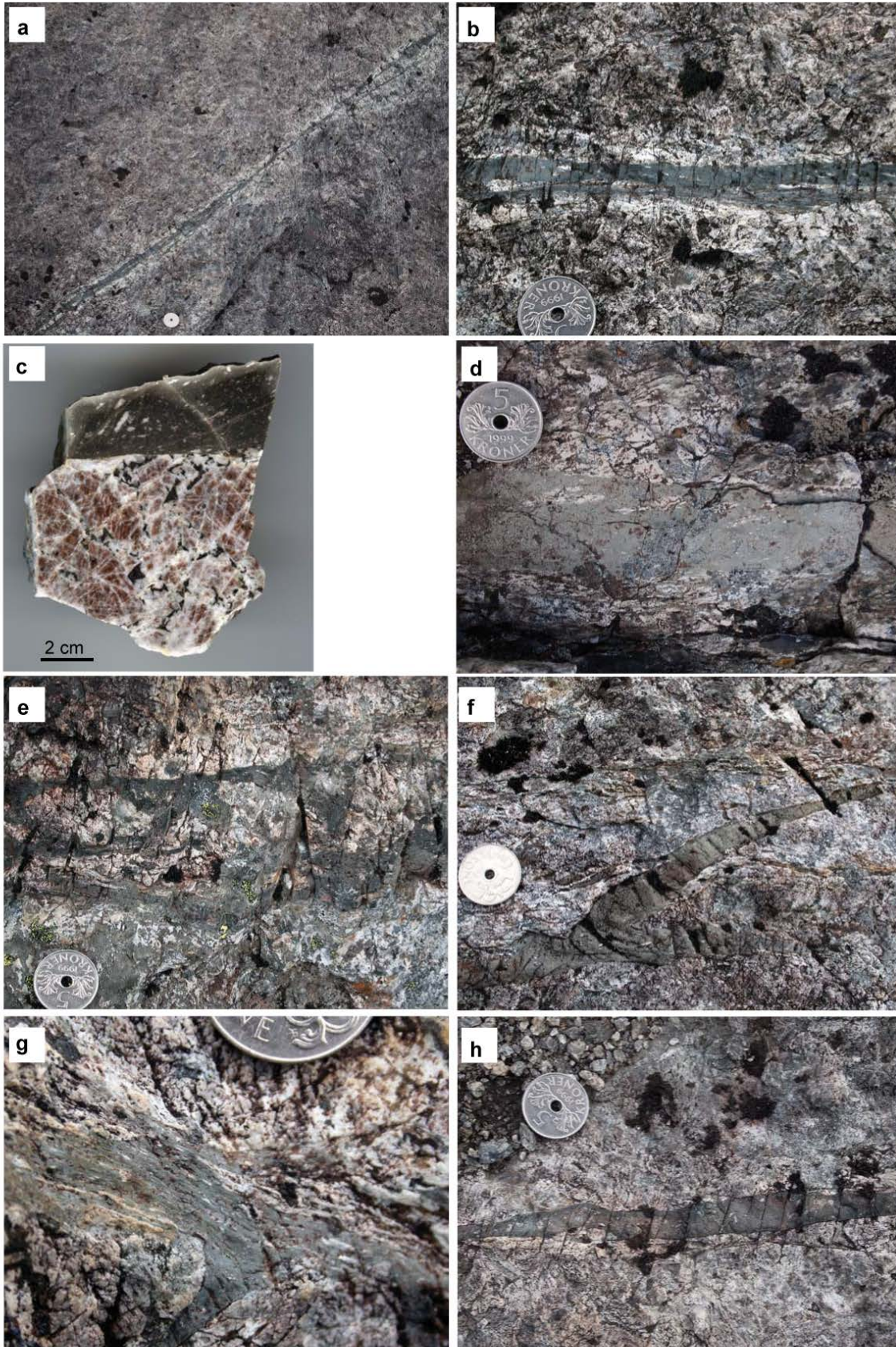
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Table 1. Representative whole rock XRF analysis of anorthosites, of pristine-, and of mylonitized pseudotachylytes. GPS coordinates are relative to WGS84, zone 33W.

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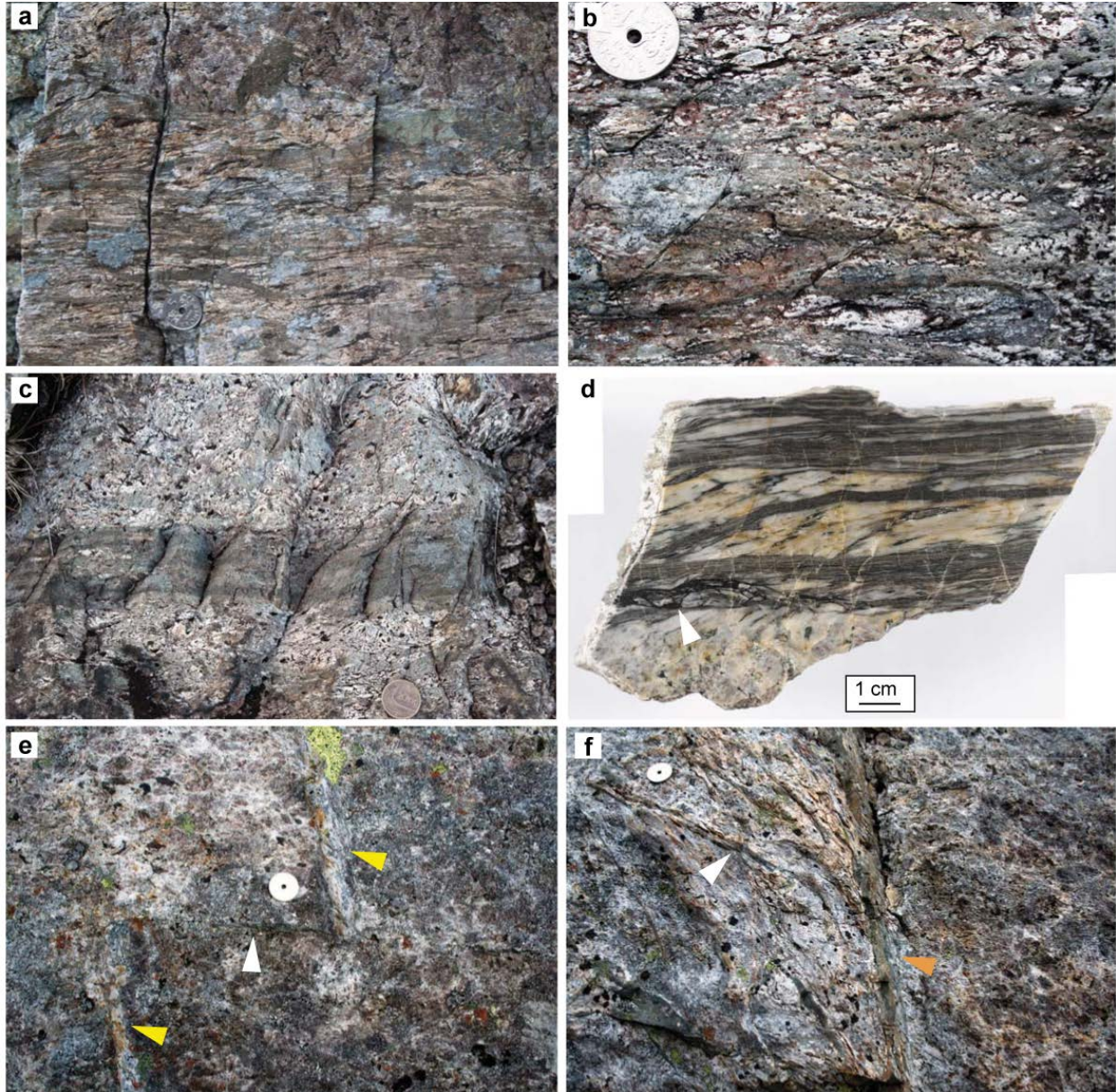
768 **Figures**

769
 770 **Figure 1.** Geological setting. (a) Simplified geological map of Flakstadøy, Lofoten. Modified after Steltenpohl et al.
 771 (2006). The dashed rectangle encompasses the area shown in (b). (b) Mosaic of aerial photographs of the Nusfjord
 772 ridge. The three main sets of lineaments, defined in the field by ductile shear zones, are marked with white dashed
 773 lines. The yellow rectangle encompasses the area shown in (c). (c) Field area investigated in this study. Arrowheads
 774 indicate examples of first order shear zones. Pole figures (lower hemisphere, stereographic projection) showing the
 775 orientation of set 1 pristine- and mylonitized pseudotachylytes (poles to planes), and of the stretching lineation
 776 measured in mylonitized pseudotachylytes. N = number of measurements.
 777



779 **Figure 2.** Field structures of 2nd order shear zones and of pseudotachylyte. All GPS coordinates are relative to
780 WGS84, zone 33W. (a) Thin (<1 cm) ductile shear zone surrounded by a bleached halo within the darker coarse
781 grained anorthosite. GPS coordinates: 0431864 East, 7549944 North. (b) Detail of the shear zone in (a) showing the
782 fine grained, dark green, sharply bounded mylonitic core surrounded by a partially sheared bleached halo. (c)
783 Polished sample of a weakly deformed dark-green shear zone in sharp contact with the host anorthosite. The coarse
784 (cm-sized) purple-coloured magmatic crystals of anorthosite show bleaching along a pervasive network of
785 microfractures. GPS coordinates: 0431685 East, 7549992 North. (d) Relatively weakly deformed pseudotachylyte
786 with lithic clasts of the bleached anorthosite deformed to elongated domains stretched oblique to the
787 pseudotachylyte (shear zone) boundary. Same locality as (c). (e) Pseudotachylyte breccia with bleached lithic clasts
788 and host rock; (f) Pristine pseudotachylyte injection vein protruding from a fault vein; GPS coordinates: 0431808
789 East, 7550038 North. (g) Sheared pseudotachylyte fault vein preserving an undeformed injection vein in the host
790 anorthosite. GPS coordinates: 0432073 East, 7550013 North. (h) Pristine pseudotachylyte preserving chilled
791 margins and small equant clasts of bleached plagioclase. Same locality as (f). Coin for scale (2.5 cm in diameter) in
792 all the field photographs.

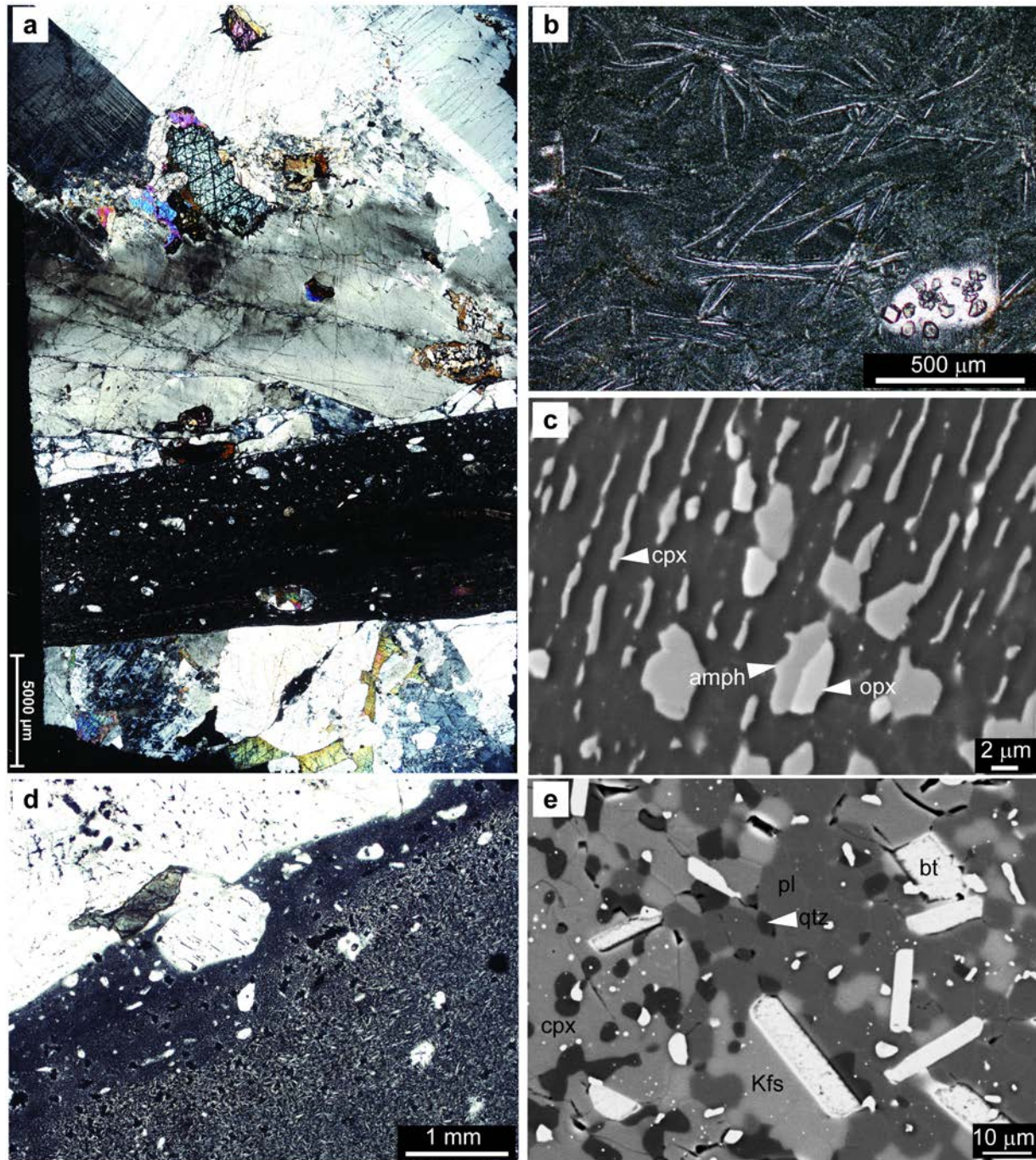
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795 **Figure 3.** Field structures of major shear zones and associated pseudotachyrites. All GPS coordinates are relative to
 796 WGS84, zone 33W. (a) Mylonitic foliation with alternating thin bands of green and whitish colour. Note that some
 797 green layers are still locally discordant to the foliation and that pristine pseudotachylyte injection veins are present in
 798 the host anorthosite. GPS coordinates: 0432042 East, 7549520 North. (b) Large domain of preserved
 799 pseudotachylyte breccia with lithic clast of the bleached host anorthosite free of mylonitic foliation. GPS
 800 coordinates: 0432009 East, 7550172 North. (c) Similar as (b) but also showing a thick central pseudotachylyte layer
 801 localizing the deformation. Same locality as (b). (d) Polished sample of a domainal foliation delineated by green and
 802 whitish layers, derived from pseudotachylyte and bleached anorthosite. In the lower part of the sample, a
 803 pseudotachylyte exploited the shear zone boundary and the mylonitic foliation was brecciated within a side wall
 804 ripout. Same locality as (a). (e-f) A pristine pseudotachylytes (indicated with a white arrowhead) sharply displaces a
 805 whitish localized shear zone (yellow arrowheads) (e) and is in turn dragged into the mylonitic foliation along a dm-

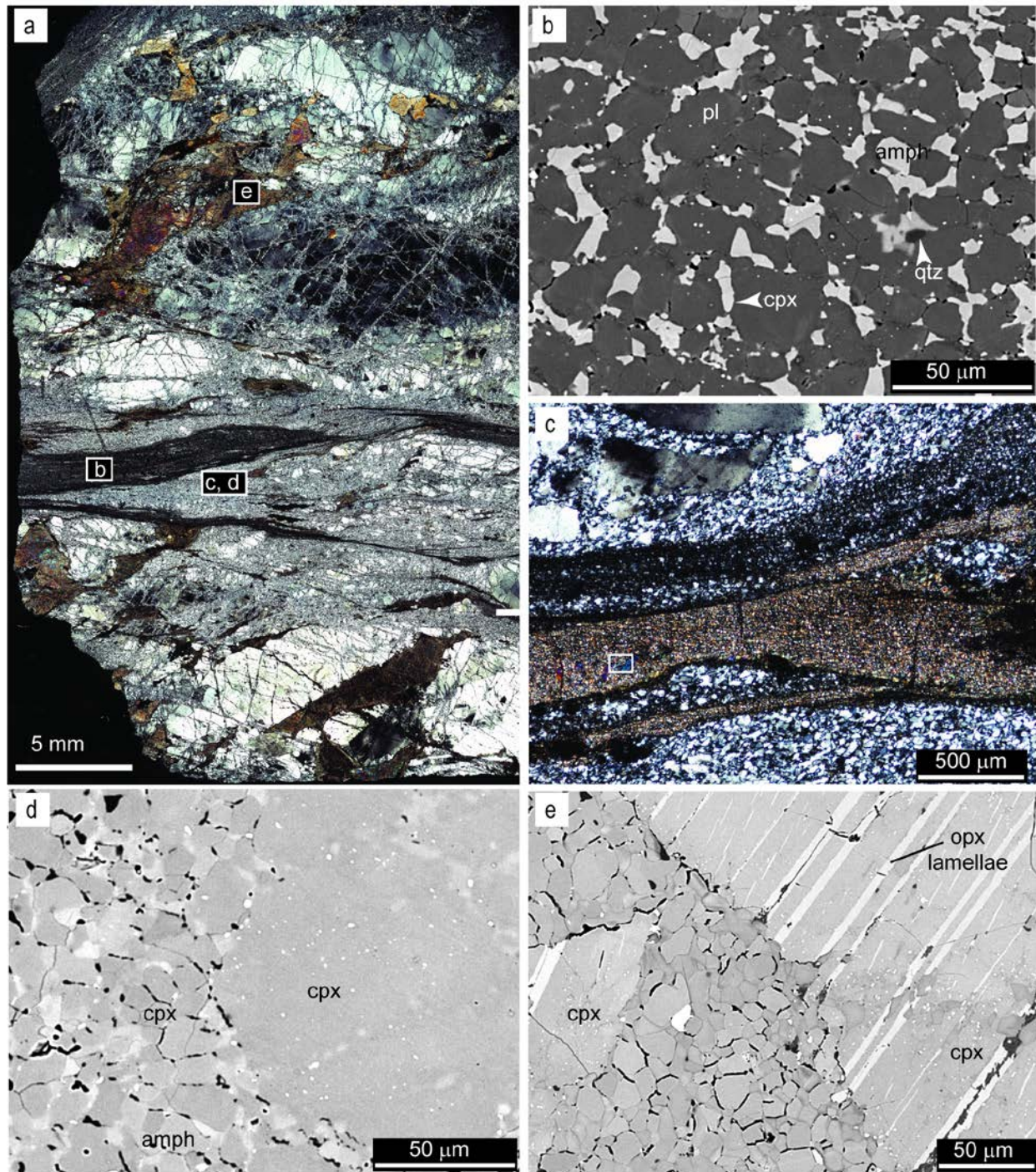
806 thick shear zone with domainal green-whitish layering (orange arrowhead) (f). GPS coordinates: 0432232 East,
 807 7549970 North. Coin for scale (2.5 cm in diameter) in all the field photographs.



808

809 **Figure 4.** Representative microstructures of pristine pseudotachyrites. Mineral abbreviations: amph = amphibole,
 810 bt = biotite, cpx = clinopyroxene, Kfs = K-feldspar, pl = plagioclase, qtz = quartz. (a) Thin section of a pristine
 811 pseudotachylyte vein and of the anorthosite host rock. Crossed polarizers. (b) Plagioclase survivor clast and
 812 microlitic microstructure. Note the cluster of garnet grains overgrowing the plagioclase survivor clast. Parallel
 813 polarizers. (c) SEM backscatter electron image of microlites of clinopyroxene, orthopyroxene, amphibole and

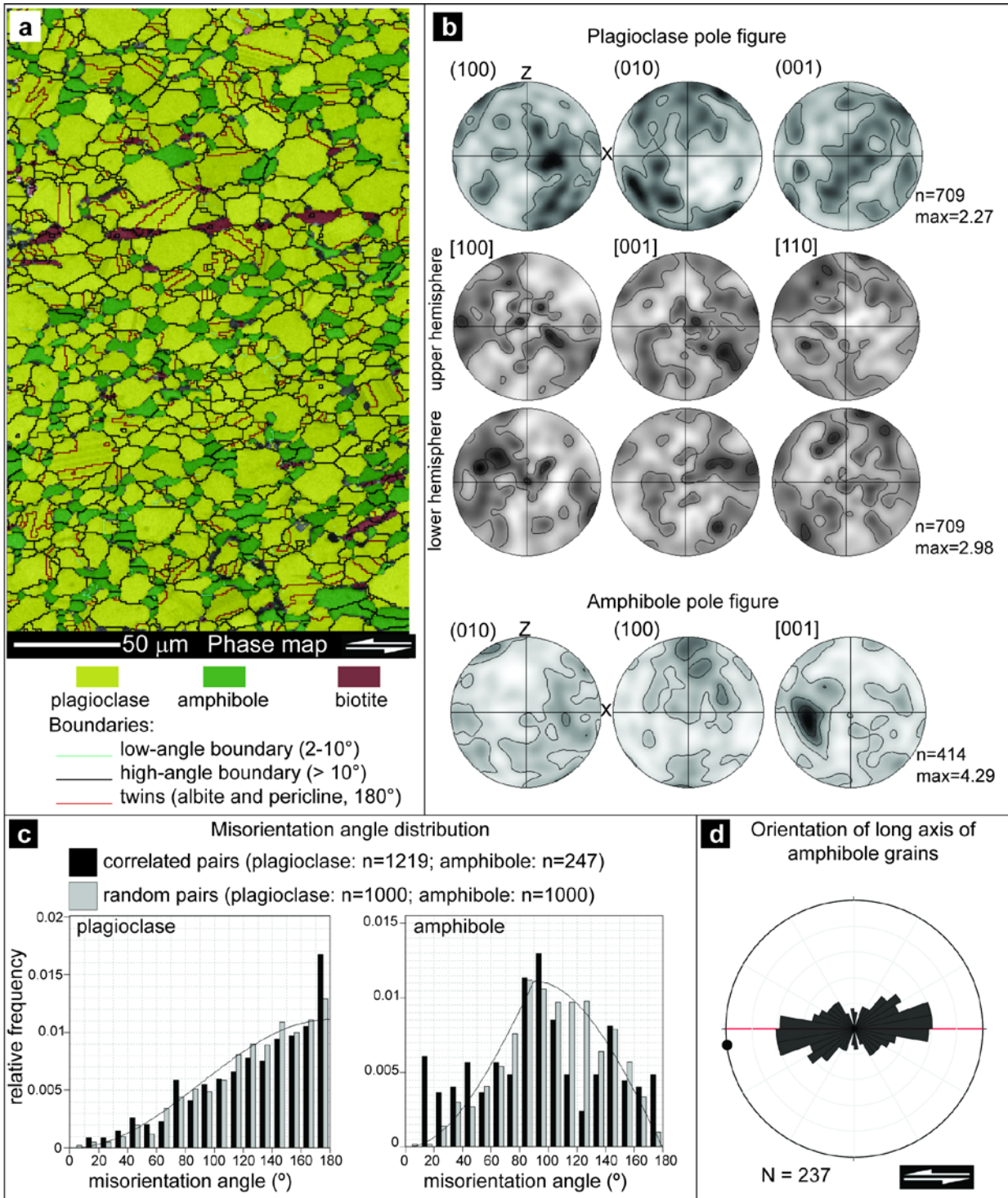
814 plagioclase (dark grey phase). (d) Example of chilled margin of a pseudotachylyte vein. Parallel polarizers. (e) SEM
 815 backscatter electron image of the fine-grained phase mixture found in the recrystallized pseudotachylytes.



816

817 **Figure 5.** Representative microstructures of ductile shear zones. Mineral abbreviations like in Fig. 4. (a) Thin
 818 section of a ductile shear zone consisting of mylonitized pseudotachylytes (transposed and sheared dark veins) and
 819 of variably fractured and recrystallized anorthosite. Black rectangles show the microstructural position of figures 5b-
 820 5e. Crossed polarizers. (b) SEM backscatter electron image of the fine-grained mylonitized pseudotachylyte vein
 821 shown in (a). (c) Bands of recrystallized plagioclase (with relict fragments of magmatic grains) alternating with
 822 clinopyroxene-derived aggregate (brown band) along the shear zone foliation. The clinopyroxene-derived

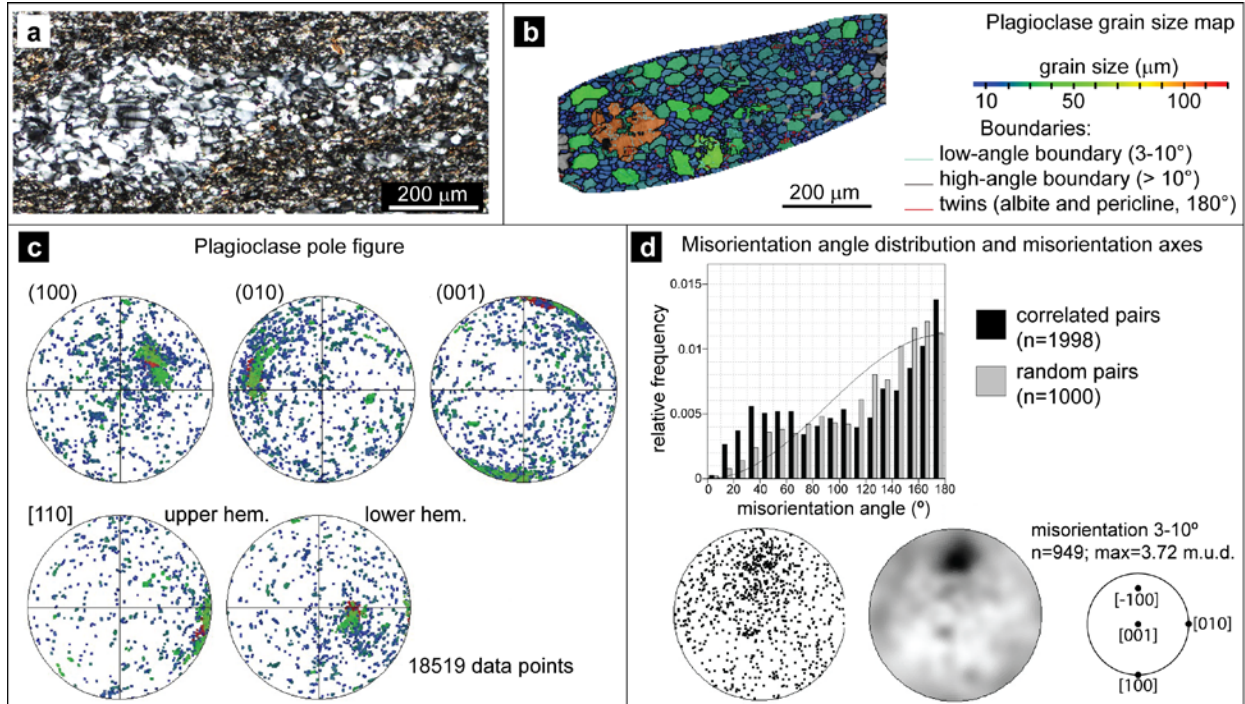
823 aggregate consists of clinopyroxene porphyroclasts embedded in a fine-grained matrix of clinopyroxene, amphibole
 824 and quartz. Light microscope image, crossed polarizers. The white rectangle shows the approximate position of Fig.
 825 5d. (d) SEM backscatter electron image of a detail of the site shown in (c). (e). Fractured clinopyroxene in the
 826 amage zone flanking the shear zone. The intracrystalline fracture is filled with small grains of clinopyroxene and
 827 amphibole.



828 **Figure 6.** EBSD analysis of a mylonitized pseudotachylyte vein. (a) Phase map of a fine-grained mylonitized
 829 pseudotachylyte. (b) Contoured pole figures (half width 15° , data clustering 5°) of plagioclase and amphibole from
 830

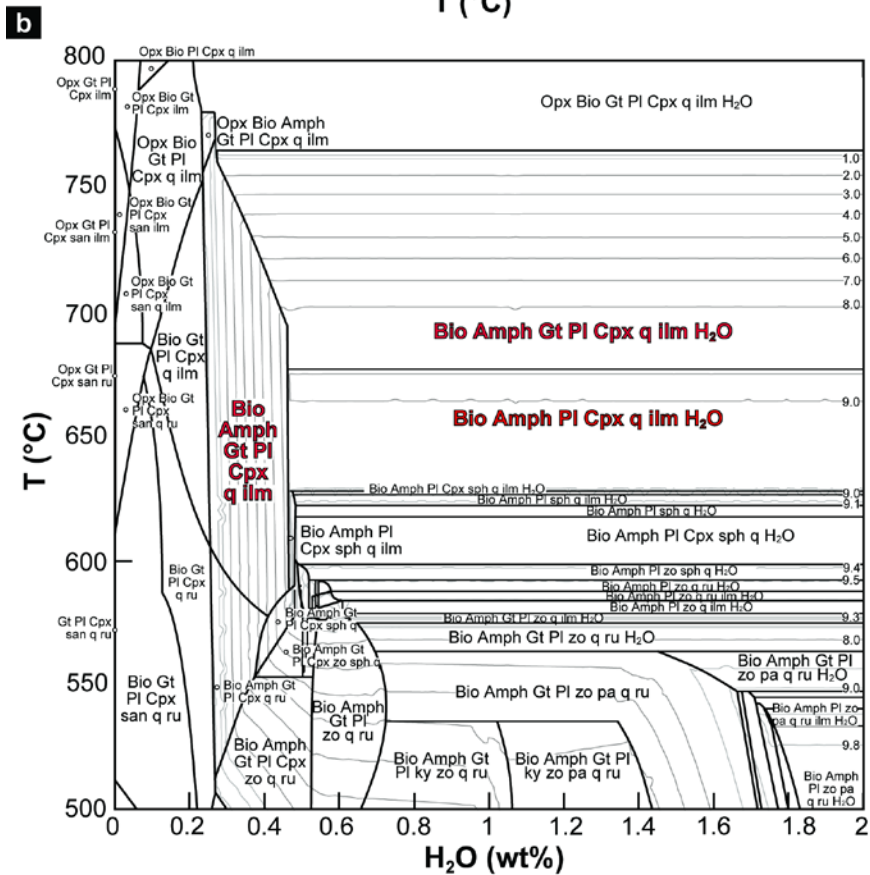
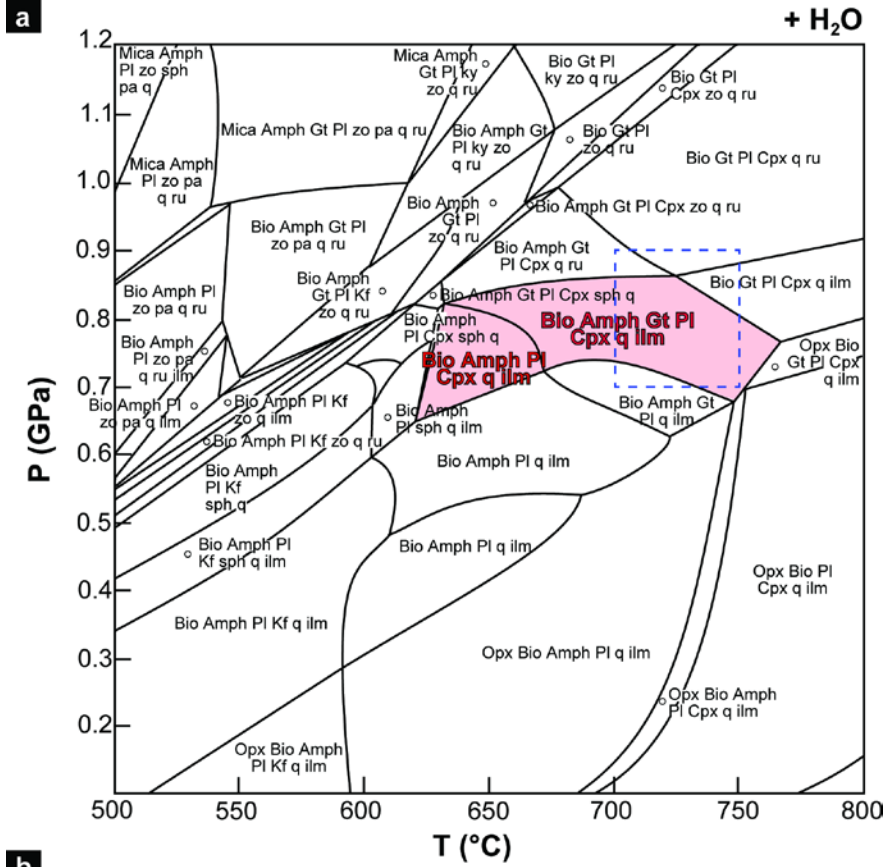
831 the site shown in (a), plotted as one point per grain (n = number of grains). Black colour marks maxima, also given
832 as multiples of the uniform distribution. (c) Histograms of misorientation axis distribution of plagioclase and
833 amphibole from the site shown in (a). The theoretical curves for a random orientation are also shown. (d) Rose
834 diagram showing the orientation of the long axis of amphibole grains included in (a). The red line is parallel to the
835 trace of the shear zone boundary.

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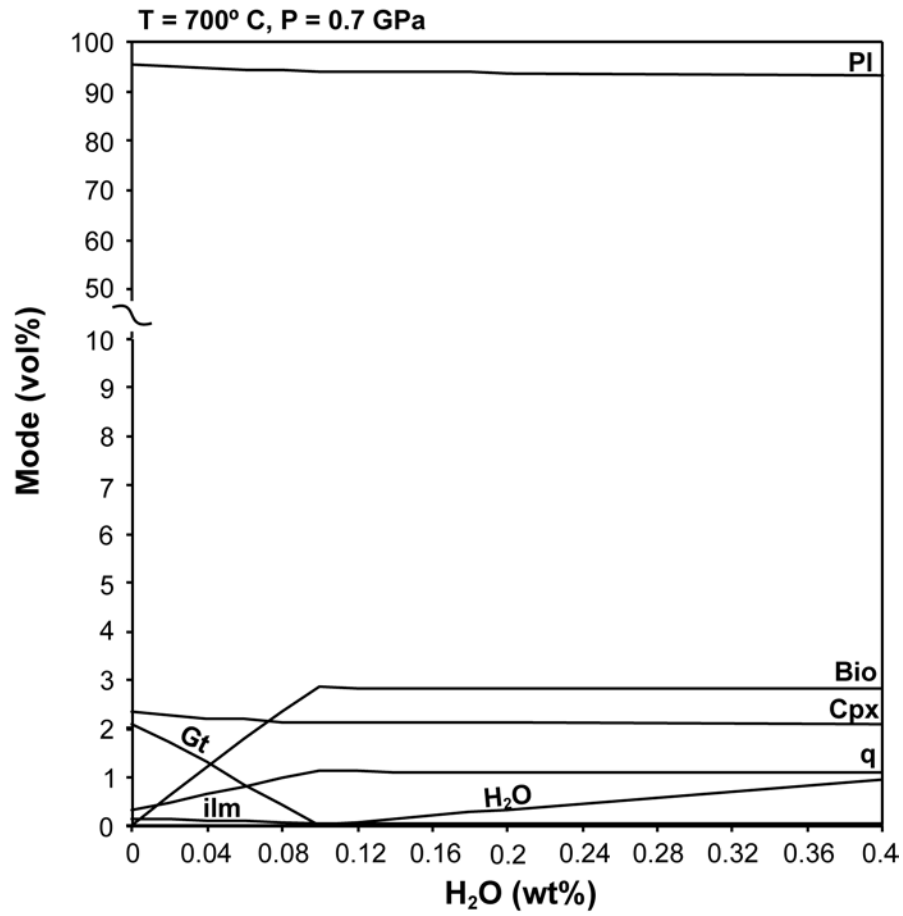
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Figure 7. EBSD analysis of a recrystallized plagioclase aggregate embedded in a mylonitized pseudotachylite vein. (a) Microstructure of the analysed aggregate. Crossed polarizers. (b) Grain size map of the plagioclase aggregate shown in (a). The large grain color-coded orange in interpreted as a relict fragment of the primary plagioclase grain. (c) Plagioclase pole figure of the grains included in (b), and colour coded like the grain size map shown in (b). (d) Misorientation angle distribution histogram of plagioclase and misorientation axis for misorientations of 3-10° plotted in crystal coordinates, both as point data and as contoured plot. The theoretical curve for a random distribution is also shown.



847 **Figure 8.** (a) Isochemical P - T section for mylonitized pseudotachylyte calculated in the system SiO_2 (53.03) –
848 Al_2O_3 (24.84) – MgO (1.87) – CaO (9.96) – Na_2O (4.24) – K_2O (0.89) – TiO_2 (0.31) – FeO (3.32) – Fe_2O_3 (0.11).
849 Values in brackets are oxide wt.%. The red area shows the hypothetical stability field of the mylonite paragenesis, as
850 evidenced by microstructural and mineral chemistry analyses. Abbreviations for solid solution models: Amph –
851 amphiboles; Bio – biotite; Cpx – clinopyroxene; Gt – garnet; Kfs – alkali feldspar; Mica – phengite; Pl – ternary
852 feldspar. Abbreviations for phases: zo – zoisite; sph – titanite (sphene); q – quartz; pa – paragonite; ky – kyanite; ru
853 – rutile; ilm – ilmenite. The dashed blue rectangle shows the P , T conditions estimated with amphibole-plagioclase
854 geothermobarometry. (b) Isochemical T - H_2O section (same composition) calculated at 0.7 GPa, showing the stable
855 paragenesis from dry conditions up to 2 wt% of water. In the temperature range of interest (600-750 °C) the system
856 becomes H_2O saturated (i.e. free water) when the H_2O of the bulk composition is higher than 0.3-0.5 wt%
857 (depending on the temperature). The grey lines and the corresponding numbers on the right-hand side of the plot
858 indicate the amount of amphibole (vol%) occurring in the mineral assemblage.

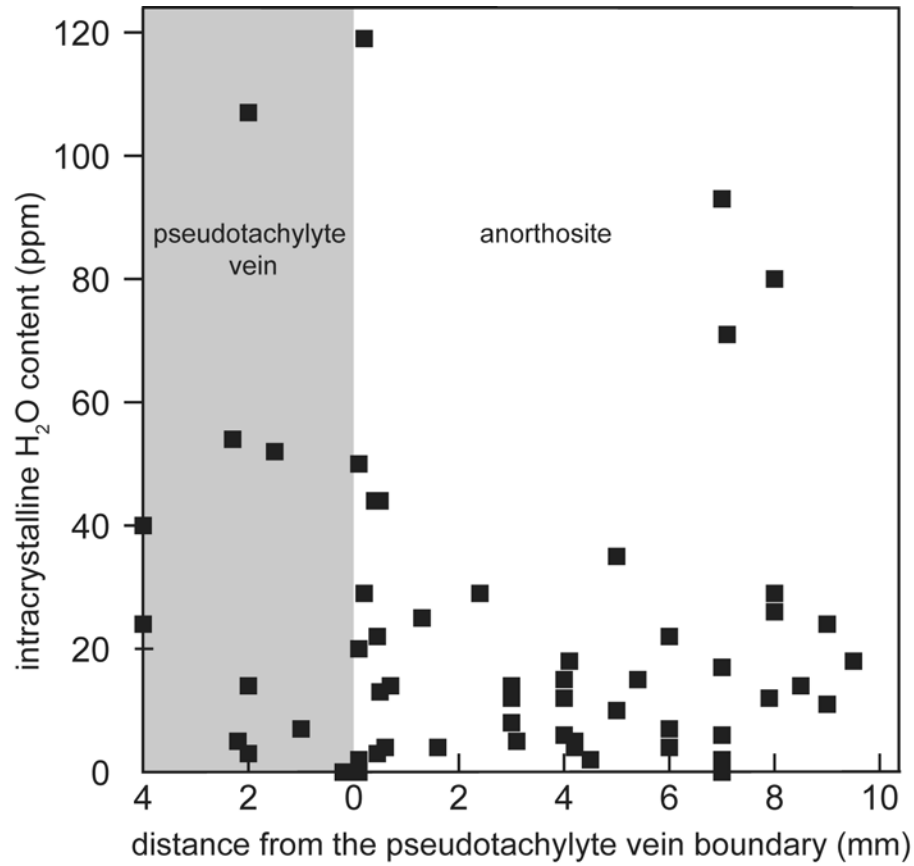
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861 **Figure 9.** Modal phase abundances (vol%) of the isochemical section calculated at 0.7 GPa and 700 °C for the
 862 pristine anorthosite, calculated in the system SiO₂ (53.83) – Al₂O₃ (27.22) – MgO (0.41) – CaO (10.75) – Na₂O
 863 (4.75) – K₂O (0.61) – TiO₂ (0.14) – FeO_{tot} (1.23). Mineral and solid solution abbreviations are the same as in Fig.8.

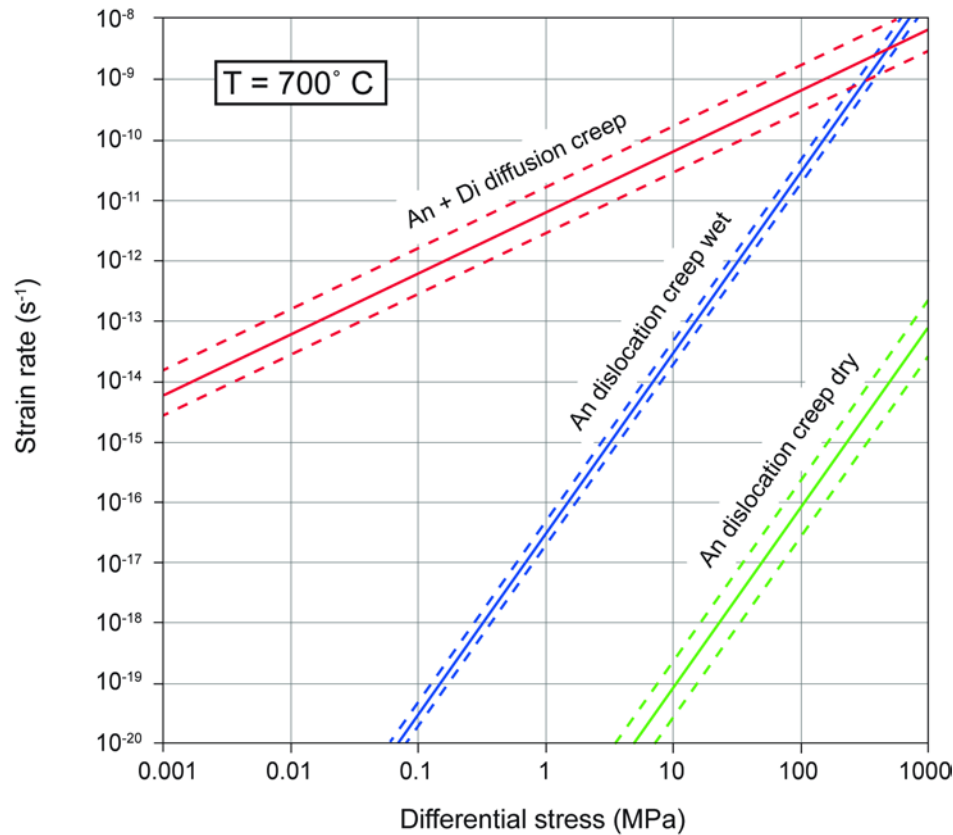
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866 **Figure 10.** Results of the ion microprobe measurements of intracrystalline water content in plagioclase. The area
867 shaded in grey corresponds to the interior of a pristine pseudotachylyte vein.

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870 **Figure 11.** Plot of strain rate (s^{-1}) versus differential stress (MPa) to model the rheology of the mylonitized
 871 pseudotachylytes at a temperature of 700°C. The following flow laws were used: anorthite + diopside wet diffusion
 872 creep (the $An_{25}Di_{35}W$ flow law of Dimanov and Dresen, 2005), anorthite wet dislocation creep (Rybacki and
 873 Dresen, 2000; water content = 0.07 wt%), anorthite dry dislocation creep (Rybacki and Dresen, 2000; water content
 874 = 0.004 wt%). Dashed lines represent uncertainties in the pre-exponential term A and in the activation energy Q of
 875 each flow law. We used the average stress exponents derived experimentally ($n = 1$ for the diffusion creep flow law,
 876 $n = 3$ for the dislocation creep flow laws).