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1 The northwest-directed "Bretonian phase" in the French Variscan Belt (Massif  
2 Central and Massif Armoricain): a consequence of the Early Carboniferous  
3 Gondwana-Laurussia collision

4

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13

#### 14 **ABSTRACT**

15 In the Variscan French Massif Central and Massif Armoricain, the tectonic  
16 significance of a widespread NW-SE trending stretching lineation, coeval with medium  
17 pressure-medium temperature metamorphism, is an open question. Based on a structural  
18 analysis in the Southern part of the Massif Central, we show that this top-to-the-NW shearing  
19 is a deformation event, referred to as D2, that followed a D1 top-to-the-South shearing  
20 Devonian phase, and was itself re-deformed by a Late D3 Viséan-Serpukhovian southward  
21 thrusting event. We date the D2 phase at 360 Ma (Famennian-Tournaisian boundary). In the  
22 Massif Armoricain, D2 is the "Bretonian phase" recorded in the metamorphic series and  
23 sedimentary basins. Geodynamically, D2 is related to a general northwestward shearing  
24 during the Laurussia-Gondwana collision, which occurred after the closure of the Rheic  
25 Ocean, as indicated by the emplacement of the Lizard ophiolitic nappe in Britain. The left-

26 lateral Nort-sur-Erdre fault accommodated the absence of ductile shearing in Central  
27 Armorica.

28

29 **Keywords:** Variscan belt, Bretonian phase, French Massif Central, Massif Armoricain

30

## 31 **1. Introduction**

32

33 It is widely acknowledged that the Variscan Belt of Western Europe formed as a result  
34 of the multiple collisions that occurred between Gondwana to the South and Laurussia to the  
35 North, from **Early Devonian to Carboniferous** times. These collisions stacked together  
36 continents and micro-blocks such as Armorica or Mid-German Crystalline Rise that had been  
37 previously drifted from Gondwana (e.g. Holder and Leveridge, 1986; Pin, 1990; Matte, 1986,  
38 2001; Franke, 2000; Faure et al., 2005, 2008; Lardeaux et al., 2014). In the French Massif  
39 Central (FMC, Fig. 1A), a polyphase syn-metamorphic deformation is well documented (e.g.  
40 Faure et al., 2009 and enclosed references). By analogy with the Himalayas, it was initially  
41 proposed that a ductile south-directed shearing, active from Late Devonian to Middle  
42 Carboniferous accommodated the nappe stacking with younging from North to South (e.g.  
43 Mattauer 1974; Mattauer and Etchecopar, 1977; Matte, 1986; Ledru et al., 1989). Top-to-the-  
44 South shearing is indeed well established in the southern part of the French Massif Central  
45 (Montagne Noire and Cévennes, Fig. 1) –where it is referred to as a “D3 event”, but more to  
46 the North and West, in Lot, Rouergue, Lyonnais, Sioule and Limousin, the dominant structure  
47 is instead a NW-SE stretching lineation with a top-to-the-NW shearing. The tectonic  
48 significance of this NW-SE deformation and its relations with the widespread southward D3  
49 shearing are not clear. According to Brun and Burg, (1982) and Burg et al., (1987), the top-to-  
50 the-NW shearing would be due to a combination of southward thrusting and sinistral  
51 wrenching during the ibero-armoricain oroclinal bending. Mattauer et al., (1988) considered

52 this deformation as a consequence of extensional tectonics, whereas Bouchez and Jover  
53 (1986), Friedrich et al., (1988), Faure et al. (2005), and Bellot and Roig (2007) argued that it  
54 corresponds to a nappe stacking. This event is referred to as the "D2 event" (Faure et al.,  
55 2009).

56 Here we present structural and dating data in the Rouergue-Albigeois (Fig. 1B) area,  
57 which allow us to discriminate and date the S- and NW-ward displacement events (D3 and  
58 D2). We report similar observations in the Massif Armoricain. We propose a geodynamic  
59 interpretation of the D2 event in the two massifs in the general framework of the Laurussia-  
60 Gondwana continental collision.

61

62

## 63 **2. The architecture of the Southern Massif Central**

64 The French Massif Central stack of metamorphic nappes belongs to the northern  
65 Gondwana margin. From top-to-bottom, the following units are recognized: 1) the Upper  
66 Gneiss Unit, 2) the Lower Gneiss Unit, 3) the Para-Autochthonous domain, 4) the Fold-and-  
67 Thrust Belt, and 5) the Southern Foreland. South of the Margeride pluton, the nappe  
68 architecture exhibits several particularities, which we describe below (Figs. 1, 2).

69 *2.1 The Upper Gneiss Unit* (UGU) crops out as several klippes (Marvejols, Vibal,  
70 Lévézou, Najac, and Decazeville). It is made of an association of felsic-mafic-ultramafic  
71 rocks and sedimentary rocks, called the leptynite-amphibolite complex. This complex  
72 experienced a high pressure-medium temperature (HP/MT) metamorphism (Nicollet and  
73 Leyreloup, 1978; Burg et al., 1986, 1989; Bodinier and Burg, 1980-1981) at  $415\pm 5$  Ma  
74 (U/Pb) or  $408\pm 7$  Ma (Sm/Nd), thus in Late Silurian-Early Devonian (Pin and Lancelot, 1982;  
75 Paquette et al., 1995). During their exhumation, most of the eclogites were retrogressed in the  
76 amphibolite facies and the melting of the Al-rich part produced migmatite. Due to its circular  
77 shape in map view and the abundance of anatexites in its center, the southernmost Lévézou

78 klippe has long been considered as a diapiric dome (Collomb, 1970; Burg, 1987; Collomb and  
79 Meyzindi, 1991). However, structural and gravimetric data show that the amphibolites and  
80 migmatites are not rooted in the Lower Gneiss Unit but rather overlie it (Matte, 1986; Bayer  
81 and Hirn, 1987). The UGU klippe have thus been displaced to the SW. As a matter of fact,  
82 northwestward overturned folds deform the UGU foliations. As we explain below, they are  
83 attributed to the D2 deformation phase.

84       2.2. *The Lower Gneiss Unit (LGU)* consists of metagreywackes and metapelites that  
85 never experienced any HP/MT metamorphism. Several plutons that intrude the LGU, such as  
86 the Caplongue diorite ( $557 \pm 10$  Ma, U/Pb on zircon, Lafon, 1984), the Rodez alkaline granite,  
87 and the Pinet and Comps porphyritic monzogranites, have been transformed into augen  
88 orthogneiss.

89       2.3. *The Para-autochthonous Unit (PAU)*, well exposed in Cévennes, consists of a  
90 thrust sheet-imbriation of greenschist facies metapelites, quartzites, and metagrauwackes  
91 with subordinate layers of conglomerate, felsic and mafic lava, and rare intrusions (Pin and  
92 Marini, 1993; Caron, 1994). In the Albigeois area, the PAU has been subdivided into the S<sup>t</sup>-  
93 Sernin-sur-Rance and the S<sup>t</sup>-Salvi-de-Carcavès nappes (Guéranzé-Lozes and Guéranzé, 1984;  
94 Guéranzé-Lozes, 1987; Guéranzé-Lozes and Burg, 1990). The lithostratigraphy of these two  
95 nappes is similar, with, from base to top, Cambrian greywacke, rhyolite and ignimbrite,  
96 Ordovician white quartzite, and black metapelite. However, a biotite-garnet  $\pm$ staurolite  
97 assemblage is fairly common in the former but is absent in the latter.

98       2.4. *The Fold-and Thrust Belt*, developed in the Montagne Noire and the Viganais, is  
99 composed of unmetamorphosed fossiliferous Paleozoic sedimentary rocks terminated by a  
100 Late Viséan to Serpukhovian turbiditic basin. The series have been deformed, by South-  
101 directed thrust sheets and km-scale recumbent folds (Gèze, 1949; Arthaud, 1970; Engel et al.,  
102 1980-81). The Montagne Noire axial zone is a migmatitic dome that cross-cuts the folds and  
103 thrusts. This area is beyond the scope of this paper.

104

105

### 106 **3. Polyphase shearing in the Southern Massif Central**

107       The pervasive foliation developed in the UGU, LGU, and PAU units results from  
108 three synmetamorphic ductile deformations, D1, D2, D3, each associated with specific  
109 kinematics and P-T conditions. The three deformation phases are described below from the  
110 youngest to the oldest.

111       *3.1. The D3 Carboniferous southward thrusting phase.* Spectacular km-scale,  
112 southward overturned recumbent folds characterize the Paleozoic Fold-and-Thrust Belt (e.g.  
113 Arthaud, 1970, Arthaud and Matte, 1977). In the Southern Montagne Noire, folding is coeval  
114 with horizontal cleavage and rare N-S to N30E trending mineral and stretching lineation (Fig.  
115 1). A more dominant lineation exists, that strikes NE-SW (Lee et al 1988; Echtler and  
116 Malavielle, 1990). This NE microstructure is a late feature since it overprints the already  
117 deformed series. The NE lineation is coeval with a HT/LP metamorphic event that has been  
118 related to the doming of the axial zone (e.g. Courtillot et al., 1986; Echtler and Malavieille,  
119 1990; Faure et al. 2014). In the Northern Montagne Noire and the S<sup>t</sup>-Salvi-de-Carcavès  
120 nappe, the submeridian stretching lineation is coeval with a top-to-the-S sense of shear  
121 (Brunel et al., 1974; Guérangé-Lozes, 1987). In map view, the basal thrust of the D3 S<sup>t</sup>-Salvi-  
122 de-Carcavès nappe truncates the bedding surface (Figs. 1, 2). The D3 event recognized in the  
123 PAU series in Cévennes and at the base of the UGU series in Marvejols is dated at ca 335-  
124 325 Ma and 340-335 Ma, respectively (Pin and Lancelot, 1982; Caron, 1994; Faure et al.,  
125 2001). Therefore, the D3 event is clearly younging southward. Moreover, the D3 structures  
126 are locally overprinted by a top-to-the-north shearing related to the Late Carboniferous  
127 extensional tectonics (Arnaud and Burg, 1993) that are not considered here (cf. Faure et al.,  
128 2009 for details).

129           3.2. *The D2 Early Carboniferous top-to-the-NW shearing phase.* The main  
130 deformation of the S<sup>t</sup>-Sernin-sur-Rance nappe is characterized by a flat-lying foliation and a  
131 NW-SE trending mineral and stretching lineation ascribed to the D2 phase. Near Naucelle  
132 (Fig. 1), top-to-the NW shear criteria coeval with a biotite-garnet-staurolite metamorphism is  
133 observed (Fig. 3C, E, F; Burg et al., 1986; 1989). A flat-lying mylonitic fabric with NW-SE  
134 trending folds with axes parallel to the stretching lineation developed during the ductile  
135 shearing that accommodated the tectonic superposition of the S<sup>t</sup>-Sernin-sur-Rance nappe  
136 upon the LGU (Fig. 3A, B). Further east, the dip of the foliation increases progressively to  
137 become SW east of Naucelle, and vertical south of the Lévézou klippe. The kinematic  
138 consistency from a top-to-the-NW shearing developed in the flat-lying foliation to dextral one  
139 observed in vertical surfaces with subhorizontal stretching lineation suggests a wrench-thrust  
140 structure. The D2 deformation phase is not recognized in the S<sup>t</sup>-Salvi-de-Carcavès nappe and  
141 farther south, but in contrast, it is clearly observed in the north in the LGU, and to a lesser  
142 extent, in the UGU units.

143           3.3. *The D1 early southwestward shearing phase.* The D2 top-to-the-NW shearing  
144 event was preceded by an earlier deformation phase that is well observed in the UGU and  
145 LGU series, in the S<sup>t</sup>-Sernin-sur-Rance and S<sup>t</sup>-Salvi-de-Carcavès nappes in the Rouergue  
146 area. Where the units are not deformed by the later D2 and D3 events, a N-S to N30°E  
147 trending stretching lineation is observed, associated with intrafolial folds and top-to-the-SW  
148 shearing (Figs. 1, 3D). Yet, at many places, the D2 NW-SE trending folds deformed this early  
149 mineral lineation (Fig. 3A,B D). In the S<sup>t</sup>-Sernin-sur-Rance nappe, km-scale S-verging  
150 recumbent folds appear re-deformed by the D2 deformation, and hence are attributed to the  
151 D1 phase (Guérangé-Lozes, 1987). The D1 top-to-the-S stretching lineation is observed in the  
152 Rodez orthogneiss of the LGU series, and in the UGU gneiss, amphibolite and migmatite of  
153 the Lévézou and Vibal klippen.

154 The foliated Pinet orthogneiss has recorded both the D1 and the D2 deformation  
155 events. Although the D2 NW-SE lineation is dominant in many places (Fig. 1, Burg and  
156 Teyssier, 1983), an older N-S trending D1 mineral lineation is observed where the D2  
157 overprint is lacking. Petro-structural analysis documents a top-to-the-south shearing coeval  
158 with a high-temperature deformation, whereas low-temperature top-to-the-NW shear bands  
159 are ascribed to the D2 phase (Fig. 3D, Duguet and Faure, 2004).

160

161

#### 162 **4. Chronological constraints**

163 *4.1 Timing of the polyphase shearing in the Southeast Massif Central.* Due to their  
164 similar submeridian trend and top-to-the-South sense of shear, the D1 and D3 deformation  
165 events have been so far considered as the result of a single, long duration intracontinental  
166 shearing period, operating during the "Himalayan thrusting" (Mattauer, 1974, Mattauer and  
167 Etchecopar, 1977). However, we have shown that the D1 and D3 events are diachronous  
168 since they are separated in time by nearly 40 Ma. The age of the D3 event is well constrained  
169 as this event deforms the Late Viséan-Early Serpukhovian (330-325 Ma) syntectonic flysch  
170 basin of the Southern Montagne Noire. Muscovite populations from the Mendic orthogneiss  
171 and the S<sup>t</sup>-Salvi-de-Carcavès nappe metasandstone yield <sup>40</sup>Ar/<sup>39</sup>Ar ages of 330±3 Ma and  
172 333±4 Ma, respectively (Costa, 1990). These Viséan ages are similar to those available in the  
173 Cévennes area (Caron, 1994; Najoui et al., 2000).

174 The biotites, muscovites and amphiboles aligned along the NW-SE lineation reveal  
175 <sup>40</sup>Ar/<sup>39</sup>Ar ages in the range 354-346 Ma. Since the closure temperature of micas is lower than  
176 that of the metamorphic climax, the above ages provide the minimum age of the D2 event. In  
177 the S<sup>t</sup>-Sernin-sur-Rance nappe, muscovites from quartzites yield a 345±3.5 Ma age. In the  
178 UGU and LGU series, biotites, muscovites and amphiboles yield consistent <sup>40</sup>Ar/<sup>39</sup>Ar ages in



179 the range 354-346 Ma (Costa, 1990). The age of the D1 event is still poorly constrained with  
180 only one  $^{40}\text{Ar}/^{39}\text{Ar}$  muscovite age of  $380\pm 10$  Ma (Guérangé-Lozes, 1987).

181         Whatever the precise absolute age of the three D1, D2 and D3 events, their relative  
182 timing allows us to draw an interpretative general cross-section (Fig. 2). The top-to-the-SW  
183 D1 structures were folded in the Late Devonian-Early Carboniferous by the northwestward  
184 D2 deformation. Later, during the Middle Carboniferous, the D3 southward shearing event re-  
185 deformed the whole stack. This new interpretation makes the Pinet orthogneiss a pre-D1  
186 pluton.

187         *4.2. New zircon U-Pb age of the Pinet orthogneiss.* The Pinet orthogneiss is a  
188 monzogranite or syenogranite foliated in a variable way, that includes both nearly  
189 undeformed facies at local spots, and mylonitic to ultramylonitic shear zones at other spots. A  
190 detailed petro-structural study of the Pinet orthogneiss has revealed that the pluton  
191 experienced two deformation phases, ascribed to the D1 and D2 events (Duguet and Faure,  
192 2004). The early deformation phase is coeval with a pervasive post-solidus planar fabric and a  
193 N-S striking lineation. The second deformation phase is characterized by a local foliation  
194 cross-cutting the previous one, and a NW-SE striking lineation with top-to-the-NW shearing.  
195 Zircon populations in the granite yielded a TIMS U-Pb age of  $ca\ 360\pm 20$  Ma (Pin, 1981),  
196 whereas a biotite single grain was dated at  $346\pm 7$  Ma using the  $^{40}\text{Ar}/^{39}\text{Ar}$  method (Maluski  
197 and Monié, 1988).

198         We separated zircon concentrates from two samples (FR 41 and FR 42) of the  
199 porphyritic orthogneiss, located along the Tarn river, at  $44^{\circ} 03' 16.27''/002^{\circ} 45' 56.35''$  and  
200  $44^{\circ} 02' 43.57''/002^{\circ} 46' 47.70''$ , respectively. We applied analytical procedures and data  
201 reduction as in Li et al. (2009) and Do Couto et al. (2016). Cathodoluminescence images of  
202 the zircon grains show a well developed oscillatory zoning representative of a magmatic  
203 origin of the grains (Fig. 5).

204 Zircons from the two rock samples yield consistent Ordovician ages of  $477\pm 5$  Ma and  
205  $475\pm 4$  Ma (Fig. 5); these ages are interpreted as dating the emplacement of the porphyritic  
206 granite. This new result shows that the Pinet orthogneiss is not a syn-kinematic pluton  
207 emplaced during the D1 or D2 events, but instead formed in the Early Ordovician pre-  
208 orogenic magmatism phase that is widespread in the Variscan belt. The ca 350 Ma  $^{40}\text{Ar}/^{39}\text{Ar}$   
209 age of post-folial biotite corresponds to the D2 event.

210

211

## 212 **5. Discussion-conclusion**

213 *5.1. Significance of the three D1, D2 and D3 tectono-metamorphic events.* The D1  
214 event is recognized in the Lyonnais, Sioule, Limousin, and Plateau d'Aigurande areas of the  
215 Massif Central. It is coeval with the Middle Devonian (385-375 Ma) migmatization, that  
216 was interpreted as the result of the exhumation of the HP rocks of the UGU series (cf Faure et  
217 al., 2008 for details).

218 The Late Devonian-Early Carboniferous (360-350 Ma) D2 event with a top-to-the-  
219 NW shearing is also recognized in many places of the Massif Central (Fig. 6). In the LGU  
220 series of Marvejols (Fig. 1), biotites from metadiorite and micaschist, aligned along the NW-  
221 SE lineation, yield consistent  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of  $352\pm 2$  and  $351\pm 3$  Ma (Costa, 1990). The  
222 Limousin, Sioule and Plateau d'Aigurande series are affected by a consistent medium  
223 temperature-medium pressure metamorphism coeval with a top-to-the-NW shearing (e.g.  
224 Floc'h 1983; Bouchez and Jover, 1986; Friedrich et al., 1988; Faure et al., 1990, 1993, 2005;  
225 Roig and Faure, 2000; Bellot and Roig, 2007). Since the  $^{40}\text{Ar}/^{39}\text{Ar}$  dates are very sensitive to  
226 the temperature, the monazite U-Th-Pb chemical chronometer that records high temperature  
227 events has also been carried out in the UGU, LGU, and PAU metapelites of Limousin and  
228 Sioule areas (Melleton et al., 2009; Do Couto et al., 2016). The 360-350 Ma ages yielded by  
229 the syn-D2 metamorphic monazites confirm the Famennian-Tournaisian age of the D2 event.

230 Furthermore, in eastern Massif Central, the Brévenne ophiolitic nappe was emplaced from SE  
231 to NW under amphibolite facies conditions, before the unconformable deposition of the Early  
232 Visean (i.e. 345 Ma) Le Goujet, terrigenous formation (Leloix et al., 1999). Thus the  
233 Brévenne ophiolitic nappe formed also during the D2 event.

234 The D3 event is identified only in the South Massif Central. The Visean to  
235 Serpukhovian southward migration from middle to shallow crustal levels is acknowledged  
236 since a long time as the consequence of the post-collisional nappe stacking (Matte, 1986;  
237 Ledru et al., 1989, Faure et al., 2009).

238

239 *5.2. Geodynamic significance of the D2 event in the French Variscan segment.* In the  
240 southern Massif Armoricaïn that belongs to the metamorphic zone of the Variscan chain, the  
241 D2 synmetamorphic shearing event is recognized in Vendée and south Brittany coast (Burg,  
242 1981; Cannat and Bouchez, 1986; Fig. 6). There, an E-W to NW-SE striking mineral lineation  
243 develops on a flat-lying foliation, coeval with a MP/MT metamorphism characterised by a  
244 biotite-garnet-staurolite assemblage. Top-to-the-NW shearing shown by asymmetric pressure  
245 shadows and shear bands is conspicuous.

246 More to the North, in Central Armorica, top-to-the-NW shearing is described only in  
247 the S<sup>t</sup>-Georges-sur-Loire and Lanvaux units (Bouchez and Blaise, 1976; Diot et al., 1983;  
248 Cogné et al., 1983; Faure and Cartier, 1998; Cartier et al., 2001; Cartier and Faure, 2004).  
249 The pre-Visean age of the deformation is attested by the age of the terrigenous deposits of the  
250 Ancenis basin that did not record the D2 event. Therefore, the top-to-the-NW shearing, that is  
251 widespread in the Massif Armoricaïn and South Massif Central, is a major feature of the  
252 Variscan orogeny.

253 The MP/MT D2 metamorphism dated at ca 360 Ma, is coeval with the Devonian-  
254 Carboniferous "Bretonian phase" long recognized in the central part of the Massif  
255 Armoricaïn, through the Tournaisian unconformity and the widespread erosion of the Late

256 Devonian formations (e.g. Stille, 1929 in Rolet, 1982). In the 70's and 80's, on the basis of the  
257 gentle folding of the pre-Tournaisian formations, and the small unconformity angle (lower  
258 than 20°), the Bretonian phase was considered as a km-scale upright folding coeval with  
259 vertical movements, responsible for erosion and clastic sedimentation in the Laval and  
260 Châteaulin basins (e.g. Cogné 1965, 1974; Pelhâte, 1971; Paris et al., 1982; Houlgatte et al.,  
261 1988, Le Gall et al. 1992; Fig. 6). However, recumbent folds, ductile shear zones, and low  
262 angle thrust faults identified in the Brest area (Rolet et al., 1986) allow us to reassess the  
263 importance of the Bretonian phase.

264         The Variscan orogeny is the consequence of the closure of the Rheic ocean, and  
265 subsequent collision of Laurussia that was driven by a southward oceanic and then  
266 continental subduction below Gondwana. The Rheic suture is today hidden below the English  
267 Channel, but the Lizard ophiolitic nappe of SW Britain that overthrusts to the NNW the  
268 Laurussia foreland at ca 360 Ma is a remnant of the Rheic Ocean (e.g. Holder and Leveridge,  
269 1986; Le Gall and Darboux, 1986; Sandeman et al., 1994). Therefore we argue here that the  
270 ductile and synmetamorphic, top-to-the-NW, shearing observed in the French Massif Central  
271 and South Armorica, and more broadly the Famennian-Tournaisian deformation ascribed to  
272 the Bretonian phase, are the result of the collision between Laurussia and Gondwana. **Top-to-**  
273 **the NW shearing observed at the microtectonic scale results of the Variscan collision.**

274         Although close to the Rheic suture, (Fig. 6), the syn-metamorphic ductile deformation  
275 does not exist in Central Armorica. A possible explanation would be to consider the  
276 contrasted mechanical behaviour between the Massif Central - South Armorican domain with  
277 a widespread D2 event, and the Central Armorican domain where D2 is absent. Due to its  
278 crustal rigidity, possibly related to the Neoproterozoic Cadomian orogeny, and the thick sub-  
279 continental mantle lithosphere, the Central Armorican domain behaved as a rigid plate  
280 characterized by sinistral wrenching, responsible for the opening of the Châteaulin and Laval  
281 basins (Fig. 7). In its southern margin, along the Nort-sur-Erdre strike-slip fault, and the S<sup>t</sup>-

282 Georges-sur-Loire and Lanvaux units, the sinistral shearing accommodated the decoupling  
283 between the ductile and brittle domains (Cogné et al., 1983; Diot et al., 1983; Lardeux and  
284 Cavet, 1994; Faure et al., 1997; 2005; Cartier et al., 2001; Cartier and Faure, 2004). This  
285 interpretation, which is at variance to the previous geodynamic models (e.g. Ballèvre et al.,  
286 2009), provides a satisfactory solution for the significance of the D2 event in the French  
287 Variscan orogen.

288

289

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597

598

599 Figure Captions

600 Fig. 1: A Insert showing the Variscan belt in France. NASZ: North Armorican Shear Zone,  
601 NBSASZ: North Branch of the South Armorican Shear Zone, SBSASZ: South Branch of the  
602 South Armorican Shear Zone, NPF: North Pyrenean Fault, MGCR: Mid-German Crystalline  
603 Rise. B: simplified structural sketch of the French Massif Central (purple areas: Upper Gneiss  
604 Unit; the Lower Gneiss Unit and Para-Autochthonous Unit are not distinguished). C: Structural  
605 map of the S part of the French Massif Central with emphasis on the compressional  
606 structures. Syn- to late orogenic extensional structures are not considered here. Captions are  
607 the same as for Fig. 2.

608

609 Fig. 2: Crustal scale cross-section through the Montagne Noire-Rouergue area (location in  
610 Fig. 1). The bulk architecture results of three shearing events, D1, D2, and D3, with different  
611 ages and displacement directions.

612

613 Fig. 3: Field and thin section photographs showing D1 and D2 deformation in the S<sup>t</sup>-Sernin-  
614 sur-Rance nappe and Lower Gneiss Unit. A: NW-SE trending D2 fold (L2) reworking an  
615 early NE-SW trending L1 lineation; B: NW-SE trending "a-type" fold deforming the L1

616 lineation; C: Low-temperature D2 top-to-the-NW shear zone reworking the foliated Pinet  
617 granodiorite; D: N-S striking intrafolial D1 fold; E: Thin section of staurolite porphyroblast  
618 with top-to-the-NW, D2, asymmetric biotite pressure shadows; F: Garnet with top-to-the NW,  
619 D2, pressure shadows.

620

621 Fig. 4: Cathodoluminescence images of representative zircons for SIMS U–Pb dating. White  
622 ellipses indicate the in-situ analytical spots of  $^{206}\text{Pb}/^{238}\text{U}$  age (shown nearby the spot). SIMS  
623 spots are 30  $\mu\text{m}$  in length for scale.

624

625 Fig. 5: Zircon ion-probe U-Pb age of the Pinet orthogneiss. Samples FR 41 and FR 42 are  
626 located at  $44^\circ 03' 16.27''/002^\circ 45' 56.35''$ , and  $44^\circ 02' 43.57''/002^\circ 46' 47.70''$ , respectively.

627

628 Fig. 6: Structural map of the French-Britain segment of the Variscan belt with emphasis on  
629 the D2 NW-SE shearing developed from Albigeois up to the Lizard nappe. The Late  
630 Carboniferous, south-directed, D3 event develops in southern Massif Central (Cévennes and  
631 Montagne Noire). At the scale of the entire Variscan belt, a top-to-the-North-directed  
632 deformation, characterized by thrust and faults, deforms the Ardenne Massif and the northern  
633 Variscan front.

634

635 Fig. 7: 3D Schematic geodynamic interpretative model to account for the tectonic difference  
636 between the Massif Central-Massif Armorica areas dominated by the top-to-the D2  
637 synmetamorphic ductile shearing, and the Central Armorican Domain, where D2 deformation  
638 is represented by brittle structures only. This difference in rheological behaviour can be  
639 explained by a tectonic decoupling between the two domains. The Neoproterozoic Cadomian  
640 orogeny, and the thick subcontinental lithosphere mantle may explain the rigid behaviour of  
641 the Central Armorican Domain in which the brittle deformation is associated with the opening

642 of the Châteaulin and Laval basins. The left-lateral Nort-sur-Erdre strike-slip fault can be  
643 seen as a transfer fault between the S. Gondwana and Central Armorican domains.

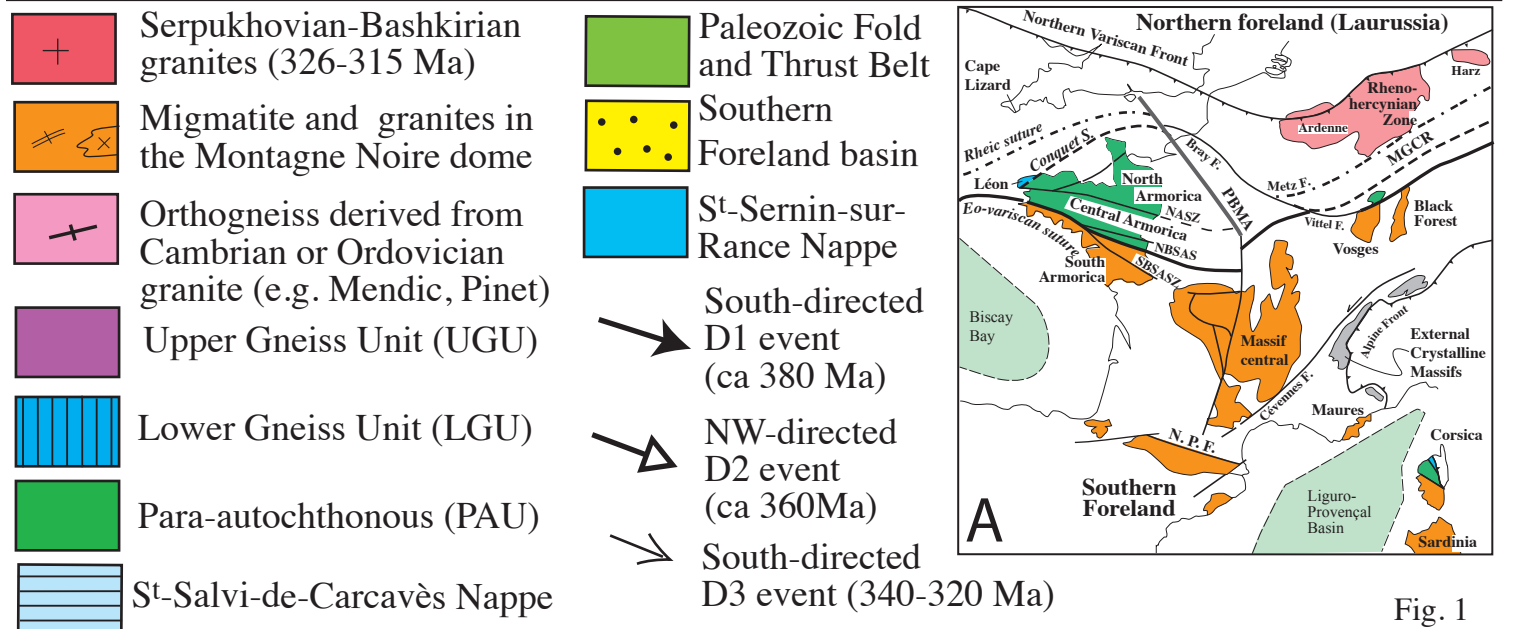
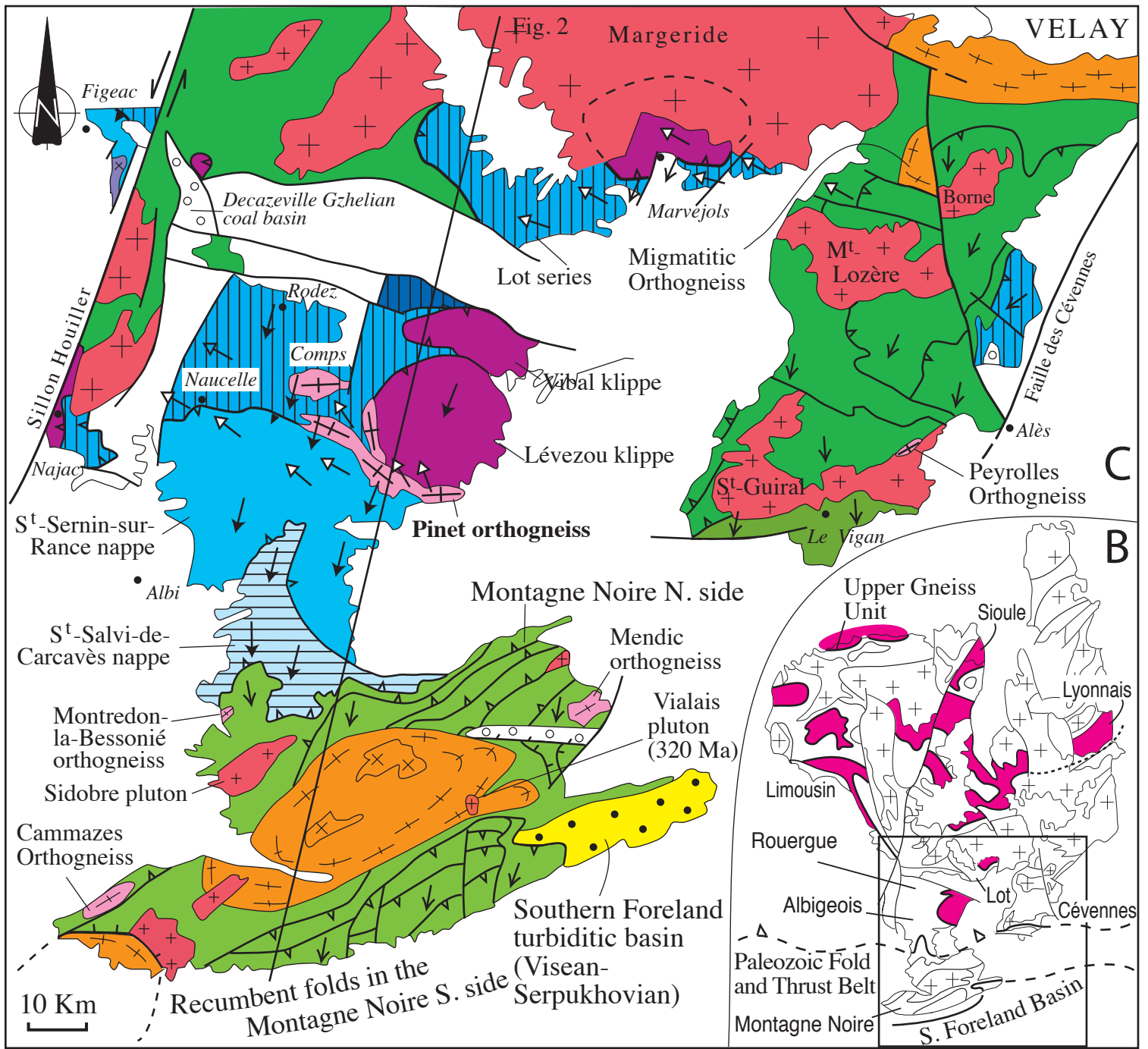


Fig. 1

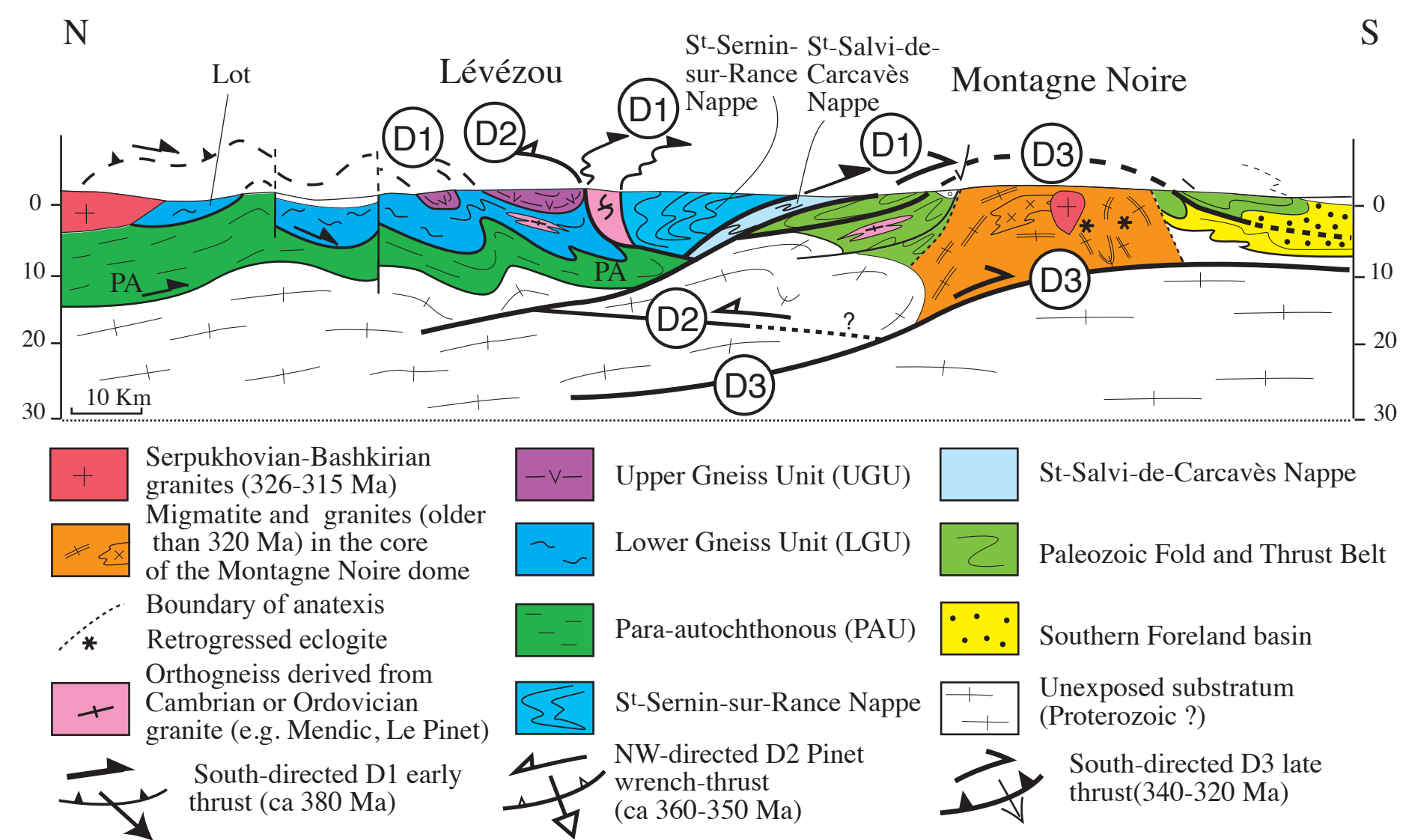


Fig. 2

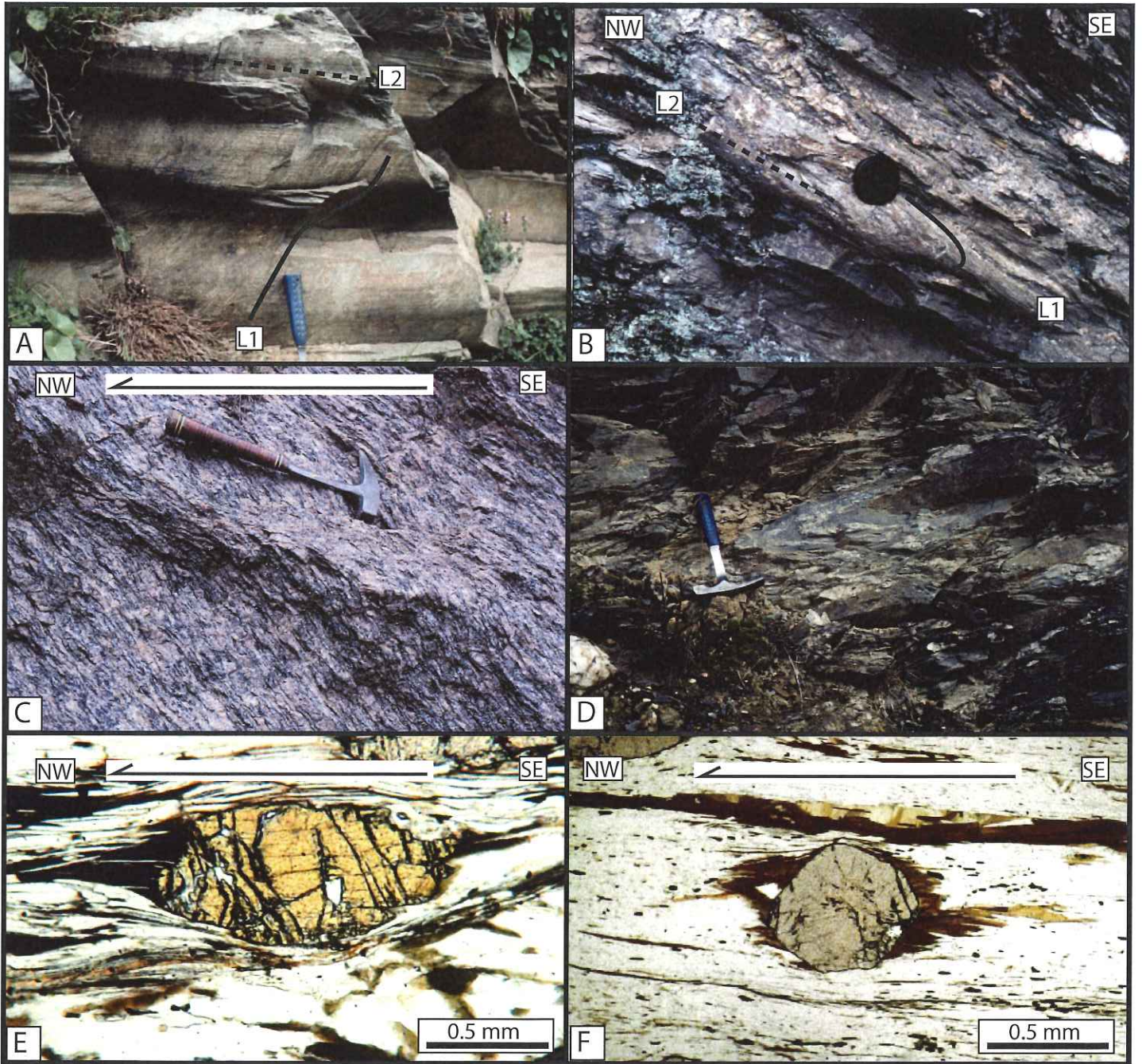
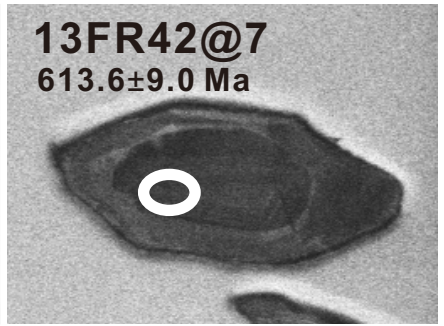
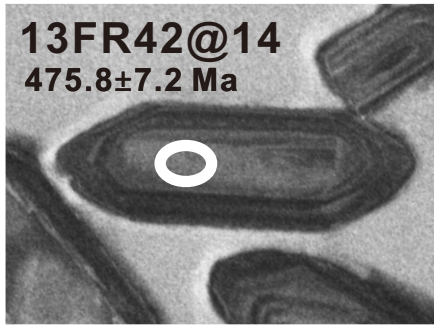
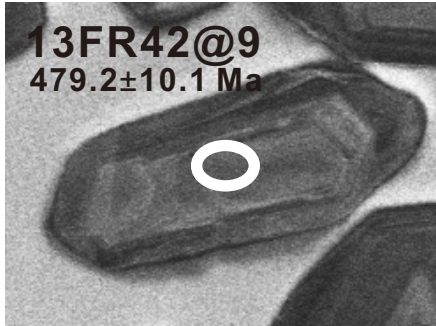
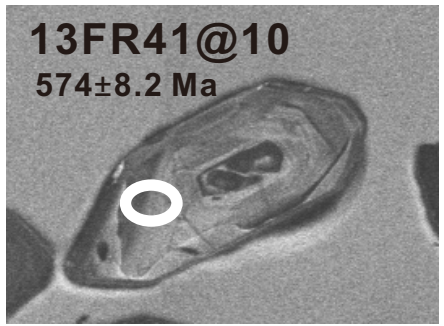
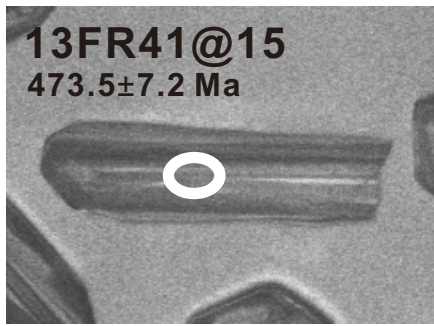
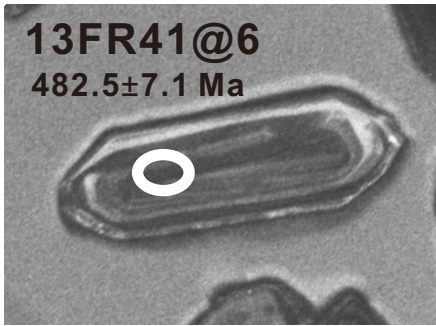


Fig. 3



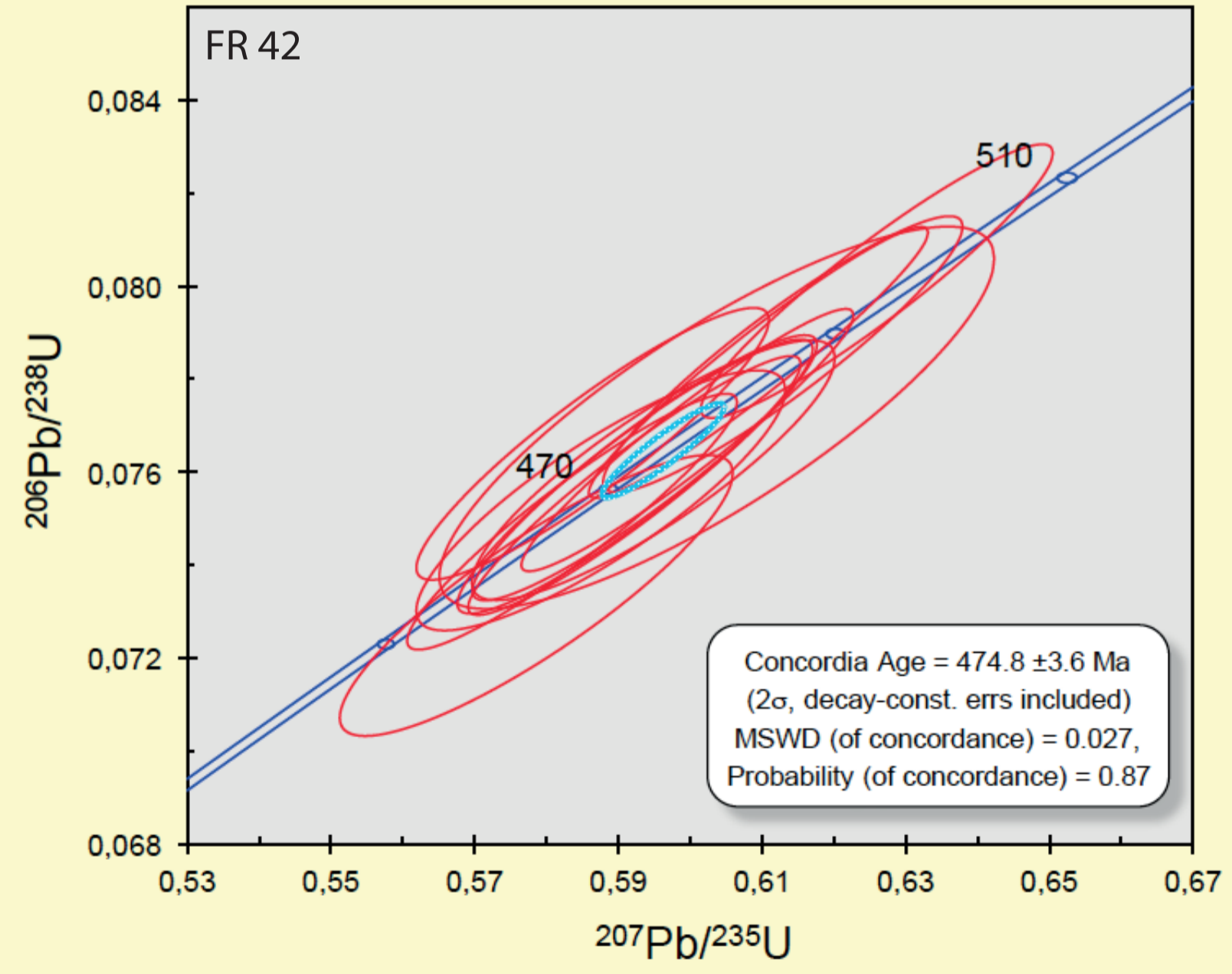
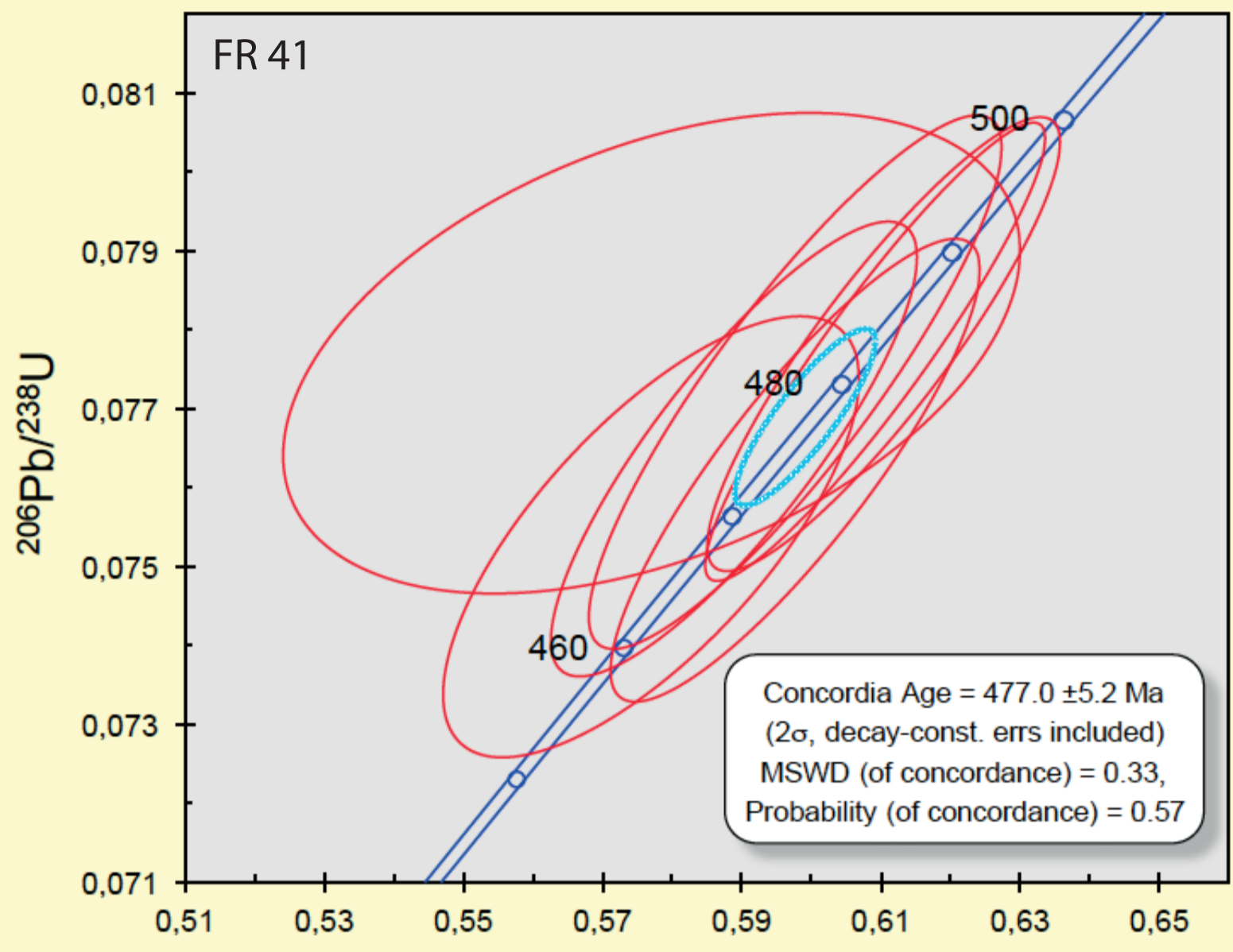
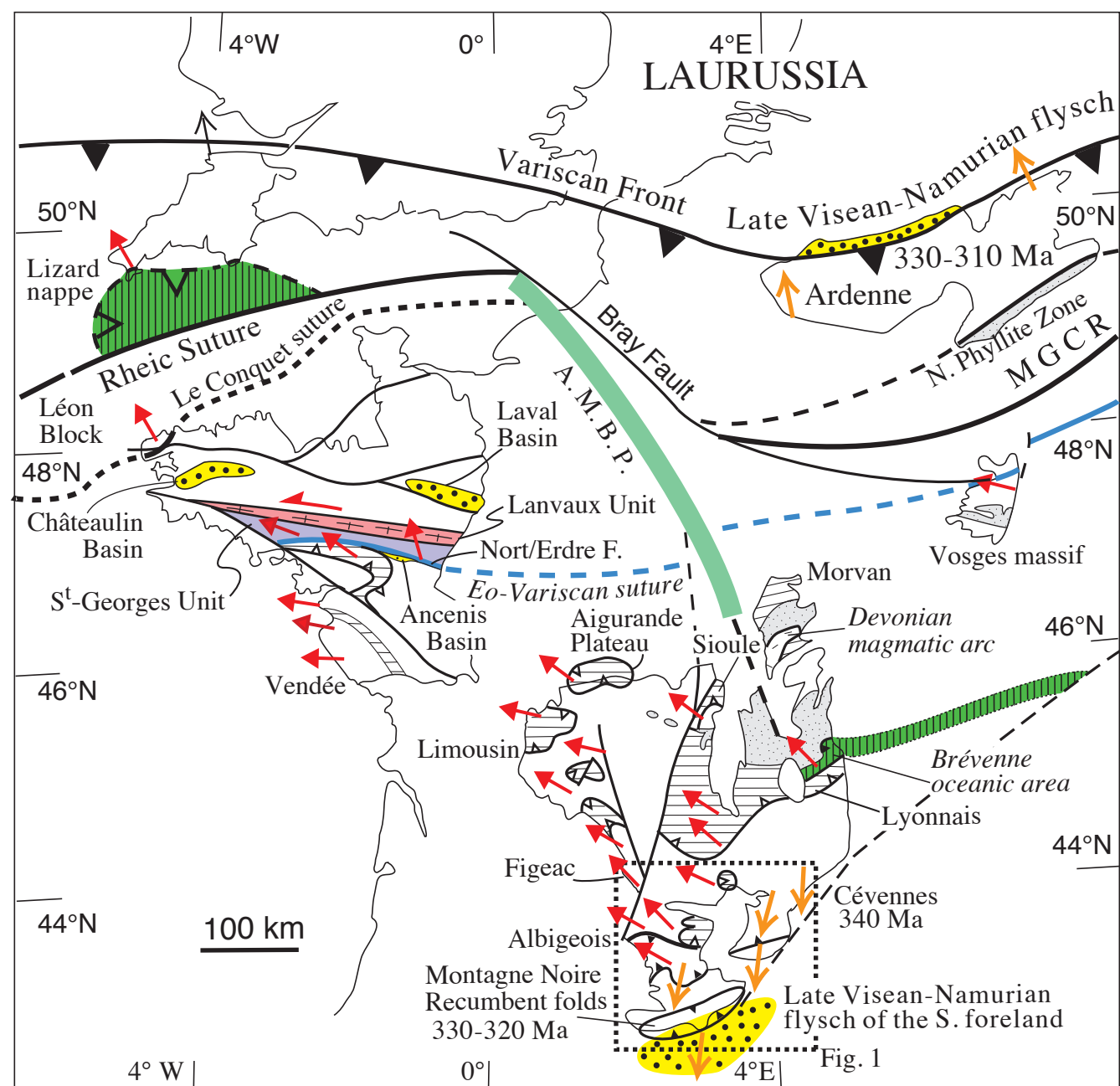


Fig. 5





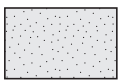


-  Post-orogenic formations:  
Late Visean "tufs anthracifères"  
Red granites and microgranites
-  360 Ma lineation related to the D2 event
-  Upper Gneiss Unit with  
420 Ma high pressure rocks
-  Late Visean-Namurian thrust with  
nappe displacement sense (D3 event)

Fig. 6

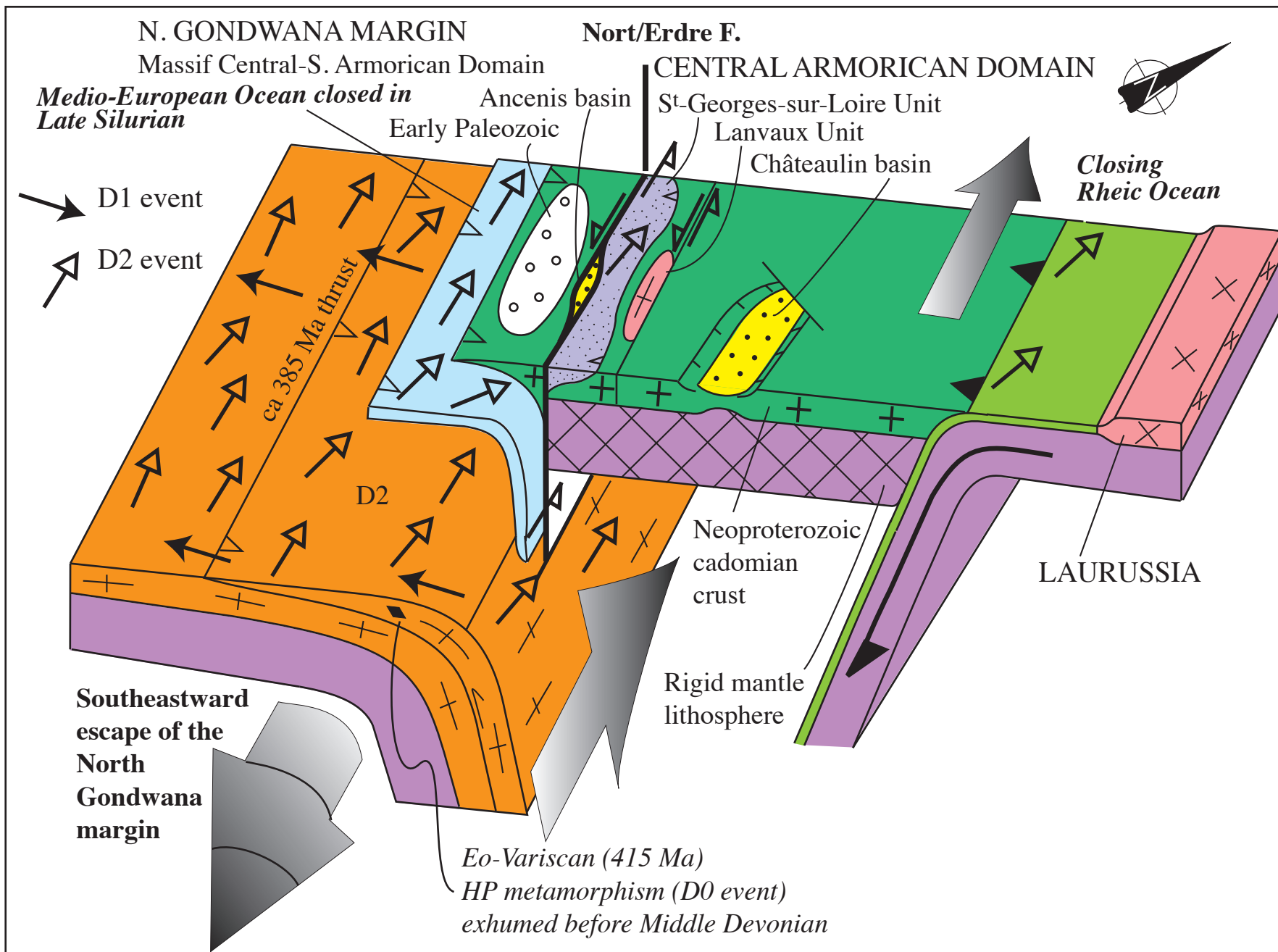


Fig. 7