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Shortening of the axial zone, pyrenees: Shortening sequence, upper crustal mylonites and crustal strength

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33 phyllonites but also more generally in the upper crust early during the shortening history, with
34 implications for the shortening style. Estimates of the geothermal gradient suggest that
35 inherited thermicity may also have influenced the style of shortening. We propose that the
36 upper crust was very weak before or at the onset of its shortening due to high-thermal gradient
37 and fluid circulation that induced large-scale sericitization in greenschist facies conditions.
38 This has strong implications on the rheological evolution of the upper crust submitted to
39 metamorphic alteration in the greenschist facies and below.

40

41 **Keywords:** collision, rheology, Pyrenees, basement, greenschist

42

43 **1. Introduction**

44 The strength of the continental crust is usually estimated by extrapolating flow laws derived
45 from laboratory experiments on quartz (Brace and Kohlstedt, 1980) extrapolated to large
46 scales of space and time. This approach coupled with Byerlee's law (Byerlee, 1978) provides
47 a "simple" two-layer rheological profile including a brittle upper- and a ductile lower part
48 (Ranalli and Murphy, 1987), although it can be more complex. This model is consistent with
49 field observations, showing that deformation is rather localized in fault zones with cataclastic
50 rocks in the upper crust and more distributed in shear zones with mylonites in the lower crust
51 (e.g., Scholz, 1989), and with a transitional regime at the brittle-ductile boundary (e.g., Handy
52 et al. 2007; and references therein). Quartz being the most abundant mineral in the crust, its
53 mechanical behavior is generally used to model the strength of the ductile crust. However, if
54 interconnected layers (Handy, 1994) of weaker phases are present, the mechanical behavior of
55 the polymineralic crustal rocks is dramatically altered. These weak phases often comprise
56 phyllosilicates (e.g., Janecke and Evans, 1988; Shea and Kronenberg, 1993; Wintsch et al.,
57 1995; Gueydan et al., 2003).

58 Sericitization, i.e. the breakdown of feldspar into much smaller-sized mica (e.g., Wintsch et
59 al., 1995), is a common process during crustal deformation under greenschist facies
60 conditions (e.g. Wibberley, 1999). It is generally accompanied by crystallization of kaolinite,
61 chlorite, phrenite, pumpellyite, calcite and albite, by-products of the hydrothermal alteration
62 of granitoids (e.g., Ciesielczuk and Janeczek 2004). Sericitization of feldspars triggers both an
63 increase of the rock ductility and a decrease of its strength. The deformation of mica-rich

64 rocks (such as phyllonites) often results in penetrative schistosity, and S/C-C' type fabrics
65 (e.g., Jefferies et al., 2006). The strength of such rocks is significantly lower than that of
66 quartzo-feldspatic rocks (e.g. Wintsch et al., 1995) and experiments on anisotropic rocks have
67 shown that even a small amount of phyllosilicates (e.g., talc, Collettini et al., 2009a;
68 Niemeijer and Spiers, 2005; Niemeijer et al., 2010) has a large effect on bulk rock strength.
69 Phyllosilicate fabric in fault zones may therefore explain the weakness of some crustal faults,
70 such as the San Andreas fault (Lachenbruch and Sass, 1980; Zoback et al., 1987; Holdsworth
71 et al., 2011; Carpenter et al., 2011), the Zuccale Fault (Collettini et al., 2009b), and the
72 Median Tectonic Line (Jefferies et al., 2006).

73 The initiation of syn-kinematic phyllonitization increases the contrast between the strength
74 of the fault zone and the strength of the adjacent rocks, enhancing strain localization. Less
75 studied is the case where pre-kinematic rock alteration may decrease the rock strength. Such
76 case can be found in context where distributed deformation is observed. However, distributed
77 deformation of upper crustal basement rocks is generally poorly studied (except in e.g., Butler
78 et al., 2002; Butler and Mazzoli, 2006; Carminati, 2009, Mazzoli et al., 2009) mainly because
79 markers are usually lacking and polyphased tectonic history complicates strain analysis. In the
80 basement units of the Axial Zone in the Pyrenees (**Fig. 1**), such a study is possible thanks to
81 late Variscan granitoids, that appear to be low-deformed at solid state during late Variscan
82 times that bear Alpine deformation. Moreover, the Triassic cover crops out in many places
83 above the basement in the Axial Zone. This provides a marker to investigate the Alpine
84 shortening in terms of localization and magnitude, and to study the related deformation
85 mechanisms of crustal rocks at temperature around and below 300°C. Indeed, in the Axial
86 Zone, basement units are usually considered to have been buried and deformed during
87 Tertiary below 300°C (e.g., Fitzgerald et al., 1999; Sinclair et al., 2005; Jolivet et al. 2007;
88 Gibson et al., 2007; Metcalf et al., 2009). Shortening is accompanied by metamorphism in
89 greenschist facies conditions that affected both the Paleozoic sedimentary and
90 metasedimentary rocks and the late-Variscan granites and granodiorites.

91 The present contribution first reports on the structural evolution of the Axial Zone basement
92 units. For this purpose, we present a structural study consisting of a regional, balanced cross-
93 section and an outcrop-scale analysis of the shortening kinematics. Second, this contribution
94 documents the microstructural evolution of basement rocks and the weakening processes that
95 they undergo before and during deformation. A thermometric and thermochronological study
96 was carried out to document the thermal and metamorphic history. Eventually, the effects of

97 widespread sericitization of the granite/granodiorite and of the inherited hot geothermal
98 gradient and their implications on crustal rheology are discussed.

99

100 **2. Geological setting**

101 The Pyrenean mountain belt results from the closure of a rifted domain located in the
102 present-day North Pyrenean Zone (NPZ, **Fig.1**). This rift opened during Albo-Cenomanian
103 times at the eastern end of the Bay of Biscay (Puigdefabregas and Souquet, 1986) and was
104 subsequently shortened during upper Cretaceous (around 84 Ma) to Oligocene times (Olivet,
105 1996). During the collision, north-verging crustal thrusts deformed the NPZ, while south-
106 verging ones affected the Axial Zone and the South Pyrenean Zone (SPZ, **Fig. 1, 2**). While in
107 most of the SPZ, the structural style is thin-skinned, the Axial Zone is composed of few
108 basement units, namely the Lakora, Gavarnie, Bielsa-Millares, and Guarga in the western part
109 (**Fig. 2**) and Gavarnie/Nogueres, Orri, and Rialp in the East (Teixell, 1996; Munoz, 1992;
110 Roure et al., 1989; Mouthereau et al., 2014). The nappe stack is usually described as an
111 antiformal stack, where the lowest unit is the youngest. This antiformal stack geometry is
112 indeed clear in the western axial zone (Jolivet et al. 2007; Teixell, 1998; Teixell et al., 2018)
113 but was questioned in the central axial zone (Cochelin et al., 2018) contrarily to what was
114 classically proposed (Munoz, 1992; Beaumont et al., 2000; Roure et al., 1989). As this study
115 focuses on the western part, this controversy will not be discussed in the contribution.

116 In the westernmost part of the axial zone (Teixell, 1996), the Lakora thrust (**Fig. 1, 2**)
117 emplaced the Iguntze massif over the Gavarnie unit during at least Lutetian times (Teixell,
118 1996) during the deposition of the Hecho group turbidite series (Labaume et al., 1985) in the
119 Jaca foreland basin. In the footwall of the Lakora thrust, the Larra thrust affected the
120 Mesozoic cover of the Gavarnie unit and, later on, the Gavarnie thrust affected the basement,
121 triggering the southward migration of the foreland basin during late Eocene-early Oligocene.
122 During Oligocene times, the Guarga unit was emplaced below the turbiditic basin and
123 controlled its southward migration and evolution into a molasse-type overfilled basin (**Fig. 2**).

124 Further east, the Lakora unit is ill-defined, especially its southern boundary. The
125 Gavarnie/Lakora unit, at the longitude of our study, is affected by a distributed zone of
126 deformation including the Pic du Midi thrust, the Eaux Chaudes shear zones (Lamouroux et
127 al., 1979; Henderson and McCaig, 1996; Ingles et al., 1999), and the Pic Long thrust, from

128 North to South, respectively (see Jolivet et al., 2007). This large zone is probably the eastward
129 continuation of the Lakora thrust (**Fig. 2**).

130 In the southern part of this diffuse zone of deformation, the Lakora/Gavarnie unit is
131 composed of the Late Variscan Neouvielle granitic massif and surrounding Paleozoic rocks
132 (Gleizes et al., 2001; Roman-Berdiel et al., 2004) and emplaced above the Gavarnie thrust. In
133 the Neouvielle massif, alpine reverse shear zones linked to the Eaux Chaudes thrust system
134 were dated at about 47 +/- 8 Ma or 48 +/- 2 Ma (Wayne and McCaig, 1988). These ages are
135 similar to the probable age of reactivation of the Merens shear zone activity further East
136 around 50 Ma (McCaig and Miller, 1986). It should be noted that the formation of the
137 numerous shear zone of the Pyrenees lead to several discussions about the age of their
138 formation, variscan or alpine. Although some authors have proposed that some of these shear
139 zone belonging to a variscan anastomosed network that have influenced the evolution of the
140 variscan gneiss dome systems (Denèle et al., 2007; Cochelin et al., 2017) and the dynamic of
141 pluton emplacement (Gleizes et al., 1998; Denèle et al., 2008), they consider also that these
142 shear zone have been clearly reactivated during alpine orogenesis as in the Neouvielle pluton
143 (Gleizes et al., 2001) and the Merens zone (Denèle et al., 2008) that could explain alpine ages
144 of synkinematic phyllosilicates (see above). Apatite Fission-Tracks (AFT) in Neouvielle ages
145 range between 22 and 27 Ma while in northern massif (Bordères Louron) the ages range
146 between 31 and 36 Ma (Morris et al., 1998; Jolivet et al., 2007).

147 In the footwall of the Gavarnie thrust, the Bielsa/Millares unit is composed of the Late
148 Variscan Bielsa granitic massif and surrounding Paleozoic rocks (**Fig. 3**). This unit can be
149 subdivided into two sub-units: the Millares and the Bielsa unit. A basement thrust is probably
150 present below the Bielsa unit (Casas et al., 2003; Jolivet et al., 2007) although it does not crop
151 out. Before this thrust initiation (Miocene, Jolivet et al., 2007), the Bielsa/Millares unit was
152 part of a larger basement unit, namely the Guarga unit (**Fig. 2**) (Teixell, 1996), which
153 emplacement tilted the Lutetian Hecho turbidites in the Jaca basin (Labaume et al., 1985;
154 Teixell, 1996) probably during Late Oligocene times (Jolivet et al., 2003). In the Bielsa
155 granite, AFT ages range between 10 and 20 Ma (Morris et al., 1998; Jolivet et al., 2007) and
156 have been interpreted as witnessing the out-of-sequence initiation of the Bielsa thrust (Jolivet
157 et al., 2007), i.e. after the Guarga thrust initiation.

158 South of the Bielsa unit, the Monte Perdido unit is a sedimentary nappe that might root in
159 the Eaux Chaudes shear zones (Labaume et al., 1985; Teixell, 1996).

160 In the central Axial Zone, further East, in the Gavarnie unit, a zircon fission-track (ZFT) age
161 of 49 Ma was provided by Sinclair et al. (2005). AFT ages range between 26 to 45 Ma
162 (Fitzgerald et al., 1999; Sinclair et al., 2005), whereas apatite (U-Th)/He ages are between 17
163 and 37 Ma (Gibson et al., 2007).

164 In the Orri unit, structurally below the Gavarnie unit, AFT ages in the Maladeta massif
165 range between 17 and 35 Ma (Morris et al., 1998; Fitzgerald et al., 1999; Sinclair et al., 2005;
166 Gibson et al., 2007), apatite (U-Th)/He ages range between 10 and 35 Ma (Gibson et al.,
167 2007). ZFT data present both Mesozoic (104 and 159 Ma) and Cenozoic ages (49 Ma),
168 (Sinclair et al., 2005). These ages show a rapid exhumation during early Oligocene times,
169 following and followed by a slower exhumation during Eocene and late Oligocene times
170 (Fitzgerald et al., 1999).

171

172 **3. Structural and microstructural results**

173

174 3.1. Regional cross-section and structural study

175 As shown by the cross-section (**Fig. 3, 4**), the lower Triassic layers are still attached to the
176 basement top and both are folded. There are up to about 9 antiforms; their wavelength is
177 about 1 km. Six steep faults are associated to some of these antiforms (**Figs. 3, 4**) and the
178 synforms are marked by outcrops of Triassic layers (**Fig. 5, 6**). These layers are offset by
179 steep faults, located on the southern side of the synforms. The faults are either normal or
180 reverse: the southernmost and the northern most ones appear to be normal faults, while the
181 other ones are reverse faults. The synforms are usually asymmetrical generally with steeper
182 limbs on the faulted side (**Figs. 5, 6, 7b,c**).

183 The cross-section has been restored considering that the basement top (i.e. the lower Triassic
184 layers) has a constant length during shortening. The Triassic layers were probably deformed
185 during shortening (e.g., lengthening of the layers), by a probably very low amount, which is
186 negligible for restoration purposes. The second main assumption concerns the geometry of the
187 sub-vertical faults. Here, we chose to restore these faults (**Fig. 4**) as we consider them to be
188 pre-orogenic for the following reason: the Fobillons and Punta Suelza klippe are clearly not
189 as folded as the underlying Triassic layers and basement top (**Figs. 4 and 7a**). This means that

190 folding and the vertical faults predated the end of the burial of Bielsa unit below Millares and
191 Gavarnie. If the vertical faults were coeval with the Bielsa thrust, the Millares thrust would
192 have been folded too, but this is not the case. Assuming this, the restoration provides an
193 amount of shortening of 1.7 km (13%). If the sub-vertical faults were syn-shortening, the
194 amount of shortening would have been slightly higher. The faults may be Cretaceous in age
195 (either pre- or syn-Cenomanian) as suggested by Johnson et al. (1996).

196 The northernmost synform (**Fig. 3, 4, 5**, Urdiceto synform) varies structurally along strike.
197 East of the Lago Urdiceto, the synform is pinched and associated to significant deformation in
198 the basement (**Fig. 6a**), as shown by a penetrative schistosity (**Fig. 6, 8a**), with some shear
199 zones (**Fig. 6, 9**). This schistosity is especially well expressed where the basement shows a
200 thick altered/weathered upper horizon (several tens of meters; **Fig. 8**). The basement
201 weathering seems to predate the shortening as unaltered areas present a less penetrative
202 schistosity (**Fig. 8b**). This weathering is most likely Permian in age as the Triassic cover is
203 not affected. West of the Lago Urdiceto, shortening was less intense and the basement
204 presents less deformation (**Fig. 5, 6**).

205 Within this synform, several features attest for shortening in the cover. Top-to-the-South
206 kinematic indicators are observed above the basement-cover interface on the northern side of
207 the synforms (**Fig. 5, site 37/38; Fig. 10a**): there, no large-scale décollement occurred but the
208 schistosity is deflected in a sigmoidal pattern due to shearing along the bedding planes.
209 Further up in the cover, tight folds are accompanied by coeval sub-vertical schistosity (**Figs.**
210 **10b, c**) along with a vertical stretching attested by numerous sub-horizontal veins (**Fig. 10b**)
211 with sub-vertical fibrous crystals of calcite, quartz, and chlorite. These cover folds with
212 wavelength much smaller than those of the basement top implies that there is a slight
213 décollement above the basement top unconformity.

214 Within the synform further south (**Figs. 3, 4**), the border fault dips south and is reverse
215 (**Figs. 4, 6, 7c**). Folds are associated with a steep cleavage (**Fig. 7c**) and the cover is affected
216 by reverse faults, at shallower dip than the main border fault (**Fig. 11**).

217 Between the synforms, shear zones are widespread in the basement. They consist of thin (<
218 1 m) zones dipping 45 to 70° N with both south- and north-verging kinematics (mainly south-
219 verging, **Fig. 6, 9**). Quartz crystals are affected by cataclastic deformation within shear zones,
220 while chlorites are affected by a more ductile-like deformation and present sigmoidal shapes

221 (**Fig. 12 and 13**). Alpine schistosity and shear zones are also characterized by interconnected
222 layers of white micas (see below).

223 The steep faults (**Fig. 5, 6, 7b, c, 11a**) bordering the synforms have been sampled and show
224 typical cataclastic textures (**Fig. 12a, b**). The quartz are broken up in millimetric clasts. These
225 structures are linear and affect a hydrated protolith: phases such as chlorites and white micas
226 (in feldspars, see below) are widespread in thin sections but do not affect significantly the
227 distribution of cataclasites (**Fig. 12a, b**). The cataclasites do not either seem to control the
228 distribution of the hydrated phases.

229

230 3.2. Sericitization

231 In samples of un-deformed rocks (**Fig. 13a, 14a**), the mineral assemblage is composed of
232 quartz, K feldspars, plagioclase, chlorites, biotites, white micas as well as accessory minerals
233 such as alanites, zircons, titanites, rutiles, pyrites, apatites. Quartz and feldspars form
234 subhedral to anhedral millimetric grains (**Fig. 12**). Quartz commonly contains fluid inclusion
235 planes. Chlorites are large (few tens of mm) and they overgrow biotites, with titanites and
236 rutiles inclusions. The chlorites are also found smaller interlayered with white micas.

237 In undeformed and weakly deformed samples, K feldspar is replaced by albite that grows in
238 fractures and along mineral cleavage (**Fig. 14a**), where white micas and carbonates are also
239 observed. With increasing destabilization, in more deformed samples, both albite and white
240 micas develop in these K feldspars (**Fig. 14b**) and totally replace them (**Fig. 14d, e**). Locally,
241 prehnites and epidotes can be observed. Moreover, white micas are also associated to
242 chlorites that derive from the biotite destabilization (**Fig. 14c**).

243 In samples from deformed rocks, the amount of albite and white mica increases and a
244 schistosity appears, formed by the alignment of micas and chlorites (**Fig. 14d**). The rock
245 mainly consists of quartz, white mica, and albite. Quartz is mainly affected by undulose
246 extinction and some fractures, but only little or no dynamic recrystallization (**Fig. 13b**).

247 In very deformed rocks, the assemblage is mainly composed of deformed quartz and white
248 micas (**Fig. 14e**). Thus, the amount of white micas has increased with deformation. This is
249 most likely due to the total sericitization of feldspars that is probably favored by mineral
250 fracturing, hence fluid infiltration. White micas are often aligned and define micro-shear

251 zones and schistosity (**Fig. 13c**). Quartz clasts and chlorites are embedded in a white mica
252 “matrix” (**Fig. 13c, 14e**), with some calcite close to faults affecting the basement (that may
253 have provided fluids with calcium). In the white mica “matrix”, very small scale aggregates of
254 poorly oriented grains can be observed (**Fig. 14e**).

255

256 **4. Geo-thermometry**

257 Phyllosilicates provide thermobarometry constraints on samples from the Bielsa massif (**Fig.**
258 **15**). Thermodynamic modelling using multi-equilibrium thermobarometry yields
259 crystallization temperatures in the range 250-300°C for micas (Dubacq et al., 2010) and
260 chlorites (Vidal et al., 2006) (both +/- ~50°C).

261 The Raman Spectrometry of Carbonaceous Material (RSCM) method (Beysac et al., 2002;
262 Lahfid et al., 2010) is based on the quantitative study of the degree of (irreversible)
263 graphitization of carbonaceous material (CM), which provides the maximal temperature
264 (Tmax) attained during the rock history. Tmax can be determined in the range 200–650°C
265 (Beysac et al., 2002; Lahfid et al., 2010), with absolute precisions of ca. 50°C between 650
266 and 300°C and ca. 25°C between 330 and 200°C. Relative uncertainties between samples are
267 probably much smaller (around 10-15°C according to Beysac et al., 2004, 2007; Gabalda et
268 al., 2009; Vitale Brovarone et al., 2013). Raman spectra (Supplementary Material) were
269 measured and analyzed following Beysac et al. (2002, 2004, 2007) and Lahfid et al. (2010).
270 Raman spectra were obtained using a Renishaw InVIAReflex microspectrometer (BRGM,
271 Orléans, France) with a 514-nm argon laser on polished thin sections from CM-rich Mesozoic
272 limestones (Cenomanian, see **Fig. 3** for location).

273 The RSCM study yields Tmax between 236 and 264 °C +/- 30°C for the cover of the Bielsa
274 massif (**Table 1**). The samples are part of the Bielsa unit, from the Cenomanian cover, mostly
275 from the southern part of the area. There, the cover is slightly detached from the basement and
276 the Triassic layers. However, Tmax obtained from these samples may be confidently assigned
277 to the Bielsa unit. In the Gavarnie unit, one RSCM analysis indicates Tmax of 328°C +/-
278 30°C (**Table 1, Fig. 16** for location).

279

280 **5. Zircon fission-track data and thermal history**

281

282 5.1. Zircon fission track data

283 In the Neouvielle massif, ZFT ages were obtained from three samples, NAZ-1, NAZ-7 and
284 NAZ-6, collected along an elevation profile at 1881 m, 2400 m, and 2700 m, respectively
285 (**Table 2, Fig. 16**). No major faults were mapped between these samples, although some shear
286 zones are present (Fig. 17). These shear zones do not present very high displacement and
287 were dated to be active around 48 Ma (Ingles et al., 1999; Wayne and Mc Caig, 1998). All
288 three samples pass the χ^2 -test and have central ages at around 45-46 Ma (**Fig. 16, Table 1**).
289 Nonetheless, the dispersion of single grains ages shows an interesting trend, decreasing from
290 16% at 1881 m elevation to 0% at 2700 m elevation. Furthermore, minimum ages, which
291 indicate the youngest coherent age population in a grain age distribution (Galbraith and
292 Laslett, 1993), are 38 Ma, 41 Ma and 45 Ma for samples NAZ-1, NAZ-7 and NAZ-6
293 respectively (**Fig. 16, Table 2**). In the Bielsa massif, 2 ZFT ages were obtained, with central
294 ages of 143 and 161 Ma, for elevations of 1680 and 2700 m, respectively (**Fig. 16, Table 1**).
295 Between these two samples no major faults were mapped (**Fig. 3, 16**)

296

297 5.2. Interpretation of ZFT data

298 The dataset from the Neouvielle massif provides Cretaceous to Cenozoic apparent cooling
299 ages (**Fig. 16**). The difference between the minimum ages and central ages of samples NAZ-1
300 and NAZ-7 may be related to slightly different closure temperatures in zircons that have
301 accumulated higher amounts of radiation damage because of high U concentrations. At least
302 the age and U concentration information shown in the radial plots display a weak trend of
303 higher U concentration grains having younger apparent cooling ages. The interpretation could
304 be the following: Cenozoic cooling started at around 46 Ma given the very similar central
305 ages of all three samples, but the increasing age dispersion and trend to younger minimum
306 ages for the lower elevation samples indicate that these rocks may have been affected by very
307 minor partial annealing, possibly resetting in the upper part of ZFT the partial annealing zone
308 for a short time before cooling to temperatures below the partial annealing zone.

309

310 5.3. Neouvielle thermal history

311 The thermal history for the Neouvielle massif (Gavarnie unit) is defined as follows. During
312 Mesozoic times, the temperature probably slightly increased due to rifting (the amplitude of
313 the temperature increase being difficult to constrain, see discussion below on the inherited
314 thermal structure) around 100 Ma and due to upper Cretaceous sedimentation. During Eocene
315 times, the temperature increased as a consequence of both the turbiditic basin formation
316 (Hecho group) and the overthrusting of the Lakora unit. The latter effect was probably limited
317 given the small amount of displacement on this thrust (see cross sections in Teixell, 1996).

318 The RSCM data are too scarce to be conclusive, but they provide a T_{max} estimate of about
319 330°C (**Table 1**), also consistent with Jolivet et al. (2007), who estimated T of about 300°C
320 for the Neouvielle massif, mainly obtained from geochronological data. Thus, we set the
321 thermal peak of this unit to about 300°C reached before 50 Ma, given the ZFT central ages
322 around 46 Ma. From the rather limited ZFT data set that we have for the Neouville massif, it
323 seems that cooling started during mid-Eocene cooling. AFT ages for the Neouville massif
324 (Jolivet et al., 2005) indicate that cooling continued during Oligocene times (Fig. 18). Recent
325 U-Th/He data on zircons (Pik et al., in prep) will further constrain the Oligo-Miocene thermal
326 history. The cooling so far is constrained by AFT data from Jolivet et al. (2007) and their
327 thermal path and the age of activity of the Gavarnie thrust (**Fig. 17**)

328

329 5.4. Bielsa thermal history

330 The Jurassic ZFT ages obtained for the Bielsa massif (**Fig. 16**) indicated a different history.
331 Either the ages indicate a thermal event during Jurassic times or there has been a partial reset
332 during the mid-Cretaceous times (rifting). However, the grain age distribution does not
333 support the latter interpretation (they pass the χ^2 -test and have 0% age dispersion). Due to the
334 limited number of samples and number of grains per samples, we cannot conclude about the
335 Mesozoic thermal history. However, it indicates that the Pyrenean shortening was not
336 responsible for any reset of this samples. It suggests a lower burial and/or a lower geothermal
337 gradient than, for the Neouvielle massif.

338 On **Fig. 17**, the Bielsa thermal path is colder than in the Neouvielle Massif as, in the Bielsa
339 massif, the analyzed samples are from a structural level slightly higher than the Neouvielle
340 massif. In the former samples are from the basement top, while in the latter there were
341 probably few km of basement above the samples. During Eocene times, heating of the Bielsa

342 massif is probably due to foreland sedimentation (Hecho group, **Fig. 17**) and possibly by the
343 emplacement of the Gavarnie unit.

344 The T_{max} of about 250°C according to the RSCM data (possibly Cenozoic in age) is in line
345 with the Mesozoic ages ZFT we obtained: such temperature did not imply a Cenozoic reset of
346 ZFT ages. We cannot derive from our data a complete thermal history but the exhumation
347 during the last 35 Ma may be similar to the exhumation of the Neouvielle unit. The final
348 thermal path is constrained from Jolivet et al. (2007) and the age of activity of the Guarga
349 (**Fig. 2**) thrust that might have exhumed the Bielsa unit (**Fig. 17**).

350

351 **6. Discussion**

352 Our study shows that the small amount of shortening in the Bielsa unit is distributed at
353 kilometer scale, as attested by the “basement antiforms” mapped in the field and shown in
354 **Figures 6 and 7**. This distributed mode of shortening is also observed in the distributed
355 ductile strain within both cover and basement units, where it may be explained by a pre-
356 kinematic sericitization of the granitic basement.

357

358 6.1. Shortening kinematics and restoration

359 In our structural study, we restored the cross-section to its pre-orogenic stage, including the
360 present-day steep faults, inferred to predate the shortening. Casas et al. (2003) proposed that
361 these steep faults were rather late during the shortening history. Their main argument is that
362 these faults do not seem to be overprinted by the contractional deformation. Indeed, the small-
363 scale, low-angle reverse shear zone or faults do not crosscut the steep faults (**Fig. 11**),
364 however, the latter may be pre-orogenic for several reasons.

365 (1) The steep faults appear to have various dip and sense of dip: some of them are normal
366 faults others are reverse ones (**Fig. 4**). This geometry is not predicted in the model by Casas et
367 al. (2003), where steep faults are due to distributed deformation during fault bend folding. On
368 the contrary, if these faults are inherited, they may have been deformed and curved during
369 shortening, especially if they were not reactivated (e.g., Marquer et al., 2006; Lafosse et al.,
370 2016).

371 (2) In the Casas et al. (2003) kinematical model, the steep faults form lately in the basement,
372 after the initiation of the underlying Bielsa basement thrust. The steep faults form because the
373 basement unit deforms as a fault-bend fold over a flat-ramp system. In this model, to initiate
374 the steep faults, the displacement along the thrust must be at least about the distance between
375 the northernmost and the southernmost steep fault. This would imply a displacement of at
376 least 5 km that is unlikely for a thrust that is poorly constrained by structural mapping. Recent
377 structural interpretations do not include any flat structures (e.g., Jolivet et al., 2007).

378 (3) The Bielsa unit is affected by horizontal deformation characterized by basement-cover
379 interface antiforms associated with distributed basement deformation. These deformations
380 cannot result from fault-bend fold-associated deformation as in Casas et al. (2003), as such
381 kinematics do not require the formation of a pervasive axial plane foliation. It appears that
382 such contractional deformation is quite systematically present in the hangingwall of steep
383 faults (or northern block when the faults are reverse and north-verging). This suggests that
384 there might be a genetic link between the steep faults and the basement-cover interface
385 antiforms: the latter may localize close to the former. In this case, it would be clear that the
386 steep faults have to be inherited to influence the location of the Cenozoic contractional
387 structures.

388 (4) In the model of Casas et al. (2003), the steep faults being late, they must deform the
389 overlying nappe (Millares). However, in the field the latter does not appear as
390 folded/deformed by steep faults (**Fig. 7a**): The basal thrust of the Millares nappe is only
391 affected by open, large-wavelength folding although below, the Triassic layers are strongly
392 folded.

393 (5) Extension is known to have occurred in the Pyrenees Axial Zone at various times: during
394 the Stephano-Permian (Saura et al., 2006) and during the mid-Cretaceous (Johnson et al.,
395 1996). These extensions may have triggered the formation of steep normal faults. Johnson et
396 al. (1996) showed that mineralizations in veins, in the same basement unit further East in the
397 Pyrenees, formed during Mesozoic extension with fluid flowing downward along normal
398 faults.

399 To sum up, we have shown that the steep faults may have been inherited from pre-orogenic
400 stages, thus supporting our cross-sectional restoration (**Fig. 4**). However, we cannot exclude
401 the fact that these steep faults have been reactivated during the late basement deformation as
402 suggested in Casas et al. (2003).

403

404 6.2. Inherited thermal structure

405 RSCM data, chlorite thermometry as well as thermochronologic data show that the thermal
406 peak is about 300 +/- 30 °C for the Gavarnie unit and around 250 +/- 30 °C for the Bielsa,
407 consistent with Jolivet et al. (2007) and with Ar/Ar step heating data on late Variscan biotites
408 and their likely partial resetting during the Pyrenean orogeny (Jolivet et al., 2007), in addition
409 to temperatures deduced from fluid inclusions and petrography in the Eaux Chaudes shear
410 zones (Henderson and McCaig, 1996; Ingles et al., 1999).

411 These temperatures are very high considering the inferred sedimentary and tectonic burial of
412 both massifs. For the Bielsa unit basement, reasonable estimates of sedimentary burial are
413 about 1 km of Mesozoic and Paleocene formations plus 3 km of Eocene turbidites (Labaume
414 et al., 1985). Such a thin overburden would require a very high geothermal gradient, of about
415 70°C/km, in order to attain 300° C at the cover-basement interface. Thus, it is very unlikely
416 that the recorded Tmax characterizes the temperature at the cover base during the turbidite
417 basin formation. If we consider the Eocene tectonic burial following the turbiditic basin, a
418 thickness of about 5 km (basement and cover of the Gavarnie unit, see cross-sections in
419 Jolivet et al., 2007; with an uncertainty of about 1 km) plus Bielsa cover (only 1 km as the
420 Eocene turbidites were detached before the Gavarnie unit emplaced above the Bielsa unit), we
421 obtain a burial of about 6 km +/- 1 km; this implies a geothermal gradient of at least 45 +/- 10
422 °C/km.

423 Similarly, the sedimentary cover of the Gavarnie basement is too thin to explain the
424 estimated Tmax. The tectonic burial may have been about 2 km +/- 0.5 km (see cross-sections
425 in Teixell, 1996). If we add the thickness of the Gavarnie unit above the analyzed samples and
426 below the Lakora unit (4 km +/- 1 km), we obtain a maximum thickness of about 6 km +/- 1.5
427 km, hence a geothermal gradient of 50 +/- 15 °C/km.

428 Thus, it appears that the geothermal gradients are quite high, the Neouvielle gradient being
429 probably higher than the Bielsa one. However, it is possible that both gradients are similar
430 given the uncertainties. Our results still strongly suggest that both gradients are higher-than-
431 normal (i.e., higher than 25-30 °C/km) and that the Neouvielle gradient is slightly higher than
432 the Bielsa one.

433 Such geothermal gradients are Eocene to post-Eocene in age as they were determined from
434 thermochronologic data yielding Eocene to Oligocene ages and from thermo(baro)metry on
435 phyllosilicates that are at least partly Alpine in age, given that their deformations are
436 kinematically consistent with the deformation in the Triassic meta-sediments. In the North
437 Pyrenean Zone (NPZ), an early Eocene gradient of about 70-80 °C/km was determined
438 (Vacherat et al., 2014) and interpreted as the consequence of Albo-Cenomanian thinning that
439 occurs in the NPZ. The thermal anomaly reached a climax as oceanic spreading started in the
440 Bay of Biscay at 100 Ma. This led to a profound and durable change in the thermal structure
441 of northern Iberia margin lithosphere (Rat et al., 2019) until the plate convergence started.
442 Our results suggest that the thermal anomaly due to mid-Cretaceous rifting processes lasted
443 long enough to be still active during Eocene shortening and that it was large (in space)
444 enough to affect the Iberian crust that was only moderately thinned (i.e., in the future Axial
445 Zone, **Fig. 18**).

446 The rifting event probably stopped around 95 Ma and convergence started just after this
447 event (Mouthereau et al., 2014 and references therein). Considering that shortening recorded
448 in the western Axial Zone probably started around 55 Ma, the thermal anomaly inherited from
449 the rifting stage lasted about 40 Ma. This time interval being quite long, it may suggest that a
450 small-scale sub-lithospheric convection was active in this area, at least until the convergence
451 started at about 80 Ma. This may indeed occur at the rift margins as proposed by Boutillier
452 and Keen (1999) and may have triggered partial melting (van Wijk et al., 2001).

453 The fact that the crust may have suffered a high thermal gradient while it was not
454 significantly thinned may be explained as follows: the lithosphere base is most likely thinned
455 with a wavelength much higher than that affecting the crust, as it is thicker. Such
456 characteristic length is probably about 200-300 km. As a consequence, the lithospheric
457 thermal gradient, controlled by the lithosphere base isotherms, may be affected by thermal
458 anomalies in an area much larger than the area of crustal thinning (**Fig. 18**). This suggests that
459 the thermal anomaly in a lithosphere due to thinning may not be restricted to the area of
460 crustal thinning, i.e., the margin. Such interpretations are in line with models of depth
461 dependent extension (Royden and Keen, 1980) where the crust is less stretched and thinned
462 than the mantle in some margins as suggested by subsidence data (Royden and Keen, 1980;
463 Davis and Kusznir, 2003) and reproduced in thermo-mechanical models (e.g., Huisman and
464 Beaumont, 2011). In the distal part of the rift, the crust is more thinned than the mantle (as in
465 other distal margins, e.g., Peron Pinvidic et al., 2013 and references therein) as attested by

466 mantle exhumation in the rift center (Lagabrielle and Bodinier, 2008; Lagabrielle et al., 2010;
467 Clerc et al., 2012; 2014; Jammes et al., 2009; Masini et al., 2014). In the more proximal part
468 of the rift, the crust is less thinned than the mantle as in many margins worldwide (e.g., Davis
469 and Kuszniir, 2004 and references therein).

470

471 6.3. Thermal and structural evolution

472 The Eaux Chaudes shear zone was active at about 48 +/- 2 Ma (Wayne and McCaig, 1998).
473 Similarly, further East the Merens shear zone was reactivated around 48 Ma (Mc Caig and
474 Miller, 1989). The Larra thrust, linked to the Lakora thrust was active until about 40-45 Ma
475 (Teixell, 1996). Thus, between 50 and 40 Ma, several shear zones/thrusts were active in the
476 Gavarnie/Lakora unit. It is noteworthy that it corresponds to the probable exhumation during
477 Eocene times suggested by ZFT data (see above). The emplacement of these thrust/shear
478 zones (Lakora and Eaux Chaudes) did not thicken the crust significantly (see cross-sections in
479 Teixell, 1996). Moreover, it implies that these thrusts were probably not the main control on
480 the Hecho turbidite basin development. It is possible that this basin was a lateral basin (West
481 of the proto-Pyrenees) due to crustal thickening further east in central Pyrenees where the
482 orogen started to grow earlier (see references above). Since 50-55 Ma, the Gavarnie unit was
483 buried below the Hecho turbiditic basin. However, at about 50 Ma, we record the first
484 shortening, implying that the Gavarnie unit was probably the northern part of the turbiditic
485 basin already slowly exhuming and deforming (Eaux Chaudes shear zones, Wayne and
486 McCaig, 1998) during the deposition of the turbidites. Recently, the Pic de Port Vieux,
487 associated to the Gavarnie thrust has been dated to 37 Ma by Ar/Ar step heating on
488 muscovites (Elmola et al., 2018). This shows that distributed shortening along with
489 exhumation predated the localization of the shortening on the Gavarnie thrust at 37 Ma (**Fig.**
490 **17, 19**).

491 The emplacement of the Gavarnie unit also influenced the burial of Bielsa unit when the
492 Gavarnie thrust became active, around 37 Ma. At this time, the intra-Bielsa deformations (the
493 distributed shortening described in this study) were probably already active (30-40 Ma, **Fig.**
494 **19**). The Guarga thrust activation and the exhumation of the Guarga-Bielsa probably started
495 around 30 Ma. The age of 30 Ma also corresponds to the basinward localization of the Jaca
496 basin, in the Rio Guarga syncline interpreted as the consequence of the Guarga unit
497 emplacement (see Teixell, 1996). This suggests that in the western axial zone a distributed

498 phase of shortening generally predated the activation of each main thrust (Gavarnie and
499 Guarga).

500 Such distributed then localized mode of shortening is similar to the modes of shortening
501 characterized in the Alpine External Crystalline Massifs (ECM) where distributed shear zones
502 were active during Oligo-Miocene times (e.g., Marquer et al., 2006; Rolland et al., 2008;
503 Simon-Labric et al., 2009; Sanchez et al., 2011; Bellahsen et al., 2012; 2014; Bellanger et al.,
504 2015, Lacombe and Bellahsen, 2016; Egli et al., 2016; Egli et al., 2017; Mair et al., 2018)
505 before the activation of frontal crustal thrusts (Bellahsen et al., 2014). In the ECM, the
506 deformation occurred under green schist facies metamorphism, at around 300-350°C (Crouzet
507 et al., 2001; Bellanger et al., 2015; Boutoux et al., 2016), thus at temperatures slightly higher
508 than in the Axial Zone. In the External Alps, such mid-crustal temperatures and the associated
509 distributed shortening are due to significant burial below the internal units that thermally
510 weakened the crust (e.g., Butler et al., 2006; Bellahsen et al., 2012). Indeed, in the External
511 Alps, the Alpine mylonites consist of brittle-ductile shear zones that affect various Variscan
512 rocks (migmatic gneisses, micaschists, granites ...) where quartz are both fractured and
513 ductily deformed (e.g., Leloup et al., 2005; Rolland et al., 2008; Bellahsen et al., 2012;
514 Bellanger et al., 2014). In the Pyrenean Axial Zone, the mylonites are also brittle-ductile but
515 the rock ductility is mainly due to the diffuse deformation of small-grain micas aggregates.
516 This suggests that periods of distributed shortening may be inherent to early times of
517 mountain building, predating the activity of the main thrusts.

518

519 6.4. Sericitization and rheological implications

520 White micas here have crystallized following destabilization of K-feldspar and plagioclase.
521 Such destabilization is commonly described in shear zones, even in the upper crust (e.g.,
522 Wibberley, 1999), where it reduces the strength and increases the ductility of deforming
523 rocks. Such syn-kinematic micas are observed in thin sections presented above (**Fig. 13c**).
524 Undeformed, sericitized samples show that a significant amount of micas likely formed
525 before a deformation, as suggested by sericitized, undeformed samples (**Fig. 13a**). The white
526 micas in these undeformed samples are in textural equilibrium with chlorites (**Fig. 14c**),
527 showing temperatures of about 250°C (**Fig. 16, Table 1**). Therefore, these micas formed most
528 likely after tectonic burial, at the thermal peak, before significant deformation.

529 The fluids responsible for the feldspar destabilization are probably of meteoric origin as is
530 usually the case in the upper crust (e.g., Sibson et al., 1975) and suggested for the area
531 (Henderson and McCaig, 1996). These surficial fluids may have circulated in the rocks
532 through Mesozoic (extensive) fractures (Johnson et al., 1996). The fluids may also be the
533 product of prograde dehydration processes but the protolith contained quite little hydrated
534 phases.

535 Having stressed that sericitization was pre-kinematic implies that the crust was weak since
536 the onset of shortening. Indeed, localization of deformation along interconnected layers of
537 micas reduces the bulk strength of the granodiorites to values close to those of pure mica
538 aggregates (Handy, 1994), that are much lower than those of quartz and feldspar (**Fig. 20**).
539 Given the geothermal gradient determined above, the “brittle” crust was probably about 7-8
540 km thick and its strength may have been controlled by the one of interconnected layers of
541 weak materials (e.g., Bos and Spiers, 2002). Considering a high geothermal gradient and the
542 associated flow law, the crustal strength in the Pyrenean Axial Zone during Cenozoic might
543 have been much lower than a “normal” crustal strength (**Fig. 20**). This weak crustal strength
544 is a likely cause for the distributed deformation that may have been active before the
545 localization of the shortening on crustal ramps (**Fig. 19**), although we still lack
546 geochronologic data to confirm it.

547

548 **Conclusions**

549 This contribution presents field data on the structural evolution of the Bielsa basement unit
550 together with new thermometric and thermochronologic data to constrain the thermal peak. In
551 the basement of the Bielsa unit, shortening is distributed. It consists of several short-
552 wavelength (<1 km) antiforms and synforms affecting the basement top. At outcrop scale,
553 shortening is accommodated by brittle and brittle-ductile shear zones active under greenschist
554 metamorphic conditions. These conditions are attested by temperatures around 250-300°C
555 given by thermometric data and RSCM data, consistent with ZFT ages. At microscale, the
556 protolith is characterized by hydrated phases (mostly phyllosilicates) that strongly influence
557 the mode of shortening at this scale: feldspars are sericitized even in undeformed samples.
558 Sericitization also increases with increasing strain attesting for syn-kinematic feldspar
559 destabilization. We conclude that the high thermal gradient, together with phyllosilicate

560 crystallization, probably induced crustal weakening that controlled the style and timing of
561 shortening.

562

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566

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859

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863

864 **Figure caption**

865

866 Figure 1: Simplified structural map of the Pyrenees. The Axial Zone is located between the
867 North Pyrenean Zone (NPZ) and the South Pyrenean Zone (SPZ). The two main units in the
868 western Axial Zone are the Gavarnie and Bielsa units. The Guarga unit does not crop out and
869 is located below the Bielsa unit and the Jaca basin. The line represents the trace of the cross-
870 section presented in Fig. 2, the box the map in Fig. 3.

871

872 Figure 2: Regional cross-section from Jolivet et al. (2007). Four main units are represented
873 in the Axial Zone: the Lakora, the Gavarnie, the Bielsa, and the Guarga units, from north to
874 south, respectively. Shortening in the cover of the South Pyrenean Zone is linked to basement
875 thrusts in the Axial Zone or underneath the Jaca basin.

876

877 Figure 3: Geological map of the Bielsa unit after Casas et al. (2003). The Bielsa unit is
878 structurally below the Millares unit. On Fig. 2, Bielsa and Millares units are grouped under
879 Bielsa unit name. The Bielsa unit is composed of a Variscan Carboniferous
880 granite/granodiorite intruding the Paleozoic cover. The Triassic layers unconformably overly
881 the basement. Several WNW-ESE to ENE-WSW antiforms and synforms affect the cover and
882 the basement top. Two klippen of Millares unit (Fobillons and Punta Suelza) are cropping out
883 above the folded cover of Bielsa unit. The line represents the trace of the cross-section
884 presented in Fig. 4, the box the map in Fig. 5.

885

886 Figure 4: Cross-section of the Bielsa unit and its restoration. The shortening is estimated
887 considering constant length for Triassic and non-reactivated inherited faults. Nine antiforms
888 are mapped along with 6 steep faults. The restoration is done assuming inherited steep faults
889 (see text for explanation and discussion). However, the age of these faults is poorly
890 constrained as attested by the question marks: the Cretaceous layers are drawn syn-tectonic
891 although it is not necessarily the case. Same legend than Fig. 3.

892

893 Figure 5: Structural map of the Lago Urdiceto area. Three steep structures are mapped along
894 the southern limb of the Urdiceto, the Western, and the Punta Suelza synforms. Small scale
895 shear zones are also recognized, as well as an Alpine schistosity/cleavage (in both the
896 basement and cover) mainly steeply north-dipping. These structures and striated faults are
897 consistent with a NNE-SSW contraction. Sites of measurements and sampling are indicated.

898

899 Figure 6: Cross-sections through Urdiceto synform. See Fig. 5 for location. The synform is
900 bounded northward by a steep normal fault. It is asymmetrical especially in its eastern part,
901 with an overhanging sheared flank. The apparent folding of the top basement is associated to
902 distributed deformation within the basement (e.g., shear zones).

903

904 Figure 7: Field photographs. a) Punta Suelza klippen over the Bielsa unit with folded
905 Triassic layers suggesting a folding during the nappe emplacement. b) The Urdiceto synform
906 (see Fig. 5 and 6). c) The Western synform with the syn-S1 folding.

907

908 Figure 8: Field photographs of the Urdiceto synform basement. a) Altered top basement with
909 penetrative schistosity. b) Zoom on an unaltered granodiorite with less penetrative schistosity.

910

911 Figure 9: Field photographs of reverse top-to-the-South shear zones in the basement.

912

913 Figure 10: Field photographs of the Urdiceto synform cover. a) Basement/Triassic
914 unconformity with bedding-parallel top-to-the-South shearing. b) Sub-vertical cleavage and
915 sub-horizontal veins attesting for horizontal shortening and vertical stretching. c) Fold with
916 steep axial surfaces and associated cleavage.

917

918 Figure 11: Northern contact (steep fault) of the Western synform (see Fig. 5 for location). a)
919 Steep (reverse, South verging) fault and steep cleavage oblique to bedding in the cover. b)
920 Zoom of the cover deformation with locally cleavage parallel to bedding and small-scale
921 reverse faults (South verging).

922

923 Figure 12: Thin-sections photographs. The rocks are composed of quartz, phengite,
924 feldspars, chlorites and some contain calcite a) Polarized (sample B11-33). b) Interpretation
925 (sample B11-33, Fig. 5). The rock is affected by cataclastic deformation. c) Ductile
926 deformation with mylonitic shear zones (sample B11-31, Fig. 5). This foliated rock is

927 composed of fractured quartz, deformed chloritized biotites, calcite, and micrometric white
928 micas.

929

930 Figure 13: Photographs of thin sections. a) Undeformed (except a few fractures) granite with
931 quartz, chlorite (dark crystals), and micas partly replacing feldspars (pseudomorphs) (sample
932 B11-33, Fig. 5). b) protomylonite where quartz are embedded in aggregates of micas and
933 chlorites (dark phases) with the development of a weak schistosity (sample B11-18, Fig. 5). c)
934 Mylonite with fractured quartz, a schistosity (S1) mainly composed of micas, pressure
935 shadows of chlorite/mica around quartz, and formation of a few micro shear zones (sample
936 B11-31, Fig. 5). Between a) and c) the amount of phyllosilicates, in particular micas, strongly
937 increases.

938

939 Figure 14: SEM images of the progressive sericitization. a) Albitisation of K feldspars in a
940 weakly deformed sample (sample B10-31). b) Albitisation and phengitisation of K feldspars,
941 along with titanite and epidotes inclusions in weakly deformed samples (sample B11-33). c)
942 Biotites replaced by micas and chlorites (bright spots: apatites) in an undeformed sample
943 (Quartz and Feldspars) (sample B11-37). d) Quartz embedded in an Albite/micas “matrix”
944 (sample B10-31). e) Images of a phyllonite composed almost exclusively of micas (with
945 quartz out of the image) (sample B11-14).

946

947 Figure 15: Results of thermometric study. For 4 samples, the temperature is provided with 2
948 methods. The multi-equilibrium thermo-barometer on Quartz Micas is from Dubacq et al.
949 (2010). The multi-equilibrium thermo-barometer on chlorite is from Vidal et al. (2006).

950

951 Figure 16: Zircon fission-track data and location of samples. Plots are realized with Radial
952 Plotter (Vermeesch, 2009).

953

954 Figure 17: Speculative thermal structure of the Iberian lithosphere. In the southern part, the
955 thermal structure is considered as normal given its age and the Moho temperature set at

956 slightly more than 500°C (lithosphere with a thermotectonique age of about 200 Myrs). In the
957 northern part, the thermal structure is built using the gradient of 80 °C/km from Vacherat et
958 al. (2016) in the North Pyrenean Zone. Between these 2 zones, i.e., in the future Axial Zone,
959 the geothermal gradient is intermediate between the northern and the southern one, and fits
960 the gradients of about 50°C/km estimated in this study for the northern axial zone.

961

962 Figure 18: Thermal histories for a) Bielsa and b) Neouvielle. See text for explanations.

963

964 Figure 19: Present day cross-section of the Pyrenees and possible ages of deformation. The
965 deep structure is from Chevrot et al. (2018). The cover structure in the foreland basin is from
966 Jolivet et al. (2005). For a detailed balanced cross-section and its restored stage see Teixell et
967 al. (2016) and Espurt et al. (this issue). Bielsa thrust and internal Bielsa deformation ages are
968 deduced from discussion on the timing of the other structures (see text).

969

970 Figure 20: Rheological laws for both typical continental basement (quartz) and weakened
971 (hot) continental crust. The weakened strength of the upper crust is from Bos and Spiers
972 (2002) for strain rates of both 10^{-10} and 10^{-12} s^{-1} , as the result of sericitized feldspars. The
973 associated lower crust strength is considered for a rather high inherited geothermal gradient of
974 45 °C/km.

975

976 Table 1: Maximal temperatures obtained from RSCM method.

977

978 Table 2: Zircon Fission Track dataset from Bielsa and Neouvielle plutonic massifs.

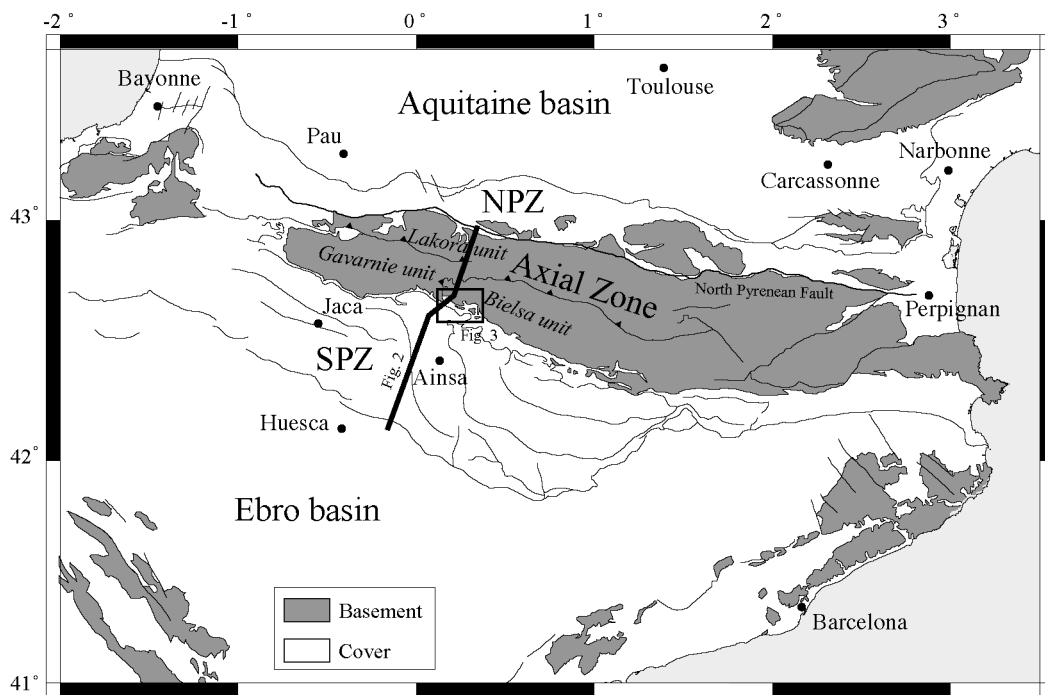


Figure 1

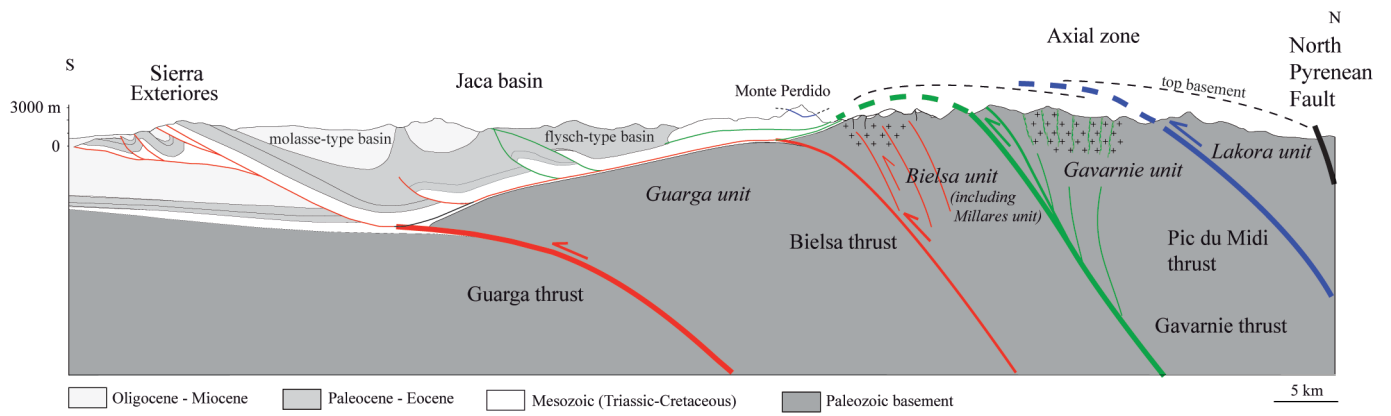


Figure 2

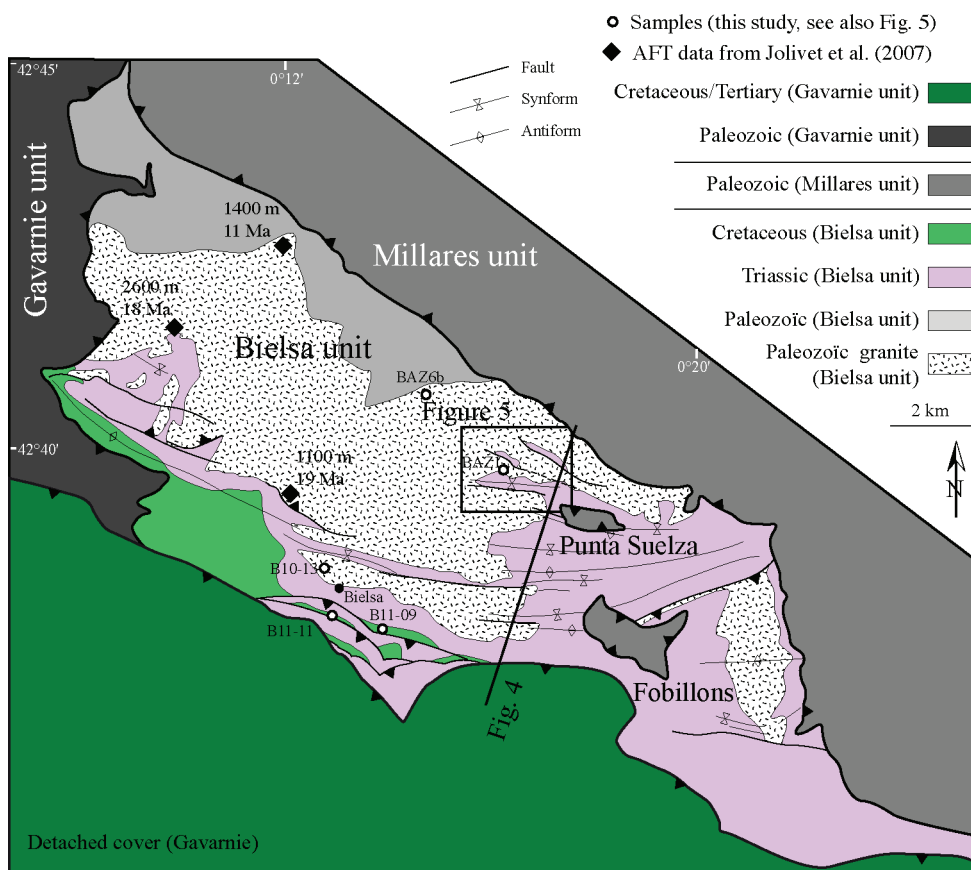


Figure 3

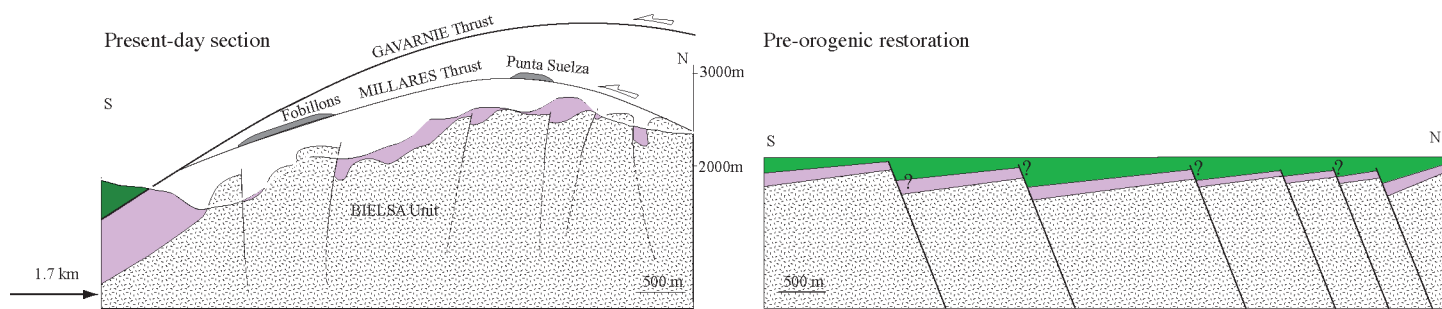


Figure 4

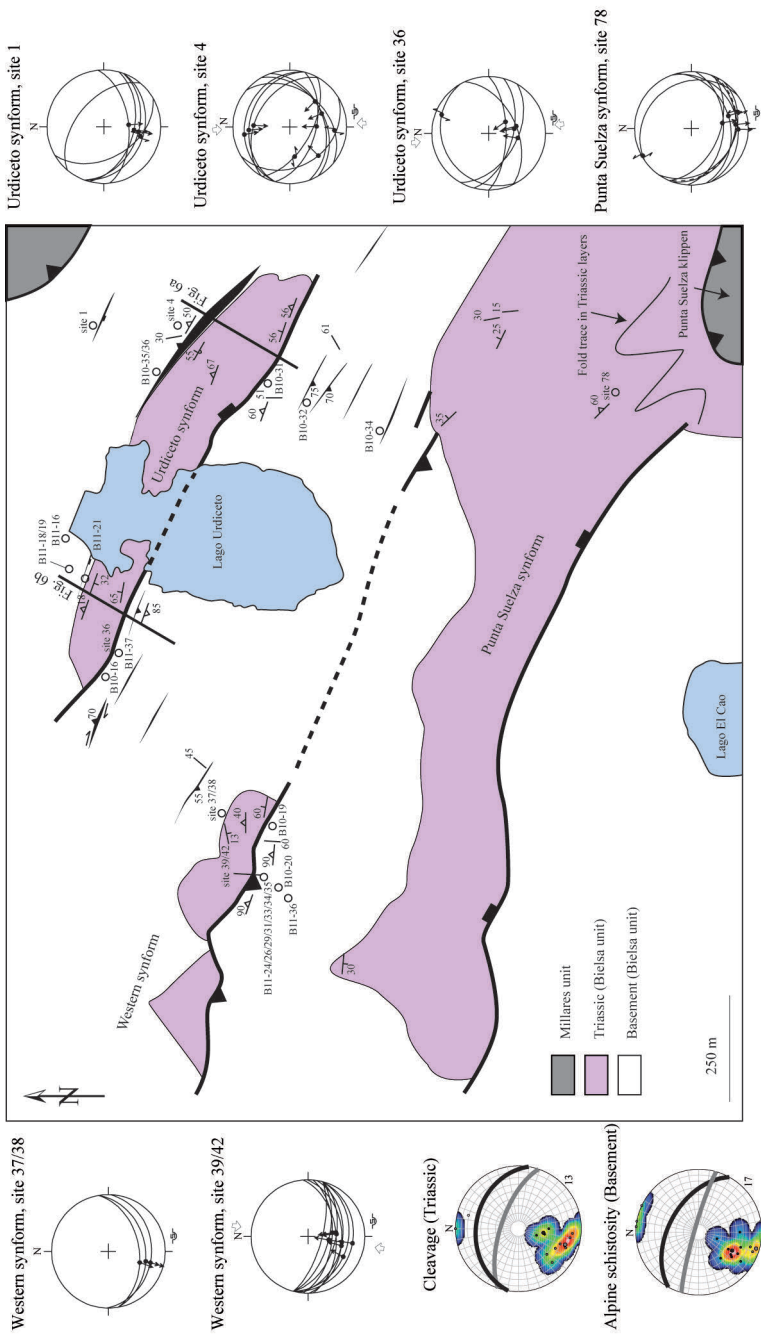


Figure 5

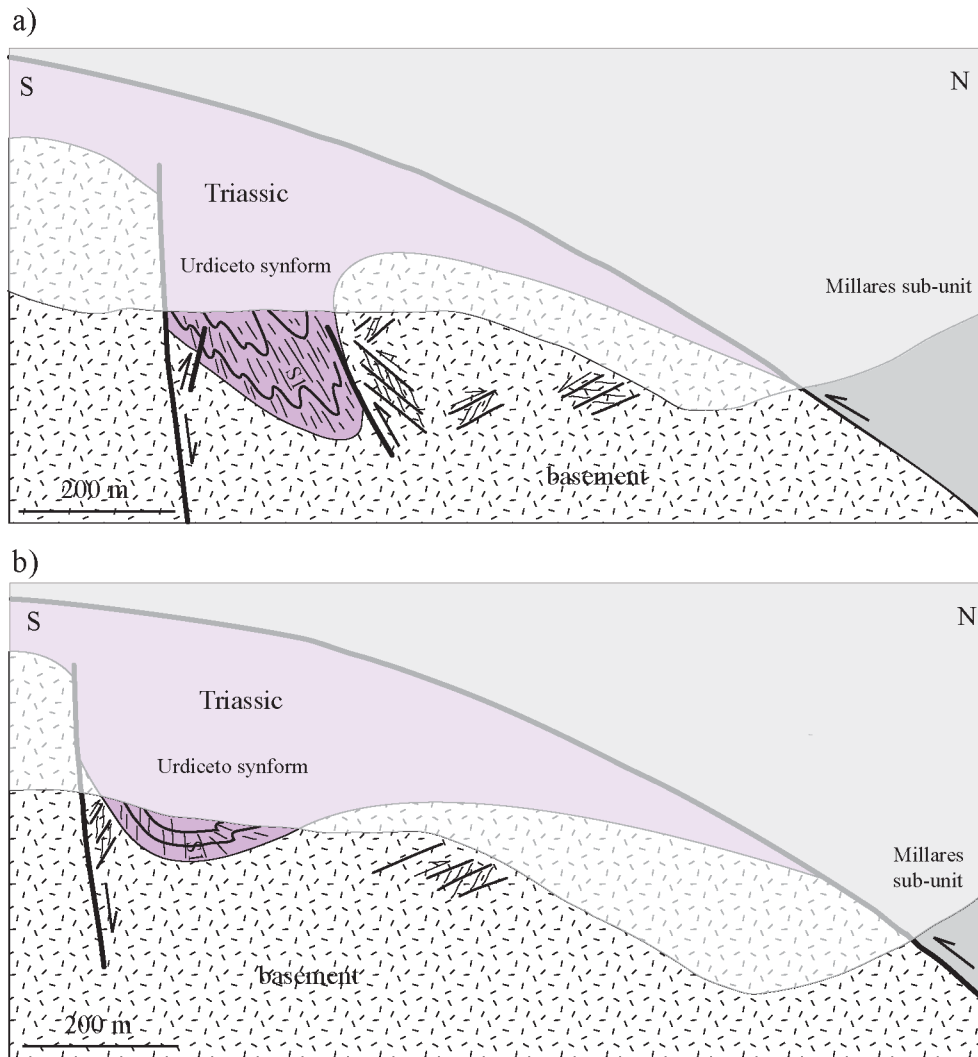


Figure 6

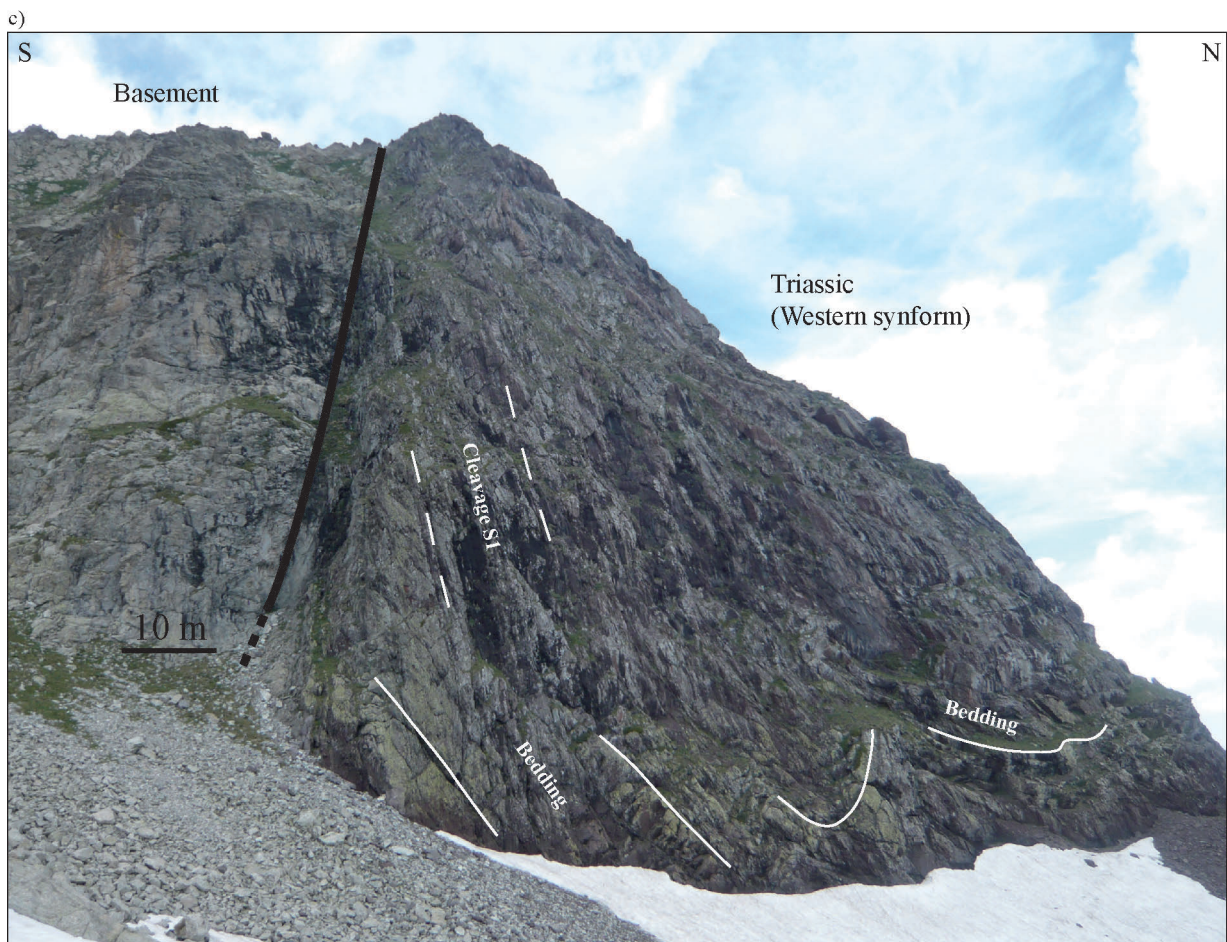
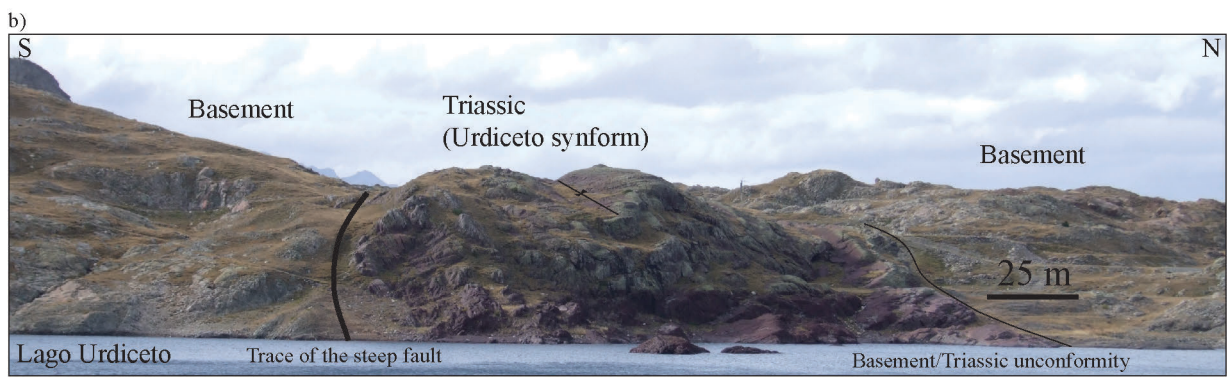


Figure 7

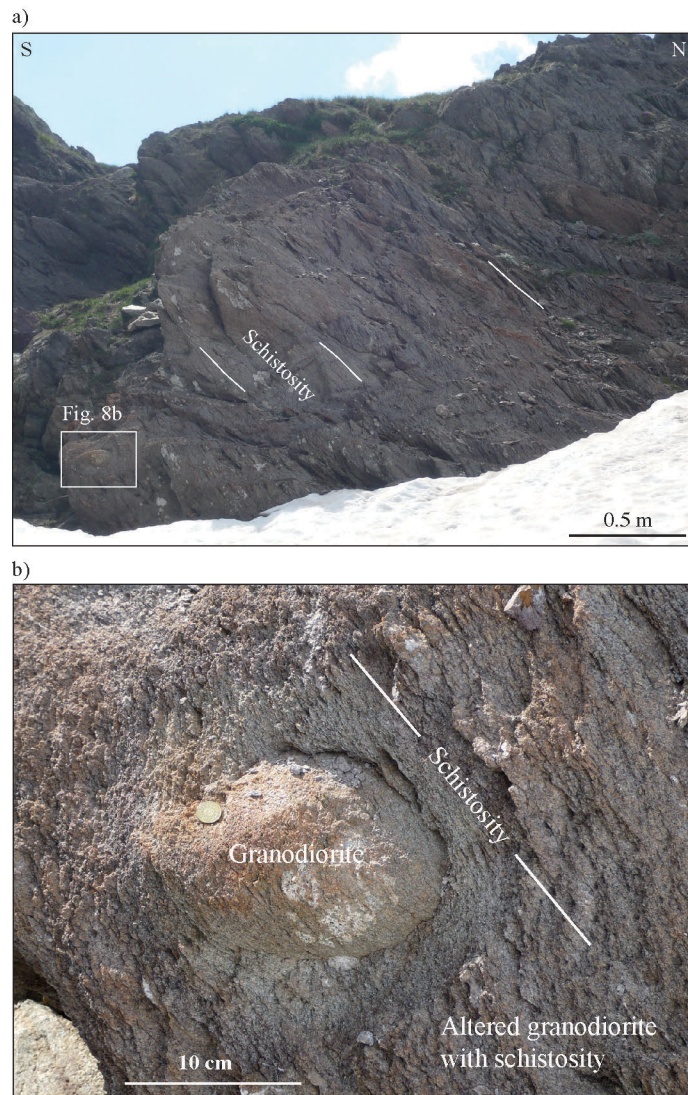


Figure 8



Figure 9

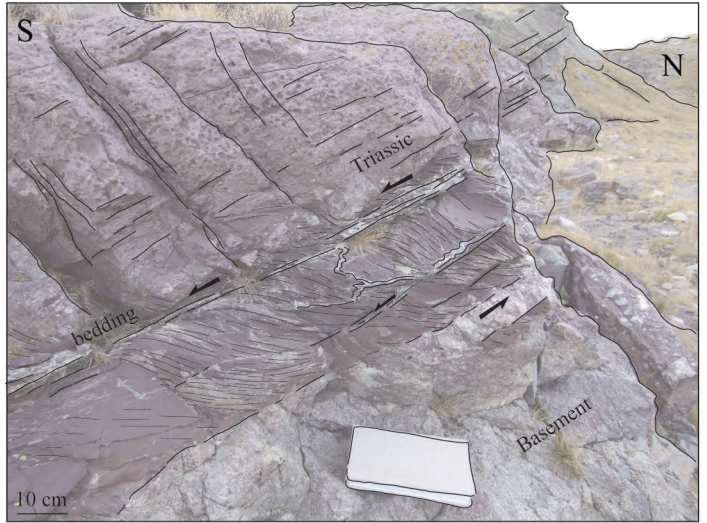
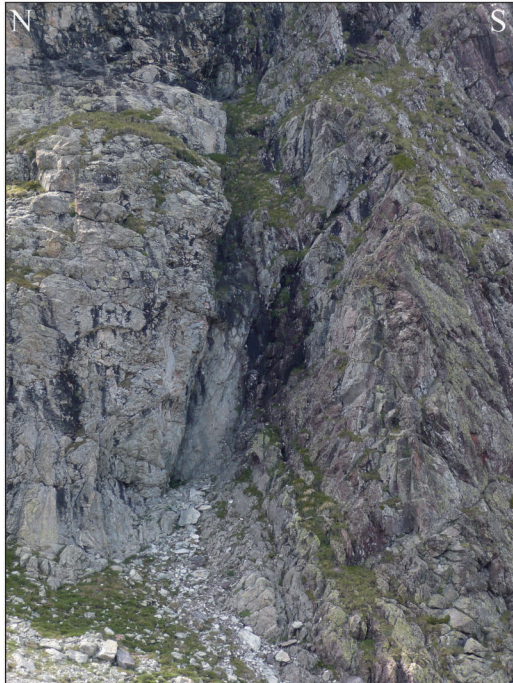


Figure 10

a)



b)

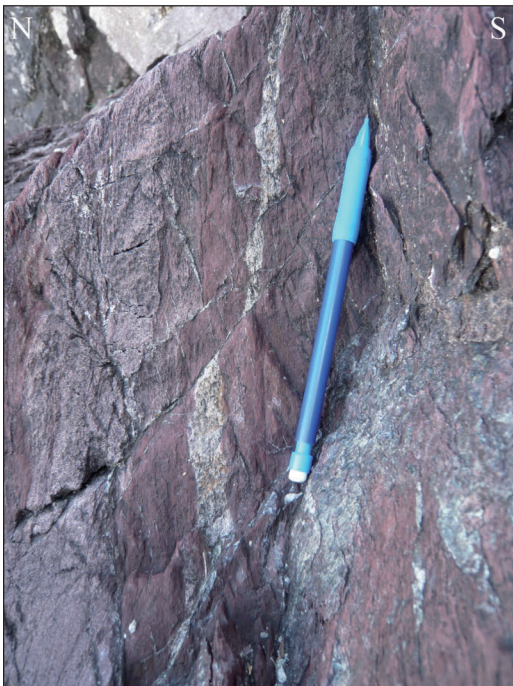


Figure 11

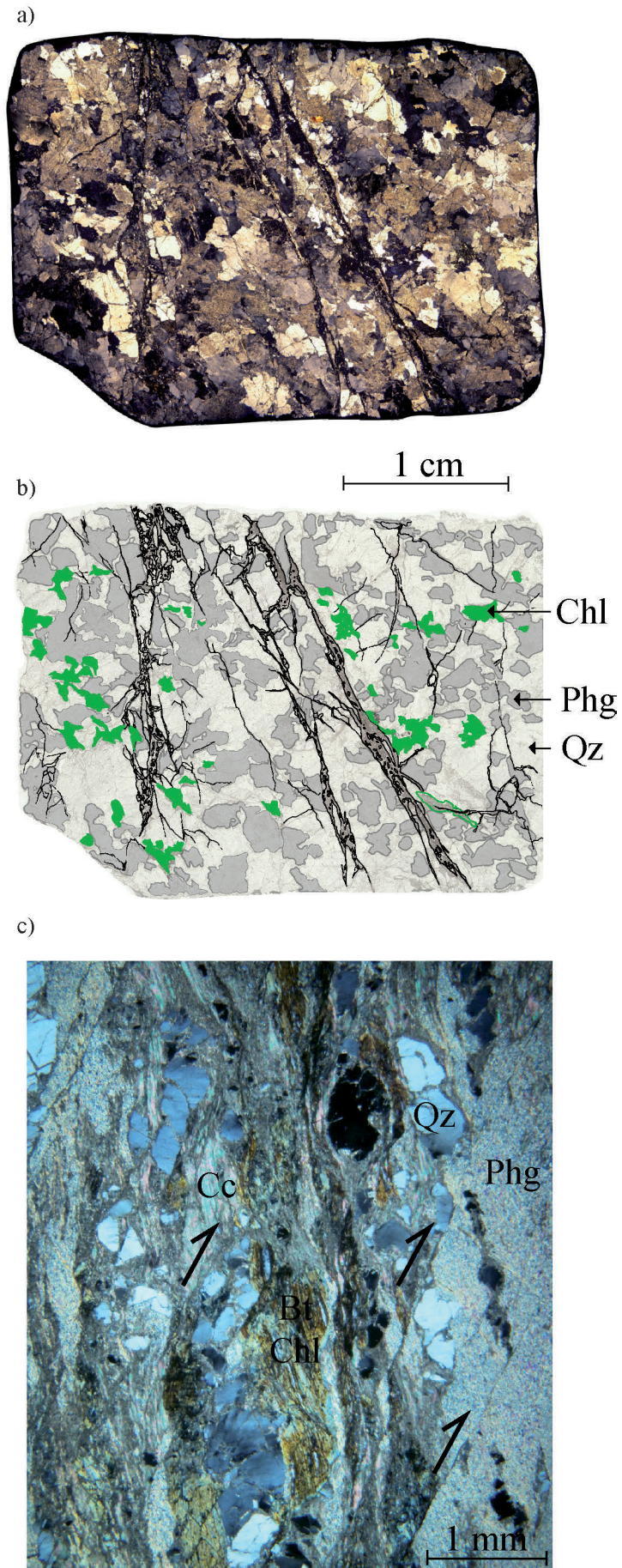
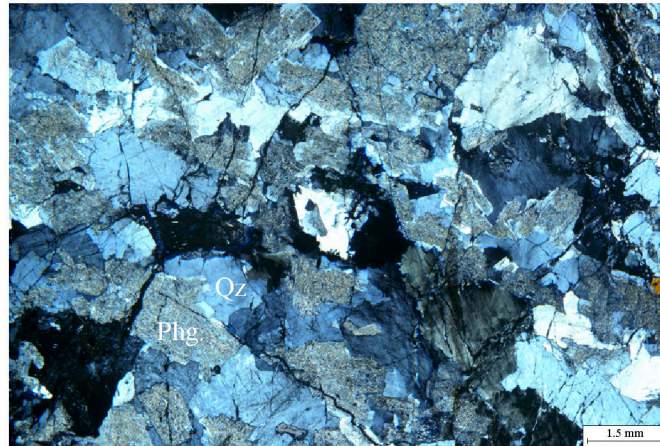
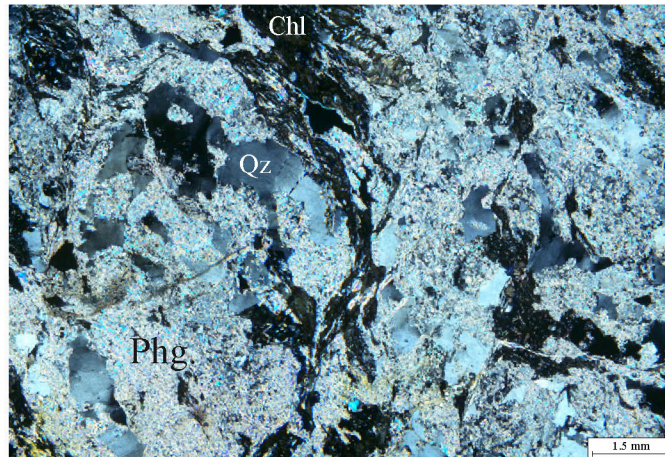


Figure 12

a)



b)



c)

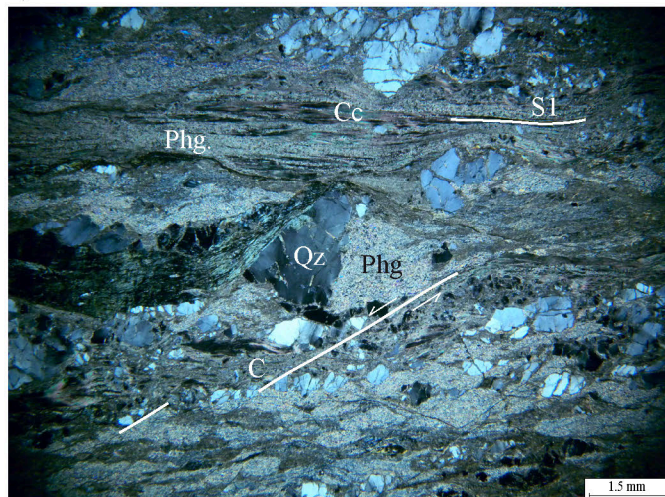


Figure 13

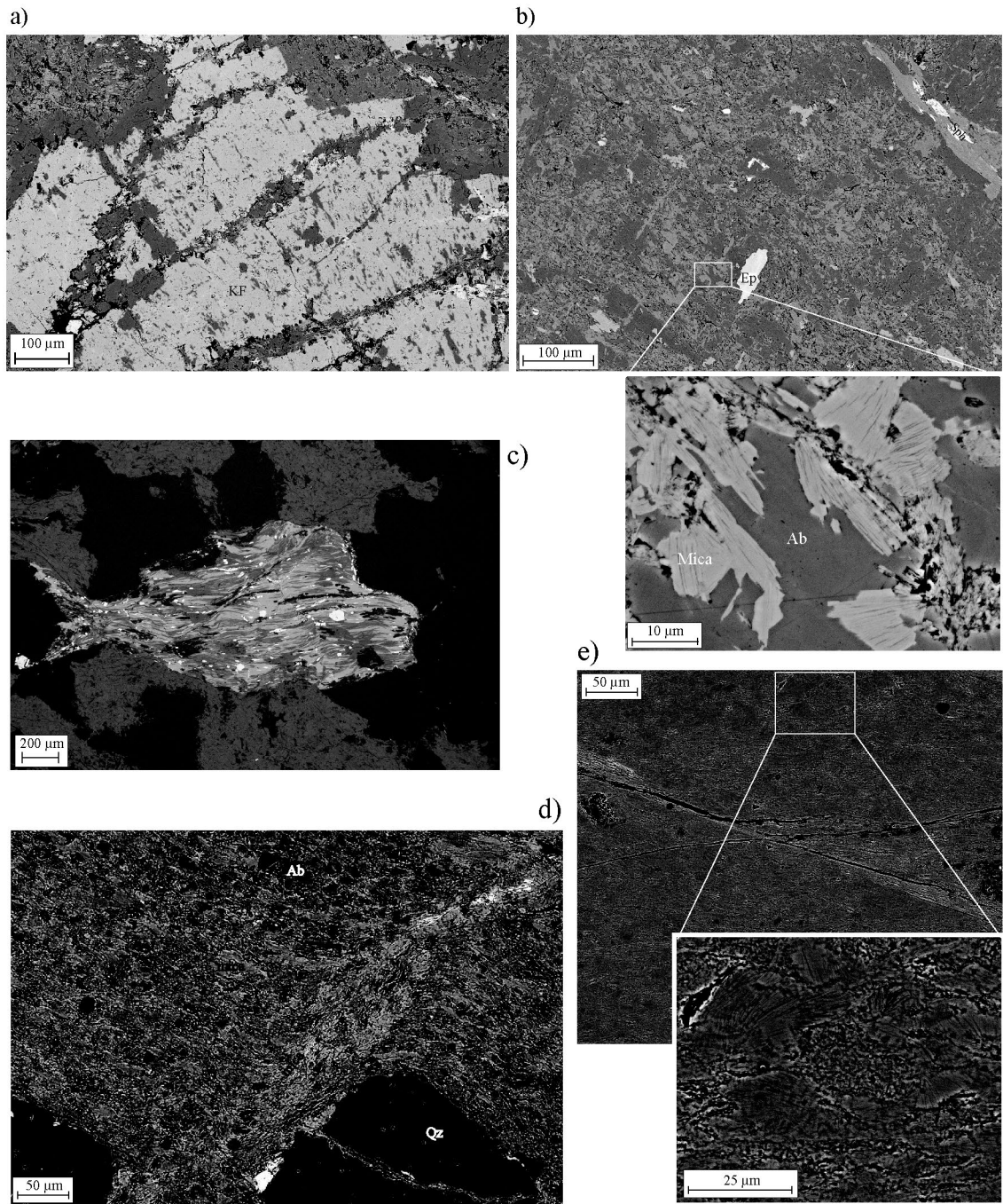


Figure 14

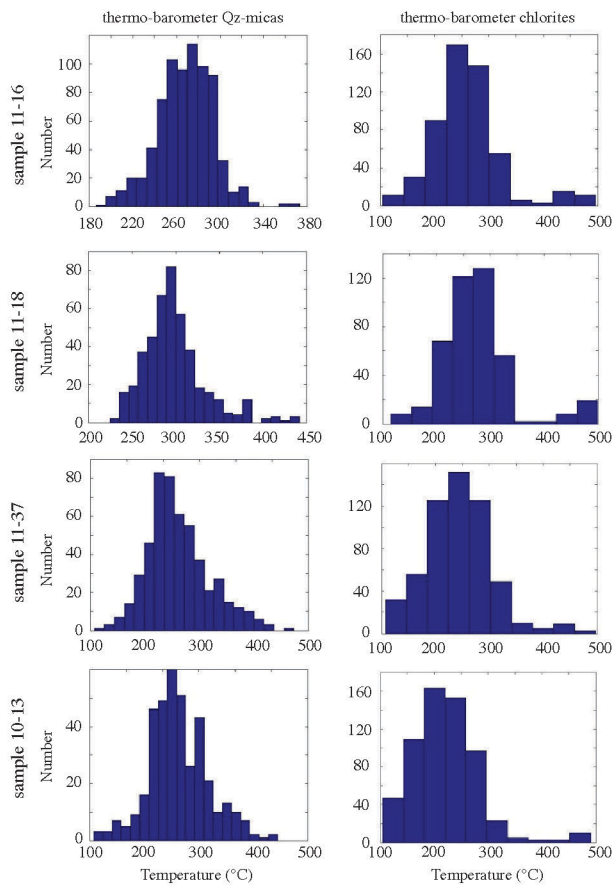


Figure 15

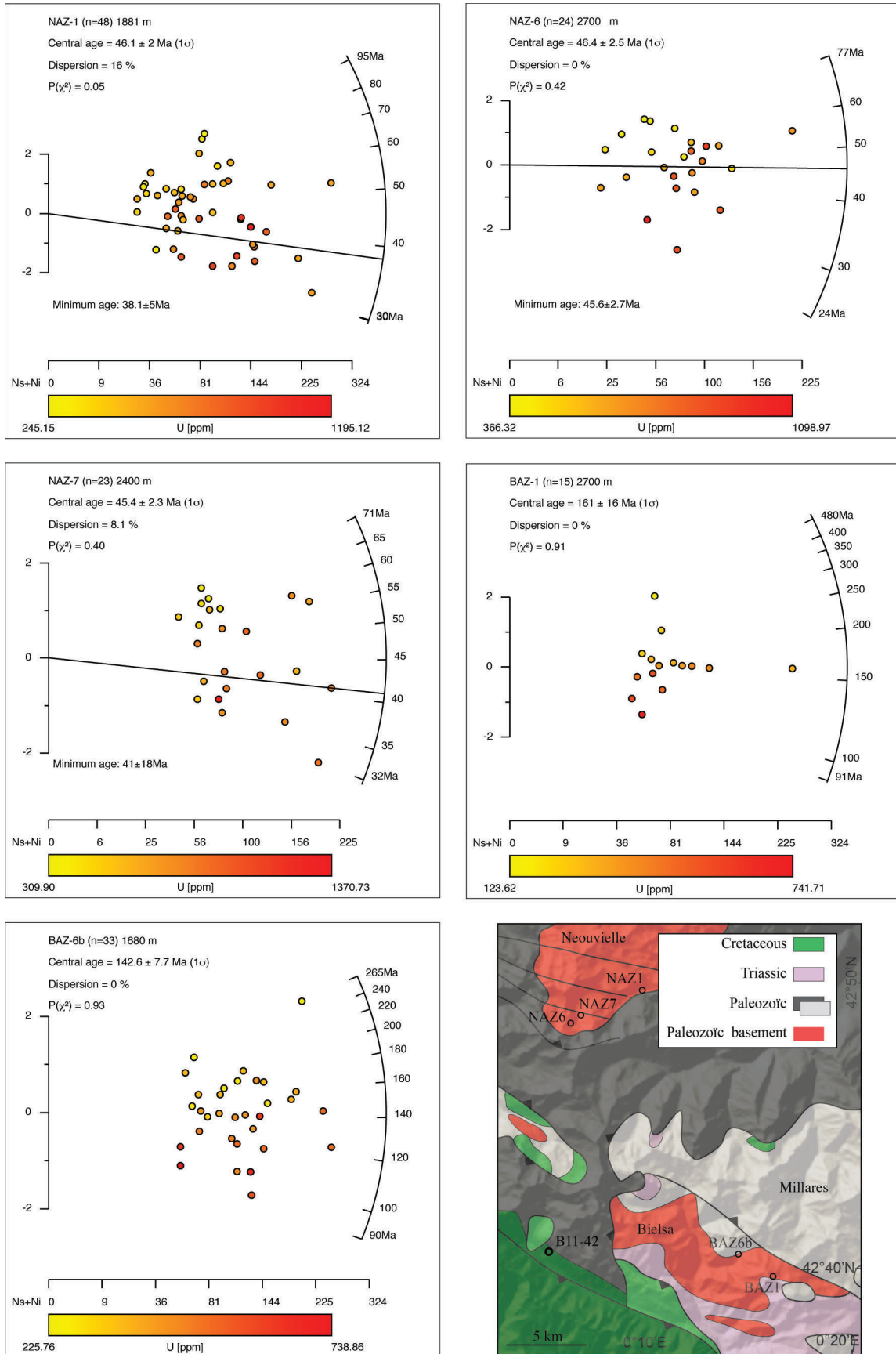


Figure 16

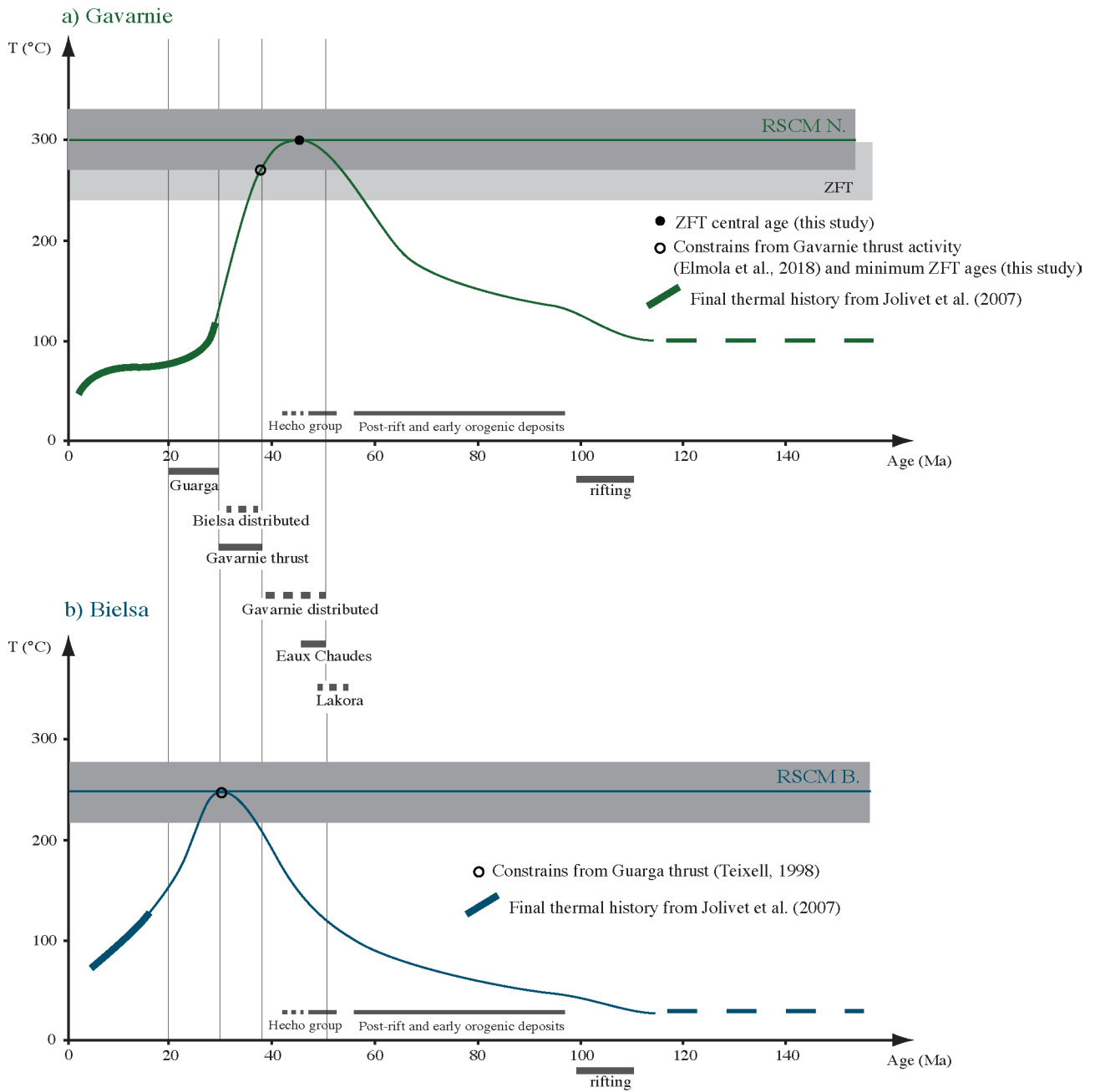


Figure 17

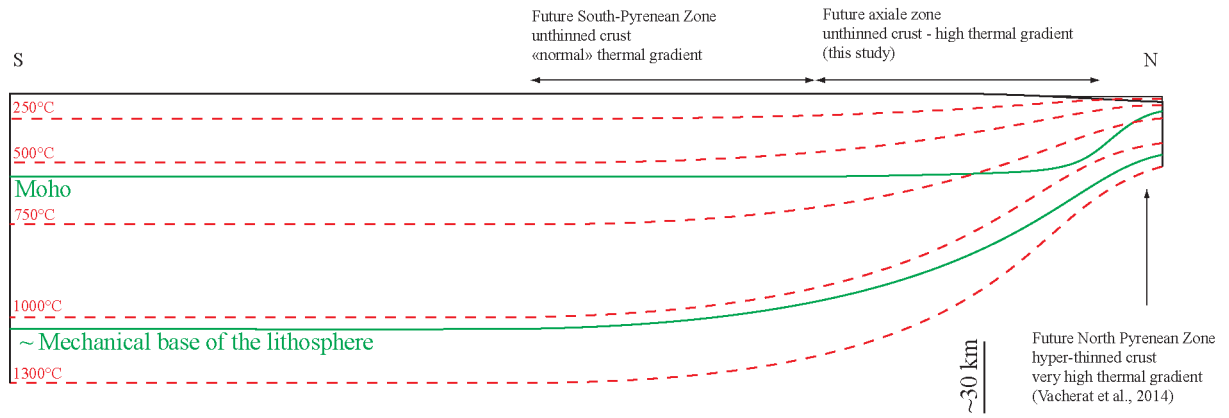


Figure 18

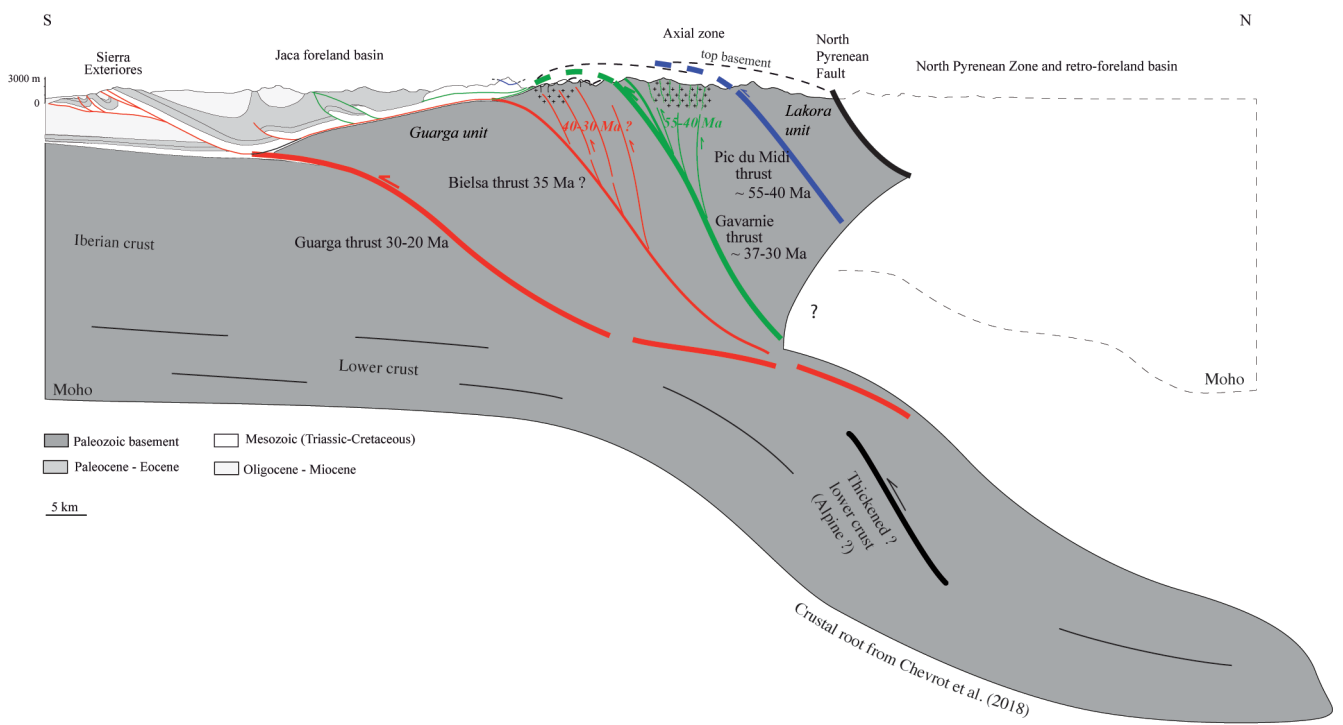


Figure 19

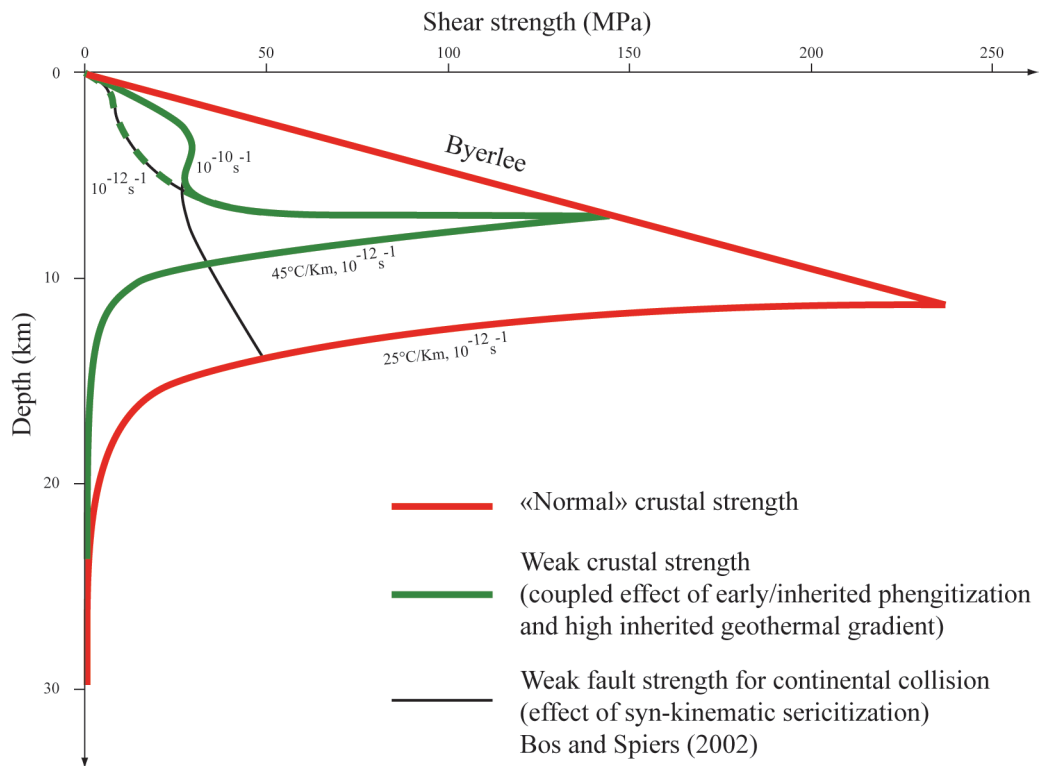


Figure 20

Sample	Raman parameter RA1 (Lahfid et al., 2010)		RSCM Temperature (°C)	
	Average	SD	average	SD
B11-9	0.60340626	0.016081064	264.0	19.6
B11-11	0.57516455	0.017642586	236.7	16.1
B11-42	0.65579392	0.005169598	327.8	6.3

Table 1: TMax obtained from RSCM method.

Sample n°	Elevation (m)	Rho-S x10 ⁶ (Ns)	Rho-I x10 ⁶ (Ni)	Rho-D x10 ⁶ (Nd)	Grains	Dispersion %	P(X ²) %	Central age Ma	Central age +/-	Minimum age Ma	Minimum age +/-
BAZ-1	2528	12.6 (1161)	1.2 (111)	0.220 (3112)	15	0	91	161	16	155	16
BAZ-6B	1532	12.7 (3541)	1.37 (382)	0.220 (3114)	33	0	93	142.6	7.7	138.8	8.1
NAZ-1	1864	5.95 (3348)	2.03 (1142)	0.219 (3116)	48	16	5	46.1	2	38.1	5
NAZ-6	2707	6.3 (1406)	2.1 (469)	0.219 (3118)	24	0	42	46.4	2.5	45.6	2.7
NAZ-7	2423	7.2 (1718)	2.4 (587)	0.219 (3120)	23	8.1	40	45.4	2.3	41	18

Zeta value 141.69 +/- 4.66

Table 2. Zircon Fission Track dataset from Bielsa and Neouvielle plutonic massifs