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1 **Subduction zone intermediate-depth seismicity: Insights from the structural analysis of**
2 **Alpine high-pressure ophiolite-hosted pseudotachylyte (Corsica, France).**

3
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10 **Highlights**

11 Alpine Corsica ophiolite pseudotachylyte formed in a subducting oceanic lithosphere

12 Displacement sense associated with pseudotachylyte fault veins is determined

13 Corsican ruptures are similar to seismic ruptures in Pacific plate beneath NE Japan

14 **ABSTRACT**

15 Pseudotachylyte in the Cima di Gratera ophiolite, Alpine Corsica, is distributed in the
16 peridotite unit and in the overlying metagabbro unit and was formed under blueschist to
17 eclogite metamorphic facies conditions, corresponding to a 60-90 km depth range. Peridotite
18 pseudotachylyte is clustered in fault zones either beneath the tectonic contact with overlying
19 metagabbros or at short distance from it. Fault zones are either parallel to the contact or make
20 an angle of 55° to it. Displacement sense criteria associated with fault veins indicate top-to-
21 the-west or top-to-the-northwest reverse senses. Cataclasite flanking most veins was formed
22 before or coevally with frictional melting and likely mechanically weakened the peridotite,
23 facilitating subsequent seismic rupture. In the basal part of the metagabbro unit, post-
24 mylonitization pseudotachylyte can be distinguished from pre-mylonitization pseudotachylyte
25 formed earlier. In the equant metagabbro above the mylonitic sole, only one episode of
26 pseudotachylyte formation can be identified. Kinematics associated with metagabbro
27 pseudotachylyte remain unknown. The geometry and kinematics of the pseudotachylyte veins
28 from the peridotite unit and to a lesser extent from the metagabbro unit are similar to modern

29 seismic ruptures of the upper parts of the Wadati-Benioff zones such as in the Pacific plate
30 beneath NE Japan.

31 KEYWORDS

32 Pseudotachylyte, intermediate-depth seismicity, peridotite, metagabbro, subduction, Alpine Corsica

33 1. Introduction

34 Subduction zone seismicity consists mainly of shallow 'megathrust-type' earthquakes (focal
35 depths < 60 km), intermediate-depth earthquakes (focal depths between 60 and 300 km), and
36 deep-focus earthquakes (focal depths > 300 km). Events from the first category typically
37 nucleate and propagate at the interface between the two lithospheric plates, while those from
38 the last two categories nucleate in the crust or the mantle of the subducting slab. In subduction
39 zones, the most destructive earthquakes are those from the first category. Their hypocenters
40 can be shallow, their magnitudes can reach or exceed 8, and they can trigger tsunamis.
41 Though less spectacular, intermediate-depth earthquakes still represent a significant threat
42 because of the proximity of their epicenters with major cities and also because of infrequent
43 but high magnitudes. Events illustrating such hazards include the 1939 Chile earthquake
44 (Frohlich, 2006) or the 2001 El Salvador earthquake (Vallée et al., 2003; Martinez-Diaz et al.,
45 2004).

46 In subduction zones, the hypocenters of intermediate-depth earthquakes tend to be clustered
47 along a so-called Wadati-Benioff seismic zone. In most cases, precise hypocentral locations
48 allow to divide the Wadati-Benioff seismic zone into two sub-zones. The separation between
49 the two sub-zones is comprised between 8 and 30 km and is a function of the age of the
50 subducting plate (Hasegawa et al., 1978a and b, 2009; Yoshii, 1979; Kao and Chen, 1995;
51 Kao and Liu, 1995; Seno and Yamanaka, 1996; Brudzinski et al., 2007). The upper sub-zone
52 appears to be located in the crust and/or in the uppermost part of the underlying mantle, while
53 the lower sub-zone lies entirely in the mantle (Igarashi et al., 2001; Preston et al., 2003; Abers
54 et al., 2013; Nakajima et al., 2013).

55 Pseudotachylyte exposed in Cape Corse (Alpine Corsica, France) is of tectonic origin and was
56 generated under blueschist to eclogite facies metamorphic conditions (Austrheim and
57 Andersen, 2004; Andersen and Austrheim, 2006; Andersen et al., 2008, 2014; Deseta et al.,
58 2014a and b). As such, it was likely formed in Cretaceous to Paleogene times in the Wadati-
59 Benioff seismic zone of a subducting Tethysian lithosphere. The aim of this contribution is to
60 analyze the geometry of the pseudotachylyte fault veins and the kinematics during their

61 formation. The results of this analysis then allows a comparison between the fossil Corsican
62 Cretaceous to Paleogene seismic ruptures and present-day Wadati-Benioff seismic zones
63 observed in cold slabs such as the Pacific plate beneath NE Japan.

64 **2. Geological setting and structure of the study area**

65 *2.1. General setting*

66 Alpine Corsica is a segment of the Alpine orogen which displays an imbrication of thrust
67 sheets composed of rocks of various origins and variably deformed and metamorphosed.
68 Peridotite, serpentinite, gabbro, basalt, calcareous schist, siliceous schist and marble represent
69 remnants of the lithosphere of the Jurassic Piemonte-Liguria oceanic basin and its pelagic
70 sedimentary cover. Radiometric dating of the oceanic peridotites and gabbros yielded Middle
71 to Late Jurassic ages between 169 and 152 Ma (Ohnenstetter et al., 1981; Rossi et al., 2002;
72 Rampone et al., 2009; Li et al., 2015). Several observations indicate that the spreading ridge
73 of the Piemonte-Liguria basin was of slow to very slow type (Rampone and Piccardo, 2000;
74 Piccardo, 2008; Manatschal and Müntener, 2009).

75 Crystalline thrust sheets (composed mainly of orthogneisses) and proximal sedimentary
76 deposits are interpreted as fragments of the stretched European continental paleo-margin and
77 its sedimentary cover (Vitale-Brovarone et al., 2011, 2014; Meresse et al., 2012). The
78 imbrication of such a variety of rocks is classically interpreted as the result of an eastward-
79 dipping Cretaceous subduction of the Piemonte-Liguria oceanic basin and a part of the
80 stretched European margin beneath the continental lithosphere of Apulia followed by an
81 Eocene collision between the European and Apulian continental lithospheres (Mattauer and
82 Proust, 1976; Mattauer et al., 1977, 1981; Warburton, 1986). Initiation of the subduction is
83 poorly dated. Paleogeographic reconstructions suggest a Late Cretaceous age (Stampfli et al.,
84 1998), but Late Cretaceous absolute ages of the HP metamorphism suggest that subduction
85 started in Middle Cretaceous times or earlier. The ophiolite-bearing Schistes Lustrés nappe
86 complex lies between the two blocks involved in the collision.

87 The classical Alpine evolutionary models were modified by taking into account the Apennine
88 orogeny (e.g., Durand-Delga and Rossi, 2002). Several authors suggested that the east-
89 dipping Cretaceous subduction ceased in Paleocene-Early Eocene times and was followed by
90 a west-dipping subduction of a young oceanic lithosphere of a back-arc basin formed further
91 east (Jolivet et al., 1998; Lacombe and Jolivet, 2005; Molli, 2008; Molli and Malavieille,
92 2010; Agard and Vitale-Brovarone, 2013). Based on new Late Eocene HP metamorphism

93 ages, Vitale-Brovarone and Herwartz (2013) suggest that the subducting oceanic plate now
94 preserved as ophiolitic thrust sheets and nappes in Alpine Corsica was of ‘Apennine’ affinity,
95 i.e., was dipping westward since the very beginning of the convergence. These authors
96 however acknowledge that more datings are needed before invalidating the eastward-dipping
97 (‘Alpine’) subduction. In this paper, the Late Cretaceous to Early Tertiary oceanic subduction
98 will be considered as an Alpine-type east-dipping subduction, before it is replaced by a west-
99 dipping Apennine-type subduction in Middle Eocene times.

100 A large part of Alpine Corsica rocks suffered from a high pressure-low temperature (HP-LT)
101 blueschist to lawsonite-eclogite facies metamorphism (Ravna et al., 2010; Vitale-Brovarone et
102 al., 2013 and references therein). This HP-LT metamorphism is of Eocene age (55-34 Ma,
103 Brunet et al., 2000; Martin et al., 2011; Maggi et al., 2012; Vitale-Brovarone and Herwartz,
104 2013) and is interpreted as the result of the subduction of continental or oceanic units at great
105 depths. A retrograde greenschist facies metamorphism is also recorded in some units and is
106 interpreted as the consequence of a late- to post-orogenic extension during late Oligocene to
107 early Miocene times (Jolivet et al., 1990, 1991, 1998; Fournier et al., 1991).

108 Non-metamorphic tectonic units (so-called upper or superficial nappes) lie at the top of the
109 structural stacking of thrust sheets. These units consist of sedimentary strata mostly of
110 Jurassic to Eocene age and ophiolitic rocks and related deposits. The rocks constituting these
111 superficial nappes were likely formed or deposited on the margin of the Apulian continent (or
112 an intervening island arc or micro-block) and subsequently transported westward during the
113 collision above the metamorphic units.

114 Structures associated with the ductile non-coaxial deformation of the metamorphic units
115 include a widespread foliation, various folds (including sheath folds) and a pervasive E-W
116 stretching and mineral lineation. The sense of shear associated with the non-coaxial
117 deformation during the prograde metamorphism is top-to-the-west (Mattauer et al., 1977,
118 1981; Faure and Malavieille, 1981; Harris, 1985; Warburton, 1986) whereas that associated
119 with the retrograde metamorphism and the late- to post-orogenic extension is mainly top-to-
120 the east (Fournier et al., 1991; Jolivet et al., 1990, 1991). The late- to post-orogenic extension
121 is also attested by numerous normal faults striking around N-S and dipping eastwards or
122 westwards.

123 *2.2. Structure of the study area*

124 The study area is located in the southern part of the Cape Corse peninsula, around the Cima di
125 Gratera peak, and consists of what will be referred to hereafter as the Cima di Gratera
126 ophiolitic nappe. Through an inferred tectonic contact called ϕ_1 , this nappe overlies ductilely
127 deformed units composed of continental basement rocks, meta-ophiolites and meta-
128 sedimentary cover rocks (Pigno-Olivaccio and Morteda-Farinole units, Fig. 1) which recorded
129 a HP-LT metamorphism (Lahondère, 1981, 1996; Vitale-Brovarone et al., 2013). The Cima di
130 Gratera nappe consists of two superimposed units (Fig. 2): a lower serpentinite-peridotite unit
131 (hereafter peridotite unit) and an upper metagabbro unit separated by a brittle/ductile flat-
132 lying contact referred to as ϕ_2 .

133 According to Vitale-Brovarone et al. (2013), the highest metamorphic conditions recorded by
134 meta-sedimentary rocks and continental units surrounding the Cima di Gratera nappe are
135 temperatures between 414 and 471°C and pressures between 1.9 and 2.6 GPa, corresponding
136 to blueschist to eclogite facies conditions. Comparable P-T conditions (1.3 GPa, $455 \pm 35^\circ\text{C}$)
137 were estimated by Lahondère and Guerrot (1997) in similar units nearby the Cima di Gratera
138 nappe. No accurate P-T conditions could be determined directly from the Cima di Gratera
139 units. Following Deseta et al. (2014a and b) and despite the presence of faults between them,
140 we suppose that the Cima di Gratera units suffered from P-T conditions comparable to those
141 of the surrounding units, that is, blueschist to eclogite facies P-T conditions.

142 *2.2.1. The peridotite unit.*

143 The peridotite unit is composed of massive or foliated serpentinite embedding fresh to
144 variably serpentinitized peridotite lenses. Near the inferred ϕ_1 tectonic contact, the base of the
145 unit consists of strongly foliated serpentinites whose foliation is severely folded or sheared by
146 C-like surfaces. This intense basal deformation is interpreted as a consequence of the
147 emplacement of the nappe over its substratum. Most serpentinites are distributed in the lower
148 part of the peridotite unit and the degree of serpentinitization generally decreases upwards from
149 ϕ_1 . Most peridotite masses are located near the ϕ_2 contact and have thicknesses between 20 m
150 and 300 m. The peridotite is massive and granoblastic. At locality 3 (Fig. 1), it is cut by a
151 gabbro dyke. The peridotite is lherzolitic in composition, and is constituted by olivine,
152 diopside, enstatite and minor plagioclase, Cr-spinel and magnetite (Deseta et al., 2014a).
153 Pyroxenite was found near the contact ϕ_2 (locality 7), but the poor exposure conditions
154 prevent to determine its actual extent and its nature (dyke, cumulate or sill). Recrystallization
155 of diopside and enstatite to clinocllore and tremolite testifies to a greenschist facies
156 metamorphism of the peridotite (Deseta et al., 2014a and b). On the other hand, and unlike the

157 metagabbro of the overlying unit (see below), no mineralogical evidence for blueschist facies
158 metamorphism could be found in the peridotite. That the peridotite itself does not contain any
159 evidence for blueschist facies metamorphism most likely reflects the fact that its composition
160 does not allow formation of minerals diagnostic of blueschist facies. Indeed (see our
161 observations below and also Deseta et al., 2014a and b), peridotite-hosted pseudotachylyte
162 veins contain omphacitic microlites indicating that the melt cooled and solidified under
163 blueschist facies to eclogite facies conditions.

164 According to Deseta et al. (2014a), the greenschist facies metamorphism occurred during two
165 events or successions of events, first *before* the formation of pseudotachylyte and then *after* it.
166 The early metamorphic event or succession of events are tentatively related to hydrothermal
167 alterations having occurred during ocean-continent hyperextension, during seafloor spreading
168 at the ridge, or near the trench where the approaching plate bends. The late metamorphic
169 event or succession of events are likely contemporaneous with slab exhumation or nappe
170 emplacement processes.

171 In several localities, especially in the northern part of the study area, thin (about 15 m) to
172 thick (> 500 m) lenses of strongly foliated meta-sedimentary rocks and metagabbros are found
173 within the serpentinites, generally at short distances from the inferred basal contact φ_1 (Fig.
174 1). These lenses likely correspond to slices of the underlying units that were incorporated in
175 the peridotite unit during nappe emplacement. The generally strong deformation (intense
176 folding, shear bands offsetting foliation) of the rocks in the lenses and of the surrounding
177 serpentinites is in favor of this interpretation.

178 2.2.2. *The metagabbro unit.*

179 The description of the metagabbro unit by Deseta et al. (2014a) is summarized below and is
180 completed by our observations. The metagabbro unit is predominantly composed of an equant
181 metagabbro. Only its basal part consists of a foliated metagabbro. The primary minerals of the
182 equant metagabbro are plagioclase, diopside, minor olivine and rare ilmenite. Alteration of
183 plagioclase into sericite and of olivine into serpentine, magnetite or iddingsite is common.
184 The texture frequently changes from micro-gabbro to coarse-grained gabbro and locally to
185 pegmatitic gabbro. A magmatic foliation is locally observed but could not be mapped because
186 of exposure scarcity. The metagabbro is intruded by dolerite dykes (locality 12, Fig. 1).
187 According to Deseta et al. (2014a), an early greenschist facies metamorphism of the gabbro is
188 responsible for the partial or total replacement of diopside by actinolite, bastite or Mg-
189 hornblende. This early greenschist facies metamorphism is followed by a blueschist facies

190 metamorphism as attested by the replacement of diopside, actinolite, Mg-hornblende and
191 plagioclase by glaucophane, barrosite, albite and epidote. A late greenschist facies
192 metamorphism is evidenced by epidote, clinocllore and pumpellyite overprinting the
193 blueschist facies minerals. Like for the peridotite unit, the early greenschist metamorphism of
194 the gabbro is tentatively related to hydrothermal alteration having occurred during ocean-
195 continent hyperextension, during seafloor spreading at the ridge, or near the trench through
196 normal faulting of the bending plate (Deseta et al., 2014a and b). The late metamorphic event
197 is considered to be contemporaneous with slab exhumation or nappe emplacement.

198 2.2.3. *The contact between the two units*

199 The contact ϕ_2 between the two units of the Cima di Gratera nappe, already described by
200 Andersen et al. (2014), can be observed at several places (localities 6, 7, 8 and 9, Fig. 1). It
201 consists in a flat-lying sharp fracture surface which undulates gently and which superimposes
202 equant or foliated gabbro over peridotite or serpentinite. Weakly marked striations or
203 corrugations trending N75°E to N120°E are preserved on the surface. The peridotite below
204 the contact surface is intensely fractured and hosts abundant pseudotachylyte fault veins, most
205 of which being parallel or slightly oblique to the surface (see description below). In the
206 localities where the footwall consists of serpentinites, no pseudotachylyte could be found in
207 these rocks.

208 Where the hanging-wall consists of equant metagabbro, up to 10 cm thick pseudotachylyte
209 veins locally outline the base of the metagabbro unit. Where the hanging-wall consists of
210 foliated metagabbro, the foliation is generally parallel to the contact but can also be oblique to
211 it, forming angles of up to 35°. The thickness of the basal foliated metagabbro is between 20
212 cm and 30 m. The mylonitic deformation progressively decreases in intensity when going
213 upward, that is away from ϕ_2 . The foliation bears a stretching lineation around N120°E.
214 Polished hand-sample sections and thin sections perpendicular to the foliation and parallel to
215 the lineation did not provide any consistent shear senses. Going upward and away from ϕ_2 ,
216 minor shear zones cross-cut the equant metagabbro. Their thickness is about 15 cm but can
217 locally reach 1.5 m. They strike about N-S and dip between 10 and 40° westward. Their
218 foliation bears a weakly marked stretching lineation along N80°E. No consistent shear senses
219 could be retrieved from hand sample sections or thin sections. The spatial as well as the
220 chronological relationships between these minor shear zones and the main basal foliated zone
221 could not be clarified.

222

223 **3. Pseudotachylyte**

224 *3.1. Pseudotachylyte in the peridotite unit*

225 *3.1.1. General observations*

226 A summary of the description of the pseudotachylyte in the peridotite unit by Andersen et al.
227 (2008, 2014), Andersen and Austrheim (2006), Austrheim and Andersen (2004) and Deseta et
228 al. (2014a and b), completed by our observations, is given in the following. Pseudotachylyte
229 veins in the peridotite are characterized by a positive relief and by an orange to yellowish
230 color which contrasts with the rusty color of the host rock (Fig. 3). Two categories of
231 pseudotachylyte veins are distinguished: fault veins and injection veins. In contrast with
232 injection veins, fault veins can be followed along several tens of centimeters up to a few
233 meters, have similar orientations, and their thickness is between a few millimeters and 30 cm.
234 Injection veins are rare, small (length < 10 cm), and often cut by microfaults or cataclastic
235 shear zones, rendering their recognition difficult. Most fault veins (about 90 %) do not occur
236 as isolated occurrences, but are clustered in fault zones in which a large number of veins
237 (commonly several tens) form anastomosed or tangled networks (Figs 3 to 6).

238 As demonstrated by hand sample polished sections or thin sections, the peridotite hosting the
239 pseudotachylyte veins, whatever these are isolated or forming networks, is cataclastic to
240 varying extents. Where cataclasis is important, the granoblastic texture cannot be recognized
241 any more. Where cataclasis is moderate, olivine crystals, yet fractured, can be distinguished.

242 In the peridotite unit, the fresh (i.e., not serpentinized) peridotite always contains
243 pseudotachylyte. Conversely, the massive serpentinite does not contain any pseudotachylyte.
244 In the vicinity of localities 4 and 12 (Fig. 1), some fault or injection veins can be recognized
245 in serpentinite but, in these occurrences, serpentinization affects both the host rock and the
246 veins, and clearly post-dates pseudotachylyte formation.

247 As already reported by Andersen et al. (2014), unlike all pseudotachylyte occurrences
248 described in non-ductilely deformed host rocks and particularly in other ultramafic rocks
249 (Obata and Karato, 1995; Piccardo et al., 2007, 2010; Souquière and Fabbri, 2010; Ueda et
250 al., 2008), the peridotite fault veins form complex anastomosed or tangled networks inside
251 which no general relative chronology between veins can be established. Indeed, cross-cutting
252 relationships between a limited number (up to 3) veins are often observed at the outcrop scale

253 and also at the thin section scale, but remain insufficient with respect to the large number of
254 veins to allow a complete chronology of a single fault zone to be reconstructed.

255 Another difference between the Cima di Gratera peridotite pseudotachylyte and most
256 occurrences from elsewhere lies in the scarcity of sharp and planar fault surfaces adjacent to
257 fault veins. Indeed, since the pioneering work of Sibson (1975), it has been recognized that
258 fault veins are always flanked, on one side or on both sides, by sharp, commonly planar, slip
259 surfaces whose length is between one and several meters or tens of meters (e.g., Allen, 2005;
260 Di Toro and Pennacchioni, 2005; Grocott, 1981; McNulty, 1995; Swanson, 1988; Wenk et al.,
261 2000; Zechmeister et al., 2007). In the case of the Cima di Gratera peridotite unit, such meter-
262 scale slip surfaces are rare. This seems to be also the case for the Lanzo peridotite
263 pseudotachylyte (Piccardo et al., 2007, 2010).

264 *3.1.2. Geometry and kinematics of pseudotachylyte-bearing fault zones*

265 As reported by Andersen and Austrheim (2006), two types of pseudotachylyte fault vein-
266 bearing zones can be distinguished in the peridotite unit: flat-lying fault zones in the upper
267 part of the unit (near φ_2), and steeply-dipping fault zones in the middle or lower part of the
268 unit (Fig. 2). Isolated fault veins also follow this geometry: flat-lying veins in the upper part
269 of the peridotite unit, steeply dipping ones in the lower part. To determine the sense of
270 displacement (seismic slip) associated with pseudotachylyte fault veins and given the fact that
271 the fault veins or associated cataclastic surfaces lack displacement direction indicators such as
272 striations, field-oriented hand samples were cut along vertical planes striking N-S, NE-SW, E-
273 W and NW-SE. For all samples, the most coherent sense of displacement criteria are observed
274 along NW-SE surfaces and to a lesser extent along E-W surfaces. This indicates that the NW-
275 SE direction is likely the displacement direction.

276 *Flat-lying pseudotachylyte-bearing fault zones*

277 Two main flat-lying fault zones are found in the peridotite unit (Fig. 2): the first one is located
278 immediately beneath the φ_2 contact where it can be followed almost continuously (localities 4
279 to 12, Fig. 1); the second one is located in the median part of the unit (localities 1 and 3, Fig.
280 1). The thickness of the fault zone beneath φ_2 varies between 25 and 250 m, while that at
281 localities 1 and 3 is about 15 m. In the fault zone beneath φ_2 , fault veins are parallel or slightly
282 oblique to φ_2 and their density unambiguously increases upwards when getting closer to φ_2 .
283 The parallelism between veins also tends to increase when getting closer to the contact.
284 Examination of outcrops, polished hand sample sections and thin sections reveals numerous

285 cross-cutting relationships between fault veins. As stated above, because of the complexity of
286 the relationships, it is not possible to establish a unique and complete chronology of seismic
287 ruptures in the flat-lying fault zones. The only clear relationship observed at all localities
288 consists in early gently to moderately ($5\sim 20^\circ$) dipping fault veins offset by late flat-lying to
289 gently dipping ($0\sim 10^\circ$) fault veins. Early veins often show blurred boundaries. Conversely,
290 late veins have sharp boundaries, and are frequently thinner than early veins. Cataclasite is
291 always associated with the early veins but is seldom observed along the late veins. For the
292 two fault zones (Figs 3 and 4), the sense of displacement associated with late veins is top-to-
293 the-northwest or top-to-the-west. The sense of displacement associated with the early veins
294 remains undetermined. Top-to-the-northwest or top-to-the-west displacement senses are
295 obtained for isolated flat-lying fault veins scattered in the peridotite unit.

296 *Steeply dipping pseudotachylyte-bearing fault zones*

297 In addition to isolated steeply dipping veins, three steeply dipping pseudotachylyte-bearing
298 fault zones are observed (localities 1, 3 and 5, Fig. 1). They consist of anastomosed networks
299 of fault veins hosted by cataclastic peridotite. Their thickness varies between 30 cm and 1 m.
300 The two first zones strike $N20^\circ E$ to $N40^\circ E$ and dip 55° eastwards. The third zone (locality 5)
301 strikes around $N80\text{--}N100^\circ E$ and dips 55° northwards. The two first fault zones display a clear
302 zonation with cataclastic peridotite predominating in the hanging-wall side and fault veins
303 predominating in the footwall side. This asymmetrical zonation is reminiscent of
304 pseudotachylyte described in different geological settings and host rocks (Fabbri et al., 2000).
305 Kinematic indicators suggest that the sense of displacement during formation of the steeply
306 fault veins was a reverse one, that is a top-to-the-northwest one (Figs 5 and 6). Flat-lying or
307 gently-dipping ($0\sim 10^\circ$) fault surfaces, most of them being coated by pseudotachylyte, cross-
308 cut the steeply dipping fault veins of the three steeply-dipping fault zones. Kinematic
309 indicators associated with these late flat-lying faults indicate a top-to-the-west or top-to-the-
310 northwest sense of displacement. The same cross-cutting relationships and kinematics are
311 obtained for the isolated steeply dipping fault veins scattered in the peridotite unit: steeply
312 dipping fault veins are cross-cut by flat-lying veins, but both types are characterized by a top-
313 to-the-west or top-to-the-northwest sense of displacement.

314 *Relative dating of flat-lying and steeply-dipping fault zones.*

315 Both flat-lying and steeply dipping fault zones were the sites of repeated seismic ruptures, as
316 attested by multiple generations of pseudotachylyte veins. Given the complexity of cross-
317 cutting relationships, it is difficult to establish a chronology of seismic ruptures within each

318 fault zone and also between fault zones. It is particularly impossible to establish a chronology
319 between flat-lying and steeply-dipping fault zones. However, the latest recorded seismic
320 ruptures are those corresponding to flat-lying or gently dipping, thin (< 5 mm)
321 pseudotachylyte veins displacing all pre-existing veins in either type of fault zones. These
322 veins are rarely associated with cataclasite, hence their sharp boundaries. The ubiquity of
323 these most recent veins suggests that activity of the flat-lying fault zones, especially that
324 beneath φ_2 (which contains a lot of such late veins), lasted for a longer time than that of the
325 steeply dipping fault zones.

326 *3.1.3. Pseudotachylyte in the peridotite unit: microscopic observations*

327 The peridotite pseudotachylyte fault veins are of two types: microlitic and annealed (Fig. 7).
328 Unlike the annealed type which is found only in the flat-lying fault zone beneath φ_2 , the
329 microlitic type is found in all vein types, whatever their relative chronology (early as well as
330 late veins).

331 The microlitic type is characterized by abundant microlites embedded in a brownish
332 amorphous or crypto-crystalline matrix (Fig. 7A, B and C). Microlites have dendritic shapes,
333 with sizes between 1 μm and 120 μm . They commonly draw a zonation in veins, with no or
334 very small microlites at the vein margins and large microlites in the median part of the veins.
335 Such zonations likely reflect quenching at the vein margins. Microlites observed in the
336 thickest veins (> 5cm) have sizes up to 1.5 mm and display spinifex textures. Microlites
337 consist mostly of olivine. Diopside and enstatite are less common. ‘Survivor’ clasts, about
338 10% in volume, consist mostly of monocrystalline olivine and minor pyroxene. Some
339 polycrystalline clasts consist of olivine and pyroxene. Clasts of pseudotachylyte are also
340 observed, especially in the flat-lying fault veins beneath φ_2 . The largest clasts are frequently
341 elongated and parallel to the vein walls. Flow folds are common in injection veins, especially
342 at their root. Although most pseudotachylyte vein boundaries are sharp, some are diffuse and
343 show a progressive transition from pseudotachylyte to cataclasites (see below). Such diffuse
344 boundaries are found on one side of the vein wall only, the other showing a sharp transition
345 from pseudotachylyte to host rock. Some microlitic veins, especially flat-lying ones located
346 near the boundary between intact peridotite and surrounding serpentinite, are serpentinitized,
347 but the textures, including microlites and clasts, are preserved, showing that serpentinitization
348 post-dates melting.

349 The annealed pseudotachylyte type (Fig. 7D and E) was observed only in the flat-lying fault
350 zone of locality 4. The annealed type fault veins have thicknesses between 0.8 and 1.5 cm.
351 They are crossed by numerous cooling cracks perpendicular to the vein walls. The matrix
352 consists of entirely recrystallized olivine with a granoblastic annealed texture. The crystal size
353 is homogeneous, except at the margins, with a mean size of *ca.* 10 μm . Crystal junctions are
354 triple and typically define 120° angles. Chilled margins are thin (< 0.5 mm thick). Survivor
355 clasts are rare (< 5% in volume) and consist of monocrystalline or polycrystalline olivine.
356 Injection veins are not entirely recrystallized, their central part showing a microlitic texture.
357 This suggests that the annealed-type veins are originally microlitic type veins that were
358 subsequently recrystallized.

359 An important characteristic of the peridotite fault veins, especially the early ones, is that they
360 are frequently flanked, on either side, by cataclastic peridotite (Figs 8 and 9). Cataclasite can
361 be found in association with late veins from flat-lying fault zones, but never with late veins
362 from the steeply dipping fault zones. Where cataclasite is present, a progressive evolution
363 from proto-cataclasite to cataclasite and to ultra-cataclasite can be observed. The cataclasite
364 usually remains on the same side of the fault vein, but can also shift to the opposite side,
365 through a progressive decrease of the cataclastic domain on one side and a correlative
366 increase on the other side.

367 The kinematics determined from the observation of outcrops or of polished hand sample
368 sections (top-to-the-west or top-to-the-northwest displacement senses) are also observed at the
369 thin section scale. Figure 10 shows examples of criteria of displacement senses associated
370 with early and late veins of steeply dipping fault zones and with late veins from flat-lying
371 fault zones.

372 *3.2. Pseudotachylyte in the metagabbro unit*

373 In the metagabbro unit, pseudotachylyte veins are common in the mylonitic sole and abundant
374 just above it (in the equant metagabbro). Their density decreases upwards. At a distance larger
375 than 300 m above ϕ_2 , no more veins can be found. As already noted by Andersen and
376 Austrheim (2006), within the mylonitized sole, some veins were formed *before* mylonitization
377 and others were formed *after* it (Fig. 11).

378 *3.2.1. Pre-mylonitization pseudotachylyte in the foliated sole*

379 In the foliated metagabbro, dark bluish to blackish fault veins are parallel or slightly oblique
380 to the foliation. Their thickness is less than 5 mm. Because of mylonitization-related
381 stretching and pinch-and-swell, they show a poor lateral continuity and cannot be followed
382 over more than a few centimeters. In some instances, they are flanked by injection veins
383 which are offset by millimeter-thick shear zones parallel to the foliation. No clear kinematics
384 is associated with the pre-mylonitization veins.

385 Most textures typical of pseudotachylyte are obliterated by the penetrative foliation (Fig. 12).
386 Flow structures, chilled margins and microlites are no longer recognizable. Newly formed
387 minerals are aligned in the foliation and consist of glaucophane, albite, epidote and ilmenite.
388 Survivor clasts, mostly plagioclases, are elongated in the foliation and are flanked with
389 pressure shadows. They are smaller (maximum 50 μm , 30 μm on average) than the clasts in
390 the non-mylonitized veins.

391 *3.2.2. Post-mylonitization pseudotachylyte in the foliated sole*

392 Post-mylonitization fault veins are greenish to greyish and are not foliated (Fig. 11). Most
393 fault veins are parallel or nearly parallel to the foliation. The parallelism between most of the
394 fault veins and the foliation likely reflects a mechanical influence of the foliation on the
395 seismic rupture propagation and hence on the resulting attitude of the veins. Post-
396 mylonitization fault veins as well as injection veins are also found to cross-cut pre-
397 mylonitization veins. No clear kinematics can be attributed to the post-mylonitization
398 pseudotachylyte veins.

399 Post-mylonitization fault veins are usually less than 5 mm thick, but can reach 10 mm. Clasts
400 consist mostly of plagioclase, with minor tremolite or pyroxene. The largest clasts, especially
401 the pyroxene ones, show embayments. The matrix is cryptocrystalline or glassy, and includes
402 microlites of albite and glaucophane. The length of the microlites is between 10 and 20 μm ,
403 but decreases to less than 5 μm near the vein boundaries (chilled margins). Flow folds are
404 abundant and are defined by alternating layers of microlites of different sizes. Injection veins
405 are rare. They include less clasts than the fault veins do. The cryptocrystalline or glassy
406 matrix contains microlites which are slightly larger (40 μm) than in the fault veins. The
407 mineralogical nature of clasts or microlites is the same as in the fault veins.

408 *3.2.3. Pseudotachylyte in the equant metagabbro above the foliated sole*

409 The appearance of pseudotachylyte veins in the equant metagabbro is similar to that of post-
410 mylonitization veins in the foliated sole. In particular, cross-cutting relationships between
411 veins are common and indicate polyphase seismic rupturing. A notable difference is that
412 lenses of pseudotachylyte-supported breccias are observed in equant metagabbro but not in
413 the foliated sole. These lenses are up to 15 cm thick and include rounded fragments of equant
414 metagabbro and foliated metagabbro. They are interpreted as local accumulations in so-called
415 dilational or releasing bends located along slipping surfaces (Sibson, 1986). Another
416 difference is that the fault vein attitudes in the equant metagabbro are scattered. Particularly,
417 veins parallel or slightly oblique to φ_2 are scarce.

418 The metagabbro pseudotachylyte matrix is glassy or crypto-crystalline. Using X-ray
419 diffraction synchrotron, Deseta et al. (2014b) produced Laue patterns without diffraction
420 points, showing unambiguously the presence of glass in the matrix. Survivor clasts are well
421 rounded and show the same mineralogical nature as in post-mylonitization veins. Feldspar
422 clasts are numerous and small whereas pyroxene and olivine clasts are scarce and large.
423 Microlites consist mainly of fibrous or acicular, rarely spherulitic, pyroxene. Deseta et al.
424 (2014b) report the presence of blueschist facies microlites, namely Al-rich omphacite, high-
425 Fe anorthite and accessory ilmenite. A few thin sections show different microlite
426 assemblages, suggesting different stages of vein formation. Some veins contain omphacite but
427 no tremolite or actinolite, some others contain tremolite or actinolite but no omphacite.
428 Omphacite-bearing veins likely formed under high-pressure conditions while amphibole-
429 bearing veins formed in shallower conditions (greenschist facies). Unfortunately, such
430 observations are too scarce to allow a reliable sorting of the equant metagabbro veins.

431 Pseudotachylyte veins in the equant metagabbro and post-mylonitization veins in the foliated
432 sole can be considered as contemporaneous. However, a part of the veins in the equant
433 metagabbro, especially those cut by other veins, could be older and could have been formed
434 coevally with the pre-mylonitization veins. Only absolute dating of the veins could help
435 clarify the relationships between post-mylonitization veins in the foliated metagabbro and
436 veins in the non-foliated metagabbro.

437

438 **4. Discussion: Analysis of the seismic ruptures fossilized in the study area and** 439 **comparison with present-day subduction zone seismology**

440 *4.1. Significance of the φ_2 contact and formation of metagabbro mylonite and pseudotachylyte*

441 The φ_2 contact is interpreted as an ancient deformed Moho interface by Andersen et al. (2008,
442 2014). Alternatively, φ_2 can be interpreted as an ancient low-angle detachment fault or shear
443 zone dating back to the initial stretching of the continental lithosphere before formation of the
444 Piemonte-Liguria basin (Meresse et al., 2012), following the model of Manatschal and
445 Müntener (2009). It can also be interpreted as an ancient low-angle detachment fault located
446 at or near the accretion ridge of the Piemonte-Liguria basin (Vitale-Brovarone et al., 2014)
447 following the models of Tucholke and Lin (1994) or Cannat et al. (2009). If φ_2 is an ancient
448 detachment (at the ridge or in the ocean-continent transition), it may not coincide with the
449 Moho interface. The supposed detachment could have been reactivated during plate
450 convergence or during subsequent collision and nappe emplacement. Whatever its origin, the
451 detachment should have been localized beneath the bulging sides of a gabbro diapiric pluton
452 emplaced in the uppermost mantle, to account for the gabbro-over-peridotite succession
453 observed in the study area. Lastly, the two scenarios (reactivated Moho or detachment in the
454 uppermost mantle) are not contradictory. Both involve a low-angle shear zone in the upper
455 part of the lithosphere.

456 The ductile deformation of the metagabbro sole may result from one or several of the
457 following settings: (1) normal shear along a crustal-scale detachment following continental
458 lithosphere breakup (e.g., Meresse et al., 2012; Vitale-Brovarone et al., 2014), (2) normal
459 shear along an axial detachment fault near the spreading ridge of the Piemonte-Liguria basin,
460 (3) reverse shear along the crust-mantle boundary (Moho) of the subducting slab. The lack of
461 kinematic indicators in the metagabbro sole prevents distinguishing stages (1) or (2) (normal
462 sense of shear) from stage (3) (reverse sense of shear).

463 Whatever the setting, the ductile deformation was achieved at a place where the ambient
464 temperature was higher than the brittle/ductile transition temperature of gabbro. Since the
465 plagioclase modal content in the metagabbro is about 50%, a minimum estimate of the
466 brittle/ductile transition temperature is *ca.* 550°C (e.g., Molli, 1994; Hansen et al., 2013).
467 Given the abundance of pyroxenes (about 35 % modal content), the actual transition
468 temperature of the metagabbro is likely higher. The maximum metamorphic temperatures in
469 the study area are 414-471°C (Vitale-Brovarone et al., 2013), that is less than 550°C. This
470 could rule out the possibility of a ductile deformation in the subducting slab. However, by
471 admitting that the metagabbro underwent some hydrothermal alteration (as can be expected if
472 φ_2 is a reactivated detachment), its brittle/ductile transition temperature would have been
473 lower (*ca.* 300°C, Stünitz, 1993), thus permitting ductile deformation.

474 Calling upon ductile deformation of the metagabbro sole along a detachment in the ocean-
475 continent transition or at the mid-oceanic ridge (and hence generation of pre-mylonitization
476 pseudotachylyte at the same place) requires that these early structures will then be transported
477 until the subduction zone at intermediate depths where they will be overprinted by (post-
478 mylonitization) pseudotachylyte veins. In particular, it means that (pre-mylonitization)
479 pseudotachylyte veins formed away (200~600 km away according to Guerrera et al., 1993;
480 Stampfli et al., 1998; Rosenbaum et al., 2002; Marroni and Pandolfi, 2007; Turco et al., 2012)
481 from the subduction zone will be overprinted by (post-mylonitization) pseudotachylyte veins
482 formed at depth (> 60 km) in a subduction zone. Though not impossible, such a coincidence
483 does not seem very plausible.

484 Figure 13 suggests a simpler scenario in which both pseudotachylyte and mylonite are formed
485 in the subducting slab at shallow to intermediate depths. Pre-mylonitization pseudotachylyte
486 is formed at depths shallower than the gabbro brittle/ductile transition isotherm. It can be
487 formed near the trench, along the Moho (Fig. 13A), following a scenario proposed by Singh et
488 al. (2008, 2011) to account for the location of the hypocenters and rupture propagation
489 geometries of the 2004 Sumatra and 2010 Pagai events. A similar possibility of seismic
490 ruptures inside the oceanic crust of the subducting Philippine Sea plate beneath SW Japan was
491 also suggested by Tsuji et al. (2009, 2013). Pre-mylonitization pseudotachylyte can
492 alternatively be formed deeper, before being subsequently mylonitized when passing through
493 the brittle/ductile transition isotherm. Continuing shear along the crust-mantle boundary
494 would result in σ_2 -parallel seismic ruptures in the brittle peridotite and in σ_2 -parallel foliation
495 in the metagabbro sole (Fig. 13B). That post-mylonitization pseudotachylyte cross-cuts the
496 foliated metagabbro suggests oscillations of the gabbro brittle/ductile transition isotherm.
497 Such oscillations are possible. Indeed, numerical simulations of long-term equilibrium state of
498 the subduction interplate show that the brittle/ductile transition is almost parallel to the crust-
499 mantle boundary of the subducting slab (Arcay, 2012). An alternative view is that post-
500 mylonitization pseudotachylyte veins, especially those perpendicular or highly oblique to the
501 foliation, result from seismic ruptures having nucleated in the underlying mantle and having
502 propagated upwards across the ductile metagabbro (Fig. 13D).

503 *4.2. Formation of peridotite pseudotachylyte*

504 *4.2.1. Weakening mechanisms facilitating seismic ruptures at intermediate depths*

505 Intermediate-depth seismicity as well as deep-focus seismicity are puzzling. Indeed, given the
506 high stresses expected at depths > 60 km, brittle fracturing or frictional sliding along pre-
507 existing fractures require unrealistic rock strengths or over-pressurized pore fluids which
508 could reduce stresses. Yet earthquakes occur. To solve this paradox, three mechanisms have
509 been proposed (Green and Houston, 1995; Hacker et al., 2003; Frohlich, 2006; Houston,
510 2015): dehydration embrittlement, ductile shear instability and transformational faulting.

511 Transformational faulting calls for the formation of anticracks during phase transformation of
512 olivine to denser phases such as β - or γ -spinel (Green and Burnley, 1989; Kirby et al., 1991;
513 Wiens et al., 1993; Schubnel et al., 2013). The applicability of transformational faulting as a
514 possible mechanism accounting for *intermediate-depth* seismicity in the subducting oceanic
515 lithosphere is questionable (e.g., Hacker et al., 2003) because (1) the expected reactions are
516 too slow compared with earthquake timescales, (2) olivine remains stable at the considered
517 depths, and (3) the metamorphism of basalts or gabbros does not involve the polymorphic
518 reactions required in transformational faulting processes.

519 Dehydration embrittlement is based on a pore fluid pressure increase leading to fracture
520 formation or reactivation by decrease of the otherwise high normal stresses. The pore fluid
521 pressure increase would result from fluid accumulations following dehydration reactions of
522 hydrated minerals. Seismological observations and data modeling, laboratory experiments,
523 field observations and thermal and thermodynamic computations have pointed out possible
524 links between intermediate-depth seismicity and dehydration reactions affecting hydrated
525 rocks or minerals of the subducting slab such as basalt, gabbro, chlorite, antigorite, talc or
526 brucite (Raleigh and Paterson, 1965; Rutter and Brodie, 1988; Green and Burnley, 1989;
527 Green et al., 1990; Green and Houston, 1995; Kirby, 1995; Seno and Yamanaka, 1996; Davis,
528 1999; Peacock, 2001; Seno et al., 2001; Dobson et al., 2002; Wang, 2002; Hacker et al., 2003;
529 Preston et al., 2003; Yamazaki and Seno, 2003; Jung et al., 2004; Wang et al., 2004;
530 Brudzinski et al., 2007; Hirose and Bystricky, 2007; Rondenay et al., 2008; Hasegawa et al.,
531 2009; Nakajima et al., 2009; Angiboust et al., 2012; Abers et al., 2013; Nakajima et al., 2013;
532 Houston, 2015). A strong argument justifying the link between intermediate-depth seismicity
533 and dehydration reactions in the subducting slab lies in overlaps between predicted
534 dehydration reaction isotherms and location of hypocenters (e.g., Peacock, 2001 or Hacker et
535 al., 2003). Regarding the *upper* Wadati-Benioff seismic sub-zone, the source of fluids would
536 lie in dehydration reactions transforming basalts or gabbros into blueschist or eclogite-facies
537 rocks (Hacker et al., 2003; Preston et al., 2003; Yamazaki and Seno, 2003; Kita et al., 2006;
538 Nakajima et al., 2009). Regarding the *lower* Wadati-Benioff sub-zone, the source of fluids

539 should be searched for in dehydration reactions of chlorite or antigorite since these two
540 minerals are thought to be present in the mantle of subducting slabs (Seno and Yamanaka,
541 1996; Peacock, 2001; Seno et al., 2001; Hacker et al., 2003). Dehydration reactions have been
542 invoked to account for secondary olivine crystallization along natural fault zones in ophiolitic
543 rocks from the Voltri complex in Italy (Hoogerduijn-Strating and Vissers, 1991; Scambelluri
544 et al., 1991). According to Hoogerduijn-Strating and Vissers (1991), fluid overpressures
545 would nearly reach lithostatic values.

546 Ductile instability, also called thermal runaway or thermal shear instability, postulates that the
547 temperature-dependent viscosity of a highly localized, ideally fine-grained, creeping ductile
548 shear zone is progressively reduced by the heat provided by continuing creep (Ogawa, 1987;
549 Kameyama et al., 1999; Braeck and Podladchikov, 2007; Kelemen and Hirth, 2007). This
550 positive feedback between continuing creep and temperature rising can eventually lead to
551 seismic failure. This mechanism was evoked in the case of the 1994 M_w 8.2 Bolivian deep-
552 focus earthquake (depth = 637 km) by Kanamori et al. (1998), who further suggested that
553 failure could have led to melting along the newly nucleated fault surface. Indeed, these
554 authors calculated that, for a starting shear zone thinner than 1 cm, temperature elevation in
555 the shear zone would exceed 10,000°C, which is far higher than melting temperature of any
556 rock. Source parameter scaling and energy budget of clusters of intermediate-depth (140-160
557 km) M_w 4-5 earthquakes beneath northern Colombia led Prieto et al. (2013) to suggest that
558 propagation of these ruptures was caused by a thermal runaway mechanism. A similar
559 mechanism was suggested by Wiens and Snider (2001) to account for deep (550-600 km)
560 earthquakes in the Tonga slab. From the analysis of closely associated mylonitic zones and
561 pseudotachylyte across a gabbro metamorphosed under high- to ultrahigh-pressure conditions
562 in Norway, John et al. (2009) suggested that co-seismic melting was contemporaneous with
563 ductile shear and is the result of a self-localizing thermal runaway process along the shear
564 zones. Another example of possible thermal runaway frozen in the geological record is
565 provided by Andersen et al. (2008, 2014) and Deseta et al. (2014a), as is discussed below.

566 *4.2.2. Did dehydration embrittlement facilitate seismic failure in the peridotite unit?*

567 According to Deseta et al. (2014a), the metagabbro pseudotachylyte was formed under
568 eclogite facies P-T conditions (presence of omphacite microlites in the matrix) before being
569 retrogressed under blueschist facies P-T conditions (presence of glaucophane microlites).
570 These authors estimate that the P-T conditions of crystallization of glaucophane and
571 omphacite microlites in the metagabbro pseudotachylyte are between 430 and 550°C and 1.8

572 GPa and 2.6 GPa, corresponding to blueschist to eclogite facies conditions. These values are
573 close to those obtained in the units around the Cima di Gratera nappe by Vitale-Brovarone et
574 al. (2013) which are 414-471°C and 1.9-2.6 GPa. With a mean rock density of 3000 kg/m³
575 and assuming a lithostatic equilibrium, the pressure range corresponds to depths between 60
576 and 90 km. P-T conditions in the peridotite pseudotachylyte cannot be ascertained. Indeed, the
577 microlites (diopside, olivine, enstatite and clinocllore) that crystallized during cooling of the
578 melt do not bring any constraints on the pressure conditions during pseudotachylyte
579 formation. However, if considering the peridotite unit as *attached* to the metagabbro unit
580 during subduction, then the same metamorphic conditions should apply to both units. In other
581 words, it can be assumed that pseudotachylyte in the peridotite unit formed under blueschist
582 to lawsonite-eclogite facies conditions, as suggested by Deseta et al. (2014a).

583 As mentioned above, antigorite, which is common in the serpentinites or serpentitized
584 peridotites of the Cima di Gratera nappe, is a candidate to account for dehydration
585 embrittlement and subsequent seismicity in subducting slabs. The P-T conditions for
586 antigorite dehydration are known from experiments and are between 550 and 720°C for
587 pressures between 1 and 3 GPa (Ulmer and Trommsdorff, 1995; Wunder and Schreyer, 1997;
588 Dobson et al., 2002; Perrillat et al., 2005; Hilairet et al., 2006; Padron-Navarta et al., 2010).
589 The peak temperatures supposedly recorded by the peridotite pseudotachylyte (430-550°C,
590 see above) or in the surrounding units (414-471°C, Vitale-Brovarone et al., 2013) are lower
591 than the temperatures required for dehydration of antigorite-bearing serpentinite. This
592 temperature difference renders dehydration of the mantle unit antigorite unlikely. In addition
593 to these temperature issues, no optical microscope or SEM observations of secondary olivine
594 newly crystallized at the expense of primary antigorite, such as the assemblages described by
595 Hoogerduijn-Strating and Vissers (1991) or by Scambelluri et al. (1991), could be found in
596 the pseudotachylyte veins or in their vicinity, confirming that antigorite dehydration did not
597 occur in the peridotite. Andersen et al. (2014) and Deseta et al. (2014b) confirm that they did
598 not observe secondary anhydrous minerals resulting from the dehydration of serpentine, talc,
599 clinocllore or amphibole in the Cima di Gratera metagabbros or peridotites. For the time
600 being, we consider that dehydration embrittlement was not an operative mechanism during
601 seismic failure in the peridotite unit.

602 *4.2.3. Did self-localizing thermal runaway facilitate seismic failure in the peridotite unit?*

603 Deseta et al. (2014a) report thin-section scale (20 mm to < 100 µm range) ductile (plastic)
604 deformation structures inside or along pseudotachylyte fault veins. These structures include

605 (1) elongated wallrock clasts in gabbro-hosted pseudotachylyte veins, (2) crystal plastic
606 deformation of the host metagabbro along the boundaries of pseudotachylyte veins, (3) grain
607 boundary alignment in prolate and lozenge-shaped grains suggesting grain boundary
608 migration in some peridotite-hosted pseudotachylyte veins, and (4) plastic ribbons in gabbro-
609 or peridotite-hosted ultracataclasites. Additionally, by analyzing dislocation slip systems in
610 olivine from peridotite wall rock or from clasts in the pseudotachylyte with the help of
611 electron backscatter diffraction, Andersen et al. (2014) and Deseta et al (2014b) suggested
612 that ductile deformation preceded pseudotachylyte formation. Based on these microscale
613 ductile precursors to seismic faulting found along pseudotachylyte fault veins in the Cima di
614 Gratera peridotites and gabbros, Andersen et al. (2014) and Deseta et al. (2014a and b)
615 suggested that seismic ruptures were facilitated or triggered by a self-localizing thermal
616 runaway process.

617 The possible activity of this process in the basal foliated metagabbro sole cannot be
618 demonstrated nor discarded. Indeed, the parallelism between (post-mylonitization)
619 pseudotachylyte fault veins and the foliation of the basal metagabbro probably results from
620 the influence of the pre-existing foliation on the propagation of the seismic rupture, as often
621 invoked in other settings (e.g., Grocott, 1981; Swanson, 1988; Allen, 2005; Zechmeister et al.,
622 2007). More generally, pseudotachylyte veins preserved inside mylonitic zones are quite
623 common (e.g., Sibson, 1980; Passchier, 1982), and their formation, although influenced or
624 guided by the pre-existing planar heterogeneity as stated above, does not necessarily depend
625 on a precursory softening shortly before seismic rupturing, as required in the ductile
626 instability mechanism.

627 Unlike Deseta et al. (2014a and b), we did not observe any ductile shear zones along the
628 peridotite fault veins, despite a large number of thin sections prepared with samples from the
629 peridotite unit. We rather observe a quasi-ubiquitous association of pseudotachylyte veins
630 with cataclastic peridotite. Consequently, ductile instability does not appear as a predominant
631 mechanism associated with seismic ruptures in the peridotite unit of the study area.

632 *4.2.4. Did cataclasis facilitate seismic failure in the peridotite unit?*

633 Peridotite-hosted pseudotachylyte fault veins are almost always flanked by cataclasite (Figs 8,
634 9 and 10). Cataclasis may predate or postdate frictional melting, as shown by cataclasite zones
635 crossed by pseudotachylyte veins or by fragments of pseudotachylyte included in cataclasites.
636 Similar pseudotachylyte-cataclasite associations were reported from natural occurrences
637 (Maddock, 1992; Magloughlin, 1992; Swanson, 1992; McNulty, 1995; Obata and Karato,

638 1995; Curewitz and Karson, 1999; Fabbri et al., 2000; Rowe et al., 2005; Di Toro and
639 Pennacchioni, 2004, 2005; Piccardo et al., 2007, 2010) and also from rock friction
640 experiments (Spray, 1995; Del Gaudio et al., 2009; Hirose et al., 2012). Swanson (1992)
641 considered the cataclasite as the result of the propagating seismic rupture front, frictional
642 melting occurring during seismic slip behind the front. Curewitz and Karson (1999) proposed
643 that cataclasite results from slip surface leveling by asperity grinding and abrasion. Since a
644 cataclastic peridotite is obviously mechanically weaker than an intact peridotite, one can
645 expect cataclasis to be a precursory weakening mechanism facilitating ensuing seismic
646 rupture. Unlike dehydration-derived over-pressurized fluids, cataclasis *per se* does not
647 contribute to counterbalance the high stresses expected at depths > 60 km. Though this
648 mechanism does not bring any answer to the enigma of earthquakes at great depths, it
649 however provides a plausible way to mechanically weaken strong rocks.

650 *4.3. Co-seismic displacement kinematics frozen in the peridotite unit compared with present-* 651 *day Wadati-Benioff zone earthquakes*

652 Figure 14 shows a comparison between co-seismic kinematics frozen in the peridotite unit
653 and the intermediate-depth seismicity of the Pacific plate presently subducting beneath NE
654 Japan. The Pacific plate is taken as representative of a cold slab. It is also a slab for which
655 high-resolution seismological data are available. The choice of the intermediate-depth
656 seismicity is justified by the fact that the peridotite-hosted pseudotachylyte were formed at
657 depths between 60 and 90 km, as suggested by the metamorphic pressure conditions in the
658 overlying metagabbro unit, supposed attached to the underlying peridotite unit. Since the
659 kinematics of gabbro-hosted pseudotachylyte are undetermined, the discussion will be largely
660 based on peridotite-hosted occurrences, with the assumption of an Alpine-type east-dipping
661 subduction (Section 2.1).

662 Given the large (55°) angle between flat-lying fault zones and the earliest veins of the steeply-
663 dipping fault zones in the peridotite unit, no unique stress tensor can account for simultaneous
664 slip along the two zones, suggesting that when one fault zone was active, the other was not
665 active. The pervasive intermingling between flat-lying and steeply-dipping veins suggests that
666 the two types of fault zones were active alternatively and under oscillating stress conditions.
667 Such changing stress conditions could result from the near-surface interplate seismic cycle
668 and periodic unlocking of the shallow plate interface during large earthquakes as suggested by
669 Astiz et al. (1988).

670 Several authors showed that in the case of cold slabs, the upper surface of the intermediate-
671 depth Wadati-Benioff zone is characterized by a so-called downdip compression, meaning
672 that P axes of earthquakes are parallel or almost parallel ($< 30^\circ$) to the dipping direction of the
673 subducting slab (Isacks and Molnar, 1971; Apperson and Frohlich, 1987; Green and Houston,
674 1995; Kao and Liu, 1995; Igarashi et al., 2001; Chen et al., 2004). The reverse kinematics
675 observed in the steeply-dipping fault zones agree with this configuration. Indeed, in slab
676 coordinates, a 55° dipping earthquake fault plane would have a ‘favorably’ oriented P axis at
677 30 to 45° to the fault plane, that is, at 10° to 25° to the slab upper surface (taken parallel to ϕ_2).

678 Refining the comparison with cold subducting plates, a series of events having occurred in the
679 early 2000s in the Pacific plate beneath NE Japan gives some insights on the geometry and
680 sense of slip of the Corsican paleo-ruptures (Fig. 14B and C). First, the November 3, 2002, M
681 6.1 earthquake which occurred along the upper Wadati-Benioff plane beneath NE Japan
682 (Okada and Hasegawa, 2003) can constitute an analog to seismic ruptures along *flat-lying*
683 fault zones in the peridotite unit or in the base of the metagabbro unit, on either side of ϕ_2 .
684 Indeed, the actual fault plane of this event was parallel to the alignment defining the upper
685 Wadati-Benioff plane (i.e., parallel to the crust-mantle boundary) and the sense of slip was
686 reverse. The analogy is somewhat limited by the shallow focal depth (38 km) of this event
687 (Hasegawa et al., 2007), meaning that it is not strictly speaking an intermediate-depth
688 earthquake. Second, the May 26, 2003 M 7.1 earthquake that occurred in the Pacific plate
689 (focal depth 68 km) can be an analog to seismic ruptures along *steeply-dipping* fault zones in
690 the peridotite unit. This event was located near the upper Wadati-Benioff plane, close to the
691 crust-mantle boundary (Okada and Hasegawa, 2003). Its hypocenter was 50 km away from
692 the November 3, 2002 event. Okada and Hasegawa (2003) further showed that aftershocks
693 were distributed along the steeply-dipping nodal plane, straddling both the oceanic crust and
694 the uppermost mantle. Sense of slip was reverse. The angle between the fault plane and the
695 crust-mantle boundary is 50° (Hasegawa et al., 2007), a value quite comparable to the 55°
696 angle between the flat-lying and the steeply dipping fault zones observed in the peridotite
697 unit. The analogy between the Pacific slab upper Wadati-Benioff seismic events and the
698 Corsican configuration is depicted on Fig. 14. Slab-boundary parallel seismic ruptures would
699 be analogous of the November 3, 2002 earthquake off NE Japan and, more generally, could
700 be deeper equivalents of the low-angle thrust fault (“LT”) type events of Igarashi et al. (2001).

701 In summary, the reverse senses associated with the seismic ruptures frozen in the study area
702 can be compared with earthquakes with reverse-type focal mechanisms or ‘down-dip
703 compression’ events. More particularly, the lack of normal kinematics associated with

704 pseudotachylyte generation suggests that normal-type events such as those corresponding to
705 reactivation at depth of normal faults formed in the slab before it starts subducting (e.g., Jiao
706 et al., 2000; Barnhart et al., 2014) either did not occur in the Corsican subduction zone or did
707 not leave any imprint.

708 **5. Conclusion**

709 The structural analysis of pseudotachylyte in the Cima di Gratera ophiolitic nappe leads to the
710 following results.

711 Pseudotachylyte veins in the peridotite unit are either isolated and scattered in the unit or
712 clustered in fault zones. Isolated veins as well as fault zones are horizontal (flat-lying type) or
713 dip about 55° (steeply dipping type). In fault zones, the abundance of fault veins likely
714 reflects a large number of repeating seismic ruptures, among which some may correspond to
715 small magnitude events like aftershocks. The lack of clear cross-cutting relationships suggests
716 that the flat-lying fault zones and their steeply dipping equivalents were active alternatively,
717 as a consequence of oscillating stress states possibly resulting from periodic unlocking of the
718 shallow plate interface during the seismic cycle. The activity of the flat-lying fault zones
719 probably lasted for a longer time than the steeply dipping fault zones. The sense of
720 displacement associated with steeply dipping fault zones and with most of the flat-lying fault
721 zones is top-to-the-west or top-to-the northwest. Cataclasite flanking most of the veins was
722 formed before or coevally with frictional melting and likely mechanically weakened the
723 peridotite, facilitating subsequent seismic rupture.

724 The base of the metagabbro unit is mylonitic. The origin of the mylonitization remains
725 undetermined. The scenario retained here suggests that the ductile deformation was achieved
726 in the subducting slab below the brittle-ductile transition depth of the gabbro. The isotherm
727 corresponding to the brittle-ductile transition depth of the gabbro can be taken as parallel to
728 the crust-mantle boundary in the subducting slab. Pseudotachylyte veins in the metagabbro
729 are distributed in the lower part of the unit, in the foliated sole as well as in the equant
730 metagabbro above. They are not as well organized as their equivalents in the peridotite unit.
731 Flat-lying veins are abundant near the contact with the underlying peridotite, steeply dipping
732 veins are scattered in the lower part of the unit. The ductile deformation affecting the base of
733 the unit allows to distinguish pre-mylonitization pseudotachylyte formed above the brittle-
734 ductile transition depth of the gabbro from post-mylonitization veins formed below this depth.
735 In the equant metagabbro, it is no longer possible to distinguish more than one episode of
736 pseudotachylyte formation. No information regarding the sense of displacement associated

737 with seismic ruptures could be retrieved from gabbro pseudotachylyte veins, whatever their
738 positions or attitudes. It is furthermore not possible to establish a relative chronology between
739 pseudotachylyte formed on either side of the contact ϕ_2 between the peridotite and
740 metagabbro units.

741 Depth constraints provided by the metamorphic conditions recorded by metagabbro
742 pseudotachylyte (1.9-2.6 GPa pressure range, Deseta et al., 2014 a and b) and geometry as
743 well as kinematics data from peridotite pseudotachylyte show similarities with well
744 documented seismic ruptures occurring in the Wadati-Benioff zone of the Pacific plate
745 beneath NE Japan. These similarities allow to propose a scenario of formation of
746 pseudotachylyte which encompasses shallow seismic ruptures along the crust-mantle
747 boundary as suggested by Singh et al. (2008) for the 2004 Sumatra earthquake and deeper
748 ruptures in the Wadati-Benioff zone (60-100 km depth range). In this scenario, seismic
749 ruptures in the subducting mantle would always occur under brittle conditions while those in
750 the lower part of the subducting crust would partly be coeval with ductile deformation. No
751 relative chronology between pseudotachylyte of either side of ϕ_2 (peridotite vs. metagabbro)
752 can be firmly established. A part of the metagabbro post-mylonitization veins could be the
753 result of large ruptures nucleated in the peridotite unit and propagated upward across ϕ_2 and
754 through a metagabbro sole under ductile conditions.

755 Deciphering intermediate-depth seismicity from ophiolite-hosted pseudotachylyte is a
756 complex task because rocks may have recorded earthquakes elsewhere than in the subducting
757 slab. The rocks may have been deformed, at least partly, at the axial ridge during oceanic
758 accretion or in the continent-ocean transition, during initial crustal thinning. Final ophiolite
759 emplacement, whatever by obduction or collision, and subsequent episodes (e.g., late- to post-
760 orogenic extension) are also responsible for additional deformation. All these deformation
761 episodes may contribute to clutter the final picture. However, like for the continental
762 lithosphere (Swanson, 1992; Obata and Karato, 1995; Allen, 2005; Di Toro and Pennacchioni,
763 2005; Ueda et al., 2008), pseudotachylyte is a valuable tool to improve our understanding of
764 the mechanics of the seismic ruptures in the oceanic lithosphere, in complement to
765 geophysical studies.

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1081 **Figure captions**

1082 **Figure 1.** Structural map of the study area (modified after Faure and Malavieille, 1981,
1083 Lahondère, 1996 and Meresse et al., 2012) and lower-hemisphere equal-area projections of
1084 poles to pseudotachylyte fault veins in the peridotite unit. Arrows indicate the sense of shear
1085 associated with pseudotachylyte fault veins. The locations of cross-sections A-A' and B-B'
1086 (see Fig. 2) are shown. CdG: Cima di Gratera; PdM: Punta di Muzzelli.

1087 **Figure 2.** Geological cross-sections of the study area (location in Fig. 1).

1088 **Figure 3.** Examples of outcrop-scale top-to-the-west or top-to-the-northwest displacement
1089 sense criteria from the flat-lying fault zones in the peridotite unit. A, B and C: West-dipping
1090 or northwest-dipping Riedel-like pseudotachylyte-coated normal faults (labeled by R)
1091 offsetting earlier fault veins. A and B from locality 6, C from locality 5. D: Southeast-dipping
1092 pseudotachylyte-coated reverse faults offsetting earlier fault veins at locality 3.

1093 **Figure 4.** Polished surface (A) and corresponding sketches (B and C) of a peridotite hand
1094 sample from the upper flat-lying fault zone at locality 2 (Fig. 1) showing two stages of
1095 pseudotachylyte formation. The kinematics associated with the early pseudotachylyte veins is
1096 undetermined, while that associated with the late pseudotachylyte veins is top-to-the-west
1097 (N280°E). Rectangle on (A) corresponds to the thin section scanner image of Fig. 10C.

1098 **Figure 5.** Detailed field view (A) and corresponding sketch (B) of a steeply-dipping reverse
1099 fault zone (locality 2) showing anastomosed fault veins crossing cataclastic peridotite.

1100 **Figure 6.** Polished surface (A) and corresponding sketch (B) of a peridotite hand sample from
1101 the steeply-dipping fault zone of locality 2 showing three stages of pseudotachylyte

1102 formation. Senses of shear of the intermediate and late seismic ruptures are top-to-the-
1103 northwest (N320°E). Rectangle on (A) corresponds to Fig. 10D.

1104 Figure 7. Thin section scanner images, SEM images and photomicrographs of microlitic-type
1105 and annealed-type pseudotachylyte veins in the peridotite. A: Thin section scanner image of a
1106 microlitic-type fault vein showing a zonation parallel to the boundaries with the host rock
1107 (locality 3). B: Chilled margin (C.m) of a microlitic-type fault vein showing a sharp decrease
1108 in microlite size. Microlites consist of olivine and pyroxene (locality 1). C: Pyroxene
1109 microlites in an injection vein (locality 4). D: Annealed-type fault vein from locality 4.
1110 Arrows indicate cooling cracks. Square is for E. E: SEM image of D showing olivine with a
1111 granoblastic annealed texture.

1112 Figure 8. Examples of pseudotachylyte-cataclasite associations in the peridotite-hosted
1113 steeply dipping fault zone at locality 2. A: Photomicrograph. (B) Thin section scanner image.
1114 Ct: cataclasite; Pct: proto-cataclasite; Pst: pseudotachylyte; Uc: ultra-cataclasite.

1115 Figure 9. SEM images of associated cataclasites and pseudotachylytes in peridotite from
1116 locality 6. A: From top left to bottom right, juxtaposition of proto-cataclasite (PCt), cataclasite
1117 (Ct), pseudotachylyte (Pst) and moderately fractured peridotite (wall). B: Detail of the Ct-Pst-
1118 host rock zoned domain of (A). C: Detail of the Ct domain of (A), showing angular clasts. D:
1119 Detailed view of a cataclasite-wall rock contact.

1120 Figure 10. Thin section scanner images and photomicrographs of fault veins from the steeply
1121 dipping fault zone at locality 2 (A and B) and from the flat-lying fault zone at locality 3 (C, D
1122 and E) showing top-to-the-west or top-to-the-northwest displacement senses. A: Parallel
1123 polarized thin section scanner image showing an anastomosed network of steeply dipping
1124 cataclasite zones with a reverse displacement sense. B: Crossed-polar enlarged image from
1125 (A) showing cataclastic zones offsetting olivine crystals in a reverse sense. C: Parallel-polar
1126 thin section scanner image of a pseudotachylyte vein and associated sheared peridotite
1127 suggesting a top-to-the-west sense of displacement. D: Parallel-polar thin section scanner
1128 image showing an early steeply dipping pseudotachylyte vein left-laterally offset by a late
1129 flat-lying pseudotachylyte vein. E: Detail of D showing that the flat-lying vein is younger
1130 than the vein dipping to the right.

1131 Figure 11. Outcrop aspect of pre- and post-mylonitization pseudotachylyte veins in the
1132 foliated metagabbro sole at locality 4 and attitudes of nearby post-mylonitization veins. A:
1133 Foliated metagabbro showing a post-mylonitization fault vein secant on pre-mylonitization

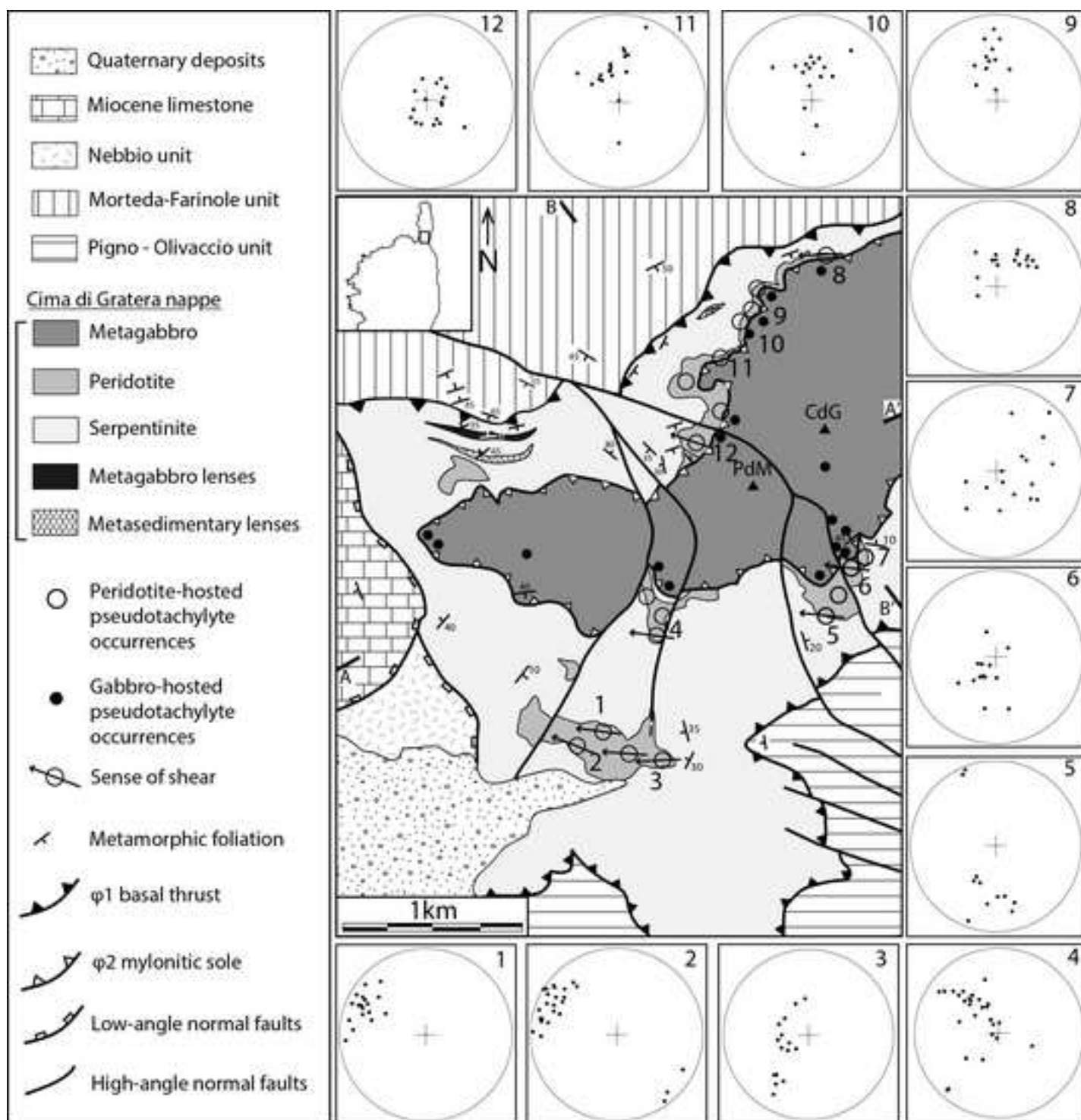
1134 fault veins and on foliation. The foliation is outlined by the dashed red line. Red arrows point
1135 at pre-mylonitization fault veins. B: Lower-hemisphere equal-area projection of poles to post-
1136 mylonitization fault veins (solid circles) and poles to foliation (red triangles).

1137 Figure 12. Microscopic aspect of pre- and post-mylonitization pseudotachylyte veins in the
1138 foliated sole of the metagabbro unit. A: Parallel-polar thin section scanner image of the
1139 foliated metagabbro. B: Detail of A showing a mylonitized pseudotachylyte vein. The
1140 foliation inside the vein is slightly oblique to the foliation outside, likely because of some
1141 obliquity of the vein with respect to the deformation axes. C: Parallel-polar thin section
1142 scanner image showing a post-mylonitization fault vein, a post-mylonitization injection vein
1143 and a foliated pre-mylonitization vein. D: sketch of C. Pre-myl. Pst: pre-mylonitization vein;
1144 Post-myl. Pst: post-mylonitization vein. Dashed lines outline the foliation trace.

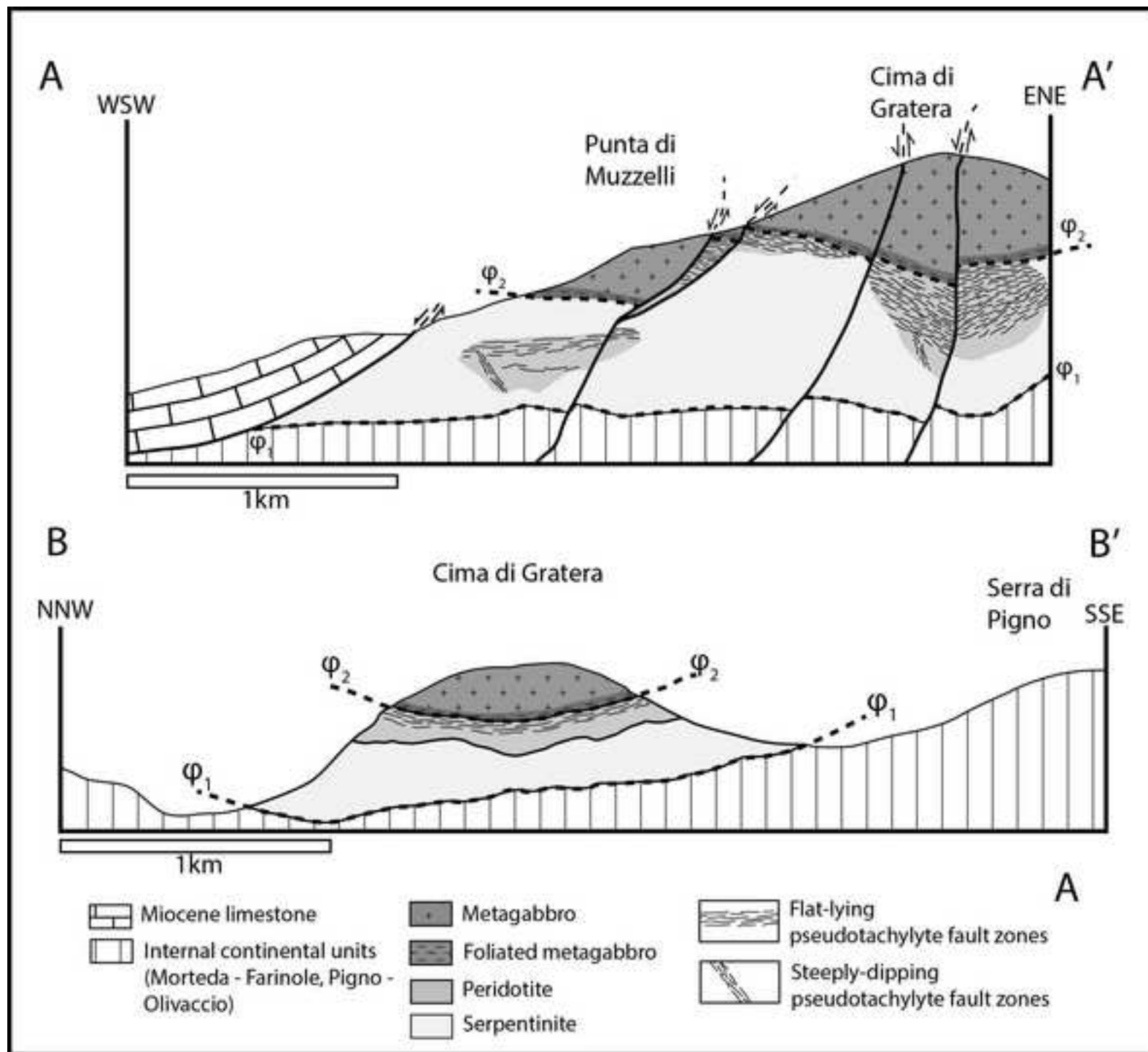
1145 Figure 13. Multi-stage scenario of formation of pseudotachylyte and mylonite in the Cima di
1146 Gratera nappe. A: General sketch showing the east-dipping subduction of the Piemonte-
1147 Liguria oceanic basin beneath an arc or a micro-continent in Cretaceous times. Also depicted
1148 are the hypocenters (asterisks) of the Wadati-Benioff seismic zone. B: Formation of pre-
1149 mylonitization pseudotachylyte at shallow depth at or near the mantle-crust boundary. C:
1150 Ductile deformation of the base of the oceanic crust and coeval formation of pseudotachylyte
1151 in the underlying peridotite. D: Formation of post-mylonitization veins in the ductilely
1152 deforming metagabbro by seismic ruptures nucleated in the peridotite and having propagated
1153 upward across and beyond the foliated metagabbro.

1154 Figure 14. Geometrical and kinematic similarities between the present-day seismic activity of
1155 the Wadati-Benioff zone beneath NE Japan (A and B) and the Corsican fossil seismic ruptures
1156 (C and D). A: Possible location of the Corsican seismic ruptures (rectangle) in a cold slab
1157 thermal model (isotherms after Peacock, 2001 and Hacker et al., 2003). B: Hypocenters and
1158 kinematics of the 2002-2003 seismic activity in the uppermost part of the Pacific plate off NE
1159 Japan (Hasegawa et al., 2007). The red rectangle delineates the possible equivalent of the
1160 seismic fault zones frozen in Corsica. C: Sketch summarizing the geometry and kinematics of
1161 the Corsican fossil seismic ruptures in the peridotite unit (approximate width: 5 km). D:
1162 Detail of C emphasizing large ruptures propagating upwards across the crust-mantle boundary
1163 and beyond the foliated basal metagabbro (approximate width: 1 km).

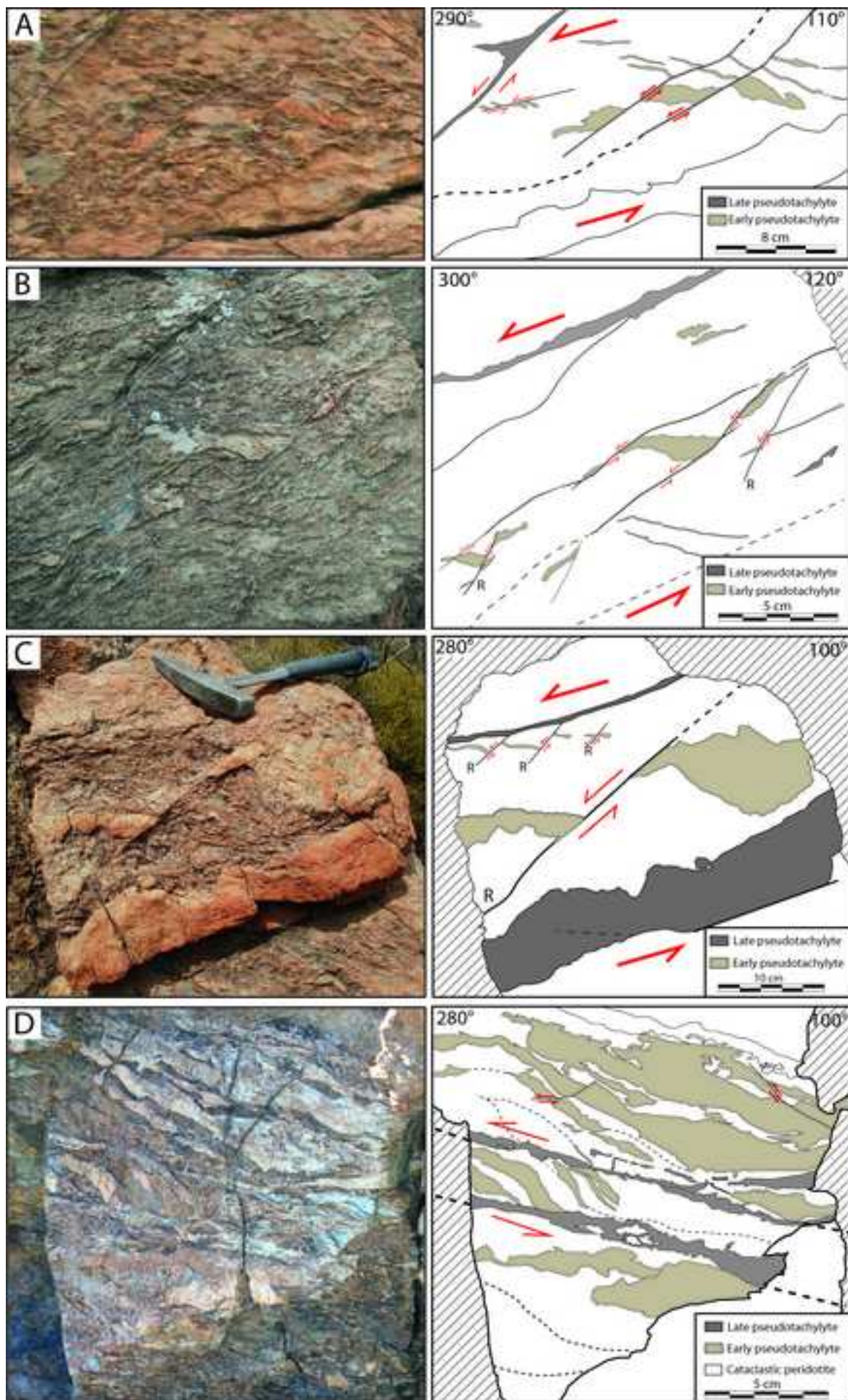
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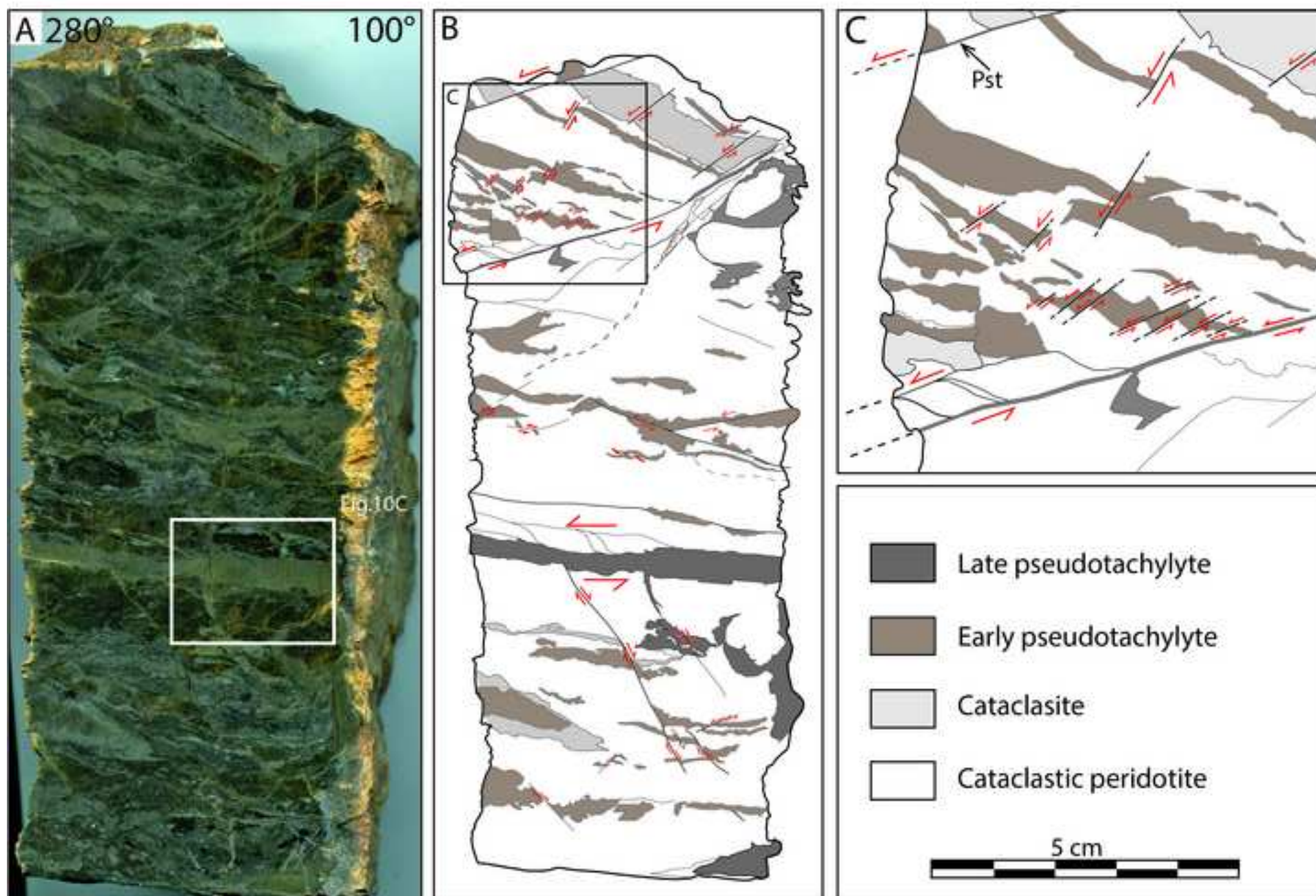


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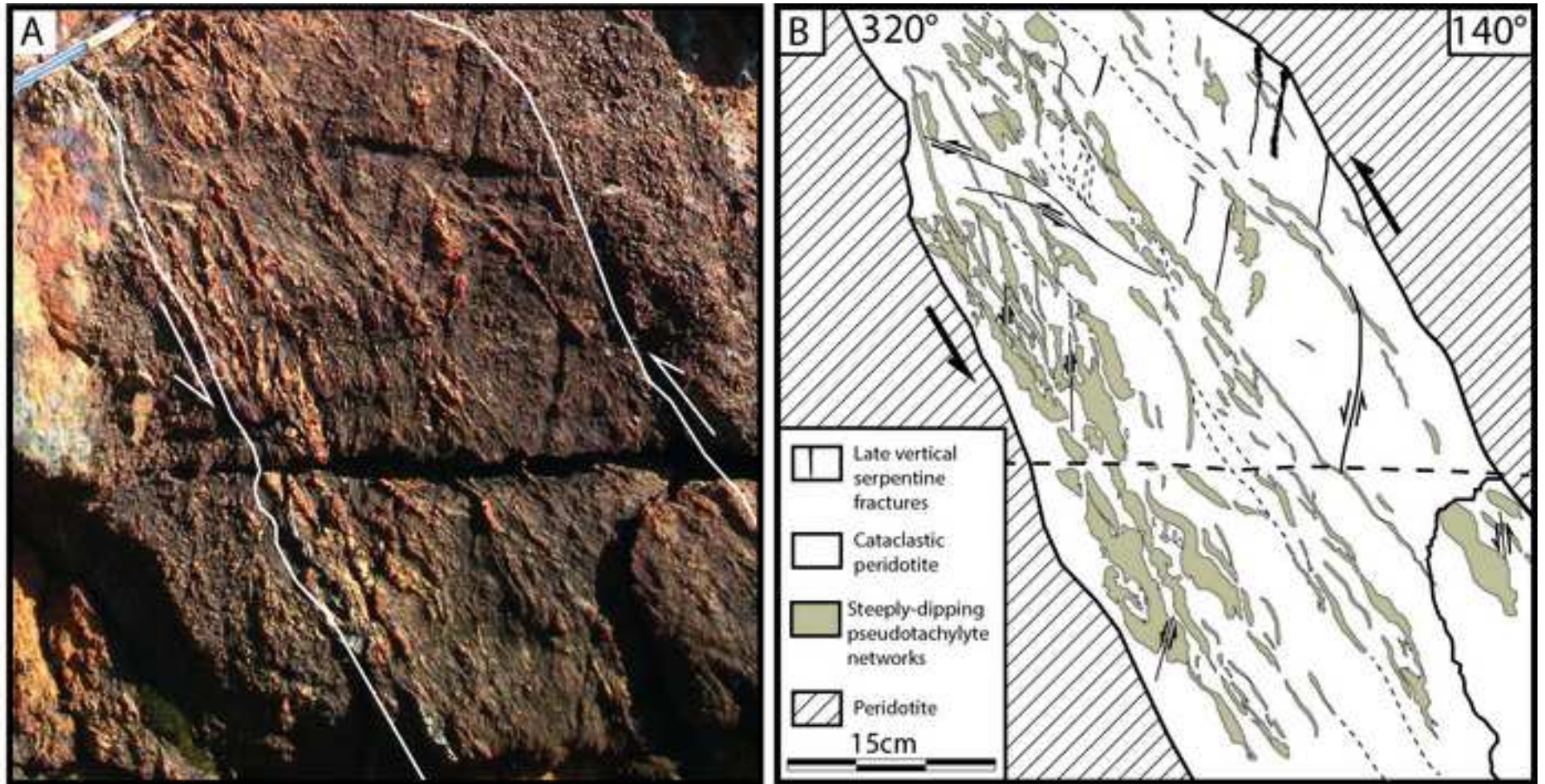


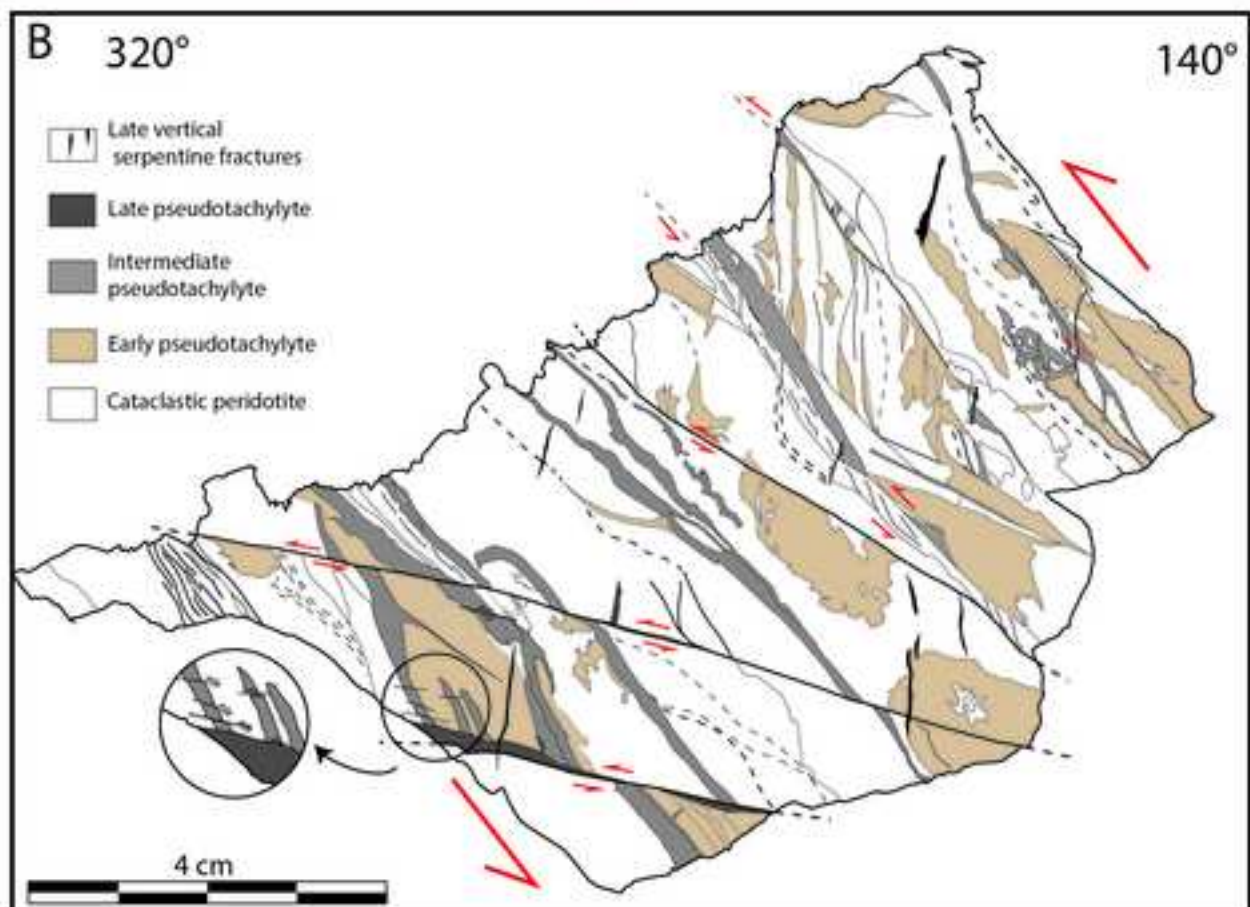
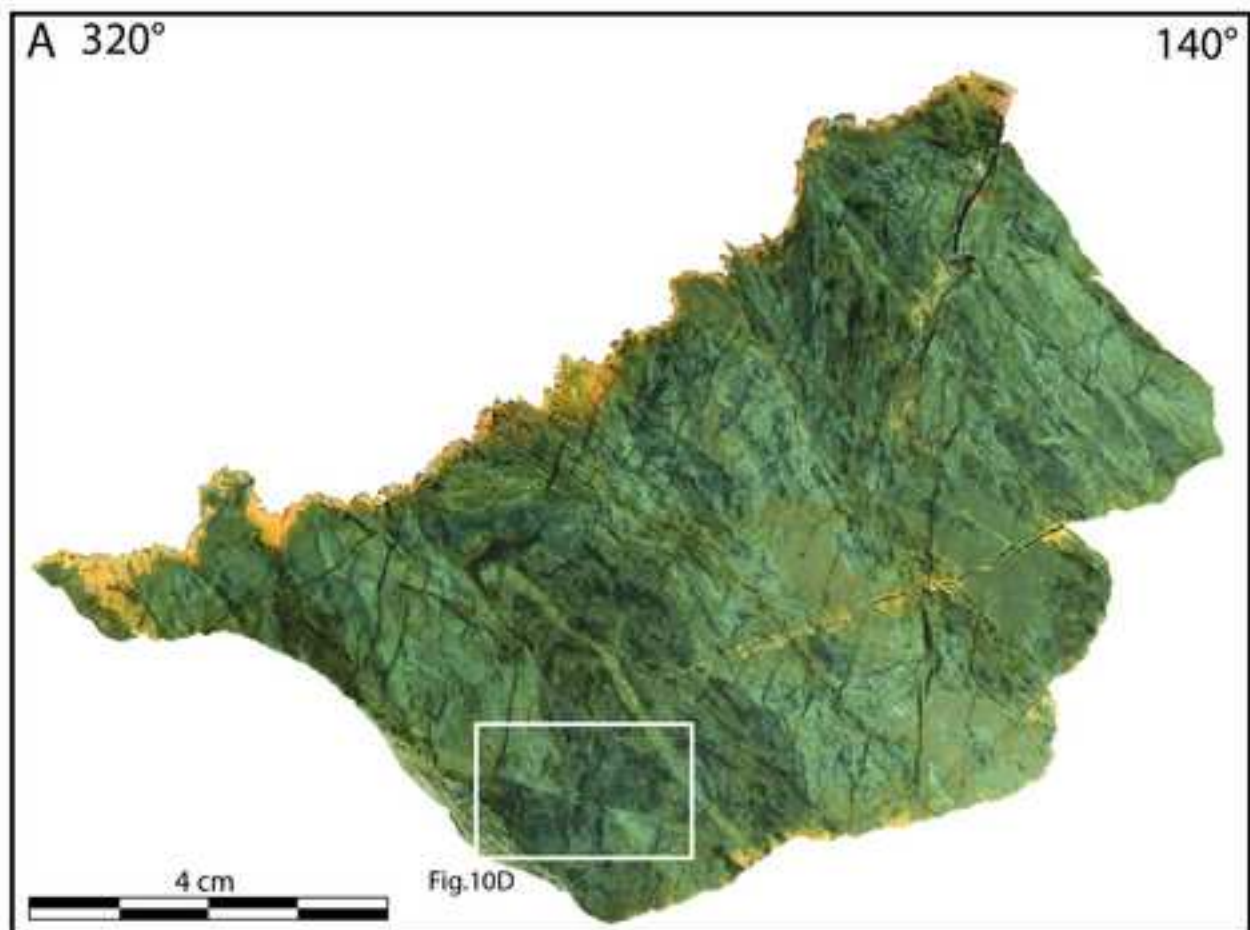
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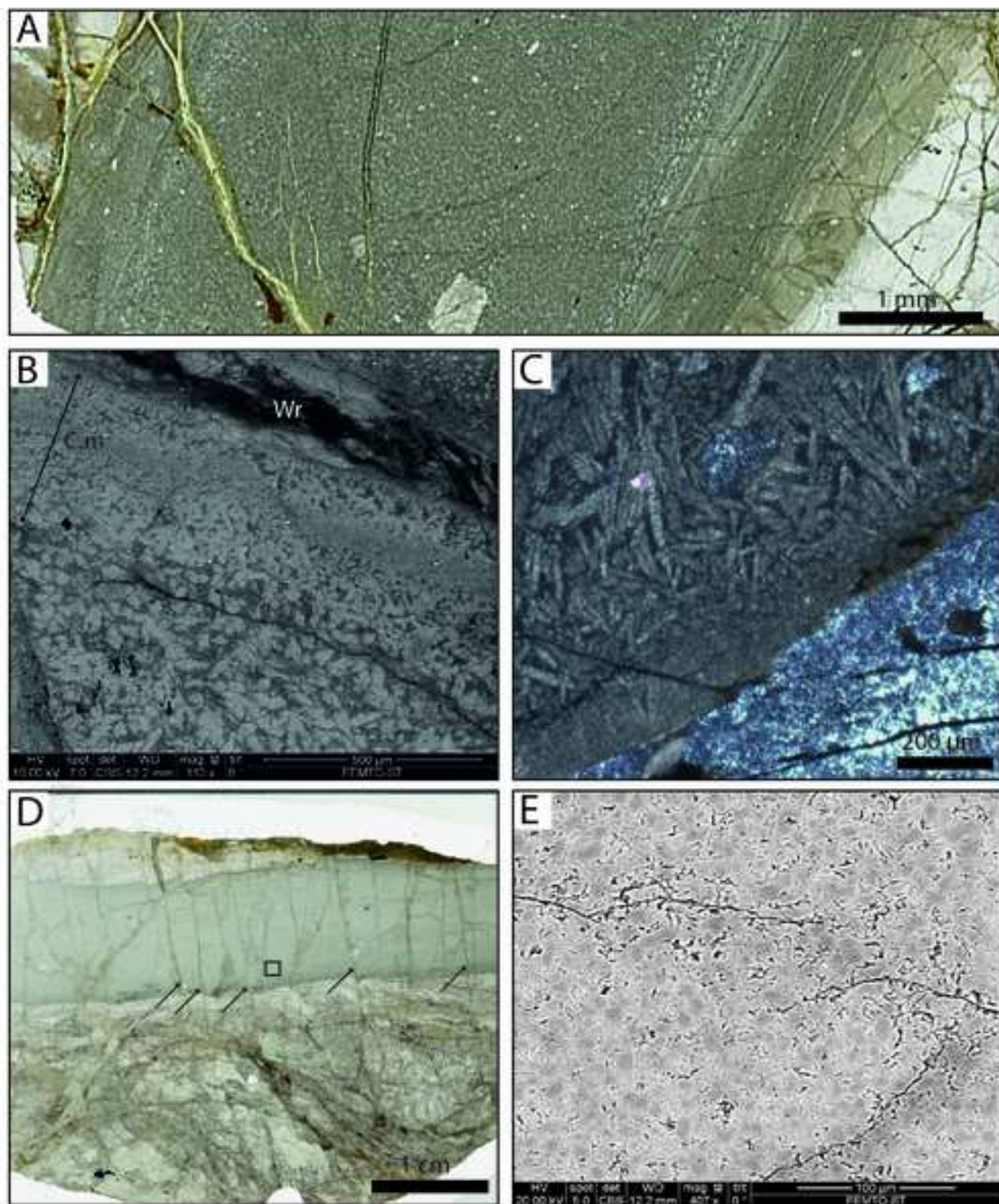


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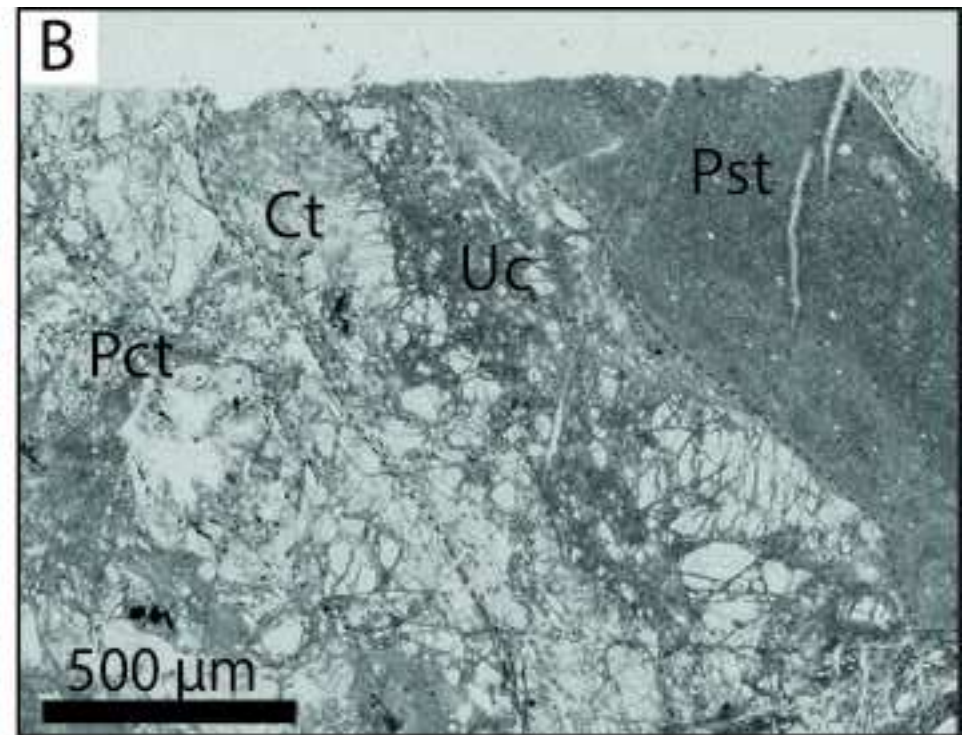
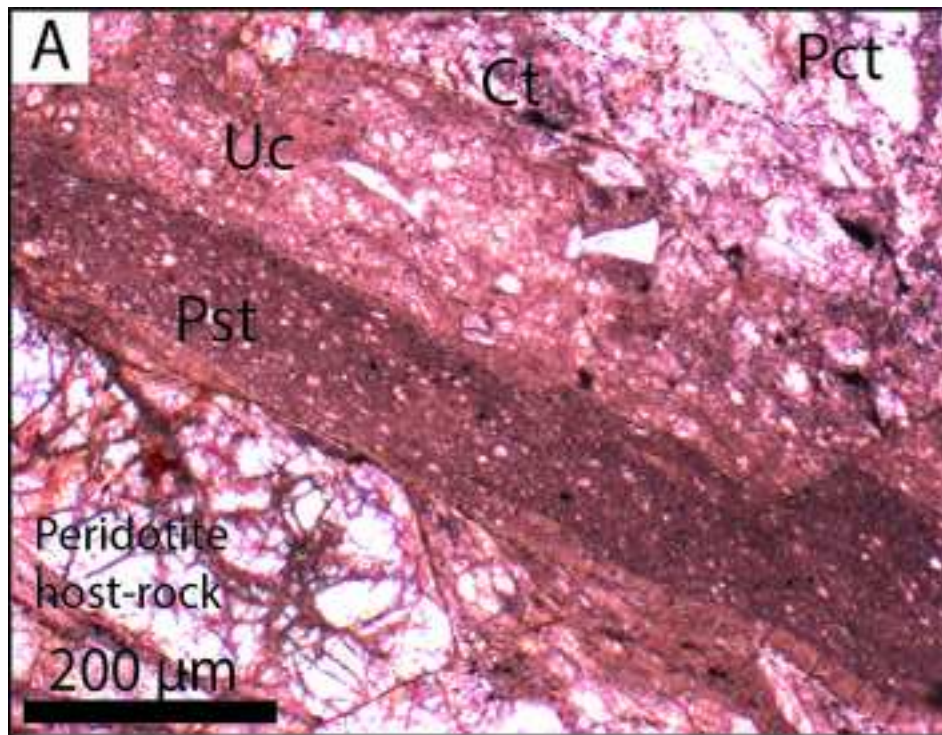




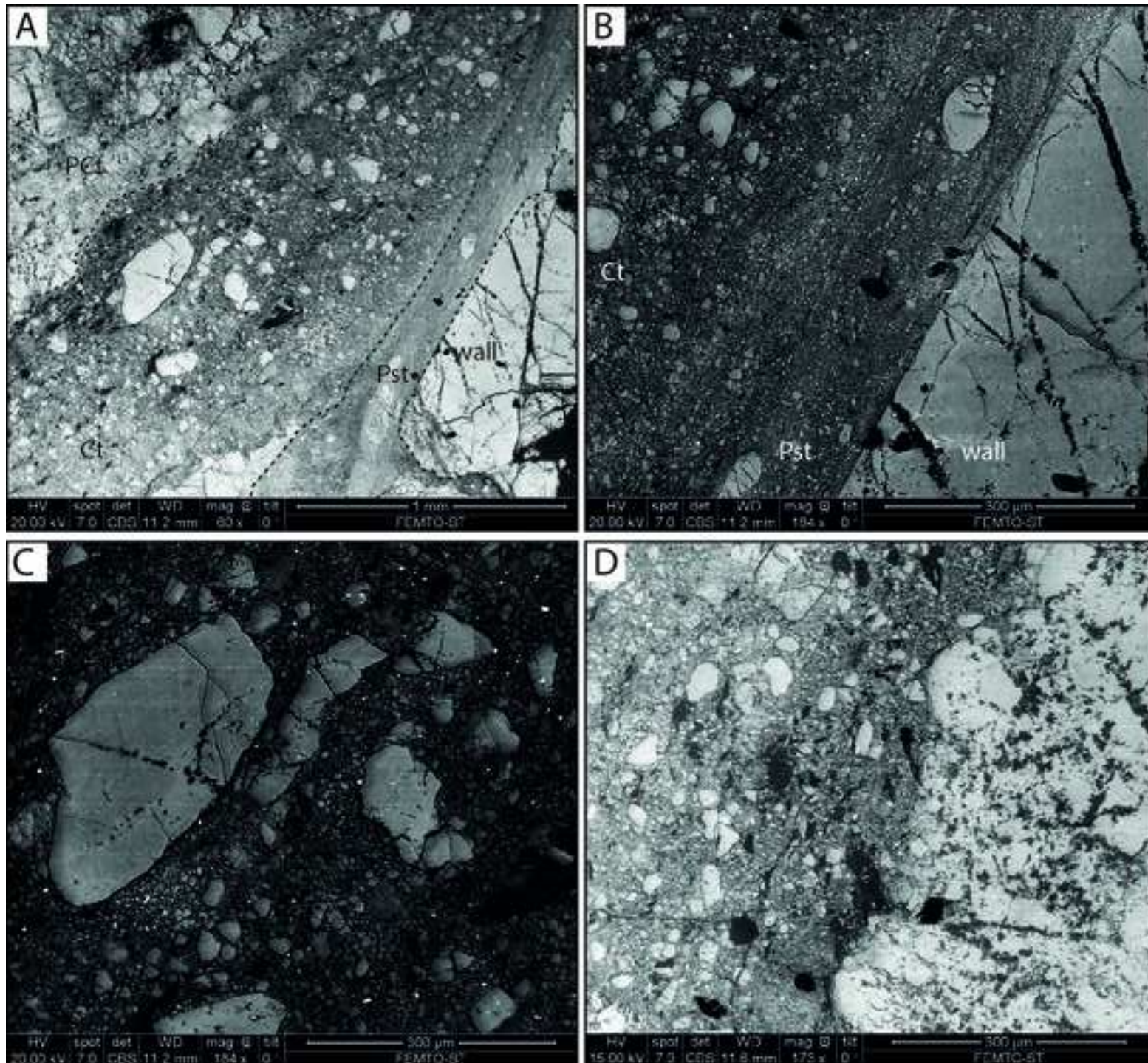
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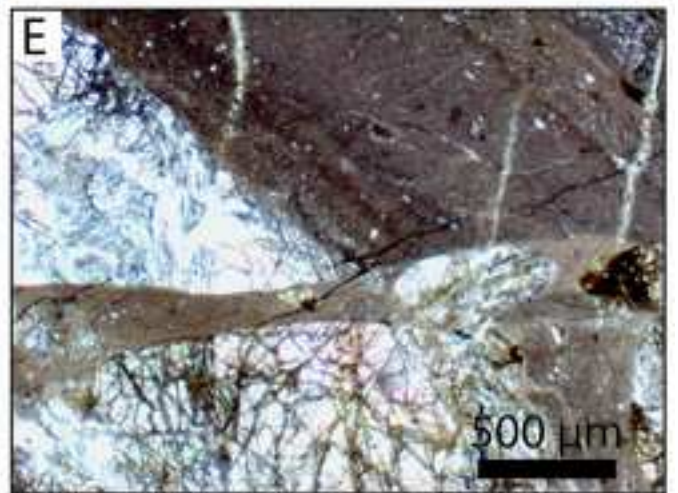
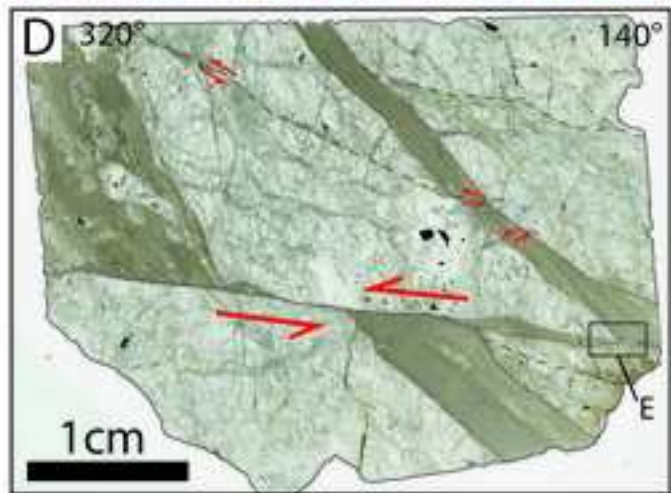
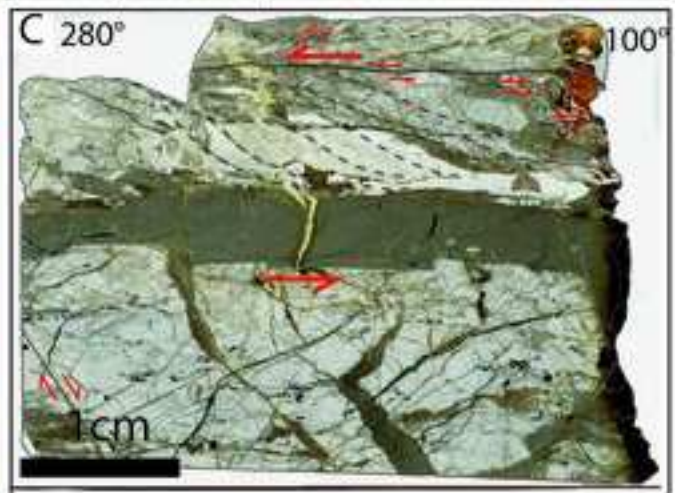
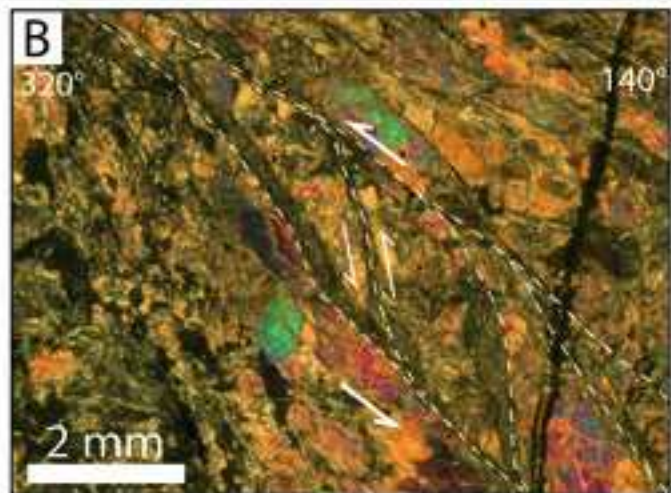
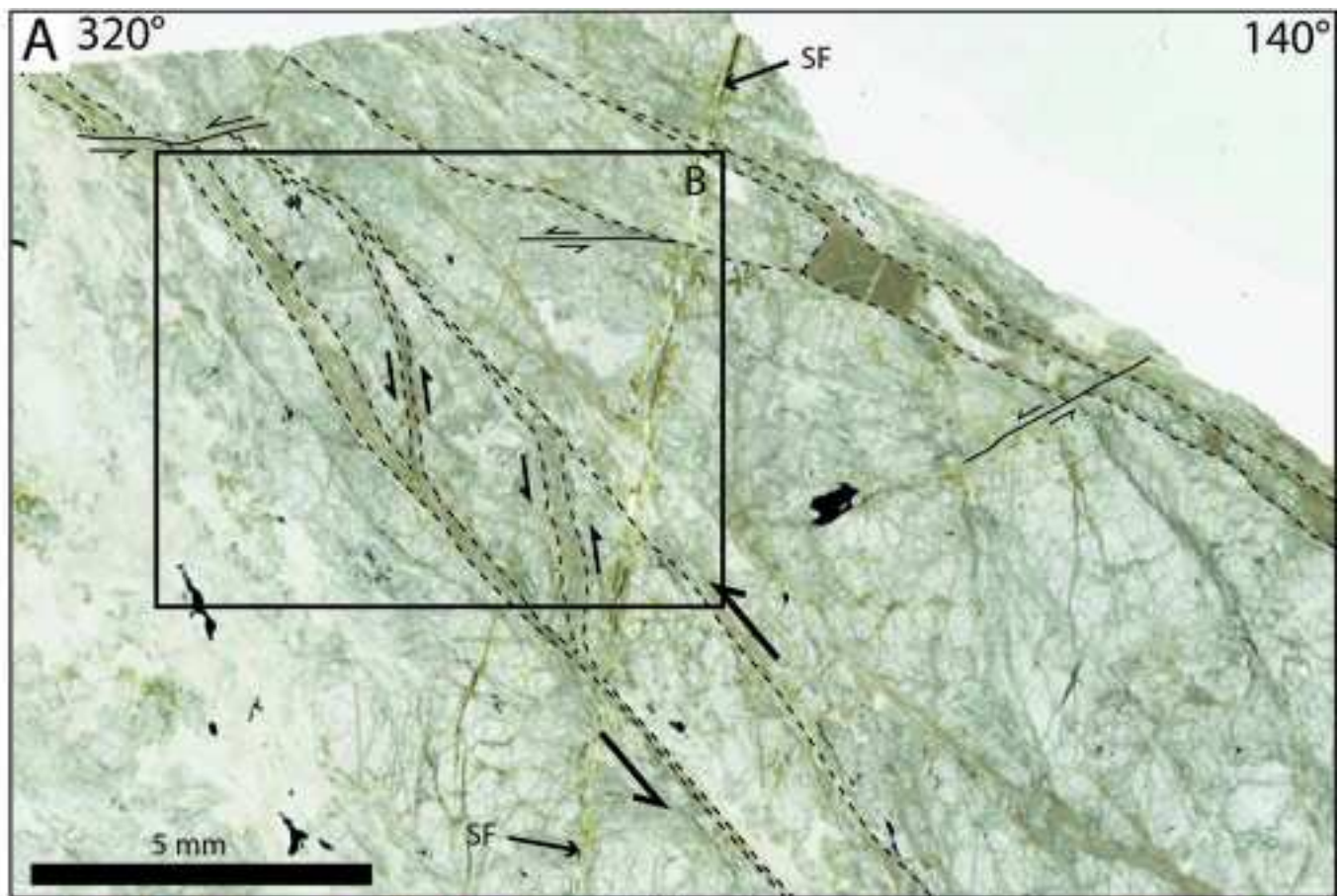
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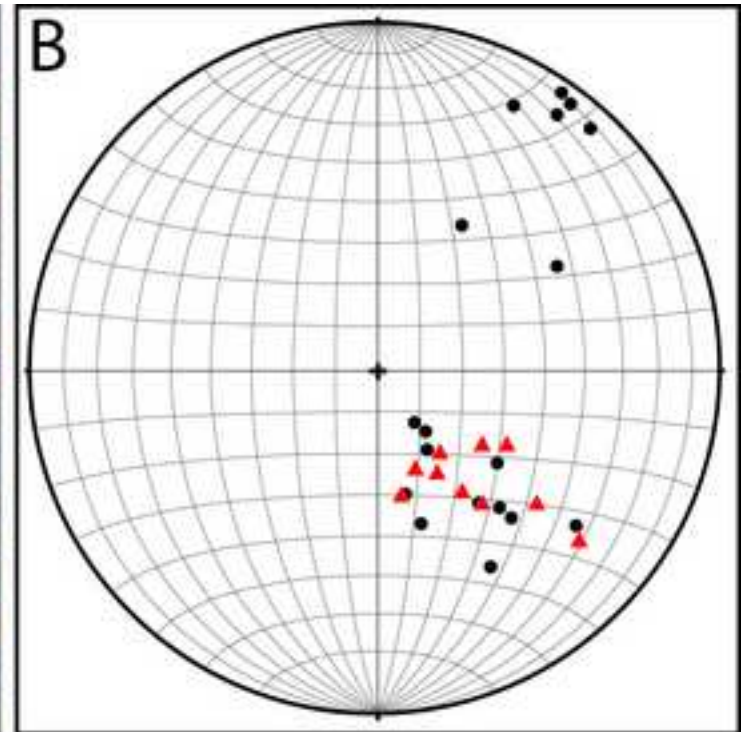
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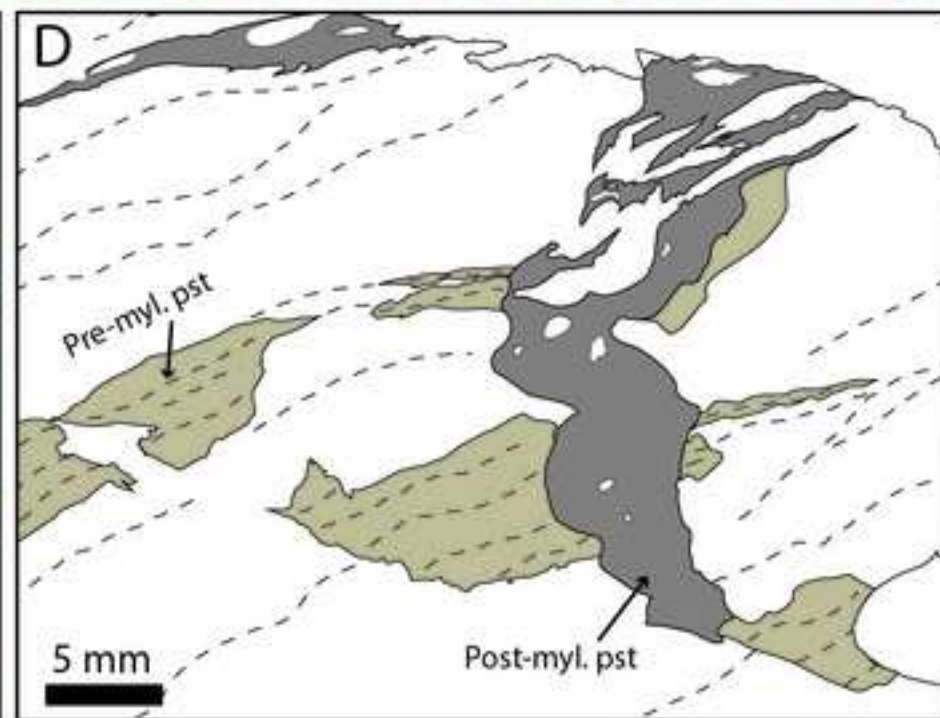
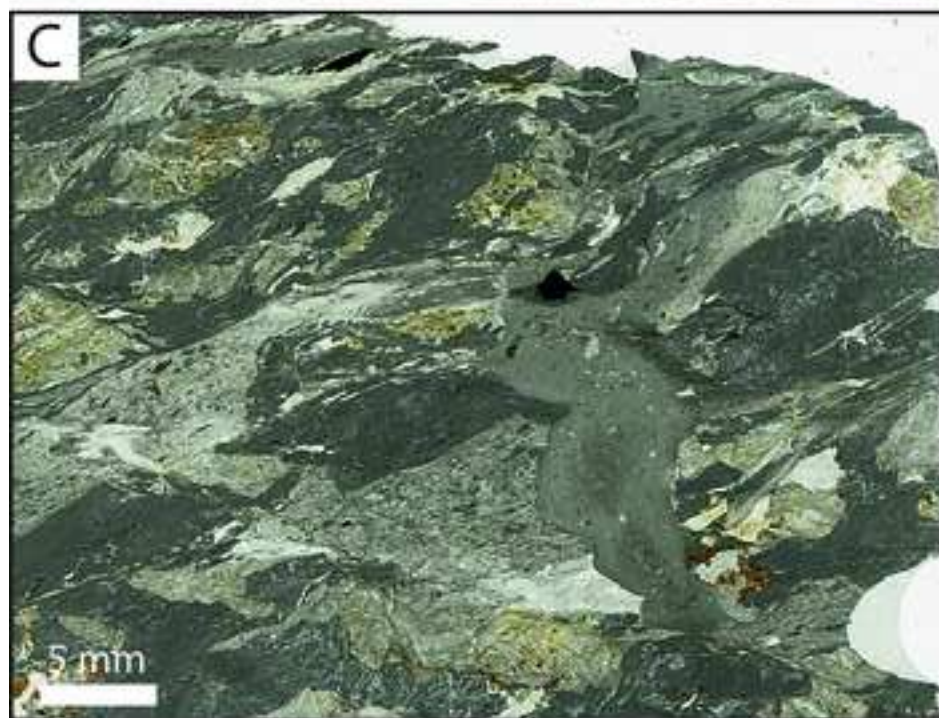
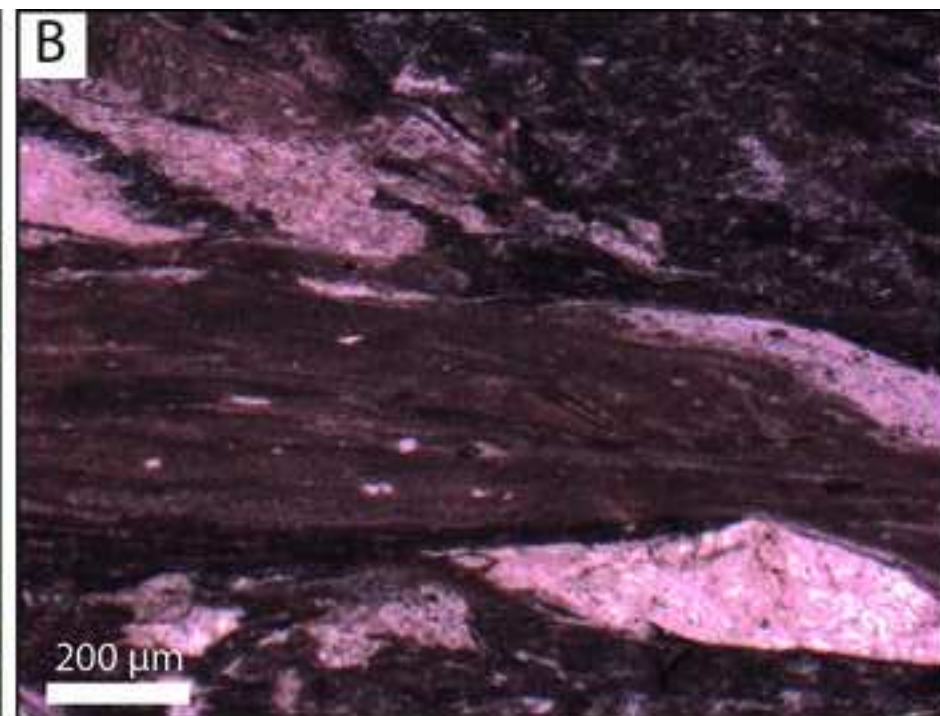
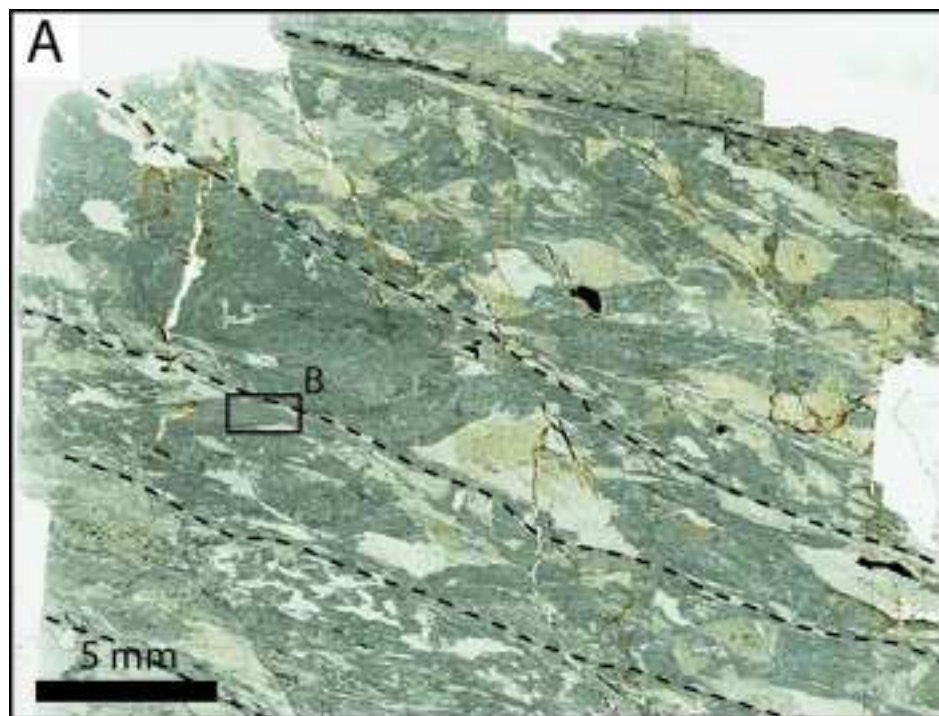
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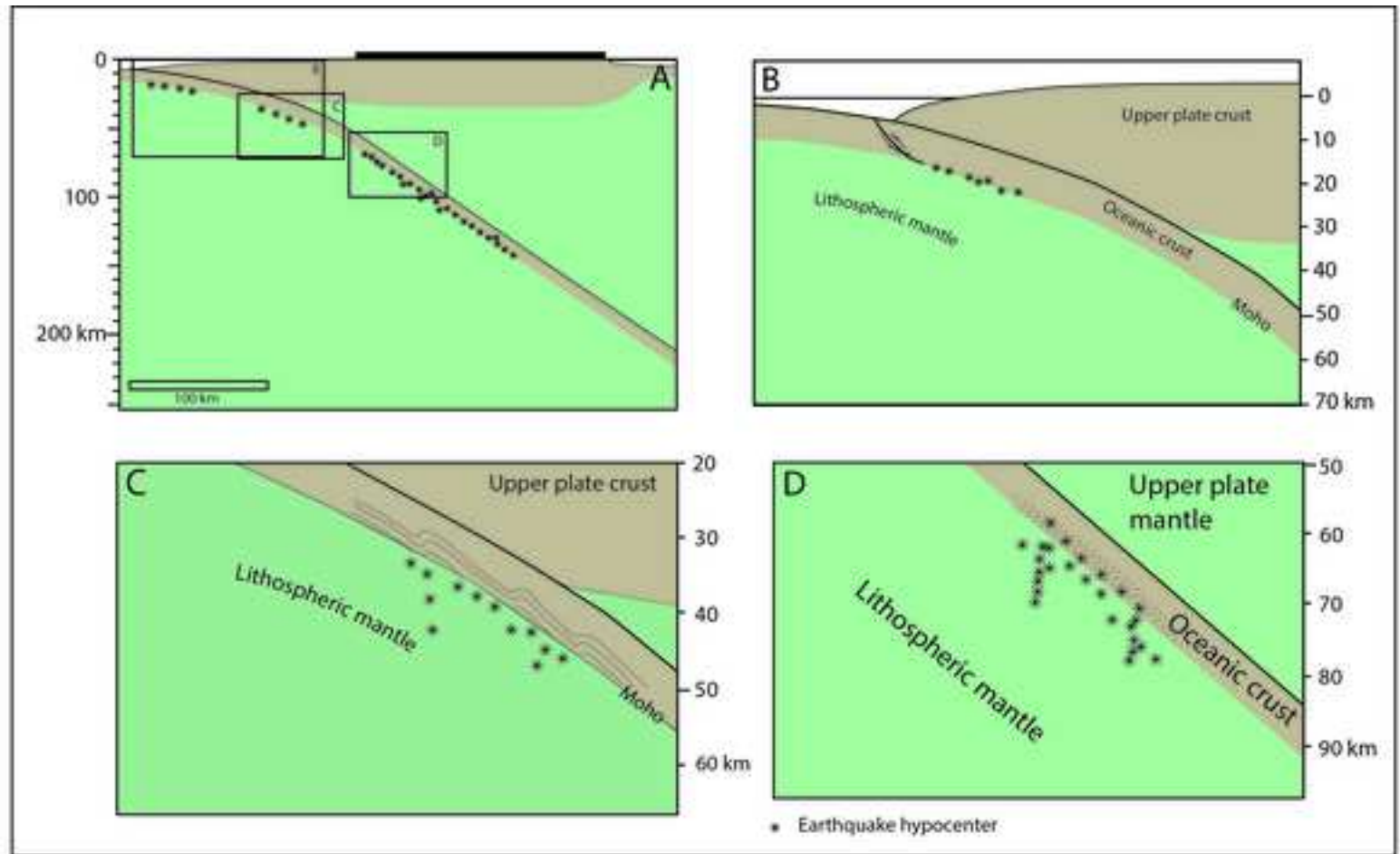
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