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# Seismic properties across an amphibolite- to greenschist-facies strain gradient (Neves area, eastern Alps, Italy): new considerations for shear zone imaging

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*Keywords: Seismic anisotropy, shear zone, EBSD, ultrasonic measurements*

*Highlights :*

1. Non-linear evolution of P- and S- wave anisotropy with ductile strain
2. Unexpected anisotropy destruction in mylonite due to a change in mica texture
3. Magnitude and geometry of seismic anisotropy controlled by metamorphic conditions
4. Bulk shear zone seismic properties controlled by the S plane-dominated margins

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## 1 ABSTRACT

2 Detecting the presence and geometry of crustal shear zones by geophysical methods relies on our  
3 understanding of the intrinsic parameters controlling the seismic properties of these deformed rocks,  
4 over the range of pressure-temperature conditions expected in the Earth's crust. To this end, we aimed  
5 to track changes in P-wave propagation velocity ( $V_P$ ), anisotropy ( $AV_P$ ) and S-wave splitting ( $AV_S$ )  
6 across a natural shear zone using experimental and electron backscatter diffraction methods. Five  
7 samples were collected across a meter-scale shear zone developed under amphibolite- to greenschist-  
8 facies in the Neves metagranodiorite (Tauern Window, Eastern Alps). With increasing strain, seismic  
9 properties show a non-linear evolution resulting from interactions among structural, textural and min-  
10 eralogical changes. Geometry and magnitude of seismic anisotropy are controlled by the interference

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11 of strongly anisotropic phyllosilicates with less anisotropic but more abundant quartz and plagioclase.  
12 The amphibolite-facies shear zone margin ( $> 500^{\circ}\text{C}$ ) is free of major mineral transformations, and  
13 rather characterized by the development of S-C structure, inducing an orthorhombic symmetry to the  
14  $V_P$  distribution and up to 12 % AVP, with the fast axis parallel to lineation.  $AV_S$  is maximized (6 %) in  
15 the two directions parallel to the C-planes. Further deformation under greenschist-facies ( $450\text{-}500^{\circ}\text{C}$ )  
16 induces significant crystallization of muscovite, chlorite, and albite. A dominant hexagonal symmetry  
17 develops, with a fast VP plane and a fast S-wave polarization plane along the C-planes.  $AV_P$  and  
18  $AV_S$  exceed 10 %. The computed elastic tensor of the whole shear zone emphasizes the importance of  
19 the strain gradient on the seismic signature of the overall structure. In our model, the bulk seismic  
20 properties are dominated by the orthorhombic symmetry of the protomylonitic zone rather than by  
21 the ultramylonite core with hexagonal symmetry (C-plane), which may explain why seismic studies of  
22 shear zones commonly involve a fast velocity plane oblique to the layer boundaries.

## 23 **1. Introduction**

24 Characterizing the structure and the geometry of shear zones is crucial for better understanding  
25 the overall behavior of the Earth's lithosphere. Indeed, one of the main challenges in geosciences  
26 is to improve our knowledge of the interactions between the different layers of the lithosphere, and  
27 shear zones play a key role in these interactions since they constitute the deep roots of transcrustal  
28 discontinuities (e.g. [Sibson, 1977](#)). Involving both brittle and ductile deformation mechanisms (e.g.  
29 [Hayman and Lavier, 2014](#); [Segall and Simpson, 1986](#)) and acting as fluid channels allowing important  
30 chemical mass transfer (e.g. [Goncalves et al., 2012](#); [Oliot et al., 2010](#)), they thus make a chemical  
31 and rheological link between the processes operating in the shallower crust and in the lithospheric  
32 mantle. The vertical link with the seismogenic upper crust is also crucial to investigate, since shear  
33 zones can host brittle-ductile cyclic processes that still have poorly understood roles in the overall  
34 seismic cycle (e.g. [Hayman and Lavier, 2014](#); [Fussey and Handy, 2008](#)). However, imaging these  
35 highly deformed systems with sufficient resolution is still challenging. Among geophysical techniques,  
36 receiver functions (RF) have been successful in detecting velocity contrasts and anisotropic domains

37 highlighting the presence of deep crustal shear zones (e.g. [Phinney, 1964](#); [Burdick and Langston, 1977](#);  
38 [Schulte-Pelkum and Mahan, 2014](#)). One of the limitations of these methods is due to the use of  
39 simplified geological models for shear zones consisting of homogeneous anisotropic fabrics surrounded  
40 by isotropic protoliths (e.g. [Schulte-Pelkum et al., 2005](#)), while the relationship between strain and  
41 seismic properties within shear zones is made more complex by the variety of intrinsic parameters  
42 (mineralogy, crystal preferred orientation, structure). For example, [Schulte-Pelkum et al. \(2005\)](#) used  
43 RFs to image a crustal decollement beneath the Himalaya and quantified an angle of  $\sim 40^\circ$  between the  
44 inferred anisotropic foliation and the expected dip of the shear zone boundaries. This suggests that  
45 the seismic anisotropy is not solely controlled by the highest strained part of the shear zone, where  
46 the ultramylonitic foliation is presumed to be parallel to the shear zone boundaries. Other parameters  
47 must thus be considered for explaining some seismically anisotropic crustal-scale shear zones that may  
48 have more heterogeneous geometries (e.g. [Fossen and Cavalcante, 2017](#); [Fusseis et al., 2006](#)).

49 In this paper, we propose an improved geological model to be considered for geophysical investiga-  
50 tions of shear zones. A mineralogical and microtextural analysis is combined with the quantification  
51 of seismic properties on an exposed meter-scale Alpine shear zone from the Tauern Window in Eastern  
52 Alps (Neves area, Tyrol, Italy). Along the ductile strain gradient, a suite of 5 samples were collected  
53 with various degree of deformation (from the protolith to the ultramylonitic core). Seismic properties  
54 of each sample were quantified by the means of two independent methods: (1) laboratory measure-  
55 ments of elastic wave velocities on cylindrical samples (P wave velocity  $V_P$ , P wave anisotropy  $AV_P$  and  
56 shear wave splitting  $AV_S$ ) and (2) calculation of the bulk elastic properties of the full polycrystalline  
57 aggregate and wave velocities from crystal orientation measurements derived from electron backscat-  
58 ter diffraction (EBSD) ([Mainprice, 2007](#)). Estimation of deformed rocks' seismic properties by the  
59 experimental (e.g. [Barruol et al., 1992](#); [Jones and Nur, 1982](#); [Leclère et al., 2015](#); [Wenning et al., 2016](#))  
60 or numerical methods (e.g. [Punturo et al., 2014](#); [Tatham et al., 2008](#)) has previously been performed,  
61 mostly focusing on P wave anisotropy. However, they have rarely been compared with each other. P  
62 and S wave velocity estimated in mylonites are about  $6.0 \pm 1.0$  and  $3.5 \pm 0.5 \text{ km.s}^{-1}$ , respectively, and

63 always lower when the propagation direction is normal to the foliation (e.g. [Chroston and Max, 1988](#);  
64 [Wenning et al., 2016](#)).  $AV_P$  and  $AV_S$  in mylonites have been estimated at 8-11 % (e.g. [Fountain, 1976](#);  
65 [Jones and Nur, 1982](#)) and 6-7 %, respectively (e.g. [Ji and Salisbury, 1993](#)).  $AV_S$  is usually lower when  
66 the propagation direction is normal to the foliation (e.g. [Barruol et al., 1992](#); [Ji and Salisbury, 1993](#)).  
67 Although the modal proportion and fabrics of micas are generally assumed as the main source of seis-  
68 mic anisotropy in mylonites, the role of other constituting phases of the lower crust, like amphibole,  
69 has also been highlighted ([Tatham et al., 2008](#)).

70 Although these two methods have been widely used independently, their application together on the  
71 same natural samples is less common. In this work, the combination of both methods is used to quantify  
72 the relative contributions of structure, texture and crystal preferred orientation and mineralogy on  
73 the seismic properties of shear zones (e.g. [Almqvist et al., 2013](#); [Barruol et al., 1992](#)). Particular  
74 attention is paid to the interferences between the strongly anisotropic phase (e.g. phyllosilicates)  
75 and less anisotropic but more abundant rock-forming phases like quartz and plagioclase. Calculated  
76 bulk elastic stiffnesses and wave velocities of each sample are combined, re-oriented and weighted to  
77 elucidate the seismic behavior of a theoretical shear zone consisting of a protolith, an intermediate  
78 zone with a foliation at high angle to the shear plane, and a high strain core oriented parallel to the  
79 shear plane. This analysis is used to discuss the overall strength and geometry of seismic anisotropy  
80 generated by a crustal shear zone, and to add a geological constrain to the initial models used to  
81 predict their geophysical signature.

## 82 **2. Geological context and sampling strategy**

83 Within the austroalpine Eastern Alps, the Tauern Window exposes Variscan continental basement  
84 intruded by a late-Variscan batholith, the Zentralgneise (Fig.1a; e.g. [Morteani, 1974](#); [Schmid et al.,](#)  
85 [2013](#)). This complex was buried and deformed during the Alpine orogeny (c. 28-30 Ma), reaching  
86 peak conditions of 550-600°C and 0.5-0.7 GPa, corresponding to amphibolite-facies metamorphism  
87 (e.g. [Christensen et al., 1994](#)). The Variscan metasediments were pervasively mylonitized, while the  
88 more resistant magmatic intrusions were only weakly affected by deformation (e.g. [Morteani, 1974](#);

89 [Glodny et al., 2008](#)). Thus today, the Tauern Window shows a structural pattern that is typical of  
90 the ductile deformation of granitic regions (e.g. [Choukroune and Gapais, 1983](#)), which is inherited  
91 from these rheological heterogeneities. It is characterized by an anastomosing geometry defined by  
92 low-strain granodioritic domains surrounded by km-scale mylonitized Variscan metasediments. Only  
93 the low-strain granodioritic domains embedded in this crustal-scale shear zone system preserve the  
94 m-scale strain gradients that are of interest for our analysis.

95 The study area is located in the south-western part of the Zentralgneise, 4 km northeast of Neves  
96 lake (South Tyrol, Italy). It is a glacier-polished region of nearly 500 000  $m^2$ , exposed by the retreat  
97 of the Mesule glacier (Fig.1b). It has been described in detail by [Mancktelow and Pennacchioni \(2005\)](#)  
98 and [Pennacchioni and Mancktelow \(2007\)](#). The dominant lithology is a late-Variscan metagranodiorite,  
99 affected by many magmatic dykes. In the core of the low-strain domain, alpine deformation occurs  
100 as localized cm- to m-wide shear zones. The shear zones localized on viscosity contrasts, such as  
101 the margins of aplite dykes or brittle fractures, affecting the metagranodiorite. The studied outcrop  
102 is a steeply dipping, dextral alpine shear zone of nearly 2 m in width, oriented east-west (Fig.1c).  
103 In order to study the role of ductile strain localization on the seismic properties, 4 homogeneously  
104 deformed samples were collected from the shear zone, corresponding to 4 successive deformation stages:  
105 protomylonite 1 (18NEV01A), protomylonite 2 (18NEV01B), mylonite (13NEV01C) and ultramylonite  
106 (13NEV01D) (Fig.1d). As a reference frame, one sample was collected in the surrounding undeformed  
107 protolith (13NEV01A), which preserves an inherited magmatic north-south fabric at high angle to the  
108 shear zone and allows us to investigate the role of the protolith fabric on the anisotropy of the whole  
109 shear zone.

### 110 **3. Methods**

#### 111 *3.1. Elastic-wave velocity measurements*

112 Velocity measurements of ultrasonic waves (P and S) were performed in a high-pressure hydrostatic  
113 apparatus ( $\sigma_1 = \sigma_2 = \sigma_3$ ) located in the Rock Deformation Laboratory at the University of Liverpool

114 (UK) (Armitage et al., 2011; Blake et al., 2013). The apparatus uses low-viscosity silicon oil (10 cS)  
115 as the confining medium and is equipped with an ultrasonic active acquisition system. Experiments  
116 were carried out in room-dry conditions and at room temperature on cores of 20 mm in diameter and  
117 between 20 and 90 mm in length. Cores of the ultramylonite could not be recovered from the field  
118 samples due to the friable nature of the material. Cores were squared and then placed in a 3-mm thick  
119 PVC jacket to prevent infiltration of the confining silicon oil into the samples. Velocity measurements  
120 were conducted at 250 MPa, which is the maximum confining pressure supported by the apparatus.  
121 At these conditions, we expect to close most of the open microcracks and measure the intrinsic seismic  
122 properties of mineral aggregates (see section 4.3). Indeed, many experimental studies have described  
123 an asymptotic evolution of seismic properties with increasing confining pressure, consisting in a fast,  
124 non linear velocity increase (and anisotropy decrease) at lower pressure, followed by a more progressive,  
125 linear evolution (e.g. Birch, 1960; Christensen, 1965; Kern and Wenk, 1990; Blake et al., 2013). This  
126 behavior has been interpreted as the closure of most of the open microcracks, until a crack-closing  
127 pressure that ranges from 100 to 250 MPa (e.g. Christensen, 1965; Kern and Wenk, 1990)). It is  
128 also noteworthy that at high confining pressure, the signal-to-noise ratio is maximized, thus reducing  
129 uncertainties in arrival time picking.

130 Jacketed cores were clamped between the top and the bottom loading platens on which P and  
131 S 1-MHz piezoelectric ceramic transducers are glued. P- or S-seismic waves are emitted from the  
132 bottom platen with  $P_{bottom}$  and  $S_{bottom}$  transducers and recorded by P- and S-top transducers. The  
133 top S-transducer records on two components,  $S_1$  and  $S_2$ , which are at  $90^\circ$  from each other, and point  
134 toward East and North, respectively in the (N, E, d) coordinate system depicted in Supplementary  
135 Material S1. The  $S_{bottom}$  emitting transducer is polarized N45°E and  $P_{bottom}$  is vertical. This source-  
136 receiver geometry allows the recording of P- and S-wave velocity, as well as shear-wave splitting (see  
137 Supplementary Material S2 for the pulser-receiver settings, after Blake, 2011). Cores were oriented in  
138 the sample assembly according to the orientation of the foliation or the lineation which are aligned  
139 with  $S_1$  (Supplementary Material S1).

140 Data acquisition was performed using the pulse-transmission technique and by averaging 512  
 141 stacked signals, to optimize the signal-to-noise ratio (see Supplementary Material S3a-d). Seismic  
 142 velocities are calculated by picking the first seismic phase arrival time (see Supplementary Material  
 143 S3e). Picking error is estimated with the method of Diehl et al. (2009), in which the arrival ( $a$ ) of the  
 144 seismic phase is considered as a probabilistic function  $P_a(t)$ . The arrival time ( $t_A$ ) is expressed as the  
 145 most likely arrival time ( $P_a(t) = (max)P_a$ ), included in a temporal uncertainty window defined by the  
 146 earliest ( $t_E$ ) and the latest ( $t_L$ ) arrival times, where  $P_a$  tends to zero.  $t_E$  and  $t_L$  are defined using the  
 147 Amplitude-based Signal-to-Noise Ratio (ASNR). The error increases as the interval between  $t_E$  and  
 148  $t_L$  increases. Velocity anisotropy is calculated as follows (e.g. Birch, 1960; Mainprice and Humbert,  
 149 1994) :

$$A(\%) = 200 \times \frac{(V_{max} - V_{min})}{(V_{max} + V_{min})} \quad (1)$$

150  $AV_P$  is inferred from the maximum and minimum velocity among the  $V_P$  measured along the X-,  
 151 Y- and Z- cores of each rock sample.  $AV_S$  is estimated for each core, from the maximum and minimum  
 152 velocity among  $V_{S1}$  and  $V_{S2}$ .

### 153 3.2. Calculating rock aggregate seismic properties from petrofabrics

154 Crystallographic preferred orientations (CPO) were measured using electron backscatter diffraction  
 155 (EBSD) with a Hitachi SU3500 scanning electron microscope equipped with a W source electron gun at  
 156 the University of Colorado Boulder (USA). Analytical settings are summarized in the Supplementary  
 157 Material S4. CPOs are plotted in the structural reference frame of the sample, where X-, Y- and Z-  
 158 axes correspond to the three principal directions of the strain ellipsoid. Fabric strength is expressed  
 159 with the J index (Bunge and Esling, 1982). The J index has a value of one for random orientations and  
 160 scales infinitely with increasing texture strength, although typical "high" values for natural polycrys-  
 161 tals range from 10 to 100. Indexing phyllosilicates by EBSD is very challenging (Prior et al., 2009).  
 162 EBSD data acquired on micas is biased by the dependence of their indexing rate on their crystallo-  
 163 graphic orientation: mica indexing is higher on the basal (001) plane. To avoid the overestimation of



164 this population, we selected regions of interest where the relative proportion of the underestimated  
165 population (basal plane at high angle to the thin section) is sufficient to balance the better indexing  
166 of the first one (see Supplementary Material S5). It results in a more representative overall fabric.  
167 Calculations of elastic tensor and seismic velocities of polymineralic aggregates (Mainprice and Casey,  
168 1990; Mainprice and Humbert, 1994) are based on the volume fraction, CPO and the elastic tensor of  
169 each mineral phase, from Ogi et al. (2006) (quartz), Brown et al. (2016) (plagioclase), Waesermann  
170 et al. (2016) (K-feldspar), Vaughan and Guggenheim (1986) (muscovite), Aleksandrov and TV (1961)  
171 (biotite and chlorite); Ryzhova et al. (1966) (epidote) and Jiang et al. (2004) (garnet). Single crystal  
172 elastic tensors are rotated according to their CPO and then weighted according to their modal pro-  
173 portions. Seismic properties of the rock aggregate are calculated using the Voigt-Reuss-Hill average  
174 (VRH; Hill, 1952), which is the arithmetic mean of the Voigt (iso-strain boundary; Voigt, 1928) and  
175 Reuss (iso-stress boundary; Reuss, 1929) approximations. The VRH approximation is widely used  
176 to estimate the elastic properties of anisotropic polycrystalline aggregates since it provides consistent  
177 values with laboratory measurements of seismic velocities (e.g., Mainprice and Humbert, 1994; Bunge  
178 et al., 2000). Seismic properties are calculated with the MATLAB software toolbox MTEX, developed  
179 by D. Mainprice (Mainprice, 1990; Mainprice et al., 2011; Hielscher and Schaeben, 2008).

## 180 4. Results

### 181 4.1. Structural and mineralogical evolution

182 The magmatic assemblage of the Neves metagranodiorite consists of plagioclase (48 vol %), quartz  
183 (28 vol.%), biotite (14 vol.%) and minor K-feldspar (9 vol.%) (Fig.2a-b). Igneous plagioclase ( $Pl_0$ ),  
184 occurs as coarse ( $\sim 1$  mm) subhedral grains, and is saussuritized, i.e. altered to a more Na-rich  
185 plagioclase containing a fine-grained assemblage of white mica and epidote (1 vol.%). These inclusions  
186 measure from a few tens to a few hundreds of  $\mu\text{m}$  and are randomly oriented within the host crystal  
187 (Fig.2c). Quartz occurs as aggregates of large grains of 500  $\mu\text{m}$  to 1 mm in size, with irregular,  
188 serrate-to-lobate boundaries. Quartz grains are affected by weak intracrystalline deformation revealed

189 by discrete undulose extinction and rare recrystallized bands (Fig.2c). K-feldspar occurs as a late-  
190 crystallized interstitial phase in plagioclase-rich domains (Fig.2b). Igneous biotite ( $Bt_0$ ) is surrounded  
191 by recrystallized fine-grain metamorphic biotite ( $Bt_1$ ) (Fig.2d). Magmatic plagioclase and biotite  
192 define a weak magmatic foliation that trends north-south (Fig.2a).

193 Protomylonitic deformation is characterized by the development of a C-S-C' fabric, as observed in  
194 protomylonite 2 (Fig.3a). The intermediate stage protomylonite 1 shows similar but less-developed  
195 features, and will not be extensively described here (see Supplementary Material S6). Quartz-rich  
196 domains (26 vol.%) form recrystallized aggregates with an elongated, asymmetrical sigmoidal shape,  
197 with preferred stretching parallel to C, consistent with dextral shearing (Fig.3a). These aggregates  
198 are recrystallized to grains with a size of a few hundred  $\mu\text{m}$ , showing interlobate boundaries (Fig.21b),  
199 suggesting that deformation occurred via dislocation creep accommodated by grain boundary migration  
200 (GBM, Guillope and Poirier, 1979; Urai et al., 1986; Stipp et al., 2002). Quartz  $\langle c \rangle$  axes show three  
201 maxima between Z- and Y-axes ( $J=3.8$ ), which is consistent with a basal  $\langle a \rangle$  slip system assisted by  
202 the understated intervention of prismatic  $\langle a \rangle$  slip (Fig.3c, e.g. Mainprice et al., 1986; Stipp et al.,  
203 2002). Plagioclase occurs as a fine-grained recrystallized matrix in which the grain size does not exceed  
204 100  $\mu\text{m}$  (Fig.3a). These recrystallized plagioclase aggregates form elongated interconnected weak layers  
205 surrounding coarse quartz lenses. Saussuritized plagioclase phenocrysts, described in the protolith, are  
206 lacking. All these observations suggest that during the onset of strain localization, plagioclase is likely  
207 to have been the weakest phase. Although recrystallized plagioclase shows scattered crystallographic  
208 orientations ( $J=1.9$ ), poles (010) and (100) define a diffuse maxima at low angle to the X- and Z-axes,  
209 respectively (Fig.3c), consistent with a (100) $\langle 010 \rangle$  slip system. Interconnected thin layers of biotite  
210 (13 vol.%) and to a lesser extent muscovite (5 vol.%), define sigmoidal lenses within the plagioclase  
211 matrix and outline the quartz lenses boundaries (Fig.3a). They also highlight the development of a  
212 secondary shear plane C', that was not present in protomylonite 1, where biotite remained scattered  
213 and did not form interconnected ribbons yet (see Supplementary Material S6). A few grains of  $Bt_0$   
214 are preserved as porphyroclasts with cleavage at high angle to the foliation planes (Fig.3a), likely

215 at the origin of the diffuse maxima of  $\langle c \rangle$  axes in the center of the pole figure (Fig.3c). Biotite  
216 shows the strongest fabric ( $J=19.1$ ), which is due in part to the weaker contribution from this diffuse  
217 intermediate maxima corresponding to magmatic fabrics (Fig.3c), with respect to protomylonite 1 (see  
218 Supplementary Material S6). Neo-crystallization of metamorphic muscovite is observed and is likely  
219 related to K-feldspar breakdown (2 vol.%).

220 Mylonite 13NEV01C (Fig.4) is characterized by the development of a gneissic layering parallel  
221 to the shear plane. It consists of mm-thick quartz ribbons (33 vol.%) interlayered with plagioclase-  
222 (46 vol.%), epidote- (7 vol.%) and mica-rich (13 vol.%) domains (Fig.4a). The plagioclase matrix is  
223 very fine-grained (Fig.4a,b) with a weak CPO ( $J=1.4$ ). A dominant slip system is now very difficult  
224 to infer from the CPO (Fig.4d). Recrystallized quartz (Fig.4b) has a smaller grain size than in  
225 the protomylonites (40-500  $\mu\text{m}$ ). All quartz grains show evidence for very mobile grain boundaries  
226 (GBM). Quartz grains are elongated with a shape-preferred orientation (SPO) at  $\sim 35^\circ$  to the C-planes,  
227 consistent with dextral shearing (Fig.4b). Quartz  $\langle c \rangle$  axes still show maxima at low angle from Z,  
228 indicating basal  $\langle a \rangle$  dextral slip (Fig.4d, Mainprice et al., 1986). There is no evidence for prismatic  
229 slip. We observe a decrease of the biotite modal proportion (9 vol.%) and fabric strength ( $J=6.4$ ). The  
230 diffuse maximum of the  $\langle c \rangle$  axes close to the Y axis cannot be explained by the presence of magmatic  
231 relicts, since they completely recrystallized (Fig.4d). Biotite and muscovite (4 vol.%) ribbons delimit  
232 the epidote-rich plagioclase matrix from the quartz domains (Fig.4c).

233 Ultramylonite 13NEV01D (Fig.5) shows a significant change in mineralogy, with the complete  
234 breakdown of biotite ( $<1$  vol.%), epidote, garnet and K-feldspar, the production of a more albite-rich  
235 plagioclase, the crystallization of muscovite (21 vol.%) and, to a lesser extent, chlorite (3 vol.%). Mus-  
236 covite and chlorite occur as ribbons defining a very narrow foliation ( $\sim 500 \mu\text{m}$ , Fig.5a,b). Plagioclase is  
237 less abundant than in the less deformed rocks (30 vol.%). Cathodoluminescence imaging of plagioclase  
238 (Fig.5c) shows two distinct populations, based on the Na/Ca ratio : a Ca-richer population (oligoclase,  
239 green in CL) and a Na-richer one (albite, red in CL). The two populations of plagioclase show a weak  
240 CPO, with the (100) planes parallel to the foliation. Oligoclase does not show any clear slip direction,

241 but albite (010) poles are close to the lineation, suggesting a (100)<010> slip system (Fig.5d). Quartz  
242 (43 vol.%) occurs as either thin recrystallized ribbons or mixed with plagioclase. Grain size ranges  
243 from 40 to 100  $\mu\text{m}$ , while the largest quartz grains can reach 250  $\mu\text{m}$  (Fig.5b). Quartz fabric is weak  
244 ( $J=1.8$ ) compared to less deformed rocks and the <c> axes form a girdle between the two poles of the  
245 foliation, with a maximum located between the Z- and Y-axes (Fig.5d). We suggest that quartz in the  
246 ultramylonite deformed mostly via the rhomb <a> slip system (e.g. [Mainprice et al., 1986](#)).

#### 247 4.2. Pressure-temperature conditions of deformation

248 The main mineralogical change described along the strain gradient occurs between the mylonite and  
249 the ultramylonite and consists of the breakdown of biotite and plagioclase to produce muscovite, albite  
250 and chlorite. To constrain the P-T conditions of this transition, phase relations were modeled. The  
251 phase diagram shown in Fig.6 is derived from a pseudosection (isochemical phase diagram) computed  
252 with the bulk composition of the ultramylonite, in the H<sub>2</sub>O saturated Na<sub>2</sub>O–CaO–K<sub>2</sub>O–FeO–MgO–Al<sub>2</sub>O<sub>3</sub>–SiO<sub>2</sub>  
253 system with *Perple<sub>X</sub>* 6.8.8 ([Connolly and Kerrick, 1987](#); [Connolly, 2005](#)). This phase diagram shows  
254 an univariant equilibrium  $\text{Pl} + \text{Kf} + \text{Bt} + \text{Qz} + \text{H}_2\text{O} = \text{Mu} + \text{Chl} + \text{Ab} + \text{Ep}$  at  $\sim 450\text{-}475^\circ\text{C}$ . For the  
255 selected bulk composition, four divariant equilibria are stable over the P-T space associated with the  
256 above equilibrium (white fields in Fig. 6). Among them, those located on the high pressure side (Chl)  
257 and (Kf) are relevant for the studied sample. Along a cooling path, the (Chl) divariant equilibrium is  
258 responsible for the breakdown of the high temperature assemblage Pl Kf Bt Ep Qz into a Bt Mu Ep  
259 Ab Pl Qz assemblage. With further cooling, the divariant (Kf) equilibrium is crossed and the assem-  
260 blage Pl + Bt is broken down to produce the Mu Chl Ab-bearing assemblage that is characteristic of  
261 the ultramylonite. This phase relation modeling shows that the development of the meter-wide strain  
262 gradient occurred during a cooling path from  $\sim 500$  to  $\sim 475^\circ\text{C}$  at a pressure greater than 0.5-0.6 GPa.  
263 The high-temperature portion of this cooling path is consistent with the thermal peak of the "Tauern  
264 metamorphism", and to the amphibolite-facies conditions of deformation inferred from mineralogy  
265 and quartz deformation mechanisms by previous metamorphic studies (e.g. [Christensen et al., 1994](#);  
266 [Mancktelow and Pennacchioni, 2005](#)).

267 The evolution of mineralogy, microtexture and metamorphic conditions are summarized in Fig.7.  
268 Modal proportions are inferred from the combination of the EBSD data, scanning electron microscopy  
269 and cathodoluminescence imaging. In the following, the strain gradient will be divided in three parts:  
270 (1) the undeformed, metastable protolith bearing an inherited magmatic foliation, (2) the amphibolite-  
271 facies margin of the shear zone, consisting of protomylonite 1, protomylonite 2 and mylonite, char-  
272 acterized by a stable mineralogy and a widespread recrystallization, and (3) the greenschist-facies  
273 high-strain core of the shear zone, the ultramylonite, marked by the important crystallization of mus-  
274 covite, chlorite and albite. It reflects the evolution of the P-T conditions during the formation of the  
275 shear zone, which started under amphibolite-facies conditions and continued under retrograde condi-  
276 tions related to the exhumation episode, recorded in the greenschist-facies assemblage in the core of  
277 the shear zone.

#### 278 4.3. Measured seismic properties

279 Measured and calculated seismic properties along the strain gradient are summarized in Table 1.

280  
281 Figure 8 shows the evolution of the measured  $V_P$  (Fig.8a),  $AV_P$  (Fig.8b), and  $AV_S$  (Fig.8c) with  
282 strain, from the protolith to the mylonite. In the protolith,  $V_P$  ranges from  $5.97 \pm 0.9$  to  $6.25 \pm 0.08$   
283  $km.s^{-1}$ , which corresponds to  $4.6 \pm 0.3$  % of P-wave anisotropy. In one of the three measured cores,  
284 S-wave anisotropy reaches a maximum value of  $5.9 \pm 1.6$  %. The three perpendicular cores of the  
285 protolith are not oriented with respect to a reference frame, therefore the  $AV_P$  and  $AV_S$  values should  
286 not be treated as maximum values. In any case, the protolith already presents seismic anisotropy.

287 In the amphibolite-facies shear zone, from protomylonite 1 to mylonite, seismic P-wave velocities  
288 are not significantly different with respect to the protolith (Fig.8a). In all samples, the fast P-wave  
289 propagation direction is always parallel to the lineation direction (X-axis) and ranges from  $V_P(X) =$   
290  $6.22 \pm 0.09$  to  $6.44 \pm 0.14$   $km.s^{-1}$ . In contrast, P-waves propagating normal to the foliation and shear  
291 planes (Z-axis) are systematically the slowest and do not vary significantly with strain ( $V_P(Z) = 5.70$   
292  $\pm 0.08$  to  $5.90 \pm 0.25$   $km.s^{-1}$ ). In the protomylonites, P-waves propagating along the Y-axis have an

Table 1: Measured and calculated seismic properties across the Neves meter-scale strain gradient.

	Measured seismic properties			Calculated seismic properties		
<i>Protolith</i>	<i>C1</i>	<i>C2</i>	<i>C3</i>	<i>C1</i>	<i>C2</i>	<i>C3</i>
$V_P$ ( $km.s^{-1}$ )	6.25±0.08	6.11±0.04	5.97±0.9	6.41	6.20	6.25
$AV_P$ (%)		4.6±0.3			4.6	
$V_{S1}$ ( $km.s^{-1}$ )	3.41±0.14	3.40±0.26	3.47±0.16	3.73	3.70	3.74
$V_{S2}$ ( $km.s^{-1}$ )	3.41±0.15	3.32±0.17	3.68±0.33	3.58	3.58	3.70
$AV_S$ (%)	0.0	2.4±0.6	5.9±1.6	4.1	3.3	1.1
<i>Protomylonite 1</i>	<i>X</i>	<i>Y</i>	<i>Z</i>	<i>X</i>	<i>Y</i>	<i>Z</i>
$V_P$ ( $km.s^{-1}$ )	6.34±0.04	6.19±0.12	5.80±0.41	6.80	6.64	6.50
$AV_P$ (%)		8.9±1.4			5.2	
$V_{S1}$ ( $km.s^{-1}$ )	3.78±0.15	3.77±0.25	3.24±0.35	4.08	4.08	4.00
$V_{S2}$ ( $km.s^{-1}$ )	3.65±0.13	3.65±0.39	3.28±0.27	3.98	3.98	3.92
$AV_S$ (%)	3.5±0.5	3.3±1.1	1.2±0.5	2.48	4.00	2.02
<i>Protomylonite 2</i>	<i>X</i>	<i>Y</i>	<i>Z</i>	<i>X</i>	<i>Y</i>	<i>Z</i>
$V_P$ ( $km.s^{-1}$ )	6.44±0.14	6.14±0.17	5.70±0.08	6.80	6.65	6.42
$AV_P$ (%)		12.2±0.9			7.3	
$V_{S1}$ ( $km.s^{-1}$ )	3.79±0.29	3.90±0.40	3.35±0.18	4.05	4.05	3.88
$V_{S2}$ ( $km.s^{-1}$ )	3.55±0.16	3.63±0.32	3.36±0.17	3.85	3.80	3.80
$AV_S$ (%)	6.6±1.6	7.0±2.8	0.5±0.1	5.06	6.37	2.08
<i>Mylonite</i>	<i>X</i>	<i>Y</i>	<i>Z</i>	<i>X</i>	<i>Y</i>	<i>Z</i>
$V_P$ ( $km.s^{-1}$ )	6.22±0.09	6.24±0.08	5.90±0.25	6.90	6.69	6.67
$AV_P$ (%)		5.6±0.6			4.7	
$V_{S1}$ ( $km.s^{-1}$ )	3.82±0.25	3.78±0.32	3.34±0.59	4.14	4.12	4.09
$V_{S2}$ ( $km.s^{-1}$ )	3.69±0.17	3.57±0.34	3.34±0.53	4.07	4.02	4.01
$AV_S$ (%)	3.6±0.8	5.6±2.0	0.3±0.2	1.70	2.47	1.98
<i>Ultramylonite</i>	<i>X</i>	<i>Y</i>	<i>Z</i>	<i>X</i>	<i>Y</i>	<i>Z</i>
$V_P$ ( $km.s^{-1}$ )	-	-	-	6.60	6.60	6.0
$AV_P$ (%)		-			10.6	
$V_{S1}$ ( $km.s^{-1}$ )	-	-	-	4.02	4.02	3.72
$V_{S2}$ ( $km.s^{-1}$ )	-	-	-	3.74	3.74	3.64
$AV_S$ (%)	-	-	-	7.22	7.22	0.54

293 intermediate velocity leading to an orthorhombic symmetry ( $X \neq Y \neq Z$ ) for velocity anisotropy.  
294 In the mylonite,  $V_P(X)$  decreases down to  $6.22 \pm 0.09 \text{ km.s}^{-1}$  such that  $V_P(X) = V_P(Y)$ . Thus, the  
295 symmetry of the velocity distribution in the mylonite is hexagonal ( $X = Y \neq Z$ ) with a slow symmetry  
296 axis corresponding to the Z axis. The magnitude of velocity anisotropy shows large variations with  
297 strain (Fig.8b).

298 In the protomylonite,  $AV_P$  increases significantly and reaches  $12.2 \pm 0.9 \%$  in protomylonite 2. The  
299 mylonite is characterized by a decrease of anisotropy with an unexpected low  $AV_P = 5.6 \pm 0.6 \%$ ,  
300 comparable to the undeformed protolith. Shear wave splitting is almost null along the Z-axis ( $AV_S$   
301  $< 1.2 \pm 0.5 \%$ ). In contrast, shear wave anisotropy is maximum when S wave-propagation direction  
302 is parallel to the foliation or shear plane. The lineation does not influence shear wave anisotropy (i.e.  
303  $AV_S(X) = AV_S(Y)$ ). Similar to  $AV_P$ ,  $AV_S$  is at its maximum in protomylonite 2 ( $6.6 \pm 1.6 \%$  and  $7.0$   
304  $\pm 2.8 \%$  along X- and Y-axes, respectively), then it decreases to ca. 3 to 4 % in the mylonite.

#### 305 4.4. Calculated seismic properties

##### 306 4.4.1. Aggregate seismic properties

307 Figures 9 to 12 show for each sample of the strain gradient: (a) the modal abundance of each  
308 phase, (b) the CPO of the crystallographic direction corresponding to the fast, intermediate and slow  
309 P-wave propagation directions for each phase, (c) the seismic properties ( $V_P$  distribution,  $AV_P$ , and  
310  $AV_S$  distribution) computed for each phase and, finally, (d) the seismic properties of the whole rock  
311 aggregate. The location of EBSD maps is showed in the Supplementary Material S7. The results for  
312 protomylonite 1 are shown in the Supplementary Material S8. The stereographic projections showing  
313 the computed distributions of  $V_{S1}$  and  $V_{S2}$  are available in the Supplementary Material S9. The bulk  
314 elastic tensors calculated for each sample are summarized in the Supplementary Material S10.

##### 315 *Protolith*

316 In the protolith (Fig.9), the distribution of  $V_P$  is mainly controlled by plagioclase ( $AV_P(\text{Plagioclase})$   
317  $= 6 \%$ ), which represents 48 % of the rock volume (Fig.9a), and more particularly by the orientation of  
318 its fast direction (010) normal to the twin planes, and thus to the magmatic foliation (Fig.9b). Although

319 biotite has the highest single phase  $AV_P$  (9.7 %), the low modal abundance (14 %) combined with  
320 lack of preferred crystallographic orientation means that it exerts little systematic control on the bulk  
321  $V_P$  distribution and  $AV_P$  (Fig.9c). Due to its weak seismic anisotropy, quartz does not contribute to  
322 increase that of the aggregate. As a result of plagioclase and biotite contributions, the fastest P-wave  
323 velocities ( $\sim 6.3 \text{ km.s}^{-1}$ ) of the aggregate are concentrated at high angles to the magmatic foliation  
324 with a bulk  $AV_P$  of 4.6 % (Fig.9d).  $AV_S$  locally reaches 4.5 % but, in contrast to  $V_P$ , these maxima  
325 do not show any link with the igneous structure.

326 *Amphibolite-facies shear zone: protomylonites and mylonite*

327 The calculated  $V_P$  of the two protomylonites and mylonite, belonging to the amphibolite-facies  
328 strain gradient, are higher than that of the protolith and range from 6.4 to 7.0  $\text{km.s}^{-1}$  (Fig.10,11,  
329 see Supplementary Material S8 for protomylonite 1). The  $V_P$  anisotropy of the protomylonites and  
330 mylonite have an orthorhombic symmetry with the fast propagation axis parallel to lineation (X) and  
331 the slow propagation axis normal to the foliation (Z) (Fig.10d).  $AV_P$  for the aggregate increases with  
332 strain and reaches a maximum value of 7.3 % in protomylonite 2 (Fig.10d). In the mylonite,  $AV_P$   
333 decreases to 4.7 % (Fig.11d). The absolute values and the distribution of  $AV_S$  show the same evolution  
334 with strain. In the protomylonites,  $AV_S$  is maximum when the S-wave propagates within the foliation  
335 plane. There is almost no splitting perpendicular to the foliation plane ( $AV_S < 2$  %). The maximum  
336 value of  $AV_S$  reaches 6.5 % in protomylonite 2 and decreases to 4.0 % in the mylonite. The distribution  
337 of  $AV_S$  in the mylonite is complex without any clear relationship with the strain ellipsoid.

338 The anisotropy of wave speeds in protomylonite and mylonite samples directly reflect the behavior  
339 of phyllosilicates. Biotite and muscovite represent almost 20 % of the rock volume in the protomylonites  
340 and show the highest values of seismic anisotropy for both P- (38.7 and 40.3 % for biotite and muscovite  
341 respectively in protomylonite 2) and S-waves (up to 40 %, Fig.10c).

342 While in protomylonite 1, biotite fabric was influenced by the presence of magmatic relicts oriented  
343 at high angle to the shear plane (see Supplementary Materials S6 and S8), magmatic biotite is less  
344 abundant in protomylonite 2, leading to a strong fabric strength (J index = 19 for biotite, Fig.10b).



345 This evolution in phyllosilicate fabric strength is likely responsible for the progressive transition from  
346 orthorhombic to hexagonal symmetry of the whole aggregate and the increase in the magnitude of the  
347 seismic anisotropy .

348 The weakening of seismic anisotropy magnitude in the mylonite ( $AV_P = 4.7\%$  and  $AV_S < 4\%$ ) is  
349 due to (1) a decrease in the modal proportion of mica (from 18 to 13 %, Fig.11a) and to (2) a decrease  
350 in fabric strength of mica (J index  $< 6.5$  and 4 for biotite and muscovite, Fig.11b) due to a relatively  
351 large number of grains with their slow c-axes parallel to Y, in addition to metamorphic phyllosilicates  
352 parallel to the foliation plane (slow c-axis parallel to Z). As in the protomylonite, biotite oriented at  
353 high angles to the foliation plane is likely due to magmatic relicts. However, muscovite oriented at such  
354 high angles to the foliation plane cannot be magmatic since muscovite is essentially metamorphic and  
355 well-parallelized to the foliation. We speculate that these muscovite data are misindexed magmatic  
356 biotite.

357 Although plagioclase and quartz have a much weaker CPO and seismic anisotropy ( $AV_P$  and  $AV_S$   
358  $< 5\%$ ), their large modal abundance ( $\sim 75\%$  together) counteracts and attenuates the effect of the  
359 phyllosilicates on the bulk seismic properties. Because the fast propagation axis of plagioclase (010)  
360 is nearly parallel to the lineation, it contributes in the same way as micas to the velocity distribution.  
361 In contrast, quartz CPO results in slow  $V_P$  along or near the foliation plane. As a result, the overall  
362 anisotropy is nearly five times lower than the intrinsic anisotropy of biotite.

### 363 *Greenschist-facies ultramylonite zone*

364 The transition to the greenschist-facies ultramylonite (Fig.12) is characterized by striking changes of  
365 seismic properties (seismic wave velocities and anisotropy). Absolute P-wave velocities of the aggregate  
366 decrease ( $6.0 < V_P < 6.6 \text{ km.s}^{-1}$ ) and the strength of anisotropy increases to 10.6 % (more than  
367 twice the value calculated in the mylonite, Fig.12d). The ultramylonite has a well-defined hexagonal  
368 symmetry with a fast plane of P-wave propagation ( $6.5\text{-}6.6 \text{ km.s}^{-1}$ ) matching the foliation and a slow  
369 symmetry axis ( $6.0 \text{ km.s}^{-1}$ ) corresponding to the normal to the foliation plane (Z axis).  $AV_S$  follows  
370 the same trend with a doubling of anisotropy up to 10 % in the foliation plane, while splitting is null

371 when the S-wave propagates along Z. The ultramylonite is characterized by the complete breakdown  
372 of biotite and the crystallisation of more than 20 % muscovite with a very well-developed CPO parallel  
373 to the foliation ( $J=13.8$ , Fig.12a,b). As a result, muscovite shows the highest single phase seismic  
374 anisotropy, reaching 46.9 % and 50 % of  $AV_P$  and  $AV_S$ , respectively, and it controls the bulk anisotropy  
375 of the aggregate (Fig.12c). Seismic anisotropy of quartzofeldspathic phases is on average ten times  
376 lower than that of muscovite. Quartz ( $AV_P = 4.2$  %) shows a slow plane at low angle from the foliation  
377 and a fast axis related to the orientation of the c-axis. Plagioclase, which represents 30 % of the rock  
378 volume, is recrystallized to oligoclase and albite. Both phases have a very weak fabric and therefore  
379 low seismic anisotropy (oligoclase ( $AV_P = 3.6$  %) and albite ( $AV_P = 5.3$  %, Fig.12b,c). Thus, in the  
380 ultramylonite, the main effects of quartz and oligoclase + albite is to attenuate the effect of muscovite  
381 on the bulk seismic anisotropy. The variation of the calculated seismic properties across the strain  
382 gradient is summarized in Fig.13.

#### 383 4.4.2. Modelled seismic properties of a meter-scale shear zone

384 The aim of this section is to model the seismic properties of the meter-scale shear zone, from which  
385 the studied samples were collected, using the seismic properties calculated for each deformation stage.  
386 The natural shear zone shown in Fig.1c and d is simplified into a conceptual composite tabular shear  
387 zone (e.g. Fossen and Cavalcante, 2017, Fig.14). This conceptual model is divided into three parts:  
388 (1) the undeformed protolith, bearing a magmatic foliation normal to the shear zone margins ; (2) the  
389 amphibolite-facies shear zone, corresponding to protomylonites and mylonite, where the angle between  
390 the metamorphic foliation and the shear plane progressively decreases, from  $60^\circ$  in protomylonite 1  
391 to  $20^\circ$  in the mylonite ; (3) the greenschist-facies ultramylonite core of the shear zone, where the  
392 metamorphic foliation is parallel to the shear plane. The high angle between foliation and shear plane  
393 in protomylonites is due to the presence of the inherited magmatic foliation, at  $90^\circ$  to the shear zone.  
394 The respective seismic properties ( $V_P$ ,  $AV_P$ ,  $AV_S$ ) of the different deformation stages, shown on the  
395 left side of Fig.14, are from Fig.9d, 23d, 10d, 11d and 12d, and are rotated such that the reference  
396 frame matches the orientation of the foliation plane. The incident wave at the bottom of the model is

397 a cone of ray paths at  $15^\circ$  on either side of the vertical axis and travels normally to the shear zone.  
398 If we consider a teleseismic wave, the shear zone would be horizontal. When considering a teleseismic  
399 P-wave propagating through a full shear zone, four velocity contrasts are crossed:  $V_P$  increases by  $\sim 8$   
400 % between the protolith and the amphibolite-facies shear zone, and decreases by  $\sim 10$  % between the  
401 mylonite and the greenschist-facies ultramylonite core.

402 To obtain the bulk seismic properties of the shear zone, the rotated seismic properties of each  
403 component were weighted depending on its relative volume proportion. In order to discuss the role  
404 of the three defined domains of the conceptual shear zone, three different geometries were taken into  
405 account with three distinct high strain core versus amphibolite shear zone ratios (5, 15 and 30 %  
406 of high strain zone, respectively ; Fig.15a). The thickness of the amphibolite shear zone zone is  
407 kept constant. The geometries could also be considered as three distinct strain and time steps of  
408 the development of a shear zone (Fossen and Cavalcante, 2017; Means, 1995; Hull, 1988; Vitale and  
409 Mazzoli, 2008).

410 Results of the modelling are shown in Fig.15. For geometry I (5 % of high strain zone),  $V_P$   
411 anisotropy can be approximated with orthorhombic symmetry, with fast and slow axes at  $\sim 30^\circ$  and  $90^\circ$   
412 to the shear plane, respectively. With the progressive development of the high strain core, geometries II  
413 and III in Fig.15a show that the anisotropy progressively evolves towards a more hexagonal symmetry.  
414 The fast plane tends to become closer to the shear plane, but even when the shear zone represents  
415 approximately 50 vol.% of the considered volume (geometry III), there is still an obliquity of  $15^\circ$   
416 between the fast plane and the shear plane (Fig.15b). The strength of anisotropy  $AV_P$  does not  
417 significantly increase with strain. It ranges from 4.2 % in geometry I to 4.7 % in geometry III. Although  
418 the highly anisotropic strain zone volume increases by 25 %, the increase in bulk  $AV_P$  does not exceed  
419 12 % (Fig.15b,d). The obliquity between the anisotropic fabric and the shear plane is also reflected by  
420  $AV_S$ . Two  $AV_S$  maxima, aligned on a plane oblique to the shear plane, are observed in geometries I  
421 and II. With the development of the ultramylonite, these maxima become more connected, defining a  
422 plane, still oblique, along which splitting is maximized, although not homogeneous (Fig.15c).  $AV_S$  is

423 not modified by the increasing ultramylonite versus marginal zone ratio (Fig.15d).

## 424 5. Discussion

### 425 5.1. Consistency between measured and calculated seismic properties

426 In the studied shear zone from the Neves area, the experimental and calculated results show good  
427 consistency, with  $V_P$ ,  $AV_P$  and  $AV_S$  having similar trends for both methods. Experimentally measured  
428  $V_P$  ranges from 5.5 to 6.5  $km.s^{-1}$  (Fig.8a), which is in good agreement with published experimental  
429 results (e.g. Jones and Nur, 1982 in the Columbia River fault zone; Barruol et al., 1992 in the  
430 San Barthelemy massif; Kästner et al., 2020 in the Seve Nappe Complex; Leclère et al., 2015 in the  
431 Argentera fault zone) and numerical studies (e.g. Barruol et al., 1992; Punturo et al., 2014 in the Kavala  
432 shear zone). Although experiments have been performed under high pressure conditions, experimental  
433  $V_P$  is consistently lower than those calculated by microtextural EBSD analysis, by  $\sim 3\%$  (Fig.8a and  
434 13, Table 1). In contrast,  $AV_P$  is higher when estimated in laboratory than by EBSD, as previously  
435 observed by Almqvist et al. (2013) and Barruol et al. (1992) (Fig.8b, 13b, 1). This discrepancy may  
436 be due to the presence of aligned grain boundaries and microcracks that are not completely closed  
437 (Almqvist et al., 2013). It is likely that the confining pressure of 250 MPa is not sufficient to close  
438 all the microcracks, especially those with a low aspect ratio. This may cause a slight overestimation  
439 of the measured anisotropy magnitude, but since microcracks are likely to be parallel to the preferred  
440 orientation of the grains, the presence of residual microcracks should not influence the symmetry of the  
441 anisotropic fabrics in mylonites. We also cannot rule out that the mismatch between both methods  
442 may be due to polishing issues and misindexing of phyllosilicates that tend to underestimate the  
443 contribution of phyllosilicates on modeled seismic P-wave speed anisotropy. The Y-axis concentration  
444 of mica c-axes, which contributes to reducing the overall anisotropy magnitude, can be due to the  
445 preferential indexing of the grains showing this orientation (Dempsey et al., 2011), even if the indexing  
446 strategy may have reduced this effect (see Supplementary Material S5). Furthermore, the modelling  
447 approach used to calculate bulk elastic data and infer seismic properties of the rock aggregate is based  
448 on the Voigt-Reuss-Hill average (arithmetic mean) and therefore does not take into account the effect

449 of the spatial arrangement of grains (microstructure) (Vel et al., 2016). Compositional layering, with  
450 aligned phyllosilicate grains is also known to affect seismic wave speed anisotropy (Naus-Thijssen et al.,  
451 2011). In our case, this effect could be neglected because there is no layering in the igneous protolith  
452 and deformation does not produce a layering with strong mineralogical contrast.

453 Measured and modeled  $AV_S$  (Fig.8c and 13c, Table 1) are overall very consistent, except for the X  
454 core of protomylonite 2. Our results contrast with (Almqvist et al., 2013) who observed experimental  
455  $AV_S$  values consistently higher by 3-7 % than those calculated from a combination of EBSD and neutron  
456 diffraction data. Combining these two methods allows to test the accuracy of the reconstruction  
457 of seismic properties and to investigate the mineralogical and textural contributions of the effective  
458 properties measured in laboratory.

## 459 5.2. Evolution of bulk seismic properties with strain and metamorphism

460 The studied shear zone is described in three sections based on deformation and metamorphic style:  
461 (1) a protolith, (2) an amphibolite-facies strain gradient and (3) a greenschist-facies high strain zone.  
462 In this contribution, we show that the bulk seismic properties do not evolve linearly with strain.

463 Although the shear zone is developed in a mineralogically homogeneous plutonic rock without  
464 layering, this host rock initially exhibits a bulk seismic anisotropy that cannot be neglected from a  
465 geophysical perspective (4.6 %  $AV_P$  and 4.1 %  $AV_S(max)$  calculated). It results from the presence  
466 of a weak magmatic foliation, visible in the field and mostly defined by abundant and aligned plagioclase  
467 phenocrysts. The crystallographic preferred orientation of magmatic plagioclase is such that the  
468 magmatic foliation corresponds to a slow  $V_P$  plane, oriented at high angle ( $\sim 90^\circ$ ) to the shear zone  
469 boundaries. In the protolith, quartz and biotite grains are randomly oriented and do not contribute  
470 significantly to the bulk seismic anisotropy.

471 The development of the shear zone via solid state deformation under amphibolite facies is not  
472 associated with a significant change in the modal abundance of the main rock-forming phases (quartz,  
473 plagioclase and biotite) except for the crystallization of  $\sim 5$  % muscovite. Indeed, at the inferred P-T  
474 conditions of the onset of deformation (ca. 500°C and 0.6 GPa) there is no major phase transition

475 (Fig.6) that could explain the evolution of seismic properties between the protolith and the deformed  
476 zone. The major change is the development of a composite structure that consists of foliation and  
477 shear planes (S-C structure) combined with the recrystallization of plagioclase, quartz and biotite.  
478 With increasing strain, the fabric becomes very well defined with the development of interconnected  
479 plagioclase weak layers, alignment of phyllosilicates and the progressive rotation of the foliation plane  
480 (S) towards the shear plane (C) orientation. This structural change induces an increase of P-wave  
481 fastest velocities and anisotropy, which reaches maximum values of 12 % for  $AV_P$  and 7 % for  $AV_S$ ,  
482 in the protomylonite. However, in the higher strain mylonite, the strong decrease of  $AV_P$  and  $AV_S$ ,  
483 which drop down to  $\sim 5$  % and  $\sim 3-5$  % respectively, cannot be directly accounted for by either the  
484 fabric transition or the mineralogical evolution.

485 We suggest that this non-linear evolution of seismic properties with strain, and more particularly  
486 the decrease of anisotropy in the mylonite, is related to the evolution of crystallographic preferred  
487 orientations of the rock-forming minerals. Specifically, it is due to the interference between strongly  
488 anisotropic minerals (e.g phyllosilicates) and less anisotropic but more abundant phases (e.g. quartz  
489 and plagioclase). The interaction between quartz and muscovite has already been explored by [Naus-](#)  
490 [Thijssen et al. \(2011\)](#) and [Ward et al. \(2012\)](#), which demonstrated, using either synthetic or natural rock  
491 samples, that muscovite-induced anisotropy can be overwhelmed by the orientation of less anisotropic  
492 matrix grains and more particularly in mica-poor rocks. In the amphibolite-facies section of the studied  
493 shear zone, the destructive effect of quartz is shown by the geometry of P-wave velocities  $V_P$  and S-  
494 wave anisotropy  $AV_S$  (Fig.23,10,11). With 19 to 13 volume % of phyllosilicates, we could expect a  
495 hexagonal symmetry of both  $V_P$  and  $AV_S$  but the active slip system for quartz (basal  $\langle a \rangle$ ) results in  
496 the alignment of the fast c-axis perpendicular to the fast cleavage plane of phyllosilicates. This results  
497 in an orthorhombic symmetry for all the samples belonging to the amphibolite-facies shear zone section,  
498 with a fast axis corresponding to the lineation (X) and a slow axis perpendicular to the foliation (Z) and  
499 an intermediate axis (Y) that corresponds to the intersection of S- and C-planes. Although plagioclase  
500 represents almost  $\sim 50$  % of the rock volume of the amphibolite-facies shear zone, its CPO is so weak

501 that it could be considered as quasi-isotropic with respect to seismic wave speed behaviour. Even  
502 if the (100)<001> slip system aligns its fast axis with the lineation, the main effect of plagioclase  
503 is to mute to some extent the high values of seismic anisotropy associated with phyllosilicates. The  
504 unexpected decrease of seismic anisotropy magnitude in the mylonite is incompletely understood. It  
505 is not related to a strengthening of the quartz contribution: in contrast with protomylonites, where  
506 deformation is more likely controlled by dislocation creep, it has been shown that diffusion creep can  
507 become more important in the higher strained parts of shear zones, producing weaker CPO (e.g. [Condit  
508 and Mahan, 2018](#); [Oliot et al., 2014](#); [Wehrens et al., 2016](#)). In particular, it can explain the absence  
509 of significant CPO in plagioclase, where granular flow is likely to be dominant (e.g. [Oliot et al., 2014](#)).  
510 This variation in the dominant deformation mechanism could thus play a role in the loss of seismic  
511 anisotropy in mylonites. But despite the potential effect of deformation mechanisms, this decrease  
512 seems more probably related to a weakening of the phyllosilicate contribution. First, the amount of  
513 phyllosilicates in the mylonite decreases to 13 % (instead of 19-18 % in the protomylonite). Second,  
514 and perhaps most importantly, we observe a weakening of their fabric strength and therefore of their  
515 contribution to the bulk seismic anisotropy (Fig.11). Indeed, the breakdown of biotite due to the  
516 evolution of the metamorphic conditions initiates in the mylonite, and is not counterbalanced yet by  
517 the crystallization of muscovite, only observed in the ultramylonite. Our observations are consistent  
518 with numerical and experimental studies showing that seismic anisotropy can decrease in the most  
519 deformed stages, unless a sufficient modal fraction of well-oriented micas could balance this effect (e.g.  
520 [Jones and Nur, 1982](#); [Barruol et al., 1992](#); [Tatham et al., 2008](#); [Wenning et al., 2016](#)).

521 The evolution of seismic properties in the high strain core of the shear zone, corresponding to an  
522 increase of  $AV_P$  and  $AV_S$  up to 10 and 7 %, respectively, in the ultramylonite, is consistent with  
523 experimental and numerical values (6-18 %) obtained on mylonitized granites ([Jones and Nur, 1982](#),  
524 [1984](#); [Barruol et al., 1992](#); [Lloyd et al., 2009](#)). This evolution observed along the gradient is directly  
525 related to change in syn-kinematic metamorphic conditions and therefore mineralogy. In granitic  
526 rocks, the transition from amphibolite- to greenschist-facies conditions during cooling from 500 to

527 450°C corresponds to the univariant reaction in the NaCaKFMASH system : plagioclase + biotite  
528 + K-feldspar + quartz + H<sub>2</sub>O = muscovite + chlorite + albite + epidote (Fig.6). This transition  
529 results in the crystallization of more than 20 % muscovite and chlorite and the recrystallization of  
530 plagioclase into a fine grain aggregate of albite and oligoclase. It is also accompanied by a switch in  
531 the dominant active slip system in quartz (from basal <a> to rhomb <a>). Among all these changes,  
532 the crystallization of very well aligned muscovite in the shear plane has the most profound effect on  
533 seismic properties (Fig.12). The geometry of the P-wave velocity distribution and the anisotropy of  
534 S-wave exhibits a strong hexagonal symmetry with a fast  $V_P$  plane, and strong S-wave splitting with  
535 fast polarization along the shear plane. Quartz and plagioclase do not modify the geometry of the  
536 seismic properties but attenuate the intensity of anisotropy of the bulk aggregate.

### 537 *5.3. Imaging of crustal shear zones based on seismic anisotropy*

538 Only shear zones with a thickness higher than  $\sim 2$  km can be properly imaged by teleseismic  
539 technics like receiver functions (Schulte-Pelkum and Mahan, 2014). Many structural studies carried  
540 on exhumed km-scale shear zones have described them as very complex and heterogeneous deformation  
541 regions. They consists of a network of anastomosed high-strain zones surrounding low-strain lenses,  
542 as described for example in Cap de Creus (Spain; Carreras, 2001, e.g.) or in the East Tenda shear  
543 zone (Corsica, France; Molli et al., 2006, e.g.). This anastomosed geometry can be observed at all  
544 scales, from resistant feldspar lenses embedded in micas-rich matrix at the microscale, to plutonic  
545 domains of several hundreds of kilometers surrounded by kilometer-wide mylonite zones (e.g. Fossen  
546 and Cavalcante, 2017). Modelling the teleseismic signature of anastomosed crustal-scale shear zone  
547 is therefore a complex task that depends widely on the geometry. In this contribution, we focus on  
548 the seismic signature of a simplified single strain gradient, from the weakly deformed domains to the  
549 high-strain zone.

550 Our reconstruction of the seismic properties of the whole strain gradient highlights the major  
551 role of the amphibolite-facies marginal zone, which induces an orthorhombic symmetry in the spatial  
552 distribution of  $V_P$  and an obliquity between the shear plane and the fast P-wave propagation plane



553 (Fig.15). While some prior studies have also emphasized the potential importance of shear zone strain  
554 gradients (e.g. [Rey et al., 1994](#)), others have assumed that the seismic properties of shear zones are  
555 controlled by the ultramylonite core (e.g. [Jones and Nur, 1984](#); [Christensen and Szymanski, 1988](#);  
556 [McDonough and Fountain, 1988, 1993](#)), which shows a strong velocity contrast with the surrounding  
557 rock. In a natural shear zone, the progressive rotation of the foliation plane in the protomylonitic  
558 margin causes a more gradual decrease of  $V_P$  in a direction normal to the shear plane, marked by  
559 two velocity contrasts, between the protolith and the shear zone margins, and between the shear zone  
560 margins and the ultramylonite core.

561       Imaging crustal shear zones requires to compute synthetic seismograms associated with a geological  
562 model that shows the spatial variations of physical properties of the rocks (e.g. [Schulte-Pelkum and](#)  
563 [Mahan, 2014](#)). Most models used in geophysical studies assumed an homogeneous anisotropic fabric  
564 within shear zones, reflecting either the oblique fabric of the marginal orthorhombic zone (Fig.16a, e.g.  
565 [Zandt et al., 2004](#); [Schulte-Pelkum et al., 2005](#)) or the parallelized fabric of the ultramylonite (Fig.16b,  
566 e.g. [Schulte-Pelkum and Mahan, 2014](#)). The aim of the present contribution is to include also the role of  
567 strain gradient of the shear zones. The integrated  $AV_P$  estimated in this study for the whole shear zone  
568 including the amphibolite-facies margins with an orthorhombic symmetry does not exceed 5 %, despite  
569 an  $AV_P$  higher than 10 % in the ultramylonite. Thus, the role of the S plane-dominated marginal zone  
570 may need to be taken into account in the elaboration of the synthetic seismograms. Furthermore, the  
571 role of the protolith cannot be neglected, since it can bear an inherited anisotropic fabric which may  
572 impact the strength and the orientation of the seismic properties of the whole shear zone. We thus  
573 propose a new model based on the spatial variability of the seismic properties observed across the  
574 strain gradient. Our model includes (1) an anisotropic protolith, and a deformed zone composed of (2)  
575 an amphibolite-facies marginal zone characterized by an orthorhombic symmetry which fast axis is at  
576  $30^\circ$  from the shear zone boundaries and (3) a greenschist-facies ultramylonitic core with a hexagonal  
577 symmetry which fast plane is parallel to the shear plane (Fig.16c). This model does not take into  
578 account all the complexity of the transition zone, i.e. the progressive rotation of the foliation and the

579 decrease of anisotropy in mylonite. However, it highlights a more realistic view of shear zones and  
580 could contribute to solving uncertainties in the interpretation of geophysical imaging. For instance,  
581 [Schulte-Pelkum et al. \(2005\)](#) have observed an obliquity between the expected and measured fast planes  
582 within the Main Himalayan Thrust. Actually, the detected anisotropic fabric is oriented  $40^\circ$  from the  
583 thrust plane. According to our model, and assuming that it can be upscaled as explained above, this  
584 detected anisotropy could be explained by the presence of the transition zone which presents similar  
585 anisotropy magnitude as the ultramylonitic core, and most likely controls the global orientation of  
586 the anisotropic fabric. However, we have shown that the metamorphic conditions of deformation,  
587 i.e. the reequilibration into greenschist-facies conditions, strongly influence the seismic properties of  
588 the studied shear zone by modifying texture and mineralogy in the ultramylonite zone. Thus, the  
589 seismic properties of crustal-scale strain zones spanning a wide range of pressure and temperature  
590 conditions must integrate the effect of the various metamorphic conditions through the full extent of  
591 the discontinuity.

## 592 **6. Conclusions**

593 Ultrasonic laboratory measurements and calculations from petrofabrics exhibit a complex transi-  
594 tion in the seismic properties across the studied shear zone strain gradient. The expected continuous  
595 increase of the magnitude of seismic anisotropy with strain is disrupted by a large reduction of the  
596 anisotropy in the mylonite, despite the progressive development of the foliation parallel to the shear  
597 plane. The seismic properties are mainly controlled by the behavior of phyllosilicates and by the de-  
598 structive interference from the more weakly anisotropic quartzo-feldspathic phases. The results also  
599 highlight the importance of the metamorphic conditions. Deformation in protomylonites and mylonite  
600 developed under amphibolite-facies conditions without significant major mineral transformation. Here,  
601 a S-C composite structure is defined by metamorphic biotite and influenced by locally preserved mag-  
602 matic biotite. In contrast, the retrogression under greenschist-facies conditions in the ultramylonite  
603 induces a C-dominated structure, characterized by the complete replacement of biotite by muscovite  
604 and chlorite. In summary, under amphibolite- to greenschist-facies conditions, the seismic properties

605 of shear zones may be largely controlled by the orthorhombic symmetry preserved in the transitional  
606 margin of the strain gradients, consistent with oblique anisotropic fabrics associated with shear zones  
607 imaged in some geophysical studies. While the high-strain ultramylonites have often be regarded as  
608 the main source of seismic anisotropy within shear zones, this study highlights the important contri-  
609 bution of the protolith and the shear zone margins, which should be taken into account to predict the  
610 geophysical signature of crustal-scale shear zones.

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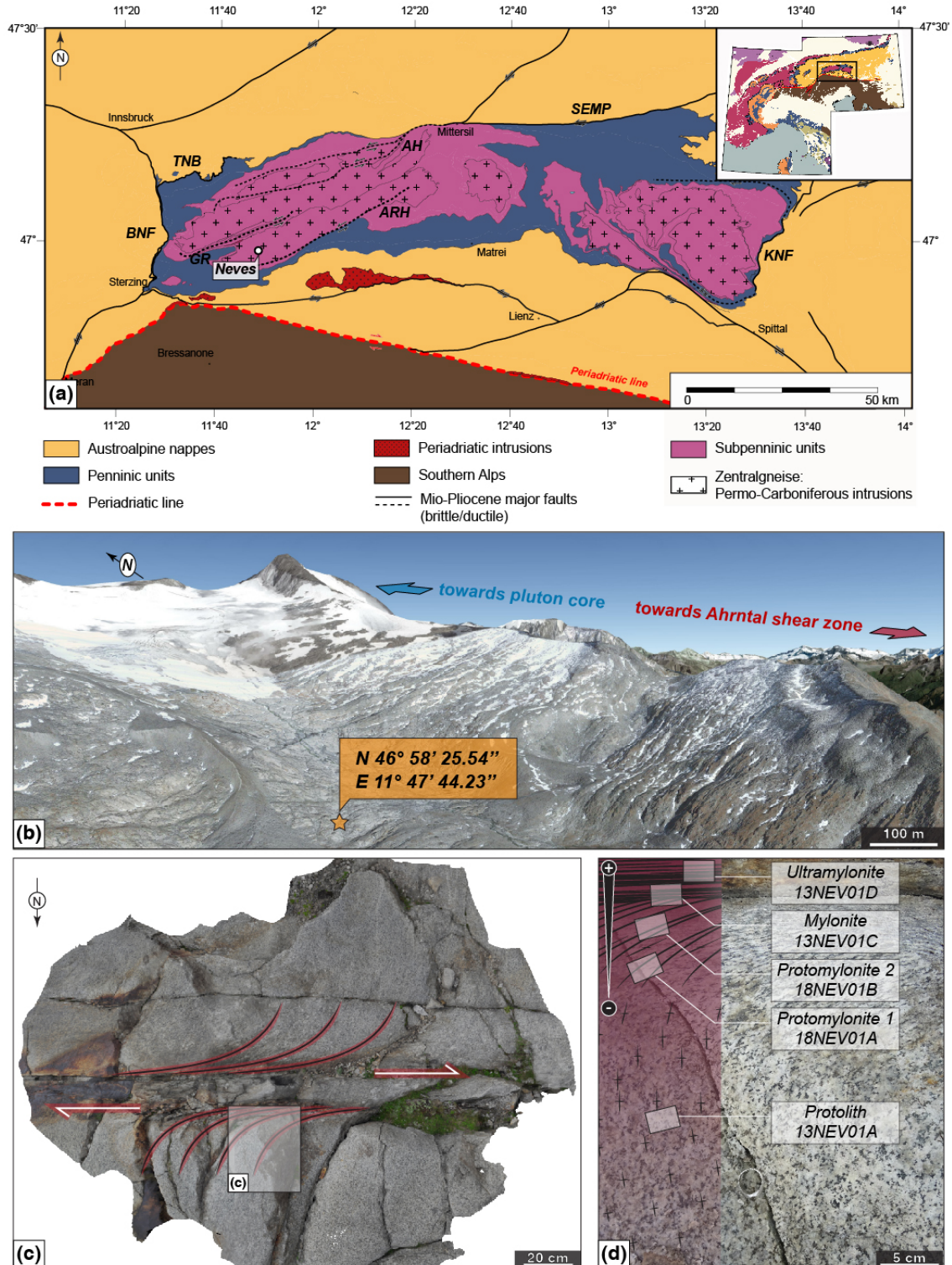


Figure 1: (2-column fitting image) Sampling area (a) Simplified tectonic map of the Tauern Window (modified from Schmid et al., 2013). The sampling area of Neves is highlighted by the white circle. AH Ahorn shear zone, AHR Ahrntal shear zone, BNF Brenner Normal Fault, KNF Katschberg Normal Fault, SEMP Salzach-Ennstal-Mariazell-Puchberg fault, TNB Tauern Northern Boundary fault. (b) View of the Neves area. The orange star and the coordinates indicate the location of the studied outcrop. (c) Photogrammetry picture of the studied outcrop, a m-wide dextral shear zone (top view). The white arrows indicate the kinematics, and the black lines materialize the metamorphic foliation. The square indicates the location of (d). (d) Samples location on the ductile strain gradient. The black arrow on the left

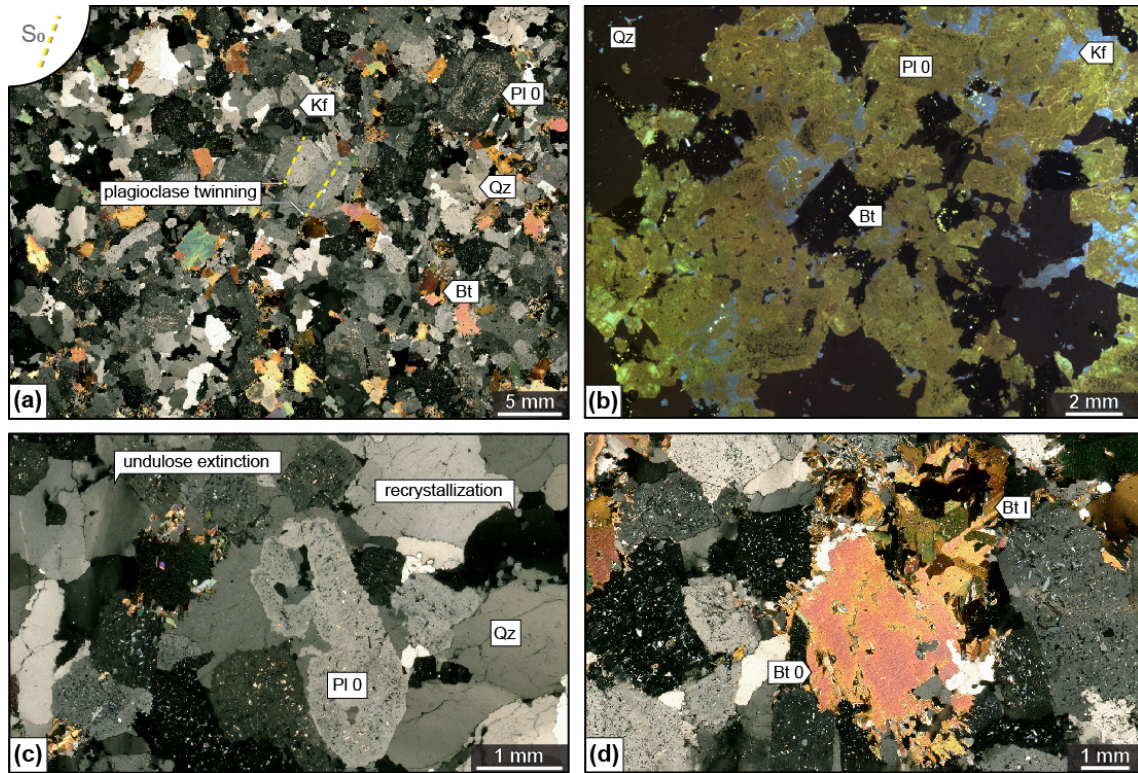


Figure 2: (2-column fitting image) Mineralogy and microstructure of the Neves metagranodiorite (13NEV01A). (a) Cross-polarized photomicrograph of the thin section. The magmatic foliation  $S_0$  is highlighted by the aligned twin planes in plagioclase (the yellow dashed lines). (b) Cathodoluminescence (CL) picture. Quartz (Qz): dark blue, plagioclase ( $Pl_0$ ): greenish, potassic feldspar (Kf): light blue, biotite (Bt): dark. (c) Quartz microstructures. (d) Recrystallization of magmatic biotite ( $Bt_0$  in new metamorphic grains ( $Bt_1$ )). Mineral abbreviations are from [Whitney and Evans \(2010\)](#)



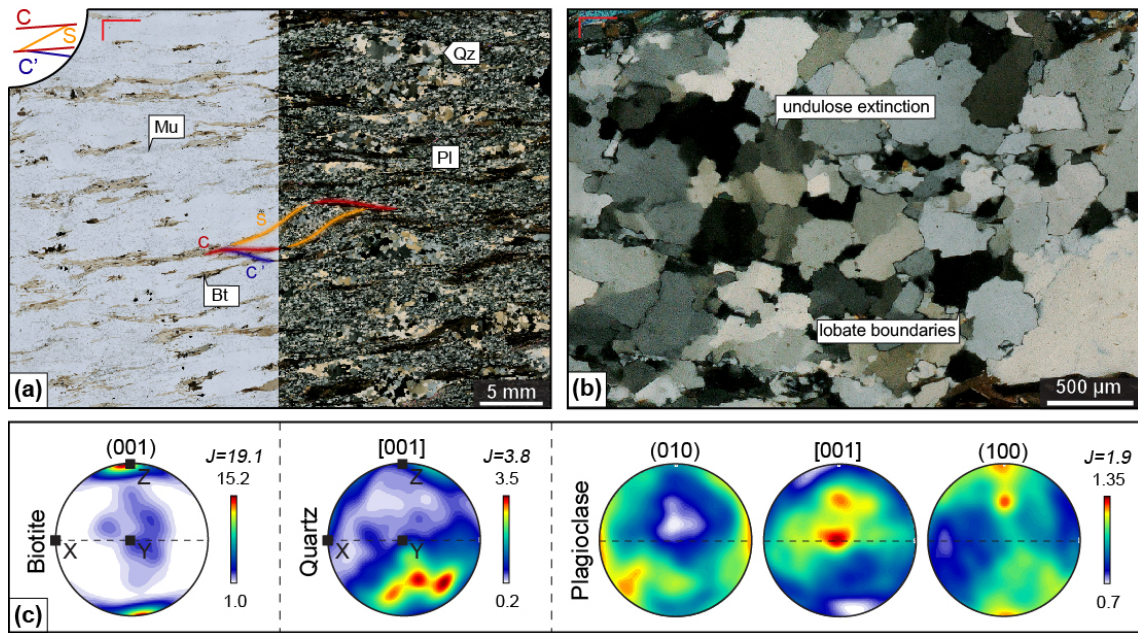


Figure 3: (*2-column fitting image*) Mineralogy and microtextures of protomylonite 2 (18NEV01B) (a) Plane-polarized (left) and cross-polarized (right) photomicrographs of the thin section. The sigmoid highlights the C-, S- and C'-planes, in red, orange and blue, respectively. (b) Quartz microstructures. (c) CPO of quartz, plagioclase and biotite (lower hemisphere).

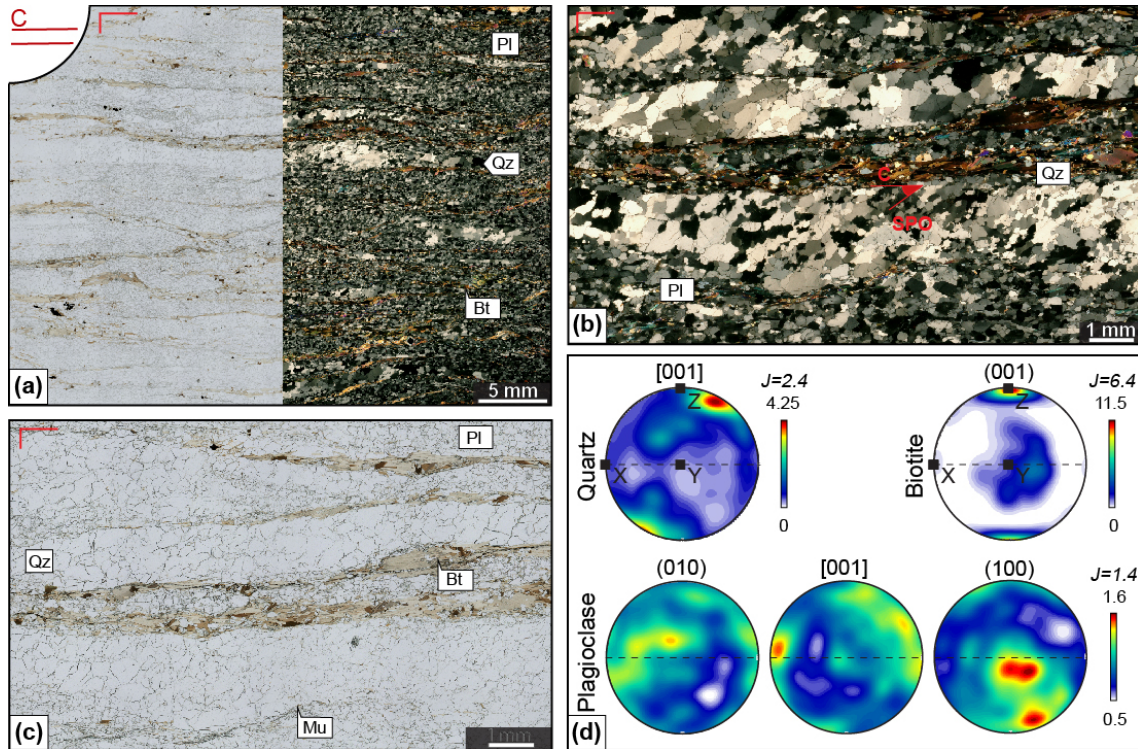


Figure 4: (*2-column fitting image*) Mineralogy and microtextures of the mylonite (13NEV01C) (a) Plane-polarized (left) and cross-polarized (right) photomicrographs of the thin section. (b-c) Metamorphic layering defined by the alternance of quartz, plagioclase, and micas ribbons. Shape-preferred orientation (SPO) of quartz at  $\sim 35^\circ$  from the C-plane is highlighted on (b). (d) CPO of quartz, plagioclase and biotite (lower hemisphere).

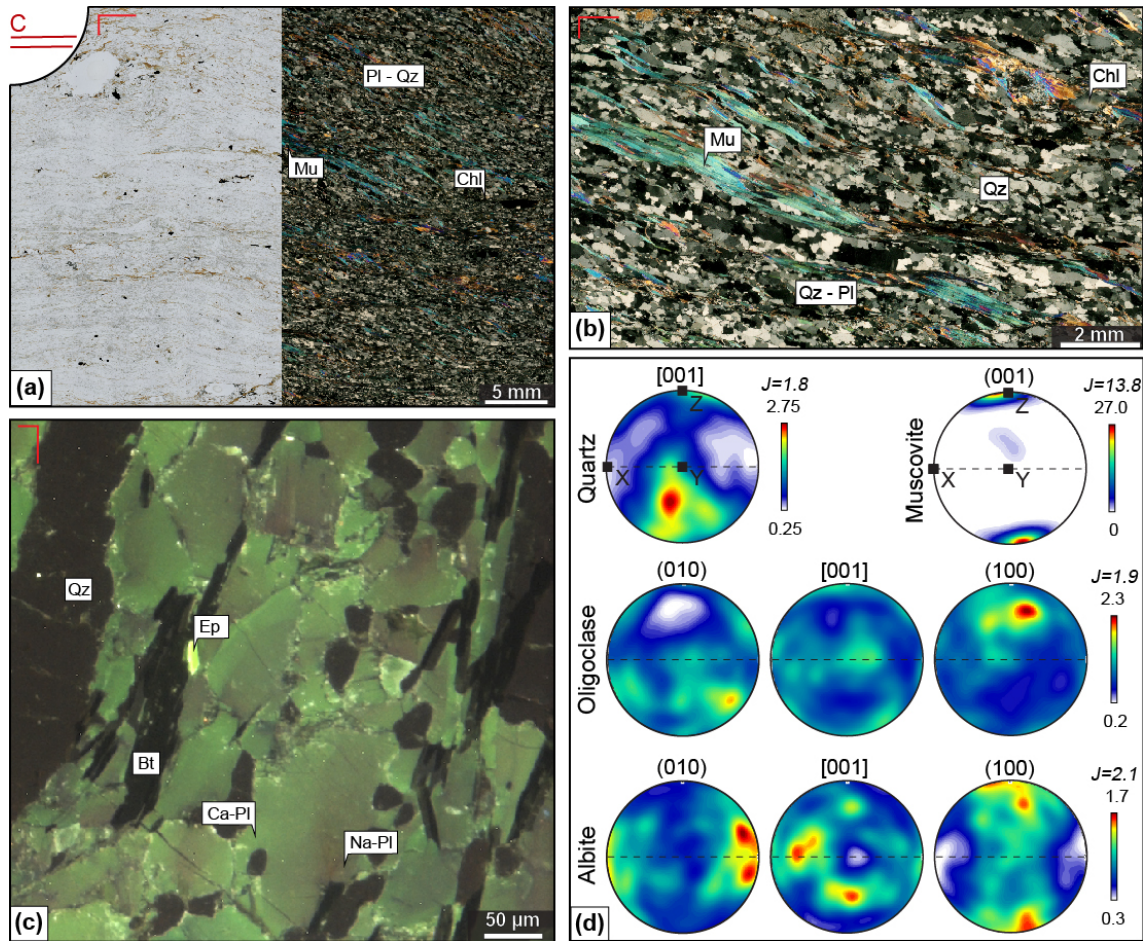


Figure 5: (*2-column fitting image*) Mineralogy and microtextures of the utramylonite (13NEV01D) (a) Plane-polarized (left) and cross-polarized (right) photomicrographs of the thin section. (b-c) Intense metamorphic layering defined by the alternance of muscovite ribbons and mixed plagioclase-quartz domains. (c) CL imaging highlighting the two populations of plagioclase, CL colors traducing the Na-Ca ratio (Na-rich: reddish, Ca-rich: greenish). (d) CPO of quartz, muscovite, Ca-rich (oligoclase) and Na-rich (albite) plagioclase (lower hemisphere).

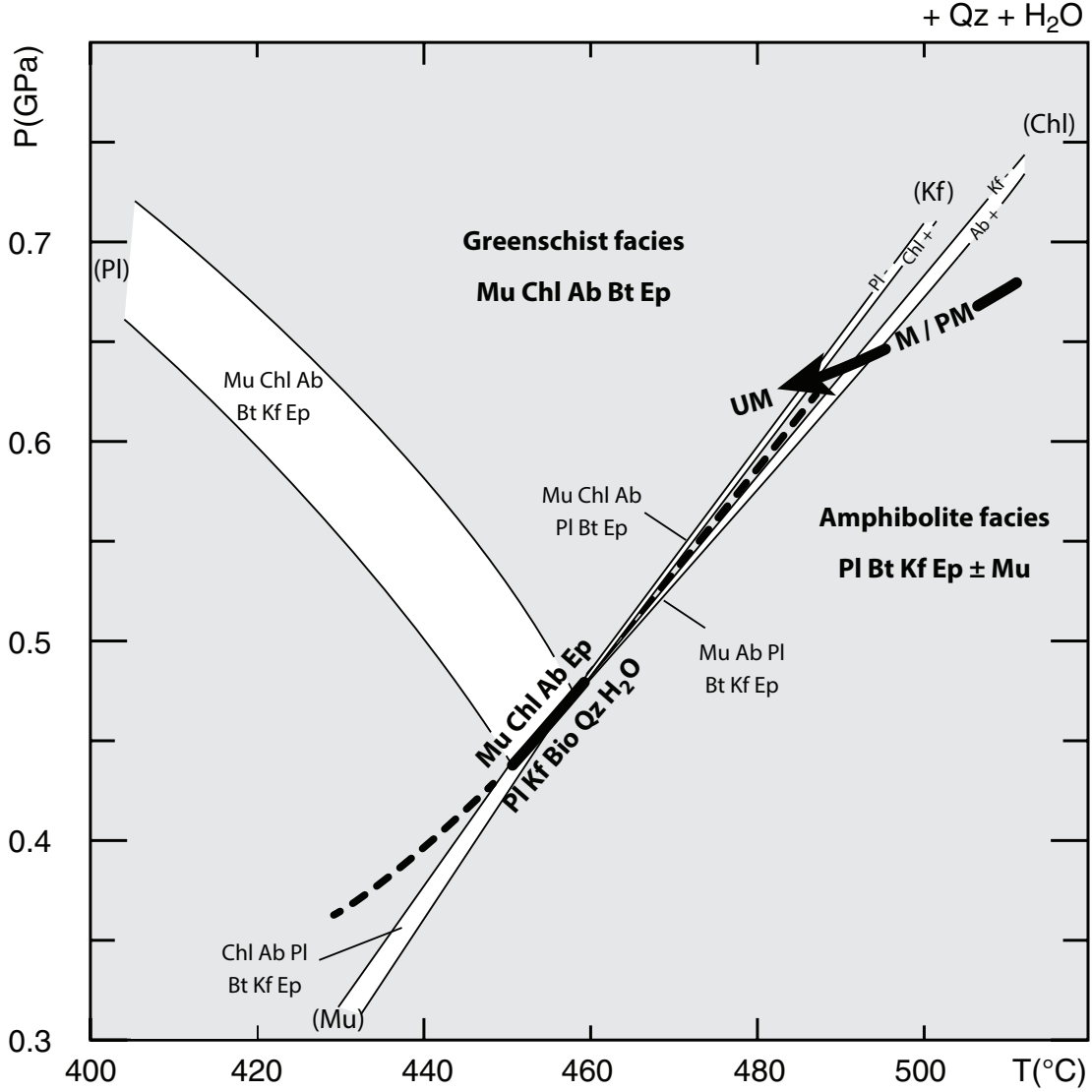


Figure 6: (1-column fitting image) Pressure-temperature conditions of deformation. P-T phase diagram section computed in the NaCaKFMASH system from the bulk chemical composition of the Neves metagranodiorite 13NEV01A ( $SiO_2$  71.43 :  $Al_2O_3$  11.05 : FeO 4.31 : MgO 1.97 : CaO 4.99 :  $Na_2O$  4.79 :  $K_2O$  1.46, in mol %). Divariant equilibria (in white) are named with the missing phase in parenthesis. The black arrow materializes the evolution of the metamorphic conditions during deformation, from the amphibolite-facies deformation related to protomylonites (PM) and mylonite (M) (high-temperature field  $\geq$  500 °C, where Bt, Pl and Kf are stable), to the greenschist-facies ultramylonite (UM, lower-temperature field 450-500 °C, where Mu, Ab and Chl are stable). The transition between the two fields is the univariant reaction represented by the thick black line. The thick dashed line corresponds to the metastable extension of this reaction.

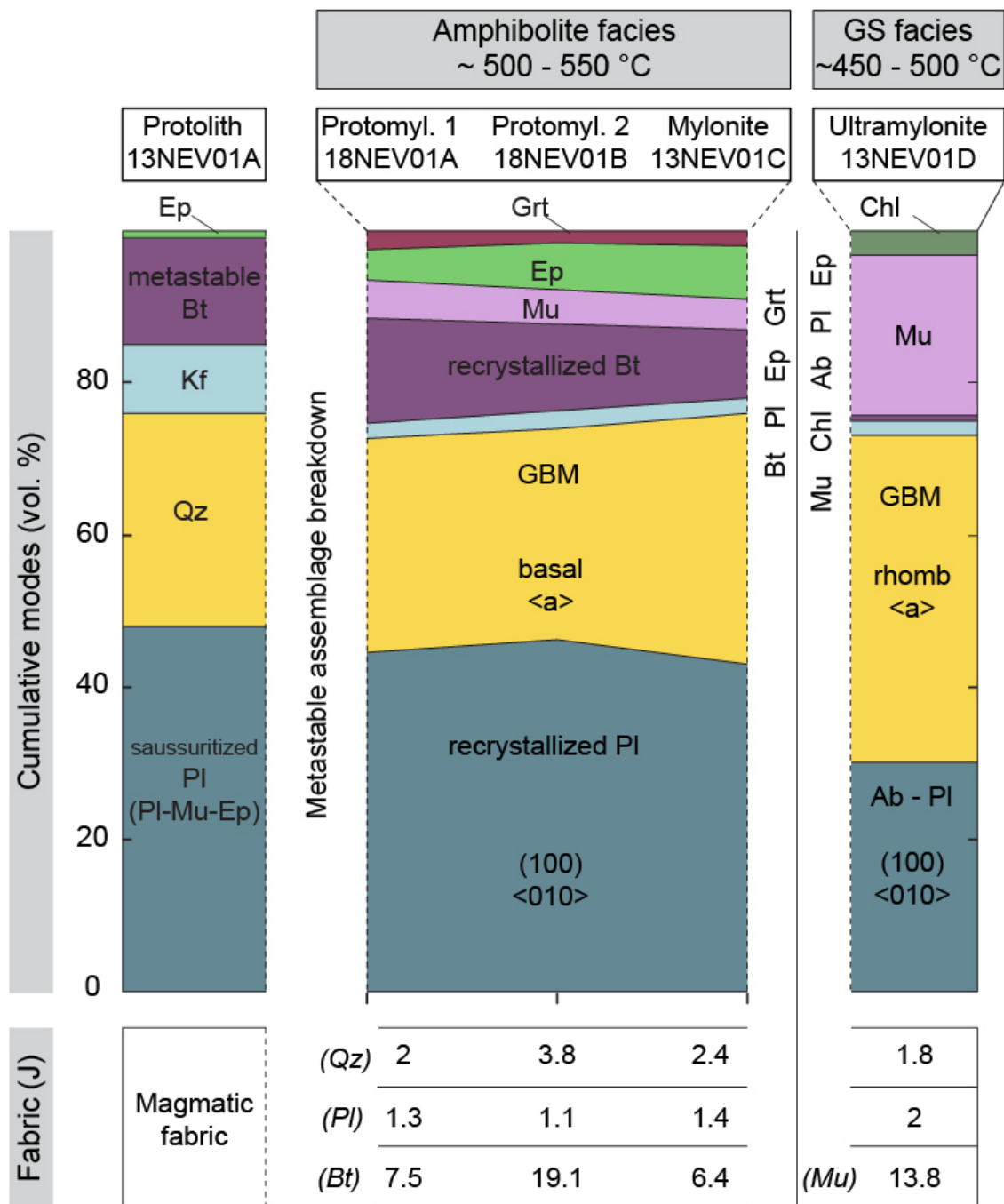


Figure 7: (1-column fitting image) Evolution of cumulative modes and fabrics with strain. The color coding refers to that of the phase maps obtained by EBSD (see Supplementary Material S7).

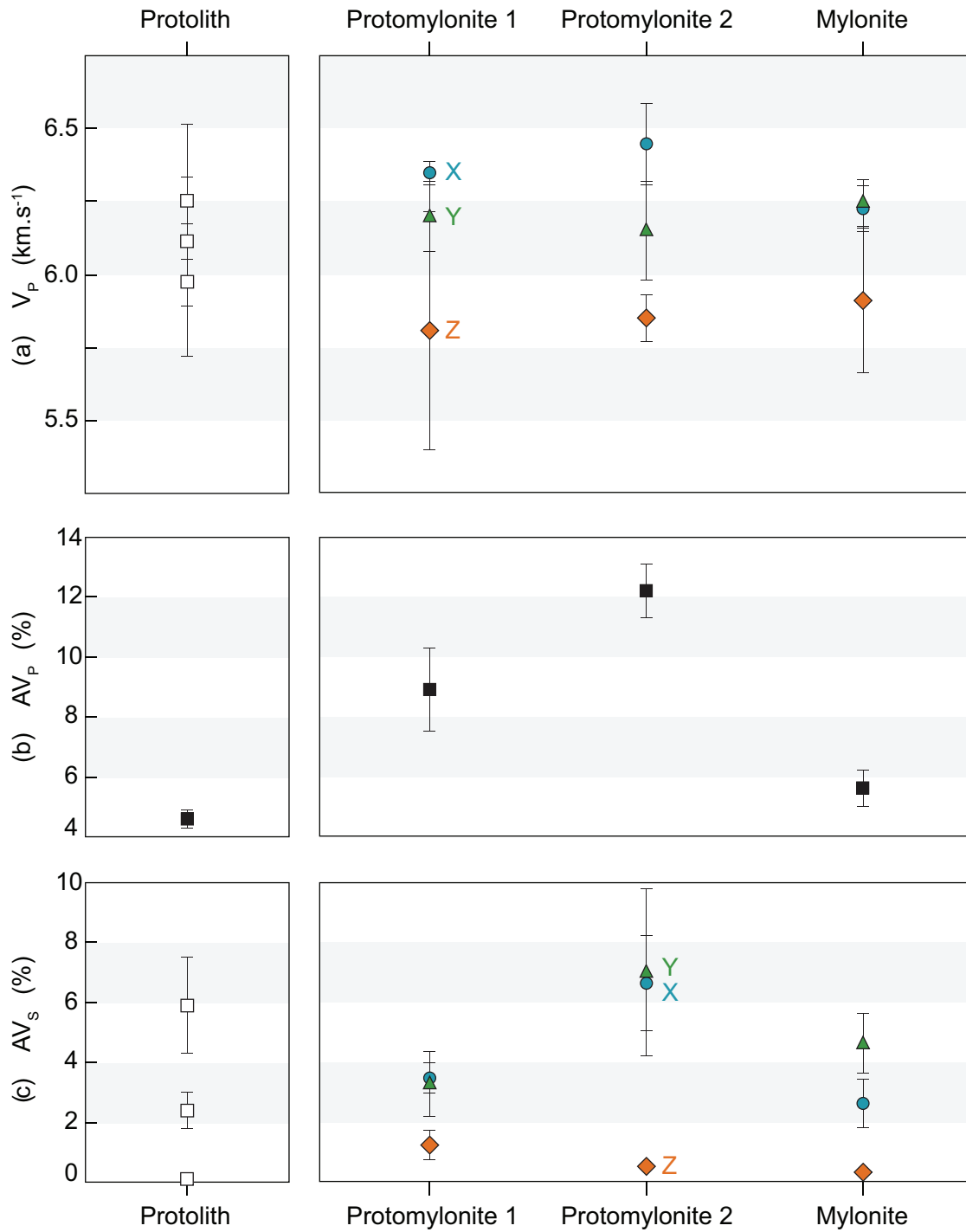
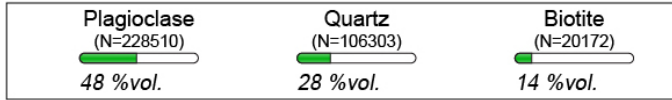


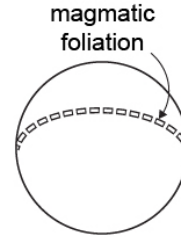
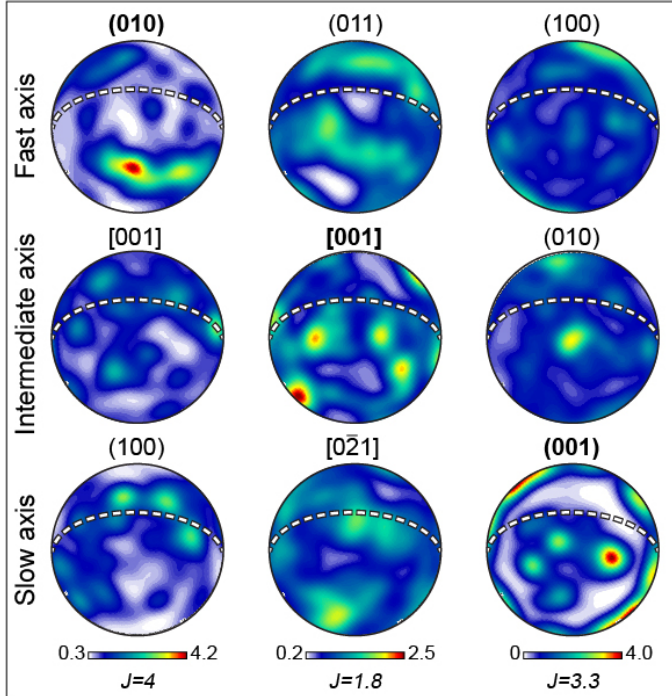
Figure 8: (*2-column fitting image*) Measured  $V_P$  (a),  $AV_P$  (b) and  $AV_S$  (c). The white squares are the results obtained in the protolith, along undifferentiate directions. Blue circles, green triangles and orange diamonds are the results obtained along the X-, Y- and Z-axes of the strain ellipsoid, respectively, in the deformed samples. There are no data for the ultramylonite.

**PROTOLITH - 13NEV01A**

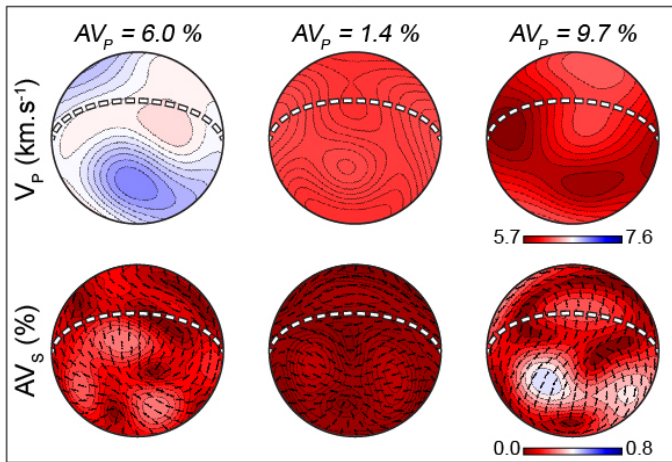
**(a) Modes**



**(b) CPO**



**(c) Monomineralic seismic properties**



**(d) Full aggregate seismic properties**

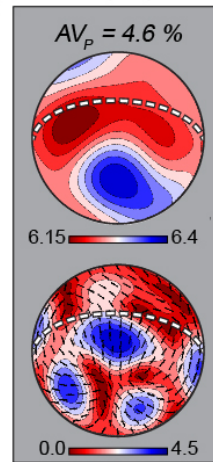
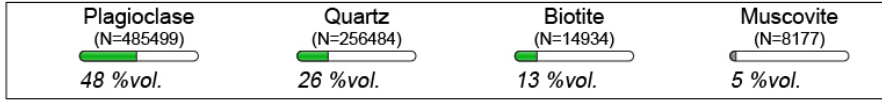


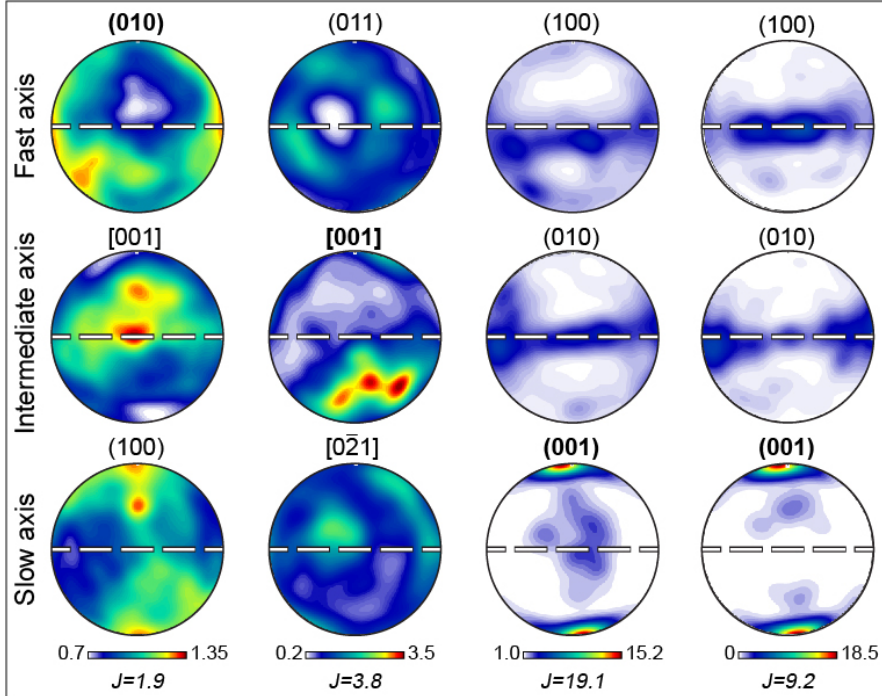
Figure 9: (2-column fitting image) Texture and calculated seismic properties of the protolith 13NEV01A. The dashed white line materializes the magmatic foliation. (a) Modal proportions of the main mineral phases. N is the number of points measured by EBSD. (b) CPO. For each phase, CPO is expressed for the fastest, intermediate and slowest P-wave propagation axes. Phases are ranked by descending order of abundance. Axes in bold font are the main axes described in the text. (c) Calculated seismic properties ( $V_p$ ,  $AV_p$  and  $AV_s$ ) of the monocrystalline aggregates. Stereographic normalized plots represent the spatial distributions of  $V_p$  (top) and  $AV_s$  (bottom). (d) Calculated seismic properties of the whole rock aggregate.

**PROTOMYLONITE 2 - 18NEV01B**

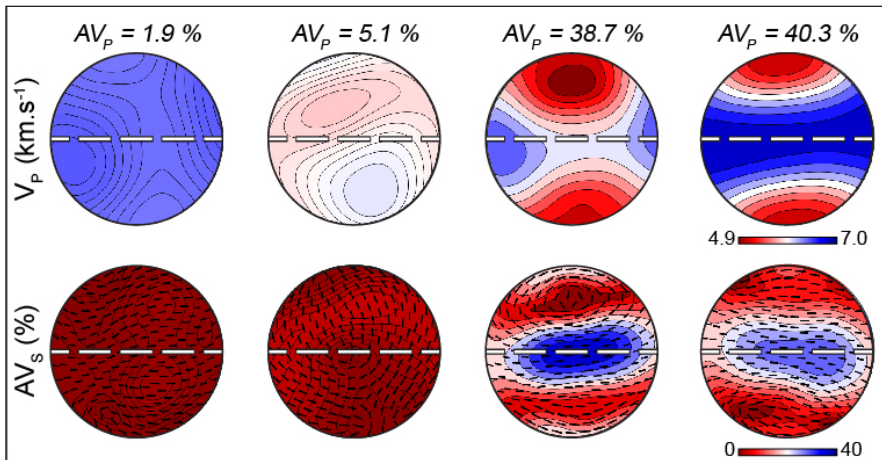
(a) Modes



(b) CPO



(c) Monomineralic seismic properties



(d) Full aggregate seismic properties

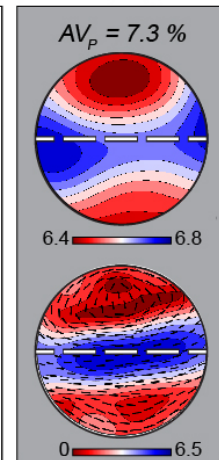
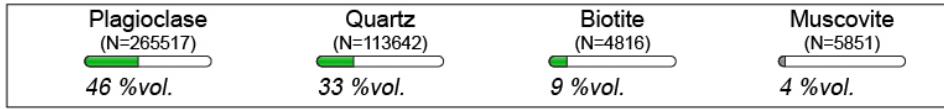


Figure 10: (2-column fitting image) Texture and calculated seismic properties of protomylonite 2 18NEV01B. (a) Modal proportions of the main mineral phases. N is the number of points measured by EBSD. (b) CPO. For each phase, CPO is expressed for the fastest, intermediate and slowest P-wave propagation axes. Phases are ranked by descending order of abundance. Axes in bold font are the main axes described in the text. (c) Calculated seismic properties ( $V_P$ ,  $AV_P$  and  $AV_S$ ) of the monocrystalline aggregates. Stereographic normalized plots represent the spatial distributions of  $V_P$  (top) and  $AV_S$  (bottom). (d) Calculated seismic properties of the whole rock aggregate.

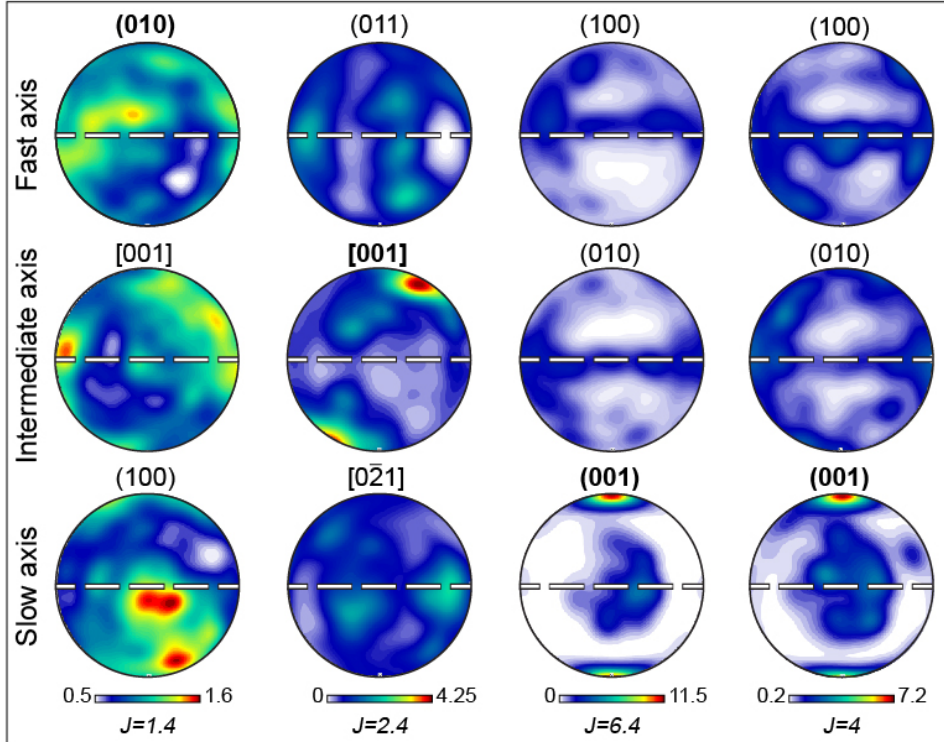


# MYLONITE - 13NEV01C

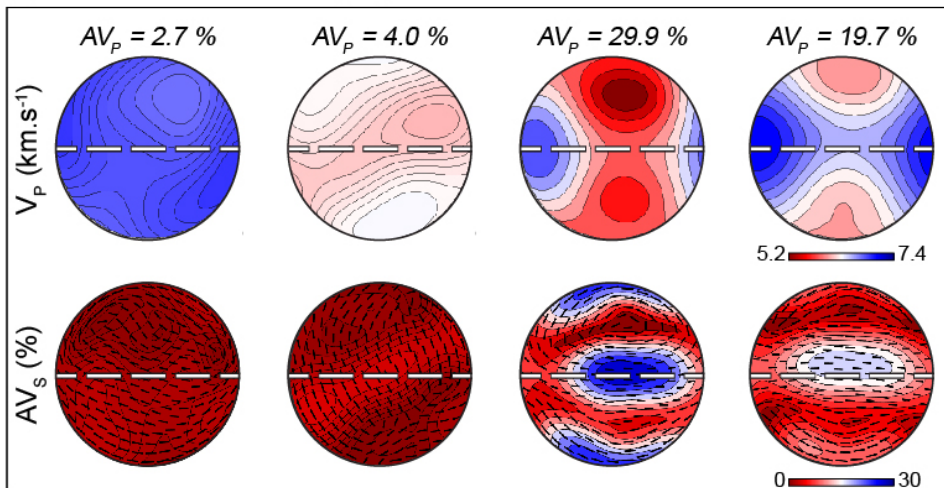
## (a) Modes



## (b) CPO



## (c) Monomineralic seismic properties



## (d) Full aggregate seismic properties

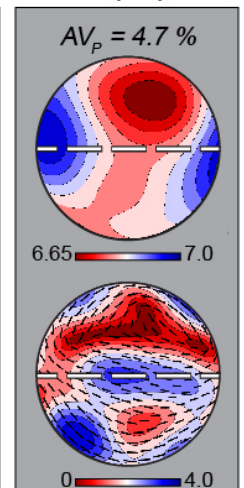
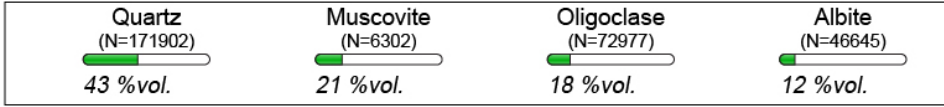


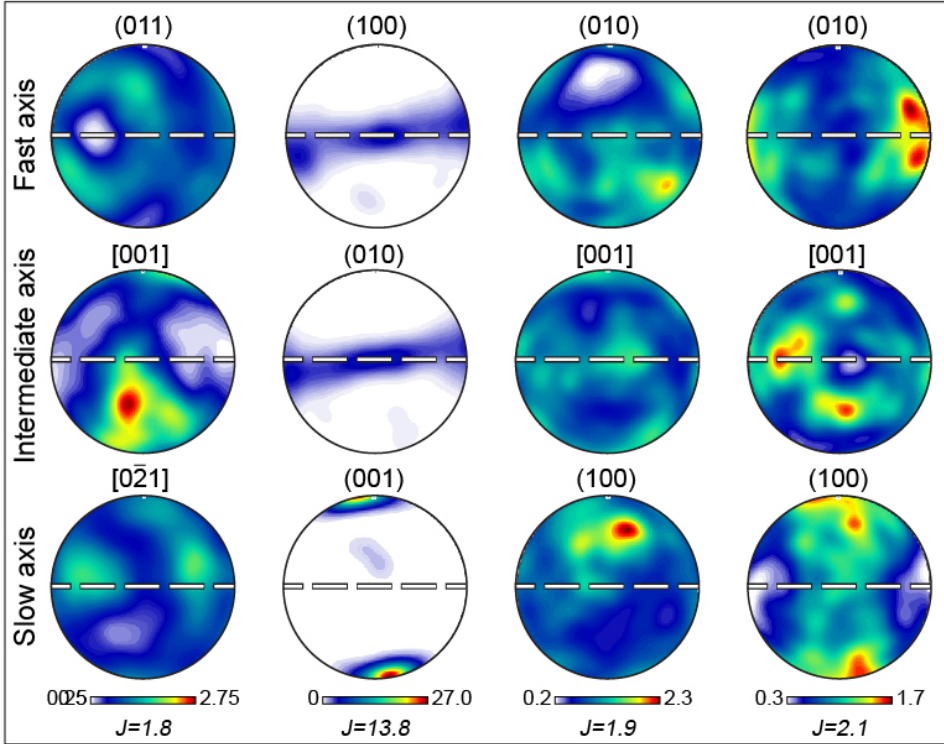
Figure 11: (2-column fitting image) Texture and calculated seismic properties of the mylonite 13NEV01C. (a) Modal proportions of the main mineral phases. N is the number of points measured by EBSD. (b) CPO is expressed for the fastest, intermediate and slowest P-wave propagation axes. Phases are ranked by descending order of abundance. Axes in bold font are the main axes described in the text. (c) Calculated seismic properties ( $V_p$ ,  $AV_p$  and  $AV_s$ ) of the monocrystalline aggregates. Stereographic normalized plots represent the spatial distributions of  $V_p$

# ULTRAMYLONITE - 13NEV01D

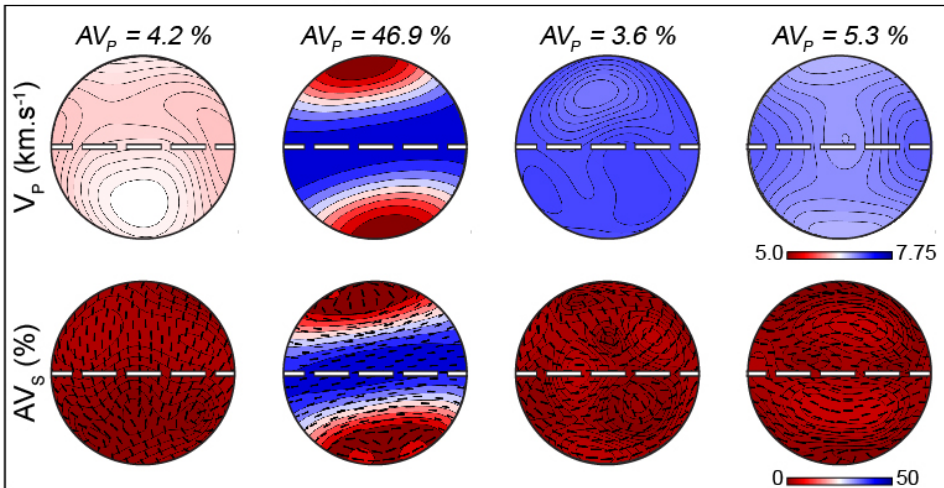
## (a) Modes



## (b) CPO



## (c) Monomineralic seismic properties



## (d) Full aggregate seismic properties

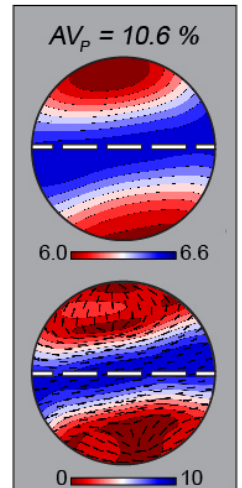


Figure 12: (2-column fitting image) Texture and calculated seismic properties of the ultramylonite 13NEV01D. (a)

Modal proportions of the main mineral phases. N is the number of points measured by EBSD. (b) CPO. For each phase, CPO is expressed for the fastest, intermediate and slowest P-wave propagation axes. Phases are ranked by descending order of abundance. Axes in bold font are the main axes described in the text. (c) Calculated seismic properties ( $V_p$ ,  $AV_p$  and  $AV_s$ ) of the monocrystalline aggregates. Stereographic normalized plots represent the spatial distributions of

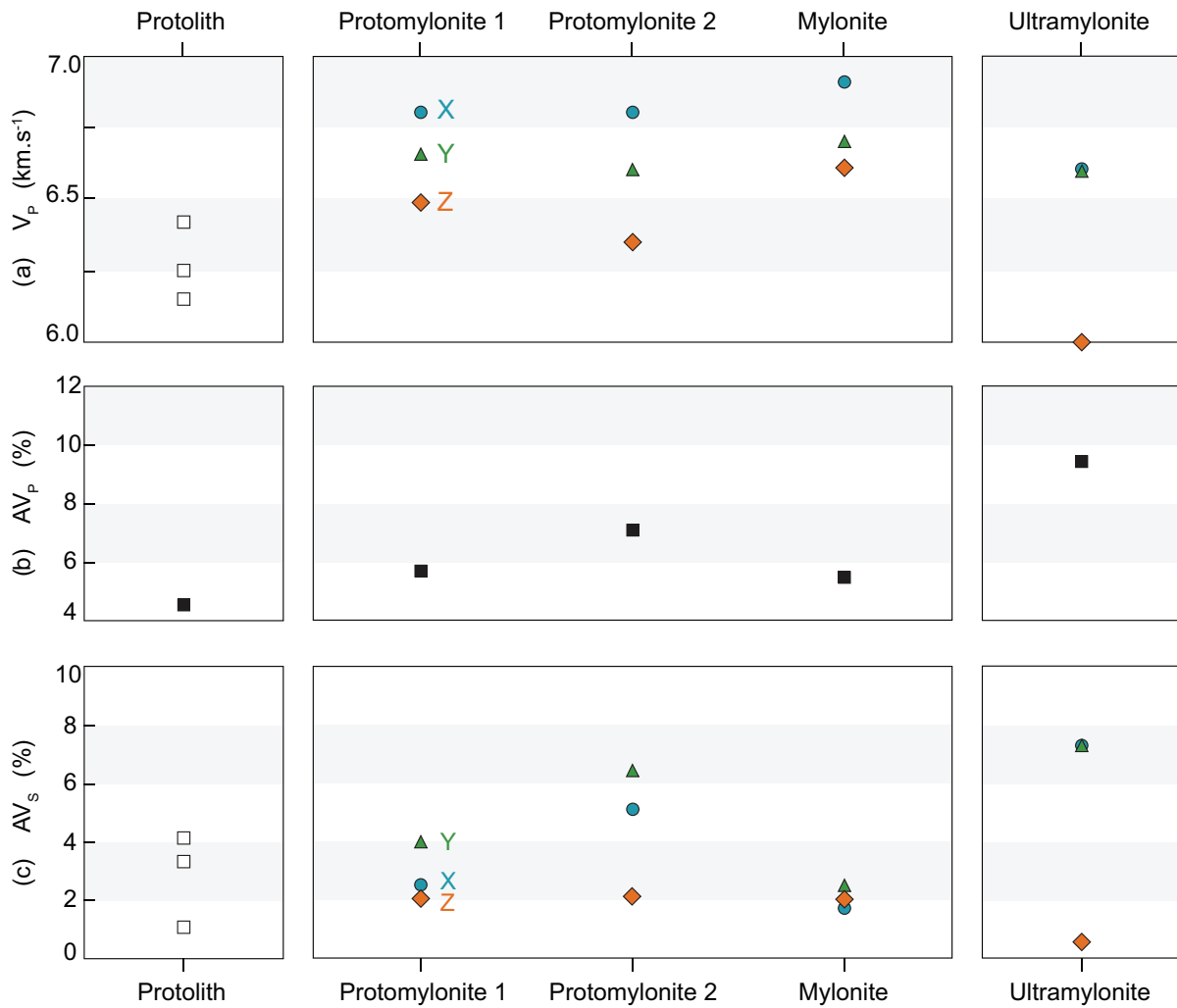


Figure 13: (*2-column fitting image*) Calculated  $V_P$  (a),  $AV_P$  (b) and  $AV_S$  (c). The white squares are the results obtained in the protolith, along undifferentiate directions. Blue circles, green triangles and orange diamonds are the results obtained along the X-, Y- and Z-axes of the strain ellipsoid, respectively, in the deformed samples.

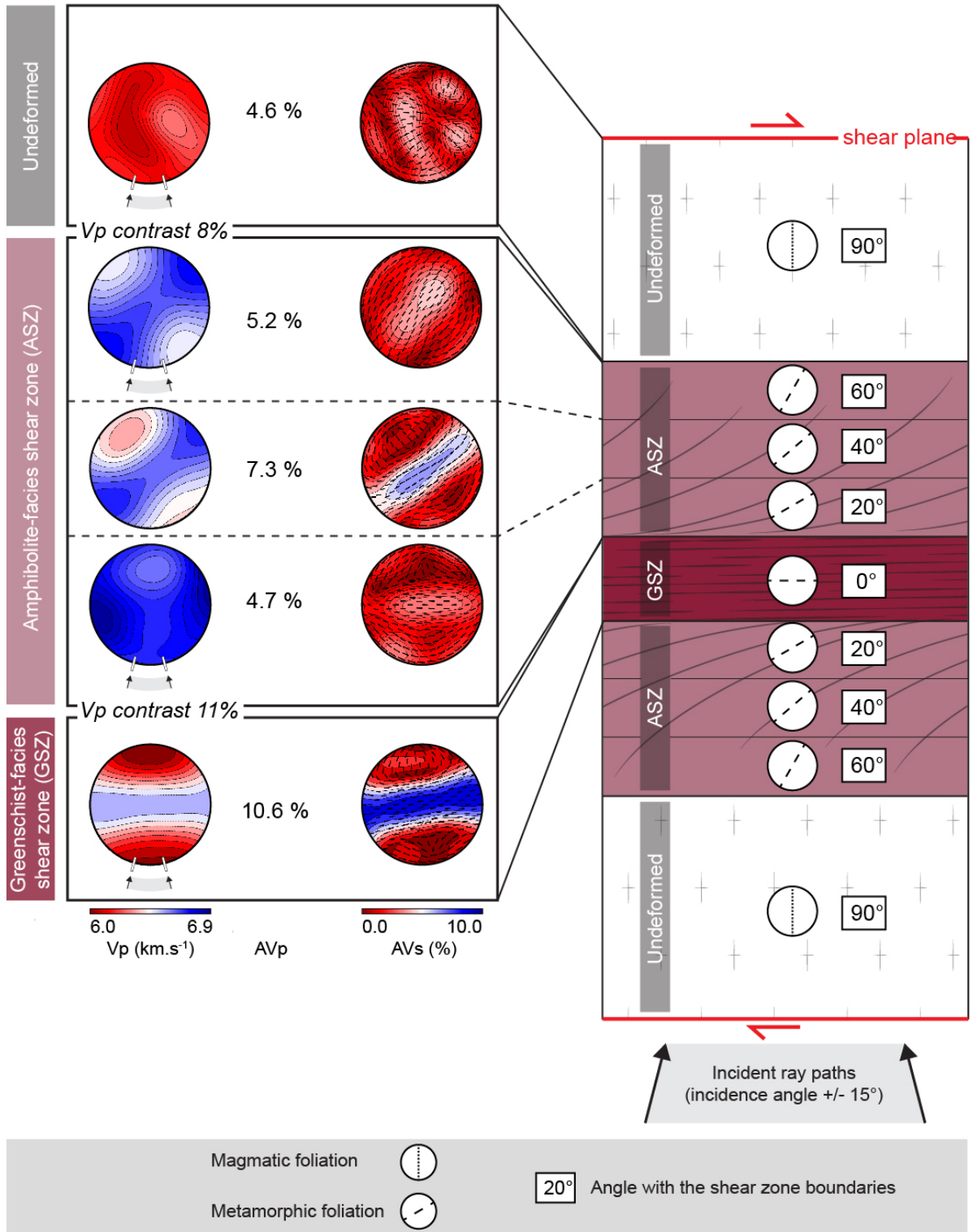


Figure 14: (2-column fitting image) Simplified model of the dextral meter-scale shear zone. On the right, the circular plots and the angle value represent the orientation of the magmatic/metamorphic foliation. On the left, the pole figures represent the normalized spatial distributions of  $V_P$  and  $AV_S$  for each stage, which have been rotated following the orientation of the foliation.  $AV_P$  is indicated in percent. Teleseismic P-waves propagating through the horizontal shear zone are represented below the model and the circular plots, as a cone of incident ray paths at  $15^\circ$  on either side of the

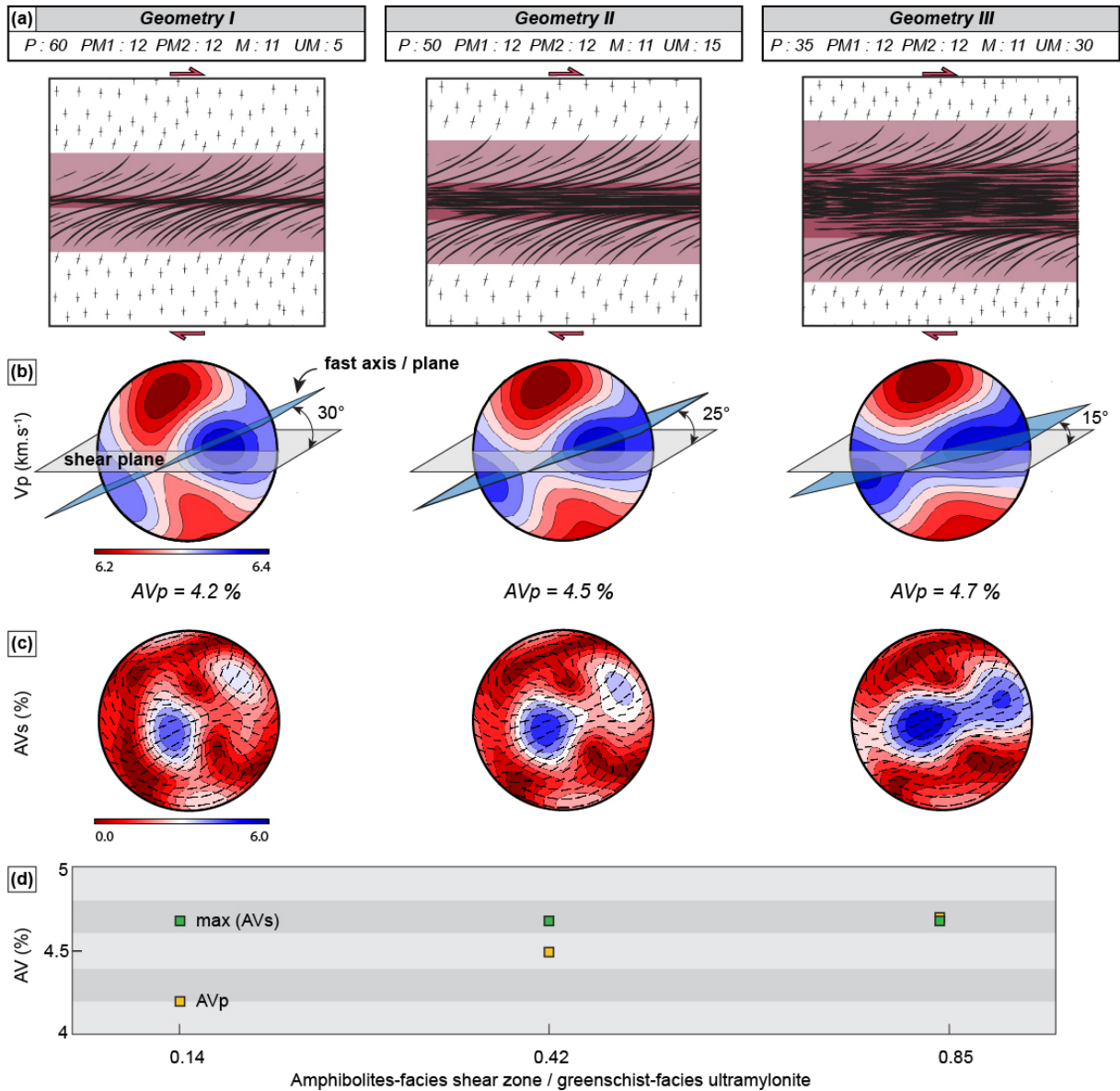


Figure 15: (*2-column fitting image*) Calculated seismic properties of a whole shear zone. (a) Relative proportions of the protolith, the amphibolite-facies marginal zone and the greenschist-facies ultramylonite zone of the three modelled geometries, illustrated in the cartoons. (b) Spatial distributions of  $V_p$ . The angle between the C-plane and the fast P-wave propagation plane is represented. (c) Spatial distributions of  $AV_s$ . (d) Evolution of  $AV_p$  and  $AV_s(max)$  with the amphibolite versus greenschist shear zone ratio.

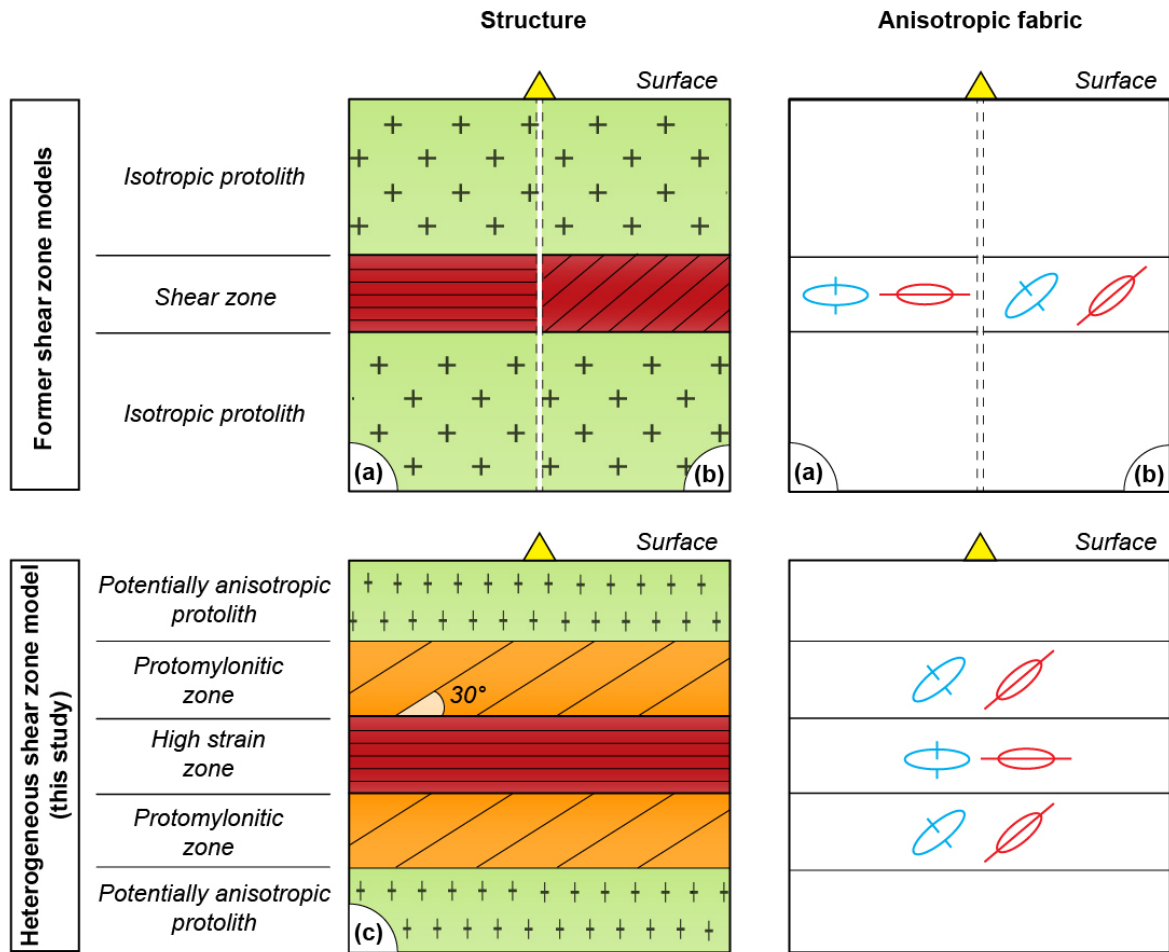
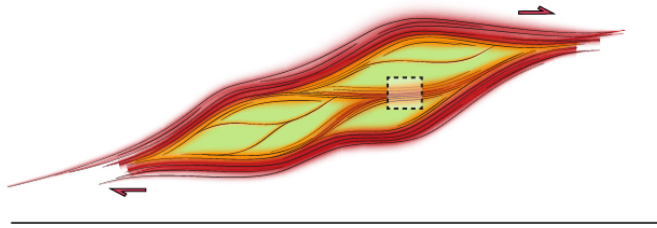


Figure 16: (1.5-column fitting image) Model of spatial variability of the seismic properties across a crustal shear zone. (a-b) Former geophysical models for shear zones, consisting in an homogeneous anisotropic layer with the fast plane (a) parallel (e.g. Schulte-Pelkum & Mahan, 2014) or (b) oblique (e.g. Schulte-Pelkum et al., 2005; Zandt et al., 2004) to shear zone boundaries. (c) New model (this study), taking both into account the oblique fabric of the marginal zone with an orthorhombic symmetry and the parallelized fabric of the ultramylonite zone with a hexagonal symmetry. Left-side cartoons represent the structure of the models, right-side cartoons, the orientation of the anisotropic fabric.

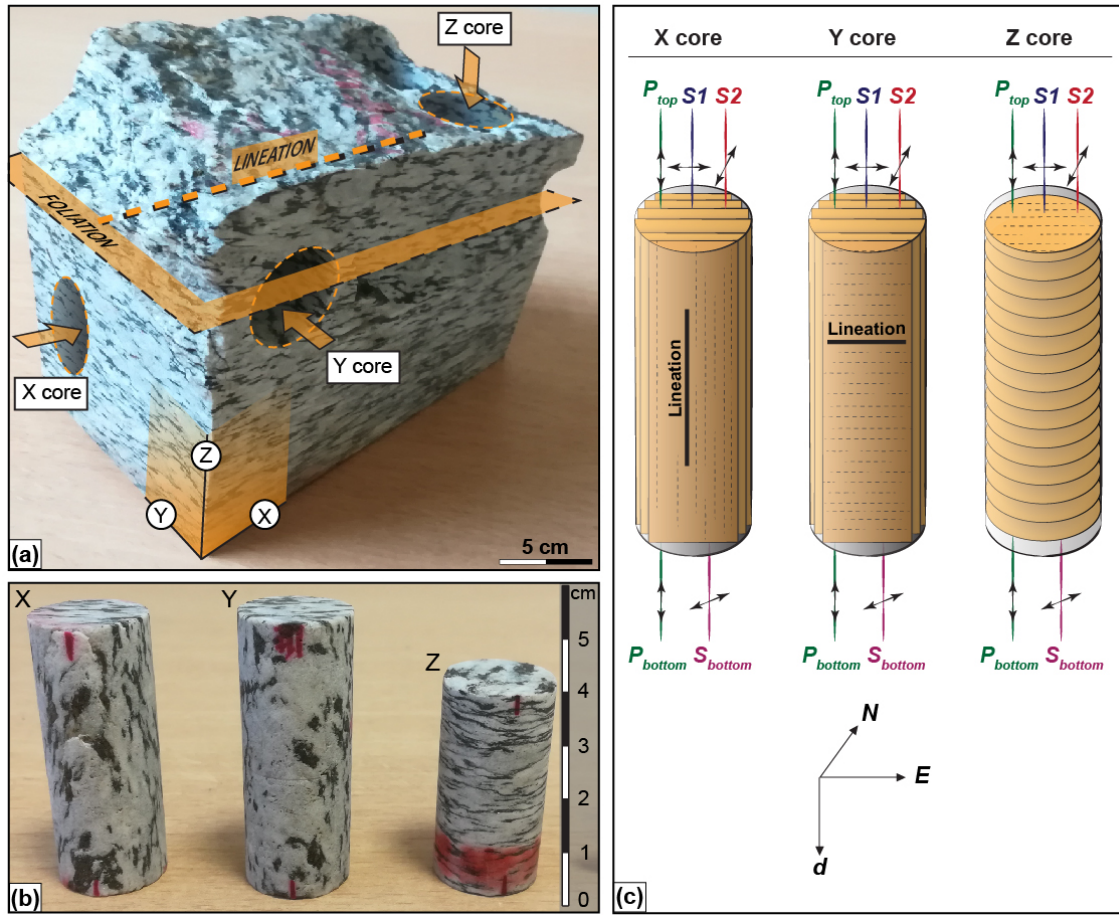


Figure 17: (2-column fitting image) (a) Drilling directions within a homogeneously deformed sample : X (along the lineation), Y (parallel to the foliation and perpendicular to the lineation) and Z (normal to the foliation). (b) The three cores, X, Y and Z. (c) Orientation of the ductile structure in the X, Y and Z cores, and its position with respect to the polarization of the piezoelectric transducers. The orange plane is the foliation plane, the dashed line is the lineation. Acoustic wave are emitted by  $P_{bottom}$  and  $S_{bottom}$  and received by  $P_{top}$ ,  $S_1$  and  $S_2$ . The  $P_{bottom}$  and  $P_{top}$  transducers are both polarized along  $d$  (depth). The  $S_{bottom}$ ,  $S_1$  and  $S_2$  transducers are polarized towards N45°E, E (East) and N (North), respectively.

**S2. Pulsed-receiver settings**

<b>Receiver</b>	
Bandwidth (MHz)	35
Gain (dB)	10
High pass filter	Out
Low pass filter	35

<b>Pulser</b>	
Type	DPR300
Trigger	Internal
Damping (Ohms)	331
Energy	4
PRF (kHz)	100
Echo/Through mode	Through
Voltage (V)	207
Impedance	High



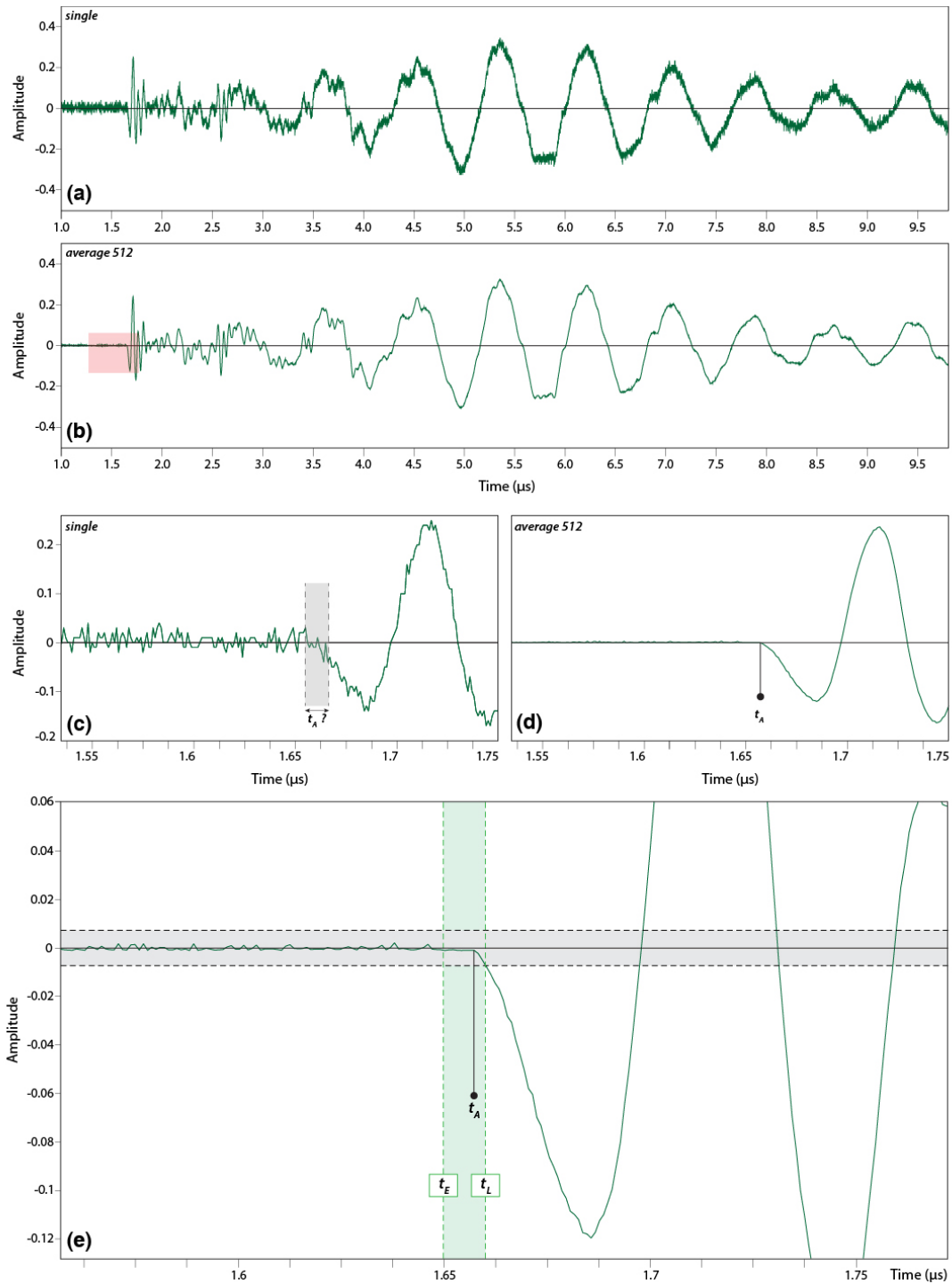


Figure 18: S2. (a-d) Effect of the acquisition method on the signal-to-noise ratio (case of a P wave propagating through a mylonite sample). (a-c) "Single" mode: signal obtained from a single electric pulse. (b-d) "Average" mode: average of the signals generated by 512 electric pulses. (c) and (d) are zooms of (a) and (b), respectively, near the seismic phase arrival. The red square is the location of the time window used in for the picking of the seismic phase arrival. (e) Picking method. The dashed black lines are the ASNR (Amplitude-based Signal-to-Noise Ratio) thresholds, defining a grey band in which the earliest ( $t_E$ ), latest ( $t_L$ ) and most likely ( $t_A$ ) arrival times are picked.  $t_E$  is the last point with a null derivative,  $t_L$  is the intersection between the signal and the ASNR threshold.  $t_A$  is picked within this time window, at the basis of the first inflection.

**S4. Analytical settings of EBSD mapping**

<b>Sample</b>	<b>13NEV01A</b>	<b>18NEV01A</b>	<b>18NEV01B</b>	<b>13NEV01C</b>	<b>13NEV01D</b>
Resolution (width) (px)	908	1009	1357	955	532
Resolution (height) (px)	682	695	849	752	750
Step size (μm)	33.3	10.6	12.4	10.1	10.6
Field width (mm)	30.2	10.7	16.8	9.62	5.64
Field height (mm)	22.7	7.37	10.5	7.57	7.95
Number of points	619256	701255	1152093	718160	399000
Hit rate (%)	70.1	77.0	79.8	67.1	78.3
Accelerating voltage (kV)			15.00		
Working distance (mm)	27.3	23.7	21.1	16.7	14.7
Detector insertion distance (mm)	169.1	169.1			
Reanalysis time (h:m:s)	00:10:16	00:08:54	00:18:07	00:11:05	00:06:00
Speed of reanalysis (Hz)	782.8	1091.9	922.8	980.4	1064.2
Specimen tilt (°)			0		
EBSD camera binning mode (px x px)			336x256		
EBSD camera exposure time (ms)	11.7	5.1	3.0	4.6	2.6
EBSD camera gain			1		
Static background correction			Activated		
Auto background correction			Activated		
Ban detection mode			Centers		
Number of bands detected			8		
Indexing mode			Optimized-TKD		

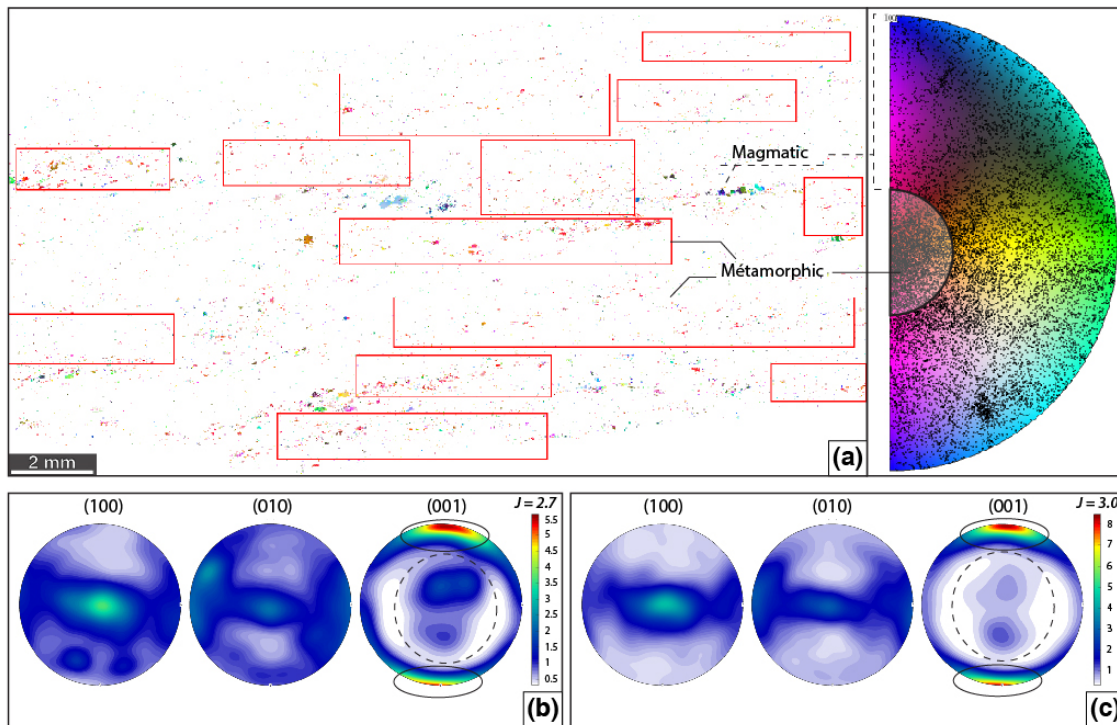


Figure 19: S4. Strategy for phyllosilicates indexing by EBSD. (a) Selection of regions of interest on muscovite EBSD mapping. The legend on the right correspond to the inverse pole figure showing the Y axis (perpendicular to lineation) orientation in the muscovite crystal referential. CPO of (100), (010) and (001) of muscovite, before (b) and after (c) the selection of regions of interest.

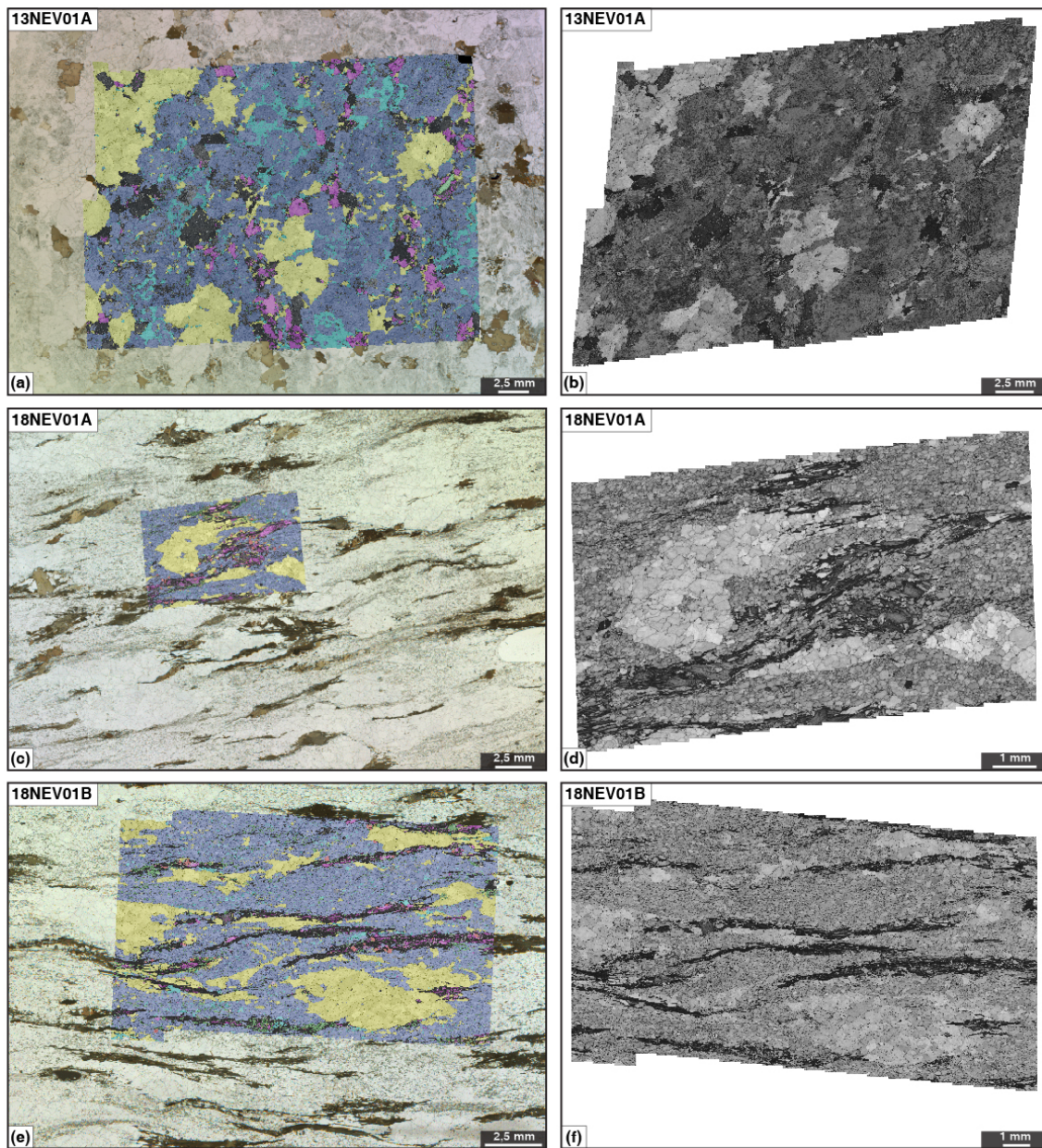


Figure 20: S5. Location of the EBSD map on the thin section (left) and Band contrast map (right) of the protolith 13NEV10A (a-b), protomylonite 1 18NEV01A (c-d), protomylonite 2 18NEV01B (e-f). Yellow: quartz, dark blue: plagioclase, light blue: K-feldspar, light purple: muscovite, dark purple: biotite, red: garnet, light green: epidote, dark green: chlorite.

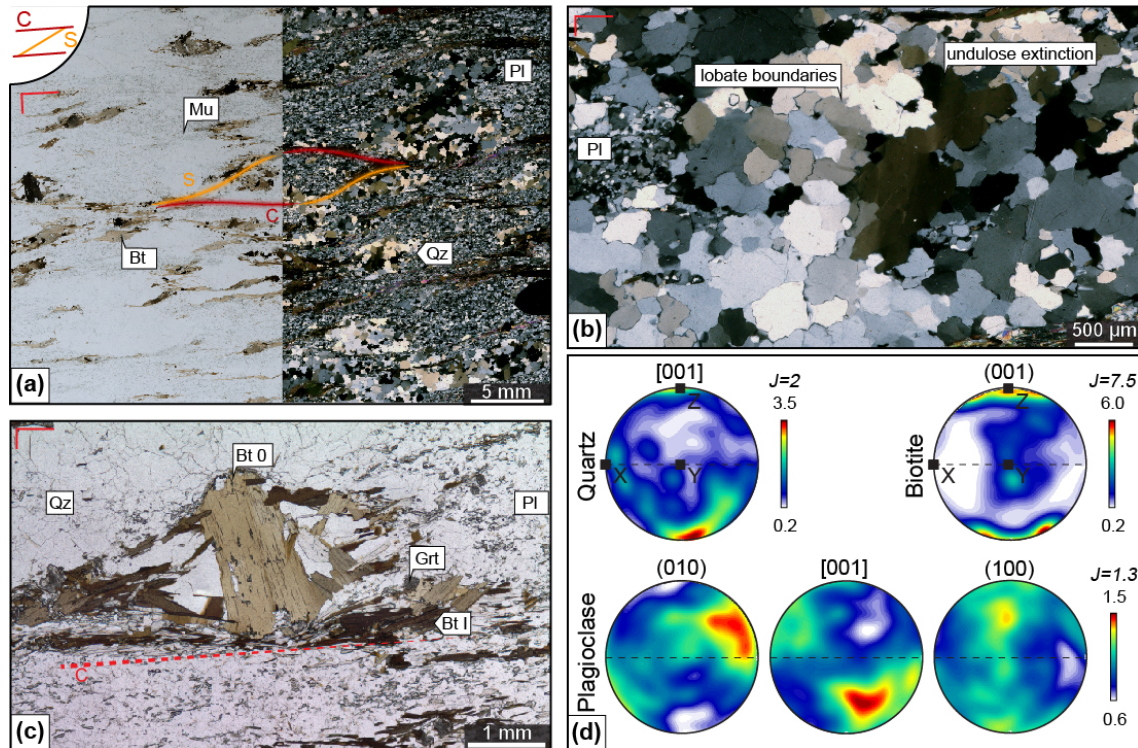


Figure 21: (*2-column fitting image*) Mineralogy and microtextures of protomylonite 1 (18NEV01A) (a) Plane-polarized (left) and cross-polarized (right) photomicrographs of the thin section. The sigmoid highlights the C- and S-planes, in red and orange, respectively. The red symbols indicate the orientation of X- (long) and Z- (short) axes. (b) Quartz microstructures. (c) Relict of  $Bt_0$  preserved in a sigmoid defined by metamorphic  $Bt_1$ . Grt: garnet, Mu: muscovite. (d) CPO of quartz, plagioclase and biotite (lower hemisphere). Fabric strength is indicated by the J-index.

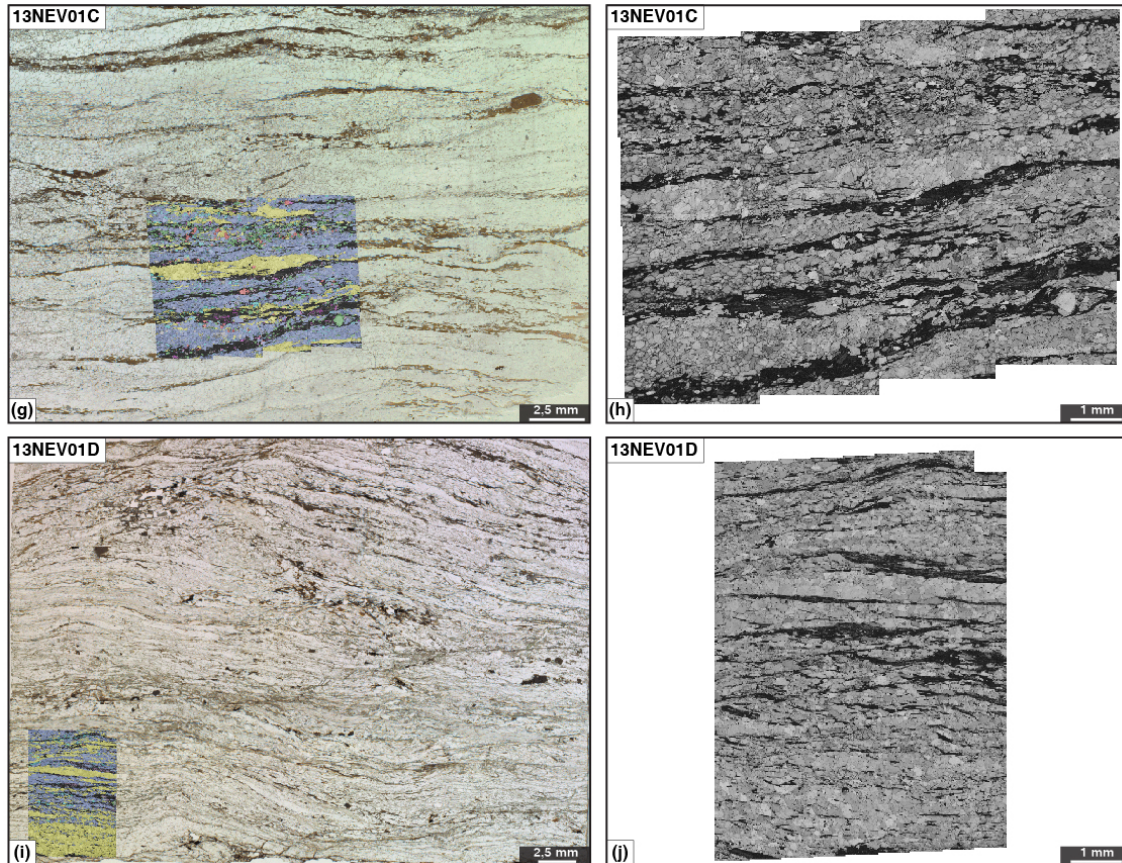
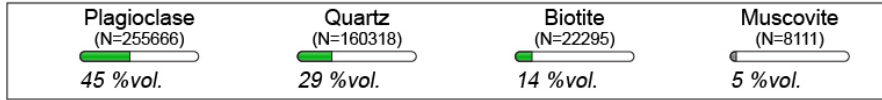


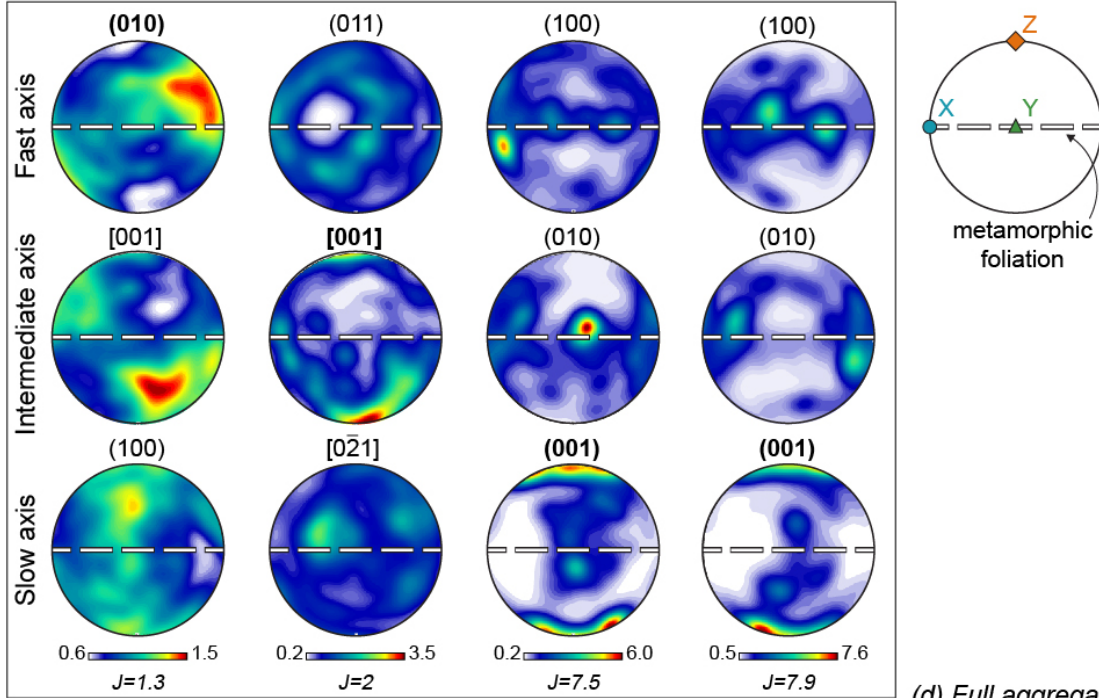
Figure 22: S5.(following) Location of the EBSD map on the thin section (left) and Band contrast map (right) of the mylonite 13NEV01C (g-h) and the ultramylonite 13NEV01D (i-j). Yellow: quartz, dark blue: plagioclase, light blue: K-feldspar, light purple: muscovite, dark purple: biotite, red: garnet, light green: epidote, dark green: chlorite.

**PROTOMYLONITE 1 - 18NEV01A**

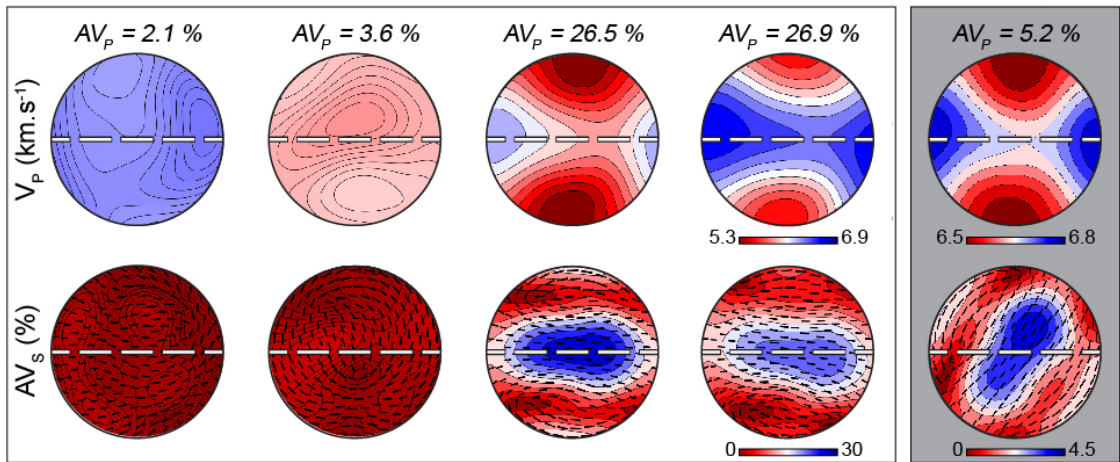
(a) Modes



(b) CPO



(c) Monomineralic seismic properties



(d) Full aggregate seismic properties

Figure 23: (*2-column fitting image*) Texture and calculated seismic properties of protomylonite 1 18NEV01A. (a) Modal proportions of the main mineral phases. N is the number of points measured by EBSD. (b) CPO. For each phase, CPO is expressed for the fastest, intermediate and slowest P-wave propagation axes. Phases are ranked by descending order of abundance. Axes in bold font are the main axes described in the text. (c) Calculated seismic properties ( $V_p$ ,  $AV_p$  and  $AV_s$ ) of the monocrystalline aggregates. Stereographic normalized plots represent the spatial distributions of  $V_p$  (top) and  $AV_s$  (bottom). (d) Calculated seismic properties of the whole rock aggregate.

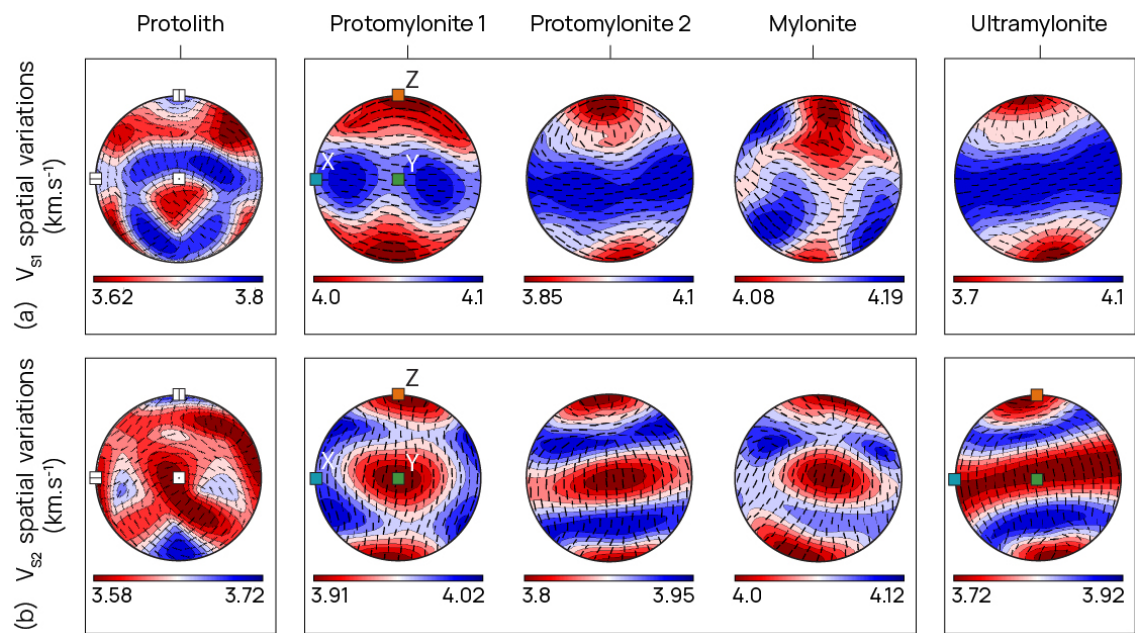


Figure 24: S6. Spatial distributions of  $V_{S1}$  and  $V_{S2}$  of the rock aggregates, calculated from petrofabrics.



**S10.** Bulk elastic tensors ( $C_{ij}$ ) calculated by EBSD for the samples of the strain gradient developed in the Neves metagranodiorite, and additional informations (lithology, P-T conditions, step size of EBSD measurements).

Sample	Rock type	Metamorphic grade	Location	$C_{ij}$						Step size ( $\mu\text{m}$ )	Density	P (GPa)	T ( $^{\circ}\text{C}$ )
13NEV01A	Weakly deformed metagranodiorite	Amphibolite	Zillertäl massive (Tauern Window, E. Alps)	107.45	30.87	31.63	0.03	-0.94	0.37	33.3	2.8	0.5-0.8	500-550
				30.87	105.65	30.63	0.49	-0.28	0.08				
				31.62	30.63	105.59	-2.41	-2.2	0.54				
				0.03	0.49	-2.41	35.88	0.97	0.16				
				-0.94	-0.28	-2.2	0.97	36.49	0.03				
0.37	0.08	0.56	0.16	0.03	35.63								
18NEV01A	Protomylonite 1			129.74	32.74	32.28	0.42	-0.08	0.55	10.6			
				32.74	117.15	32.03	-0.22	-0.19	0.26				
				32.28	32.03	124.17	-0.26	0.39	-0.14				
				0.42	-0.22	-0.26	42.81	0.41	0.06				
				-0.08	-0.19	0.39	0.41	46.66	0.22				
0.55	0.26	-0.14	0.06	0.22	44.65								
18NEV01B	Protomylonite 2			128.27	31.97	32.95	0.09	0.63	1.5	12.4			
				31.97	114.32	31.25	1.21	0.18	-0.02				
				32.95	31.25	123.93	1.92	0.18	0.46				
				0.09	1.21	1.92	40.37	0.6	0.22				
				0.63	0.18	0.18	0.6	46.1	1.21				
1.5	-0.02	0.46	0.22	1.21	42.13								
13NEV01C	Mylonite			133.47	32.6	33.77	0.15	1.33	-1.21	10.1			
				32.6	127.44	33.16	1.16	-0.04	-0.59				
				33.77	33.16	126.84	0.72	0.6	-0.24				
				0.15	1.16	0.72	45.15	-0.72	0.4				
				1.33	-0.04	0.6	-0.72	47.7	0.7				
-1.21	-0.59	-0.24	0.4	0.7	46.89								
13NEV01D	Ultramylonite	Upper Greenschist		120.14	24.35	27.07	-0.1	0.42	3.27	10.6			450-500
				24.35	99.81	23.59	1.36	-0.17	0.46				
				27.07	23.59	121.08	0.15	0.04	0.56				
				-0.1	1.36	0.15	39.14	1.28	-0.01				
				0.42	-0.17	0.04	1.28	46.91	0.5				
3.27	0.46	0.56	-0.01	0.5	39.48								

**S10 (continued).** Numer of points indexed by EBSD (N) and modal proportions of each phase used in tensor calculation. N.B. : modal proportions are estimated by combining EBSD reports, X-ray mapping, cathodoluminescence imaging, since the number of points indexed by EBSD is often underestimated, especially for phyllosilicates.

Sample	Albite		Biotite		Epidote		Garnet		Jadeite		K-feldspar		Oligoclase		Quartz		White mica	
	N	Mode (%)	N	Mode (%)	N	Mode (%)	N	Mode (%)	N	Mode (%)	N	Mode (%)	N	Mode (%)	N	Mode (%)	N	Mode (%)
13NEV01A	-	-	20172	14	1754	1	-	-	-	-	39678	9	228510	48	106303	28	-	-
18NEV01A	-	-	22295	14	20351	4	8545	2	-	-	10252	2	255666	45	160318	29	8111	5
18NEV01B	-	-	14934	13	38519	4	3947	1	-	-	26964	3	485499	48	256484	26	8177	5
13NEV01C	-	-	265517	9	30889	7	7521	2	-	-	18389	3	265517	46	113642	33	5851	4
13NEV01D	72977	12	-	-	-	-	-	-	-	-	9530	3	72977	18	171902	43	6302	21