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1    **Replacement reactions and deformation by dissolution and precipitation processes in**  
2    **amphibolites**

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8    **Short title: dissolution-precipitation in amphibolites**

## 9    **Abstract**

10    The deformation of the middle to lower crust in collisional settings occurs via deformation  
11    mechanisms that vary with rock composition, fluid content, pressure and temperature. These  
12    mechanisms are responsible for the accommodation of large tectonic transport distances during nappe  
13    stacking and exhumation. Here we show that fracturing and fluid flow triggered coupled dissolution-  
14    precipitation and dissolution-precipitation creep processes, which were responsible for the formation  
15    of a mylonitic microstructure in amphibolites. This fabric is developed over a crustal thickness of  
16    >500 m in the Lower Seve Nappe (Scandinavian Caledonides). Amphibolites display a mylonitic  
17    foliation that wraps around albite porphyroclasts appearing dark in panchromatic  
18    cathodoluminescence. The albite porphyroclasts were dissected and fragmented by fractures  
19    preferentially developed along the (001) cleavage planes, and display lobate edges with embayments  
20    and peninsular features. Two albite/oligoclase generations, bright in cathodoluminescence, resorbed  
21    and overgrew the porphyroclasts, sealing the fractures. Electron backscattered diffraction shows that  
22    the two albite/oligoclase generations grew both pseudomorphically and topotaxially at the expenses  
23    of the albite porphyroclasts and epitaxially around these. These two albite/oligoclase generations also  
24    grew as neoblasts elongated parallel to the mylonitic foliation. The amphibole crystals experienced a  
25    similar microstructural evolution, as evidenced by corroded ferrohornblende cores surrounded by  
26    ferrotschermakite rims that preserve the same crystallographic orientation of the cores. Misorientation  
27    maps highlight how misorientations in amphibole are related to displacement along fractures  
28    perpendicular to its c-axis. No crystal plasticity is observed in either mineral species. Plagioclase and  
29    amphibole display a crystallographic preferred orientation that is the result of topotaxial growth on  
30    parental grains and nucleation of new grains with a similar crystallographic orientation. Amphibole  
31    and plagioclase thermobarometry constrains the mylonitic foliation development to the epidote  
32    amphibolite facies (~600°C, 0.75-0.97 GPa). Our results demonstrate that at middle to lower crustal  
33    levels the presence of H<sub>2</sub>O-rich fluid at grain boundaries facilitates replacement reactions by coupled

34 dissolution-precipitation and favours deformation by dissolution-precipitation creep over dislocation  
35 creep in plagioclase and amphibole.

## 36 **Keywords**

37 Replacement reactions, dissolution-precipitation processes, Caledonides, electron backscatter  
38 diffraction, X-ray mapping.

## 39 **1. INTRODUCTION**

40 The thermo-mechanical properties of the middle to lower crust exert a fundamental control on the  
41 structure of orogenic belts, and on the amount and style of shortening during continental collision  
42 (e.g. Jackson, Austrheim, McKenzie, & Priestley, 2004; Mouthereau, Watts, & Burov, 2013). In  
43 particular, how strain is distributed vertically and horizontally in orogenic belts is one of the more  
44 important questions in crustal dynamics, and one that can be addressed by investigating the  
45 deformation mechanisms associated with the accumulation of hundreds of km tectonic transport  
46 distances along thrust faults during mountain building processes (e.g. Fousseis & Handy, 2008; Gilotti,  
47 1989; Mouthereau, Lacombe, & Vergés, 2012; Northrup, 1996; Royden, 1996).

48 In particular, the deformation processes and rheology of mafic shear zones are the subject of  
49 considerable debate, because their main mineral constituents (e.g. plagioclase, amphibole,  
50 clinopyroxene) are expected to be rheologically strong at middle to lower crustal conditions (e.g.  
51 Bürgmann & Dresen, 2008). Thus, the weakening of mafic assemblages along major thrust faults  
52 developed at middle to lower crustal conditions seems to critically depend on the occurrence of  
53 metamorphic reactions, which can result in the formation of rheologically weaker phases, or in the  
54 formation of fine-grained material able to deform by grain size sensitive creep, or in both (e.g. Rutter  
55 & Brodie, 1992; Brander, Svahnberg, & Piazzolo, 2012; Okudaira, Shigematsu, Harigane, & Yoshida,  
56 2017). Furthermore, the presence of H<sub>2</sub>O-rich fluid at the grain boundary typically facilitates  
57 dissolution and precipitation processes, which have been identified as the main deformation



mechanisms in different mid-crustal lithologies and up to high temperature and high pressure conditions (Carmichael, 1969; Giuntoli, Lanari, & Engi, 2018; Gratier, Dyshe, & Renard, 2013; Imon, Okudaira, & Fujimoto, 2002; Imon, Okudaira, & Kanagawa, 2004; McAleer et al., 2017; Menegon, Pennacchioni, & Spiess, 2008; Mukai, Austrheim, Putnis, & Putnis, 2014; Putnis, 2009; Rutter, 1983; Stokes, Wintsch, & Southworth, 2012; Wassmann & Stöckhert, 2013; Wassmann, Stöckhert, & Trepmann, 2011; Wintsch & Yi, 2002). Two main dissolution and precipitation processes can be distinguished: coupled dissolution-precipitation and dissolution-precipitation creep. Coupled dissolution-precipitation results in the pseudomorphic (maintaining the size and shape of the pre-existing phase) and topotaxial (using the orientation of the pre-existing phase) replacement of a parent phase by a product phase from a reaction interface (e.g. Putnis, 2002; Ruiz-Agudo, Putnis, & Putnis, 2014). Dissolution-precipitation creep includes the transport of the chemical constituents from the dissolution sites, of locally high normal stress, to the precipitation site, of locally low normal stress (e.g. Imon et al., 2002; Imon et al., 2004; Mukai et al., 2014; Wassmann et al., 2011). Both processes invariably require the presence of a fluid.

Porosity is another essential requirement for the operation of dissolution and precipitation processes, as it provides the necessary fluid pathways. Porosity may develop during replacement reactions due to molar volume differences between the dissolved and the precipitated minerals (Engvik, Putnis, Gerald, & Austrheim, 2008), or from dehydration reactions and associated fracturing (Plummer, John, Podladchikov, Vrijmoed, & Scambelluri, 2017). Other porosity-generating mechanisms in metamorphic environments include dilatancy at grain boundaries (Tullis, Yund, & Farver, 1996), fracturing (e.g. Brander et al., 2012), and creep cavitation in fine-grained ultramylonites deforming by grain size sensitive creep (Fusseis, Regenauer-Lieb, Liu, Hough, & De Carlo, 2009; Menegon, Fusseis, Stünitz, & Xiao, 2015).

Here we show that mineral reactions and deformation in amphibolites occurred mainly by coupled dissolution-precipitation and dissolution-precipitation creep at epidote-amphibolite facies conditions,

83 and that fracturing was the most efficient porosity-generating mechanism assisting deformation.  
84 Metamorphism and deformation in the amphibolites resulted in the development of a > 500 m thick  
85 mylonitic foliation during Caledonian nappe thrusting.

## 86 **2. GEOLOGICAL SETTING**

87 The Scandinavian Caledonides developed due to the closure of the Iapetus Ocean in the Ordovician,  
88 and the subsequent subduction and continent collision of the Baltica plate below the Laurentia plate  
89 in the Silurian to early Devonian (e.g. Gee, Fossen, Henriksen, & Higgins, 2008; Roberts, 2003;  
90 Roberts & Gee, 1985; Stephens, 1988). In the Scandinavian Caledonides, tectonic units were  
91 transported up to 400 km to the east (Gayer, Rice, Roberts, Townsend, & Welbon, 1987; Gee, 1975;  
92 Gee, Juhlin, Pascal, & Robinson, 2010; Rice & Anderson, 2016; Roberts & Gee, 1985) as a result of  
93 the collision, creating a nappe stack of several allochthons units on top of Autochthons Baltic Shield  
94 (Figure 1a-b). After emplacement, the nappe stack was folded into north-trending synforms and  
95 antiforms, possibly related to the crustal extension and normal faulting occurring during the latest  
96 orogenic phases (Bergman & Sjöström, 1997). By virtue of their deep erosional level, the internal  
97 parts of the Scandinavian Caledonides expose middle and lower crustal sections involved in  
98 subduction-exhumation history and nappe stacking.

99 The Scandinavian Caledonides are subdivided, from top to bottom, into an Uppermost Allochthon,  
100 an Upper Allochthon, a Middle Allochthon, and a Lower Allochthon based on tectonostratigraphy  
101 (Figure 1; Gee & Sturt, 1985; Strand & Kulling, 1972; Strömberg et al., 1984). The Middle  
102 Allochthon, the target of this study, includes several basement units and associated metasediments  
103 representing the outermost Baltica margin and possibly including units derived from an ocean-  
104 continent transition zone (e.g. Andréasson, 1994; Gee et al., 2008; Gee, Janák, Majka, Robinson, &  
105 van Roermund, 2013; Roberts, 2003; Stephens, 1988). The upper tectonic unit of the Middle  
106 Allochthon is the Seve Nappe Complex (SNC; e.g. Sjöström, 1983), which, in the central Scandes,

107 outcrops over a N-S distance of ~1000 km and an W-E distance of ~200 km (Figure 1; Andréasson,  
108 1994).

109 In the Jämtland region, the SNC can be further subdivided into Lower, Middle and Upper Seve Nappe  
110 by the presence of internal thrust sheets (Zachrisson & Sjöstrand, 1990). The Lower Seve Nappe is  
111 mainly composed of micaschists, quartzites and metapsammities with gneisses, metabasics and with  
112 minor peridotites and serpentinites (Figure 1c). The Middle Seve Nappe is composed of similar  
113 lithotypes, but overprinted by a pervasive migmatization. Several parts of the Lower- and Middle  
114 Seve preserve evidence of high pressure (HP) to ultrahigh pressure (UHP-) metamorphism (summary  
115 in Figure 4 of Klonowska, Janák, Majka, Froitzheim, & Kościńska, 2016) spanning from ~1.1 GPa  
116 and 600°C up to 4 GPa and 800°C, within the stability field of coesite and diamond (Brueckner &  
117 van Roermund, 2004; Gilio, Clos, & van Roermund, 2015; Janák, van Roermund, Majka, & Gee,  
118 2013; Klonowska et al., 2016; Klonowska et al., 2017; Majka et al., 2014; Van Roermund, 1985,  
119 1989). The HP-UHP metamorphism is the manifestation of the Ordovician subduction of the SNC  
120 (Brueckner & Van Roermund, 2007; Ladenberger et al., 2013; Root & Corfu, 2012). It is worth noting  
121 that to date, no evidence of (U)HP metamorphism has been recorded in the Lower Seve Nappe in the  
122 central Jämtland. In the Middle Seve Nappe, the granulite and amphibolite facies metamorphism  
123 appears to postdate the HP-UHP stage, and produced partial melting at 442–436 Ma (Ladenberger et  
124 al., 2013). A recent study has related the amphibolite facies metamorphic “stage” to the exhumation  
125 and lateral extrusion of the SNC (Grimmer, Glodny, Drüppel, Greiling, & Kontny, 2015). In the  
126 Lower Seve Nappe, a pervasive amphibolite facies foliation overprints the (U)HP fabric where  
127 present; where not present it represents the main metamorphic fabric. In the Åreskutan area, the  
128 amphibolite facies metamorphic stage was constrained at 550°C and 0.2 to 0.5 GPa (Arnbom, 1980).

129 The “Collisional Orogeny in the Scandinavian Caledonides (COSC-1)” borehole (Lorenz et al., 2015;  
130 see location in Figure 1b-c) is located in the central Jämtland region, near Åre (Sweden). The drill  
131 core provides an almost complete section (recovery rate higher than 99%) through the Lower Seve

132 Nappe. In detail, the core comprises alternating layers of felsic gneisses, calc-silicates and  
133 amphibolites displaying narrow (mm-cm) and localized shear zones from the surface down to 1700  
134 m (Hedin et al., 2016). The rocks show strongly deformed fabrics from 1700 m to 2500 m (the end  
135 of the core; Lorenz et al., 2015), with the development of mylonitic fabrics. The lowermost portion  
136 of the core is composed of strongly deformed metasediments. Acoustic televiewer data indicates that  
137 the foliation is generally shallow and trends N-S, but there are remarkable exceptions, where the  
138 foliation has dips to the E or W with angles  $>50^\circ$ , related to recumbent folds and boudinage (Wenning  
139 et al., 2017).

### 140 **3. METHODS**

#### 141 **3.1 Scanning electron microscopy (SEM)**

142 All the SEM analyses were performed on carbon-coated polished thin sections cut perpendicular to  
143 the foliation and parallel to the stretching lineation of the sample. Backscattered electron (BSE) and  
144 cathodoluminescence (CL) analyses were performed at the Open University (UK), using an FEI  
145 Quanta 200 three-dimensional SEM on carbon coated thin sections. Analyses were conducted under  
146 high vacuum, using an accelerating voltage of 10 kV, a beam current of 3.3 nA, a working distance  
147 of 13 mm, and an electron source provided by a tungsten filament. The panchromatic CL detector  
148 used is a Centaurus Deben with a photo multiplier tube (Hamamatsu R316) characterized by  
149 sensitivity in the range of 400-1200nm.

150 Electron backscattered diffraction (EBSD) analyses were conducted with a Jeol-7001FEG SEM at  
151 the Electron Microscopy Centre, Plymouth University (UK). EBSD patterns were acquired with a  
152  $70^\circ$  tilted sample geometry, 20 kV accelerating voltage, 18-23 mm working distance and 1.3-1.7  $\mu\text{m}$   
153 step size. Diffraction patterns were automatically indexed using AZtec (Oxford Instruments). The  
154 indexing match units used for the analysed phases were “anorthite” (Laue group -1) for albite and  
155 oligoclase, and “hornblende” (Laue group 2/m) for tschermakite-ferrotschermakite. Both match units

were taken from the American Mineralogist database. Raw maps were processed with HKL Channel 5 (Oxford Instruments), using the noise reduction procedure tested by Prior, Wheeler, Peruzzo, Spiess, and Storey (2002). Wild spikes were removed, and un-indexed points were replaced by the average orientation of the neighbour points. Grains smaller than 3 times the step size were not considered in the analysis. The mean angular deviation values were 0.3 for amphibole and 0.4-0.6 for plagioclase; the raw indexing rate ranged between 90% and 95%. Crystallographic directions were plotted on pole figures (upper and lower hemisphere of the stereographic projection), with X parallel to the stretching lineation and Z parallel to the pole of the mylonitic foliation. The grain orientation spread maps (GOS maps) were calculated as the average misorientation between every pixel in the grain and the grain's average orientation.

### **3.2 Electron probe micro-analyser (EPMA)**

EPMA analyses were conducted at the Open University (UK), using a Cameca SX100 connected to five spectrometers. Wavelength dispersive spectrometers (WDS) were used for both spot analyses and X-ray maps, the latter acquired following the procedure of Lanari et al. (2013). Spot analyses were acquired first for each mineral phase, before performing the X-ray maps on the same area. Spot analyses were performed with 20 KeV accelerating voltage, 20 nA specimen current and 2  $\mu\text{m}$  beam diameter. Ten oxide compositions were measured, using natural standards: K-feldspar ( $\text{SiO}_2$ ,  $\text{Al}_2\text{O}_3$ ,  $\text{K}_2\text{O}$ ), bustamite ( $\text{CaO}$ ,  $\text{MnO}$ ), hematite ( $\text{FeO}$ ), forsterite ( $\text{MgO}$ ), jadeite ( $\text{Na}_2\text{O}$ ), rutile ( $\text{TiO}_2$ ), apatite ( $\text{P}_2\text{O}_5$ ). A ZAF matrix correction routine was applied; uncertainty on major element concentrations was <1%. X-ray maps were acquired with 15 KeV accelerating voltage, 100 nA specimen current, dwell times of 70-100 ms and step size of 5  $\mu\text{m}$ . Ten elements (Si, Ti, Al, Fe, Mn, Mg, Na, Ca, K and P) were measured at the specific wavelength in two series. Intensity X-ray maps were standardized to concentration maps of oxide weight percentage using spot analyses as internal standard. X-ray maps were processed using XMapTools 2.2.1 (Lanari et al., 2014).

### **3.3 Geothermobarometry**

### 3.3.1 Amphibole-plagioclase thermobarometry

Temperature (T) was estimated using Holland and Blundy (1994) geothermometer. In detail, the Holland and Blundy (1994) thermometer is based on element exchange between amphibole and plagioclase pairs in equilibrium; the calibration reaction edenite + albite = richterite + anorthite was constrained from experimental and natural data for silica-saturated and silica-rich igneous and metamorphic rocks in the range of 0.1-1.5 GPa and 400-1000 °C.

Pressure (P) was estimated using two geobarometers: Bhadra and Bhattacharya (2007) and Anderson and Smith (1995). The former is based on element distribution between amphibole and plagioclase pairs in equilibrium. Experimental data were conducted on silica-saturated assemblages in the P-T range of 0.1-1.5 GPa and 650-950 °C and on the reaction tremolite + tschermakite + 2 albite = 2 pargasite + 8 quartz. The latter is based on the increase of Al content in hornblende with increasing P and is calibrated on experimental data at 675 and 760 °C, accounting for the effects of T and  $f_{O_2}$ . P and T were calculated using the Plagioclase–Hornblende Thermobarometry spreadsheet (Anderson, Barth, Wooden, & Mazdab, 2008) and the spreadsheet of Wallis, Phillips, and Lloyd (2014) (the latter was used to derive P from the calibration of Bhadra & Bhattacharya, 2007).

### 3.3.2 Chlorite+Quartz+H<sub>2</sub>O thermometry

Chlorite+Quartz+H<sub>2</sub>O thermometry was performed using the program CHLMICA EQUI (Lanari, 2012). The crystallization temperature of chlorite and the  $X_{Fe^{3+}}$  were computed at a fixed pressure of 0.7 GPa in the temperature range of 100 to 550°C from the convergence of four equilibria involving five chlorite end-members, quartz and H<sub>2</sub>O (Lanari et al., 2012; Vidal, Lanari, Munoz, Bourdelle, & De Andrade, 2016) using standard state properties and solid solution models of Vidal, Parra, and Vieillard (2005; 2006).

## 4. RESULTS

### 4.1 Petrography and microstructure

205 In the middle and lower portions of the COSC-1 borehole (~1500-2300 m deep), amphibolites show  
206 a mylonitic foliation. The studied sample (International Geological Sample Number:  
207 ICDP5054EX8E601) was extracted from a depth of 2206.97 metres from the surface and is similar  
208 to several amphibolite samples present between ~1600 and ~2300 m of depth. In this sample the  
209 mylonitic foliation is subhorizontal, it wraps around plagioclase porphyrocrysts and is defined by  
210 amphibole, plagioclase (with smaller grain size, details in the following), chlorite, quartz, epidote and  
211 ilmenite (Figure 2). Quartz and calcite occur mainly in veins subparallel to the mylonitic foliation or  
212 as fine grains dispersed in the mylonitic foliation, typically along the phase boundaries between  
213 plagioclase and amphibole.

214 Plagioclase is present as zoned porphyrocrysts (up to 0.5-1 cm in size) wrapped by the mylonitic  
215 foliation, and as finer (up to hundreds of  $\mu\text{m}$  in size) albite/oligoclase grains elongated parallel to the  
216 mylonitic foliation (aspect ratios up to  $\sim 5$ ; mean  $\sim 2$ ). Plagioclase porphyrocrysts have albite cores  
217 that appear turbid due to abundant fine-grained (up to 30  $\mu\text{m}$  in size) inclusions of ilmenite, rutile and  
218 epidote. Ilmenite crystals are also included with bigger grain size (100 microns in size); quartz is  
219 present as inclusions with variable grain size (tens to hundreds of  $\mu\text{m}$  in size; Figures 3 and 4). The  
220 inclusions locally define a rotated internal foliation (upper plagioclase porphyrocryst in Figure 3a).  
221 These cores are porphyroclastic, and are cut by fracture systems and some are dismembered parallel  
222 to the foliation (see details in section 4.2). The cores are always surrounded by albite/oligoclase rims.  
223 Pores are evident in the albite/oligoclase rims: they are few microns in size and occur in trails parallel  
224 to fracture systems that dissect the plagioclase core (Figure 4). Several mineral inclusions also occur  
225 along such trails, ranging from a few microns to tens of microns in size. These are muscovite, epidote  
226 and calcite, with rare Ba-rich muscovite (Figure 4). Such inclusions confer a turbid aspect to  
227 plagioclase rim in plane-polarized light micrographs (Figure 3a, Table 1).

228 Amphibole is pleochroic, with absorption colours ranging from light brown to dark green-blue; in  
229 some bigger crystals absorption colours highlight a paler core and a darker rim. Amphibole displays

230 the C-axis parallel to the stretching lineation and a maximum grain size of >1 mm. Chlorite has a  
231 pleochroism varying from light yellow to light green, a negative optical sign and grain size similar to  
232 the amphibole. Asymmetric pressure shadows around plagioclase porphyroclasts indicate both  
233 dextral and sinistral sense of shear. In these pressure shadows amphibole, albite/oligoclase grains and  
234 chlorite crystallize intergrown and elongated parallel to the mylonitic foliation. C' planes display a  
235 dominant sinistral sense of shear and are defined by the same minerals found along the mylonitic  
236 foliation (Figures 2, 3a-b). Opaque minerals include mostly ilmenite, up to 1 mm in size, with minor  
237 magnetite and pyrite. Ilmenite crystals are elongate and lie parallel to the mylonitic foliation.

#### 238 **4.2 Microstructure and chemistry of plagioclase**

239 BSE images highlight two plagioclase generations: dark cores with an albitic composition (P11) are  
240 surrounded by brighter plagioclase rims with higher anorthite content (P12 and P13, as defined based  
241 on chemistry in the next paragraph; Figures 3c and 4). Furthermore, the cores are cut by fractures  
242 filled with plagioclase with the same composition as the plagioclase rims. The distinction of these  
243 plagioclase generations is more evident in the CL images, where plagioclase cores appear moderately  
244 luminescent and are surrounded by bright plagioclase rims (Figure 3d). The fractures dissecting the  
245 cores are as bright as the plagioclase rims. The plagioclase cores (P11) are locally fragmented and  
246 display lobate edges and embayments that are typically surrounded by bright plagioclase (Figures 3d,  
247 7a and 8b). In some areas, the plagioclase rims can be further subdivided into two generations based  
248 on the brightness of their CL response: a brighter plagioclase generally surrounded by a darker one  
249 (P12 and P13, respectively).

250 Compositional maps of plagioclase display a perfect match with the CL images (Figure 3e). The  
251 maps highlight three plagioclase compositions, numbered from the older to the younger: P11 (XAn 0-  
252 0.05), P12 (XAn 0.13-0.25), P13 (XAn 0.05-0.13; Figures 5a, 7b, Table 2). The plagioclase core (P11)  
253 has an albite composition and is overgrown by the two albite/oligoclase rims (P12 and P13). P12 is  
254 present only locally and is overgrown by P13, as shown in the top right corner of Figure 3e.



255 Furthermore, Pl3 seals the fractures inside Pl1. Pl1 forms porphyroclasts wrapped by the mylonitic  
256 foliation, whereas Pl2 and Pl3 occur in the pressure shadows of Pl1.

### 257 **4.3 Microstructure and chemistry of amphibole and chlorite**

258 Compositional maps of amphibole crystals display corroded cores, higher in Mg#  
259 ( $Mg\# = Mg / (Fe^{2+} + Mg)$ ; Amp1: Mg# 0.5-0.44, Si apfu 7-6.5), and rims, lower in Mg# (Amp2: Mg#  
260 0.44-0.36, Si apfu 6.5-6.1; Figures 3f and 5b, Table 2). Based on the classification of Leake et al.  
261 (1997) the Amp1 is at the compositional boundary between tschermakite-ferrotschermakite-  
262 magnesiohornblende-ferrohornblende; Amp2 is a ferrotschermakite (Figure 5c). There are local  
263 exceptions to this zoning pattern, in which the two peripheral areas have higher Mg# (Figure 9c).  
264 Amp2 and Pl3 grains displaying mutual intergrowths and elongation parallel to the foliation (Figure  
265 3 c-f)

266 Chlorite is characterized by a homogeneous Mg# (Mg# 0.59-0.56, Chl1), except along grain  
267 boundaries and the C' planes, where some lower values are visible (Mg# 0.55-0.52, Chl2; Figure 5c,  
268 Table 3). Chl1 grains occur as intergrown with Pl3 and Amp2 along the mylonitic.

### 269 **4.4 Amp-Pl thermobarometry**

270 Thermometric estimates for the Pl1 and Amp1 pair yield 400°C; thermobarometric estimates for the  
271 Pl2 and Amp1 pair and the Amp2 and Pl3 pair yield 615°C and 0.97 GPa and 605°C and 0.74 GPa  
272 respectively ( $\pm 50^\circ\text{C} \pm 0.2\text{ GPa}$ , Table 4; see Section 5.1 for discussion on the Pl and Amp growth  
273 zones that we consider to be in equilibrium).

### 274 **4.5 Chlorite+Quartz+H<sub>2</sub>O thermometry**

275 Chlorite+Quartz+H<sub>2</sub>O thermometry was performed on the two different chlorite groups, Chl1 and  
276 Chl2 (Figure 5c), at a fixed pressure of 0.7 GPa. Chl1 displays two peaks in the T-frequency diagram  
277 (Figure 6a): one close to the T limit of the thermometer (550°C-450°C) and one at lower T (400°C-

278 250°C). Chl2 yields T in the range of 200-350°C, with a peak at ~250°C (Figure 6b), with only one  
279 grain yielding a higher T (~500°C).

#### 280 **4.6 EBSD analysis**

281 Figures 7-8 show the results of EBSD analysis of two microstructural domains (see Figures 2 and 3  
282 for the location of EBSD maps): domain 1 includes a plagioclase porphyrocryst with a core of Pl1  
283 composition rimmed by Pl2 and Pl3 compositions. The porphyroclast Pl1 is cut by a network of  
284 intracrystalline microfractures oriented NW-SE and filled with Pl2-Pl3. Domain 1 also includes  
285 clusters of amphibole and plagioclase grains around the plagioclase porphyrocryst (Figure 7). Domain  
286 2 contains an aggregate of amphibole grains, mostly elongate parallel to the foliation, and several  
287 grains of plagioclase showing a core-and-rim zoning in CL images (Figure 8).

288 The comparison between EBSD maps, CL images and compositional maps shows that the Pl1  
289 porphyroclasts, which EBSD identifies as one individual grain (i.e. it does not contain high-angle  
290 boundaries with misorientation  $> 10^\circ$ ), actually include areas with Pl2 and Pl3 compositions, which  
291 maintain the original crystal shape and crystallographic orientation of Pl1 (Figures 7 and 8). The  
292 internal distortion of the compositionally zoned porphyrocrysts, as evaluated with the GOS, is rather  
293 low (GOS values lower than  $2^\circ$ : Figures 7c and 8e). Low-angle boundaries (misorientations  $2\text{-}10^\circ$ )  
294 are preferentially distributed along the microfracture. The analysis of local misorientations within Pl1  
295 cores indicates that Pl2 and Pl3 areas have misorientations of up to  $2^\circ$  with respect to the Pl1 cores  
296 (Figure 7d). Higher misorientations (up to  $3^\circ$ ) are found in the plagioclase porphyrocryst of the  
297 domain 2 (Figure 8d).

298 In domain 1, the compositionally zoned plagioclase porphyrocryst is surrounded by Pl2-Pl3 grains  
299 that range in size between 30 and 200  $\mu\text{m}$ , delimited by high-angle boundaries with misorientation  $>$   
300  $10^\circ$  from the porphyrocryst. The Pl2 and Pl3 grains have mean GOS values lower than  $1^\circ$ , with one  
301 maximum value of  $4^\circ$ .

302 Amphibole generally shows GOS values lower than  $2^\circ$ , with a few maxima of up to  $7^\circ$  (Figures 7e,  
303 8g, 9d). Misorientations progressively increase towards fractures perpendicular to the crystal  
304 elongation and, locally, towards very few low angle boundaries oriented subparallel to the crystal  
305 elongation (Figures 7f, 8h, 9f). The misorientation profile in Figure 9e highlights a jump in  
306 misorientation of up to  $7^\circ$  across the low angle boundary.

307 The crystallographic orientation of the new grains of P12 and P13 displays a variable degree of overlap  
308 with the one of the P11 core that they overgrow (Figures 10a-b and 11a-b). In some cases, all the new  
309 grains inherit the crystallographic orientation of the P11 core (Figure 10a-b), whereas in other cases,  
310 the overlap is more limited and the new grains show a significant dispersion of crystallographic  
311 directions with respect to the P11 cores (Figure 11). The crystallographic inheritance is more evident  
312 if we consider the crystallographic orientation of the P12-P13 grains in direct contact with the P11 core  
313 (Figures 10a and 11b). The P12-P13 grains dispersed in the mylonitic matrix show a wide range of  
314 crystallographic orientations, which do not necessarily overlap with those of the P11 cores included  
315 in the map (Figures 10b and 11a). EBSD analysis also shows that the most common fracture set in  
316 the P11 porphyroclasts is parallel to the (001) plane, which is a perfect cleavage plane in plagioclase  
317 (Figure 11a).

318 Amphibole has a strong CPO, with the (100) and [001] subparallel to the foliation and to the stretching  
319 lineation, respectively (Figures 10c and 11c). The misorientation angle distribution of amphibole  
320 displays the strongest peaks between  $2^\circ$  and  $25^\circ$  and around  $50^\circ$  and  $180^\circ$ , with higher values for the  
321 correlated pairs (Figure 11d). The misorientation axes of amphibole, plotted in crystal coordinates,  
322 show maxima around the c-axis for misorientations up to  $30^\circ$ , which, in sample coordinates, is  
323 oriented subparallel to the stretching lineation of the mylonite (Figure 11d).

## 324 5. DISCUSSION

### 325 5.1 P-T conditions of metamorphism and deformation

326 The Anderson and Smith (1995) geobarometer calibration was based on rocks that equilibrated in the  
327 presence of melt. There is no evidence of melt in our sample, therefore we favour the pressure results  
328 from Bhadra and Bhattacharya (2007; Table 4). However, we kept both results, as the Plagioclase–  
329 Hornblende Thermobarometry spreadsheet (Anderson et al., 2008) that we used to compute P and T  
330 utilises a convergence between the results of Anderson and Smith (1995) geobarometer and Holland  
331 and Blundy (1994) geothermometer. As a test, we used the Bhadra and Bhattacharya (2007) P results  
332 as input for the Holland and Blundy (1994) geothermometer and we obtained coincident T results,  
333 within error.

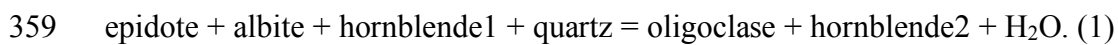
334 The oldest metamorphic stage preserved in the sample is represented by the albite cores (Pl1) with  
335 their oriented inclusions of epidote, ilmenite and rutile. In particular, rutile inclusions suggest  
336 relatively high pressure, but the coexistence with albite crystals would constrain the maximum  
337 pressure of this metamorphic stage below the albite breakdown reaction (albite = jadeite + quartz;  
338 Newton & Smith, 1967). However, the exact conditions of this first metamorphic stage were not  
339 constrained in this study.

340 The microstructural relationships between Pl1 and Amp1 are equivocal, therefore it is difficult to  
341 prove that they grew in equilibrium. Their compositions suggest that had they grown in equilibrium,  
342 they would have crystallised at 400°C (Table 4). These results, however, are far away from the lower  
343 limits of the calibration of Bhadra and Bhattacharya (2007; 650–950°C) and Anderson and Smith  
344 (1995; 675–760 °C): for this reason, pressures were not computed for the Amp1-Pl1 pairs.  
345 Furthermore, Pl1 is out of the compositional range of the Bhadra and Bhattacharya (2007)  
346 equilibration (Table 10 of that contribution).

347 Within uncertainty, the T results of Amp1-Pl2 and Amp2-Pl3 pairs lie just within the lower limits of  
348 the two calibrations. The T and P estimates from the Amp1-Pl2 and Amp2-Pl3 pairs are coincident  
349 within error ( $\pm 50$  °C and  $\pm 0.2$  GPa). This is due to the subtle chemical differences existing between  
350 these two generations. Thermobarometric calculations suggest that Pl2 may have grown

synchronously with Amp1 at 615°C and 0.97 GPa. The crystallisation of Pl2 marks the first stage of the mylonitic foliation development, as evidenced by the presence of Pl2 in the pressure shadows of Pl1 (Figures 3, 5a and 7). The mylonitic foliation continued to develop under conditions of ~ 600°C and 0.75 GPa (Amp2-Pl3 pair), as supported by the microtextural observation of Amp2, Pl3 and Chl1 intergrowths within the Pl1 pressure shadows (Section 4.1). This T result is coincident, within error, with the Chlorite+Quartz+H<sub>2</sub>O thermometry results for Chl1 (550°C; higher T peak in Figure 6a).

The XAn increase from Pl1 to Pl2 and Pl3 could reflect the following reaction from Apter and Liou (1983) for P of 0.7 GPa:



The abundant Ep inclusions in Pl1 and its scarcity and corroded aspect in the matrix of the sample would support this hypothesis. Pl2 likely crystallized at temperatures >600°C in the (High-T) Ep-amphibolite facies field (Apter & Liou, 1983; Liou, Kuniyoshi, & Ito, 1974; Miyashiro, 1968; Winkler, 1980), and Pl3 at lower temperatures, as the associated Amp2 and Chl1 limit the maximum T to 550-600°C (Figures 5 and 6a). A final retrograde stage is recorded by the Chl2 compositions suggesting a T range of 350°C-200°C (using the Chlorite+Quartz+H<sub>2</sub>O thermometry, Figure 6b), and by the overprinted Chl1 values yielding the low T peak of figure 6a. Indeed, Chl2 appears to be associated with the development of C' shear bands overprinting the mylonitic foliation.

## **5.2 The origin of luminescence in plagioclase**

BSE and CL images provide important insight into feldspar textures (e.g. Lee, Parsons, Edwards, & Martin, 2007; Parsons & Lee, 2009; Parsons, Steele, Lee, & Magee, 2008). Plagioclase luminescence has been related to many causes (summarized in Götze, 2012): the Mn<sup>2+</sup>, Ti, Fe<sup>3+</sup> content (Götze, Habermann, Kempe, Neuser, & Richter, 1999; Mariano & King, 1975), trace and REE elements (Götze, Habermann, Neuser, & Richter, 1999; Mariano & King, 1975), lattice defects (e.g. Al-O-Al bridge; Finch & Klein, 1999) and/or mineral inclusions (Smith & Stenstrom, 1965). The CL signal is

375 most commonly linked to Ti concentrations (Lee et al., 2007; Parsons et al., 2008), with a contribution  
376 from Fe (Lee et al., 2007). Furthermore, Parsons et al. (2008) has noted that CL emissions appear to  
377 be linked to Ca zoning, but they did not find a direct link between Ca concentration (or other trace  
378 element concentrations) and CL intensity.

379 In our samples, TiO<sub>2</sub> concentrations appear uniform across all Pl generations (with a content of  
380 0.01wt%; Table 2). This is also the case for the MnO concentrations, with a scatter between 0.002  
381 and 0.006 wt%. The FeO (Fe total) concentrations in Pl1 and Pl2 are the same (0.03 wt%; Table 2),  
382 but increases in Pl3 (0.06 wt%; Table 2). The CL signal in our samples appears to be related to the  
383 Ca\Na ratio (with maybe some contribution from Fe): the higher the ratio the brighter the growth  
384 zone, as visible from the perfect match between the SEM-CL and EPMA compositional maps  
385 (Figures 3 and 7), even though we cannot rule out a contribution from trace and REE elements or  
386 lattice defects. Luminescence induced by calcite inclusions in PL2 and 3 can be excluded, as the  
387 calcite veins in the sample do not luminesce.

### 388 **5.3 Replacement reactions by coupled dissolution-precipitation and deformation by** 389 **dissolution-precipitation creep in plagioclase and amphibole**

390 Our analysis highlights a strong correlation between CL images and both BSE images and EPMA  
391 compositional maps of plagioclase. Textural and chemical features similar to those reported in this  
392 study are recognised in several amphibolites present over > 500 m of the COSC-1 core (from ~1600  
393 to ~2300 m deep), implying that the deformation mechanisms and the mineral replacement reactions  
394 discussed below are important for the development of the middle to lower crustal thrusts and  
395 associated tectonic transport during Caledonian nappe stacking.

396 The plagioclase porphyrocrysts deformation history is summarized in Figure 12a. During the first  
397 stage Pl1 grew including a foliation defined by Qz, Ep, Ilm and Rt. Subsequently, fractures developed  
398 along the (001) cleavage planes (Ague, 1988; Brander et al., 2012; Brown & Macaudière, 1984;

399 McLaren & Pryer, 2001) enhanced fluid infiltration that triggered replacement reactions occurring  
400 via coupled dissolution of P11 and precipitation of P12 (Brander et al., 2012; Marti, Stünitz,  
401 Heilbronner, Plümper, & Drury, 2017) in equilibrium with the changed P and T conditions of 615°  
402 C, 0.9 GPa. These replacement processes continued precipitating P13, which rimmed P11 and P12 and  
403 sealed the fractures in P11. The slightly different chemistry of P13 compared to P12 and the  
404 microstructural observation that P13 overgrows P12 suggest that these two growth zones were closely  
405 related in time or crystallisation reaction (Section 4.2).

406 The growth zones of P12 and P13 inherit the crystallographic orientation of the P11 cores, with  
407 maximum misorientation of 5° (Figure 7c-d). Thus, the precipitation of P12 and P13 on P11 is an  
408 example of pseudomorphic and topotaxial growth, as often observed during coupled dissolution-  
409 precipitation processes (Engvik et al., 2008; Hövelmann, Putnis, Geisler, Schmidt, & Golla-  
410 Schindler, 2010; Plümper et al., 2017; Putnis & Putnis, 2007; Spruzeniece, Piazzolo, & Maynard-  
411 Casely, 2017). The small (generally < 2°) and only local difference in crystallographic orientation  
412 between P11 and P12-P13 presumably results from the slightly different unit cell parameters between  
413 albite and oligoclase. The few low angle boundaries associated with P12-P13 growth zones are  
414 typically found along intracrystalline fractures (Figure 7d), and we interpret them as the evidence of  
415 slightly rotated fractured fragments of P11 that re-equilibrated to P12-P13 compositions via coupled  
416 dissolution-precipitation processes. Pore trails occur in the plagioclase porphyrocrysts parallel to the  
417 two fracture systems, together with several micrometric mineral inclusions resulting in a turbid aspect  
418 of the plagioclase (Figures 3a-b and 4). Those trails probably represented the main pathways of fluid  
419 circulation during coupled P11 dissolution and P12 and P13 precipitation. These mineral inclusions  
420 may represent the signature of transient porosity during plagioclase replacement reactions (e.g.  
421 Plümper et al., 2017; Plümper & Putnis, 2009; Putnis, 2015; Walker, Lee, & Parsons, 1995).

422 Crystallographic continuity indicative of epitaxial overgrowth is also observed between plagioclase  
423 porphyrocrysts and P12 and P13 grains around them, but to different extents (compare Figure 10a with

424 Figure 11b). Thus, the crystallographic orientation of new Pl2 and Pl3 grains may be inherited from  
425 the Pl1 parent grain due to epitaxy, as described for several minerals deforming by dissolution-  
426 precipitation creep at different crustal levels (Engvik et al., 2008; Imon et al., 2002; Imon et al., 2004;  
427 Jiang, Prior, & Wheeler, 2000; Mukai et al., 2014; Spruzeniec et al., 2017; Wassmann & Stöckhert,  
428 2012; Wassmann et al., 2011). The crystallographic continuity seems to decrease away from the  
429 plagioclase porphyrocrysts (Figures 10b and 11a), presumably reflecting the heterogeneous  
430 nucleation of Pl2 and Pl3 neoblasts in the surrounding matrix together with Amp2 and Chl1.  
431 Additionally, these neoblasts may have also undergone grain-boundary sliding during deformation,  
432 given their fine grain size, which can further disperse their inherited crystallographic orientation  
433 (Okudaira et al., 2017).

434 Thus, we conclude that metamorphism of plagioclase at  $\sim 600^{\circ}\text{C}$  and 0.75-0.9 GPa occurred by  
435 coupled dissolution-precipitation processes with pseudomorphic and topotaxial replacement of Pl1  
436 by Pl2 and Pl3, and deformation was accommodated by dissolution-precipitation creep with  
437 nucleation of Pl2-Pl3 grains around the plagioclase porphyrocrysts (epitaxial) and in the surrounding  
438 matrix. It is worth noting that the replacement of Pl1 porphyroclasts by coupled dissolution-  
439 precipitation processes generally occurred concentrically (Figure 7). Consequently, at least part of  
440 this replacement process outlasted the deformation. The similar chemistry of Pl3 replacing Pl1 and  
441 the Pl3 neoblasts in the surrounding matrix suggest that there was no major change in metamorphic  
442 conditions throughout the timing of growth of Pl3.

443 No significant contribution of crystal plasticity was observed, although deformation occurred at PT  
444 conditions at which crystal plasticity in plagioclase is expected to occur (e.g. Gerald & Stünitz,  
445 1993; Pearce, Wheeler, & Prior, 2011). The porosity necessary to maintain fluid transport during  
446 coupled dissolution-precipitation was generated mostly by fracturing, as shown by the common  
447 occurrence of Pl2-Pl3 growth zones along fractures parallel to the (001) perfect cleavage planes.  
448 Finally, the sharp chemical transition (across  $\sim 1\text{-}5\text{ }\mu\text{m}$ ) between all the Pl generations, visible from



449 the EPMA and CL maps (Figures 3 and 7), suggest that chemical equilibration did not occur by solid-  
450 state diffusion (e.g. Hövelmann et al., 2010).

451 The textural features of amphibole suggest a similar deformation history to plagioclase (Figure 12b).  
452 Amp1 (high Mg<sup>#</sup>) grains are preserved mostly as relict cores and displays embayments, lobate edges  
453 and truncated chemical zoning patterns (Figures 3f and 9c; Bukovská, Wirth, & Morales, 2015;  
454 Gratier et al., 2013; Hyppolito, García-Casco, Juliani, Meira, & Hall, 2014; Passchier & Trouw, 1996;  
455 Rutter, 1983; Stokes et al., 2012; Wassmann & Stöckhert, 2013; Wintsch & Yi, 2002). These textural  
456 features suggest that Amp1 underwent coupled dissolution and Amp2 precipitated on Amp1. As for  
457 plagioclase, the growth of Amp2 on Amp1 was pseudomorphic and topotaxial (Figure 9d-f). Amp2  
458 grew also as smaller neoblasts (maximum few hundreds of  $\mu\text{m}$  in size), elongated parallel to the  
459 mylonitic foliation and preferentially elongated parallel to their c-axis, due to dissolution-  
460 precipitation creep. These crystals do not display Amp1 cores in the compositional maps (Figures 3f  
461 and 5b).

462 The CPO and shape-preferred orientation of the amphibole can be acquired via different mechanisms:  
463 dissolution-precipitation creep (Bons & den Brok, 2000; Imon et al., 2004; Pearce, Wheeler, & Prior,  
464 2011), oriented grain growth and passive rotation after growth (Berger & Stünitz, 1996; Kanagawa,  
465 Shimano, & Hiroi, 2008) and/or diffusion creep (Getsinger & Hirth, 2014). In the studied sample the  
466 CPO displayed by Amp2 is mostly inherited due to the pseudomorphic and topotaxial growth on  
467 Amp1. The small misorientations ( $< 3^\circ$ ) evident in the EBSD map (Figure 9f) are attributed to  
468 fractures that developed perpendicular to the crystal elongation (Figure 9a). Some amphibole crystals  
469 display more complex zoning (e.g. central crystal in Figure 9c), probably due to a preferential  
470 replacement of the central part of the crystal (e.g. Hyppolito et al., 2014). The CPO of Amp1 grains  
471 was presumably formed via oriented grain growth during an earlier deformation event. Moreover, as  
472 presented in section 4.6, the misorientation axes of amphibole show maxima around the c-axis, which  
473 is oriented subparallel to the stretching lineation of the mylonite (Fig. 11d). This geometry is not

474 consistent with dislocation creep on the prism  $\langle c \rangle$  slips system of amphibole, as the misorientation  
475 axis cannot coincide with the Burgers vector (e.g. Kruse, Stünitz, & Kunze, 2001; Lloyd, Farmer, &  
476 Mainprice, 1997). Instead, we interpret the cluster of misorientation axes around  $\langle c \rangle$  as the evidence  
477 of topotaxial growth of elongated amphibole grains that preferentially share their c-axis. As the data  
478 come from all the grains included in the map (and not only from the rims of Amp2), we think that  
479 oriented growth was the dominant deformation mechanisms of amphibole throughout the deformation  
480 history.

481 In the studied sample amphibole, like plagioclase, shows no evidence of deformation via crystal  
482 plasticity, such as high intracrystalline misorientations, misorientation bands, subgrains. Instead it  
483 appears to have deformed by fracturing and coupled dissolution-precipitation, as also suggested in  
484 other studies (Berger & Stünitz, 1996; Brodie & Rutter, 1985; Lafrance & Vernon, 1993; Nyman,  
485 Law, & Smelik, 1992; Pearce et al., 2011). Crystal plasticity is potentially a more effective  
486 deformation mechanism at higher temperatures (e.g. Skrotzki, 1992).

487 It is worth noting that the synkinematic reaction that produced Pl2-Pl3 and Amp2 (reaction 1) could  
488 have been a dehydration reaction. If this was the case, the aqueous fluid necessary to sustain coupled  
489 dissolution-precipitation processes did not necessarily infiltrate from an external source, but may have  
490 been released internally. The role of dehydration reactions in the rheological evolution of crustal  
491 rocks has received little attention so far, as reaction weakening is commonly associated with hydration  
492 reactions during retrogression (e.g. Gueydan, Leroy, Jolivet, & Agard, 2003). However, dehydration  
493 reactions that release fluids at grain boundaries can also potentially result in weakening and strain  
494 localization during burial and nappe stacking, if the released fluids facilitate the activation of coupled  
495 dissolution-precipitation creep and of diffusion creep.

## 496 **6. CONCLUSIONS**

497 Middle to lower crustal mylonites from the COSC-1 drill core (Lower Seve Nappe) were investigated  
498 with EPMA compositional maps, CL images and EBSD maps to constrain the mechanism(s)  
499 responsible for their formation. The data suggest that fracturing, coupled dissolution-precipitation  
500 and dissolution-precipitation creep were responsible for the development of the mylonitic fabric in  
501 amphibolites at conditions of  $\sim 600^{\circ}\text{C}$ , 0.75-0.97 GPa, in the epidote-amphibolite facies, over a  
502 thickness of  $> 500$  m. No evidence of deformation via crystal plasticity is present in either plagioclase  
503 or amphibole in the analysed sample, even though deformation occurred at pressure and temperature  
504 conditions at which plagioclase is expected to deform by dislocation creep. The presence of  $\text{H}_2\text{O}$ -rich  
505 fluid at the grain boundaries appears to have enhanced replacement reactions and facilitated  
506 dissolution and precipitation processes, which in turn considerably decreased the strength of this  
507 middle to lower crustal shear zone. Importantly, replacement reactions of plagioclase only occurred  
508 by coupled dissolution-precipitation at grain boundaries and along fractures, and were otherwise  
509 sluggish.

510 Our study shows that crystallographic preferred orientation in plagioclase and amphibole can be  
511 inherited from parental grains due to pseudomorphic and topotaxial growth during coupled  
512 dissolution-precipitation during deformation. Thus, care must be taken when considering CPOs in  
513 deformed rocks as evidence of deformation by dislocation creep.

514 The development of a mylonitic fabric by coupled dissolution-precipitation and dissolution-  
515 precipitation creep in amphibolites over a thickness  $> 500$  in the Lower Seve Nappe suggests that the  
516 strength of amphibolites can be significantly low in the presence of grain-boundary aqueous fluid  
517 during nappe thrusting in the middle to lower crust.

518

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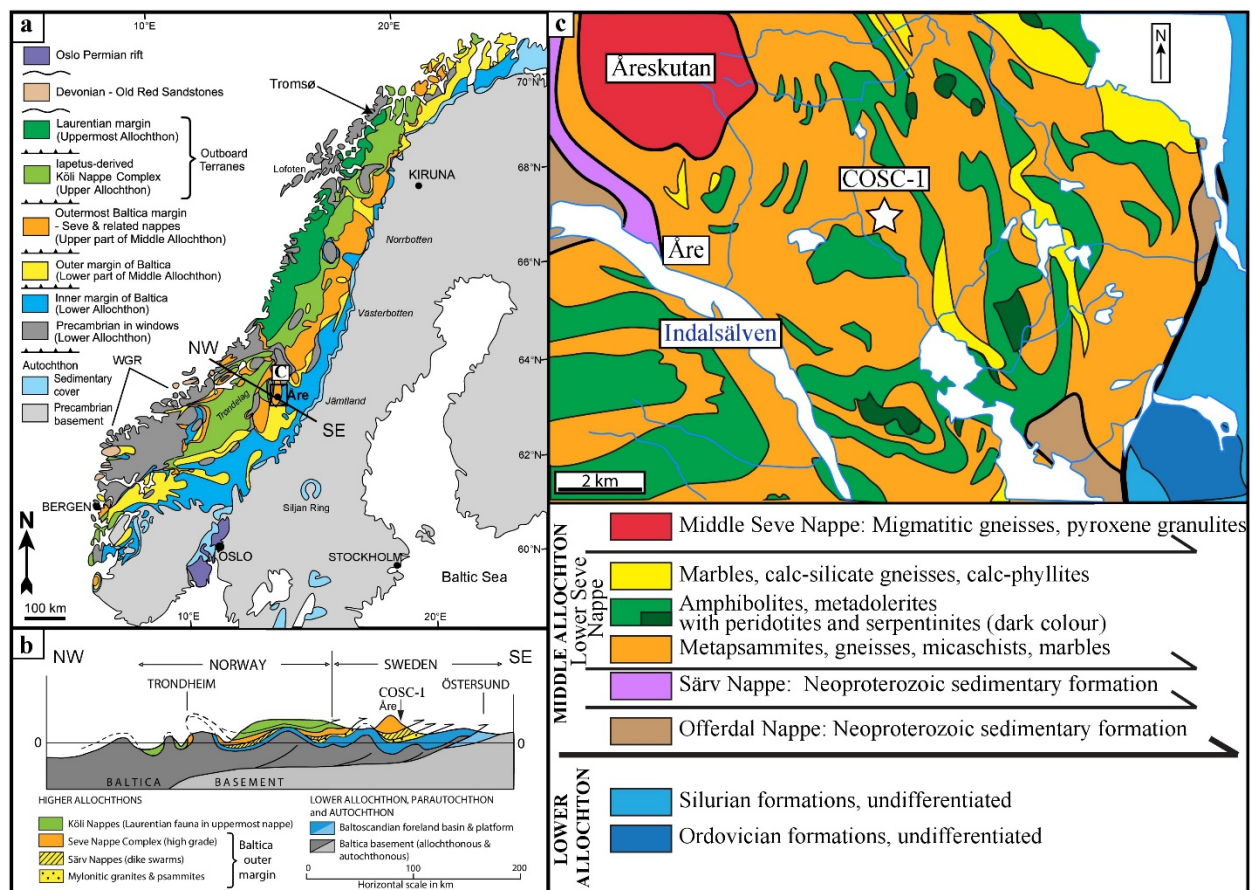
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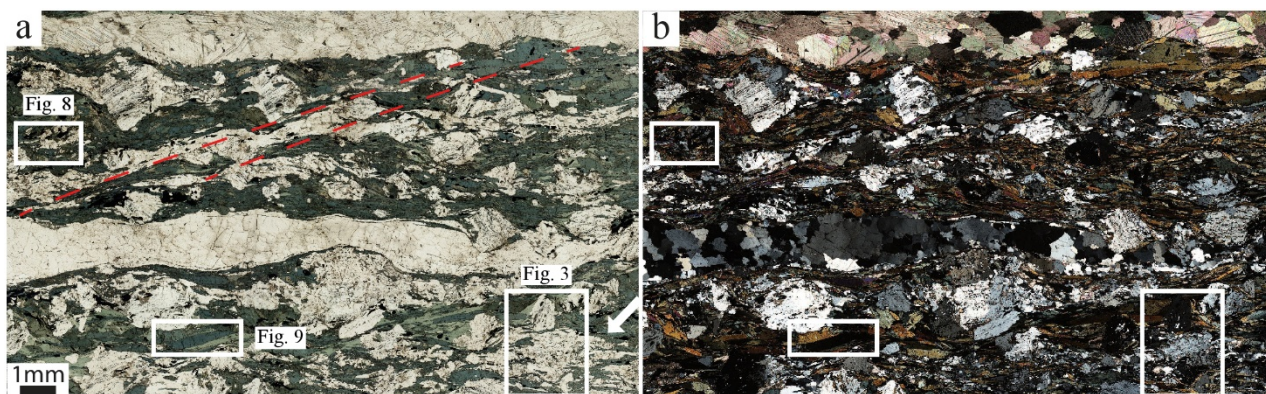
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855

856 **FIGURE 1** Geological setting of the Scandinavian Caledonides. (a) Tectonic map with inferred  
 857 paleogeography of the nappes (modified after Gee et al., 2010). (b) Cross section marked in (a) with  
 858 vertical exaggeration of 5 x and approximate location of the COSC-1 borehole (modified after Gee  
 859 et al., 2010). (c) Detail of the study area with location of the COSC-1 drilling site (modified after  
 860 Strömberg et al., 1984).

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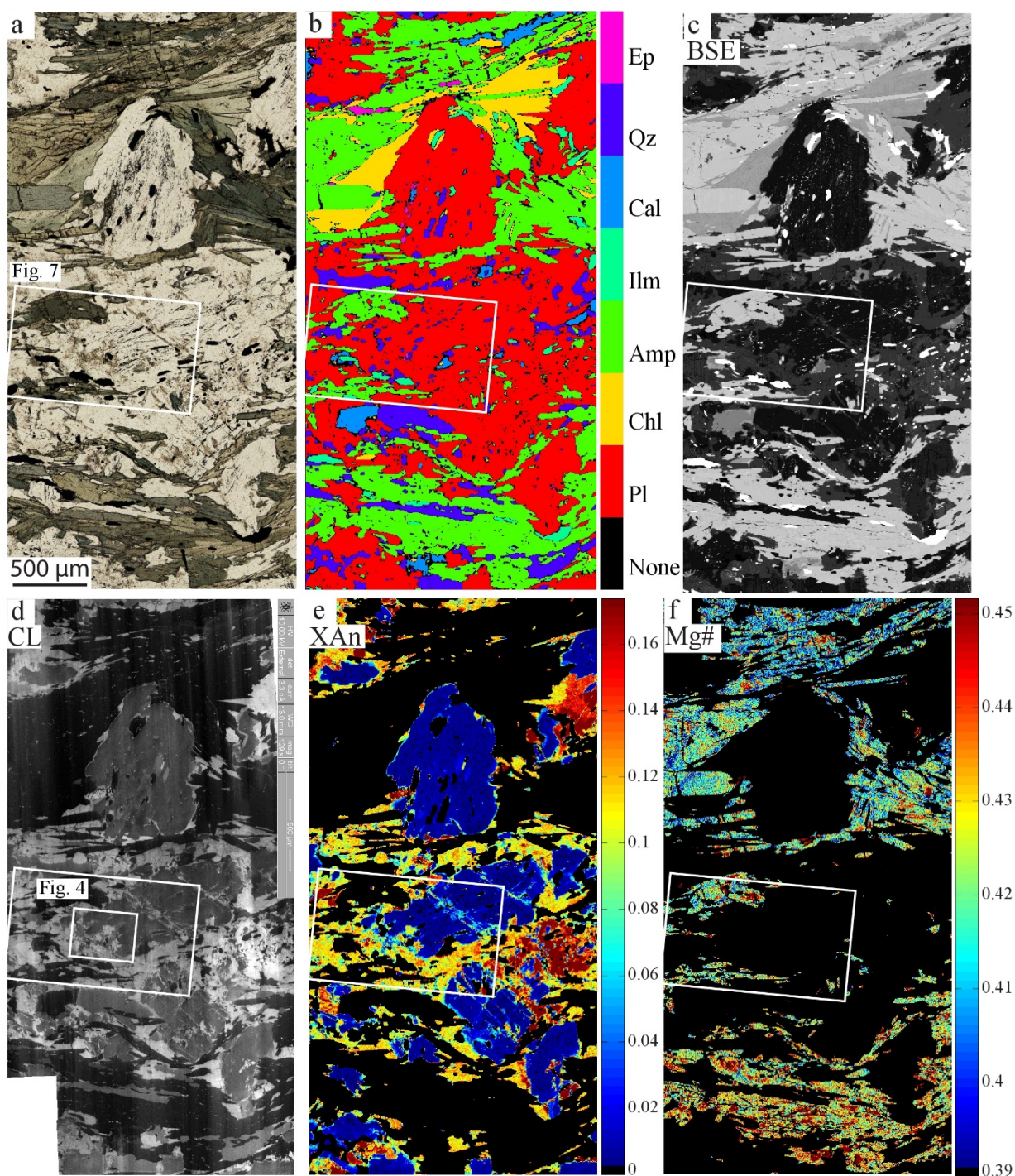


**FIGURE 2** Amphibolite with Pl porphyroclast displaying a mylonitic foliation defined by Amp, Pl, Chl, Qz, Ep and Ilm. Cal and Qz ribbons, presumably representing transposed veins, are parallel to the foliation. Both dextral- (clasts close to Figure 3 rectangle, marked by the white arrow) and sinistral sense of shear (C' planes, red dashed line) are visible. (a): *plane-polarized light*; (b): *crossed-polarized light*. Mineral abbreviation from Whitney and Evans (2010).

Minerals	Pre-main foliation	Main foliation	C' Planes
Amp	-----	————	
Pl	Ab Core ————	Ab/Olig Rims ————	
Chl		————	-----
Quartz	————	————	-----
Cal	-----	————	-----
Epidote	————	-----	
Ilmenite	————	————	

**TABLE 1:** Metamorphic and deformation evolution of sample ICDP5054EX8E601.



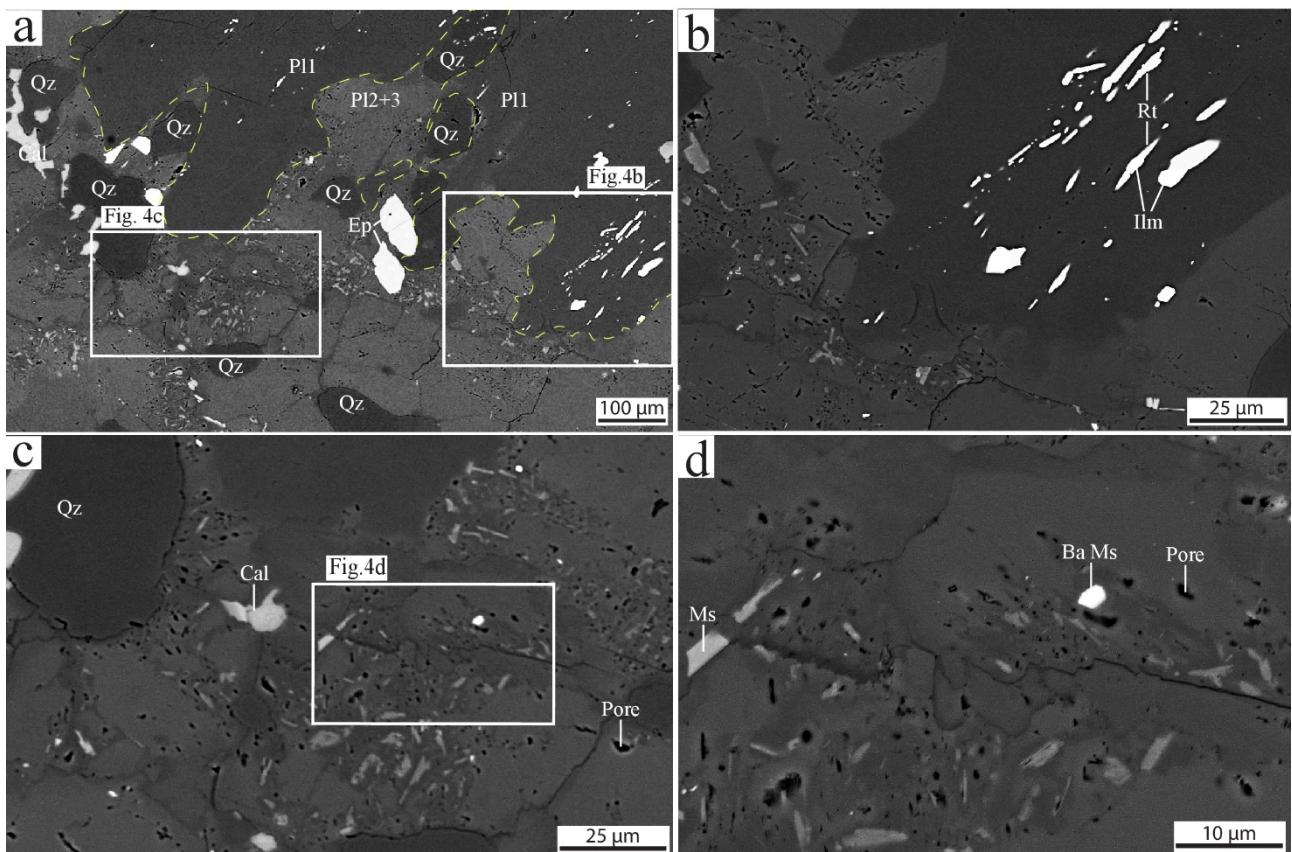


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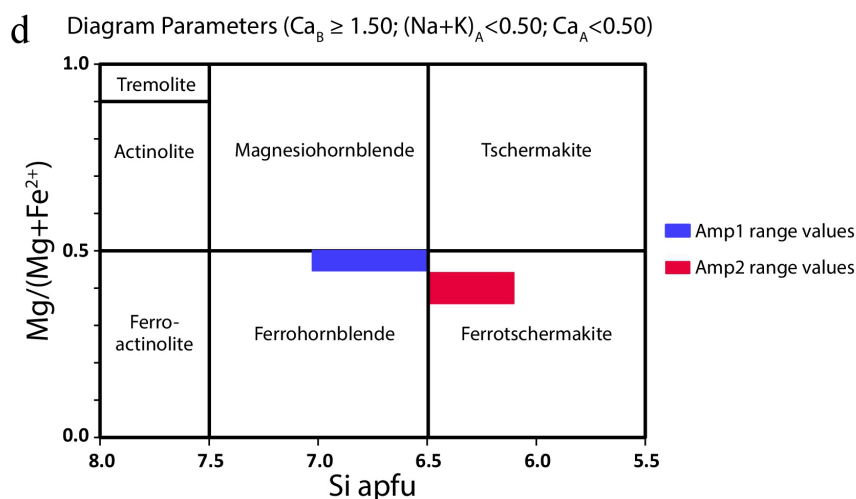
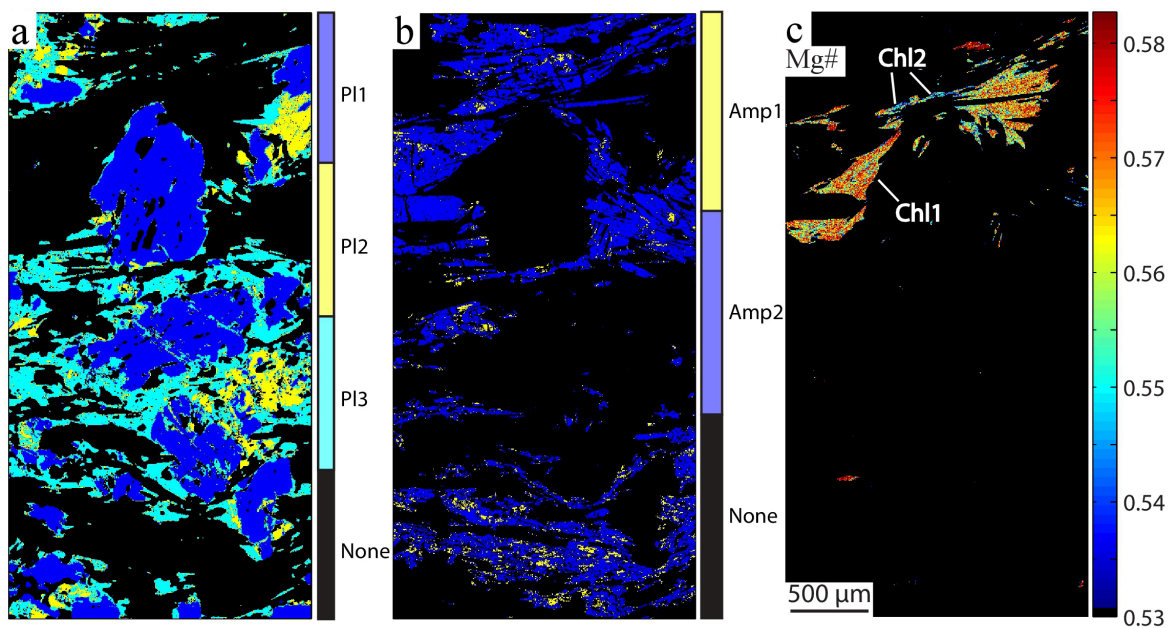
871 **FIGURE 3** More detailed areas of sections shown in Figure 2. (a) Ab porphyroclasts with dark trails  
 872 of inclusions of Ep and Cal few  $\mu\text{m}$  in size. The Ab porphyroclasts are wrapped by the mylonitic  
 873 foliation defined by Amp, Olig and Chl. The white rectangle indicates the site of the EBSD map  
 874 shown in Figure 7 (*plane polarized light*). (b) X-ray map showing Pl, Amp, Chl and Ep that crystallize  
 875 in the pressure shadows of Ab porphyroclasts. (c) BSE image showing zoned Pl with dark cores (P11)  
 876 and brighter rims (P12 and P13). Bright inclusions inside Pl are Ilm crystals. (d) CL image highlighting



877 the difference between Ab cores (dark, P11) rimmed by Pl with higher An content (bright, P12 and  
 878 P13). This bright Pl crystallises also in the pressure shadows and in the fractures that dissect the Ab  
 879 cores. (e) Standardized X-Ray map of the X Anorthite (XAn) in the Pl. Note the fractures in P11  
 880 sealed by P13 and sheared-off fragments of P11 rimmed by P12 and 3 (see text and Figure 5 for  
 881 distinction). (f) Standardized X-Ray map of Amp crystals displaying relic cores higher in Mg#  
 882 (Amp1) and rims lower in Mg# lengthened as the main foliation (Amp2). A minor but consistent shift  
 883 toward higher Mg# is visible from the top to the bottom of the picture, probably related to an  
 884 analytical artefact.



886 **FIGURE 4** Details of the microstructures shown in Figure 3c. BSE images. (a) Dark P11, highlighted  
 887 by the dashed yellow line, with lobate edges and peninsular features rimmed by P12 and P13. (b) Iso-  
 888 oriented Ilm and Rt inclusions in P11. Note the two perpendicular trails of pores in P12 and P13. (c)  
 889 Detail of the pore trails and associated Ms and Cal inclusions. (d) Close-up of (c) highlighting pores,  
 890 abundant Ms inclusions and rare Ba-rich Ms.



**FIGURE 5** Mineral chemistry of Pl, Amp and Chl. The subdivision in generations was made based on microtextures. (a) Pl groups based on the XAb and XAn content: PI1 XAn 0-0.05; PI2 XAn 0.05-0.13; PI3 XAn 0.13-0.25. Compare with Figure 3e. (b) Amp groups based on the Mg# and Si apfu content: Amp1 Mg# 0.5-0.44, Si apfu 7-6.5; Amp2 Mg# 0.44-0.36, Si apfu 6.5-6.1. Compare with Figure 3f. (c) Compositional map of Chl Mg# for 0% Fe<sup>3+</sup>. The crystals display homogeneous compositions (Chl1 Mg# 0.59-0.56), except at grain boundaries and along C' band where lower values are present (Chl2 Mg# 0.55-0.52) (d) Diagram of classification of calcic amphiboles with plotted the average chemical compositions of Amp1 and Amp2 (from Leake et al., 1997).

	Pl			Amp	
	Core (Pl1)	Rim1 (Pl2)	Rim2 (Pl3)	Core (Amp1)	Rim (Amp2)
SiO <sub>2</sub>	68.21	62.95	64.99	43.52	41.62
TiO <sub>2</sub>	0.01	0.01	0.01	0.22	0.35
Al <sub>2</sub> O <sub>3</sub>	19.08	24.37	21.55	13.21	15.42
FeO	0.03	0.03	0.06	18.41	18.56
MnO	0.00	0.00	0.00	0.17	0.17
MgO	0.00	0.00	0.00	8.34	7.37
CaO	0.11	4.30	2.30	11.42	10.72
Na <sub>2</sub> O	12.34	8.93	10.47	1.72	2.02
K <sub>2</sub> O	0.04	0.05	0.05	0.21	0.35
Sum	99.83	100.64	99.43	97.22	96.57
	Formulae based on 8 O			Formulae based on 23 anhydrous O	
Si	2.99	2.76	2.88	6.49	6.26
Ti	-	-	-	0.02	0.04
Al	0.99	1.26	1.12	2.33	2.73
Fe <sup>3+</sup>	-	-	-	0.41	0.49
Fe <sup>2+</sup>	-	-	-	1.89	1.85
Mn	-	-	-	0.02	0.02
Mg	-	-	-	1.86	1.65
Ca	0.01	0.20	0.11	1.83	1.73
Na	1.05	0.76	0.90	0.50	0.59
K	0.00	0.00	0.00	0.04	0.07
Sum	5.04	4.99	5.01	15.38	15.42
Mg#	-	-	-	0.45	0.41
XAn	0.01	0.21	0.11	-	-
XAb	0.99	0.79	0.89	-	-

901 **TABLE 2** Representative average composition analysis (wt%) of Pl and Amp.

902

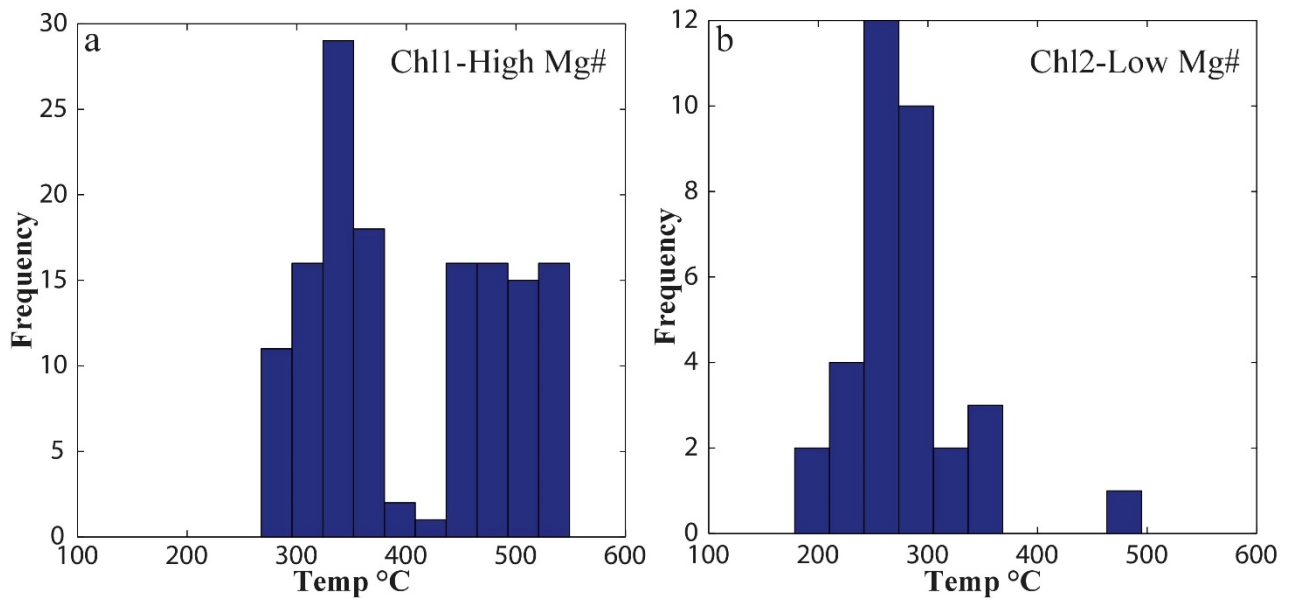


Chl	Chl1-High Mg#		Chl2-Low Mg#	
	0% XFe <sup>3+</sup>	30% XFe <sup>3+</sup>	0% XFe <sup>3+</sup>	30% XFe <sup>3+</sup>
SiO <sub>2</sub>	25.70	25.70	26.52	26.69
Al <sub>2</sub> O <sub>3</sub>	22.33	22.34	21.70	21.69
FeO	22.00	21.98	22.60	22.41
Fe <sub>2</sub> O <sub>3</sub>	-	0.02	-	0.21
MnO	0.15	0.15	0.15	0.14
MgO	16.43	16.43	15.37	15.42
CaO	0.02	0.02	0.03	0.03
Na <sub>2</sub> O	0.01	0.01	0.01	0.02
K <sub>2</sub> O	0.00	0.00	0.01	0.01
Sum	86.63	86.65	86.38	86.62
<b>Formulae based on 14 anhydrous O</b>				
Si	2.69	2.63	2.78	2.74
Al	2.75	2.70	2.68	2.62
Mg	2.56	2.51	2.41	2.36
Fe <sup>3+</sup>	0.00	0.57	0.00	0.58
Fe <sup>2+</sup>	1.92	1.32	1.98	1.35
Sum	9.92	9.73	9.85	9.65
Mg#	0.57	0.66	0.55	0.64

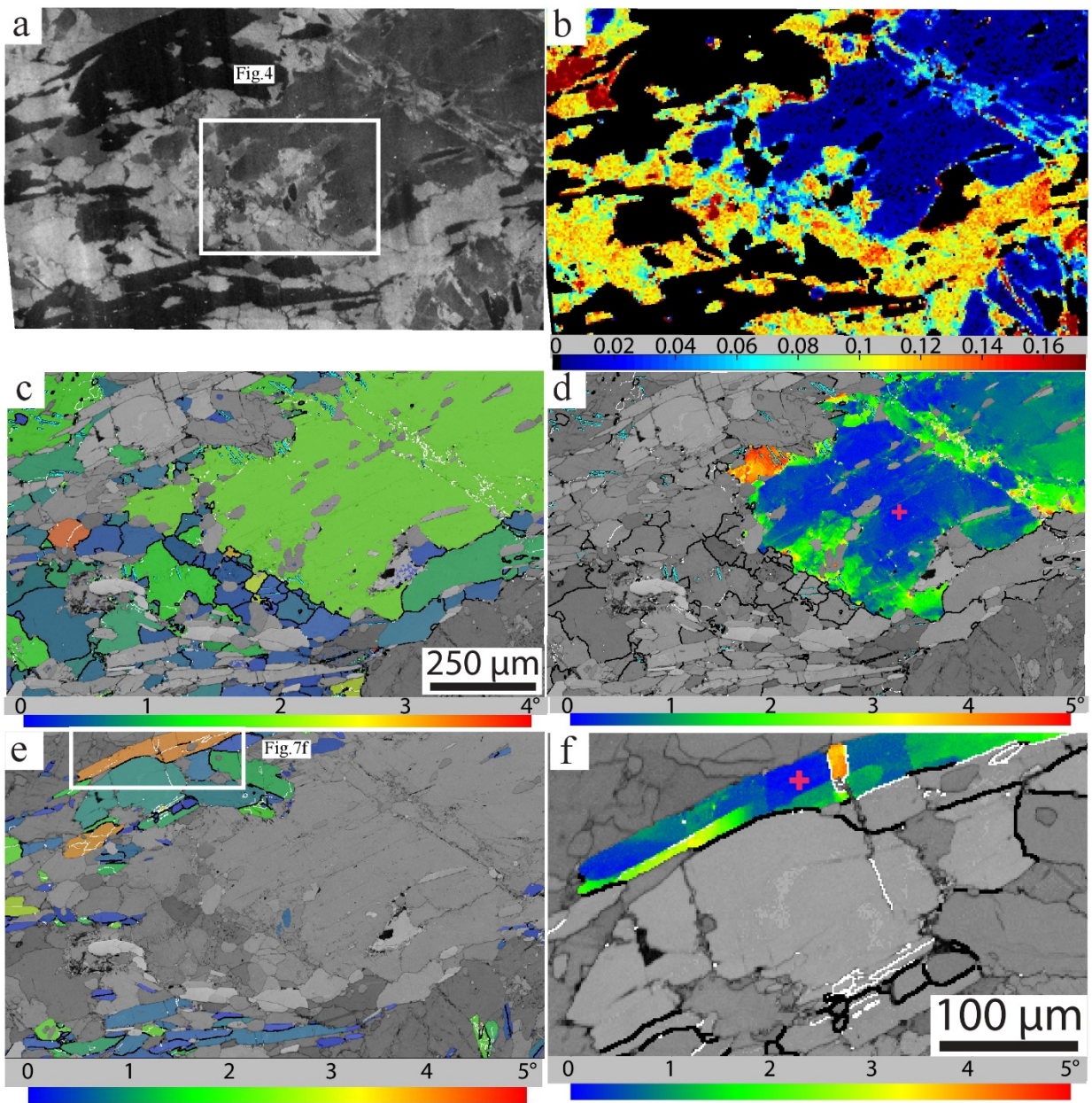
**TABLE 3** Representative average composition analysis (wt%) of Chl.

Amp-Pl couples	Thermometer	Barometer	
	HB	BB	AS
<b>Amp1-Pl1</b>	<b>392 °C</b>	-	-
<b>Amp1-Pl2</b>	<b>615 °C</b>	<b>0.97 GPa</b>	0.87 GPa
<b>Amp2-Pl3</b>	<b>605 °C</b>	<b>0.74 GPa</b>	1.09 GPa

**TABLE 4** Results of Amphibole-Plagioclase geothermobarometry computed from the values of Table 2. Thermometer abbreviation: HB: Holland and Blundy (1994). Barometer abbreviations: BB: Bhadra and Bhattacharya (2007); AS: Anderson and Smith (1995). The favoured results are highlighted in bold (see Section 5.1 for details).



**FIGURE 6** Chlorite+Quartz+H<sub>2</sub>O thermometry results calculated at a pressure of 0.7 GPa and a range between 0-50 % of Fe<sup>3+</sup>.

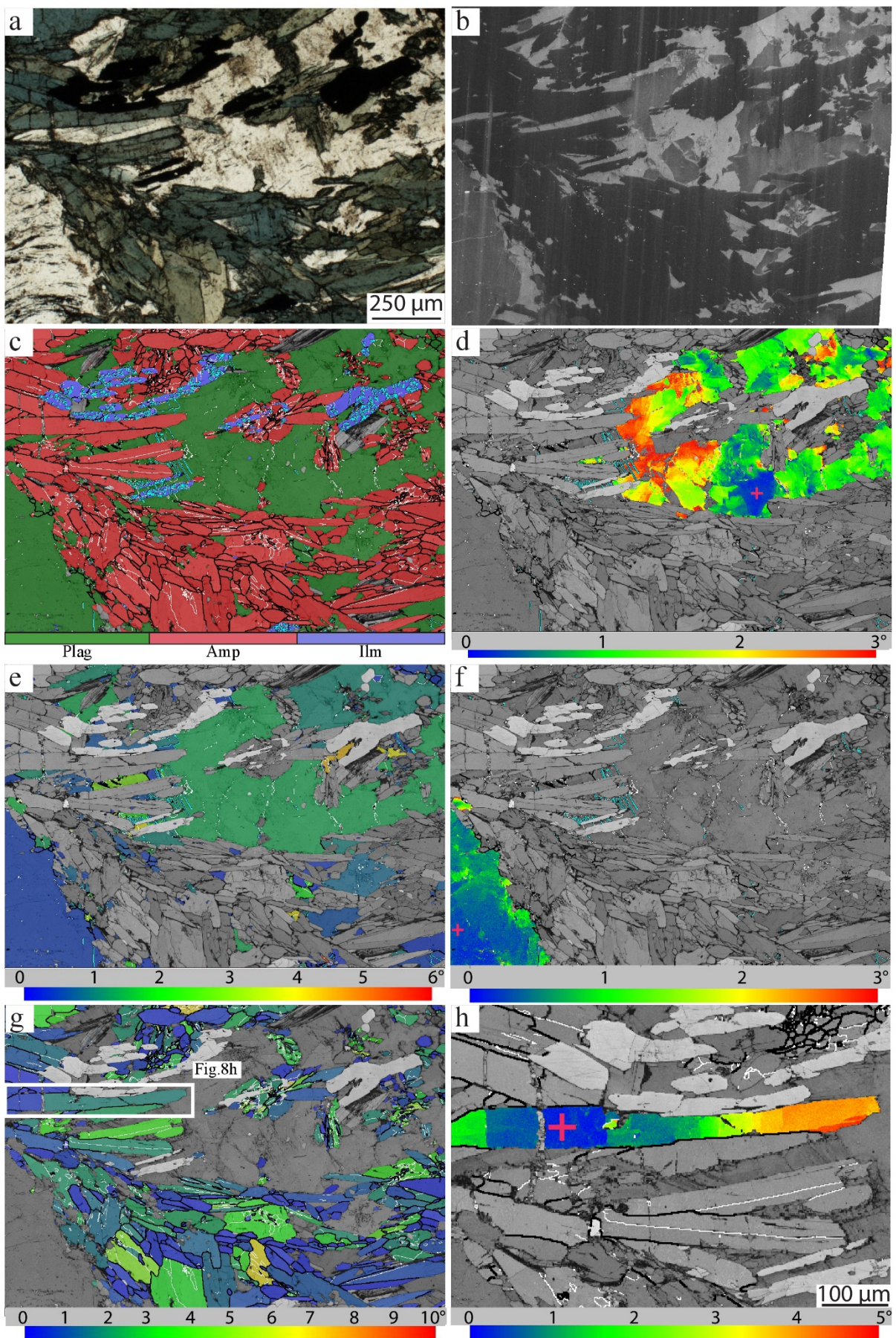


915

916 **FIGURE 7** Details of the microstructure shown in Figure 3 (referred to as domain 1 in the text). (a)  
 917 CL image of a dark P11 porphyroblast with fractures and rims containing the bright P12-3. (b)  
 918 Standardized X-Ray map of the XAn content showing the overlap between the bright CL areas and  
 919 the P12-P13 compositions. (c) EBSD GOS map superposed to the band contrast (BC) map of the same  
 920 area shown in (a). White lines: low angle boundaries 2-10°. Black lines: high angle boundaries > 10°.   
 921 Light blue lines: Twin boundaries in Pl. (d) EBSD texture component map (TCM) of the P11  
 922 porphyroblast, showing the misorientation from the reference point marked by the red cross. White  
 923 lines: low angle boundaries 2-10°. Black lines: high angle boundaries > 10°. Light blue lines: Twin

924 boundaries in Pl. (e) EBSD GOS of Amp from the area shown in (a). White lines: low angle  
925 boundaries 2-10°. Black lines: high angle boundaries > 10°. (f) EBSD TCM of Amp, showing the  
926 misorientation from the reference point marked by the red cross. White lines: low angle boundaries  
927 2-10°. Black lines: high angle boundaries > 10°.

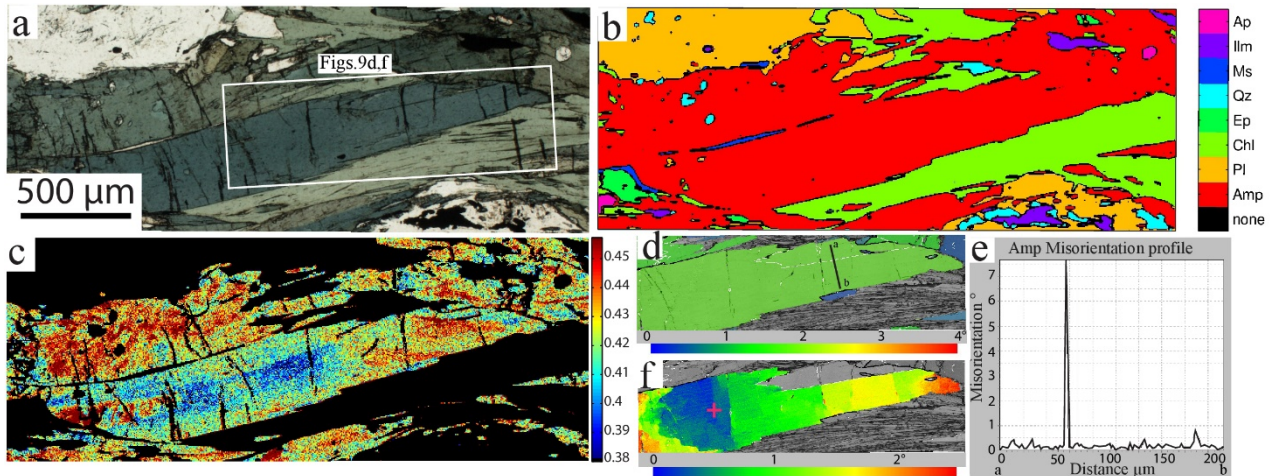






929 **FIGURE 8** Details of Figure 2 (referred to as domain 2 in the text). (a) Light microscopy  
 930 microstructure of the site analysed with EBSD. (b) CL image of the site shown in (a). Dark P11  
 931 porphyroclasts are overgrown by bright P12 and P13. (c) EBSD phases map of the site shown in (a).  
 932 Note the epitaxial growth of P12 and P13 on P11, as indicated by the lack of high angle boundaries  
 933 separating P11 from P12 and P13, and the very few low angle boundaries in the P11 porphyroclast. (d)  
 934 EBSD TCM showing the misorientation from reference point (red cross) in P11 porphyroclasts. Note  
 935 the correlation between higher misorientation and P12-P13 overgrowth on the right-hand side of the  
 936 reference point. (e) EBSD GOS map suggesting that the Pl crystals are very low internal strain. (f)  
 937 EBSD TCM showing the misorientation from reference point of another P11 porphyroclast. (g) EBSD  
 938 GOS map of Amp. (h) EBSD TCM showing the misorientation from reference point in an Amp  
 939 crystal elongate parallel to the foliation.

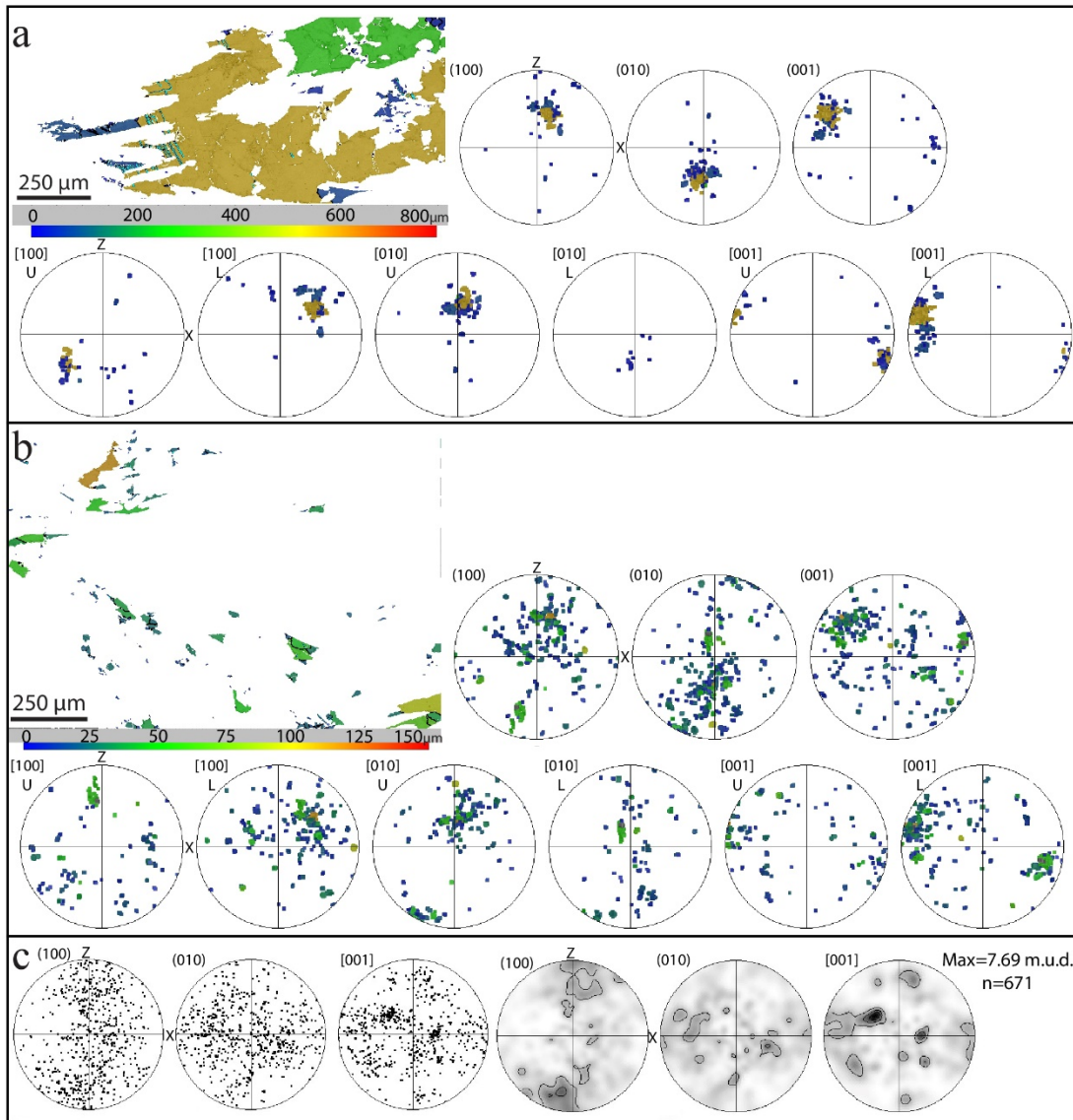
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941

942 **FIGURE 9** Details of Figure 2. (a) Sigmoidal grain of Amp surrounded by Chl along the foliation  
 943 wrapping around Pl porphyroclasts. Several brittle fractures perpendicular to the Amp elongation are  
 944 visible (*plane-polarized light*). The white rectangle encompasses the site of the EBSD maps shown  
 945 in (d) and (f). (b) X-ray map of the mineral phases in (a). (c) Standardized X-Ray map showing the  
 946 variation in Mg# of Amp grains (scale bar on the right-hand side). Amp displays cores higher in Mg#  
 947 (Amp1) and rims lower in Mg# (Amp2) elongate parallel to the foliation (compare with Figure 3f).

948 (d) EBSD GOS map of Amp. The black line is the trace of the misorientation profile a-b shown in  
 949 (e). See (a) for the location of the map. (e) Misorientation profile a-b drawn across a low-angle  
 950 boundary. (f) EBSD TCM showing the misorientation from the reference point marked by a red cross.  
 951



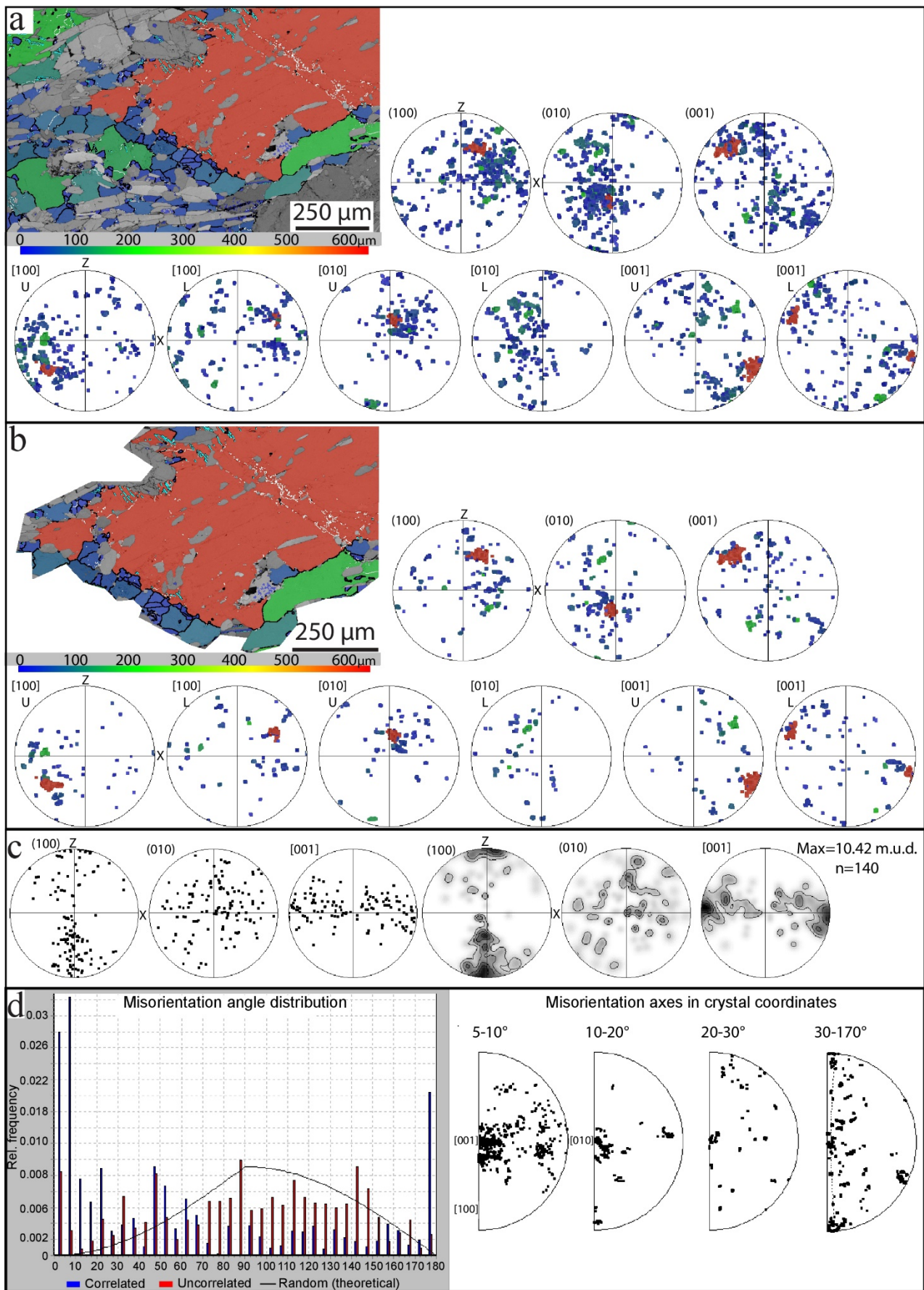
952

953 **FIGURE 10** Pole figures of the crystallographic orientation data of PI (colour-coding as in the grain  
 954 size maps) and Amp; same site as maps shown in Figure 8. X is the extensional instantaneous  
 955 stretching axis, Z is the pole of foliation, stereographic projections, lower hemisphere if not specified;  
 956 U: upper hemisphere, L: lower hemisphere. (a) Subset of the PI porphyroclast and adjacent grains  
 957 sharing the same crystallographic orientation, with the exception of a few data points. The PI displays

958 a CPO with the (100) and (001) maxima approximately at  $45^\circ$  from X. The (010) maximum is  
959 perpendicular to X. a: 189029 Pl data points. (b) Subset of the Pl crystals of the matrix (Pl2 and 3)  
960 not adjacent to the porphyroclast. A weak CPO similar to Figure 11a is visible. 34676 Pl data points.  
961 (c) Amp displaying a CPO with the (100) and [001] perpendicular and parallel to X, respectively. 671  
962 Amp data points (one-point-per-grain). n=number of grains. Half width  $10^\circ$  and cluster size  $5^\circ$ ,  
963 maximum value is given. Contouring is 1.

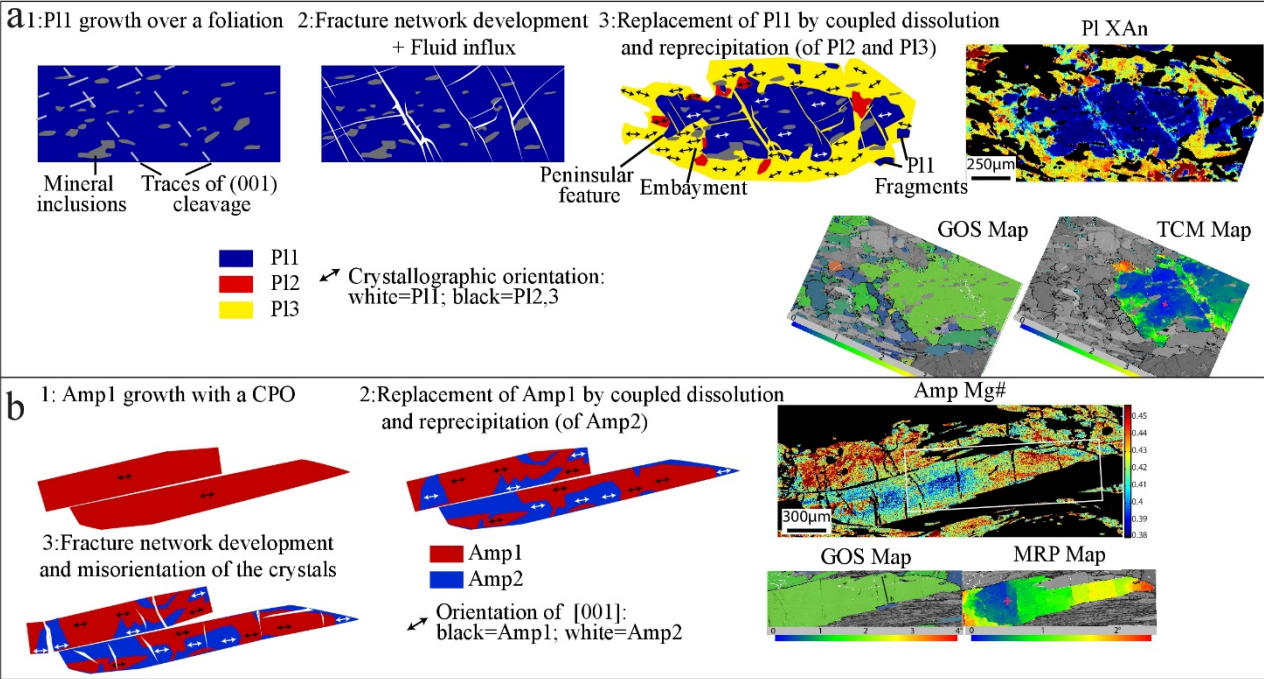
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966 **FIGURE 11** Pole figures of the crystallographic orientation data of Pl (colour-coding as in the grain  
 967 size maps) and Amp; same site as maps shown in Figure 7. (a) The Pl displays a CPO with the (100)  
 968 and (001) at low to medium angles to X. Note how the maximum of the values coincides with the  
 969 porphyroclast values, in red. 357353 Pl data points. (b) Subset of the Pl porphyroclast and adjacent  
 970 grains; some of the latter display a similar crystallographic orientation as the porphyroclast. 238333  
 971 Pl data points (c) Pole figures of Amp displaying a strong CPO with the (100) and [001] perpendicular  
 972 and parallel to X, respectively. 140 Amp data points (one-point-per-grain). n=number of grains. Half  
 973 width 10° and cluster size 5°, maximum value is given. Contouring is 2. (d) Histogram of distribution  
 974 of misorientation angles and misorientation axes of amphibole plotted in crystal coordinates. 3295  
 975 Amp data points. Dashed line: forbidden zone limit.

976



977

978 **FIGURE 12** Idealized sketch of the deformation history of Pl and Amp, colour-coding as in the  
 979 compositional maps. (a) 1: P11 porphyrocrysts grew including a foliation marked by Ep, Qz and Ilm.  
 980 2: Fracturing of the P11 porphyrocrysts, mostly imposed along the (001) plane. 3: Fluid infiltration  
 981 triggered mineral replacement by coupled dissolution-precipitation with topotaxial and

982 pseudomorphic growth of Pl2 and, successively, Pl3 on Pl1. Replacement occurred mostly along and  
983 in proximity of the fractures as well as at the edges of the crystals. Pl3 nucleated also as newly grown  
984 grains in the matrix. (b) 1: Amp1 grew with a CPO. 2: Replacement of Amp1 by Amp2 by coupled  
985 dissolution-precipitation; topotaxial growth is suggested by the same CPO shared by the two  
986 generations. 3: Development of misorientation due to displacement along the fractures and along the  
987 cleavage planes.