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1 **Structural inheritance in the Central Pyrenees: The Variscan to Alpine tectono- metamorphic**
2 **evolution of the Axial Zone**

3

4 Bryan COCHELIN ^{a,b}*, Baptiste LEMIRRE^a, Yoann DENELE^a, Michel de SAINT
5 BLANQUAT^a, Abdeltif LAHFID ^b, Stéphanie DUCHENE ^a

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8 ^a Géosciences Environnement Toulouse, Université de Toulouse, CNRS, IRD, UPS, CNES, F-
9 31400, France

10 ^b BRGM, 3 avenue Claude Guillemin, F-45060 Orléans, France

11

12 * corresponding author: bryan.cochelin.get@gmail.com

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18 **Abstract**

19 Estimating structural inheritance in orogens is critical to understand the manner in which
20 plate convergence is accommodated. The Pyrenean belt, which developed in Late Cretaceous to
21 Paleogene times, was affected by Cretaceous rifting and Variscan orogeny. Here we combine a
22 structural and petrological study of the Axial Zone in the Central Pyrenees to discuss structural
23 inheritance. Low-grade Paleozoic metasedimentary rocks were affected by a Variscan
24 transpressional event that produced successively: (i) regional-scale folds; (ii) isoclinal folding,
25 steep pervasive cleavage and vertical stretching, synchronous with peak metamorphism; (iii) and
26 strain localization into ductile reverse shear zones. The persistence of a relative flat envelope for
27 the Paleozoic sedimentary pile and Variscan isograds, and the absence of Alpine crustal-scale
28 faults in the core of the Axial Zone, suggest that the Axial Zone constitutes a large Variscan
29 structural unit preserved during Pyrenean orogeny. This configuration seems to be inherited from
30 Cretaceous rifting, which led to the individualization of a large continental block (future Axial
31 Zone) against a hyper-extended domain along the North Pyrenean Fault zone. This study places
32 the currently prevailing model of Pyrenean belt deformation in a new perspective and bears
33 important implications for crustal evolution and inheritance in mountain belts more generally.

34
35 **Keywords:** Structural inheritance, Pyrenean belt, Variscan orogeny, Passive margin inversion,
36 Central Pyrenees, RSCM thermometry

37 **Supplementary materials:**

38 - Table with RSCM data. - Figure illustrating peak-fitting of the Raman spectrum of carbonaceous
39 material and Raman spectra from the various samples of the Pallaresa cross-section

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42 The crustal architecture of a mountain belt narrowly depends on the rheology and relative
43 thickness of the continental plates involved in the collision. If the two plates were passive
44 margins, both were affected by rifting before convergence, leading to variably thinned domains
45 of continental crust, dismembered by normal faults or detachments. At the initiation of
46 convergence, the thickness and structure of the two margins exert first-order controls on the
47 architecture of the mountain belt (Mouthereau & Lacombe, 2006; Wrobel-Daveau et al., 2010;
48 Mohn et al., 2012; Mouthereau et al., 2012; McIntosh et al., 2013; Masini et al., 2014; Bellahsen
49 et al., 2014). The thinnest and thermally weaker continental margin is preferentially affected by
50 shortening during inversion of the rift systems and favourably orientated structures, which are
51 inherited from rifting, could be reactivated (Beaumont et al., 2000; Masini et al., 2011;
52 Mouthereau et al., 2012, 2013; Mesalles et al., 2014; Vacherat et al., 2014). At the paroxysm of
53 collision, when the two margins are accreted into the orogenic prism, new thrusts (unrelated to
54 the rift) are generated to balance shortening, and pre-rift structures within the basement may
55 undergo reactivation (Cloke et al., 1997; Allen et al., 1998; Soulaïmani & Burkhard, 2008).
56 However, the importance of these long-term inherited structures is most of the time overlooked
57 because of the polyphase tectonic history or because of the absence of structural markers to
58 constrain it.

59 The importance of rift-related structures has been previously illustrated in the Pyrenees
60 (Choukroune, 1992), where the European and Iberian plates began to collide in Late Cretaceous
61 times (Dubois & Seguin, 1978; Roest & Srivastava, 1991; Olivet, 1996). Early Cretaceous rifting
62 led to the formation of an ultra-thinned northern margin and to the exhumation of subcontinental
63 mantle (Lagabrielle and Bodinier, 2008; Lagabrielle et al., 2010; Masini et al., 2014; Clerc &
64 Lagabrielle, 2014; de Saint Blanquat et al., 2016; Vacherat et al., 2016). While the European
65 margin was hyper-extended in Cretaceous times, no evidence of this type of extension has been
66 detected in the Iberian plate, i.e. in the hinterland and in the southern foreland of the Pyrenean

67 belt. This hinterland, which corresponds to the so-called Axial Zone, is mainly made of Paleozoic
68 metasediments and magmatic bodies involved in the Variscan orogeny (Fig. 1a & 1b). Deep
69 exhumation of the Axial Zone is responsible for the erosion of post-Variscan deposits, which
70 form the main structural marker of the Cretaceous rifting event and Pyrenean collision. Since the
71 ECORS deep seismic profile (Roure et al., 1989; Choukroune, 1989; Choukroune et al., 1990),
72 the Pyrenees were interpreted as an asymmetric, doubly vergent collisional wedge with
73 numerous stacked crustal units in the Axial Zone (Fig. 1c). Whereas this geometry is in
74 agreement with structural observations made in the Central to Western Pyrenees, where the
75 crustal-scale north-dipping Gavarnie and Les Eaux Chaudes/Lakhora thrusts are observed (Fig.
76 1b), the continuity of these structures in the Central and Eastern Pyrenees has remained debated
77 (Carreras and Debat, 1996; Soler et al. 1998; Laumonier, 2015). Moreover, the anticlinal nappe-
78 stack model developed by Muñoz (1992) implies that the Variscan crust of the Central Pyrenees
79 was deeply affected by both Cretaceous rifting and Alpine orogeny, implying large-scale bloc
80 rotations, internal deformation and thrusting (Berastegui et al., 1993; Beaumont et al., 2000;
81 Mouthereau et al., 2014). Nevertheless, detailed structural studies performed at local scale on the
82 Variscan crust in the last 20 years tend to indicate only limited reactivation of Variscan structures
83 (Bons, 1988; Carreras & Debat, 1996; Capella and Carreras, 1996; Soler et al., 1998; Mezger &
84 Passchier, 2003; Denèle et al. 2008; Clariana & García-Sansegundo, 2009; Laumonier, 2015).

85 In order to resolve this apparent paradox, we try here to better constrain the deformation
86 of the Central Pyrenees through time. We combine a structural, petrological and
87 geothermometric study of the Variscan crust along the ECORS seismic profile. This study allows
88 us to discuss the structure of the hinterland of the central Pyrenean belt and the relative impact
89 of structures inherited from the Lower Cretaceous rifting event and from the Variscan orogeny.
90 We also propose a new model of the eastern Pyrenean crustal wedge in which a pop-up structure
91 corresponding to the Axial Zone was thrust over the southern foreland along moderately

92 dipping ramps. This constitutes an alternative view by comparison to the currently prevailing
93 model (Muñoz et al., 1992), which advocates several steeply-dipping thrusts cross-cutting the
94 Axial Zone.

95 **Geological setting**

96 The Axial Zone of the Pyrenean range is classically described as being affected by thick-
97 skinned tectonics involving several basement thrust sheets (Fig. 1c), namely the Lakhora,
98 Gavarnie and Bielsa units in the western part (Teixell et al., 1998), and the Noguères, Orri, and
99 Rialp units in the east (e.g. Vergés et al., 1995). The crustal-scale, north-dipping Gavarnie and
100 Les Eaux Chaudes/Lakhora thrusts that placed basement rocks over Cretaceous cover constitute
101 a major feature of the Pyrenean range and have been studied in detail. In the field, these two
102 thrusts form hundred metre-thick high-strain zones, mainly consisting of brittle–ductile
103 mylonites and breccia (Déramond, 1979; Majesté-Menjoulas, 1979). The Noguères unit (Fig. 1c)
104 is defined in the northern part of the Axial Zone (Muñoz, 1992) as the rooting zone of the so-
105 called "têtes plongeantes" or "plunging noses" defined to the south of the Axial Zone, that
106 consists of overturned synforms of Devonian to Permo-Triassic series (Vergely, 1970; Séguret,
107 1972; Choukroune & Séguret, 1973). This basement sheet is interpreted as the lateral continuity
108 of the Gavarnie unit separated from the Orri unit to the south by the Gavarnie thrust (Beaumont
109 et al., 2000; Mouthereau et al., 2014). While this thrust is clearly identified in the western Axial
110 Zone (Majesté-Menjoulas, 1979), its lateral continuity in the Central Pyrenees is discussed, as
111 this zone is devoid of post-Variscan deposits to identified Pyrenean thrusting, except one outcrop
112 of Triassic deposits, deformed in a fault zone located north on the Maladeta massif (Fig. 2 & 3,
113 see Soler et al., 1998).

114 The Axial Zone consists of Precambrian and Paleozoic rocks that were affected by the
115 Variscan orogeny between 330 Ma and 290 Ma (Denèle et al., 2014 and references therein). The

116 Variscan orogeny is marked by Permo-Carboniferous deformation, high temperature–low
117 pressure (HT–LP) regional metamorphism, and calc-alkaline plutonism. The Variscan crust of
118 the Axial Zone is classically interpreted as being formed by two contrasting structural levels (de
119 Sitter & Zwart, 1962; Carreras & Capella, 1994): (i) an upper crustal level, also called
120 “Superstructure”, made of low-grade metamorphosed Paleozoic sediments and defined by tight
121 to upright folds and steep axial-plane cleavage; (ii) a lower crustal level, or “Infrastructure”,
122 mostly characterized by Precambrian to Lower Paleozoic rocks affected by HT–LP
123 metamorphism and exposed into domal structures with shallow-dipping foliation planes. Recent
124 studies suggest that the formation of these two domains was coeval and occurred in a dextral
125 transpressive regime (Gleizes et al., 1998b; Mezger, 2009; Denèle et al., 2014).

126 In the Central Pyrenees, rocks affected by the Variscan orogeny are made of Cambrian
127 to Carboniferous metasedimentary rocks (Fig. 2). Cambrian to Ordovician rocks occupy the
128 central part of two regional-scale antiforms, the Pallaresa and Orri anticlines (Fig. 2). Whereas
129 the stratigraphy of Cambrian to Lower Ordovician rocks remains poorly constrained because of
130 its azoic character, the stratigraphic continuity between Upper Ordovician conglomerates and
131 lower series in the southern flank of the Pallaresa anticline seems to confirm that the older rocks
132 occupy the core of the anticline (Zandvliet, 1960; Hartevelt, 1970; Laumonier et al., 1996). The
133 lower terms are made of sandstone, locally microconglomerate and limestone at the top,
134 described as belonging to the Cambrian Evol Formation (Laumonier et al., 1996). The upper
135 series, belonging to the Cambrian Jujols Formation (Laumonier et al., 1996), is mainly made of
136 sandstone and greenish to dark schist. The upper Paleozoic series outcrops occur in the Couflens
137 syncline, bordering the North Pyrenean Fault, and in the Llavorsi syncline, pinched between the
138 Pallaresa and Orri anticlines (Fig. 2). These Upper Paleozoic rocks are made of (i) thin levels of
139 Silurian black shale and limestone; (ii) a succession of Devonian sandstone, limestone and schist

140 with significant local variations of facies (see García-Sansegundo et al., 2011 and references
141 therein); and (iii) Middle Carboniferous flysch.

142 The Cambrian to Paleozoic metasedimentary rocks of the Central Pyrenees are affected
143 by multi-scale folds and pervasive axial-plane cleavage, developed under low-grade
144 metamorphic conditions (Zwart, 1979; Bons, 1988; Capella and Carreras, 1996). The Paleozoic
145 metasedimentary rocks of the Superstructure are intruded by voluminous calc-alkaline plutons
146 such as the Maladeta pluton in the west, the Bassiès, Marimanha and Ribérot plutons in the north,
147 with U/Pb ages ranging between 312 and 298 Ma (Evans, 1993; Paquette et al., 1997; Denèle et
148 al., 2014). The intrusion of these plutons is considered as coeval with the late Carboniferous
149 dextral transpressional event (Bouchez & Gleizes, 1995; Evans et al., 1997; Gleizes et al.,
150 1998a). The Infrastructure marked by HT–LP metamorphic rocks and flat-lying foliations can be
151 observed in the Aston gneiss dome directly to the east of the Pallaresa anticline (Fig. 2) (Mezger,
152 2009; Denèle et al., 2009).

153 The central part of the Axial Zone is cut by several shear zones and faults generally
154 considered as Variscan in age, with reactivation during the Pyrenean collision. However, no real
155 consensus exists about the age and relative importance of activation and reactivation of these
156 various faults (Figs. 2 & 3). The older one is the Port de Salau fault, an east–west steeply north-
157 dipping fault. In its current position, its footwall consists of Cambrian rocks belonging to the
158 Pallaresa anticline, and its hangingwall of Siluro-Devonian rocks (Fig. 2). This fault seems to be
159 folded by the late Variscan transpressional event, and is interpreted by some authors as an early
160 Variscan thrust with a probable southward displacement (Bodin and Ledru, 1986; Losantos et
161 al., 1986). Farther south the Lladorre shear zone, localized in Cambrian limestones, is considered
162 by Capella & Carreras (1996) as a late reverse Variscan shear zone. This shear zone constitutes
163 the western continuity of the Variscan kilometre-scale Mérens shear zone (Fig. 2), which was
164 reactivated during the Alpine orogeny as a localized fault (McCaig, 1986; Denèle et al., 2008;

165 Mezger et al., 2012). Other authors (Berastegui et al., 1993; Beaumont et al., 2000; Vissers &
166 Meijer, 2012) consider this shear zone as an Alpine thrust that could be a hypothetical candidate
167 for the lateral continuity of the Gavarnie thrust, and renamed it the Pallaresa thrust. It is supposed
168 to be responsible for the stacking of the Nogueres sheet over the Orri sheet, with a proposed
169 displacement around 15–20 km. Farther south, the Estarón thrust was considered by Casas et al.,
170 (1989) to represent a Variscan thrust, while other authors consider it as another candidate for the
171 continuity to the east of the main Alpine Gavarnie thrust (Metcalf et al., 2009; Mouthereau et al.,
172 2014). Similarly, the Llavorsi thrust is considered to be a Variscan thrust in some studies (Casas
173 et al., 1989; Clariana, 2001), but an Alpine thrust belonging to the Orri sheet in others studies
174 (Zwart, 1986; Vergés et al., 1995, 2002). Its subtractive character, placing Devonian over
175 Cambro-Ordovician rocks, has led some authors to consider it as a normal fault, either Early
176 Carboniferous (Capellà & Bou, 1997), or Late Carboniferous to Cretaceous in age (Casas et al.,
177 2007). Further south, the Orri and Nogueres thrusts are indisputably Alpine in age, putting
178 Paleozoic rocks on Permian to Triassic sediments (Fig. 2). The Orri thrust accommodated 10–15
179 km of displacement during the collision (Séguret, 1972; Berastegui et al., 1993; Beaumont et al.,
180 2000; Mouthereau et al., 2014) and is supposed to be a reactivated Cretaceous normal fault
181 (Berastegui et al., 1993; Beaumont et al., 2000).

182 Discussion about the age of shear zones and faults in the Axial Zone of the Pyrenees is
183 due to the scarcity of post-Variscan markers. Syn-kinematic mica in some ductile shear zones
184 has been tentatively dated using $^{40}\text{Ar}/^{39}\text{Ar}$ methods and provided ages ranging between the
185 Eocene and the Jurassic (e.g. McCaig & Miller, 1986; Monié et al., 1994; Wayne & McCaig,
186 1998; Vissers et al., 2017). Eocene ages suggest a partial Pyrenean reactivation of some shear
187 zones, as for instance in the Néouvielle massif (Wayne & McCaig, 1998). Nevertheless,
188 $^{40}\text{Ar}/^{39}\text{Ar}$ ages on mica cannot be readily used as proxies for the absolute age of faults or shear
189 zones. The closure temperature of the Ar/Ar system in mica is around 300 °C (Harrison et al.,

190 1985), but the Jurassic to Cretaceous ages in Pyrenean shear zones suggest a partial to total
191 rejuvenation of argon isotopic systems during the widespread hydrothermal events ($250 < T <$
192 550 °C) that occurred between Permian and Cretaceous times (Boulvais et al., 2007; Poujol et
193 al., 2010; Fallourd et al., 2014; Boutin et al., 2016, Boutin, 2016). The difficulties linked to
194 interpreting $^{40}\text{Ar}/^{39}\text{Ar}$ ages in the Pyrenees have been emphasized by Jolivet et al. (2007), Maurel
195 et al. (2008) and Metcalf et al. (2009), who recognized the effects of the Mesozoic hydrothermal
196 events in the syn-kinematic micas of the Variscan Maladeta and Néouvielle plutons, and in the
197 Canigou massif.

198 **Methodology**

199 *Measurement and sampling*

200 This study is based on 330 sites distributed along N–S cross-sections, following the main
201 valleys of the mountains belt (Fig. 2). In order to reconstruct the large-scale geometry of folds,
202 planar and linear strain fabrics and kinematic criteria affecting metasedimentary rocks were
203 systematically measured, as well as the relationship of planar fabric with original bedding. All
204 these measurements are synthesized in Figs. 3 and 4.

205 Nineteen samples were collected along the Pallaresa cross-section. They were selected in
206 the whole metasedimentary sequence from Cambrian to Devonian (Fig. 2, red dots). At least two
207 samples were collected in each structural domain of the cross-section. Dark-coloured samples rich
208 in organic matter were favoured for Raman Spectroscopy of Carbonaceous Materials (RSCM).

209 Polished thin sections were prepared normal to the schistosity and parallel to the lineation
210 (XZ planes) for petrographic and microstructural observation.

211 *Raman Spectroscopy of Carbonaceous Materials (RSCM)*

212 Raman Spectroscopy of Carbonaceous Material (RSCM) method is used to estimate the
213 maximum temperature (up to 650 °C) recorded by the metasediments during metamorphism
214 (Beyssac et al., 2002; Lahfid et al., 2010). Delchini et al. (2016) confirmed the applicability of
215 RSCM for studying domains with polyphased metamorphic histories, such as the Pyrenees. Raman
216 analyses were performed at the BRGM, Orléans, using a Ranishaw inVia Reflex system with
217 argon-ion laser source excitation of 514.5 nm. The laser beam is focussed on the sample through
218 a Leica DM2500 microscope specially adapted for the system using a x100 lens, with power of
219 around 0.5 mW at the surface of the thin section. Before each measurement session, the
220 spectrometer was calibrated using the 520.5 cm⁻¹ line of a silicon standard. The signal obtained
221 after elimination of Rayleigh diffusion using Edge filters was dispersed using 1800 lines.mm⁻¹
222 grating before being analyzed by a deep depletion CCD detector (1024 x 256 pixels). About 1520
223 Raman spectra of particles were recorded to check data consistency. Raman parameters, peak
224 temperatures obtained by RSCM and Raman spectra are provided in the Data Repository.

225

226 **Structures and kinematics**

227 In the study area we observed two types of schistosity. The first generation is penetrative,
228 observed in all Variscan rocks, and called hereafter S1. The second generation is only locally
229 observed and corresponds to a crenulation cleavage. To the south of the Axial Zone, the Cambrian
230 series belonging to the Orri anticline are unconformably overlapped by undeformed Permian and
231 Triassic red beds that crosscut the S1 cleavage (Fig. 5a). These observations attest, in accordance
232 with previous studies (e.g. Zandvliet, 1960; Mey, 1968; Lucas, 1985; Zwart, 1986; Bichot, 1986;
233 Carreras & Debat, 1996) that the regional penetrative cleavage S1 is Variscan in age. By contrast,
234 a discrete crenulation cleavage is observed in Permian and Triassic pelitic levels suggesting that

235 the second schistosity in the basement units could be Alpine in age (Izquierdo-Llavall et al., 2013).
236 It is hereafter named Sa.

237

238 As a first approximation, the Central Pyrenees area is defined by two large and open
239 antiforms forming the Pallaresa and Orri anticlines and the southward verging and pinched
240 Llavorsi syncline (Fig. 3a). The flat attitude of bedding in anticline cores is perturbed by
241 multiscale open to isoclinal folds (Fig. 4) associated with pervasive axial-plane cleavage (i.e. S1;
242 Fig. 5b & b'). S1 shows a homogeneous N95–110E direction (Fig. 3b) with trajectories roughly
243 parallel to the Lladorre–Mérens, Estarón and Llavorsi mylonitic shear zones (Fig. 3a). In the
244 northern half of the Axial Zone, S1 is vertical and dips around 50–60°N in the southern part,
245 excepted 5 km to the north to the Nogueres Unit, where it is subhorizontal or displays shallow
246 dips to the north (Figs 3a & 4). S1 bears steeply-plunging stretching lineations with a
247 homogeneous N150–160° orientation (Figs 3a & 3c).

248 Structural studies have highlighted three domains of deformation limited by the
249 Lladorre–Mérens and Llavorsi mylonitic shear zones. The northern domain is characterized by
250 a vertical cleavage S1 and an apparent heterogeneous deformation, with highly transposed
251 Devonian to Cambrian limestone and schist and gently folded thick Cambrian
252 microconglomerate and sandstone. This domain is affected by an apparent coaxial deformation
253 attested by the coexistence of “north-side up” and “south-side up” shear criteria on subvertical
254 shear planes. This domain is limited to the south by the 200-m-thick mylonitic shear zone of
255 Lladorre, localized in interbedded sandstones and marbles showing non-coaxial “north-side up”
256 kinematics (Figs 3a, 4 & 5c). The central domain is mostly characterized by the complete
257 transposition of original bedding and associated non-coaxial “top-to-the-south” shear bands (Fig.
258 5d). This high strain domain corresponds to the southern overturned limb of the Pallaresa
259 anticline and to the pinched Llavorsi syncline, with apparent strain localization in the highly

260 laminated Silurian black shales. The southern domain, which corresponds to the Orri anticline,
261 is poorly deformed, marked by discrete S1 cleavage oblique to the original bedding. Few discrete
262 C' shears with non-coaxial "top-to-the-south" kinematics were observed, as previously described
263 by Mey (1968).

264 The S1 trajectories are locally disturbed in the contact aureole of plutons where
265 asymmetric schistosity triple points can be defined (Fig. 3a). Stretching lineations are
266 subhorizontal in these domains, with C' shear bands showing dextral senses of movement (Fig.
267 6a). Shear bands strike between N100–N130°E and concentrate in narrow mylonitic corridors
268 such as the western part of the Estarón thrust, between the Maladeta and Marimanha plutons (Fig.
269 3a), or around the Bassiès pluton (Fig. 4).

270 Over the entire area, the Paleozoic basement is locally affected by late Alpine deformation,
271 mainly localized within 1- to 100-m-thick corridors (Fig. 4). In such corridors, new 0.1- to 1-m-
272 scale folds (Fa), with N80°- to N140°-striking fold axes, affect previously schistose rocks forming
273 kink-bands and chevron folds (Figs. 3b & 3c and 6b & 6c). Crenulation cleavage Sa is common
274 in such corridors. This crenulation cleavage, axial plane of late folds, is sub-vertical in the northern
275 third part of the Axial Zone, striking at a low angle to the regional schistosity S1. In the central
276 part of the Axial Zone, Sa dips steeply to the south (Fig. 6b) and can be easily distinguished from
277 S1, which dips northward (Fig. 4, between Lladorre and Llavorsi). In the southern domain, which
278 corresponds to the Orri anticline, the late crenulation cleavage dips at low angles to the north but
279 seems to be less conspicuous than elsewhere (Fig. 4). Away from these corridors, only a
280 subhorizontal east–west discrete crenulation lineation affects the regional structures. Evidence of
281 brittle deformation is otherwise scarce in the Pallaresa cross-section. Some fault zones have been
282 observed in the Silurian black shales of the pinched Llavorsi syncline. These faults are associated
283 with reverse drags that affect the regional penetrative structures over distances of ~20 m.

284 **Petrological study**

285

Raman Spectroscopy of Carbonaceous Materials (RSCM)

286 The Raman spectroscopy results are synthesized in Fig. 7. Palaeotemperatures range
287 between 350 °C and 550 °C with a temperature around 350 °C in the Couflens syncline, the
288 Llavorsi syncline and the Orri anticline and an increase up to 550 °C in the Pallaresa anticline.
289 Two significant steps of around 50 °C and 100 °C were observed in the vicinity of the Estarón
290 thrust and the Lladorre shear zone, respectively (Fig. 7a). Isotherms were deduced from the Raman
291 temperature (Fig. 7a). Small steps of maximum 30 °C were evidenced in the Couflens syncline,
292 but these variations cannot be considered as significant in regard of the analytical error on the
293 measurements.

294

Metamorphic parageneses and microstructures

295 The optical analysis of the thin sections allowed us to determine the metamorphic
296 parageneses and microstructures and to gain insights into the thermal conditions of deformation in
297 addition to RSCM results. The Devonian sequence in the north (16BL44, 16BL47, 16BL51,
298 16BL55, 16BL59), close to the NPF, corresponds to limestone with pelitic intercalations. Silurian
299 deposits that belong to the Couflens syncline are dark schists rich in carbonaceous materials
300 (16BL60, 16BL52, 16BL53), with few small chlorite crystals. The northernmost Cambrian samples
301 consist of an alternation of thin pelitic and carbonate levels (16BL56) and quartzo-pelitic schist
302 (TP442). The following two samples of the Pallaresa anticline (15BL104, 15BL106) correspond
303 to quartz-rich schist with Ms-Bt-Pl-Qtz (see Kretz, 1983 for the mineral abbreviations). Biotite
304 porphyroblasts have grown parallel to S1. Quartz-rich levels show well-developed dynamic
305 recrystallization controlled by grain boundary migrations, attested by a pinning effect on the
306 migrating boundaries (Fig. 8a, Song and Ree, 2007). These types of microstructure and the
307 observation of secondary biotites in schistosity planes suggest a temperature close to 500 °C during
308 deformation (Hirth and Tullis, 1992; Stipp et al., 2002). Such a temperature is consistent with the

309 apparition of biotite and the absence of andalusite, cordierite and staurolite, which appear above a
310 temperature of 550–600 °C in the pseudosections obtained for similar protoliths from the western
311 Aston dome termination (Mezger & Régnier, 2016) and the Albères massif (Vilà et al., 2007).
312 Within the Lladorre shear zone (samples 15BL110 & 15BL112), well-preserved quartz veins in
313 mylonitic marble show dynamic recrystallization of quartz with an association of sub-grain
314 rotation and grain boundary migration. The transition between these two mechanisms suggests a
315 temperature of deformation at about 500–550 °C (Stipp et al., 2002). A sheared conglomerate that
316 belongs to the upper Ordovician of the southern overturned limb of the Pallaresa antiform and
317 located within the Estarón shear zone (sample 15BL118, Fig. 5d) is composed of Ms-Pl-Qtz±Chl
318 matrix and clasts of quartz. Here, S1 and shear bands are underlined by muscovite (Fig. 8b). Quartz
319 grains show evidence of recrystallization by bulging and sub-grain rotation (Fig 8c & d).
320 Conglomerates show a core and mantle microstructure with grain-size of quartz blasts of about 20
321 to 50 µm, suggesting a recrystallization temperature of about 400 °C (Stipp et al., 2002). Finally,
322 the two southernmost samples (15BL125, 15BL127) are Cambrian schists from the Orri anticline,
323 with muscovite and chlorite growing parallel to S1 (Fig. 8e).

324 The evolution from biotite-bearing schist in the Pallaresa anticline to chlorite-bearing schist
325 in Orri anticline, in addition to the quartz recrystallization dynamics, indicate a decrease in
326 palaeotemperature towards the south, which is in good agreement with RSCM results. All along
327 the cross-section, the second cleavage Sa is characterized by microfolds, forming microlithons that
328 affect S1 (Fig. 8e). The development of Sa implies the progressive destruction of previous fabrics
329 by pressure solution, mainly removing quartz grains while phyllosilicates show internal
330 deformation like folding and rotation. In agreement with previous work (Zwart, 1986; Bons, 1988;
331 Soler et al., 1998; Clariana et al., 2008, among others), our observations confirm that Alpine
332 structures in Central Pyrenees were produced under low temperatures, which never reached those
333 obtained during the Variscan orogeny

334

Relationship between deformation and metamorphism

335 Given that greenschist and amphibolite facies metamorphic minerals grew within shear
336 planes and underline the main schistosity plane S1, our petrological study highlights that HT–LP
337 metamorphism was synchronous with pervasive deformation. We determined the geometry of the
338 palaeo-isotherms based on the following constraints. The isotherms match the temperatures
339 retrieved from Raman spectroscopy (Fig. 7a). They also account for the absence of abrupt
340 petrological transitions across the section, with regular transitions from greenschist to amphibolite
341 facies assemblage. The isotherms are drawn parallel to each other. They match the geothermal
342 gradient previously estimated in the envelope of the Aston, Albères, Bossost and Canigou gneiss
343 domes (40 to 70 °C/km) (Mezger, 2005; Triboulet et al., 2005; Vilà et al., 2007; Mezger &
344 Régnier, 2016). We end up with a picture of relatively flat isotherms that define a
345 palaeogeothermal gradient of 45 °C/km (Fig. 7b).

346 On the cross-section in Fig. 7b, isotherms are secant on large-scale folds, such as the
347 Llavorsi syncline. This first-order observation allows us to interpret the metamorphism as
348 subsequent to the initiation of regional-scale folding. Besides, isotherms are vertically shifted by
349 the Lladore and Estarón amphibolite to greenschist facies reverse shear zones, suggesting that
350 shear zone activity ends after peak metamorphism. Our results show (i) that shear zones and
351 regional cleavage form a single and continuous fabric that simply illustrates the strain gradient in
352 the crust; (ii) that mylonites within the shear zones and rocks affected by the regional cleavage
353 exhibit the same amphibolite to greenschist metamorphic parageneses; and (iii) that the regional
354 fabrics and shear zone share the same structural character, with down-dip stretching lineations and
355 dip-slip/reverse sense of shear. Based on this evidence, we consider these shear zones as Variscan
356 in age. They formed during the same tectonic episode as the other regional structures. Because the
357 Lladorre and Estarón amphibolite to greenschist facies reverse shear zones shift the isotherms (Fig.
358 7b), we propose that their final activity occurred after peak metamorphism and represents the final

359 expression of the main Variscan deformation recorded in Pyrenees (see discussion below). This
360 interpretation as such does not refute the notion of a possible reactivation of the shear zones during
361 the Pyrenean orogeny, which was previously proposed by Mc Caig (1986) on the basis of Ar/Ar
362 dating.

363 **Variscan tectono-metamorphic evolution of the Central Pyrenees**

364 Based on new structural and petrological observations, we propose hereafter a scenario for
365 the Variscan tectono-metamorphic evolution of the Central Axial Zone of the Pyrenees.

366 During the first stage of Variscan evolution, the upper crust was affected by regional-scale folding,
367 producing kilometre-scale open to tight southward verging folds with an east–west axial plane
368 (Fig. 9a). We consider them as the first expression of Variscan transpression in the Axial Zone.

369 At this time, the mid to lower crust (Infrastructure) was probably affected by upward-propagating
370 HT–LP metamorphism (Fig. 9a). Note that faults recognized along the Pallaresa cross-section,
371 with a subtractive character and apparent “north side down” movement (Port de Salau and
372 Llavorsi faults), are folded and must have formed prior to this stage. Because of penetrative
373 deformation that occurred during the next stage and of the difficulty to evaluate the amount of
374 thickening, the thickness of each sedimentary unit in our reconstruction (Fig. 9a) remains
375 speculative. Moreover, we consider the existence of a large syn-orogenic Carboniferous basin,
376 similar to the Arreau bassin located a few kilometres to the west of the Palaressa cross-section
377 (Delvolvé, 1987).

378 During the second stage of Variscan evolution, the upper crust was affected by intense
379 horizontal shortening and vertical stretching producing multi-scale isoclinal folding and the
380 formation of pervasive cleavage (Fig. 9b). In the central and northern part of the Axial Zone, the
381 upper crust was affected by apparently coaxial shearing, while the southern part was the focus of
382 non-coaxial “top-to-the-south” reverse kinematics, illustrating strain partitioning within the upper

383 crust. Non-coaxial shearing is responsible for the amplification of the overturned Pallaresa
384 anticline (Fig. 9b). This event occurred at peak metamorphism, leading to the crystallization of
385 biotite, muscovite and chlorite in cleavage planes.

386 During a third stage, strain localization occurred with formation of 100-m-thick shear zones
387 under amphibolite–greenschist conditions (Fig. 9c). The Lladorre and Estarón reverse shear zones
388 were responsible for vertical offsets of the isotherms by ~ 2 km and 1 km, respectively. Similar
389 displacement can be deduced from the offset of the sedimentary pile for the Lladorre shear zone
390 (Fig. 4). At this stage, the dome shape of the Infrastructure was amplified by the displacement
391 along shear zones (Fig. 9c). This interpretation is in agreement with structural observations made
392 in the rest of the Axial Zone, where it has been shown that domes were lately amplified and
393 affected by steep transpressional shear zones, showing retrogressive parageneses (Denèle, 2007;
394 Vilà et al., 2007; Denèle et al., 2008, 2009).

395 **Evolution of the Central Pyrenees during Cretaceous rifting and Pyrenean collision**

396 *Cretaceous rifting*

397 The Cretaceous rifting event has strongly reworked the North Pyrenean Zone, which
398 belongs to the European plate. Markers of this event correspond to: (i) late Aptian to early
399 Cenomanian rift basins filled by flysch-type sediments (Debroas, 1987, 1990; Bodego et al., 2015;
400 Chelalou et al., 2016); (ii) regional HT–LP metamorphism with temperatures locally reaching 600
401 °C (Vielzeuf & Kornprobst, 1984; Dauteuil & Ricou, 1989; Golberg & Leyreloup, 1990; Clerc &
402 Lagabrielle, 2014; Vacherat et al., 2014) and widespread metasomatism (Schärer et al., 1999;
403 Boulvais et al., 2007; Pujol et al., 2010; Fallourd et al., 2014; Boutin et al., 2016); (iii) exhumation
404 of sub-continental mantle bodies (Fabries et al., 1991; Clerc et al., 2012; de Saint Blanquat et al.,
405 2016).

406 By contrast, these markers of Cretaceous rifting are absent from the Axial Zone. Wherever
407 present, the Mesozoic sediments unconformably deposited over the Variscan basement correspond
408 to post-rift Cenomano-Turonian flysch (Ternet et al., 1980, 2003; Clin et al., 1986; Mirouse et al.,
409 1993; Majeste-Menjoulas et al., 1999). In the Central Pyrenees, our structural and petrological
410 study coupled with RSCM shows that (i) all vertical offsets display reverse motion, and (ii) the
411 only thermal event recorded in the area is related to Variscan HT–LP metamorphism, with no sign
412 of any later thermal overprint. These first-order observations suggest that the Axial Zone was
413 relatively preserved during Cretaceous rifting. Evidence of Cretaceous rifting concern only the
414 borders of the Axial Zone and remains scarce. Indeed, in the study area, Triassic ophiolites are
415 described as intrusive into Paleozoic and Triassic sedimentary rocks in a fault zone near Couflens
416 (Fig. 2; Ternet et al., 1997). Proximity between this fault zone and evidence of metasomatism with
417 talc formation (Ternet et al., 1997) of Cretaceous age (Boutin et al., 2016) suggests fault
418 (re)activation during the rifting event (Fig. 9d). In the western part of the Axial Zone, normal
419 faulting is indicated by the remains of syn-rift deposits (Casteras et al., 1971; Ternet et al., 1980,
420 2003), which can be observed against the North-Pyrenean Fault. Along the North Pyrenean Fault
421 east of the study area, evidence of major Cretaceous metasomatism has been discovered, with
422 albitized rocks and talc occurrences (Fallourd et al., 2014). On the southern border of the Axial
423 Zone, the formation of the Organyà Basin, which is described as a piggy-back basin with syn-rift
424 Aptian to Albian deposits, is supposed to be related to a normal movement along the future Orri
425 thrust (Fig. 9d, García Senz, 2002; Martínez-Peña & Casas-Sainz, 2003; Mencos et al., 2015).

426 *Late Cretaceous to Cenozoic convergence*

427 In the southern part of the Axial Zone, alpine movement along the Orri thrust is undeniable,
428 highlighted by the Rialp window (Fig. 2 & 3), probably responsible for tilting of Variscan
429 structures in the Orri anticline (Bons, 1988; Carreras & Debat, 1996). At depth, the Rialp thrust is

430 inferred from the ECORS seismic profile but has no geological incidence at the surface. Along the
431 North Pyrenean Fault, Alpine deformation is marked by refolded Variscan structures and folds in
432 Permo-Triassic sediments (Dérámond, 1970; Ternet et al., 1997), as well as the reactivation of
433 inherited vertical faults marked by the uplift of the northern blocks over several hundred metres to
434 a few kilometres (Fig. 9d & e). Across the rest of the Axial Zone, there is no significant Alpine
435 deformation, except in narrow corridors highlighted in Fig. 4 where crenulation cleavage was
436 produced (Fig. 6b & 6c). So, except along the North Pyrenean Fault and few kilometres above the
437 Orri thrust, no significant Alpine fault or penetrative deformation are observed, leading to the
438 conclusion that Variscan structures appear continuous and mainly undisturbed in the central Axial
439 Zone (Fig. 4).

440 **Discussion**

441 The contrasted response to Early Cretaceous rifting of the European and Iberian margins,
442 which are represented respectively by the North-Pyrenean Zone and the Axial Zone, suggests an
443 asymmetric rift system (Fig. 10a). The European margin of this rift shows a large distal domain,
444 which comprises the inverted syn-rift basins of Camarade, Saint-Girons and Aulus (e.g.
445 Mouthereau et al., 2014). The quantity of extension increased southward in this distal domain and
446 has led to the exhumation close to the surface of mantle rocks in the Aulus Basin (Fig. 10a). These
447 observations show that this rift system was mature and led to the breakup of the continental crust.
448 By contrast, the Iberian margin shows a large proximal domain with the relatively preserved Axial
449 Zone block and a restricted distal domain, localized near the North-Pyrenean fault zone.
450 Asymmetrical conjugate margins have been evidenced in numerous purely extensional palaeo-rift
451 systems (Chian et al., 1995; Ranero & Pérez-Gussinyé, 2010; Blaich et al., 2011; Espurt et al.,
452 2012; Peron-Pinvidic et al., 2013; Sutra et al., 2013). However, a sinistral displacement has been
453 deduced from motion of the Iberian plate in Cretaceous times (Le Pichon et al., 1970; Choukroune

454 et al., 1973; Choukroune & Mattauer, 1978; Olivet, 1996), suggesting that the Pyrenean rift system
455 occurred under a transtensional stress regime. Furthermore, most of the syn-rift basins in the
456 North-Pyrenean Zone of the Central Pyrenees correspond to narrow pull-apart basins (Debroas,
457 1987). In transtensional settings, strain can be partitioned into purely extensional domains and
458 narrow strike-slip or transform faults (Allen et al., 1998; Wu et al., 2009), which could explain the
459 degree of asymmetry of the Pyrenean rift system. Here we advocate that the North-Pyrenean Fault
460 represents a transform fault that has accommodated the abrupt variation of extensional patterns
461 between the Iberian and European plates during Cretaceous rifting (Fig. 10a). Further west, in the
462 Basque massifs, the conjugate margins appear to be more symmetrical (Roca et al., 2011; Masini
463 et al., 2014; Tugend et al., 2014) and the the North-Pyrenean Fault has never been identified. This
464 feature highlights the segmentation of the Pyrenean rift system along the strike of the mountain
465 belt.

466 Concerning the inversion of the rift system, classical models of the Pyrenean orogenic
467 wedge have involved various basement thrust units in the Axial Zone and imply large-scale block
468 rotations and internal deformation of these units (e.g. Beaumont et al., 2000). These basement units
469 have been clearly identified in the western part of the Axial Zone, such as the Lakhora and
470 Gavarnie units (Teixell, 1996), in its southernmost part with the Orri and Rialp units in the Central
471 Pyrenees (e.g. Muñoz, 1992), and in the western Pyrenees with the Bielsa/Millares units (Teixell,
472 1996; Jolivet et al., 2007). Based on the interpretation of the ECORS seismic profile, Muñoz
473 (1992) had proposed that the Noguères sheet was rooted in the Axial Zone in the Central Pyrenees,
474 and was an eastward prolongation of the Gavarnie thrust. In this currently accepted model, the
475 Noguères sheet would have been displaced to the south over a distance of 15–20 km (Berastegui
476 et al., 1993; Beaumont et al., 2000). However, our results in the Axial Zone do not support such a
477 model. We show instead that the main localized structures of the Central Pyrenees are ductile
478 amphibolite to greenschist facies Variscan shear zones with only limited evidence of late

479 reactivation, and are associated with a motions that do not exceed 1–2 km (Figs. 4 & 7). Even if
480 the reactivation of these Variscan shear zones was responsible for these kilometre-scale offsets,
481 the evidence does not support the existence of a basement nappe rooted in the central part of the
482 Axial Zone. Furthermore, despite intense deformation of the Superstructure during the Variscan
483 orogeny, where upright isoclinal folds are common, the envelope of the stratigraphic markers
484 remained roughly horizontal, as was that of the Infrastructure's roof and its metamorphic isograds
485 (Figs. 4 & 7). Furthermore, thermochronological data in Central Pyrenees show that the Noguères
486 unit and the Orri unit (Vergés et al., 1995, Teixell et al., 1998) have shared similar exhumation
487 histories since late-Eocene times (Fig. 10c) (Fitzgerald et al., 1999; Sinclair et al., 2005).
488 Consequently, the Central Pyrenees form a single tectonic unit, delimited by the North-Pyrenean
489 Fault in the north and the Orri thrust in the south (Fig. 10b, Laumonier, 2015). In this scheme, the
490 "têtes plongeantes" or "Plunging noses" (Séguret, 1972; Choukroune and Séguret, 1973) defined
491 to the south of the Axial Zone should be reinterpreted as cover nappes originating in the Axial
492 Zone or in the North Pyrenean Zone, and transported to its southern border by successive thrusting
493 and décollements at the base of the Mesozoic cover and downward in the Silurian black shales.
494 Given that the Noguères allochthonous unit corresponds to the uppermost Paleozoic series and are
495 devoid of syn-rift sequences as well as HT–MP metamorphic rocks that characterized the North
496 Pyrenean massifs, we interpret this unit as belonging to the top of the Axial Zone. These cover
497 nappes, which share the same exhumation histories as the Maladeta and Marimanha massifs (Fig.
498 10c), were disconnected from the Axial Zone by the exhumation of the Orri units in Oligocene
499 times. If the amount of convergence of the Iberian plate, around 150 km since 83 Ma (Olivet, 1996;
500 Sibuet et al., 2004; Vissers & Meijer, 2012) is correct, the estimates and the spatial distribution of
501 shortening in the Pyrenean belt, especially within the Axial Zone, should be re-examined. Indeed,
502 the total amount of convergence has been explained by a combination of subduction and crustal
503 thickening (Beaumont et al., 2000). This later included 15–20 km of displacement of the Noguères

504 nappe when considered as rooted in the Axial Zone. Furthermore, 10–15 km of shortening by
505 pervasive internal deformation has been inferred in order to match the total convergence (Roure
506 et al., 1989; Muñoz, 1992; Mouthereau et al., 2014). Given our interpretation of structures in the
507 Central Pyrenees, these ca. 30 km of horizontal shortening by crustal thickening in the basement
508 must be found elsewhere, for example in a greater amount of subduction of the Iberian plate.

509 The relative stability of the Axial Zone block since Paleozoic times highlighted the critical
510 role of crustal-scale inherited structures from the Cretaceous rifting event on the geometry of the
511 Cenozoic orogenic wedge. Convergence was first accommodated between 85 and 70 Ma within
512 the thinned European margin, leading to the inversion of syn-rift basins and the accretion of
513 microcontinental blocks that form the North Pyrenean massifs (Mouthereau et al., 2014; Ford et
514 al., 2016; Vacherat et al., 2016) (Fig. 10b). Deformation then propagated southward during Eocene
515 times into the moderately thinned Iberian crust, but remained focused around the main normal
516 faults inherited from the Cretaceous. The reactivation of these normal faults as crustal-scale thrusts
517 seems to have been sufficient to initiate the subduction of a part of the Iberian plate under the
518 accreted European plate and the Axial Zone block, defining a pop-up structure (Fig. 10b; Brun,
519 2002). Given that the main shear zones of the Axial Zone remained preserved after Alpine
520 collision, the role of inherited structures from the pre-rift events appears to be negligible in this
521 scenario. The importance of the North-Pyrenean Fault, however, which formed a boundary
522 between two continental margins with contrasting behaviours during rifting, and later between two
523 contrasting pro- and retro-wedge domains during collision, is nonetheless also confirmed by our
524 study. Most previous reconstructions had likewise considered that this fault was inherited from the
525 Variscan orogenic cycle (Arthaud & Matte 1975; Burg, 1994; Carreras & Capella, 1994; Matte,
526 2001, Stampfli et al., 2013).

527 **Conclusion**

528 Our results provide a complete structural view of the architecture of the Axial Zone and its
529 relationship with regional metamorphism in the Central Pyrenees. The present-day structure of the
530 Central Pyrenees results from a multi-stage Variscan–Cretaceous–Alpine history. The central part
531 of the Axial Zone represents an upper crustal sequence deformed under a transpressional regime
532 in Variscan times, and is characterized by (i) regional to small-scale folds, (ii) steep penetrative
533 axial-plane cleavage, and (iii) greenschist to amphibolite facies metamorphism associated to a HT–
534 LP gradient. The flat pattern of Paleozoic stratigraphic and Variscan metamorphic markers all
535 along the north–south cross-section of the Central Pyrenees, along with the absence of Cretaceous
536 and Pyrenean crustal-scale faults in the core of the Axial Zone, highlight that this domain has been
537 preserved from both rift- and collision-related overprints. Thus, in Cretaceous times, the Axial
538 Zone formed a single continental bloc separated by the North Pyrenean Fault from the hyper-
539 extended domain that formed the northern margin. Finally, the “Têtes plongeantes” units located
540 south of the Axial Zone are re-interpreted here as cover nappes, detached from the central-southern
541 upper part of the Axial Zone by faulting between the basement and the Mesozoic cover during the
542 Cenozoic. In our model, the Axial Zone would thus correspond to a pop-up structure overthrusting
543 its southern foreland. It contradicts the widely accepted view of an antiformal nappe stack in the
544 core of the Axial Zone, previously elaborated by Muñoz (1992). Alpine shortening was restricted
545 to the northern and southern borders of the Axial Zone, and focused around structures inherited
546 from Cretaceous rifting.

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553

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1009

1010 **Figure captions:**

1011

1012 Figure 1: (a) Location of the Variscan crust of the Pyrenees in western Europe, with main
1013 Mesozoic basins. 1- Organyà basin, 2- Arzacq-Mauléon basin, 3- Parentis basin, 4- Basco-
1014 Cantabrian and le Danois basins, 5- Central Iberian basins (Tugend et al., 2014); (b) Geological
1015 sketch map of the Variscan crust of the Pyrenees. In blue: main Alpine faults; in black: main
1016 Variscan shear zones. Abbreviations: ELT: Les Eaux Chaudes/Lakhora Thrust; GT: Gavarnie
1017 Thrust; NPF: North Pyrenean Fault, LT: Llavorsi thrust, MSZ: Mérens Shear Zone). (c) Tectonic
1018 subdivision of the Axial Zone (Mouthereau et al. 2014).

1019

1020 Figure 2: Geological map of the Central Pyrenees, modified from Colchen et al. (1995) and Zwart
1021 et al. (1976) showing location of the Pallaresa cross-section presented in this work (Fig. 4). Studied
1022 outcrop and samples for RSCM measurements are also reported.

1023

1024 Figure 3: (a) Structural map of the Central Pyrenees. In grey: Upper Paleozoic rocks; in white:
1025 Cambrian to Ordovician rocks; in orange and pink: Variscan domes and granitoids, respectively.
1026 In blue: stretching lineations and kinematics from Evans et al. (1997). Data compiled for cleavage
1027 trajectories are from Mey (1968), Hartevelt (1970), Colchen et al. (1995), Evans et al. (1997),

1028 Mezger (2005) (Aston metamorphic dome), and from our observations (n=300). Foliation
1029 trajectories in plutons were deduced from AMS studies by Gleizes (1992), Evans et al. (1997) and
1030 Antolín-Tomás et al. (2009). **(b)** S1 (black) and Sa (blue) schistosity poles from our measurements.
1031 **(c)** L1 mineral-stretching lineations measured in the field in black, and Fa fold axes in blue.
1032 Stereograms: lower hemisphere, equal area projection. Contours at 2σ are calculated for black
1033 points (i.e. S1 and L1).

1034
1035 Figure 4: N–S geological cross-section of the Central Pyrenees. Only significant alpine corridors
1036 that have been observed or rigorously described in the literature are represented on this cross-
1037 section.

1038
1039 Figure 5: Field photographs illustrating the main structures described in the text. **(a)** Undeformed
1040 Lower to Middle Triassic sandstones and conglomerates; bedding is sub-horizontal and
1041 sedimentary figures attest normal polarity (Location: Port del Canto, Rubiò); Photograph **(b)** and
1042 interpretative sketch **(b')** of penetrative axial-plane cleavage in siltites and sandstones (Location:
1043 South of Alós d'Isil); **(c)** Asymmetrical boudinated sandstones in marbles of the Lladorre shear
1044 zone, showing north-side-up kinematics ; **(d)** C' shears with top-to-the-south kinematics in Upper
1045 Ordovician conglomerates (Location: Cardós valley).

1046
1047 Figure 6: **(a)** Asymmetrical boudinated quartz veins within Cambro-Ordovician sandstones showing
1048 dextral kinematics (Location: south of the Port de Marterat, western part of the country-rocks of
1049 the Bassiès pluton); **(b)** Apparent obliquity between the regional schistosity S0-1 and the late
1050 crenulation cleavage associated to kink-bands (Same location than Fig. 6a & 6b); **(c)** Longitudinal
1051 late folds and associated crenulation cleavage Sa steeply dipping to the south and affecting the
1052 regional schistosity S0-1 (Location: 400m north of Ainet de Cardós).

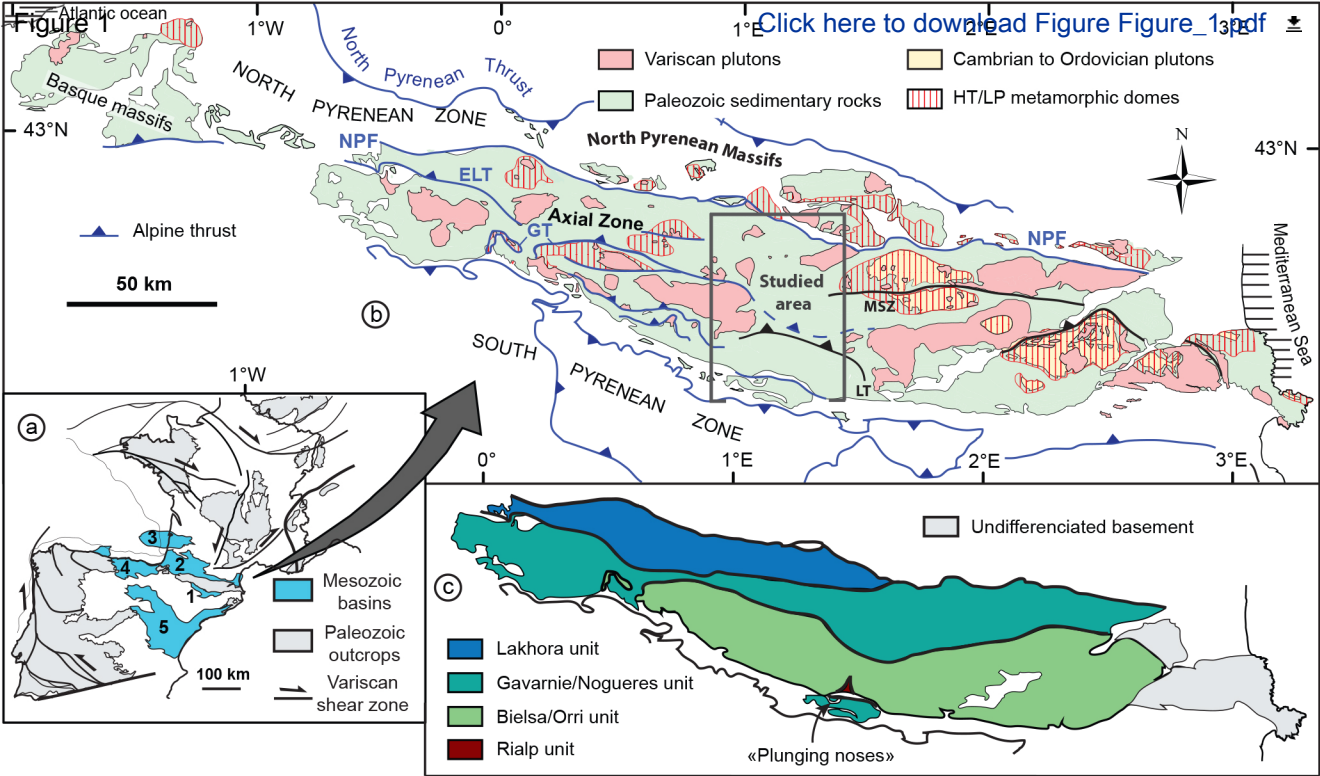
1053

1054 Figure 7: **(a)** Palaeotemperatures obtained by Raman Spectroscopy of Carbonaceous Material
1055 (RSCM). Error bars are standard deviation. X-axis corresponds to the latitude; **(b)** Interpreted
1056 position of isotherms in our cross-section deduced from RSCM data and petrological descriptions.
1057

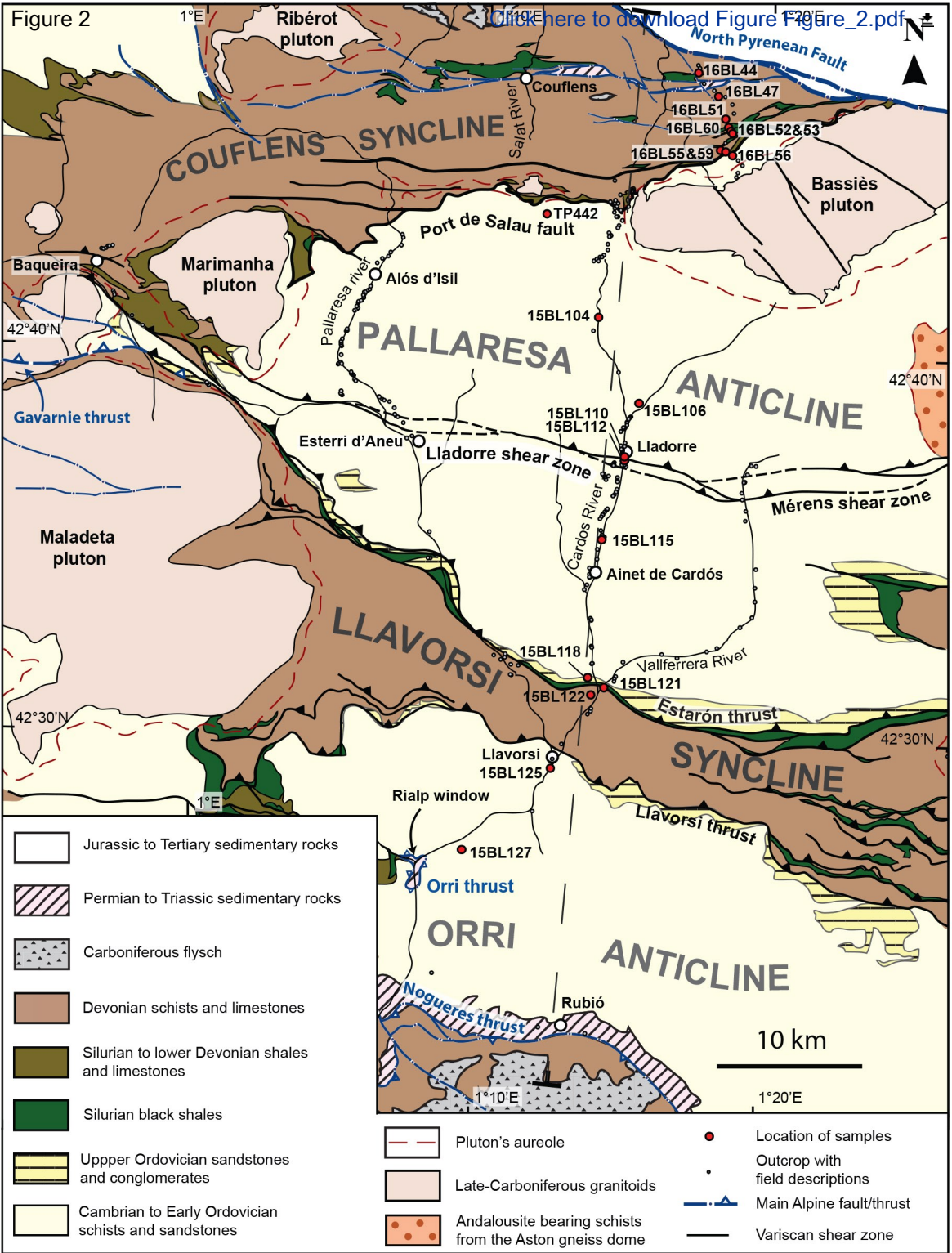
1058 Figure 8: Microphotographs of various rocks of the Pallaresa. Bt: biotite, Ms: muscovite, Chl:
1059 chlorite. Explanations are in the text. **(a)** Precambrian schist of the Palarressa anticline (sample
1060 15BL106). **(b)** Microphotograph of a muscovite-bearing schist. Schistosity, underlined by
1061 muscovite, with dip around 30° to the north and inflected by a “top-to-the-south” C’ shear band.
1062 **(c and d)** Quartz pebble of the Upper Ordovician conglomerate showing dynamic recrystallization:
1063 bugling in **(c)** and sub-grain rotation in **(d)**. **(e)** Chlorite-bearing schist of the Orri anticline. Sa:
1064 Steep alpine cleavage, S1: main schistosity (S1).
1065

1066 Figure 9: Tectonic evolution of the Axial Zone. Further explanations are in the text.
1067

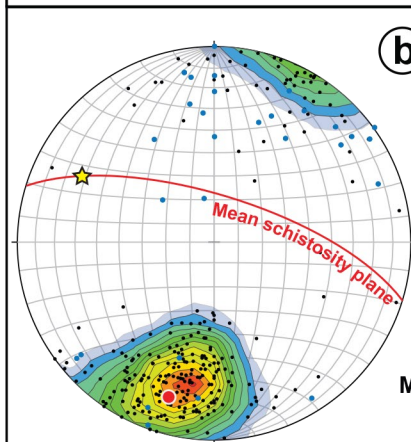
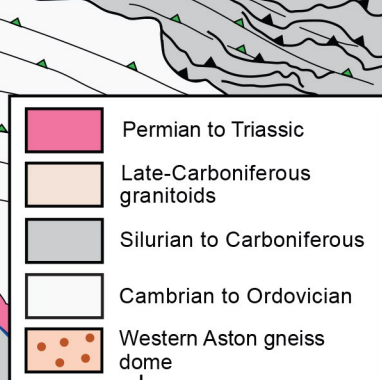
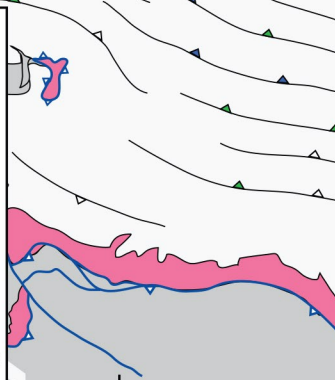
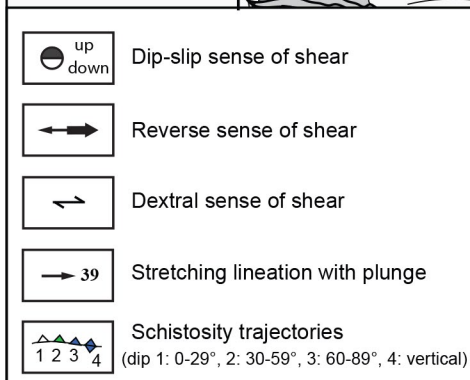
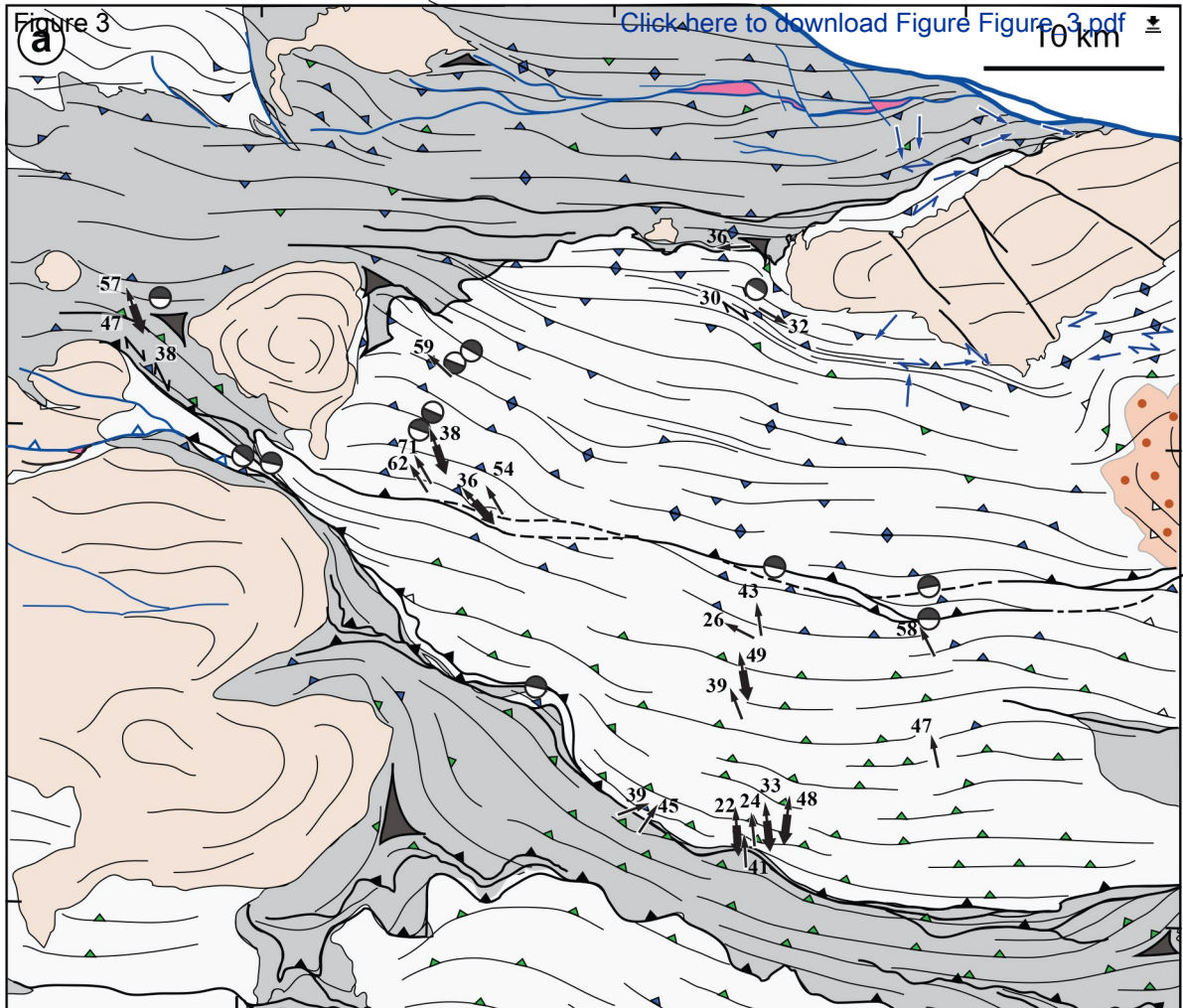
1068 Figure 10: **(a)** 3D bloc-diagram of the Cretaceous transcurrent rifting mode in the Pyrenees ; **(b)**
1069 3D bloc-diagram showing the lithospheric architecture of the Central Pyrenean belt. Interpretation
1070 of deep structures from Choukroune (1989) and Muñoz (1992), except in the Axial Zone. Massifs
1071 with apatite fission-track data in the Central Pyrenees are represented. In brown, the Noguères
1072 cover nappes; **(c)** Apatite fission-track ages obtained in the Central Axial Zone by Fitzgerald et al.
1073 (1999), Sinclair et al. (2005), and Metcalf et al. (2009), plotted together in an age–elevation plot.



Cochelin et al., Figure 1



Cochelin et al., Figure 2

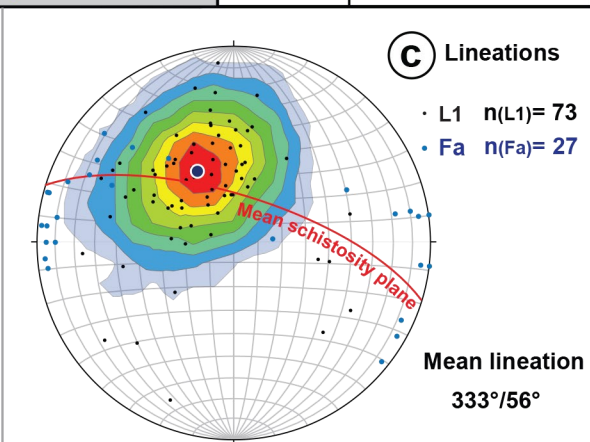


(b) Schistosity poles

- S1 $n(S1) = 260$
- Sa $n(Sa) = 30$

Best fold axis
296°/26°

Mean schistosity
N107°E, 70°N



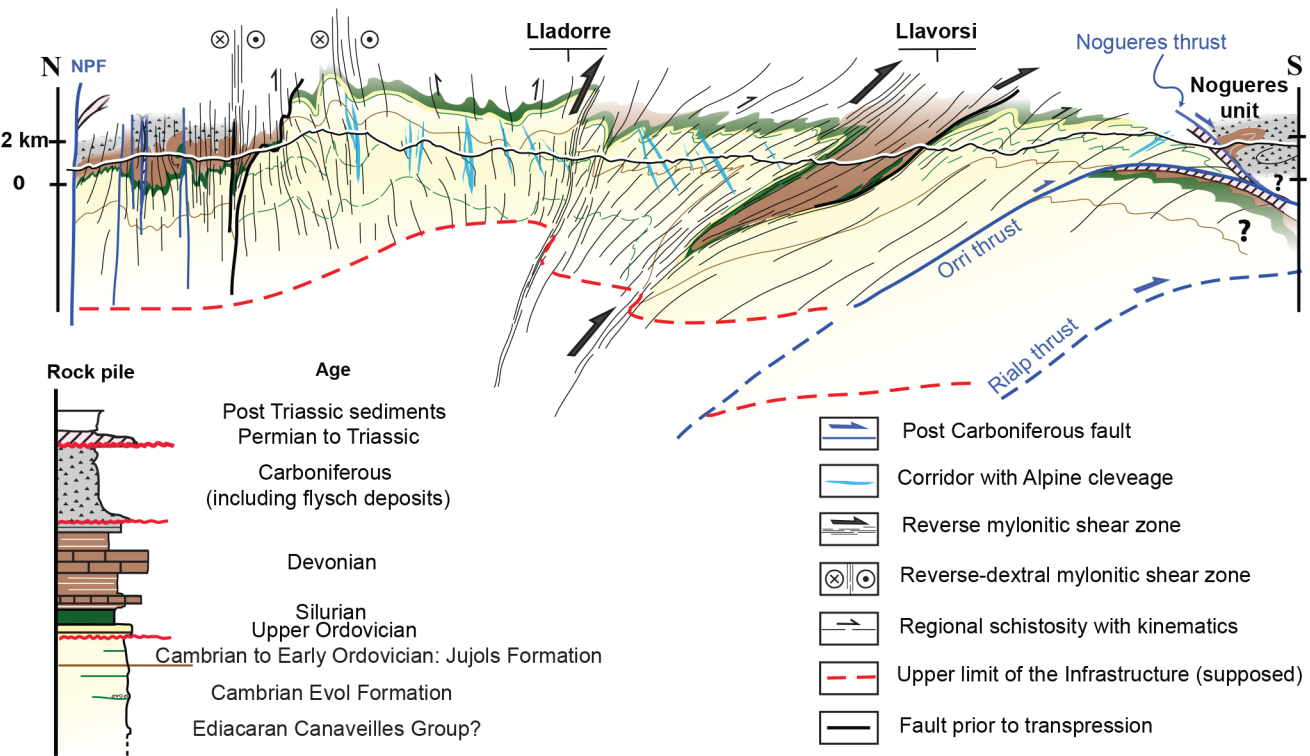
(c) Lineations

- L1 $n(L1) = 73$
- Fa $n(Fa) = 27$

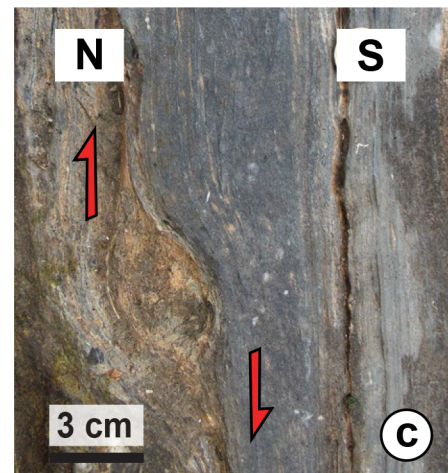
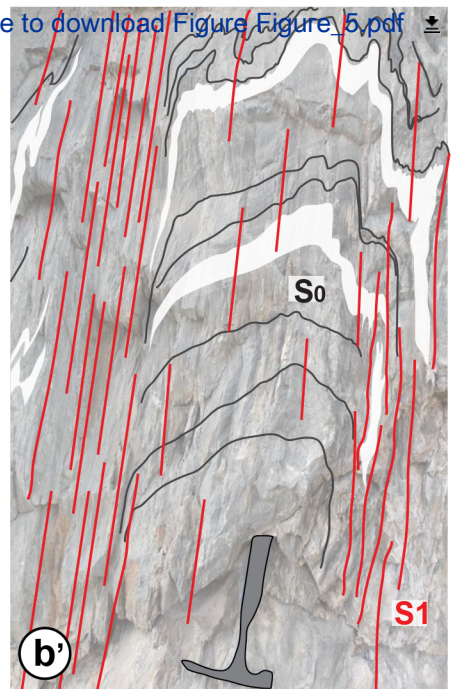
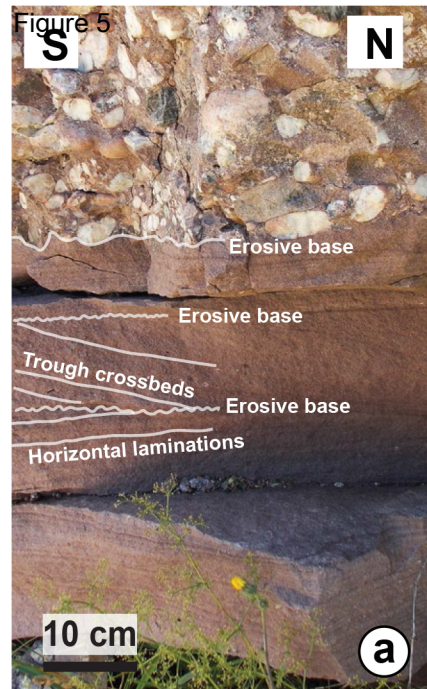
Mean lineation
333°/56°

Cochelin et al., Figure 3

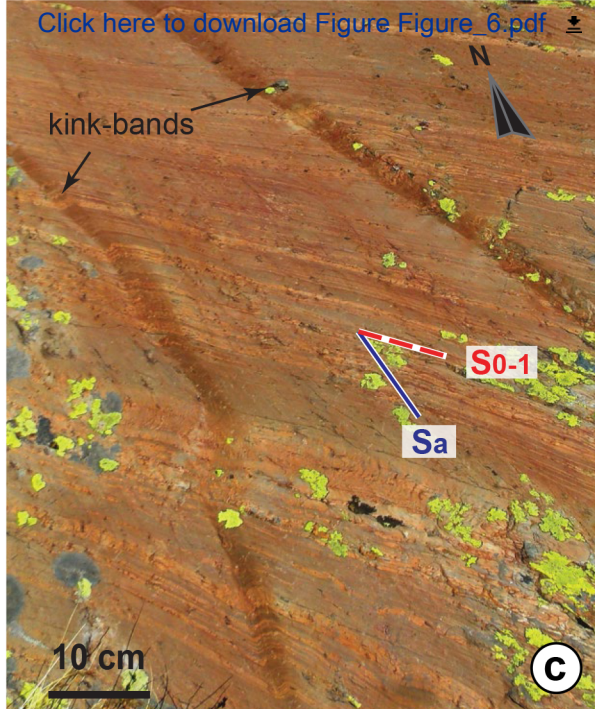
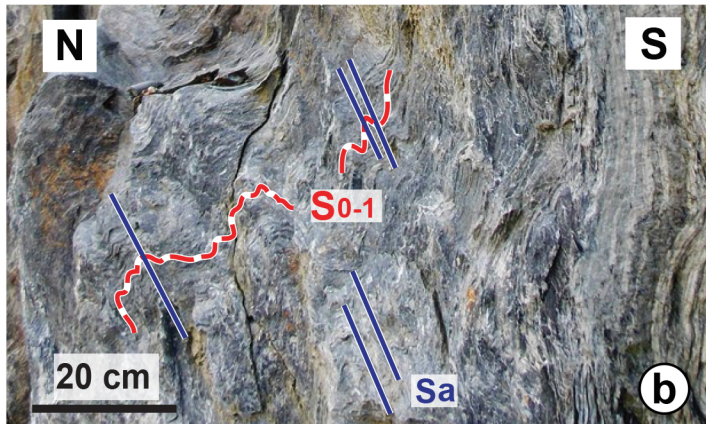
Figure 4 heterogeneous - apparently coaxial deformation High strain zone Low strain zone
Non coaxial «top to the south» shearing



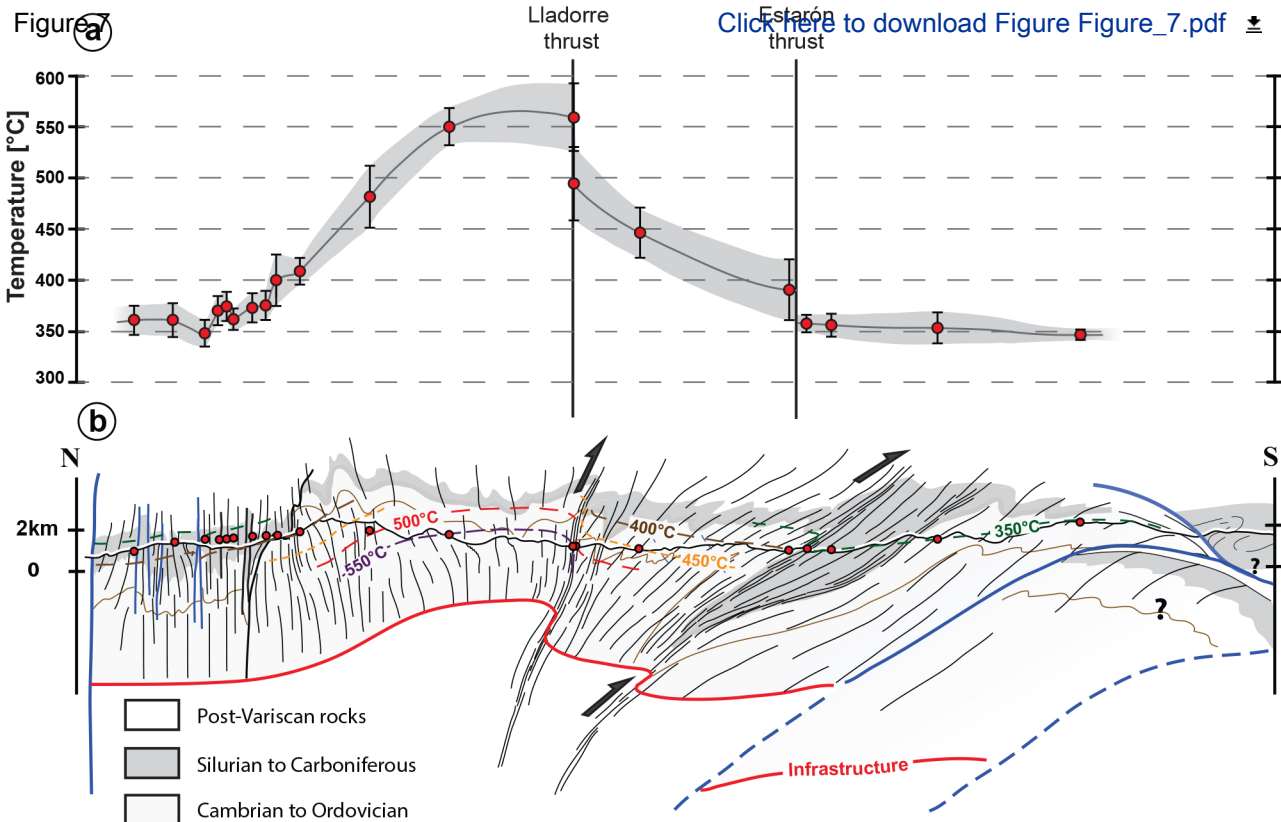
Cochelin et al. Figure 4



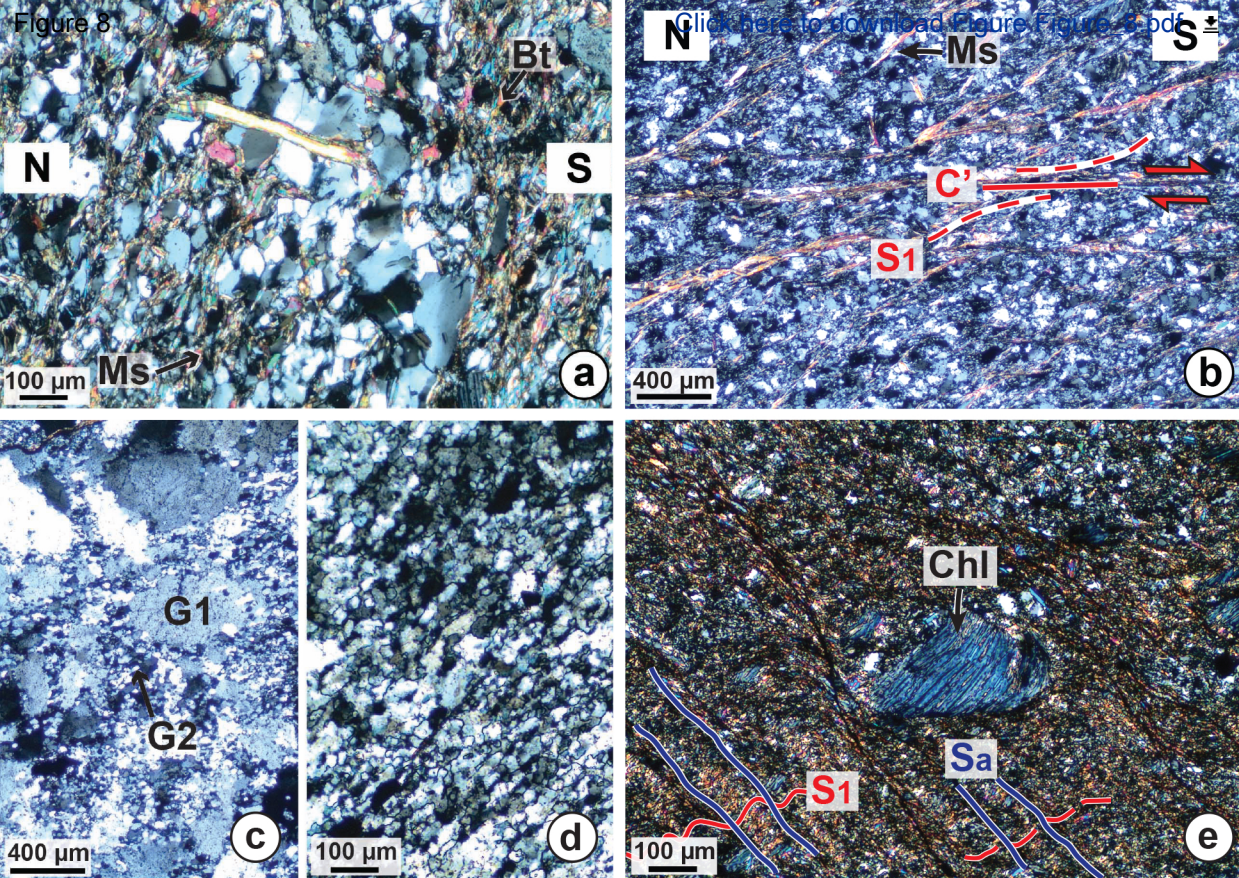
Cochelin et al., Figure 5



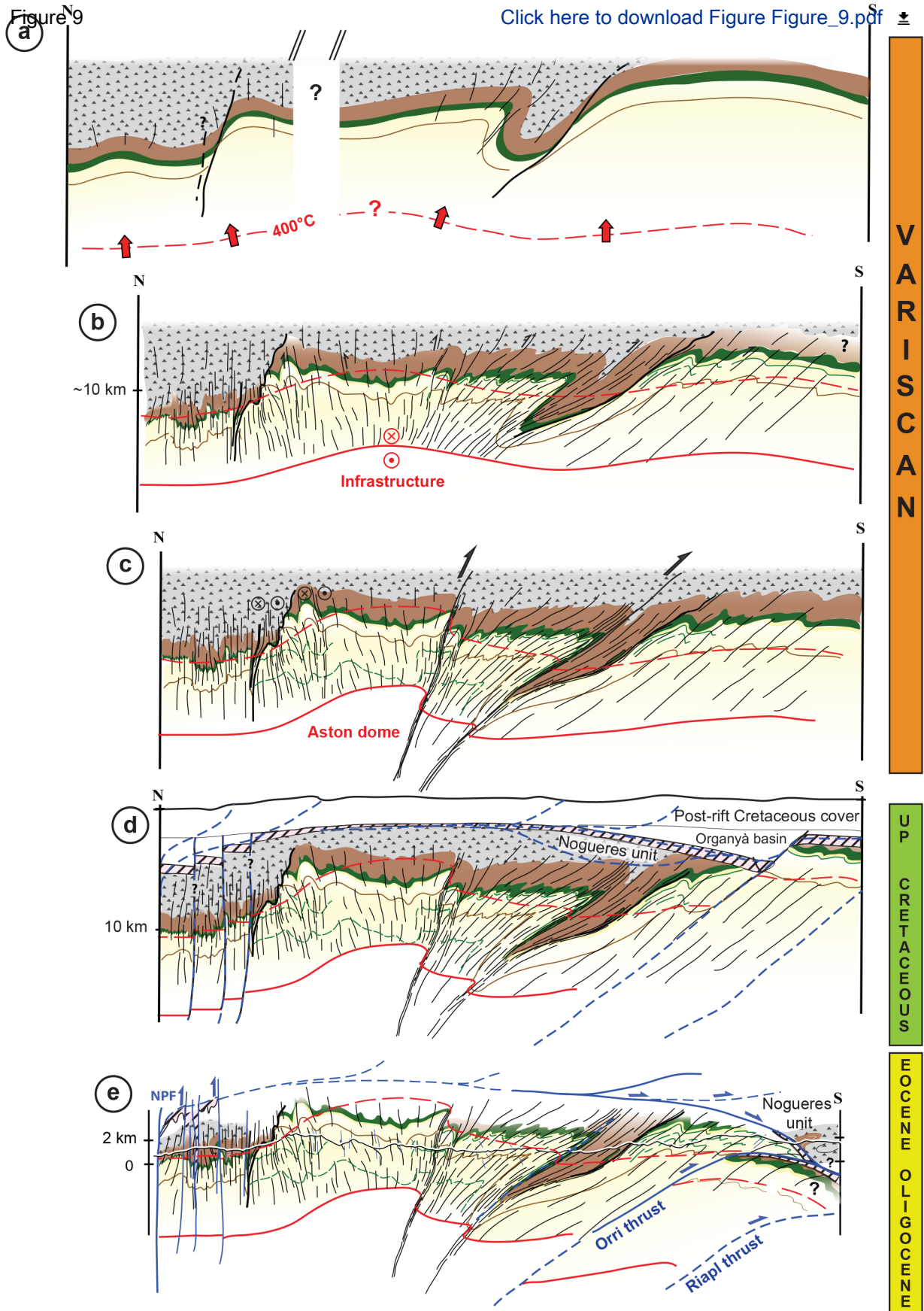
Cochelin et al., Figure 6



Cochelin et al., Figure 7

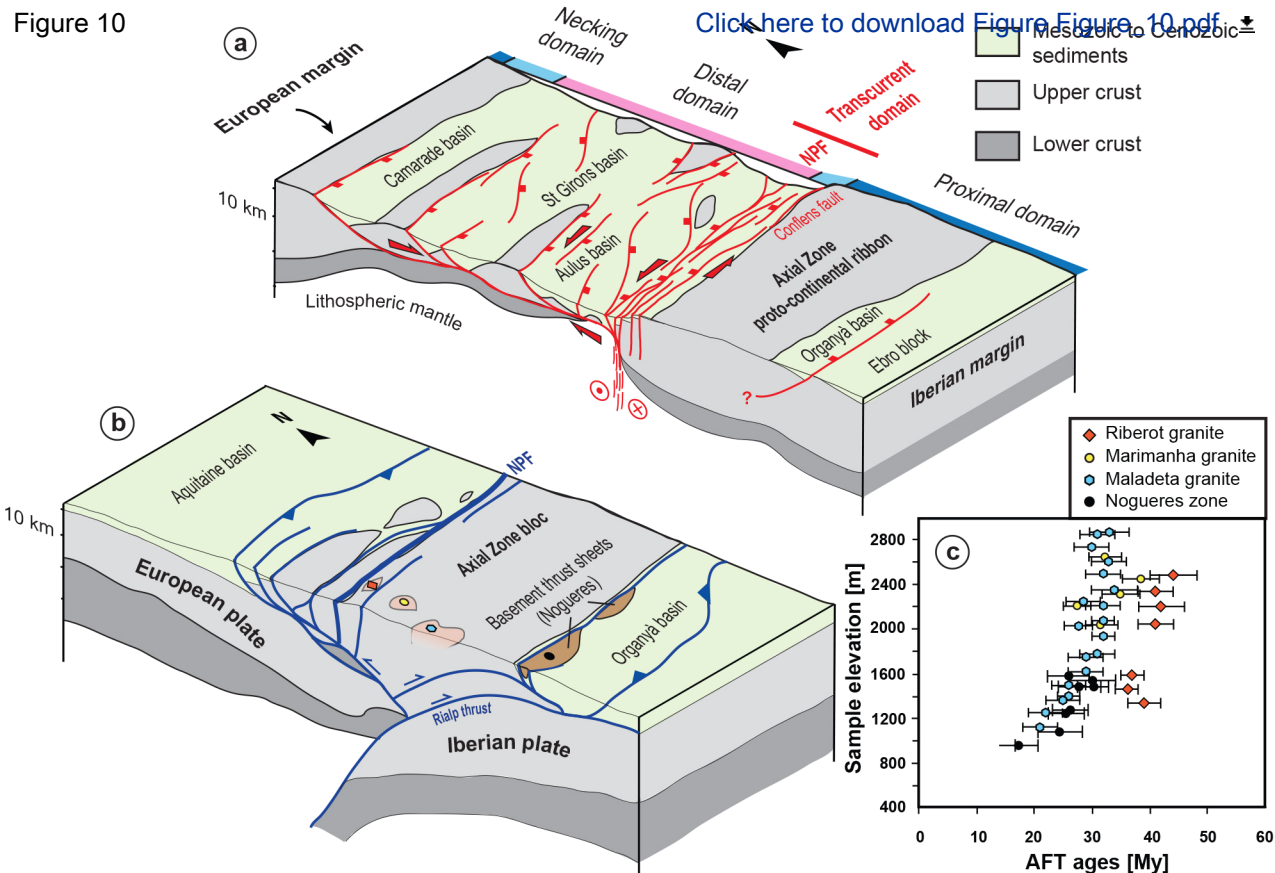


Cochelin et al., Figure 8



Cochelin et al., Figure 9

Figure 10



Cochelin et al., Figure 10