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1 **Evolution of salt structures of the Pyrenean rift (Chaînons Béarnais,**
2 **France): From hyper-extension to tectonic inversion**

3

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5

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11

12 ABSTRACT

13

14 The Chaînons Béarnais is a salt-detached fold belt in the northern Pyrenees that formerly
15 occupied the axis of the Cretaceous Pyrenean rift. Geological map revision and cross-section
16 construction from surface geology and industrial well and seismic reflection data emphasize
17 the role of salt diapirism in the folding of the belt during the Cretaceous extension and the
18 subsequent Pyrenean orogeny. Pre-rift Triassic evaporites played a fundamental role during
19 rifting, allowing the sedimentary basin lying above to detach and slide down the hyper-
20 extended margins onto a central exhumed mantle tract. Since the Early Cretaceous (and
21 locally probably since the Jurassic) a system of low-amplitude salt walls evolved in shallow
22 marine environments punctuated by episodic emersion. During the main stage of crustal
23 extension, in late Aptian to early Cenomanian times, carbonate shelves rapidly drowned giving

24 rise to deeper marine sedimentation. This was a period of major rise of salt walls, progressively
25 detached from their substratum. These salt walls enclosed minibasins that accumulated thick
26 flysch deposits arranged in growth stratal patterns. Depocenter migration and foundering of
27 previous diapiric highs controlled further flysch deposition during the Late Cretaceous, while
28 moderate extension probably persisted until the onset of the Pyrenean compression. During
29 the late Santonian to Paleogene Pyrenean orogeny, the sedimentary lid of the Chaînons
30 Béarnais basin climbed back along the Triassic detachment onto the colliding continental
31 margins, leading to salt wall squeezing and further rising. Based on the Cretaceous timing and
32 style of growth folding, we suggest that salt wall squeezing was not solely related to the
33 Pyrenean compression, but shortening affected the diapiric ridges during the syn-rift sliding
34 by gravity and crowding in the basin center, as the rifted margins were pulled apart from
35 beneath. This makes the Chaînons Béarnais belt a unique field analog for contractional salt-
36 tectonic systems in distal continental margins.

37

38 *Key words:*

39 Salt tectonics

40 Minibasin

41 Continental margin

42 Hyper-extension

43 Tectonic inversion

44 Pyrenees

45

46 **1. Introduction**

47

48 Rifted continental margins with salt diapirism and large-scale gravitational sliding above salt
49 units are commonly reported in modern settings (e.g. Rowan et al., 2004; Brun and Fort, 2011;
50 Jackson and Hudec, 2017). Abundant seismic surveys have revealed the essential structural
51 features on the salt-dominated margins, with both extensional and contractional structures
52 above common salt detachments. However, these slid structures are located in distal offshore
53 domains of modern margins, and salt detachment systems of ancient continental margins
54 observable in outcrop after tectonic inversion are rarely reported. Recent works in the former
55 peri-Iberian margins give valuable evidence of salt detachment associated to crustal hyper-
56 extension and mantle exhumation in ancient, mid-Cretaceous rifts, containing salt-related
57 extensional structures inverted during the Upper Cretaceous-Tertiary. Such features are now
58 exposed in the northern Pyrenean belt (Jammes et al., 2009, 2010; Lagabrielle et al., 2010;
59 2019a and b, 2020; Teixell et al., 2016, 2018), the Basco-Cantabrian belt (Pedrera et al., 2017;
60 Ducoux et al., 2019), or imaged by subsurface data in the offshore Parentis (Jammes et al.,
61 2010; Ferrer et al., 2012) and Columbrets (Etheve et al., 2018) basins. A particular aspect of
62 these peri-Iberian detachments is that they are located in the Middle-Upper Triassic evaporite
63 facies, i.e. a salt-bearing layer which can be considered as pre-rift with respect to the main,
64 Cretaceous-age extension. Hence, the salt-related structures potentially recorded the whole
65 rift and post-rift history, in contrast with those modern margins where detachment occurred
66 on post-rift salt (see references above) and is thus largely independent on the extensional
67 structures in the underlying crust. In the Pyrenean mountain belt, the Triassic evaporite layer
68 is commonly present in the fold and thrust belts of both the North and South-Pyrenean Zones
69 (Fig. 1). This layer is well known to have acted as the main décollement layer for the orogenic
70 thrust sheets (Séguret, 1972), and for the Mesozoic extensional structures (Jammes et al.
71 2009, 2010; Lagabrielle et al., 2010, 2019a and b, 2020; Teixell et al., 2016, 2018). It also

72 promoted diapirism both during the rifting (Canérot, 1988, 1989; James and Canérot, 1999;
73 Canérot et al., 2005; López-Mir et al., 2014; Saura et al., 2016; García-Senz et al., 2019) and
74 the Pyrenean compression (Muñoz et al., 2013; Santolaria et al., 2014). However, the detailed
75 interactions between the salt detachment and diapirism remain little discussed with the
76 exception of Jammes et al. (2010).

77 In this paper, we describe diapiric structures originated in the Cretaceous margins of
78 the west-central Pyrenean rift and inverted by the Pyrenean orogeny in the so-called Chaînons
79 Béarnais belt (Fig. 2). Moderate contractional inversion uplifted and exhumed the extensional
80 system without destroying its major features. The diapiric origin of the Chaînons Béarnais folds
81 and thrusts is put forward in previous works (Canérot, 1988, 1989; James et Canérot, 1999;
82 Canérot et al., 2005) but only local small structures have been so far described with detail
83 (Lenoble and Canérot, 1992; Canérot and Lenoble, 1993). The present study is based on
84 original field investigations complemented by well log data and a new interpretation of a
85 seismic reflection profile for the northern part of the area. The results of the study are
86 illustrated by four new balanced cross-sections of representative transects of the Chaînons
87 Béarnais belt, detailed mapping of key sectors, and a sequential restoration of the tectono-
88 sedimentary evolution during the rift opening and subsequent inversion. We propose that the
89 fold-thrust deformation observed in the pre- and syn-rift strata is not only related to the
90 Pyrenean shortening, but also was largely achieved through salt tectonics, involving diapirism
91 and possibly gravity-induced contraction, during the Mesozoic rifting. We discuss the
92 differences between this interpretation and previous works that reported the salt structures
93 of the western Pyrenees as autochthonous diapirs presently above basement normal faults
94 relict from the former continental margins (Canérot, 1988, 1989; James and Canérot, 1999;
95 Canérot et al., 2005).

96

97 **2. Geological setting**

98

99 *2.1. Tectonic framework of the Pyrenean mountain belt*

100 After the Variscan orogeny that shaped the Paleozoic basement of the Pyrenees
101 (Cochelin et al., 2017, and references therein), different episodes of rifting took place during
102 the Permian, the Triassic and the Cretaceous. While Triassic extension has been reported, to
103 which thick evaporites were associated (Biteau et al., 2006; Ortí et al., 2017; Soto et al., 2017),
104 the main rifting stage in the Pyrenean realm occurred in the early and mid-Cretaceous, leading
105 to the main basin opening episode precursor to the inversion and mountain building. The mid-
106 Cretaceous rifting in the Pyrenean realm was related to the opening of the Bay of Biscay
107 oceanic basin located more to the west and involved a major episode of crustal stretching,
108 basin subsidence and mantle exhumation during the Albian-early Cenomanian (Lagabrielle
109 and Bodinier, 2008; Jammes et al., 2009; Lagabrielle et al., 2010; Mouthereau et al., 2014;
110 Tugend et al., 2014; Teixell et al., 2016, 2018; Grool et al., 2018; Espurt et al., 2019). Most of
111 the inverted rift system now corresponds to the so-called North-Pyrenean Zone (NPZ; Fig. 1),
112 where rifting-related features comprise 1) syn-rift Albian-lower Cenomanian flysch basins (the
113 Black Flysch Group; Souquet et al., 1985), 2) peridotite bodies (the Iherzolites from their type
114 area in the central Pyrenees) originating from the exhumed mantle and included in the
115 Mesozoic series (Lagabrielle and Bodinier, 2008; Jammes et al., 2009; Lagabrielle et al. 2010;
116 2019a and b; Corre et al., 2016), 3) alkaline magmatic bodies (Montigny et al., 1986), and 4) a
117 HT-LP metamorphism of the Mesozoic series, dated from the Albian to the Coniacian (for
118 magmatism and metamorphism, see a review and references in Clerc et al., 2015). An
119 unconformity in the lower Cenomanian coinciding with a widening of the basin is often

120 considered as marking the transition to a post-rift stage (e.g. Debroas, 1987, 1990). However,
121 there is no consensus in the literature on the kinematics of the North-Pyrenean rifting, which
122 has been interpreted either in a sinistral strike-slip (Choukroune and Mattauer, 1978; Debroas,
123 1987, 1990; Canérot, 2017) or in a near orthogonal extension context (Jammes et al., 2009;
124 Masini et al., 2014; Tugend et al., 2014; Saspiturry et al., 2019), due to the lack of robust
125 geological evidence and a controversial plate kinematic framework (Bronner et al., 2011,
126 Barnett-Moore et al., 2016; Nirrengarten et al., 2018, and references therein).

127 The Pyrenean orogenic shortening spanned from the late Santonian (circa 84 Ma) to
128 the early Miocene (circa 20 Ma). First it involved the tectonic inversion of the rift system and
129 subduction of the exhumed mantle tract, and then the formation of the south-vergent crustal
130 accretionary prism corresponding to the Axial and South-Pyrenean Zones (Fig. 1), formed after
131 the Iberian upper continental margin in relation to the northward subduction of the Iberian
132 lower crust and lithospheric mantle (Jammes et al., 2009; Mouthereau et al., 2014; Tugend et
133 al., 2014; Teixell et al., 2016, 2018; Grool et al., 2018; Espurt et al., 2019). Recent
134 interpretations depict the inversion of the rift system as a pop-up structure where the North-
135 Pyrenean Fault and Lakora thrust correspond to the south-vergent southern edge of the pop-
136 up, subsequently verticalized during stacking of thrust units in the Axial Zone, while the North-
137 Pyrenean Frontal Thrust (NPFT) corresponds to the north-vergent northern edge over the
138 Aquitaine foreland basin (Lagabrielle et al., 2010; Mouthereau et al., 2014; Teixell et al., 2016,
139 2018; Espurt et al., 2019) (Fig. 1).

140

141 *2.2. The North-Pyrenean Zone in the Chaînons Béarnais area*

142

143 *2.2.1. General structure*

144 The Chaînons Béarnais belt (CBB) corresponds to a 50 km-long segment of the
145 west-central NPZ, constituted by a system of E-W trending Jurassic to Aptian carbonate
146 anticlinal or thrust ridges cored by Middle-Upper Triassic rocks (mainly Keuper facies shales at
147 the surface), and separated by synclines with thick uppermost Aptian to lower Cenomanian
148 marl and flysch deposits (Fig. 2). To the west and east, the CBB ridges plunge laterally below
149 the Albian-Cenomanian flysch of the so-called Mauléon and Ossun basins, respectively.

150 To the north, the CBB is bounded by the North-Pyrenean Frontal Thrust (NPFT).
151 The Rébénacq transverse fault zone (Fig. 2) separates an eastern segment where the NPFT is
152 emergent at the surface, and a western segment where this thrust is a blind structure marked
153 at the surface by the km-scale Oloron anticline. North of the NPFT, the Aquitaine foreland
154 basin comprises thick depocenters of Upper Cretaceous flysch sequences and Tertiary strata
155 covered by post-orogenic Upper Miocene to Quaternary molasse.

156 To the south, the CBB lies above a system of small thrust sheets, here named the
157 Bedous-Laruns thrust units (BLTU), forming the northern edge of the Axial Zone and mainly
158 comprising Upper Paleozoic strata (Silurian to Carboniferous) with a Lower Triassic sandstone
159 tegument (Canérot et al., 2001). Local remnants of Middle-Upper Triassic shales, carbonates
160 and ophite, and Aptian-Albian limestones and conglomerates are occasionally preserved in
161 these thrust sheets. To the west, the Iguntze massif corresponds to the western extension of
162 the BLTU, with a thick sequence of Albian conglomerates (the Mendibelza conglomerates)
163 onlapping the Paleozoic basement (Boirie and Souquet, 1982). The contact between the CBB
164 and BLTU corresponds to the western extension of the North-Pyrenean Fault. Between the
165 Aspe and Ossau valleys (Fig. 2), the Bergon and Tacha klippes correspond to the southernmost
166 remnants of the CBB Mesozoic succession resting on the BLTU. In the western area, the
167 Iguntze massif and BLTU are thrust southwards by the Lakora-Larra thrust system over the

168 Upper Cretaceous carbonates and flysch covering the Paleozoic of the Axial Zone (Teixell,
169 1990, 1996). The Lakora-Larra thrust system passes eastwards to the Eaux-Chaudes - Eaux-
170 Bonnes thrust system, which extends eastward into the basement of the northern Axial Zone
171 (Labaume et al., 2016).

172 The Iguntze massif-BLTU and the southern part of the CBB correspond to the
173 former Iberian margin of the Pyrenean rift, while the northern CBB (Mail Arrouy ridge; Fig. 2)
174 are ascribed to the European margin (Canérot et al., 1978; Puigdefàbregas and Souquet, 1986;
175 Combes et al., 1998; Tugend et al., 2014; Teixell et al., 2016). In the eastern CBB, a small
176 domain of scapolite-bearing Kimmeridgian marble occurs at the center of the Tres Crouts
177 structure (Casteras et al., 1970a) (Fig. 2), comparable to other occurrences of the Cretaceous
178 North-Pyrenean metamorphic zone further east (the so-called Internal Metamorphic Zone)
179 which delineate the axis of the mid-Cretaceous rift (Clerc et al., 2015). The rift axis is also
180 identified by the occurrence of seven lherzolite bodies, most of them located along the CBB
181 axis (Fig. 2). Recent studies show that high paleo-temperatures in the CBB are not restricted
182 to the Tres Crouts structure, but that most of the belt was affected by peak temperatures
183 around 350°C, with local maxima between 400 and 500°C close to some of the lherzolite
184 bodies and in the Tres Crouts metamorphic domain (Clerc et al., 2015; Menant et al., 2016;
185 Villard, 2016; Corre, 2017; Ducoux, 2017).

186

187 *2.2.2. Stratigraphy of the Mesozoic-Cenozoic*

188 A general description of the stratigraphy in the CBB area can be found in the
189 BRGM geological maps and notices (Alimen et al., 1963; Casteras et al., 1970a and b; Ternet
190 et al., 1980, 2004) and is summarized in the stratigraphic column in Figure 3.

191 Most of the rocks exposed in the CBB correspond to the Mesozoic sedimentary
192 cover, beginning with the Middle-Upper Triassic deposits. The latter consist of a disorganized
193 association of the characteristic shale/evaporite Keuper facies and occasional bodies of the
194 Muschelkalk facies, and also contain tectonic breccia (cargneule) and frequent intrusive ophite
195 bodies. Except local gypsum occurrences, evaporites are most often absent in outcrop due to
196 dissolution, but more than 2620 m of evaporite-rich facies, mostly halite, were drilled without
197 reaching their base in the Bélair well (cf. well log in <http://infoterre.brgm.fr/>), in the hanging
198 wall of the NPFT (BEL1 in Fig. 2). For convenience in the following, we collectively refer to the
199 Middle-Upper Triassic complex as Keuper according to its main component.

200 Over the Triassic is a series of platform carbonates of Jurassic – Early Cretaceous
201 age, 1500 to 2000 m-thick when complete (Lenoble, 1992; James, 1998). The Jurassic
202 corresponds to a westward-facing carbonate ramp interpreted as formed in a context of
203 moderate E-W trending extension, while the Neocomian records the formation of the E-W-
204 trending structures of the Pyrenean rift (Lenoble, 1992; James, 1998, James and Canérot,
205 1999; Canérot et al., 2005). The Lias usually begins with a polygenic breccia reworking various
206 carbonate facies, followed by dolomites and limestones passing upward to black marls. The
207 Dogger comprises Bajocian-Bathonian limestones followed by an about 400 m-thick interval
208 of black dolomite of Bathonian-Callovian age (referred to below as the Dogger dolomite).
209 Above a reduced dolomitic or calcareous Oxfordian, the Malm comprises Kimmeridgian
210 limestones with marl intervals, followed by a Portlandian dolomite.

211 The Lower Cretaceous begins by a discontinuous alterite horizon, up to ten of
212 meters thick, comprising ferruginous, locally pisolitic, breccia, sandstone and clay (“bauxite”)
213 which traduces transient emersion associated to a karstic erosion surface in the Upper Jurassic
214 (Lenoble, 1992; Combes et al., 1998; Canérot et al., 1999). The return of marine conditions is

215 marked by Barremian limestones and marls. They are followed by lower Aptian (Bedoulian)
216 marls (the Sainte-Suzanne marls) overlain by an about 300-400 m-thick layer of reefal to peri-
217 reefal limestones (Urgonian facies). These limestones are mostly upper Aptian (Gargasian) in
218 age but locally extend to the uppermost Aptian (Clansayesian) to lower Albian. Locally, the
219 erosion of the Jurassic reaches deeper stratigraphic levels and the Neocomian hiatus is wider,
220 in particular in the SW area where the Urgonian limestones rest on the Dogger dolomite and
221 in the Asasp diapir (Fig. 2) where it rests on Keuper facies (Casteras et al., 1970b).

222 The upper (Clansayesian-lower Albian) part of the Urgonian limestones shows
223 lateral transitions to argillaceous limestones and marls (the so-called Spicule marls, or Haux
224 marls in Souquet et al., 1985) which finally cover the limestone units, marking the onset of
225 platform drowning and basin deepening around the Aptian-Albian boundary. Above, the rapid
226 subsidence of the mid-Cretaceous basin during the main rifting stage is registered by the
227 middle Albian to lower Cenomanian black marls and turbidites of the Black Flysch Group
228 (Roux, 1983; Souquet et al., 1985), at least 2000 m thick. To the south, the lateral equivalents
229 of the Black Flysch in the BLTU correspond to small, mainly carbonate massifs in the eastern
230 area (Gallagos, Bazès, Arbéost, located in Fig. 2) and to the middle-upper Albian Mendibelza
231 conglomerates in the west (Iguntze massif) (Boirie and Souquet, 1982). The latter rework
232 mainly Paleozoic metasediments, as well as Permian-Triassic sandstones and Albian platform
233 carbonates.

234 The Black Flysch is followed by the Upper Cretaceous flysch sequences which are
235 preserved from erosion only in the NW part of the CBB and in the southern Aquitaine Basin.
236 These comprise about 2000 m of Cenomanian to Santonian calcareous flysch (Roux, 1983)
237 which exhibit the lower Cenomanian unconformity at the base in the Asasp area (Casteras et
238 al., 1970b) (Fig. 2). These are followed by up to 3000 m of Campanian-lower Maastrichtian

239 sandy flysch and upper Maastrichtian marls (Serrano, 2001; Biteau et al., 2006; Serrano et al.,
240 2006), which correspond to the first syn-orogenic deposits. The Tertiary, present only in the
241 Aquitaine Basin, begins with 80-100 m of lower Paleocene pelagic limestones followed by
242 middle Paleocene to Ypresian clays and sands, which are the last flysch facies (ibid). These are
243 covered by a west-prograding system of coastal sandstones and offshore marls of Ypresian to
244 lower Lutetian age, followed in turn by the upper Lutetian to Oligocene molasse recording
245 terrestrial environments in the upper part. The succession of the Aquitaine basin ends with
246 post-orogenic Upper Miocene to Quaternary detrital sediments.

247

248 *2.2.3 Basement and magmatic rocks*

249 Pre-Keuper rocks outcrops are scarce in the CBB and comprise mantle rocks
250 (more or less serpentinized Iherzolite), Paleozoic meta-sediments, and a tegument of Lower
251 Triassic sandstones. Six Iherzolite bodies occur along the axial part of the CCB, generally
252 associated with lenses of Paleozoic or Lower Triassic rocks. Five of them are embedded in
253 Keuper rocks at the core of anticlines (Tos de la Coustette, Moncaup, Saint-Pé) or thrust
254 hanging walls (Saraillé, Turon de la Técoùère), while the Urdach Iherzolite body is covered by
255 the upper Albian breccias and flysch at the western termination of the Mail Arrouy thrust
256 sheet (Fortané et al., 1986; Jammes et al., 2009; Debroas et al., 2010; Lagabrielle et al., 2010,
257 2019b; Corre, 2017) (Fig. 2). In the northeastern area, several hm- to km-scale bodies of
258 Silurian-Devonian metasediments, some of them covered by Lower Triassic sandstones, occur
259 along the emerging branch of the NPFT at the base of the hanging wall sequence (Fig. 2). These
260 basement bodies are directly covered by Aptian limestones, and one of them overlies a
261 Iherzolite body (Montaud).

262 Small bodies of Upper Cretaceous alkaline magmatic rocks occur in the northern
263 part of the CBB. They correspond to intrusive sills and dykes and submarine lava flows (pillow
264 lavas) intercalated mainly in the Black Flysch.

265

266 **3. Cross-sections of the North-Pyrenean Zone in the Chaînons Béarnais area**

267

268 We present four new cross-sections of the NPZ in the CBB area (Fig. 4),
269 constructed from original field observations and structural data complemented by
270 stratigraphic data from previously published maps (Alimen et al., 1963, Canérot, 1964; Paris,
271 1969; Casteras et al., 1970a and b; Ternet et al., 1980, 2004; Roux, 1983; Souquet et al., 1985).
272 The subsurface structures of the NPFT area were constructed from stratigraphic logs of oil
273 exploration wells and, for cross-section 4, from a new interpretation of an industrial seismic
274 profile (see references below, Section 4.3.1). The Move software was used at some steps of
275 cross-section construction. Below, we outline first the major features of the cross-sections.
276 More detailed descriptions and inferences from key-areas featuring salt structures are
277 provided below in Section 4.

278

279 *3.1. Structure of the Mesozoic cover*

280

281 We interpret that the CBB is a fold and thrust system decoupled from the
282 Paleozoic basement along the Keuper unit which cores the anticlines. The Keuper thickness at
283 antiformal cores is variable, many of them showing reduced thickness or disappearance or
284 apparent faulting indicative of salt squeezing and welding. A characteristic of the CBB
285 structure is the lack of consistent structural vergence, a feature typical of fold-thrust belts

286 detached above thick salt (Davis and Engelder, 1985; Hudec and Jackson, 2001). In the west
287 (cross-sections S1 to S3 in Fig. 4), the northern and central ridges (Mail Arrouy and Sarrance,
288 respectively) show south-vergent thrusts (Fig. 5) (the Sarrance thrust anticline passing
289 eastward to the south-vergent Aran anticline), somewhat unexpected in the North-Pyrenean
290 retrowedge, while the southern ridge (Layens-Ourdinse) corresponds to a north-vergent
291 syncline-anticline pair. The Bergon klippe, the southernmost occurrence of the CBB, features
292 also a north-vergent syncline (cross-sections S2 and S3 in Fig. 4). In the east (cross-section S4
293 in Fig. 4), the northern Saint-Pé anticline is an upright structure, the central Tres Crouts
294 structure features two synclines of opposite vergence, and the southern Estibète-Pibeste
295 ridge is thrust southwards along the NPF. The Andorre syncline, between the Tres Crouts
296 structure and the Estibète-Pibeste ridge, is an upright structure in cross-section S4 but it
297 features a northward vergence both east and west of the section. The tectono-sedimentary
298 relationships in the Mesozoic series of the CBB, involving bed fanning and thickening in the
299 synclines (growth strata) and lateral facies changes, are discussed in Section 4.

300 Most of the stratigraphic formations of the CBB succession feature a bedding-
301 parallel foliation hereafter referred to as S0-S1. It corresponds to a slaty cleavage in the
302 marly/pelitic layers (Fig. 5b), and to a schistosity marked by a grain-shape fabric of calcite in
303 limestones (Fig. 5c). The intensity of the S0-S1 is heterogeneous both laterally and vertically,
304 and it is absent in the dolomites. A later S2 is locally present with variable attitudes, probably
305 related to local deformation zones. The S0-S1 is attributed to the mid-Cretaceous syn-
306 metamorphic extensional context (Corre et al., 2016; Villard, 2016; Corre, 2017), although
307 describing and discussing the distribution of these cleavages is beyond the scope of this paper.
308 It must be noted that no widespread oblique regional cleavage related to the Pyrenean
309 compression can be observed.

310

311 *3.2. The basement of the Chaînons Béarnais Belt*

312

313 Cross-section construction implies a difference of elevation of the Paleozoic
314 basement top under the CBB of about 8000 m, from its outcrop area in the south to its deepest
315 occurrence below the northern CBB (Fig. 4). This basement is interpreted as the distal part of
316 the former Iberian margin of the Pyrenean rift, now forming the hanging wall of the NPFT
317 (Teixell et al., 2016, 2018), but its internal fault structure remains largely unconstrained. The
318 geometry of the cover structures above (tight folds and thrusts) suggests décollement in the
319 Keuper (Fig. 4), the Mail Arrouy thrust being the only one with a kilometeric displacement. On
320 the other hand, there is no evidence for large normal-faulted tilted blocks involving the
321 Paleozoic basement and its Mesozoic cover similar to the Arbailles block of the neighbouring
322 Mauléon basin (Ducasse et al., 1986; Canérot, 1989; Masini et al., 2014; Saspiturry et al.,
323 2019). Hence, we favor the interpretation of a general decoupling between the CBB fold-and-
324 thrust cover and the underlying basement. We infer a relatively smooth basement top
325 affected by thrusts faults branching to the Keuper décollement level above. To the south, the
326 thrust structures correspond to the northern part of the south-vergent BLTU, while to the
327 north, we infer north-vergent thrusts in the nearest hanging wall of the NPFT. The occurrence
328 of inherited normal faults with limited offset is also possible, but there is no evidence that
329 some of them were inverted as reverse faults breaching up to the surface as shown in some
330 published cross-sections (Canérot, 1989; Dubos-Salée et al., 2007). In order to minimize the
331 Keuper volume during section construction, our sections show primary welding of the deepest
332 synclines to the basement top.

333

334 *3.3. The structures south of the CBB: the Iguntze-Mendibelza massif, Bedous-Laruns thrust*
335 *units and Axial Zone*

336

337 To the south-west, the CBB is in contact with the easternmost part of the Iguntze
338 massif by the steeply-dipping Licq fault (cross-section S1 in Fig. 4). The Iguntze massif
339 comprises the middle-upper Albian Mendibelza conglomerates onlapping southwards a slice
340 of Paleozoic metasediments (much reduced on cross-section S1). Based on the onlap
341 geometry (Boirie and Souquet, 1982) and the frequent remnants of the Permian-Lower
342 Triassic sandstone tegument preserved along the onlap surface (Casteras and Souquet, 1970),
343 we follow the interpretation of the latter surface as the northward tilted basement top
344 (Teixell, 1993; Teixell et al., 2016; Saspiturry et al., 2019) rather than as the denuded footwall
345 of an intracrustal detachment associated to southward-tilted basement fault blocks (Johnson
346 and Hall, 1989; Masini et al., 2014). The Iguntze massif and Licq fault have been hypothetically
347 shown south of the Bergon klippe on cross-sections S2 and S3. From its main exposure west
348 of the study area, the Paleozoic slice of the Iguntze massif is interpreted as detached by the
349 short-cut of a former normal fault footwall by the Pyrenean thrust (Teixell, 1993). This differs
350 from alternative interpretations rooting the Iguntze-Mendibelza thrust in the underlying
351 crustal basement (e.g. Masini et al., 2014; Dumont et al., 2015), which do not match the
352 observed low-angle geometry of the thrust above a footwall of Keuper facies and branched to
353 the north to the Licq fault (Casteras and Souquet, 1970).

354 In cross-section S1, the southern CBB and Iguntze massif are thrust above the
355 Bedous Triassic unit, comprising Keuper facies, Muchelkalk carbonates and ophite bodies,
356 itself thrust above the Upper Cretaceous cover of the Axial Zone. The two thrusts branch
357 southward into the Lakora thrust, defining the Bedous Triassic unit as a duplex structure, while

358 a lower branch, the Larra thrust, propagated near the top of the Upper Cretaceous carbonates.
359 Both the Lakora and Larra thrusts propagated southward up to the Tertiary succession of the
360 northern Jaca basin and were subsequently folded by the Axial Zone antiform, here
361 corresponding to the Gavarnie thrust hanging wall culmination (Teixell, 1990, 1996; Labaume
362 et al., 2016; Labaume and Teixell, 2018).

363 East of the Gave d'Aspe valley, the southern CBB rests over the Paleozoic
364 basement and Lower Triassic sandstone tegument of two of the BLTU: the Bois de la Traillère
365 and Montagnon d'Iseye thrust units in cross-sections S2 and S3, respectively (Fig. 4). In cross-
366 section S3, the Montagnon d'Iseye unit is itself thrust (with intervening discontinuous slices
367 of the Bedous Triassic) above the Eaux-Chaudes km-scale recumbent fold-thrust structure
368 involving the Paleozoic metasediments and their cover of Upper Cretaceous limestone and
369 flysch (which corresponds to the northernmost structure of the Axial Zone; Caldera et al.,
370 2019). A western extension of the Eaux-Chaudes thrust-fold is shown tentatively in cross-
371 section S2. All these thrust units branch southward into the Lakora and Larra thrusts. Hence,
372 the southern CBB décollement, Iguntze massif, Bedous Triassic unit, BLTU and the Eaux-
373 Chaudes fold-thrust form a complex south-vergent duplex structure corresponding to the
374 exhumed root of the Lakora and Larra thrusts. Steep north-vergent reverse faults in the Axial
375 Zone basement correspond to small backthrusts in the Gavarnie thrust hanging wall that
376 deform the previous low-angle Lakora-Larra thrust system (Dumont et al., 2015).

377

378 *3.4. The structure north of the CBB: Grand-Rieu High and North-Pyrenean Frontal Thrust*

379

380 North of the hanging wall cut-off of the basement top by the NPFT, the Grand-
381 Rieu High (cross-sections S1 and S4 in Fig. 4) corresponds to a basement horst identified in

382 seismic profiles and wells (Bourrouilh et al., 1995; Serrano et al., 2006). It is a mid-Cretaceous
383 structure that separates the Chaînons Béarnais and Mauléon basins in the south from the
384 Arzacq-Tarbes basin in the north, and is interpreted to correspond to the upper European
385 margin of the Pyrenean rift (e.g. Lagabrielle et al., 2010; Masini et al., 2014; Teixell et al., 2016;
386 Saspiturry et al., 2019). The frontal part of the NPFT cuts the Mesozoic cover above the Grand
387 Rieu High as a blind thrust in the west and emerging in the east (cross-sections S1 and S4 in
388 Fig. 4, respectively).

389 The structure of the NPFT in cross-section S1 is deduced from surface geology
390 (Alimen et al., 1963; Casteras et al., 1970b), from four well logs located along the section
391 (OLN1, CAD2, FLAS2 and LRT1; see location of OLN1 and CAD2 on map in Figure 2, and logs in
392 Supplementary Data, Fig. S1), and the cross-section published by Lagabrielle et al. (2010). The
393 major structure is the large north-vergent Oloron anticline, cored by a thick accumulation of
394 Keuper facies. This interpretation accords with the thick accumulation of Keuper facies drilled
395 11 km to the east in the Bélair well (cf. above, Section 2.2.2) below the Jurassic of the hinge
396 zone of the Oloron anticline (see the corresponding cross-section in Biteau et al., 2006). The
397 northern limb of the Oloron anticline comprises a thick growth strata fan of Campanian-
398 Maastrichtian flysch and overlying Paleocene-Eocene flysch and shallow marine deposits. On
399 the southern Grand-Rieu High, the Jurassic-Lower Cretaceous strata were totally removed and
400 the Cenomanian to Santonian flysch is also absent. The Upper Cretaceous begins with a few
401 tens of meters thick breccia layer resting on the Keuper facies (well CAD2) or on the Lower
402 Triassic Buntsandstein facies that overlies Paleozoic metasediments (well LRT1). The breccia
403 reworks Paleozoic metasediments, Buntsandstein sandstone and Jurassic to Lower Cretaceous
404 carbonates, and is onlapped by the Campanian flysch (Supplementary Data, Fig. S1). The

405 occurrence of Keuper bodies along the upper branch of the NPFT is discussed below (Section
406 4.3.1).

407 In cross-section S1, a lower branch of the NPFT propagated 16 km northward
408 along the base of the Upper Cretaceous flysch to form the Pau anticline, which is the
409 northernmost structure of the North-Pyrenean front (see Figure 5 in Lagabrielle et al., 2010).
410 It features isopach beds in the Upper Cretaceous and growth strata in the Eocene (see Figure
411 9 in Canérot et al., 2005).

412 The structure of the NPFT in cross-section S4 (Fig. 4) is constrained by the
413 combination of stratigraphic and structural data from the surface exposure, two well logs
414 (SVT1 and LVN1; see location of wells on map in Figure 2, and logs in Supplementary Data,
415 Figure S1), and the LR5 seismic profile published by Serrano et al. (2006) (see interpretative
416 line-drawing and comments in Supplementary Data, Fig. S2). The seismic image was converted
417 to depth using the time/depth conversion in wells and isohypse maps of key stratigraphic
418 surfaces in Serrano et al. (2006). In cross-section S4, the NPFT separates the northern limb of
419 the Forêt de Mourle syncline in the hanging wall from a large northward overturned syncline
420 in the footwall. The hanging wall features a body of Paleozoic metasediments covered by
421 Aptian limestones and the Albian Black Flysch. The Paleozoic corresponds to one of the bodies
422 that occur discontinuously at the hanging wall of the NPFT in the eastern area, and it overlies
423 the Montaud Iherzolite body located 2 km west of the section (Fig. 2). The footwall syncline
424 comprises the Cenomanian to Eocene succession, with a large growth strata fan in the
425 Campanian to Eocene. In the SVT1 well, the Upper Cretaceous flysch rests on a few tens of
426 meters of Neocomian carbonates and shales following the Keuper and Buntsandstein facies
427 and Devonian metasediments of the Grand-Rieu High. More to the north, the Campanian
428 flysch onlaps an erosion surface cutting the Albian flysch of the Tarbes basin, and the Albian

429 flysch rests itself on an erosional surface cutting the Jurassic. Due to the presence of a slight
430 anticline in the Upper Cretaceous more to the north (see the complete LR5 seismic profile in
431 Serrano et al., 2006), we interpret that the base of the Upper Cretaceous flysch acted as a
432 Pyrenean thrust décollement surface, similarly to the structure observed more to the west
433 (cross-section S1 in Figure 4). The occurrence of Keuper bodies along the upper branch of the
434 NPFT and the origin of the erosional surfaces are discussed below (Section 4.3.1).

435

436 **4. Salt tectonics in the Chaînons Béarnais belt**

437

438 Salt diapirs formed during the Cretaceous extension and subsequently squeezed
439 during the Pyrenean compression have already been described in the Aquitaine basin and
440 adjacent Mauléon basin in the frame of intensive oil exploration (Mediavilla and Mauriaud,
441 1987; Bourrouilh et al., 1995; Canérot et al., 2005; Biteau et al., 2006; Serrano et al., 2006). A
442 diapiric origin of the CBB anticlines with halokinesis initiated during the Early Cretaceous was
443 first proposed by Canérot (1988), based on succinct description of thickness reduction of the
444 Barremian-Aptian carbonates at anticline hinges, locally associated to erosional surfaces, and
445 lateral facies transitions between the Aptian carbonates and marls deposited in the adjacent
446 synclines. In particular, Canérot (1988) notes the unconformity between the Barremian and
447 Jurassic carbonates at the Moncaut anticline crest (Figs. 2 and 6). Canérot (1989) and James
448 and Canérot (1999) provide more detailed descriptions and kinematic restorations of Lower
449 Cretaceous diapirs in the Mauléon basin. On the other hand, detailed description in terms of
450 salt structures in the CBB remain limited to a transverse welded diapir in the Lourdios syncline
451 (Lenoble and Canérot, 1992) (“Ponsuzon weld” in Fig. 7), and the Lauriolle breccia (located at
452 the eastern extremity of the Ourdinse ridge, Fig. 7), interpreted as related to diapir collapse

453 (Canérot and Lenoble, 1993). According to these various papers, the diapirs formed above
454 basement normal faults bounding tilted blocks of the Cretaceous rift and remained there,
455 squeezed during the subsequent Pyrenean orogeny without major décollement from their
456 Paleozoic substratum. Canérot et al. (2005) further develop description of diapirism in the
457 Aquitanian domain, including the interpretation of the NPFT as a former Cretaceous diapiric
458 ridge inverted by the Pyrenean compression.

459 In what follows we present a new description of a selection of characteristic salt
460 structures in the CBB and discuss the timing and context of their development. We first
461 present the south-western part of the CBB, which provides the largest exposure of the late
462 Aptian-Albian basins, then we describe the complex Tres Crouts structure which bears
463 evidence of a protracted development in several stages, and finally we examine the structures
464 related to the Pyrenean inversion, in particular the NPFT.

465

466 *4.1. The Lourdios and Barescou minibasins and adjacent salt ridges*

467

468 The area comprises two depocenters of upper Aptian-Albian marls and flysch,
469 bounded by Jurassic-Lower Cretaceous carbonate ridges, from south to north: the Lourdios
470 syncline between the Layens-Ourdinse ridge and the Sarrance-Aran anticline, and the
471 Barescou syncline between the latter anticline and the Mail Arrouy thrust ridge (Figs. 2 and
472 7). The map in Figure 7 is based on original mapping in the field and 3D aerial photographs.
473 Although the overall structure of the area was correctly reported in previous maps (Canérot,
474 1964; Paris, 1969; Casteras et al., 1970b; Haller and Jardiné, 1986; Ternet et al., 2004), our
475 mapping provides new structural data, more precision, and new interpretation for numerous
476 contacts, in particular for the southern limb of the Sarrance anticline (including the Iherzolite

477 massifs) and for the eastern Lourdios syncline (regarding the map pattern and stratigraphic
478 correlation of limestone-marl intercalations in the Pic Montagnon and Ourlène areas).

479

480 *4.1.1. The Layens-Ourdinse carbonate ridge: welded diapirs and overturned megaflaps*

481 The stratigraphy of the Layens-Ourdinse ridge is characterized by a deep
482 Neocomian erosion and associated hiatus, with the upper Aptian Urgonian limestones resting
483 on the Dogger dolomite which often displays an alterite horizon at its top. The structure
484 corresponds to a recumbent syncline-anticline pair with northward vergence (cross-sections
485 S1 to S3 in Fig. 4, and Fig. 7). The antiform structure is preserved from erosion at the Layens
486 summit (Fig. 8) and Ourdinse plateau (Fig. 9) where it can be observed that it corresponds to
487 a quasi-welded overhang of Keuper facies above a sub-horizontal overturned megaflap of
488 Jurassic and Urgonian carbonates. In the Layens, transverse structures interfere with the
489 general E-W trend, i.e. N-S trending folds in the western slope of the Aspe valley and the NNE-
490 SSW-trending megaflap on the eastern flank of the Layens summit (Fig. 8). The Urgonian
491 limestones are about 350 m thick in the normal limb of the syncline and thins dramatically
492 across the syncline hinge to pass to less than 100 m in the megaflap (cross-sections S1 to S3
493 in Fig. 4), attesting diapir rise during the late Aptian.

494

495 *4.1.2. The Sarrance and Aran anticlines: squeezed salt walls and salt extrusion*

496 The Sarrance anticline is cored by the Keuper facies (cross-section S1 in Fig. 4,
497 and Fig. 7). Two lherzolite bodies, the Sarailié and Tos de la Coustette lherzolites, along with
498 bodies of Paleozoic metasediments, show lateral contacts with the Keuper and are covered by
499 the Jurassic carbonates which show there a subtractive contact at the base (Fortané et al.,
500 1986; Corre et al., 2016; Corre, 2017; Lagabrielle et al., 2019a). The stratigraphy of the

501 Sarrance anticline is more complete than in the Layens-Ourdinse ridge, with the northern limb
502 comprising the Kimmeridgian and Neocomian carbonates (the latter disappearing westward
503 according to Lenoble, 1992), Sainte-Suzanne marls and thick (350-400 m) Urgonian
504 limestones. Nevertheless, this succession thins in the southern limb with strong reduction of
505 the Kimmeridgian by erosion, absence of the Neocomian and reduced thickness of the
506 Urgonian limestones (100-150 m). This reduced Urgonian is thinner than in the normal limb
507 of the Layens-Ourdinse syncline, which also indicates rise of the Sarrance ridge in late Aptian
508 times. The Sarrance ridge has an upright geometry with vertical northern limb. The central
509 part of the southern limb features a sub-horizontal south-vergent thrust carrying the Keuper
510 facies and the Saraillé massif above the Spicule marls of the Lourdios syncline with a
511 displacement not exceeding 1 km. The Saraillé massif comprises a south-vergent recumbent
512 anticline of Mesozoic carbonates wrapping in its core the Saraillé lherzolite and a lens of
513 Paleozoic metasediments (see Corre et al., 2016, Corre, 2017, and Lagabrielle et al., 2019a for
514 detailed description). East of the thrust unit, both flanks of the Sarrance ridge are welded with
515 an overturned southern flank. To the west, the southern flank features sub-vertical to steep
516 southward dips. The Tos de la Coustette lherzolite prevented welding, but the lherzolite is
517 bounded by a tear fault west of which the anticline extremity is welded.

518 The structure of the thrust unit rooted in the anticline core and the lateral
519 welding of the anticline denote a salt extrusion related to salt wall squeezing and that carried
520 part of the anticline crest (the Saraillé massif). Furthermore, the Sarrance ridge shows two
521 peculiar structural features that we interpret as also resulting from the salt wall deflation
522 during salt expulsion from its core: i) on both sides of the thrust unit, the carbonates and
523 overlying Spicule marls of the southern flank curve toward the north until their cut off at the
524 thrust, and ii) north of the thrust unit, the northern flank features a reentrant accommodated

525 by sub-vertical N-S trending faults (Fig. 7). The Spicule marls in the southern flank of the
526 Sarrance anticline dip southward, but with lower values in the thrust footwall than more to
527 the west (Fig. 7), which may also have resulted from evacuation of underlying salt. These
528 structures are quite analogous to those on several salt anticlines of the Zagros fold belt in SW
529 Iran that feature salt glacier extrusions associated with forelimb curvatures and cut-offs and
530 back-limb reentrants similar in style and dimensions to those at Sarrance (Jackson and Hudec,
531 2017).

532 To the east, a complex transverse fault zone makes the lateral transition between
533 the Sarrance and Aran ridges, with offset of their axes (referred to as the Aran transverse fault
534 zone in the following and labelled ATFZ in Fig. 7). The Aran anticline has been less exhumed,
535 with discontinuous surface exposition of Urgonian limestones covered by the Spicule marls.
536 The Aran is a tight anticline with steeply dipping limbs, the narrow shape of which suggesting
537 the absence of the Jurassic-Neocomian in the core. We interpret this structure as an ancient
538 exposed diapir where the Urgonian limestones deposited directly on the Keuper before being
539 finally squeezed. Some limestone beds at the top of the Urgonian succession pinch out in the
540 Spicule marls in the southern limb of the anticline (cross-sections S2 and S3 in Fig. 4, and Figs.
541 7 and 10), attesting its rising at late Aptian times.

542

543 *4.1.3. The Lourdios syncline: a Clansayesian (-lower Albian) minibasin*

544 The Lourdios syncline shows marked stratigraphic differences between its
545 western and eastern parts. While the area located west of the Aspe river shows the classical
546 superposition of the Spicule marls above the Urgonian limestones, the eastern area (Bois de
547 Gey, Pic Montagnon) features numerous intercalations of Urgonian-type limestones in the
548 marls (Fig. 7). The BRGM map (Casteras et al., 1970b) ascribes an Albian age to the lowest

549 layer of Spicule marls, and an upper Aptian age to the overlying marl-limestone intercalations,
550 implying a thrust between both (see also Canérot et al., 1999). Conversely, our field
551 observations and mapping confirm the stratigraphic continuity of the marl-limestone
552 intercalations with the underlying Urgonian limestones from the Aran anticline crest to the
553 Ourdinse overturned flap as was originally shown in Canérot (1964). Lenoble (1992) attributed
554 a Gargasian age to the main Urgonian limestones and an uppermost Aptian (Clansayesian) age
555 to the overlying marl-limestone intercalations, possibly reaching the Albian in the upper part.
556 Stratigraphic continuity implies that a large part of the Spicule marls in the western part of the
557 syncline has also a Clansayesian age.

558 The Lourdios syncline infill features a growth pattern. Synsedimentary folding with a
559 pouch-like geometry is best evidenced east of the Aspe river thanks to the marl-limestone
560 intercalations which show a growth pattern on both limbs (Bois de Gey-Pic Montagnon area;
561 cross-sections S2 and S3 in Fig. 4, and Figs. 7, 9 and 10). To the north, the lower marl beds
562 feature a fan-geometry against the southern limb of the tight Aran anticline. To the south, the
563 marls and intercalated limestone beds thin southwards when passing into the overturned flap
564 of the Ourdinse ridge. The along-strike thickness variations also evidence for depocenter
565 shifting probably related with transverse structures, with the maximum thickness (a hundred
566 of meters) of the lowest limestone bed in a depocenter at the west of the Montagnon
567 mountain while the upper beds merge eastward in the 200 m thick carbonate mass forming
568 the Pic Montagnon (Fig. 7).

569 West of the Aspe river only the lowest of the intercalated limestone beds is
570 present, which onlaps southwards the Urgonian limestones of the Layens ridge (Fig. 7), giving
571 a growth strata pattern similar to that described to the east. The western Lourdios syncline
572 contains mostly a thick accumulation of argillaceous limestones, marls and rare thin-bedded

573 silt to fine-grained sandstone layers all corresponding to the Spicule marls unit (Souquet et al.,
574 1985). In the western part (cf. near the village of Lourdios-Ichère; Fig. 7), the syncline is
575 strongly asymmetric with a wide northern limb where a slight southward dip reduction
576 traduces a growth strata arrangement, and a very short southern limb against the steep
577 Urgonian limestones of the Layens megaflap (cross-section S1 in Fig. 4). This growth geometry
578 indicates a progressive depocenter migration, likely by salt expulsion. North of the Lourdios-
579 Ichère village, the lower part of the succession contains several debris flow beds reworking
580 clasts of bioclastic limestone. On the other hand, the Spicule marls present in the Sarailié
581 thrust unit has a more carbonated facies than the marls of both limbs of the ridge. This thrust
582 unit representing an element of the Sarrance anticline crest, we can infer that the anticline
583 crest at Clansayesian times was a rising high where shallower bathymetry allowed carbonate
584 deposition feeding the debris flows in the adjacent deeper domain. Isolated limestone lenses,
585 up to a few tens of meters long occur in the eastern part of the western Lourdios syncline,
586 which may represent olistoliths originating either from the northern (Sarrance) or southern
587 (Layens) rising highs.

588 On the western side of the Aspe valley, an alignment of blocks of Triassic
589 carbonates and breccias defines a N-S trending welded diapir across the Lourdios syncline
590 (Lenoble and Canérot, 1992) (“Ponsuzon weld” in Fig. 7). This diapir collapsed the western
591 compartment while the top of the Urgonian limestones crops out in a small anticline on the
592 eastern side. It may have contributed to the salt evacuation discussed above for the Sarailié
593 thrust unit. A steeply dipping fault affects the Gargasian limestone and overlying marls at the
594 syncline core zone with a hundred of meters displacement, which we interpret as an ancient
595 normal fault, inverted as a reverse fault in the Aspe valley (cross-sections S2 and S3 in Fig. 4,
596 and Figs. 7, 9 and 10).

597 We underline that the stratigraphic continuity across the Lourdios syncline
598 shown in the present work does not favor the existence of a large-displacement south-vergent
599 thrust at the base of the Sarrance anticline and emerging in the syncline as suggested in a
600 previous work (Tugend et al., 2014).

601

602 *4.1.4. The Ourlène carbonate ridge and Barescou syncline: a Clansayesian-Albian depocentre*

603 In the northern limb of the Aran anticline, the Spicule marls covering the
604 Gargasian limestones are themselves covered by the Urgonian limestones of the Ourlène ridge
605 (cross-sections S2 and S3 in Fig. 4, and Fig. 7). In the eastern part, the ridge features a complex
606 stratigraphic pattern with marl intercalations, carbonate pinch-outs, erosional truncations and
607 onlaps. Locally, a northward prograding pattern of the limestones, away from the Aran
608 anticline, can be observed (Fig. 11). The geometry of the erosional truncations and onlaps as
609 well as the northward shelf edge progradation are coherent with a north-facing sedimentary
610 slope related to coeval rising of the Aran anticline. The Ourlène limestone-marl intercalations
611 merge westward in a single limestone body in continuity with the Urgonian limestones of the
612 Sarrance anticline northern limb. Below this limestone, the Aran transverse fault zone
613 juxtaposes the Sarrance Jurassic to lower Aptian succession with the Spicule marls covering
614 the Aran Gargasian limestones (Fig. 7). Tens of meters-sized bodies of partially dolomitized
615 limestone are intercalated in the Spicule marls along the fault system. We interpret this fault
616 zone as a Clansayesian syn-sedimentary cross-fault that separated a more subsiding domain
617 to the east from a less subsiding domain to the west. In this interpretation, the Gargasian
618 limestones and overlying marls of the Aran anticline are stratigraphically equivalent to the
619 lower part of the Sarrance Urgonian, while the Ourdinse limestone is the eastern extension of
620 the upper part of the Sarrance Urgonian. Bed mapping shows that the Ourdinse limestone

621 thins rapidly westward in the northern limb of the Sarrance anticline with possible lateral
622 transitions to the Spicule marls (Fig. 7). This interpretation differs from previous works which
623 assumed that the Ourlène limestones are thrust southward over the Spicule marls (Haller and
624 Jardiné, 1986; Canérot et al., 1999). However, the existence of such thrust would implicate
625 the existence of a lateral ramp separating the Ourlène and Sarrance Urgonian layers, while
626 our mapping shows the continuity of the Urgonian between the two ridges. Hence, the
627 limestone-marl intercalations at the eastern extremity of the Ourlène ridge are more logically
628 interpreted as a northern equivalent of the similar intercalations south of the Aran anticline
629 in the Bois de Gey-Pic Montagnon area discussed in the previous section.

630 North of the Sarrance anticline and Ourlène ridge, the Barescou succession forms
631 a north-dipping monocline in the footwall of the Mail Arrouy thrust (cross-sections S1 to S3 in
632 Fig. 4, and Fig. 7). The stratigraphic succession is more complete than in the Lourdios syncline,
633 with the Clansayesian-lower Albian Spicule marls followed by the middle-upper Albian Black
634 Flysch, itself subdivided into the Escot pelites and the Barescou sandstones (Roux, 1983;
635 Souquet et al., 1985). The Barescou sandstones are an upper Albian (Vraconian) turbidite
636 succession featuring up to m-thick beds of sandstone or/and conglomerate with cm-sized
637 rounded clasts of Mesozoic carbonates and Paleozoic rocks (Roux, 1983). Bed thickness and
638 grain size decrease westward, suggesting an eastern provenance. The steeply-dipping but
639 variable dips in the Spicule marls and Black Flysch succession in the Aspe valley do not show
640 an unequivocal growth folding pattern (cross-section S1 in Figure 4), but such a pattern may
641 exist in the east where the dip of the Barescou sandstones is notably lower than that of the
642 underlying Spicule marls and Escot pelites (cross-section S2 in Figure 4). The absence of
643 equivalent coarse turbidites north of the Mail Arrouy thrust suggests that the Mail Arrouy
644 already formed a ridge bounding the Barescou depocenter during the Albian.

645

646 *4.1.5. The Lauriolle diapiric breccia*

647 At the eastern end of the Ourdinse ridge, the Jurassic-Urgonian succession is
648 replaced by the about 200 m thick Lauriolle breccia (Canérot and Lenoble, 1993; Cloix, 2017)
649 (Figs. 7 and 10). On the western side, the carbonate succession shows in-situ brecciation
650 increasing eastward and passing laterally to a polygenic breccia comprising mainly Jurassic
651 elements in the lower part and Urgonian elements in the upper part (Cloix, 2017). Brecciation
652 is posterior to the syn-metamorphic foliation of the limestones observed in the clasts. The
653 presence of authigenic quartz crystals typical of the Keuper evaporites in the matrix of the
654 polygenic breccia attests to diapirism. Canérot and Lenoble (1993) interpreted the breccia as
655 resulting from diapir dissolution and collapse during the late Aptian, before being covered by
656 the Albian marls. However, this would imply the unlikely hypothesis that carbonate
657 metamorphism occurred at the sea floor. Alternatively, we postulate that, similarly to the
658 other domains of the ZNP, metamorphism occurred during the Late Cretaceous favored by the
659 blanketing effect of the flysch cover (Clerc et al., 2015, 2016), arguing for a later brecciation
660 process.

661

662 *4.1.6. Synthesis of salt tectonic evolution in the south-western CBB*

663 The descriptions above indicate that the south-western CBB recorded a
664 protracted salt tectonic activity during the Early Cretaceous. The first record of salt ridge rising
665 would be the inferred absence of the Jurassic carbonates in the core of the Aran anticline, due
666 to exposed diapir before the late Aptian. The same process may have occurred east of the
667 Ossau valley at the Béon anticline (Fig. 2) which shows a welded structure of the Urgonian
668 limestones similar to that of the Aran anticline.

669 In contrast to the Moncaut, Aran and Béon anticlines where it is limited to a
670 narrow zone at the anticline crest, the Neocomian erosion and associated depositional hiatus
671 increase southward across a km-wide zone in the south-western CBB, resulting in the
672 deposition of the Urgonian limestones above the Dogger dolomite in the Layens-Ordinse ridge
673 and the southern Tacha and Bergon klippen (Canérot et al., 1978) (cross-sections S2 and S3 in
674 Fig. 4, and Fig. 7). Previous works ascribed this to uplift of faulted basement blocks in the
675 Iberian mid-Cretaceous margin (Canérot et al., 1978; Puidefàbregas and Souquet, 1986;
676 Combes et al., 1998). Although this interpretation is consistent with the geodynamic context,
677 unequivocal evidence of contemporaneous salt wall rising at Moncaut, Aran and Béon
678 anticlines, and the presence of Keuper facies underlying the whole area suggests that the
679 rising of a wide salt inflation may have contributed to the uplift and erosion/non deposition
680 of the carbonate succession in the southern domain of the CBB.

681 The rapid thickness variations of the Gargasian limestones in the Layens-
682 Ourdinse overturned flap and at the crestal zone of the Sarrance ridge argue for rising of both
683 diapiric structures during the late Aptian. Coeval rising of the Aran anticline is also shown by
684 local pinching of the upper layers of the Gargasian limestone in the marls of the Lourdios
685 syncline.

686 Halokinesis increased during the Clansayesian to early Albian with thick
687 accumulations of Spicule marls in the Lourdios and Barescou synclines by salt withdrawal. The
688 Lourdios syncline shows the typical pouch-like geometry of a minibasin subsiding between
689 uprising adjacent diapirs, with marked growth strata on both limbs. Complex stratigraphy with
690 limestone-marl intercalations, debris flow deposits and olistolites, erosional surfaces and
691 onlaps attest important substratum mobility. The transverse Ponsuzon diapir and the
692 transition between the Sarrance and Aran anticlines show the notable role played by

693 transverse fault systems in partitioning the diapiric system. The latter transverse fault system
694 probably made the transition between the eastern diapir where the Gargasian limestones
695 deposited above the Keuper (Aran) from the less uplifted western domain where only the
696 Upper Jurassic was eroded (Sarrance). Later, the uplift trend was inverted and the transverse
697 fault system separated a more subsiding domain in the east (Aran, Ourlène) from a less
698 subsiding western domain in the Sarrance northern limb. This inversion was probably favored
699 by the previous larger rise of the salt in the east.

700 However, salt tectonics during the Clansayesian to early Albian often implied
701 moderate bed uprising. Indeed, the uppermost beds of the Lourdios and Barescou minibasins
702 are also folded and thrust (Sections 1 to 3 in Fig. 4, and Figs. 7, 8, 9 and 10). We interpret this
703 as the result of later salt-wall squeezing, traduced in the present structure by diapir welding,
704 megaflop overturning and salt extrusion. The Lauriolle breccia may also be related to diapir
705 squeezing, posterior to the Late Cretaceous metamorphism in this case. However, dating the
706 late diapir squeezing is difficult due to erosion of post-Albian strata. This dating is discussed
707 below in the frame of a general geodynamic model for the CBB evolution.

708

709 *4.2. The Tres Crouts structure: a polygonal multi-stage salt structure*

710

711 Located in the eastern part of the CBB, the Tres Crouts structure corresponds to
712 a polygonal weld resulting from the interference of E-W to ESE-WNW (the general trend of
713 the CBB) and NNE-SSW trends (Figs. 2 and 12). The latter corresponds to the transverse trend
714 which characterizes the eastern termination of the CBB at its transition to the Ossun basin
715 (Fig. 2). The map presented in Figure 12 is based on the BRGM geological map (Casteras et al.,
716 1970a) and includes structural data from Lanusse (1969), new stratigraphic and structural

717 interpretation at the NE border of the Tres Crouts structure from Cloix (2017), and precision
718 of contours and new structural data along cross-section S4 (Fig. 4) from the present work. In
719 the western part of the structure, where the weld outcrops at the level of the Lower Liassic
720 breccias, there remain lenses of Keuper facies, ophite bodies and Paleozoic metasediments
721 (Figs. 12 and 13). A remarkable feature is the centripetal vergence of the structure, i.e. the
722 inner compartment features an overturned flap with an ellipsoidal syncline axial trace,
723 resulting in two recumbent synclines with opposite vergence in profile (cross-section S4 in Fig.
724 4).

725 In the northern area, the occurrence of the subvertical Dogger dolomite against
726 the northern flank of the weld surface may result from the direct deposition of the dolomite
727 above the Keuper in a Jurassic diapir, but alternatively it may result from the tearing of the
728 Lias off the diapir margin during salt flow. Whatever the case, the dolomite layer against the
729 welded surface features pervasive brecciation cemented by dolomite and subsidiary quartz
730 (Cloix, 2017) interpreted as resulting from the diapir squeezing process. In the inner syncline,
731 the Kimmeridgian is thinner in the reverse flap than in the normal limb, suggesting a growth
732 folding geometry, and thus a possible diapiric activity during the Jurassic with accumulation
733 of the Kimmeridgian in the enclosed minibasin. Cross-section S4 in Figure 4 tentatively shows
734 continuation of growth folding up to the Albian Black Flysch.

735 In the north-western part of the Tres Crouts structure, the outer flank shows a
736 truncation surface cutting eastward the Neocomian and Upper Jurassic down to the Dogger
737 dolomite and overlain by the upper Aptian Urgonian limestones (Fig. 12). This surface is
738 identified as a fault on the BRGM geological map (Casteras et al., 1970a), but from its
739 geometry and location above the weld we consider that it more likely represents a Neocomian
740 erosional surface analog to the one at Moncaut (Fig. 6). In the north-eastern part, the Black

741 Flysch lies on another erosional surface that truncates the northern flank of the weld eastward
742 from the Urgonian limestones down to the Dogger dolomite, thus merging with the
743 Neocomian erosional surface (Fig. 12). East of a N-S trending transverse fault, the flysch is in
744 contact with the Liassic carbonates of the southern flank. Owing to the divergent polarity of
745 bedding, we interpret this contact as the eastern extension of the weld. More to the east, the
746 northern and southern welds merge in a north-vergent thrust weld extending eastward to the
747 CBB extremity (Fig. 2). Above the erosional surface and against the eastern part of the
748 northern weld, the Black Flysch comprises at its base local intercalation of conglomerate
749 reworking the underlying carbonates with up to tens of cm-sized clasts. The conglomerate is
750 not as continuous as shown on the BRGM map (Casteras et al., 1970a) and shows variable
751 clast composition: it comprises mainly Urgonian limestone clasts and rare small Dogger
752 dolomite clasts where it rests above the Urgonian limestones, and mainly Jurassic clasts with
753 subordinate Urgonian clasts against the weld, demonstrating a local origin and short transport
754 (Cloix, 2017).

755 Hence, the Tres Crouts structure results from a multi-stage evolution. Salt
756 diapirism likely begun during the Jurassic, with the Dogger dolomite depositing above a
757 piercing salt wall in the NE, and the Kimmeridgian accumulating in a minibasin enclosed in the
758 polygonal diapiric structure. The erosional surfaces record continuing salt rise during the
759 Neocomian and the late Aptian-Albian. Later deformation resulted in further fold tightening,
760 welding and local weld reactivation as a thrust. Jurassic diapirism in the CBB is not described
761 in previous works which consider that the Jurassic tectonics was controlled by N-S trending
762 faulted blocks while diapirism begun in Early Cretaceous times with development of E-W
763 trending faults (James and Canérot, 1999; Canérot et al., 2005). A few km south-east of Tres
764 Crouts, submarine slides in the Kimmeridgian of the Pibeste ridge (James et al., 1996) attest

765 to tectonic activity in the eastern CBB at that time, and the structure of the Tres Crouts
766 Kimmeridgian suggests that this activity may involve salt wall rising. The Saint-Pé de Bigorre
767 ridge, located close to the north of the Tres Crouts structure, is an E-W trending squeezed
768 diapir (cross-section S4 in Fig. 4, and Fig. 12) with a complete Jurassic-Lower Cretaceous
769 succession but with all stratigraphic intervals featuring a reduced thickness, also suggesting
770 continuous salt wall rise throughout the Jurassic and Early Cretaceous.

771 The Tres Crouts metamorphic zone occurs in the central minibasin where it
772 affects the Kimmeridgian limestones in an area about 2 X 1 km (Fig. 12), characterized by
773 marbles with scapolite crystals (replaced by calcite, quartz, chlorite and white mica) (Casteras
774 et al., 1970a; Villard, 2016). In this zone, Raman spectroscopy of carbonaceous material
775 yielded peak paleo-temperatures up to 470-490°C (Villard, 2016; Ducoux, 2017). The borders
776 of the high temperature zone are narrow (200-400 m) with very high thermal gradient, up to
777 30°C/100 m (Villard, 2016), passing in the rest of the Tres Crouts structure to temperatures
778 about 340-370°C analogous to the most common temperatures recorded elsewhere in the
779 CBB (Menant et al., 2016; Corre, 2017). The structural location of the metamorphic zone and
780 its very high lateral gradient of temperature attest to a thermal anomaly probably related to
781 fluid flow in the diapiric structure. The thermal event is not dated but it ought to be relatively
782 late in the salt tectonic history as the Pyrenean Cretaceous metamorphism was related to the
783 blanketing effect of the Upper Cretaceous flysch (Clerc et al., 2015).

784

785 *4.3. The Pyrenean inversion: The North-Pyrenean Frontal Thrust and the Licq fault*

786

787 *4.3.1. The North-Pyrenean Frontal Thrust*

788 At the eastern end of the CBB, the NPFT is marked by the Ossun diapir (Fig. 2), a
789 salt ridge formed during the Mesozoic extension and reactivated and squeezed during the
790 Pyrenean inversion -but still up to 800 m wide- (Casteras et al., 1970a; Canérot et al., 2005).
791 Westward, the diapir passes laterally to a thrust that juxtaposes the limbs of two synclines,
792 with the Black Flysch of the Forêt de Mourle syncline in the hanging wall and the Cenomanian
793 flysch in the reverse limb of the footwall syncline (Fig. 2 and cross-section S4 in Fig. 4). There
794 is no exposure of Keuper facies along the thrust, but lenses of Keuper anhydrite have been
795 drilled between 85 and 154 m depth in the St-Vincent well (see Supplementary Data, Fig. S1).
796 The juxtaposition of two syncline limbs as well as the occurrence of Keuper lenses strongly
797 suggests that the NPFT emerging between the Ossun diapir and the Rébénacq transverse
798 diapiric structure (Fig. 2) corresponds to a former salt wall that extended all along the eastern
799 part of the CBB and was squeezed and welded during the Pyrenean compression. The
800 erosional truncations at the base of the Upper Cretaceous and base of the Lower Cretaceous
801 north of the Grand Rieu High (cross-section S4 in Fig. 4, and Supplementary Data, Fig. S2) result
802 from uplifting of the northern flank of the diapir during the extensional period, similarly to the
803 model of Canérot et al. (2005) for the Ossun diapir a few km to the east of our cross-section.
804 The km-scale progressive unconformity that involves the Campanian-Maastrichtian flysch and
805 overlying Neogene strata in the thrust footwall suggests that diapir squeezing was activated
806 from the beginning of the Pyrenean convergence and that the thrust front remained localized
807 on the same structure until the Tertiary, without noticeable northward propagation. The
808 lenses of Paleozoic rocks existing at the hanging wall of the emerging thrust are interpreted
809 as short-cuts of faulted blocks of the southern edge of the Grand-Rieu High.

810 By analogy with the eastern segment, the blind segment of the NPFT thrust west
811 of the Rébénacq transverse diapiric structure can also be interpreted as a welded salt wall.

812 This salt wall may have been the feeder of the Lasseube diapir, a 10 km-long salt sheet
813 enclosed in the lower Eocene a few km east of cross-section S1 in Figure 4 (Fig. 2). We show
814 in cross-section 1 a diapir in the subsurface in a position equivalent to that of the Lasseube
815 diapir, as well as hypothetical Keuper facies lens pinched along the NPFT. Similarly to the eastern
816 segment, the western NPFT remained active during most of the Pyrenean convergence and
817 only a limited displacement (about 1 km) was transferred more to the north in the Pau
818 anticline during the Eocene.

819

820 *4.3.2. The Licq fault*

821 At the south-western border of the CBB, the Licq fault juxtaposes the Keuper
822 facies of the Layens-Ourdinse salt wall with the Albian Mendibelza conglomerates of the Flysch
823 Noir Group on the southern side (cross-section S1 in Fig. 4). The fault dips steeply southwards
824 and the Layens succession appears as an overturned footwall flat, with the Keuper layer a few
825 tens of meters thick. The Licq fault extends along tens of km westward at the southern border
826 of the Mauléon basin with the same geometry and the Mendibelza conglomerates exhibit
827 there southward dips and a ramp against the fault represented above the topographic surface
828 in cross-section S1. The apparent footwall flat in the Keuper level, as well as the occurrence of
829 several diapiric structures adjacent to the Licq fault in the southern Mauléon basin lead to
830 consider the fault as a squeezed salt wall with a corresponding flap in its northern flank (James
831 and Canérot, 1999; Teixell et al., 2016; García-Senz et al., 2019). Similarly to the NPFT, this
832 interpretation will be discussed in the frame of the geodynamic model presented below.

833

834 **5. Discussion: Hyperextended rift geodynamics and salt tectonic evolution**

835

836 In this section, we integrate the CBB salt structures described in the previous
837 sections in a basin-scale model of hyper-extended Cretaceous rifting and inversion by the
838 Pyrenean orogeny. The model is illustrated in Figure 14, based on the sequential restoration
839 of the balanced cross-section S1 in Figure 4.

840

841 *5.1. Geological basis for basin-scale restoration*

842

843 The sequential restoration in Figure 14 is based on the following essential
844 geological features and inferences deduced from them.

845 1) The CBB Mesozoic cover behaved as a continuous lid, pierced by diapirs with
846 openings not exceeding a few km wide. In the western CBB, our mapping shows the structural
847 continuity across the Lourdios syncline, and the observed lateral terminations of the Sarrance
848 anticline suggest a limited width of the diapir opening at the origin of the salt extrusion. The
849 lack of exposure precludes determining opening of the former Mail Arrouy diapir, but the
850 lateral termination of the Mail Arrouy thrust east of the Ossau valley (in the Moncaut anticline;
851 Figs. 2 and 6) also suggests a limited opening. The map structure of the eastern CBB shows
852 continuity across the whole belt with lateral termination of diapirs (Moncaut, Tres Crouts;
853 Figs. 2, 6 and 12).

854 2) As demonstrated by the structural location of most of the Iherzolite bodies
855 (Figs. 2, 4 and 7), the Mesozoic cover lid was detached on the Triassic evaporite layer and its
856 central part rested directly on an exhumed mantle tract (Jammes et al., 2009; Lagabrielle et
857 al., 2010, 2019a and b; Tugend et al., 2014; Teixell et al., 2016).

858 These features imply that the basement of the ancient continental margins had
859 a smooth-slope topography allowing for continuous cover décollement, i.e. steeply-dipping

860 normal faults were absent or had limited offsets not interrupting the continuity of the Triassic
861 décollement layer. Cover gliding also implies denudation in the proximal part of the margins.

862 3) The uppermost part of the margins could be affected by basement faults. To
863 the south, restoration of the Lakora thrust structure and the stratigraphy of the Mendibelza
864 and overlying Upper Cretaceous conglomerates and breccias observed to the west in the
865 neighbouring Mauléon basin argue for deposition against a basement fault scarp that
866 bounded the Iberian shelf domain (future Axial Zone) submitted to erosion (Durand-
867 Wackenheim et al., 1981; Boirie and Souquet, 1982). North of the fault scarp, the onlap
868 surface of the Mendibelza conglomerates corresponds to the basement top of the proximal
869 margin (i.e. necking zone) tilted to the north (i.e. basinward). Hence, this surface can be
870 interpreted as the denuded Triassic gliding surface along which the cover lid slid northward
871 (Teixell et al., 2016; Saspiturry et al., 2019). Following Teixell (1993), we infer that the future
872 Bedous Triassic unit was preserved from erosion or gliding in a fault block existing to the south.
873 To the north, the proximal European margin and shelf correspond to the Grand-Rieu horst
874 (Fig. 4)

875 4) Although a precise restoration of the basin dimensions is difficult, approximate
876 indirect estimations can be made. The footwall of the NPFT corresponds to the European
877 margin (Teixell et al., 2016, 2018), suggesting a width of around 38 km for the latter.
878 Uncertainty is greater for restoring the Iberian margin basement, as it is presently the
879 substratum of the CBB whose internal structure and hence the amount of Pyrenean
880 shortening, remain unknown. In the lack of data, we assume a relative symmetry with respect
881 to the European margin, being aware that this uncertainty impacts the estimation of the width
882 of the denuded upper margin and ultimately of the total Pyrenean shortening (see below
883 Section 5.3.4). Based on the location of the Sarrance and Mail Arrouy lherzolites, a minimum

884 width of 17 km is inferred for the exhumed mantle domain (Teixell et al., 2016). Using this
885 value implies that the cover lid glided from both margins, with the southern CBB (Layens)
886 resting on the distal Iberian margin and the northern CBB (Oloron) resting on the European
887 distal margin, with a wide denudation of both upper margins.

888

889 *5.2. Rift basement structure*

890

891 Different models of crustal deformation have been proposed for the Cretaceous
892 rifting in the CBB domain and neighbouring Mauléon basin. Teixell et al. (2016) and Asti et al.
893 (2019) adapted to the CBB the model of ductile crustal extension first proposed by Clerc and
894 Lagabrielle (2014) and Clerc et al. (2016) for the eastern and central Pyrenees, and recently
895 generalized as a “smooth-slopes basin” model to various peri-Iberian Cretaceous rifts by
896 Lagabrielle et al. (2020). Teixell et al. (2016) restoration provides a basement top geometry
897 which satisfies the basic geological features listed above.

898 Masini et al. (2014) proposed for the Mauléon basin an alternative model
899 inferring a north-dipping intracrustal detachment emerging to the south in normal faults
900 bounding tilted basement blocks (Jara-Arbailles and Mendibelza). Gomez-Romeu et al. (2019)
901 recently transposed this interpretation in an isostatically compensated kinematic modelling
902 of the CBB transverse. Similarly to the Teixell et al. (2016) interpretation, their model shows
903 continuity of the CBB Mesozoic lid and its gliding from the European margin on a 19 km wide
904 exhumed mantle domain, associated to a wide denudation of the upper margin. However, the
905 model also implies tilted basement blocks with attached Jurassic cover south of the CBB which
906 do not conform with the thin geometry of the Paleozoic in the Iguntze and Bedous-Laruns
907 thrust units, as well as with the northward tilting of the basement top in the former unit.

908 Saspiturry et al. (2019) give a new interpretation of the Mauléon basin which
909 accounts for the structure of the Iberian proximal margin, via a south-dipping intra-crustal
910 detachment causing the northward tilting of the Mendibelza onlap surface in Albian times,
911 then a new northward detachment explaining the southward-tilting of the Arbailles block
912 during the Late Cretaceous. However, such a flip-flop evolution remains hypothetical and the
913 model does not show denudation of the European margin.

914 Discussing further a crustal-scale model of rifting is out of the scope of this paper.
915 The evolutionary model used here uses a basement top geometry analogous to that of Teixell
916 et al. (2016), adapted to the observed geological features of the CBB. It does not intend to
917 preclude any alternative mechanism of crustal deformation as long as the basic geological
918 constraints are satisfied.

919

920 *5.3. Salt tectonics and basin evolution*

921

922 *5.3.1. Late Triassic to Aptian extension*

923 The Jurassic to Aptian corresponds to a period of low subsidence with
924 accumulation of ca. 1.5 km thick platform carbonates above the salt-bearing Triassic unit (Fig.
925 14, stage 1). The original thickness of the latter is unknown due to high salt mobility and
926 probable salt dissolution at exposed diapirs (James and Canérot, 1999; Canérot et al., 2005),
927 but the observed development of salt tectonics and large Keuper volumes remaining in the
928 present profiles argue for a thickness of several thousands of meters. It must also be noted
929 that, although the top Triassic Keuper facies is commonly reported as the main salt
930 detachment or diapiric unit in the Pyrenees, the occasional, but frequent existence of Middle
931 Triassic Muschelkalk limestones in the Chaînons Béarnais ridges and other Pyrenean diapirs

932 (e.g. Canérot et al., 2005; Ortí et al. 2017) suggests that the basal detachment level must
933 stratigraphically underlie the limestones. Canérot et al. (2005) proposed that it may
934 correspond to a shale level. Alternatively, it may correspond to sub-limestone evaporites such
935 as those reported from the subsurface of the Ebro basin (“middle Muschelkalk” evaporites,
936 including halite; Jurado, 1990). Hence, it is likely that the highly disrupted Muschelkalk
937 carbonate lenses in the Chaînons Béarnais were originally interbedded in shale/evaporite
938 deposits, and hence the décollement layer illustrated in our restoration is not restricted to the
939 Upper Triassic Keuper but also includes the Middle Triassic. In our restoration, an original
940 thickness of the Middle-Upper Triassic about 3000 m is deduced from balancing constraints.
941 This is consistent with the fact that the NPZ corresponds to a branch of the peri-Iberian Triassic
942 rift system which concentrated thick evaporite deposition (Soto et al., 2017; Lagabrielle et al.,
943 2020, and references therein). Curnelle (1983) reported up to 900 m of salt-bearing facies in
944 little deformed areas of the Aquitaine basin north of the CBB, while Espurt et al. (2019) show
945 an interpreted seismic profile with up to about 3000 m of Middle-Upper Triassic on an
946 undeformed tilted block of the south-central Aquitaine basin.

947 Some authors contemplate incipient rise of salt anticlines during the Jurassic in
948 the Aquitaine Basin (Mediavilla and Mauriaud, 1987; Serrano et al., 2006). We consider likely
949 that basement faulting triggered salt movements in the thick Triassic layer of the CBB during
950 this time, because as discussed above, the Tres Crouts and Saint-Pé-de-Bigore salt structures
951 suggest initial salt mobility during the Jurassic. Hence, our cross-sections in Figure 4 and
952 restoration in Figure 14 assume a slight thickness increase of the Jurassic in salt synclines.
953 Diapir rising is more evident during the Neocomian period, with erosion of the Jurassic and
954 depositional hiatus of the Neocomian on uplifted areas that can be narrow salt wall crests
955 (Moncaut, Aran and Béon anticlines) or a wider salt massif in the southern CBB. At least part

956 of the Neocomian salt walls of the CBB pierced to the surface and were overlapped by Aptian
957 deposits (Asasp, Aran, Béon), similarly to many coeval diapirs in the Aquitaine Basin
958 (Mediavilla and Mauriaud, 1987; Canérot et al., 2005; Serrano et al., 2006). Salt wall rising and
959 minibasin subsidence accelerated during the latest Aptian (Clansayesian), contemporaneously
960 to the drowning of earlier carbonate platforms and the individualization of deeper
961 depocenters. This is marked by the onset of the Spicule marls deposition, probably traducing
962 acceleration of basement stretching and crustal thinning. Carbonate sedimentation persisted
963 on the rising diapir crests until their definitive drowning in the earliest Albian (Fig. 14, stage
964 1).

965

966 *5.3.2. Albian climactic rifting*

967 The Albian-early Cenomanian period is considered as the main rifting period in
968 the NPZ, marked by the deposition of the upper part of the Spicule marls followed from the
969 late early Albian by the thick sequences of the Black Flysch Group (Souquet et al., 1985;
970 Debros, 1987, 1990). Both hyper-extended continental margins with smooth basement top
971 were formed and the mantle was exhumed in between (Fig. 14, stage 2). Down-margin gliding
972 of the cover succession resulted in the denudation of the upper margins and separation of the
973 diapirs from the basement faults on which they initiated. Eventually, they were placed on top
974 of the exhumed mantle where the Triassic salt detachment collected the peridotite blocks.

975 On the Iberian side, the sedimentary cover of the southern shelf (the future Axial
976 Zone) was eroded, and incision down to the Paleozoic fed the Mendibelza conglomerates in
977 alluvial fans onlapping up-margin the basinward tilted and denuded basement top (Boirie and
978 Souquet, 1982; Teixell, 1993; Saspiturry et al., 2019). South of the Mendibelza conglomerates,
979 in the Iberian shelf, we show the above-mentioned faulted block with preserved Keuper that

980 will become the Bedous Triassic thrust unit during the Pyrenean inversion (Teixell, 1993)
981 (cross-section S1 in Fig. 4). As an alternative, García-Senz et al. (2019) interpreted the Bedous
982 Triassic unit as a salt sheet expelled from the base of the Mendibelza conglomerates, resulting
983 in an hypothetical weld at the base of the conglomerate that we consider unlikely in the absence
984 of any remain of Keuper facies, Muschelkalk or ophite bodies which otherwise are common in
985 the Bedous unit as well as along the CBB welds (Fig. 13). No conglomerates equivalent to the
986 Mendibelza are known along cross-section S1 on the European side and we postulate that the
987 basement of the Grand Rieu High was not largely exposed to erosion, as was the Axial Zone.
988 This does not exclude local exposure at some places, as it is suggested by a conglomerate
989 reworking Paleozoic clasts at the base of the Black Flysch in the northern limb of the Forêt de
990 Mourle syncline (Fig. 2).

991 Our field observations show that folding related to diapir rising and minibasin
992 subsidence by salt withdrawal was very active during deposition of the Spicule marls and Black
993 Flysch. Restoration indicates that the Barescou depocenter corresponds to the axis of the
994 Albian basin, with the maximum thickness (up to ca. 3500 m) of Spicule marls and Black Flysch,
995 culminating with the Barescou turbidites. This location of the basin axis was already proposed
996 in the frame of a former continental rift model (e.g. Canérot et al., 1978; Combes et al., 1998).
997 Here, it is consistent with the inferred position and width of the exhumed mantle domain
998 discussed above.

999 The lower Cenomanian unconformity which marks the end of the maximum
1000 rifting stage in the NPZ is observed in the present section on the northern limb of the Asasp
1001 diapir and on the future Axial Zone, as well as, to the north, on the northern Grand-Rieu High
1002 and in the Arzacq-Tarbes basin (Canérot et al., 2005; Serrano et al., 2006; see also cross-
1003 section S4 in Fig. 4). While erosion in the south and north may be ascribed to an uplift of the

1004 rift shoulders, erosion in the deep basin is more difficult to interpret. It may traduce rising of
1005 the diapirs at the lower margin, but we see unlikely that it was related to a terrestrial emersion
1006 because the erosion surface is observed between two deep-water flysch units.

1007

1008 *5.3.3. Cenomanian to Santonian extension*

1009 Stage 3 in Figure 14 portrays the end of the Pyrenean rifting, before the onset of
1010 the Pyrenean inversion by late Santonian times. The calcareous Upper Cretaceous flysch units
1011 accumulated in the basin while carbonate platforms grew on the rift shoulders (the future
1012 Axial Zone and, north of our section, the Arzacq Basin, Biteau et al., 2006; Serano et al., 2006).
1013 The geodynamic context of the North-Pyrenean basin during the Cenomanian to Santonian
1014 period has been interpreted in different ways, either as transpressive, or continued
1015 extensional, or simply in post-rift subsidence (see discussion and references in Clerc et al.,
1016 2016, and Vacherat et al., 2016). Clerc et al. (2016) suggest from their study in the eastern NPZ
1017 that extension may have continued during this period, explaining post-metamorphic
1018 exhumation of the Mesozoic carbonates along the extensional detachments and
1019 conglomerate accumulation along some fault escarpments. This assumption is supported in
1020 the Mauléon basin by the conglomerate bodies that are deposited above the Mendibelza
1021 conglomerates up to the Santonian Ibarrodoia breccia, recording continued erosion of the
1022 Paleozoic of the future Axial Zone and its Upper Cretaceous platform limestone cover
1023 (Casteras and Souquet, 1970; Durand-Wackenheim et al., 1981; Teixell, 1993). Saspiturry et
1024 al. (2019) also interpret the stratigraphic organization of the Mauléon basin as indicative of
1025 continuous extension during the Late Cretaceous, with the formation of a new north-dipping
1026 intracrustal detachment and the lower Cenomanian unconformity as a consequence. Hence,
1027 our restoration shows a few km increase of mantle exhumation and margin extension

1028 compared to the previous stage. It also shows further sliding of the detached cover, resulting
1029 in increased width of the denuded domains at upper margins. Due to near welding at the base
1030 of the Barescou depocenter, the maximum thicknesses of Cenomanian to Santonian flysch
1031 units are accumulated laterally above the distal parts of the continental margins in association
1032 with withdrawal of the underlying Keuper and its extraction in open diapirs.

1033

1034 *5.3.4. Pyrenean inversion*

1035 Stage 4 in Figure 14 shows an intermediate stage of positive inversion of the
1036 North-Pyrenean basin at the end of the Cretaceous. Since the onset of convergence, the
1037 exhumed mantle has been subducted and the toe of the Iberian margin has begun to
1038 overthrust the European margin, initiating the NPFT. This stage corresponds to the proto-
1039 collision of Teixell et al. (2016), during which the crust recovered its original thickness before
1040 the full collisional overthickening at Eocene times. Due to basement shortening, the detached
1041 cover behaves as a thin-skinned pop-up structure climbing on both margins. Shortening of this
1042 sedimentary lid is achieved by fold tightening and diapir squeezing, mostly by contraction of
1043 the large diapiric zones of the denuded upper margins. Flexural foreland basins begin to form
1044 with the Upper Cretaceous flysch series onlapping both upper margins. The still reduced
1045 crustal thickness and continued thermal subsidence (Angrand et al., 2018) maintained most
1046 of the deformed belt in submarine conditions, but we cannot rule out the existence of local
1047 emerged reliefs as the one tentatively shown in the restoration resulting from contraction of
1048 the Licq diapir. On the European margin, the deposition of the Upper Cretaceous flysch
1049 concomitant to diapir squeezing resulted in the growth strata fan observed in the NPFT
1050 footwall. The flysch laps on the Paleozoic basement or Triassic on the southern part of the
1051 Grand-Rieu High and, more to the north, on an erosional surface which cuts the pre-

1052 Campanian succession (see Fig. 9 in Canérot et al., 2005) and fed the basal breccia of the flysch.
1053 We suggest that the tilt of the onlap surface may not be due to flexural subsidence alone, but
1054 also to the salt withdrawal and resulting welding of the cover above the Grand-Rieu High
1055 induced by the flysch accumulation, similarly to the scenario proposed by Canérot et al. (2005)
1056 for the eastern extremity of the CBB on the Ossun section (Fig. 2).

1057 Stage 5 in Figure 14 is the present structure after completion of the Pyrenean
1058 collision from late Eocene to earliest Miocene times. Squeezing of the diapiric zone of the
1059 upper European margin resulted in its welding to become the present NPFT, which remained
1060 active at least until the mid-Eocene. We interpret that diapir squeezing fed the Lasseube diapir
1061 forming a salt sheet covered by the basal Eocene strata. At the beginning of the Eocene,
1062 propagation of the cover décollement (possibly favored by welding of the NPFT at that time)
1063 resulted in the formation of the Pau anticline a few km northward (see cross-sections in
1064 Canérot et al., 2005, Lagabrielle et al., 2010, and Teixell et al., 2016). On the Iberian side, the
1065 Licq diapir was welded to form the present steeply-dipping fault between the Mendibelza
1066 conglomerates and the Black Flysch turbidites, explaining the strong facies contrast between
1067 the two formations now juxtaposed by the fault. Then, southward propagation of the CBB
1068 décollement detached the Mendibelza conglomerates and a slice of its Paleozoic substratum,
1069 the latter probably corresponding to the short-cut of the footwall of a former normal fault,
1070 thus initiating the Lakora-Larra thrust system which propagated into the northern Jaca basin
1071 and remained active until the Bartonian (Teixell, 1996; Labaume et al., 2016; Labaume and
1072 Teixell, 2018). Contrary to the NPFT where tectonic activity remained localized during the
1073 whole Pyrenean orogeny, the full collision of the European and Iberian margins during the
1074 Late Eocene-Oligocene resulted in the development of the basement thrust imbricate of the
1075 Axial Zone which uplifted and deactivated the Lakora-Larra thrust system (Teixell et al., 2016,

1076 2018). The whole CBB was tilted northward above the back-limb of the Axial Zone culmination,
1077 in the hanging wall of the NPFT.

1078 The restoration in Figure 14 implies 101 km of total Pyrenean shortening,
1079 corresponding to 75 km shortening in the North-Pyrenean wedge (including 17 km of
1080 subduction of exhumed mantle, and 1 km shortening on the Pau anticline north of the
1081 restored section) and 26 km shortening in the Axial Zone basement (Gavarnie, Broto and
1082 Guarga thrusts; Teixell, 1996; Teixell et al., 2016; Labaume and Teixell, 2018). Although these
1083 values suffer uncertainties resulting from those on the length of the pre-orogenic margins as
1084 discussed above, they are close to those deduced from kinematic modelling by Gomez-Romeu
1085 et al. (2019), and coherent with results of Espurt et al. (2019) who calculated 127 km
1086 shortening on the Nestes-Ainsa cross-section 80 km to the east, including about 12 km
1087 shortening in the Aquitaine basin north of the NPFT that is not comprised in our restoration
1088 of the CBB-Jaca cross-section. It is interesting to note that the latter implies less Axial Zone
1089 shortening than the Nestes-Ainsa section where the Iberian crustal prism is more developed,
1090 but more shortening in the northern Pyrenees. This is consistent with the fact that the
1091 Pyrenean rift was wider westward, i.e. toward the oceanic domain of the Biscay Bay, thus
1092 influencing lateral changes in the distribution of the shortening for similar values of total
1093 convergence, and explaining the more recent exhumation of the NPZ and emersion of the
1094 Pyrenean relief in the west (Vacherat et al., 2014; Bosch et al., 2016; Labaume et al., 2016;
1095 Teixell et al., 2016).

1096

1097 *5.4. Driving mechanisms of salt walls rising*

1098

1099 Restoration of the CBB structures shows a protracted history of salt ridge rising,
1100 probably beginning in the Jurassic and continuing throughout the rifting and subsequent
1101 Pyrenean inversion episodes. If the initial stages up to the Aptian can be interpreted as diapirs
1102 formed above basement normal faults (Canérot, 1988, 1989), conditions of loading changed
1103 when the diapirs were detached from their original basement during the mid-Cretaceous
1104 downslope gliding.

1105 A first consequence of the cover gliding is the denudation of the proximal
1106 margins, allowing for deposition of the Mendibelza conglomerates on the denuded basement
1107 top and Permo-Triassic sandstone tegument (Teixell et al., 2016; Saspiturry et al., 2019). We
1108 also assume that evacuation of Keuper facies under subsiding minibasins at the rift center
1109 together with extension resulted up margins in wide diapiric zones devoid of carbonate cover.
1110 To the south, such inflated zone may have separated the proximal domain of Mendibelza
1111 conglomerates from the more distal domain of flysch. Using the margin lengths discussed
1112 above, assuming that diapirs did not exceed 2-3 km width in the basin center, and length-
1113 balancing the known cover elements (Mendibelza conglomerates and CBB carbonates), we
1114 deduce a 14 km wide diapiric zone between the Mendibelza conglomerates and Layens ridge
1115 on the Iberian margin, and a 25 km wide diapir on the conjugate upper European margin.
1116 Comparable wide diapirs resulting from gravity sliding-induced denudation and/or extension
1117 have been inferred recently on the Brazilian passive margin (the “Albian Gap”; Jackson et al.,
1118 2015; Pichel et al., 2019).

1119 Furthermore, we infer that gliding of the cover lid down the margins most
1120 probably implied not only tectonic denudation of the proximal margins, but also shortening at
1121 the conjugate margin toes and over the central exhumed mantle where the slid mass
1122 accumulated. Jammes et al. (2010) discussed interactions between salt tectonics and

1123 extensional detachment systems in the Bay of Biscay and western Pyrenees. They conceptually
1124 oppose margins with post-rift salt where the deformation of the detached cover is purely
1125 gravitational, implying that contraction at the toe of the gliding cover compensates extension
1126 at the upper part (e.g. Brun and Fort, 2011), to margins with pre-rift salt layers as the Pyrenees
1127 where the salt decouples extensional deformation between the basement and cover. In the
1128 latter case, Jammes et al. (2010) argue that the cover may only feature extensional rafts
1129 because the basement and cover are extended coevally (and the salt in between is stretched),
1130 so the whole system is extended and does not need contraction. Although the Ibis anticline in
1131 the center of the Parentis basin (reported in Bois et al., 1997, and Masse, 1997) may have
1132 been related to gravitational contraction at Albian times, Jammes et al. (2010) consider that
1133 the extensional structures of the basin are not linked with compressional structures of the
1134 same age. Conversely, we consider that an important effect of gravity cannot be excluded in
1135 the slopes of actively stretching margins containing a pre-rift salt unit as the Pyrenees. If the
1136 continuity of the salt layer is not broken, the gravitational displacement can exceed tectonic
1137 stretching and hence contraction being caused by crowding at the basin axis coeval with
1138 extensional denudation upslope. In the case of the CBB, we emphasize that the salt ridge rising
1139 and folding restored at stages 2 and 3 in Figure 14 would be analogous to the fold and thrust
1140 system characterizing compressional diapirs at lower continental margins (Rowan et al., 2004;
1141 Brun and Fort, 2011; Jackson and Hudec, 2017). Figure 15 illustrates the close resemblance
1142 between the Sarrance anticline and adjacent Lourdios syncline, as well as the Tres Crouts
1143 structure (cross-sections S1 and S4 in Fig. 4), with compressional diapirs and adjacent pouch-
1144 like minibasins observed in the lower continental margin of Angola and obtained by analogical
1145 modelling (Brun and Fort, 2004).

1146 We indicate that some recent works propose that the contact of the Iherzolite
1147 and associated Paleozoic bodies with the overlying carbonates represent the original
1148 detachment contact between the exhumed mantle and the Mesozoic cover, inferring that the
1149 CBB anticlines correspond to late structures formed during the Pyrenean inversion (Lagabrielle
1150 at al., 2010, 2019a and b; Corre et al., 2016; Corre, 2017). This interpretation conflicts with
1151 the evidence that the CBB anticlines begun growing at least from the earliest Cretaceous, well
1152 before mantle exhumation during the Albian. Although we agree that the Iherzolite bodies
1153 were teared off the mantle along the detachment system described by the above-cited works,
1154 we believe more probable that the bodies outcropping at the upper part of anticlines
1155 (Sarrance, Moncaut, St-Pé-de-Bigorre) were dynamically pushed up 4-5 km to their present
1156 position by the salt ascent in the salt walls, similarly to peridotite bodies of equivalent size in
1157 the Sivas diapiric province (Kergaravat et al., 2017). The latter case corresponds to a foreland
1158 context where compression squeezed the diapirs and probably favored the uprise of dense
1159 bodies. In the CBB case, our view suggests that contraction may have acted not only during
1160 the Pyrenean compression but also earlier in the rifting context. As the Urdach Iherzolite
1161 shows evidence of being exposed to the sea floor in late Albian times (Fortané et al., 1986;
1162 Jammes et al., 2009; Debroas et al., 2010; Lagabrielle at al., 2010, 2019 b; Corre, 2017), we
1163 place the Sarailé Iherzolite high in the Sarrance salt wall in the Albian restoration of Figure 14.

1164

1165 **6. Conclusions**

1166

1167 The revisit of the Chaînons Béarnais salt structures presented in this paper and
1168 summarized in four new cross-sections and detailed mapping of key-areas emphasizes their
1169 relevance in the architecture and geodynamic evolution of the North-Pyrenean basin and

1170 subsequent thrust wedge. Our work shows that the fold-thrust structure of the Jurassic-
1171 Cretaceous sedimentary cover corresponds to salt-walls and minibasins detached and
1172 squeezed above a thick Middle-Upper Triassic evaporite-rich layer (here collectively referred
1173 to as Keuper facies). A long history of salt migration is responsible for the numerous
1174 sedimentary discontinuities of diverse ages reported in this segment of the Northern
1175 Pyrenees, and can be summarized as follows:

1176 1) Salt-walls most probably began to rise above basement normal faults during the
1177 early stages of Mesozoic extension. Following early salt movements during the Jurassic,
1178 erosions and hiatus mark the rising during the Neocomian of narrow pierced salt walls in the
1179 basin axis and a wider salt inflation in the southern basin margin.

1180 2) From the latest Aptian, the onset of margin hyperextension resulted in the
1181 acceleration of salt wall rise and minibasin subsidence with limestone shelves on ridge crests
1182 prograding toward deeper minibasins dominated by thick marl deposition. During the Albian,
1183 the main rifting stage in the North-Pyrenean basin, the limestone shelves were ultimately
1184 drowned, and the whole CBB cover glided down-margin in the domain of deep flysch
1185 sedimentation as the continental margins were pulled apart. The diapiric structures were
1186 disconnected from parent basement faults and the axial part of the cover lid was put on top
1187 of exhumed mantle, as indicated by lherzolite slices embedded in the Triassic Keuper.

1188 Meanwhile, the proximal margins were denuded, and, in the case of the southern
1189 shoulder, submitted to erosion that fed conglomerate sedimentation on the denuded
1190 proximal margin. Our margin restoration suggests that 15-25 km-wide diapiric zones may have
1191 occupied the denuded margins between the domain of conglomerate sedimentation and the
1192 slid CBB cover lid, analogous to the early Cretaceous denudation inferred in the modern
1193 Brazilian continental margin.

1194 3) Conglomerate sedimentation in the proximal Iberian margin suggests persistence of
1195 extension during the Cenomanian to Santonian. Our restoration suggests that the flysch
1196 accumulated during that period in the deep basin was thicker above the distal margins, due
1197 to withdrawal of previously inflated salt, than at the basin axis, where it was limited by primary
1198 welding of the central minibasin system on the exhumed mantle.

1199 4) From the onset of Pyrenean convergence in the late Santonian, the exhumed mantle
1200 was subducted and the Iberian margin was thrust above the European margin. The detached
1201 sedimentary lid then behaved as a pop-up structure climbing up-margins while the large
1202 diapiric zones of the denuded upper margins and the central salt walls were contracted and
1203 squeezed.

1204 Part of the salt-related folding before the Pyrenean orogeny could have been caused
1205 by gravity-driven contraction. The syn-sedimentary salt structures of the Mesozoic CBB show
1206 typical features of deep water contractional salt systems such as pouch-like minibasins, salt-
1207 wall welding, salt overhangs, and salt extrusions. The context of cover gliding during the
1208 Mesozoic extension suggests that the salt structures were affected by contraction at margin
1209 bottoms and above the exhumed mantle coeval with extension upslope, in analogy to what
1210 occurs on the gravity-dominated, Angola-type modern distal margins. However, at variance
1211 with those modern margins, the North-Pyrenean slid unit was detached above a pre-rift salt
1212 layer and formed a continuous lid between the two margins and the central basin, and
1213 probably part of the contraction was caused by crowding at the basin axis. This implies that
1214 part of the salt squeezing may have been achieved in the rifting stage and the Pyrenean
1215 inversion was only responsible for final shortening and northward tilting of the whole
1216 structure in the hanging wall of the NPZ.

1217 The salt tectonic style here described for the CBB is linked to the smooth topography
1218 of the continental margin slopes where the continuity of the décollement layer was not offset
1219 by major rift or thrust faults to prevent generalized décollement during the two tectonic
1220 phases. This shortening-by-gravity model was never applied to the Mesozoic Pyrenean basins
1221 and may find application in other segments of the belt and of the peri-Iberian basins (e.g.
1222 Parentis). The model involves formation of wide denuded domains up-margins, a concept that
1223 may be applied to other margins where gravity gliding on salt décollement layers is a major
1224 mechanism.

1225

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1227

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1240

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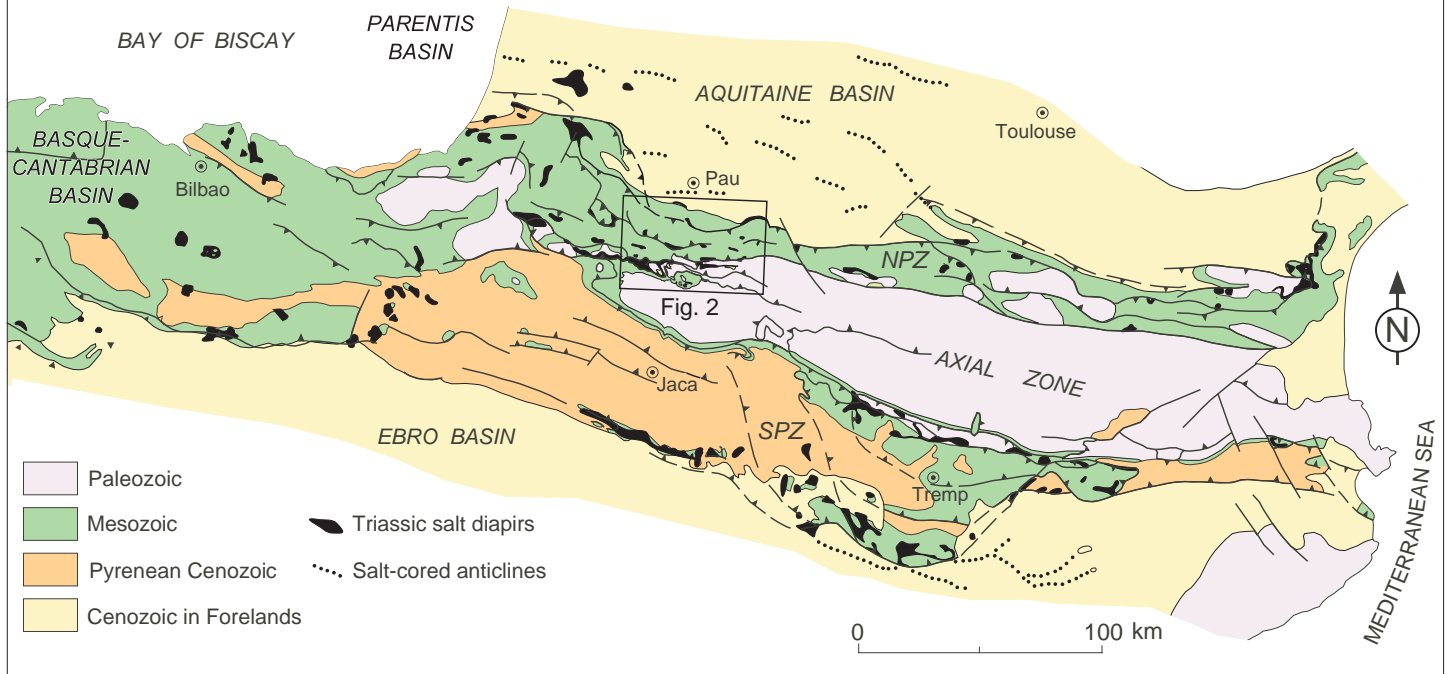


Fig. 1. Structural sketch map of the Pyrenees with location of the main outcrops of Triassic Keuper evaporitic facies. Map modified from Teixell (1996) and Saura et al. (2016). NPZ: North Pyrenean Zone; SPZ: South Pyrenean Zone.

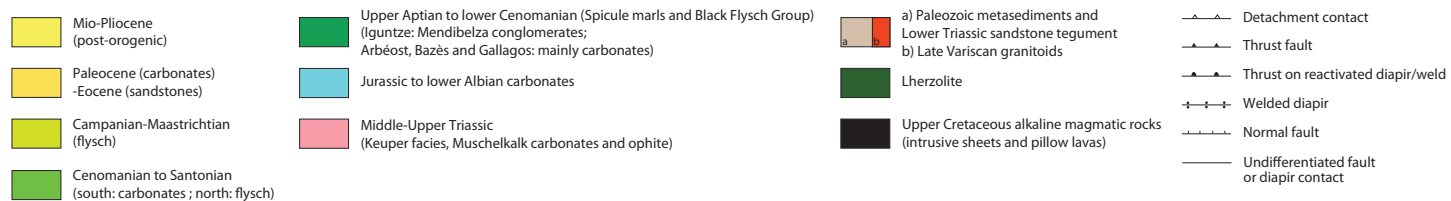
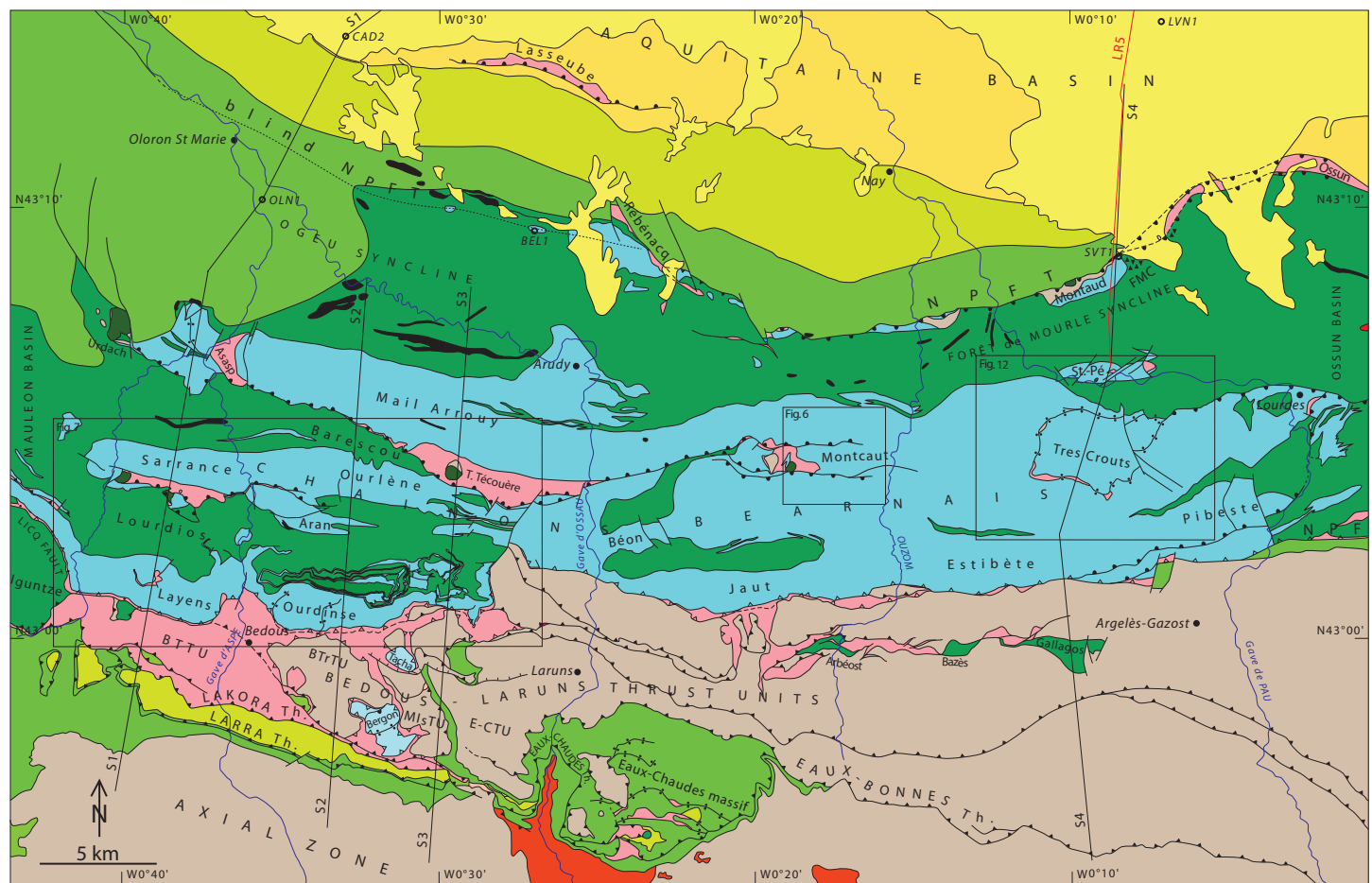


Fig. 2. Geological map of the Chaînons Béarnais belt (North-Pyrenean Zone) and adjacent areas. The so-called “Chaînons Béarnais” are the anticline and thrust ridges of Jurassic to lower Albian carbonates (in blue). BTrTU: Bois de la Traillère thrust unit; BTTU: Bedous Triassic thrust unit; E-CTU: Eaux-Chaudes fold-thrust unit; FMC: Forêt de Mourle Albian conglomerates; MlSTU: Montagnon d’Iseye thrust unit; NPF: North-Pyrenean Fault; NPFT: North-Pyrenean Frontal Thrust. S1 to S4: cross-sections in Fig. 4 (S1 extends 1.8 km north of the map boundary, in the area of post-orogenic sediment exposure). LR5: seismic profile in Supplementary Data, Fig. S2. Wells: BEL1, Béclair 1; CAD2: Cardesse 2; LVN1: Livron 1; OLN1, Oloron 1; SVT1: Saint Vincent 1 (see well stratigraphic logs in Supplementary Data, Fig. S1). Map sources: BRGM 1:50,000 scale geological maps (Alimen et al., 1963; Casteras et al., 1970a and b; Ternet et al., 1980, 2004), Souquet et al. (1985) and original mapping (in frames of Figs. 7 and 12).

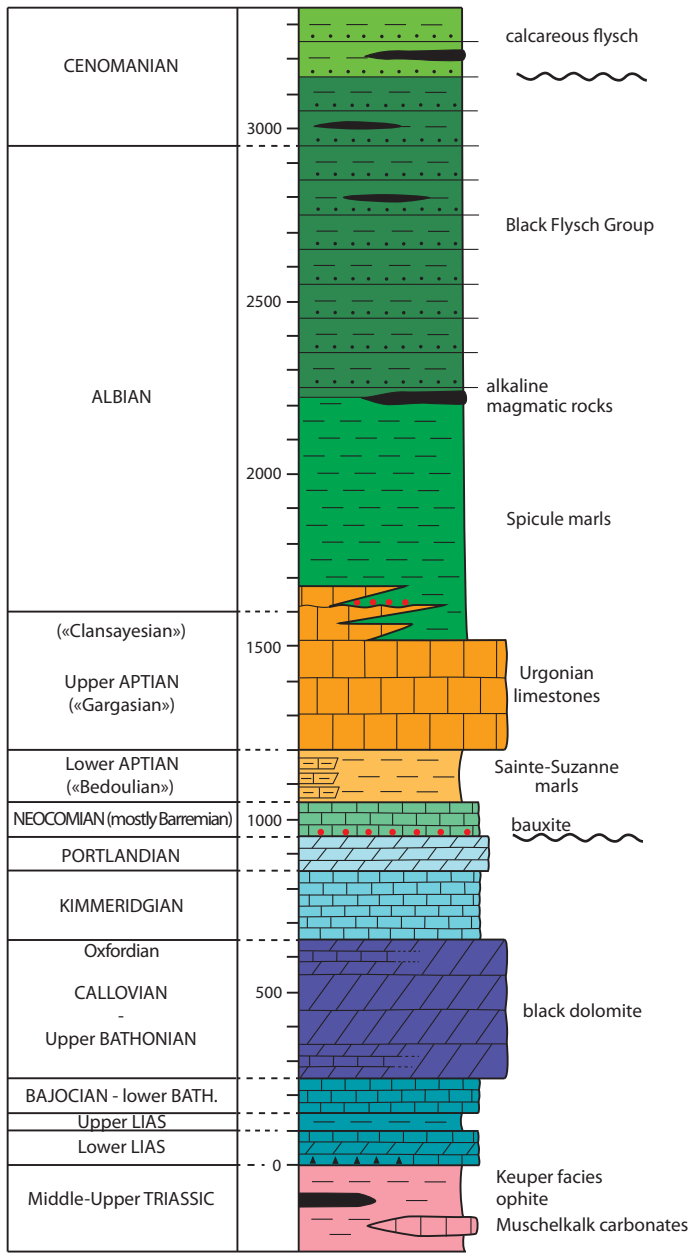


Fig. 3. Simplified stratigraphic log of the Chaînons Béarnais belt. Local unconformities are shown. Red dots: alterite horizons (including bauxite). Based on Casteras et al. (1970a and b), Souquet et al. (1985), Lenoble (1992), James (1998).

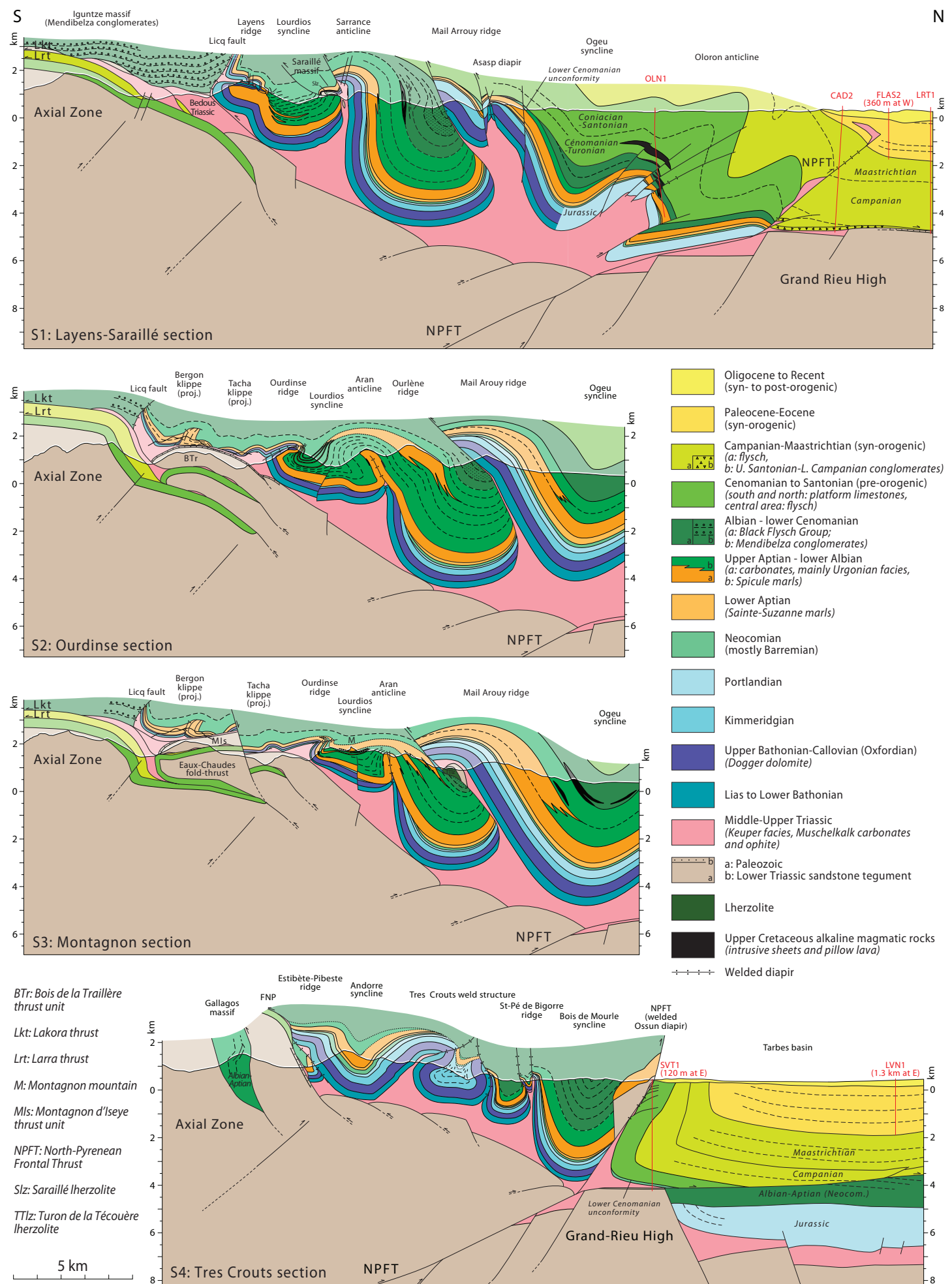


Fig. 4. Geological sections across the Chaînons Béarnais belt. See location in Fig. 2. FNP: North Pyrenean fault. Wells: CAD2: Cardesse 2; FLAS2: Lasseube 2; LRT1: Le Rouat 1; LVN1: Livron 1; OLN1: Oloron 1; SVT1: Saint Vincent 1 (see well stratigraphic logs in Supplementary Data, Fig. S1). Northern part of S4 cross-section includes interpretation of the regional seismic profile n°5 from Serrano et al. (2006) (see location in Fig. 2, and the interpreted profile in Supplementary Data, Fig. S2).

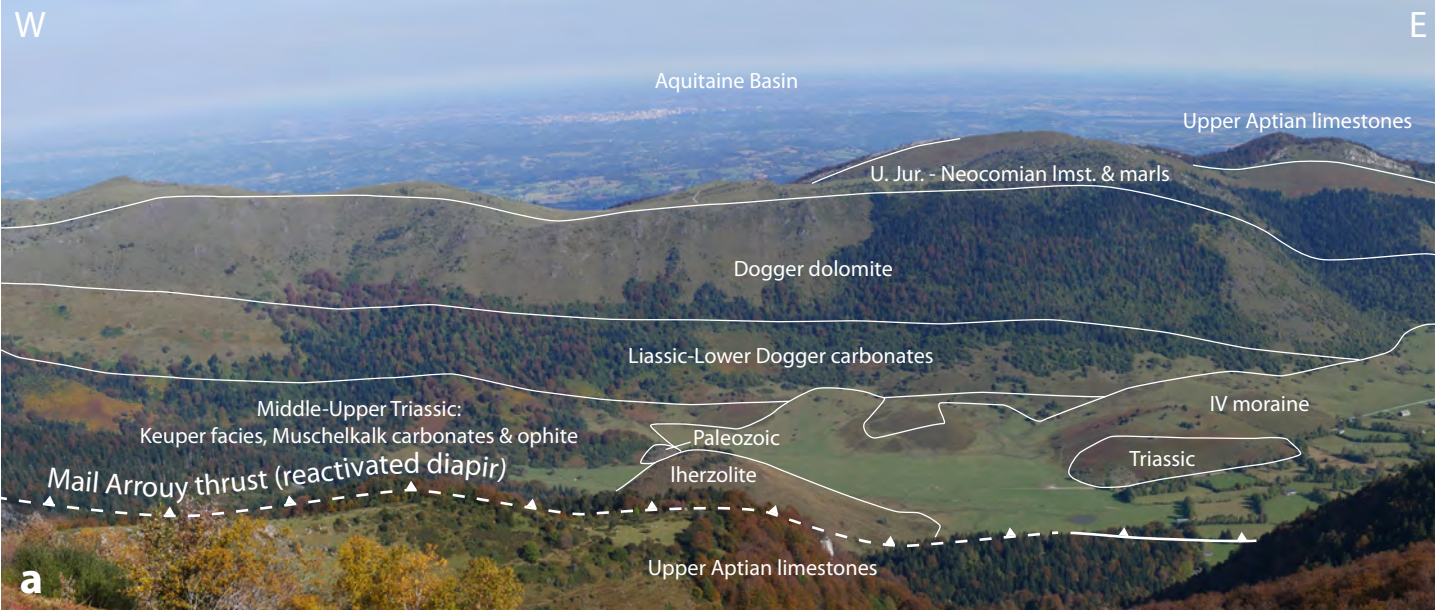


Fig. 5. Field images of the Chaînons Béarnais stratigraphic units. (a): Panorama of the Mail Arrouy carbonate ridge featuring the Middle-Upper Triassic complex and the Jurassic-Lower Cretaceous carbonate succession in the hanging wall of the Mail Arrouy thrust (cf. stratigraphic log in Fig. 3 and cross-section S3 in Fig. 4). The Triassic contains the Turon de la Técoùère Iherzolite and adjacent lens of Paleozoic metasediments. The unit is thrust southwards over the upper Aptian limestones of the Ourlène carbonate ridge (foreground). The Aquitaine Basin is in the background (with the Pau city in the middle part). Photograph taken from 703300E, 4769070N (UTM coordinates, zone 30T). (b): Syn-metamorphic bedding-parallel foliation in the upper Aptian Urganian facies limestones, featuring flattened bivalves (rudists; eastern Ourdinse ridge; 705380E, 4765900N). (c): Syn-metamorphic bedding-parallel foliation in the Albian flysch (Lourdios syncline; 689240E, 4768790N).

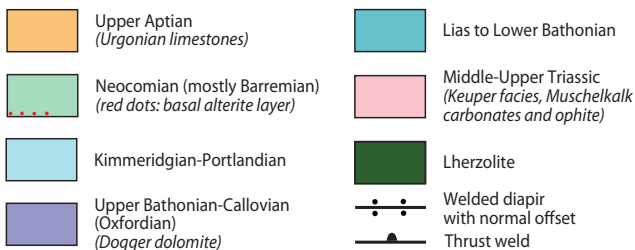
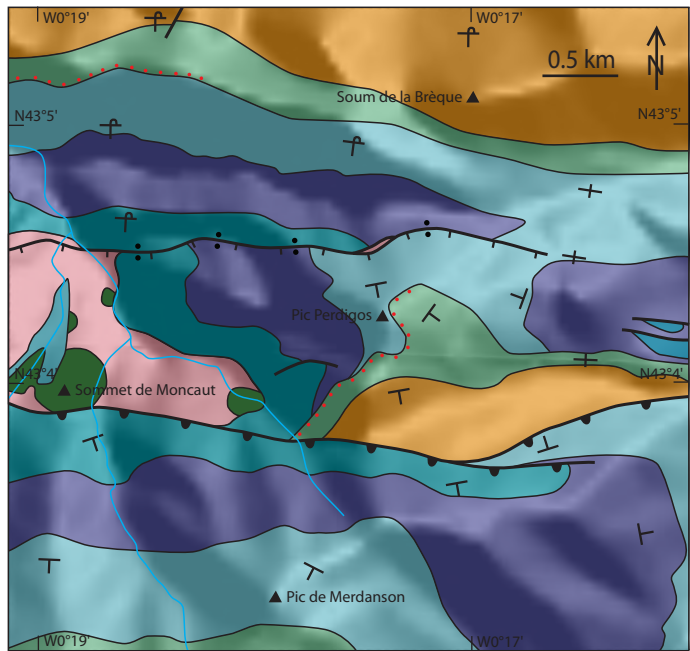


Fig. 6. Geological map of the Moncaut salt anticline showing the Lower Cretaceous unconformity on top of a former diapir crest (near Pic Perdigos). See location in Fig. 2. Map simplified from Casteras et al. (1970a).

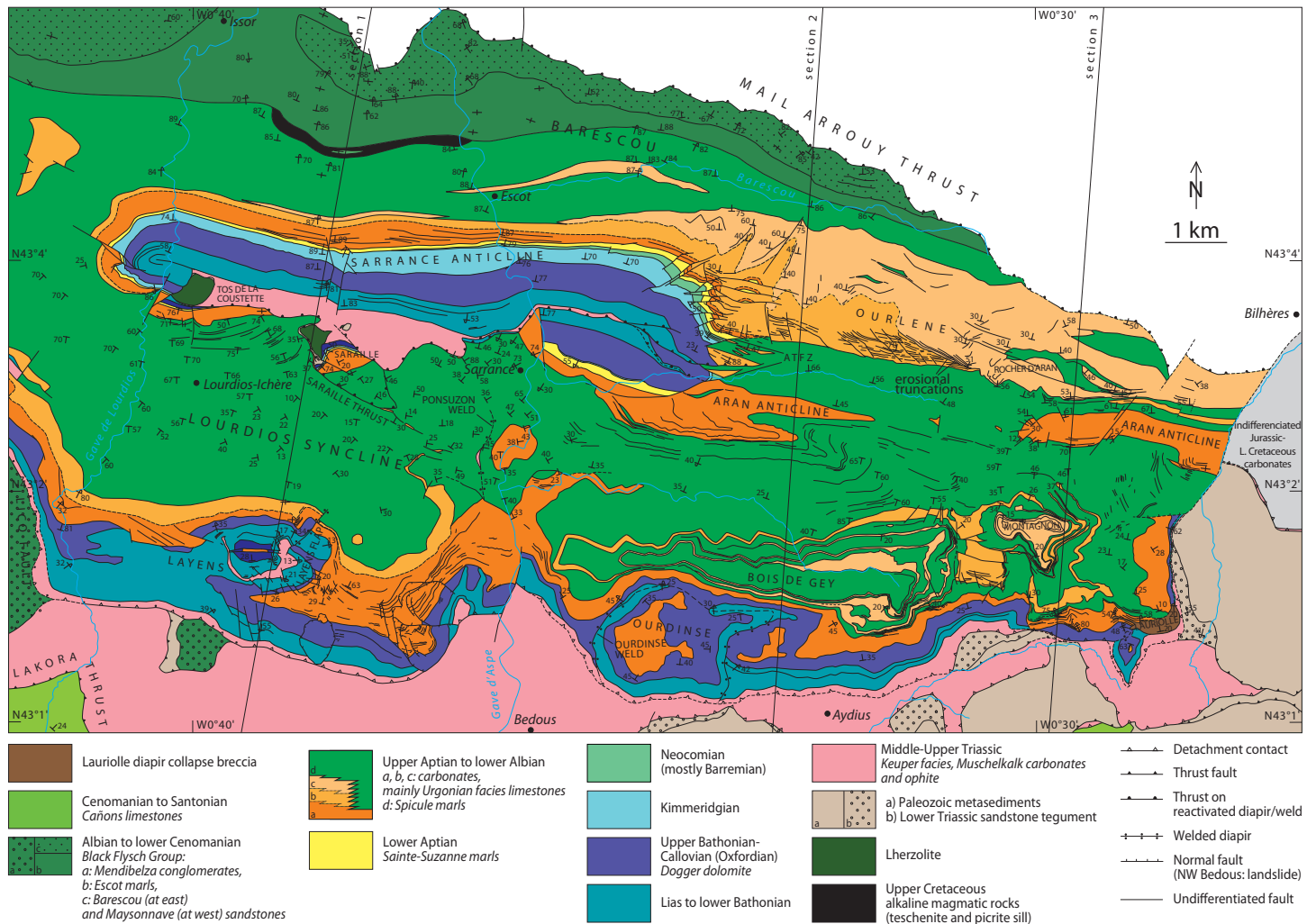


Fig. 7. Geological map of the Lourdios and Barescou minibasins and adjacent carbonate ridges (SW Chaînons Béarnais). See location in Fig. 2 and cross-sections S1 to S3 in Fig. 4. Map sources: mainly original mapping on the field and aerial photographs, with complements from Canérot (1964), Paris (1969) and Casteras et al. (1970b). The Urganian limestones are divided into three stratigraphic members. The lower interval corresponds mostly to the Gargasian and the two upper intervals to the Clansayesian, possibly reaching the lower Albian. However, the limits given to these members may not strictly correspond to time-lines across the whole map. ATFZ: Aran transverse fault zone.

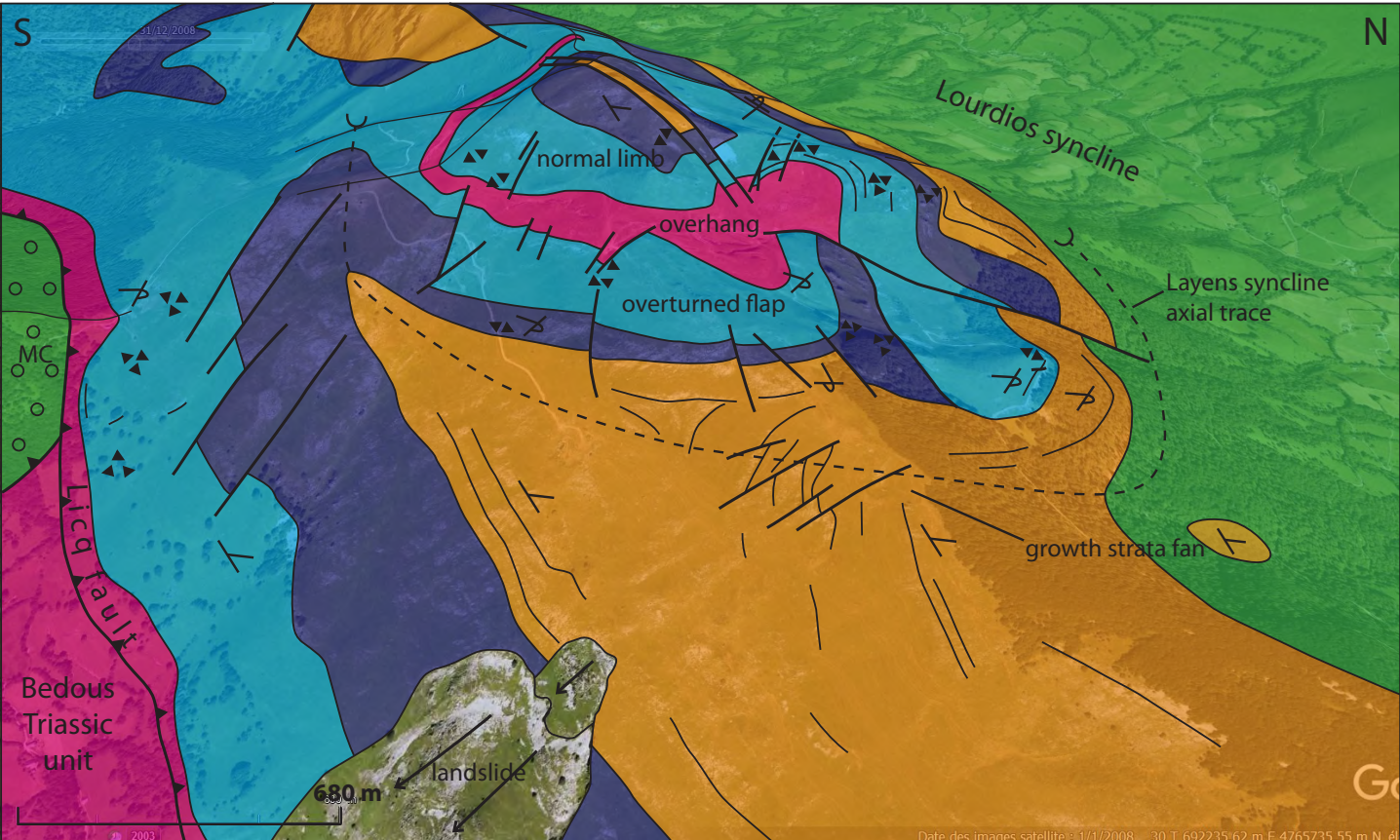


Fig. 8. Layens overhang at the southern edge of the Lourdios minibasin (Layens carbonate ridge). Google Earth oblique view from the SE. For lithostratigraphy see the stratigraphic log in Fig. 3. See location of the Layens carbonate ridge in Fig. 7, and cross-section of the diapiric structure in S1 in Fig. 4. The figure illustrates the curved axial trace of the recumbent syncline and strong thickness reduction of the Dogger dolomite and Urgonian limestones in the overturned flap and above-lying normal limb, due to the Neocomian erosion of the Dogger and to growth folding in the Urgonian, evidencing diapir rise during the Early Cretaceous. Diapir rising, collapse and erosion induced frequent brecciation in the Lias and Dogger (triangles), and steeply-dipping faults, probably related to diapir collapse, may correspond to welded diapir piercings.

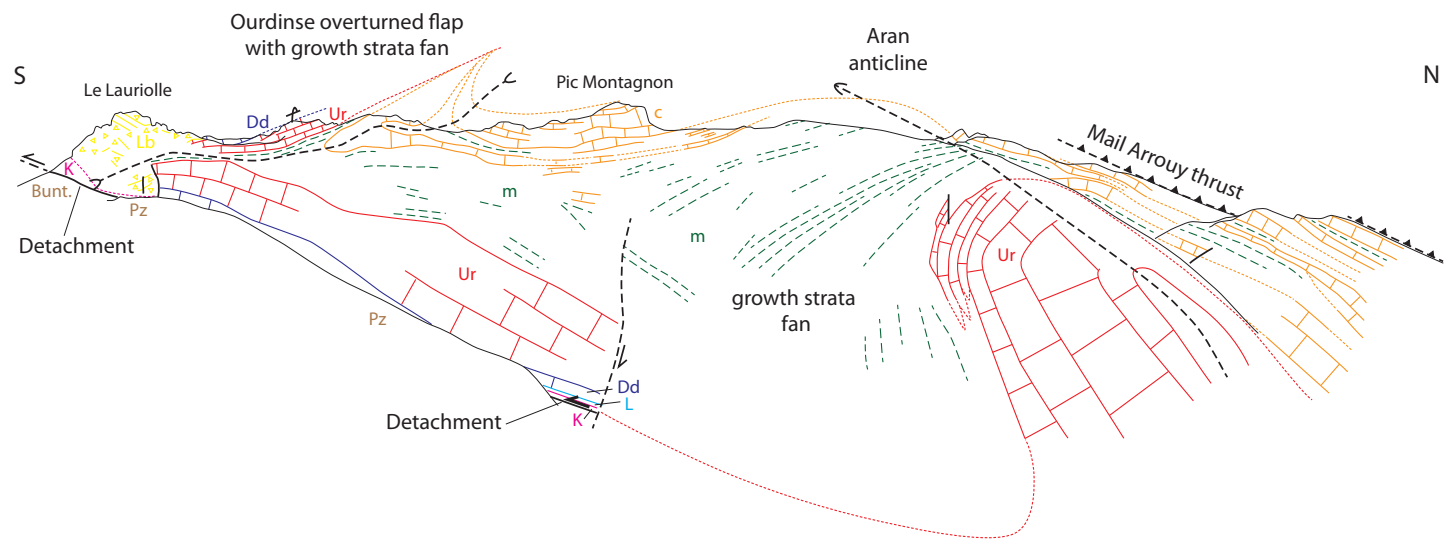


Fig. 10. Lourdios minibasin, Pic Montagnon panorama. Key: Bunt: Lower Triassic sandstone tegument (Buntsandstein facies), c: uppermost Aptian (Clansyesian) (to lower Albian?) carbonates (limestones with local dolomitization) intercalations in the marls (m), Dd: Dogger dolomite, L: Liassic limestones, Lb: Lauriolle breccia, K: Triassic Keuper facies, Pz: Paleozoic metasediments, Ur: upper Aptian limestones (Urgonian facies). See location of Le Lauriolle, Pic Montagnon and Aran anticline in Fig. 7, and cross-section of the structure in S3 in Fig. 4. View from 707538E, 4767470N. The figure illustrates the pouch-like geometry of the syncline minibasin with both limbs featuring overturned flaps and growth strata fans. The reverse fault at bottom may correspond to a tilted and deformed syn-sedimentary normal fault. At left, the Lauriolle breccia results from diapir collapse (Canérot and Lenoble, 1993; Cloix, 2017). The label “detachment” shows the contact of the Mesozoic cover above the Paleozoic and its Buntsandstein tegument, now a sub-vertical contact due to deformation along a transverse Pyrenean structure (the “Ossau thrust” of Canérot, 2017), possibly inherited from the Mesozoic extensional system.



Fig. 11. View of the northward progradation of the upper Aptian limestones at the southern edge of the Barescou minibasin (Rocher d'Aran, Ourlène carbonate ridge). S0/S1: bedding-parallel cleavage, S2: local late cleavage. See location of Rocher d'Aran and Ourlène carbonate ridge in Fig. 7, and cross-section of the structure in S3 in Fig. 4. View from 702533E, 4769493N. The figure illustrates the shelf edge progradation away from the rising crest of the Aran anticline located to the south. The contact between the limestones and underlying marls is affected by dolomitisation (dol.), similarly to most of the marl-limestone contacts on the northern limb of the Aran anticline (Canérot, 1964).

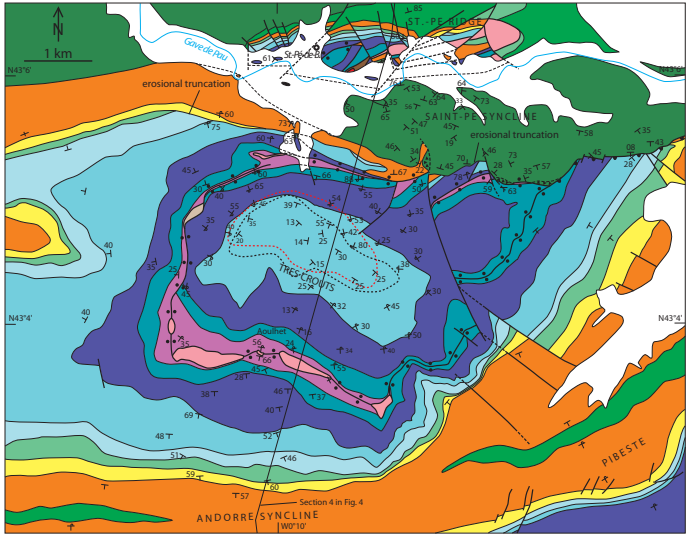


Fig. 12. Geological map of the Tres Croits polygonal weld structure (location in Fig. 2). See cross-section of the structure in S4 in Fig. 4. Map sources: BRGM 1:50,000 scale geological map (Casteras et al., 1970a), Lanusse (1969), Cloix (2017) and original mapping. The central metamorphic area is marked by marble after the Kimmeridgian limestone, with scapolite crystallization and maximum temperatures recorded up to close to 500°C (Villard, 2016; Ducoux, 2017).

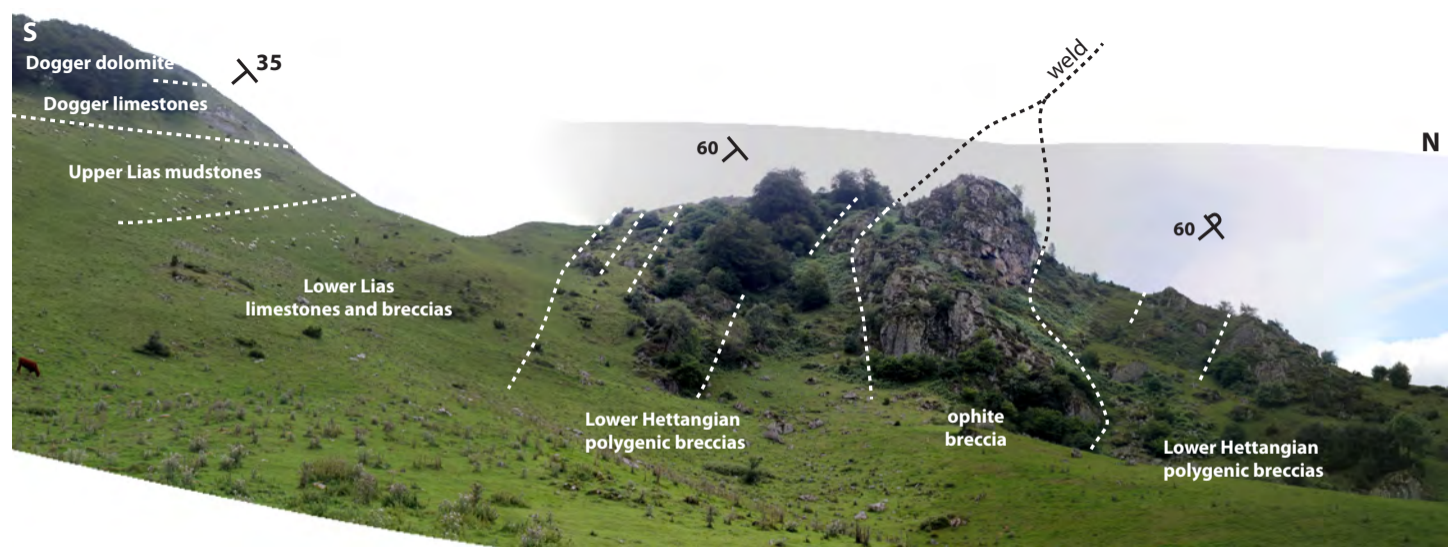
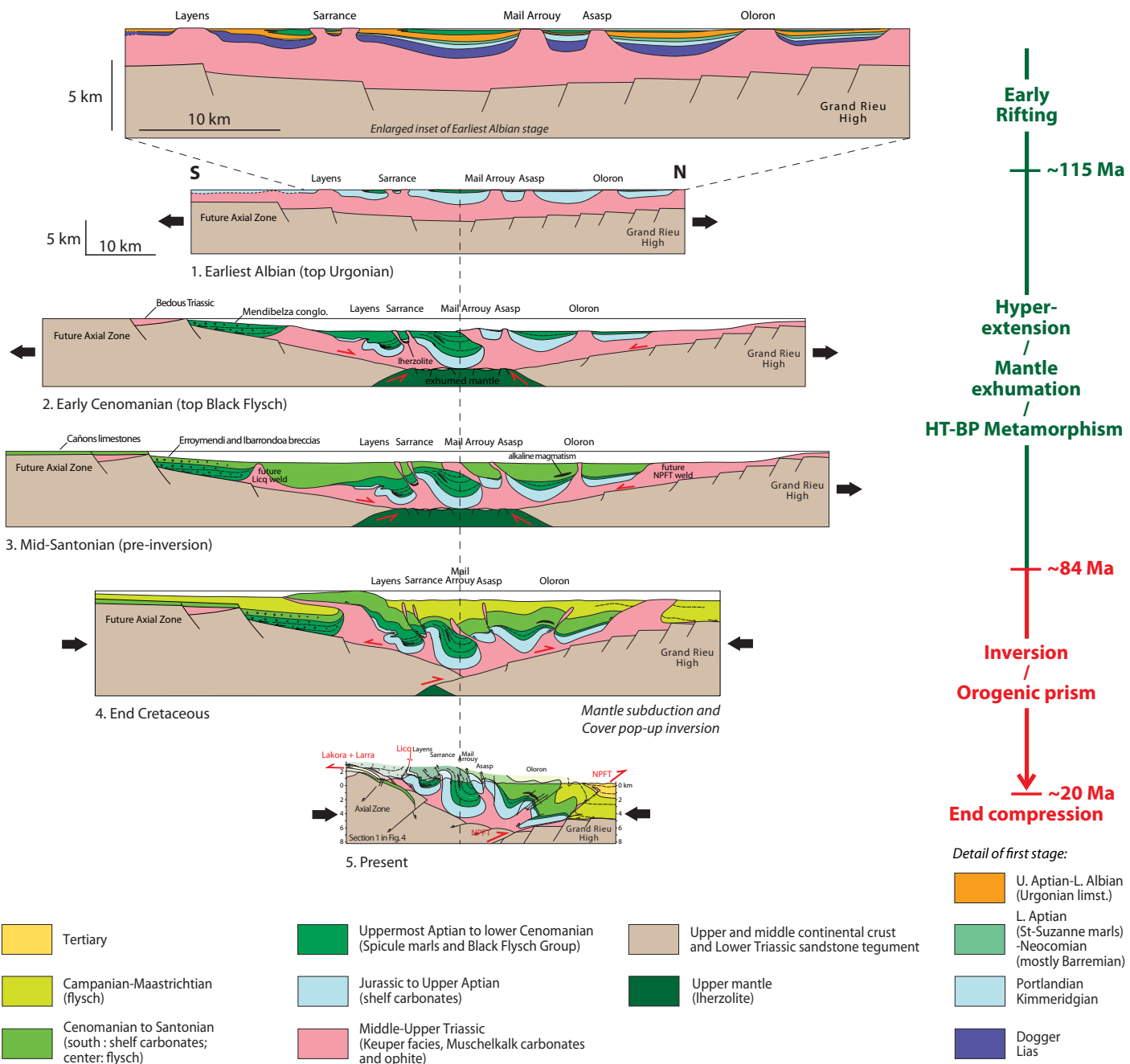


Fig. 13. Panorama of the Tres Crouts weld (southern segment at Aoulhet). See location of Aoulhet in Fig. 12 and cross-section of the structure in S4 in Fig. 4. View from 730728E, 4771609N. A hundred-meter scale block of brecciated Triassic ophite is pinched along the welded base of the Jurassic succession.



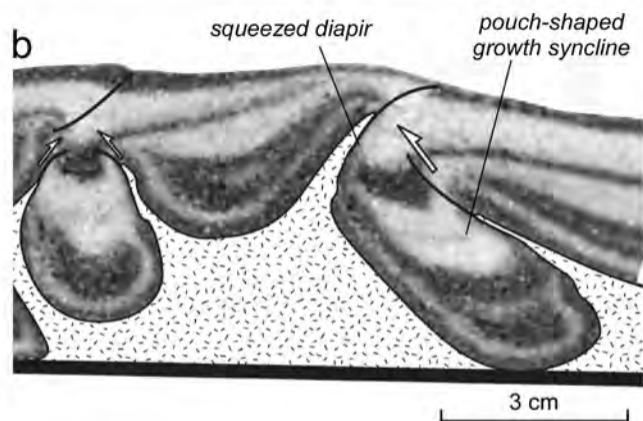
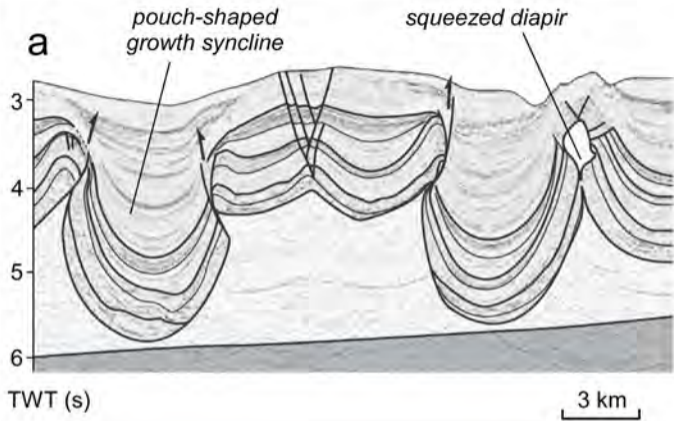


Fig. 15. Examples of (a) a seismic image of the gravity-driven contractional deformation in the Angolan margin and (b) a sandbox model of a contractional salt system (after Brun and Fort, 2004). Note the squeezed diapirs and the intervening, pouch-shaped growth synclines with irregular amplitude and wavelength that constitute good analogues for the Chaînons Béarnais salt structures shown in Figs. 4 and 14.

Section 1 in Figure 4

Section 4 in Figure 4

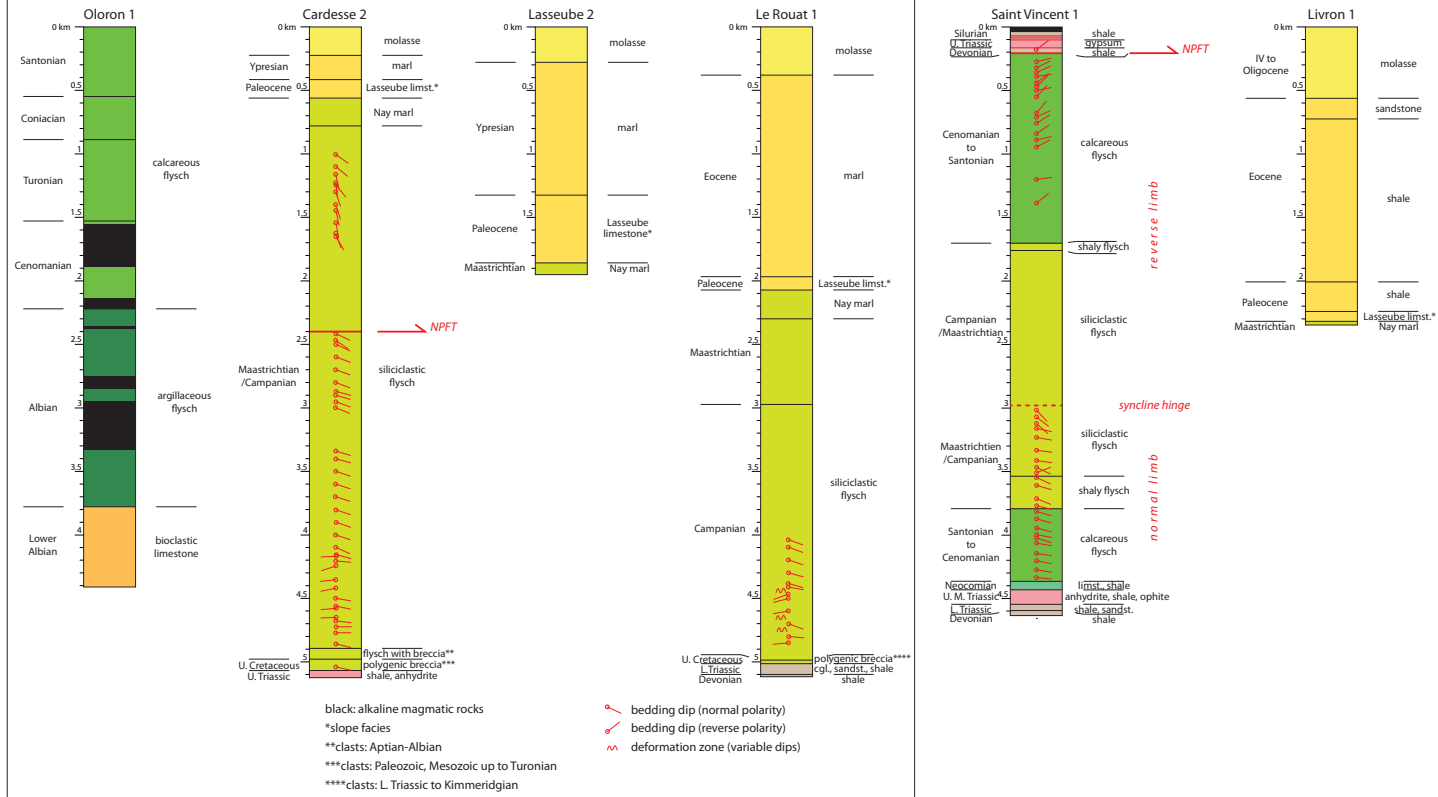


Fig. S1. Stratigraphic logs of wells used in cross-sections S1 and S4 in Figure 4. Original data are in <infoterre.brgm.fr>. NPFT: North Pyrenean Frontal Thrust.

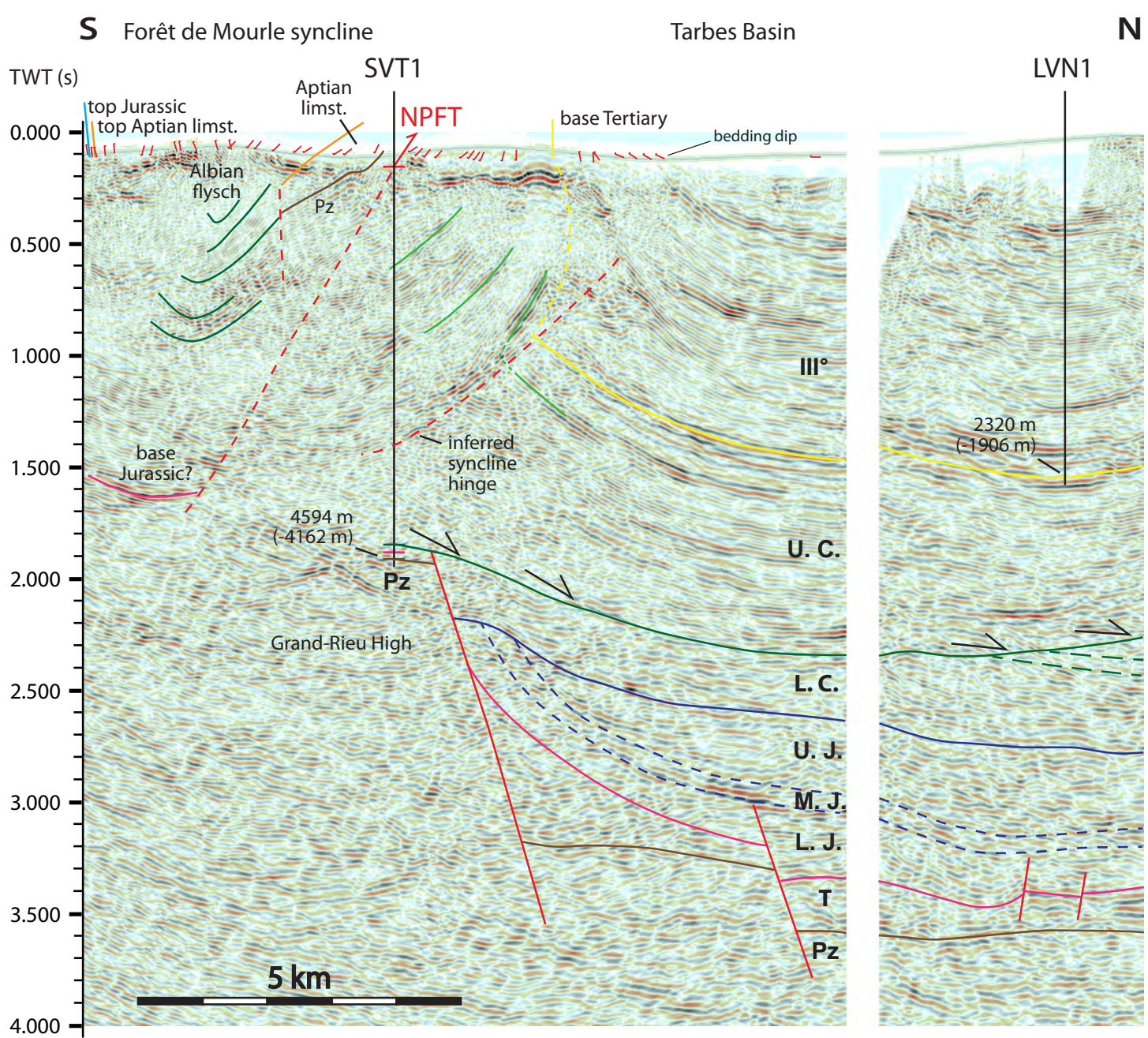


Fig. S2. Seismic reflection profile across the North Pyrenean Frontal Thrust (NPFT) in the northern part of cross-section S4 (Figure 4). It corresponds to the southern part of the Regional Line 5 processed and first published by Serrano et al. (2006). In the present work, the profile was provided by BRGM in the frame of the PYRAMID Project, and it is here visualized with the Kingdom Suite software. Depth-time conversion of wells and geological interpretation north of the Grand-Rieu High are from Serrano et al. (2006), except for the base of the Upper Cretaceous in the northernmost part where we interpret a southward dipping erosional contact in the Albian flysch and onlap of the Upper Cretaceous flysch. Geological interpretation above and south of the Grand-Rieu High and surface bedding dips are from this work. Dips were measured along the trace of the seismic line south of the NPFT and, due to the presence of post-orogenic molasse at the surface, projected from the west in the north. The depths of the basement top in well SVT1 and of the base of Tertiary in well LVN1 are reported (depths measured from the surface and, between parentheses, below sea level) (see Fig. S1). These depths show that seismic velocities are much higher in the NPFT area (well SVT1), probably due to older strata and stronger deformation, than in the younger and little deformed sediments in the Tarbes basin (well LVN1). This results in distortion of the seismic image in the transition zone, preventing good imaging of the structure of the NPFT footwall. Serrano et al. (2006) interpreted a fault between the reverse and normal limbs in the Upper Cretaceous flysch drilled in SVT1 (see bedding dips on well log in Fig. S1) However, the interpretation of an unfaulted syncline with growth strata in the Campanian to Eocene fits better with actual depths in wells, as shown in cross-section S4 in Figure 4. The latter also shows that the top of Lower Cretaceous dips gently southward. NPFT: North-Pyrenean Frontal Thrust; III°: Tertiary; L. C.: Lower Cretaceous; L. J.: Lower Jurassic; M. J.: Middle Jurassic; Pz: Paleozoic; T: Triassic; U. C.: Upper Cretaceous; U. J.: Upper Jurassic. Wells: LVN1: Livron 1; SVT1: Saint-Vincent 1. See well logs in Fig. S1 (including bedding dips for SVT1), and original well data in <infoterre.brgm.fr>.