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1 **Title: Symmetry Vs. Asymmetry of a hyper-thinned rift: example of the *Mauléon***
2 **Basin (Western Pyrenees, France)**

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9 **Abstract**

10 The aim of this study is to unravel the tectono-sedimentary evolution of a hyper-thinned rift,
11 based on the example of the *Mauléon* Basin, a basin filled by thick synrift deposits. The
12 integrated study combines field data, detailed geological mapping and seismic interpretation.
13 The field study focuses on the Iberian margin of the *Mauléon* Basin. Seismic interpretation
14 and well calibration along a N-S transect of the *Mauléon* Basin enable imaging the transition
15 with the northern conjugate margin. The synrift records are very different on either side of the
16 basin: the southern margin is composed of a proximal turbiditic s.l. siliciclastic system,
17 whereas the northern margin is characterized by a carbonate system extending from the
18 platform to the basin. We recognize the *Mauléon* rift as an apparent symmetric hyper-thinned
19 rift, related to a southward dipping Albian detachment and a northward dipping Cenomanian
20 one. Two stages of continental crustal thinning are inferred to explain the development of the
21 *Mauléon* Basin. First, a Barremian to earliest Albian “ductile pure-shear thinning phase”,
22 responsible for the lower crustal thinning and the formation of a symmetric sag basin. Second,
23 an Albian-Cenomanian simple-shear thinning phase, responsible for the onset of the
24 southward dipping *Saint-Palais* detachment faulting and for evolution to an asymmetric basin.
25 The Iberian margin appears as an upper plate and the European one as a lower plate during
26 Albian time. At Early Cenomanian time, the basin was affected by structural changes of the

27 margins resulting from shift in detachment direction, interpreted as "flip-flop detachment
28 tectonics".

29 **Key words:** Hyper-extended rift - Pyrenees - Detachment faulting - Mauléon Basin - Early
30 Cretaceous.

31 **1. Introduction**

32 Since the 1980's, conceptual rifting models have evolved from the pure-shear model
33 (McKenzie, 1978), to the simple-shear model (Wernicke 1981, 1985; Wernicke & Burchfiel
34 1982; Davis, 1983; Spencer 1984; Davis *et al.* 1986; Wernicke & Axen 1988; Brun & van den
35 Driessche 1994) and, more recently, to numerical polyphasing models (Huisman and
36 Beaumont, 2003, 2008, 2011, 2014; Tirel *et al.*, 2004; Lavier and Manatschal, 2006; Brune *et*
37 *al.*, 2014, 2016; Svartman Dias *et al.*, 2015). Much recent work describes the crustal thinning
38 and modalities of mantle exhumation at actual hyper-thinned passive margins such as the
39 Iberia-Newfoundland (Péron-Pinvidic *et al.*, 2007; Hauptert *et al.*, 2016), Angola-Brazil
40 (Unternehm *et al.*, 2010; Péron-Pinvidic *et al.*, 2015), and Australia-Antarctic (Gillard *et al.*,
41 2015) margins. These have been modelled numerically in order to reproduce the geometry of
42 the continental margins and to understand the development of crustal thinning detachment
43 faults. Although the distal part of these extensional systems is well constrained offshore by
44 seismic interpretation, few studies currently describe the proximal onshore domain. One of
45 the most studied examples is the Jurassic Adriatic margin in the Alps (Masini *et al.*, 2011,
46 2012, 2013). These authors describe the crustal detachment faults and sedimentary evolution
47 of this starved system. Unlike the situation in the *Mauléon* rift basin, the hyper-thinning
48 crustal models are based on real continental margins bordering oceanic domains. Where the
49 *Pyrenees* are concerned, development of these conceptual models has led some authors to
50 propose different models to explain the location of the Pyrenean mantle outcrops (Lagabrielle

51 & Bodinier 2008, Jammes *et al.* 2009, 2010a, 2010b; Lagabrielle *et al.* 2010, 2016; Clerc &
52 Lagabrielle, 2014; Masini *et al.* 2014; Tugend *et al.* 2015a; Corre *et al.* 2016; Hart *et al.* 2016;
53 Teixell *et al.* 2016). However, most of the hyper-thinning crustal models are based on mature
54 passive continental margins characterized by mantle exhumation at the ocean-continent
55 transition. The *Mauléon* Basin is an exceptional "laboratory" for the unravelling of the
56 tectono-sedimentary evolution of a highly subsiding, thick, sedimentary, hyper-thinned rift.
57 The evolution of this basin is synchronous to that of the hyper-thinned *Parentis* rift basin
58 (Pinet *et al.* 1987; Bois & ECORS Scientific team 1990; Jammes *et al.* 2010b, Ferrer *et al.*
59 2012; Masini *et al.* 2014) and to that of the *Bay of Biscay* continental margins, which are
60 characterized by a hyper-thinned continental crust and a probably exhumed mantle at the
61 ocean-continent transition (Ferrer *et al.*, 2008; Roca *et al.*, 2011; Tugend *et al.*, 2014). Even
62 though the *Mauléon* Basin was inverted during the Pyrenean orogeny, it has retained the
63 tectono-sedimentary record from its creation to its reactivation. Recently published
64 geodynamical models of the *Mauléon* Basin are mainly based on structural observations. The
65 aim of this article is to analyze the stratigraphic and sedimentological characteristics of the
66 syn-thinning deposits recording hyper-extension. This approach will make it possible to test
67 the validity of the various models proposed and to bring new elements of interpretation. We
68 propose a new model to explain the thinning of the continental crust beneath the *Mauléon*
69 hyper-thinned rift basin, taking into account the sedimentary evolution of the basin through
70 Cretaceous.

71 **2. Geological setting**

72 **Present day structure of the western Pyrenees**

73 The Pyrenean mountain belt results from a north-south convergence and collision between
74 Iberian and European continental blocks from the Late Santonian to the Early Miocene

75 (Puigdefàbregas and Souquet, 1986; Olivet, 1996; Rosenbaum et al., 2002; Sibuet et al., 2004;
76 Gong et al., 2008). The deformation of this intracontinental domain is linked to the northward
77 migration of the Iberian plate and to inversion of the previous Cretaceous north-Pyrenean
78 basin (Ducasse and Vélásque, 1988). The central and eastern parts of the *Pyrenees* can be
79 divided into three structural zones: the North-Pyrenean Zone (NPZ), the *Axial Zone*, and the
80 South-Pyrenean Zone (SPZ) (Fig. 1A; Choukroune, 1976). The NPZ – located between two
81 major faults, the North Pyrenean Fault (NPF) to the south, and the North Pyrenean Frontal
82 Thrust (NPFT) to the north (Fig. 1B; Choukroune and ECORS Team, 1989; Daignières et al.,
83 1994) – is composed of folded Mesozoic cover and Paleozoic units. The commonly named
84 Metamorphic Internal Zone (MIZ), along the southern part of the NPZ, corresponds to a
85 narrow, east-west, vertically metamorphosed and severely deformed zone (Ravier, 1957). The
86 MIZ is characterized by high-temperature and low-pressure metamorphism related to Albian-
87 Cenomanian rifting (Albarède and Michard-Vitrac, 1978; Montigny et al., 1986; Golberg and
88 Maluski, 1988; Golberg and Leyreloup, 1990; Boulvais et al., 2006; Clerc et al., 2015).

89 In the western *Pyrenees*, the NPZ is represented by the Cretaceous *Mauléon* Basin, thrust to
90 the north onto the *Aquitaine* Basin and bordered to the south by the *Axial Zone*, where
91 Paleozoic rocks are unconformably overlain by Late Cretaceous shallow marine carbonates
92 (Souquet, 1967). The *Mauléon* Basin is bounded to the west by Paleozoic blocks commonly
93 known as the "*Massifs Basques*" (Heddebaut, 1973; Muller and Roger, 1977). The MIZ and
94 the NPF disappear in the western part of the *Pyrenees* (Choukroune, 1976; Hall and Johnson,
95 1986; Canérot et al., 2004). The compressive phase is responsible for a greater shortening rate
96 in the central and eastern *Pyrenees* (Muñoz, 1992; Vergés et al., 1995; Beaumont et al., 2000;
97 Mouthereau et al., 2014) than in the western *Pyrenees* (Teixell, 1996, 1998), inducing better
98 preservation of the *Mauléon* Basin. The minor deformation of the western NPZ is linked to
99 diachronism at onset of the thrusting of the Pyrenean compressive phase, older to the east

100 (Santonian) and younger to the west (Middle Eocene), due to Iberian plate kinematics
101 (Souquet et al., 1977; Séguret and Daignières, 1986; Olivet, 1996). This diachronism is the
102 result of the configuration of the plates before convergence, with a more stretched domain in
103 the western *Pyrenees*. The MIZ re-appears in the commonly named “Nappes des Marbres”
104 corresponding to the center of the Albian basco-cantabric basin (Rat et al., 1983).

105 **The *Mauléon* Basin**

106 The *Mauléon* Basin corresponds to a Cretaceous subsiding domain, filled by thick Albian to
107 Upper Cretaceous deep basin deposits. The *Mendibelza-Igountze* southern Unit correspond to
108 the inverted elements of the paleo-southern *Mauléon* Basin margin (Muller and Roger, 1977;
109 Canérot et al., 1978; Puigdefàbregas and Souquet, 1986). This unit is composed of a
110 Devonian-Carboniferous sedimentary substratum (Laverdière, 1930; Paris, 1964; Casteras et
111 al., 1967; Mirouse, 1967), and a thick Albian synrift *Mendibelza* Formation (Lamare, 1946).
112 The *Mendibelza-Igountze* Unit is thrust towards the south onto the *Larrau-Saint Engrâce*
113 Triassic window (Galharague, 1966) and the Late Cretaceous carbonate cover of the *Axial*
114 *Zone*, along the *Lakhoura* Thrust (Lamare 1941; Casteras 1943; Ducasse *et al.* 1986; Teixell
115 1993).

116 The *Arbailles* Unit is separated from the *Mendibelza* Unit by the steeply dipping *Arbailles*
117 Thrust (Lamare, 1948), also called the "Licq Fault" (Teixell et al., 2016). The *Arbailles* Unit
118 is characterized by a preserved Jurassic-Lower Cretaceous carbonate sequence folded in a
119 N110° trending syncline. This syncline is bounded to the north by a major normal fault
120 separating it from the *Saint-Palais* Unit. This latter is represented by a thick (~ 6,000 m)
121 Albian to Late Cretaceous flysch sequence deformed in N110° trending folds of various wave
122 lengths. The north-eastern part of this unit is affected by the *Saint-Palais*, *Bellevue* and
123 *Sainte-Suzanne* (NPFT) northward Pyrenean thrusts (Daignières et al., 1994; Teixell, 1998).

124 During Cretaceous time, the *Mauléon* Basin is separated from the *Aquitaine* Basin by the
125 "Grand-Rieu" Ridge (Serrano et al., 2006).

126 The *Labourd* and *Aldudes* Units are composed of an Ordovician to Carboniferous sedimentary
127 basement (Laverdière, 1930; Lamare, 1944; Heddebaut, 1967, 1973), overlain respectively by
128 Permian-Triassic continental deposits (Lucas, 1985) and an Upper Cretaceous carbonate
129 platform (Souquet, 1967). The *Labourd-Aldudes* N20° Units, forming the western margin of
130 the *Mauléon* rift Basin (Fig. 2), coincide with the commonly named *Pamplona* Fault
131 identified by the authors Richard, 1986; Razin, 1989; Claude, 1990; Larrasoña et al., 2003;
132 Pedreira et al., 2007. The northern part of the *Labourd* Unit is composed of the *Ursuya*
133 granulites (Viennot and Kieh, 1928; Lamare, 1939; Boissonnas et al., 1974; Hart et al., 2016).

134 The eastern part of the *Mauléon* Basin is materialized by the "Chaînons Béarnais",
135 corresponding to a N110° anticline and syncline system, affecting a Jurassic to Early
136 Cretaceous carbonate cover (Canérot, 1988). This sedimentary cover is associated with
137 lherzolite outcrops: "Urdach, Tos de la Coustette, Sarailé, Turon de la Técoùère" (Fabriès et
138 al., 1991, 1998). The *Urdach* mantle body is reworked into Late Albian to Cenomanian
139 synrift breccias, composed of Paleozoic basement and mantle clasts (Roux, 1983; Fortané et
140 al., 1986; Jammes et al., 2009; Debroas et al., 2010; Lagabrielle et al., 2010).

141 **Tectono-sedimentary evolution of the *Mauléon* Basin**

142 The Permian-Triassic post-Hercynian continental deposits filled extensional basins, such as
143 the *Bidarray* Basin (Fig. 2), located in the *Labourd-Aldudes* Units (Bixel and Lucas, 1983,
144 1987). The Late Triassic deposits are characterized by a Lower Triassic Sandstone Unit, a
145 Middle Triassic Carbonate Unit and an Upper Triassic shale, evaporite and ophite complex
146 (Curnelle, 1983; Lucas, 1985; Rossi et al., 2003). During the Jurassic, a carbonate platform
147 developed in a relatively stable tectonic context (Delfaud and Henry, 1967; Lenoble, 1992;

148 James, 1998). The end of the Jurassic is characterized by large scale exposure of the Aquitain-
149 Pyrenean domain. The Neocomian is lacking on this part of Pyrenees and is recorded by the
150 development of the bauxites (Combes et al., 1998; James, 1998; Canérot., 2008). During
151 Early Barremian, localized subsidence of the previously emerged domain favored the
152 transgression of carbonate platform deposits that continued until the earliest Albian (Delfaud
153 and Villanova, 1967; Arnaud-Vanneau et al., 1979). The end of this period is marked locally
154 by extensive halokinetic deformation, associated with the first normal faults and the
155 development of distal spicules marls (Canérot, 1988, 1989; Canérot and Lenoble, 1993; James
156 and Canérot, 1999; Canérot et al., 2005).

157 The deposition of a thick, deep-marine conglomeratic sequence on the southern margin
158 (*Mendibelza-Igountze* Unit) of the newly formed *Mauléon* Basin is considered as marking the
159 onset of the Albian rifting basin, with a strong differentiation between an uplifted southern
160 domain being the source of the conglomerates and a strongly subsiding deep-marine basin to
161 the north (Boirie, 1981; Boirie and Souquet, 1982; Fixari, 1984; Souquet et al., 1985; Fig. 2).
162 In the basin axis, more distal turbidites of the Black Flysch group, time-equivalent to the
163 *Mendibelza* Conglomerates on the southern margin, constitute the first stratigraphic unit of the
164 North-Pyrenean rift. At the same time, along the more gently-dipping northern margin, the
165 albian shallow-marine carbonate deposits grade southwards to more distal marl-dominated
166 sedimentation (Biteau et al., 2006).

167 The *Mauléon* Basin widened during the Cenomanian to Santonian (Canérot, 2017). Active
168 carbonate turbiditic systems supplied by the Aquitain platform to the north are responsible for
169 the deposition of a thick carbonate flysch sequence at the basin axis (Razin, 1989). At the
170 southern margin, the transgressive "*Calcaires des Cañons*" carbonate platform onlaps the
171 previously emerged Paleozoic basement, currently exposed in the *Axial Zone* (Casteras and
172 Souquet, 1964; Souquet, 1967; Alhamawi, 1992). Chaotic gravity-flow sedimentation

173 characterized the southern tectonically-controlled erosional slope (Durand-Wackenheim et al.,
174 1981). From Late Santonian to Eocene times, a thick deep-water sequence deriving from
175 eastern syn-tectonic siliciclastic systems accumulated in the *Mauléon* Basin. Finally, the basin
176 was deformed and inverted by the Pyrenean compression from the Middle Eocene to the
177 Miocene (Puigdefàbregas and Souquet, 1986; Bosch et al., 2016; Labaume et al., 2016).

178 **Present day deep structure of the *Mauléon* Basin**

179 Gravimetric studies highlight a strong positive gravity anomaly under the *Mauléon* Basin
180 (Grandjean, 1992, 1994; Casas et al., 1997). This was previously considered as a dense intra-
181 crustal anomaly (Grandjean, 1994) but is now interpreted, by some authors, as being induced by
182 the proximity of the lithospheric mantle below a thinned crust (Wang et al., 2016; Wang,
183 2017). Recent research efforts focusing on mantle exhumation have linked denudation of the
184 “*Châinons Béarnais*” mantle outcrops to Albian hyper-thinning crustal mechanisms along a
185 low angle detachment fault (Jammes *et al.* 2009; Lagabrielle *et al.* 2010; Masini *et al.* 2014;
186 Tugend *et al.* 2014; Corre *et al.* 2016; Teixell *et al.* 2016). The inversion of this rift basin is
187 responsible for the Iberian northward underplating (Mattauer, 1985; Engeser and Schwentke,
188 1986; McCaig, 1988; Choukroune et al., 1990).

189 **Available geodynamic models of the *Mauléon* rift basin**

190 The crustal thinning process of the *Mauléon* rift basin during Albian times has been debated
191 recently (Jammes et al., 2009; Masini et al., 2014; Tugend et al., 2014, 2015a; Corre et al.,
192 2016; Teixell et al., 2016; Fig. 3). Rifting reconstructions differ from crustal thinning models,
193 salt tectonics and synrift filling geometries. The first models describe this basin as a classical
194 rift, thinned by basinward-dipping, crustal, normal faults, responsible for the development of
195 tilted blocks on the southern margin of the rift (Boirie and Souquet, 1982; Fixari, 1984;
196 Souquet et al., 1985; Canérot and Delavaux, 1986; Canérot, 1988; Ducasse and Vélasque,

197 1988; Souquet, 1988). In recent models of crustal hyper-thinning processes, sub-continental
198 mantle exhumation is interpreted as being related to northward-dipping detachment faults in
199 the distal part of the basin (Jammes *et al.* 2009; Tugend *et al.* 2014, 2015a; Masini *et al.* 2014;
200 Fig 3A). According to these models, the southern margin of the *Mauléon* Basin corresponds
201 to the “lower plate” and the northern *Grand-Rieu* Ridge to the “upper plate” (sensu Wernicke,
202 1985; Lister *et al.*, 1986). The *Ursuya* granulites are also considered as lower to middle crust
203 exhumed during this Albian extensional stage. Two extensional fault systems, the north and
204 south *Mauléon* detachments, could be responsible for the southward tilting of the *Mendibelza*
205 and *Arbailles* Units (Masini *et al.*, 2014). An alternative model suggests that the Albian
206 thinning of the continental crust was caused by a process of crustal boudinage and led to the
207 creation of a symmetric basin and sliding of the Mesozoic cover from the margins towards the
208 basin axis on each margin (Lagabrielle *et al.*, 2010; Corre *et al.*, 2016; Teixell *et al.*, 2016;
209 Fig. 3B).

210 Our working approach was to reconstruct a N-S cross-section of the present-day *Mauléon*
211 Basin, using a field approach on the southern margin and seismic data to illustrate the
212 northern one. The Albian reconstitution of this cross-section allows to consider the tectono-
213 sedimentary evolution of the basin, focusing on the Albian-Cenomanian rifting phase. The
214 first part of this paper focuses on a description of the southern margin of the *Mauléon* rift
215 basin based on outcrop data. The second part focuses on the deep geometry of the *Mauléon*
216 Basin, using the interpretation of a N-S reprocessed composite seismic reflection profile. This
217 line shows the deep geometry of the conjugate rift margins. The third part of the paper
218 presents a new N-S composite crustal section of the entire *Mauléon* Basin, and a
219 reconstitution at Cenomanian time. We propose a new model of crustal thinning highlighting
220 the individualization of the Iberian and European conjugate rift margins.

221 **3. The southern margin of the *Mauléon* Basin from field observations**

222 Field work has enabled updating of the 1/ 50 000 BRGM geological maps. The latter image
223 the Iberian margin of the *Mauléon* Basin: (1) *Mendibelza Unit*; and (2) *Arbailles Unit* (Fig.
224 4A). The southern margin of the *Mauléon* Basin was deformed during the Eocene alpine
225 compression along several southward-verging thrusts. The most significant of these, known as
226 the *Lakhoura* Thrust, is responsible for the under-thrusting of the *Axial Zone* beneath the
227 North-Pyrenean units (Fig. 4). Five structural units bounded by tectonic contacts have been
228 defined from south to north on this inverted margin (Fig. 4B): (1) The *Axial Zone*, formed by
229 the Paleozoic basement and its Late Cretaceous cover (over-thrust by the northern units along
230 the *Lakhoura* Thrust); (2) The *Saint-Engrâce/Bedous* Unit, composed of a complex
231 assemblage of Triassic rocks thrust onto the *Axial Zone* to the south; (3) The *Mendibelza-*
232 *Igountze* Unit, including Albian deep-marine synrift deposits onlapping the Paleozoic
233 sedimentary basement (Fig. 5); (4) The *Arbailles* Unit, represented by carboniferous rocks
234 overlain by a continental to shallow marine Triassic to earliest Albian sequence; and (5) the
235 *Saint-Palais* Unit formed, essentially, of a thick Albian to Late Cretaceous turbiditic sequence
236 overlying a deformed Jurassic to Early Cretaceous Carbonate unit.

237 **The *Axial Zone***

238 The Palaeozoic basement of the *Axial Zone* is stratigraphically covered by a shallow marine
239 carbonate sequence, Cenomanian to Santonian in age, known as the "*Calcaires des Cañons*"
240 (Souquet, 1967; Fig. 6A). This carbonate platform succession is overlain by deeper Late
241 Santonian to Maastrichtian marine sediments composed of argillaceous mudstone
242 ("calshists") containing planktonic foraminifera, grading upward and laterally to siliciclastic
243 deposits. This deep marine depositional setting continues at least until the Early Lutetian in
244 this domain. The Paleozoic rocks are not greatly deformed near the contact with the Late
245 Cretaceous carbonates (*Ehujarré* and *Kakoueta* areas).

246 **The Saint-Engrâce – Bedous Unit**

247 This unit has extensive eastward continuity (*Aspe* Valley), justifying its name *Bedous*. The
248 *Saint-Engrâce-Bedous* Unit observed in the *Larrau* tectonic window is a discontinuous thin
249 band of Triassic rocks delimited by two thrusts and squeezed between the *Axial Zone* to the
250 south and the *Mendibelza – Igountze* Unit to the north (Ducasse and Vélasque, 1988; Teixell,
251 1993). The *Saint-Engrâce-Bedous* Unit is composed of the Middle Triassic Carbonate Unit
252 and the Upper Triassic shale, evaporite and ophite complex. These Triassic rocks are affected
253 by EW-trending folds indicating a southward displacement of the *Lakhoura* Thrust system.

254 **The Mendibelza-Igountze Unit: contacts between the Paleozoic basement**
255 **and the synrift deposits**

256 The *Mendibelza-Igountze* Unit is composed of Albian deep-marine synrift deposits onlapping
257 the Paleozoic sedimentary basement. The *Mendibelza* synrift conglomeratic gravity flow
258 deposits have already been studied, by the following authors: Fournier 1905, 1908; Lamare
259 1939, 1946, 1948; Gubler *et al.* 1947; Magné 1948; Viers 1956; Paris 1964; Poignant 1965;
260 Galharague 1966; Merle 1974, Boirie 1981; Fixari 1984; Souquet *et al.* 1985. The Palaeozoic
261 substratum of the *Mendibelza* Formation (Fm.) is composed of Late Devonian to
262 Carboniferous sedimentary rocks (Fig. 5). This substratum is exposed at two highs: the
263 *Esterençuby* high to the north and the *Occabe* high to the south. The Albian conglomeratic
264 *Mendibelza* Fm. is very thick (2,000 m; Fig. 6A) at the northern block, much thinner at the
265 southern one. The pebbles in this conglomerates are made up of Paleozoic sedimentary
266 basement and Early Albian carbonate platform in the *Igountze* area (“Floridées limestones”).
267 In this area, the base of this conglomeratic sequence is composed of discontinuous calci-
268 turbidites reworking the Early Albian carbonate platform and the Paleozoic basement. In the
269 *Mendibelza* area, the base of this formation is characterized by discontinuous base-of-slope

270 sedimentary breccias reworking the Lower Triassic Sandstone Unit (Figs. 7A and B). These
271 discontinuous *Igountze* calci-turbidites and *Mendibelza* breccias materialized the base of the
272 *Mendibelza* Formation, that is overlaid by three southward back-stepping mega-sequences
273 defined by Souquet et al., 1985. The *Mendibelza* conglomerates are overlain by Cenomanian
274 to Santonian polymictic chaotic breccias reworking the “Calcaires des Cañons” carbonate
275 platform of the *Axial Zone* and angular blocks of the Paleozoic sedimentary basement. These
276 deposits attest a base-of-slope palaeogeographic boundary between the *Axial Zone* and the
277 *Mendibelza-Igountze* Unit during the Late Cretaceous. Along the N-S geological cross-section
278 (Fig. 4), the *Mendibelza-Igountze* Unit is overthrust southward by the *Lakhoura* Thrust, onto
279 the *Axial Zone* and the *Saint-Engrâce* Unit. The *Mendibelza-Igountze* Unit is also thrust
280 towards the west, onto the *Aldudes* Unit. The Cretaceous to Paleocene synrift and post-rift
281 strata are affected by the *Lakhoura* Thrust, as shown by their vertical to overturned position
282 along the front of the *Lakhoura* Thrust on the southern limit of the *Mendibelza* Unit,
283 indicating that this thrust is post Cretaceous.

284 First rift models described the *Mendibelza-Igountze* Unit as turbiditic proximal cones installed
285 on the southward tilted blocks affecting the Iberian margin of the *Mauléon* Albian Basin
286 (Boirie and Souquet, 1982; Fixari, 1984; Souquet et al., 1985; Canérot and Delavaux, 1986;
287 Canérot, 1988; Ducasse and Vélasque, 1988; Souquet, 1988). However, in the light of recent
288 understanding of crustal hyper-thinning processes, Masini *et al.* (2014) have interpreted the
289 southward tilting of the *Mendibelza-Arbailles* Unit as resulting from the development of a
290 northward detachment fault. To evaluate these different hypotheses, a detailed field analysis
291 was conducted to determine the nature of the substratum / cover contacts and the synrift
292 deposit geometry. All the contacts observed in the field between the substratum and the
293 synrift deposits are primary contacts that were not reactivated during the Pyrenean orogeny.
294 Two kinds of contacts have been distinguished: (1) depositional surfaces; and (2) syn-

295 sedimentary faults. All these surfaces have been corrected for post-depositional tilting in order
296 to reconstruct their geometries at the time of the syn-rift deposition (Fig. 8).

297 Albian depositional surfaces

298 The depositional surfaces (yellow and blue lines in Fig. 5) are in all cases characterized by an
299 angular unconformity with the Albian synrift strata. This unconformity is between 15–40°
300 (Fig. 8). Our facies analysis of the *Mendibelza* Fm. confirms that this unit consists of
301 conglomeratic gravity-flow deposits (Boirie, 1981; Boirie and Souquet, 1982; Fixari, 1984;
302 Souquet et al., 1985). The composition, the very coarse granulometry and the roundness of the
303 clasts suggest that this base-of-slope gravity system was fed by non-preserved shelf-edge fan-
304 deltas, reworking the Paleozoic meta-sedimentary units in an uplifted southern area located in
305 the present day *Axial Zone*. In this context, these deep marine conglomerates are considered to
306 have been deposited in a close to horizontal position. In addition, the *Mendibelza* Fm. is
307 currently stacked with an average regional dip of N120°–30 to 40°SW, without any wedge-
308 shaped syn-tectonic growth strata. These characteristics make it possible to reconstruct the
309 inclination of the depositional surfaces of the synrift albian deposits on the Palaeozoic
310 substratum (Fig. 8). The surfaces are generally tilted from 15° to 40° towards the NE to NNE,
311 with an average value of 30°, once the *Mendibelza* Fm. strata have been reseated in the
312 original horizontal position. This shows that the Albian substratum of the proximal part of the
313 *Mauléon* Basin was inclined towards the basin axis at the time of the synrift *Mendibelza* Fm.
314 deposition. Locally, the *Esterençuby* high is characterized by an albian surface inclined at 30°
315 to 50° towards the east.

316 Albian synsedimentary vertical normal faults

317 Two fault systems have been identified: (1) the N110°–45°NNE North *Occabe* Fault (NOF, in
318 orange in Figs. 5 and 8); and (2) the N20° vertical normal *Esterençuby* faults (in purple in Fig.

319 5). These faults have an angular relationship of between 75° and 90° with the strata of
320 *Mendibelza* conglomerates. Reconstruction of these faults indicates an original orientation and
321 dip of N120°–75 to 85° NE for the NOF and an unchanged N20° vertical for the *Esterençuby*
322 faults. The Early-Middle Albian megasequences 1 and 2 of the Black Flysch (Souquet et al.,
323 1985) were deposited during the activity of the NOF which is sealed by the upper part of
324 megasequence 2 (Fig. 5). The stratal geometries and geological mapping show that the
325 activity of the N20° *Esterençuby* fault system was contemporary with the NOF, i.e. during the
326 Early-Middle Albian.

327 Megasequence 3 overlapped the *Occabe* high during the Late Albian. The occurrence of chaotic
328 gravity flow deposits from the Late Albian to Santonian, on the Occabe domain implies the
329 presence of a tectonically controlled scarp in between the emerged to shallow marine Axial
330 Zone and the deep marine Mendibelza domain. However this fault scarp is now overprinted
331 by the Lakhoura Thrust. Therefore, these faults are diachronous and increasingly younger
332 towards the southern margin of the *Mauléon* Basin, since the North *Occabe* Fault is sealed by
333 Middle Albian deposits. These observations are in agreement with the southward back-
334 stepping of the Albian deposits (Souquet et al., 1985).

335 The Palaeozoic rocks are affected locally by low temperature weak deformations (Fig. 7C).
336 But several of these cataclasites are sealed by Permian sediments indicating a Late-Hercynian
337 age for this deformation (Fig. 5). These observations do not therefore confirm the occurrence
338 of cataclasites punctuating a detachment fault between the Hercynian basement and the
339 Albian synrift deposits, as proposed by Jammes et al. (2009), Masini et al. (2014).

340 The *Mendibelza* Unit: preliminary interpretations

341 Most previous publications describe the *Mendibelza* Unit as an Early Albian southward tilted
342 block, characterized by a stratigraphic or tectonic contact between the substratum and the

343 Albian *Mendibelza* conglomerates (Ducasse *et al.* 1986; Souquet *et al.* 1985; Ducasse &
344 Vélasque 1988; Souquet 1988; Masini *et al.* 2014). However, the present study, based on field
345 analysis and structural reconstruction, does not support these interpretations. Quite the
346 opposite in fact: it shows an absence of wedge-shaped syn-tectonic growth strata in the Albian
347 synrift sequence and depositional surface on the hercynian substratum inclined from 15 to 40°
348 towards the NE. The hercynian substratum of the *Mendibelza* Unit dips towards the *Mauléon*
349 Basin axis. The tilt of this unit towards the NE is marked by the development of diachronous
350 N120° normal faults that are increasingly younger towards the south. The present day tilting
351 of the Albian sequence, about 45° towards the south, postdates the *Mendibelza* Formation.

352 **The *Arbailles* Unit: tectonic and stratigraphic framework**

353 The contact between the *Arbailles* and the *Mendibelza* Units corresponds currently to a
354 vertical fault. Due to the overturned and northward-dipping position of the Albian sequence of
355 the *Mendibelza* Unit to the south, this fault is interpreted as a verticalized southward-directed
356 thrust fault called the *Arbailles* Thrust (Ducasse and Vélasque, 1988, Fig. 4). This thrust is
357 locally retro-thrusting towards the north, indicating a later northward displacement of the
358 *Mendibelza* Unit responsible for the verticalization of the previous *Arbailles* Thrust (Dumont
359 *et al.*, 2015). The *Arbailles* Unit is bounded northward by a set of northward-dipping to sub-
360 vertical normal faults, known as the North *Arbailles* Fault (NAF, Fig. 9). These N120° faults
361 result in a juxtaposition of Barremian - Aptian shallow-marine carbonate deposits to the
362 south, and deeper marine marl-dominated Albian deposits to the north. This set of faults
363 disappears eastward. In fact, the *Arbailles* Unit is overlapped eastward by the Albian-
364 Cenomanian *Tardets* Black Flysch group deposits (Souquet *et al.*, 1985). Westward, the
365 *Arbailles* Unit is bounded by the *Saint-Jean-Pied-de-Port* Upper Triassic Unit.

366 The *Arbailles* Unit consists of a carboniferous basement, a Late Triassic sequence and a
367 Mesozoic cover (Figs. 4 and 9). In this sector, the Late Triassic shales and evaporites overlie
368 the Lower Triassic sandstone unit, without any Middle Triassic carbonate Unit. The Upper
369 Triassic Unit is widely exposed on the south-west and consists of ophites and numerous
370 cagneules indicating a wide zone of deformation. There is a classical Jurassic sequence
371 above the Upper Triassic Unit (Lenoble, 1992). This Jurassic sequence is limited at the top by
372 an erosional surface marked by greater and greater truncation towards the south-west,
373 explaining the lack of most of the Late Jurassic in the southern part of the *Arbailles* Unit (Fig.
374 6A). This unconformity is linked to the Late Jurassic - Early Cretaceous emersion phase,
375 associated regionally with the development of bauxites (Combes et al., 1998; James, 1998;
376 Canérot et al., 1999). The geometry of this surface reflects the tilt of the Jurassic deposits
377 towards the north during this period. This emersion extends as far as the base of the
378 Barremian, as evidenced by the lack of Neocomian deposits in the *Arbailles* Unit (Fig. 6A).

379 The Barremian to earliest Albian carbonate platform deposits onlap the Jurassic carbonate
380 platform towards the south-west (Fig. 6A). This onlapping is highlighted by the southward
381 thinning of the Barremian deposits and the interbedding of transgressive conglomerates
382 including pebbles reworked from the pre-existing Kimmeridgian carbonates (Fig. 7D). The
383 Aptian carbonates (Urgonian facies) overlying the Barremian to the north backstep towards
384 the south and onlap the eroded Jurassic sequence directly in this direction. The thickness of
385 the Aptian deposits increases northward without significant facies change, indicating a
386 northward increase in subsidence (towards basin axis, Figs. 4, 6A and 9).

387 From Late Aptian to earliest Albian, algal carbonates ("Mélobésiées" limestone) onlap the
388 eroded Jurassic sequence in the southern part of the *Arbailles* Unit (Fig. 7E). Intraformational
389 breccias (Fig. 7F) attest to the presence at that time of an unstable depositional setting and
390 probably a slight inclination of the depositional profile. This is confirmed by the northward

391 change in facies from shallow marine carbonates to deeper marine spicule marls with
392 intercalations of some carbonate debrites towards the north (Fig. 7G and H). This Late
393 Aptian-Early Albian sequence is affected by two kinds of progressive unconformity: (1) a
394 generalized and progressive northward thickening of this sequence at the scale of the
395 *Arbailles* Unit, linked to a northward increase in the subsidence rate; and (2) a growth strata
396 unconformity of shorter wavelength that could result from diapiric movements along the
397 southern edge of the *Arbailles* Unit (Canérot, 1988, 1989, 2008; Canérot and Lenoble, 1993).

398 **The *Saint-Palais* Unit**

399 The *Saint-Palais* Unit is located between the *Arbailles* Unit and the *Saint-Palais* Thrust. The
400 *Saint-Palais* Unit formed the Albian - Late Cretaceous depocenter of the *Mauléon* Basin. At
401 outcrop scale, this unit is folded into N120° synclines / anticlines of pluri-kilometric
402 wavelength, composed of Early Albian spicule marls and Middle Cenomanian - Turonian
403 carbonate distal turbidites. The northern part of this unit is affected by the N110-120°
404 overturned *Saint-Palais* anticline. Field work confirms that the distal Late Cretaceous
405 carbonate turbidites are sourced from the north by the European carbonate platform, as shown
406 by the N150–200° directed paleocurrents measured in the field. These Late Cretaceous
407 carbonate deposits overlaid the Albian sequence unconformably. This is shown by the
408 southward distal onlap of the Late Cretaceous deposits on the Albian synrift sequence, on the
409 southern edge of the *Saint-Palais* Unit.

410 **4. Seismic interpretation of the *Mauléon* Basin and the southern part of the** 411 ***Aquitain* Basin**

412 Unlike the southern margin that has been observed directly in the field (*Mendibelza* and
413 *Arbailles* Units), the architecture of the central part and the northern conjugate margin of the

414 *Mauléon* Basin have been explored from seismic data and well calibration (Figs. 6 and 10).
415 This section focuses on the deep geometry of the *Mauléon* Basin from the interpretation of a
416 N-S composite seismic reflection profile, reprocessed by the BRGM in 2017 (MT104 and
417 MT112 acquired by ESSO-Rep in 1969 on the *Mauléon-Tardets* exploration permit, cf 4.2).
418 The seismic line interpreted is a merger of lines MT104, MT112 and 83HBS02 (acquired by
419 SNEAP in 1981 and reprocessed by BRGM in 2014; Fig. 10). The seismic interpretation was
420 carried out using the Geographix® Discovery Suite that allows for each CDP of the seismic
421 profile to export its X, Y, time, amplitude and depth values. Seismic/well tie is obtained by
422 using the time / depth curves acquired in the boreholes and derived from the recording of
423 checkshots. The time / depth conversion was performed using a mean velocity field calculated
424 on the basis of all the time / depth data of the calibration wells, integrating their deviations
425 (Fig. 11). The top basement has been calibrated taking into account the ECORS seismic
426 reflection profile (Daignières et al., 1994). The composite line presented in this paper is
427 calibrated with six wells (Fig. 6), from south to north: *Ainhice-1* (3,540.85 m depth); *Uhart-*
428 *Mixe-1* (1,868.8 m depth); *Bellevue-1* (6,909 m depth); *Orthez-102* (5,489.10 m depth);
429 *Amou-1* (5,543 m depth); and *Bastennes-Gaujacq-1.bis* (4,442.10 m depth). Only the *Ainhice-*
430 *1* well reaches the Paleozoic basement of the *Mauléon* Basin. The studied SSW-NNE seismic
431 line through the *Mauléon* Basin cuts across four distinct structural units, separated from one
432 another by thrust faults (Fig. 10). From south to north, the units are as follows: (1) *Saint-*
433 *Palais*; (2) *Bellevue*; (3) *Sainte-Suzanne*; and (4) *Grand-Rieu/Arzacq* (Figs. 6 and 10). The
434 *Saint-Palais* Unit corresponds to the central and deepest part of *Mauléon* Basin, while the
435 *Bellevue* and *Sainte Suzanne* Units belong to its northern margin. The *Bellevue* and *Sainte*
436 *Suzanne* thrusts involved the European basement. Using the field cross-section (Fig. 4B) and
437 well-velocity models calibrating the interpreted composite seismic reflection profile (Fig. 10),
438 we propose a depth migrated section of the entire *Mauléon* Basin (Fig. 11).

439 **The *Saint-Palais* Unit**

440 The Palaeozoic substratum below the *Saint-Palais* Unit is tilted towards the north, being at a
441 depth of around 2,900 m in the *Ainhice-1* well (Fig. 6) and of 7,000 m in the central part of
442 the basin. The base of this well is characterized by Paleozoic basement, overlain by the thick
443 Upper Triassic salt Unit materializing a major decoupling level between the basement and the
444 sedimentary cover. This Triassic salt unit is grading upward to two Mesozoic cover units
445 characterized by a normal polarity and separated by a tectonic contact. The Lower Unit, U1,
446 consists of a complete Jurassic sequence, up to the Kimmeridgian, covered by Late Aptian
447 Urgonian limestones. Unit U1 is characterized by a hiatus of the Neocomian to the Early
448 Aptian. The Upper Unit, U2, is composed of the same Jurassic but the Kimmeridgian marls
449 are thinner than in U1, grading upward to Earliest Albian carbonate platform deposits (Fig.
450 6A). These cover units are overlain by Middle to Late Albian spicule marls and siliciclastic
451 turbidites. The Albian basinal deposits thicken towards the north and reach their maximum
452 thickness in the *Saint-Palais* Anticline, clearly identified in the field. The *Saint-Palais* domain
453 corresponds to the depocenter of the Albian basin, with an estimated thickness of up to 4,000–
454 5,000 m (Roux, 1983; Fixari, 1984; Souquet et al., 1985). The upper part of the *Saint-Palais*
455 Unit is formed by the Upper Cretaceous carbonate-dominated flysch, which reaches a
456 thickness of more than 1,500 m in the *Uhart Mixe* well.

457 **The *Bellevue* Unit**

458 The *Bellevue* Unit is located between the *Saint-Palais* and the *Bellevue* Thrusts. The *Bellevue*
459 Thrust is responsible for the offset of the Jurassic-Cretaceous cover and the displacement of
460 the *Bellevue* Unit onto the *Sainte Suzanne* Unit. The Paleozoic basement is located at a depth
461 of around 8,000 m under the *Bellevue* Unit. This unit is composed of a complex structure
462 whose most obvious feature is the *Bellevue* Anticline affecting the Mesozoic cover

463 intersected in the *Bellevue* well. The northward *Bellevue* Thrust is marked by a 1,800 m thick
464 tectonized zone composed of Late Triassic salt and Barremian carbonate slices. These
465 Barremian units could be tectonically incorporated to the Triassic salt forming the overturned
466 flank of the overlying anticline evidenced in the *Bellevue* well (Fig. 6). In fact, this tectonized
467 interval is overlain by an overturned Jurassic carbonate unit composed of Dogger to
468 Kimmeridgian deposits. Above, the *Bellevue* Unit is made up of a complete Jurassic to Late
469 Cretaceous sequence in a normal position. The Early Cretaceous sequence is characterized by
470 a lack of Neocomian. The Barremian to Aptian carbonate platform unit lying unconformably
471 on the Portlandian is more than 1,500 m thick. The Barremian deposits are composed of
472 limestones with annelids and 500 m thick of more distal *Sainte-Suzanne* marls. The Aptian is
473 mainly composed of Urgonian limestones. The Albian thins towards the North and is
474 represented by spicule marls including intercalations of carbonate mudstone indicating the
475 proximity of the northern *Aquitainian* carbonate platform. The Late Cretaceous sequence is
476 made up of Cenomanian to Coniacian carbonate turbidites.

477 **The *Sainte-Suzanne* Unit**

478 The *Sainte-Suzanne* Thrust corresponds to the major north verging thrust responsible for a
479 displacement of more than 17 km of the *Mauléon* Basin deposits onto the south *Aquitain*
480 *Grand Rieu* Unit (Fig. 10). It can be considered as the North-Pyrenean Frontal Thrust in this
481 part of the *Pyrenees*. The *Sainte Suzanne* Unit is made up of a very thick Jurassic to
482 Campanian sequence, slightly inclined and thickening towards the south. It has been crosscut
483 by the *Orthez-102* well where the *Sainte Suzanne* Thrust has been identified at a depth of
484 4,320 m (Fig. 6). The Jurassic is made up of a complete sequence of marine carbonate
485 platform deposits from Hettangian to Kimmeridgian, truncated under the Early Cretaceous
486 deposits, as evidenced by the north-directed erosional truncations. The 1 000m thick Early
487 Cretaceous sequence is characterized by a lack of Neocomian. The Barremian is characterized

488 by shallow marine carbonate deposits with annelids showing that this unit is more proximal
489 than the *Bellevue* Unit. The Aptian is sub-isopachous and is composed of shallow marine
490 carbonate platform deposits (Urgonian facies). The Albian deposits are also made up of
491 carbonate platform deposits thickening towards the south. These facies are therefore more
492 proximal than the Albian deposits of the *Bellevue* Unit. A clear truncation of the Albian
493 reflections is seen towards the north, below the Late Cretaceous calcareous flysch. This latter
494 calci-turbidite sequence is thinning and onlapping to the north on top of the eroded Albian
495 sequence.

496 **The *Grand-Rieu/Arzacq* Unit**

497 The *Grand-Rieu – Arzacq* Unit is the footwall of the *Sainte Suzanne* Thrust and corresponds
498 to the southern part of the *Aquitain* Basin. From the seismic data, the Paleozoic basement
499 below the *Sainte Suzanne* Thrust can be estimated to be at a depth of 8,500 m. It is assumed to
500 be overlain by a more or less continuous Triassic Unit. The inferred Liassic Unit present in
501 the *Aquitain* Basin to the north appears to pinch-out southwards against a diapiric structure
502 associated with a normal fault. This Mesozoic tectonic structure marks the boundary between
503 the *Grand-Rieu* Ridge to the South and the commonly named *Arzacq* Basin corresponding to
504 a highly subsiding domain during Albian times. The overlying Middle to Late Jurassic
505 carbonates are more continuous and isopachous. The top of the Jurassic sequence corresponds
506 to a major exposure surface marked by north-directed erosional truncations under the
507 Barremian transgressive carbonate deposits, as indicated in the *Amou* well. This unconformity
508 is time-equivalent to the one describe in the *Arbailles* Unit on the southern margin of the
509 *Mauléon* Basin. During the latest Jurassic and the Neocomian period, the northern margin of
510 the *Mauléon* Basin was tilted towards the south, unlike the southern margin that was tilted
511 towards the north. The Barremian to Aptian sequence of the *Grand-Rieu* Unit is relatively
512 isopachous and consists mainly of shallow marine carbonate platform deposits. The Albian

513 deposits form a thick unit of fairly deep marine spicule marls indicating a strong subsidence
514 rate of the *Arzacq* Basin, as indicated in the *Amou* and *Bastennes-Gaujacq* (Fig. 6). A large
515 wave-length anticline affects the Jurassic - Albian sequence at the transition between the
516 *Grand-Rieu* Ridge and the *Arzacq* Basin (*Grand-Rieu* Anticline). A major truncation of the
517 folded strata below the Cenomanian carbonate platform deposits attests to the Late-Albian to
518 Early Cenomanian age of this deformation. In the *Grand-Rieu – Arzacq* Unit the Late
519 Cretaceous is represented by sub-isopachous carbonate platform deposits. During the Tertiary,
520 this domain was characterized by a very high subsidence rate, increasing towards the South,
521 as attested by the very thick Tertiary sequence at the front of the *Sainte Suzanne* Thrust. Local
522 southward dipping clinoforms can be observed south of the *Bastennes-Gaujacq* well,
523 probably related to the growth of this diapiric structure. The very great thickness of the
524 Tertiary sequence attests to the strong flexural subsidence of the *Grand-Rieu – Arzacq* Unit
525 but also to the large displacement and uplift of the *Mauléon* Basin along the *Sainte-Suzanne*
526 Thrust, since no Tertiary deposits have been preserved in the *Mauléon* Basin (even if they
527 were deposited over this entire flexure domain). Most of the Tertiary flexural subsidence
528 occurs before the Late Lutetian, as shown by the thickening of the Danian to Lower Ypresian
529 deposits towards the south.

530 **5. Top Jurassic to Cenomanian tectono sedimentary and geodynamic** 531 **evolution of the *Mauléon* Basin**

532 **Latest Jurassic to Neocomian uplift phase**

533 The Jurassic is characterized by an extensive and relatively continuous carbonate platform
534 spreading between the southern *Mauléon* Basin margin (*Arbailles* Unit) and the *Arzacq* Basin
535 (Fig. 12A). At its top, the platform is limited by a major unconformity resulting from an
536 emersive and erosional phase dated from the latest Jurassic to Barremian (Fig. 12B). This

537 event is evidenced by the development of bauxites and the lack of Neocomian deposits both in
538 the *Pyrenees* and the *Arzacq* Basin (Combes et al., 1998; James, 1998; Canérot et al., 1999).
539 Before the Early Cretaceous transgression, the Jurassic platform is characterized by a synform
540 morphology extending from the southern margin of the *Mauléon* Basin to the *Arzacq* Basin. It
541 is revealed respectively by the south-directed and north-directed erosional truncations
542 affecting the Jurassic in the *Arbailles* and the *Arzacq* Units (Fig. 11). This deformation stage
543 implies the uplift of these domains, corresponding partly to the Early Cretaceous Mauléon
544 Basin edges; while in the meantime, subsidence stops in the synform axis. Mechanism
545 responsible for the uplift of the entire Pyrenean-Aquitaine domain might be related to an
546 asthenosphere upwelling preceding the Early Cretaceous rifting.

547 **Barremian to earliest Albian rifting stage 1: symmetric basin**

548 After the Late Jurassic – Early Cretaceous exposure/weathering stage, onset of transgression
549 leads to the development of an extensive restricted carbonate platform during the Barremian.
550 From Barremian to Aptian time, the balance between carbonate production and creation of
551 available space favors the aggradation of 1 600m thick shallow marine carbonate facies,
552 witnessing a highly subsiding stage controlling a relative-flat depositional profile (Figs. 6 and
553 12C). This homogeneous subsidence stage evolves during Late Aptian to the earliest Albian.
554 The basin's edges were characterized by the back-stepping of the shallow carbonate platform,
555 while the *Mauléon* and *Arzacq* Basins center are composed of deeper spicule marls. In this
556 domain, creation of available space exceeded the carbonate production, resulting in the
557 relocating of the carbonate factory on the basin's edges. Paleogeography reorganization was
558 thus controlled by the differential subsidence regime responsible for a steeper depositional
559 profile.

560 As a consequence, during the Barremian - earliest Albian time the *Mauléon* Basin was a
561 symmetric basin, characterized by shallow marine carbonate platforms on each basin edges.
562 The absence of brittle deformation within the upper-crust excludes the application of the pure-
563 shear model developed by McKenzie (1978) to explain the subsidence regime of this basin. A
564 symmetric synrift sag basin with no brittle deformation in the upper crust has been proposed
565 from numerical modeling (Karner et al., 2003). In this model, the subsidence of the basin
566 results from lower crustal ductile thinning favored by decoupling between the upper and
567 lower crusts. We propose that the initiation of the *Mauléon* sag basin results from thinning of
568 the lower crust, without any major deformation of the upper crust. The isostatic response to
569 the lower crustal thinning induces regional sagging and onlapping of the syn-thinning deposits
570 on each side of the basin. The rifting stage 1 can be interpreted as a "ductile crustal pure-shear
571 thinning phase".

572 **Albian rifting stage 2: asymmetric basin**

573 At Albian time, the southern margin of the *Mauléon* Basin was affected by differential
574 vertical movements: (1) uplift of the southern part of the basin; and (2) tectonic subsidence
575 towards the north (Fig. 12E). They are responsible for the onset of the *Mendibelza* deep
576 gravity flow deposits passing distally to the Black Flysch group towards the *Saint-Palais*
577 depocenter (Fig. 12E). At this time, the Iberian margin substratum was characterized by a
578 northward dipping surface with an average slope angle of 15–30° (Fig. 8). This surface was
579 affected by syn-sedimentary albian N120° southward-propagating high-angle normal faults,
580 controlling the destabilization of the fan-delta system into the deep basin (Fig. 12E). These
581 faults were sealed by younger synrift back-stepping deposits of the *Mendibelza* Fm. towards
582 the south.

583 The conglomerates directly overlain the Paleozoic substratum of the *Mendibelza* Unit,
584 indicating the entire Mesozoic lacks (Fig. 5), while it was continuous throughout the
585 Pyrenean-Aquitaine domain. Thus, the Mesozoic cover of this unit need to be subtracted prior
586 to the sedimentation of the Albian synrift *Mendibelza* sequence. It leads to the development of
587 an emerged *Axial Zone* during the rifting stage 2. The sliding level consists of the thick Upper
588 Triassic shale, evaporite and ophite complex systematically brecciated and / or deformed
589 under the prerift cover of the Iberian margin. This mass-sliding is recorded within the
590 Ainhice-1 well Mesozoic cover unit overlap (Units U1 and U2, Fig. 6). The stratigraphic
591 differences between these two Mesozoic cover units, suggest an origin of the upper one from
592 a southern marginal part of the basin where the Jurassic-Cretaceous hiatus is greatest. Thus,
593 the Early Albian *Axial Zone* denudation is linked to the northward mass sliding of the
594 Mesozoic cover towards the *Mauléon* Basin axis in response to the Iberian margin northward
595 tilting (Fig. 12D).

596 At the same period, the European margin of the *Mauléon* Basin recorded less vertical
597 movements and tectonic activity than the southern margin. Unlike the Iberian conjugate
598 margin, this northern domain is not subject to gravitational sliding of the Mesozoic cover.
599 Indeed, no regional uplift and associated coarse-grained alluvial-derived gravity-flow deposits
600 are evidenced in the *Arzacq*, *Sainte-Suzanne* and *Bellevue* Units. The Albian period is mainly
601 represented by a shallow marine carbonate platform recognized in the *Arzacq* and *Sainte-*
602 *Suzanne* Units (Figs. 6 and 13). The slightly southward-dipping depositional profile is
603 responsible for a facies transition between shallow-marine algae limestones and deeper
604 spicule marls in the *Sainte Suzanne* Unit. In *Saint Palais* and *Bellevue* Units, the Jurassic-
605 Aptian sedimentary cover is significantly normally offset, which allows the accumulation of
606 thick deep Albian facies to the south of the *Saint-Palais* Thrust, suggesting that this latter
607 corresponds to a southward dipping Albian normal fault (Figs. 11 and 13). This fault is part of

608 the northern *Mauléon* Albian Basin's margin. The thick interval with very deformed Triassic
609 material intersected at the base of the *Bellevue* well is interpreted as an Albian diapiric ridge
610 developing on the northern edge of the Albian *Mauléon* Basin (Fig. 6).

611 Comparison between the southern and northern margins shows that the *Mauléon* Basin
612 acquired asymmetry during the Early Albian time: (1) deep basin turbiditic deposits on the
613 Iberian margin and (2) carbonate platform on the European one. According to most of the
614 models that consider a simple-shear deformation process to explain the asymmetry of a rift
615 (Wernicke, 1985; Lister et al., 1986), the *Mauléon* Basin's asymmetry is interpreted as
616 resulting from the activation of the southward dipping *Saint-Palais* Detachment during the
617 Early Albian (Fig. 13). Following Lister et al. (1986) the Albian Iberian margin corresponds
618 to the upper plate and the European one to the lower plate. The *Saint-Palais* Detachment
619 might be responsible for the northward Iberian margin tilting, this former being part of a
620 major crustal scale roll-over. The roll-over structure is accommodated by N120° oriented and
621 steeply dipping normal faults that propagate towards the south (i.e. NAF, NOF). They are
622 antithetic to the *Saint-Palais* structure. During Albian time, the width of the *Mauléon* deep
623 basin, measured from platform to platform, was around 50 km.

624 **Late Cretaceous rifting stage 3: apparent symmetric basin**

625 During the Late Cretaceous period, a carbonate platform developed on both edges of the
626 *Mauléon* Basin, and the deep basin became wider (Figs. 6 and 13). Indeed, the width of the
627 *Mauléon* deep basin, measured from platform to platform, was around 100 km wide. The
628 *Grand-Rieu* Ridge represents a major palaeogeographic boundary between the carbonate
629 platform of the European margin and the center of the *Mauléon* Basin (Razin, 1989; Serrano
630 et al., 2006). The transition between the carbonate platform and the basinal carbonate
631 turbidites is governed by a southward dipping normal fault between the *Sainte Suzanne* Unit

632 and the *Arzacq* Basin. The calcareous flysch deposited in the *Mauléon* Basin was derived
633 mainly by gravity flows coming from the northern *Aquitain* carbonate platform. These
634 calcareous turbidites onlap the previous Albian deposits towards the *Sainte-Suzanne* Unit to
635 the north and the *Saint-Palais* to *Mendibelza* Units to the south (Fig. 13). The development of
636 the *Grand-Rieu* Anticline attests of salt movement at Early Cenomanian time.

637 On the southern margin the commonly named Late Cretaceous "*Calcaires des Cañons*"
638 carbonate platform deposits lie unconformably over the previously exposed Palaeozoic
639 basement of the *Axial Zone* (Fig. 13). Towards the north, in the *Mendibelza* Unit, base-of-
640 slope chaotic breccias attest of a normal fault-controlled platform-basin transition between the
641 southern carbonate platform developing on the *Axial Zone* and the *Mauléon* Basin (Fig. 13).
642 Several arguments indicate that a second and opposite detachment occurred at Early
643 Cenomanian time: the northward verging *Lakhoura* Detachment (Fig. 13). This syn-
644 sedimentary fault, currently inverted as the *Lakhoura* Thrust (Fig. 5), is interpreted as being
645 responsible for the southward tilting of the *Mendibelza/Arbailles* Units. The Early
646 Cenomanian age of this tilting is attested by: (1) the absence of wedge-shaped syn-tectonic
647 growth strata in the Albian synrift sequence of the *Mendibelza* Unit; (2) the 98 Ma cooling
648 age obtained on the (Hart et al., 2017), responsible for the southward tilt of the Lower Triassic
649 Sandstone Unit in the *Saint-Jean-Pied-de-Port* area (Masini et al., 2014); and (3) the onlap of
650 the Late Cretaceous deposits on the tilted and eroded Albian sequence in the north of the
651 *Arbailles* Unit (Casteras et al., 1971; Ducasse and Vélasque, 1988; Fig. 9). This observation
652 implies that the northern edge of the *Arbailles* Unit becomes a structural high in the basin
653 during Early Cenomanian. This structural high is fossilized at Santonian time as highlighted
654 by the age of the first deposits passing over it.

655 **6. Discussion and Conclusion**

656 The northwestern Pyrenees present a strong positive gravity anomaly under the *Mauléon* rift
657 Basin (Grandjean, 1994; Casas et al., 1997) resulting from the presence at low depth (~10 km)
658 of lithospheric mantle (Wang et al., 2016). This high mantle body position implies the
659 existence of a very thin continental crust arguing for a hyper-thinning Cretaceous rifting
660 stage. In this work, we assume that the *Saint-Palais* and *Lakhoura* structures affect the
661 lithospheric mantle. They are responsible for the hyper-thinning of the *Mauléon* Basin
662 continental crust. However, this rift basin have not underwent oceanic spreading as classical
663 basins along the Atlantic Ocean (Manatschal., 2004; Péron-Pinvidic et al., 2007; Hauptert et
664 al., 2016), or in the Alps fossil margins (Manatschal et al., 2006, Masini et al., 2011, Mohn et
665 al., 2014 et Decarlis et al., 2015). Another significant difference with these classical hyper-
666 extended magma poor margins is the thickness of the syn-thinning deposits in the hyper-
667 extended domain (~ 8 km thick in *Mauléon* Basin vs. less than 1 km thick on the Atlantic and
668 Alps margins). The thick synrift deposits record a complex polyphasic thinning history from
669 Barremian to Early Cenomanian time.

670 During stage 1, the *Mauléon* Basin is symmetric as evidenced by the development of a
671 carbonate platform on each basin edges. The carbonate production counterbalances the high
672 accommodation, resulting in a near flat sedimentary profile. We interpret the creation of
673 accommodation as a result of ductile thinning of the lower crust, without affecting the upper-
674 crust (Fig. 14A). This thinning stage shares similarity with the crustal scale boudinage
675 process, previously proposed at Albian time (Lagabrielle et al, 2010; Corre et al, 2016; Teixell
676 et al, 2016; Fig. 3B). However, only the lower crust is affected by ductile thinning, favored by
677 the presence of a decoupling level in between the lower and upper crusts. The resulting synrift
678 sag basin morphology is quite similar to the Mesozoic *Columbrets* sag Basin, which does not

679 present any evidence of deep basin gravitary sedimentary infill (Etheve, 2016 and Etheve et
680 al, 2018).

681 At Albian time (stage 2), the rift fabric totally changed. The basin was affected by "simple-
682 shear thinning" due to the development of the southward dipping *Saint-Palais* Detachment
683 (Fig. 14B). This crustal thinning phase affected the entire lithosphere, forming an asymmetric
684 rift basin, as previously proposed in the simple-shear models (Jammes *et al*, 2009; Tugend *et*
685 *al*, 2014, 2015a; Masini *et al*. 2014, Fig. 3A). However, unlike the previous models, the
686 Albian rifting stage initiates with simple shear thinning, without distributed extension in the
687 upper crust (pure shear thinning sensus McKenzie., 1978). Lavier and Manatschal., 2006 have
688 evidenced using numerical modeling that the continental crust was already thin (~20km)
689 when simple shear thinning starts. In the case of the *Mauléon* Basin we propose that the Early
690 Cretaceous stage favors the ductile thinning of the lower crust before beginning the Albian-
691 Cenomanian simple shear thinning stage.

692 The *Saint-Palais* detachment is responsible for the northward Iberian margin roll-over. This
693 interpretation strongly contrasts with previous models arguing for an Albian southward tilt of
694 the Iberian margin along a northward dipping normal fault (Boirie., 1981; Fixari., 1984;
695 Souquet et al., 1985; Ducasse and Vélasque., 1988) or crustal scale detachments (Jammes et
696 al., 2009; Masini et al., 2014). In this scheme, we propose that rift asymmetry controls the
697 asymmetric sedimentary profile. Actually, the Iberian margin (roll-over) is composed of a
698 proximal turbiditic s.l. system whereas the European one (southward gently-dipping) is
699 characterized by a carbonate platform. Similarly, to the *Columbrets* Basin (Etheve., 2016 and
700 Etheve et al., 2018), the onset of simple shear thinning is synchronous with the basinward
701 prerift cover-sliding along the Late Triassic salt décollement. Such Early Albian gravity
702 tectonics have already been reported all along the Iberian rift margin (Bouquet, 1986; Ducasse
703 et al., 1986; Ducasse and Vélasque, 1988; Lagabrielle et al., 2010; Corre et al., 2016; Teixell

704 et al., 2016; Fig. 3B). However, one of the major discrepancies with the previous models is
705 the absence of gravitational cover sliding along the European rift margin (Fig. 3B). The
706 sliding of the Mesozoic cover led to local diapirism along the Iberian margin, as shown by the
707 lateral facies variations from shallow carbonate platform to spicule marls around the diapiric
708 structures (Canérot, 1988, 1989; Canérot and Lenoble, 1993; James and Canérot, 1999;
709 Canérot et al., 2005).

710 The paleogeography of the Early Cretaceous *Mauléon* Basin, derived from the sedimentary
711 record is not compatible with a major strike-slip drift of the Iberian plate as proposed in
712 kinematic reconstructions (Le Pichon et al., 1971; Roest and Srivastava, 1991; Olivet, 1996;
713 Rosenbaum et al., 2002; Sibuet et al., 2004; Gong et al., 2008). We assume that the Iberia
714 sinistral strike-slip motion did occur on a more southward structure (Malod., 1982; Canérot.,
715 2016) developed earlier during Late Jurassic (Tugend et al., 2015b) and / or may have been
716 evenly distributed within the Cretaceous rift system.

717 At Early Cenomanian time (final stage 3), the southward tilting of the *Mendibelza-Arbailles*
718 Units, – formerly considered as Early Albian (Boirie, 1981; Boirie and Souquet, 1982; Fixari,
719 1984; Souquet et al., 1985; Ducasse and Vélasque, 1988; Masini et al., 2014) – is induced by
720 the change of detachment vergence and the development of the *Lakhoura* northward dipping
721 Detachment. The Arbailles Unit thus becomes a structural high bordering the southern hyper-
722 thinned *Mauléon* Basin domain. Consequently, we assume that the *Arbailles* Unit becomes a
723 tilted block at Early Cenomanian time.

724 Therefore, the tectono-sedimentary evolution of the *Mauléon* rift Basin during the stages 2
725 and 3 is defined by the development of two antithetic and diachronous detachments, i.e. the
726 Albian southward dipping *Saint-Palais* detachment during Stage 2 and the Early Cenomanian
727 northward dipping *Lakhoura* one during Stage 3. The change in detachment vergence between

728 Albian and Cenomanian time can be interpreted as "flip-flop detachment tectonic" (Sauter et
729 al., 2013; Geoffroy et al., 2014; Gillard et al., 2015), highlighting the diachronous shift from
730 upper-plate to lower-plate morphology along a same margin section. This double successive
731 asymmetry is responsible for the apparent final symmetry of this hyper-thinned rift. This work
732 reconciliates the previous rift models, i.e. symmetric (Souquet, 1988; Ducasse and Vélásque,
733 1988), asymmetric (Jammes et al., 2009; Masini et al., 2014; Tugend et al., 2015a, Fig. 3A) or
734 mixed (Lagabrielle et al., 2010; Corre et al., 2016; Teixell et al., 2016, Fig. 3B). The
735 polyphase thinning history is responsible for the hyper-thinning of the continental crust and
736 the development of a relatively narrow hyper-thinned rift domain (Fig. 14C).

737 In three dimensions, the *Mauléon* Basin appears relatively complex and not cylindrical. On
738 our synthetic 2D transect (Fig. 14C), the mantle is systematically overlain by a hyper-thinned
739 crust whereas eastward (*Urdach*), it has been denuded and partly reworked into synrift
740 deposits (Roux, 1983; Duée et al., 1984; Fortané et al., 1986; Debroyas et al., 2010; Lagabrielle
741 et al., 2010; Teixell et al., 2016). We consider the *Saint-Palais* Detachment as responsible for
742 the hyper-thinning of the continental crust, while the *Lakhoura* one might be responsible for
743 partial mantle denudation. This is consistent with the *Urdach* mantle/crust contact that display
744 top north-east shear direction (Corre., 2017). Thus, it appears that the basin morphology
745 changes along strike, due to N0°-20° transverse structures such as the *Pamplona* (Richard,
746 1986; Razin, 1989; Claude, 1990; Larrasoaña et al., 2003; Pedreira et al., 2007; Díaz et al.,
747 2012), the *Saint-Jean-Pied-de-Port*, *Saison*, *Barlanès* and *Ossau* structures (Canérot, 2008,
748 2017; Debroyas et al., 2010). This N0°-20° structuration seems to clearly control the 3D
749 geometry of the *Saint-Palais* hyper-extended rift domain, and will be the purposed of future
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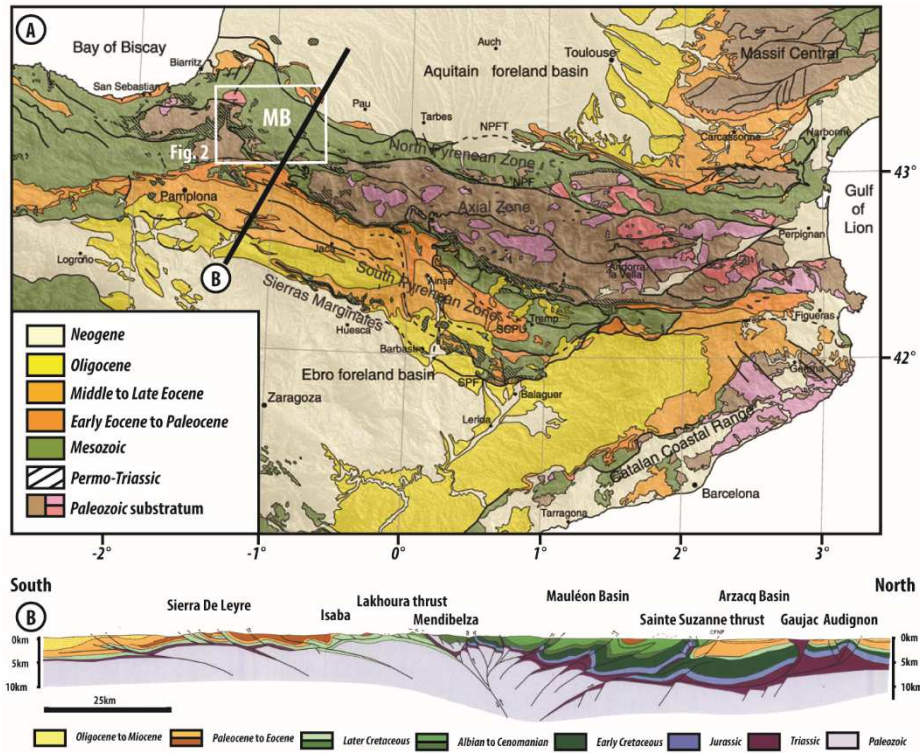
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1319 **Figures and Captions:**

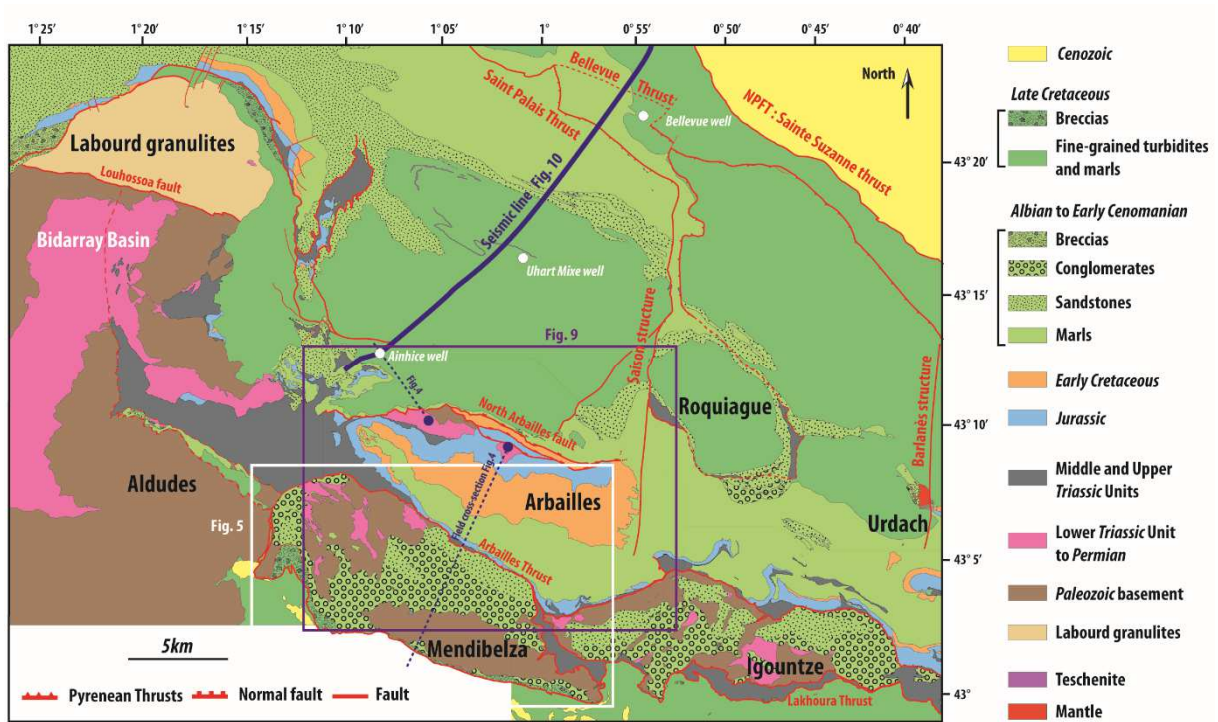
1320 **Figure. 1**



1321
 1322 (A) Geological map of the Pyrenees showing the structural domains of the belt: North
 1323 Pyrenean Zone, Axial Zone and South Pyrenean Zone (Mouthereau et al., 2014). NPFT:
 1324 North Pyrenean Frontal Thrust; MB: Mauléon Basin; (B) Geological cross-section across the
 1325 Western Pyrenees showing from north to south, the: *Audignon Ridge*, *Arzacq Basin*, *Mauléon*
 1326 *Basin*, *Mendibelza Unit*, *Lakhoura Thrust*, *Isaba Unit* and *Sierra de Leyre Unit* (AGSO and
 1327 BRGM., 2018).

1328

Figure. 2

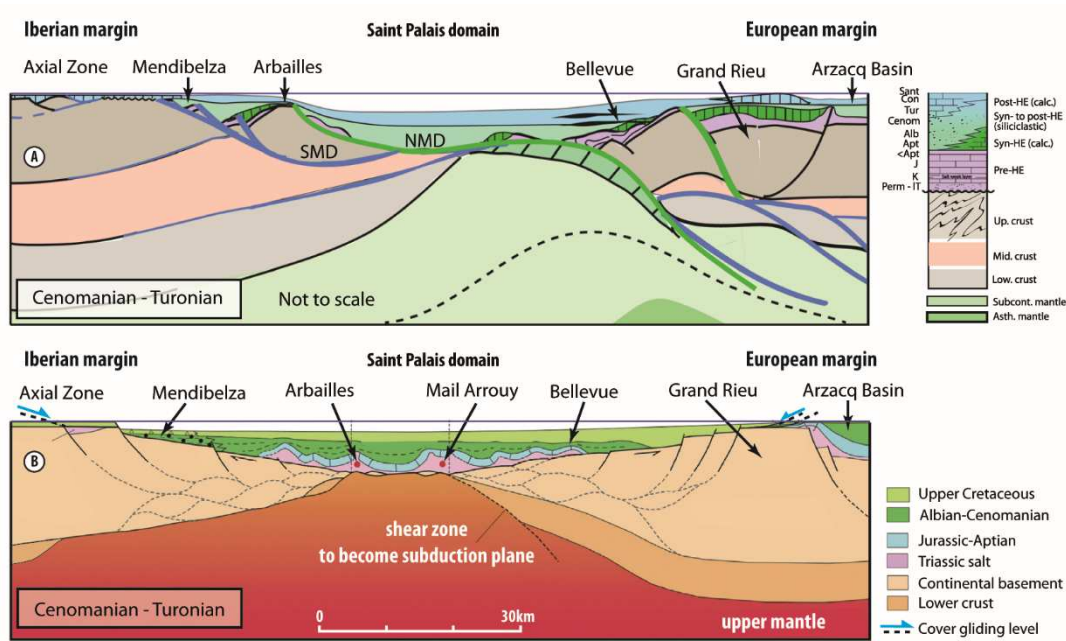


1329

1330 Geological map showing the main structural units of the Iberian *Mauléon* margin, from west
 1331 to south: *Labourd* granulites, *Bidarray* Basin, *Aldudes* Unit, *Mendibelza-Igountze* Unit. The
 1332 *Mauléon* Basin geological map presented in this paper derive from an update of the published
 1333 1/50 000 geological maps (Casteras et al., 1970, 1971; Boissonnas et al., 1974; Le Pochat et
 1334 al., 1976, 1978; Henry et al., 1987).

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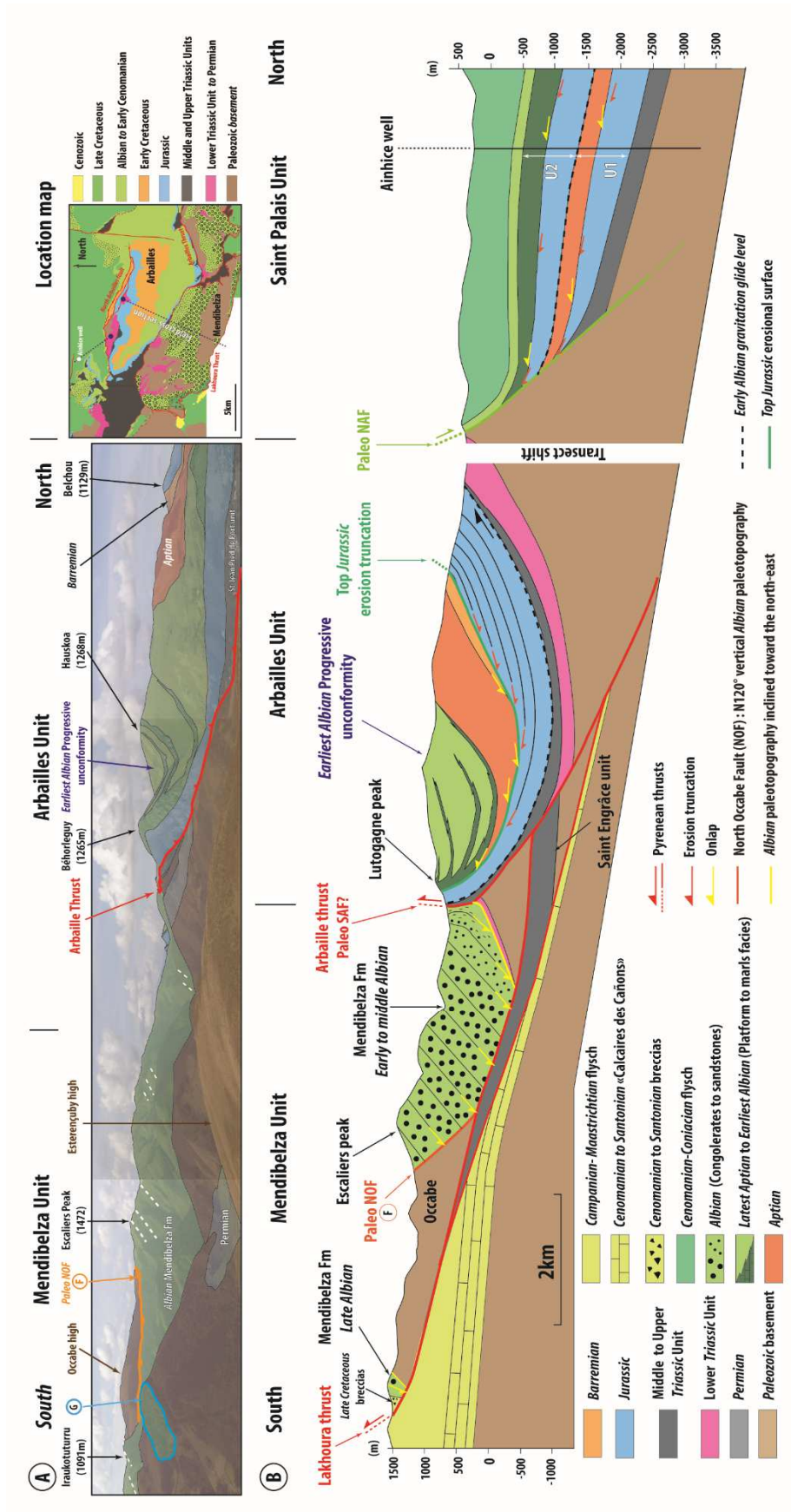
Figure. 3



1336

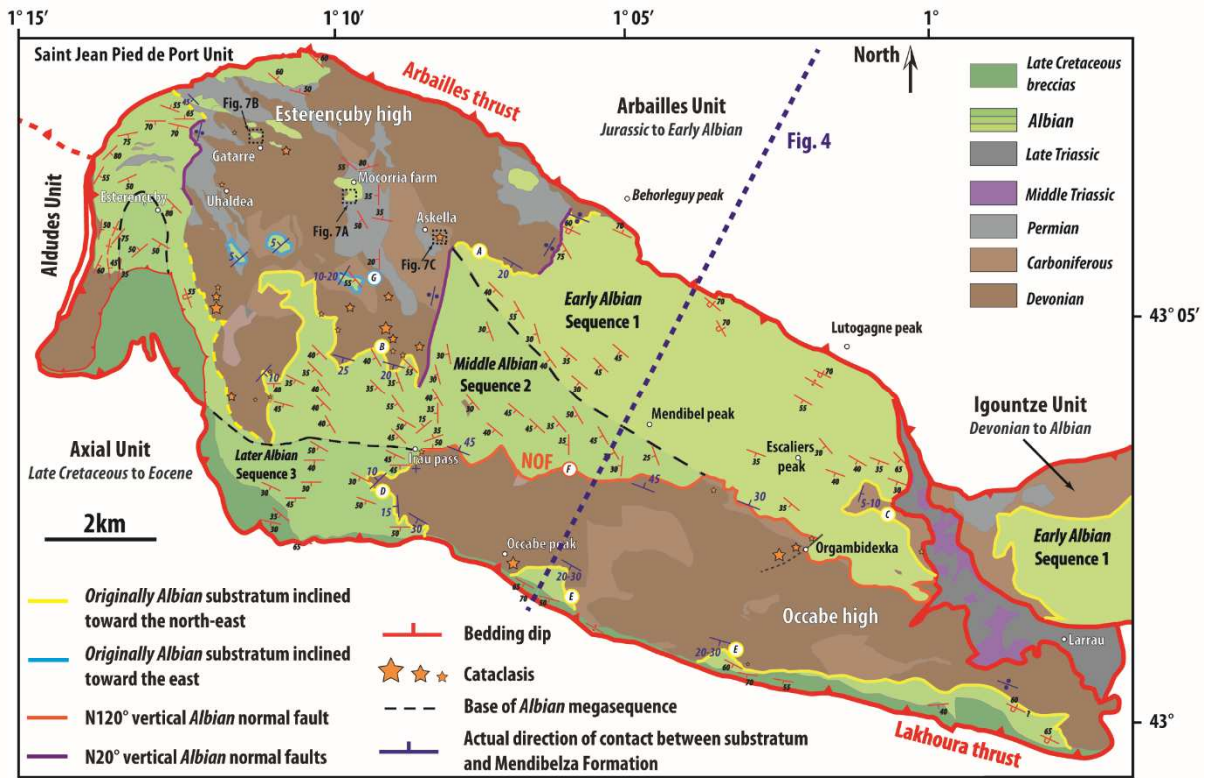
1337 (A) 2D synthetic tectono-sedimentary model of the Mauléon-Arzacq rift section during the
 1338 Cenomanian to Turonian post-tectonic stage (modified from Masini et al., 2014). The
 1339 continental crust is thinned during Albian time as the result of the development of two
 1340 diachronous northward dipping detachments. SMD: South Mauléon Detachment; NMD:
 1341 North Mauléon Detachment; HE: Hyper-extension; (B) Late Cretaceous restoration of the
 1342 Mauléon Basin. The dashed lines correspond to supposed anastomosed shear zones
 1343 contributing to continental crust thinning (modified from Teixell et al., 2016). This model
 1344 proposes a boudinage-like geometry of the crust, with narrow continental margins (<40 km
 1345 wide).

Figure. 4



1348 (A) Panoramic view of the *Mendibelza* and *Arbailles* Units. The *Mendibelza* Formation is
1349 characterized by the absence of wedge-shaped syn-tectonic growth strata in the Albian synrift
1350 sequence and a N120°-30°SW orientation. These conglomerates stop on the Albian North
1351 *Occabe* paleo-normal fault (NOF) whose current orientation is N110°-45°N. The *Arbailles*
1352 Unit is characterized by a progressive unconformity affecting the Aptian to earliest Albian
1353 deposits, showing a northward inclination of the sedimentary profile at this time; (B) Current-
1354 day geological cross-section showing from north to south the: *Saint-Palais*, *Arbailles*,
1355 *Mendibelza*, *Saint-Engrâce/Bedous* and *Axial Zone* Units. The *Arbailles* Unit is characterized
1356 by Jurassic cover eroded to the south, overlain by a southward back-stepping Early
1357 Cretaceous carbonate platform. The *Arbailles* Unit is separated from the *Mendibelza* Unit by
1358 the steep *Arbailles* Thrust. The *Mendibelza* Unit overthrust the *Axial Zone* and the
1359 *Larraul/Saint-Engrâce* Units due to the *Lakhoura* Thrust.

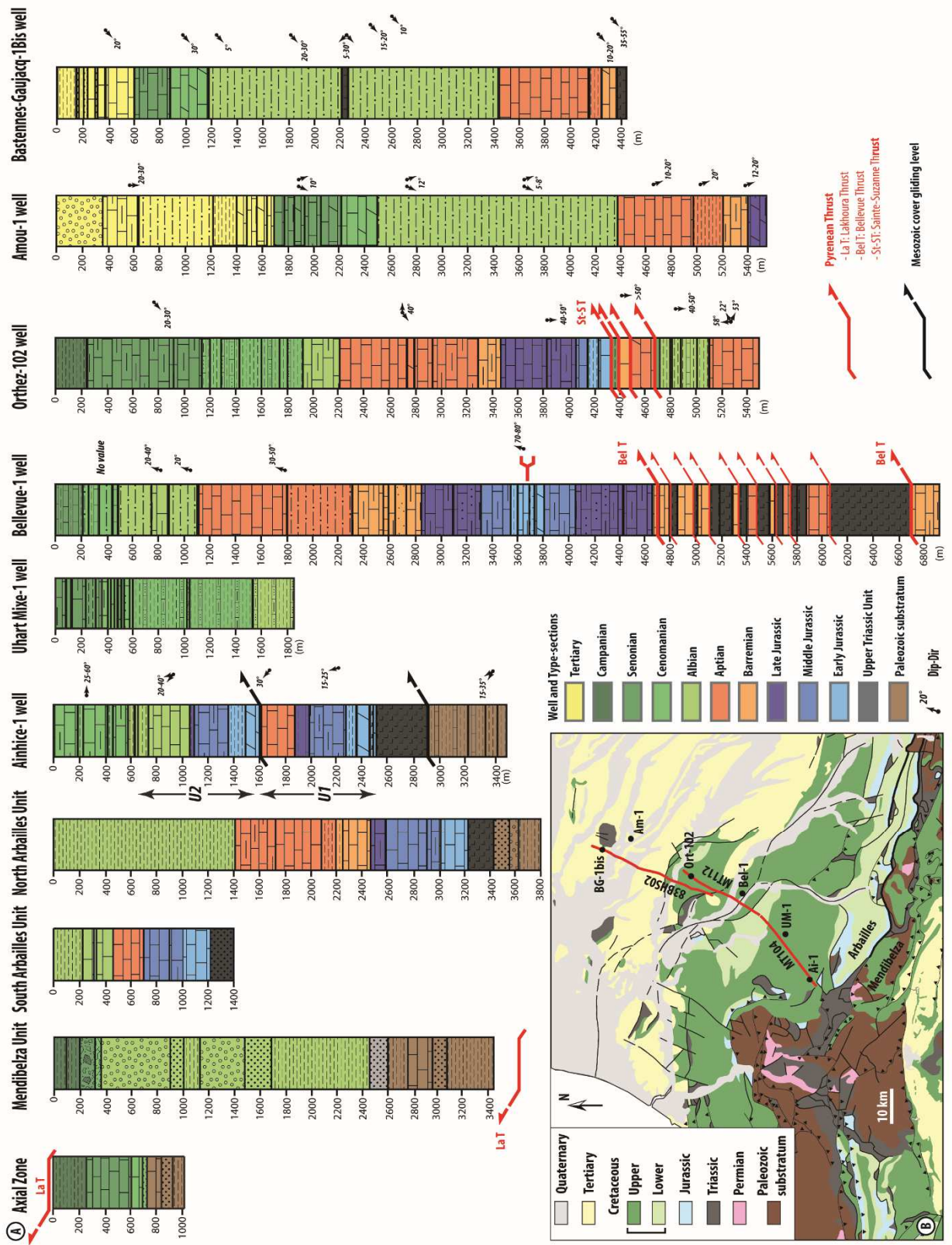
Figure. 5



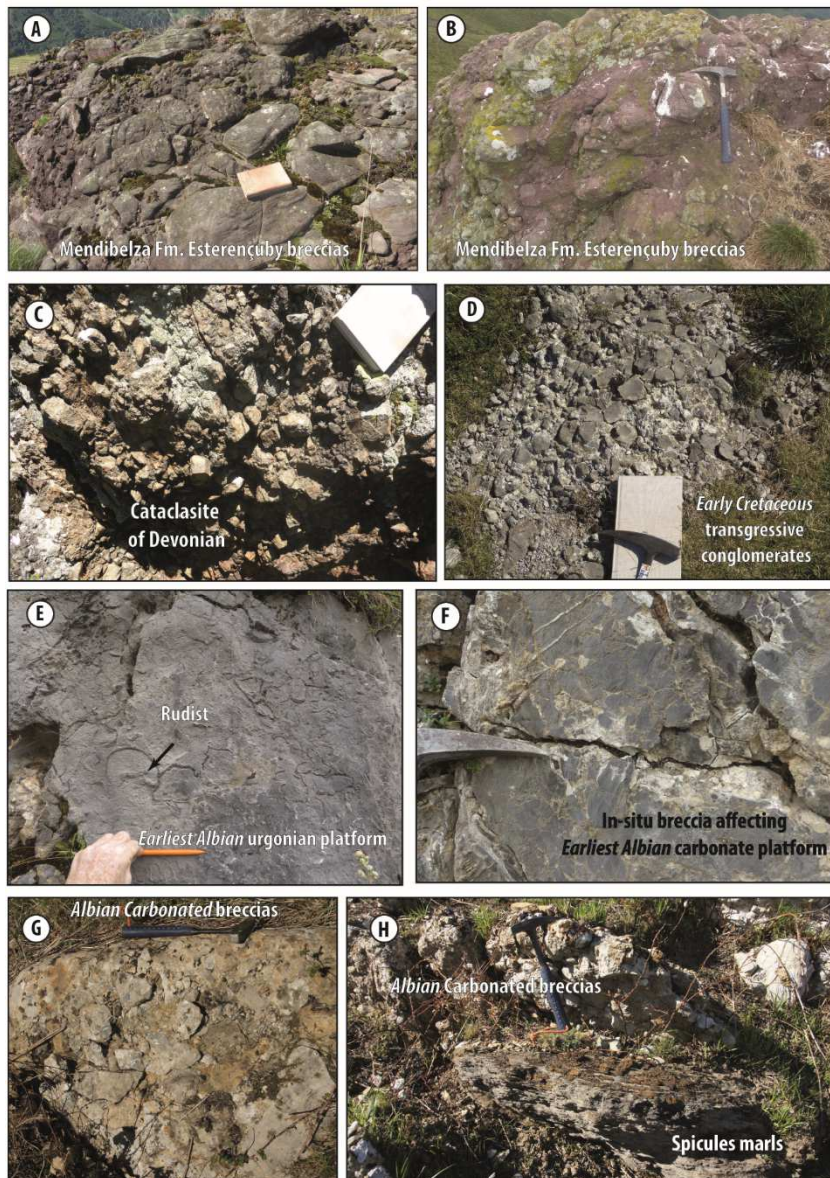
1361

1362 Geological map of the *Mendibelza* Unit, modified from the 1/50 000 BRGM geological maps
 1363 (Casteras et al., 1970, 1971; Le Pochat et al., 1978). The *Mendibelza* Formation is composed
 1364 of three main synrift mega-sequences, younger and younger towards the south, defined by
 1365 Souquet et al. (1985). The stereographic representations show the Albian orientation of the
 1366 contact between the *Mendibelza* conglomerates strata and the sedimentary basement. The
 1367 *Mendibelza* Unit is separated from the *Arbailles* Unit by the steep *Arbailles* fault and is
 1368 overthrust towards the south by the *Lakhoura* Thrust. The North Occabe Fault (NOF) is
 1369 fossilized by the Albian Megasequences 1 and 2, indicating an Early to Middle Albian age for
 1370 this fault.

Figure. 6



1373 (A) To the south: synthetic sedimentary succession of the Axial Zone, Mendibelza Unit,
1374 South Arbailles Unit and North Arbailles Unit; To the north: sedimentary facies of the six
1375 wells calibrating the interpreted N-S composite seismic reflection profile (Fig. 10). (B)
1376 Location of the N-S composite seismic reflection profile (Fig. 10) on the 1 / 400 000 BRGM
1377 geological map of Pyrenees (Baudin and Barnolas, 2008). From south to north the MT104,
1378 MT112 and 83HBS02 seismic lines. Ai-1: Ainhice-1 well, UM-1: Uhart Mixe-1 well, Bel-1:
1379 Bellevue-1 well, Ort-102: Orthez-102 well, Am-1: Amou-1 well and BG-1bis: Bastennes
1380 Gaujacq-1bis well.

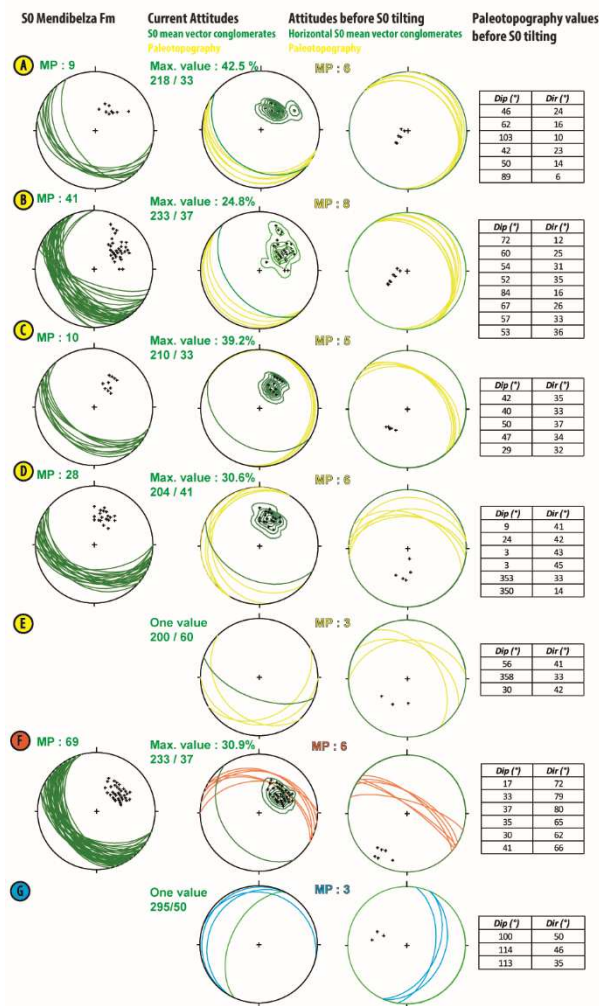
Figure. 7

1382

1383 See Figs. 5 and 9 for location. (A-B) Early Albian *Esterençuby* breccias reworking the Lower
 1384 Triassic Sandstone Unit. These discontinuous breccias are located at the base of the
 1385 Megasequence one of the *Mendibelza* Formation defined by Souquet et al. (1985); (C)
 1386 Cataclasites affecting the sedimentary Paleozoic basement of the Esterençuby high; (D) Early
 1387 Cretaceous transgressive conglomerates reworking the Kimmeridgian carbonate deposits; (E)
 1388 latest Aptian-earliest Albian “Mélobésiées” carbonate platform on the southern part of the
 1389 *Arbailles* Syncline; (F) In-situ breccias of the Late Aptian-earliest Albian carbonate platform

1390 deposits, due to the northward inclination of the sedimentary profile; (G-H) Carbonate
1391 breccias intercalated with spicule marls on the northern corner of the *Arbailles* Unit. These
1392 breccias correspond to the northward reworking of the Late Aptian-earliest Albian carbonate
1393 platform.

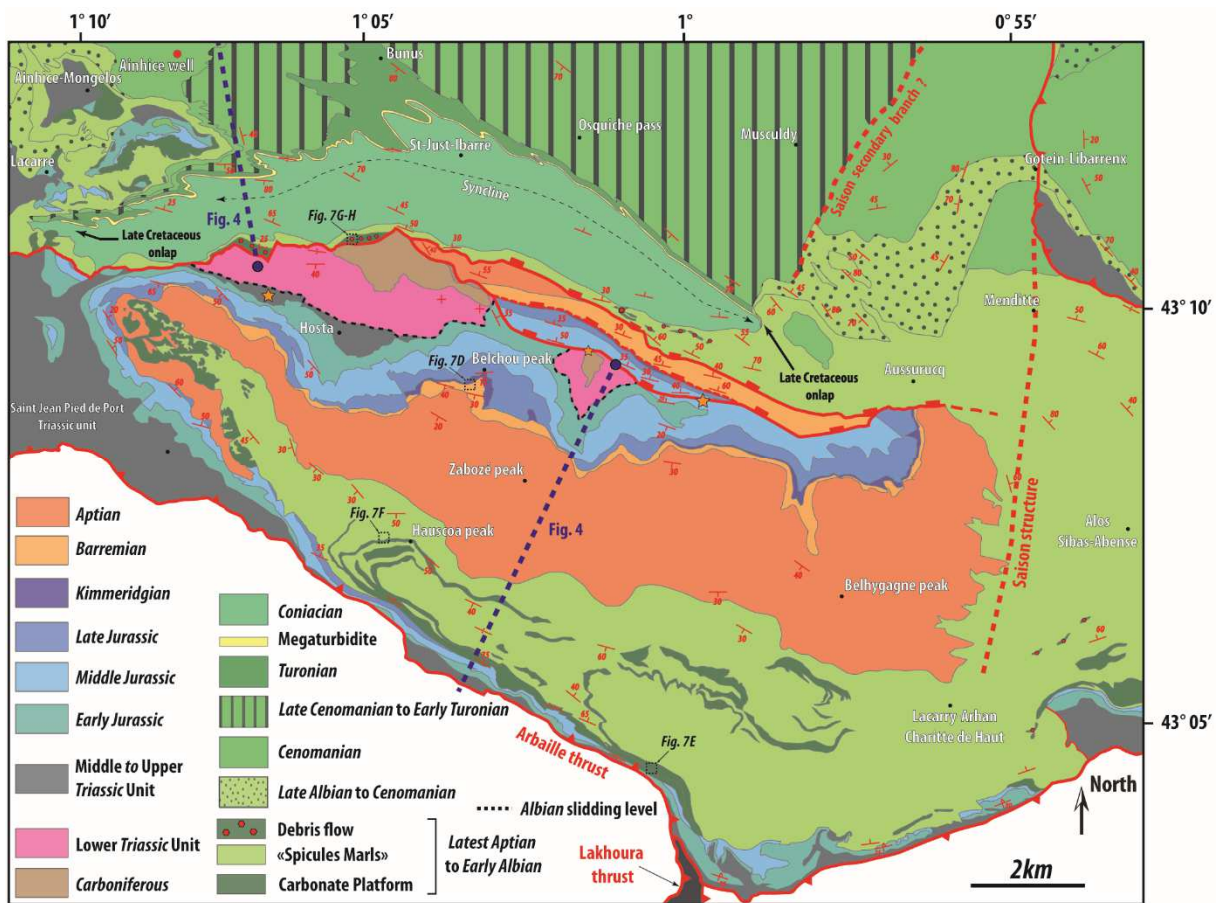
Figure. 8



1395

1396 Stereographic representation of the structural data showing the current and the Albian
 1397 orientation of the contacts between the *Mendibelza* conglomerates strata and the Paleozoic
 1398 sedimentary basement of the *Mendibelza* Unit. The sites are localized on the map Fig 5. The
 1399 first column represents the current *Mendibelza* Formation strata. Second column shows the
 1400 current orientation of the contacts between Mendibelza Formation strata and the Paleozoic
 1401 basement. The third column shows the Albian orientation of the contacts. The average value
 1402 of the Albian yellow surface is N131-35°NE. The NOF represented by the site F is
 1403 characterized by a sub-vertical Albian orientation. MP: Measurements planes.

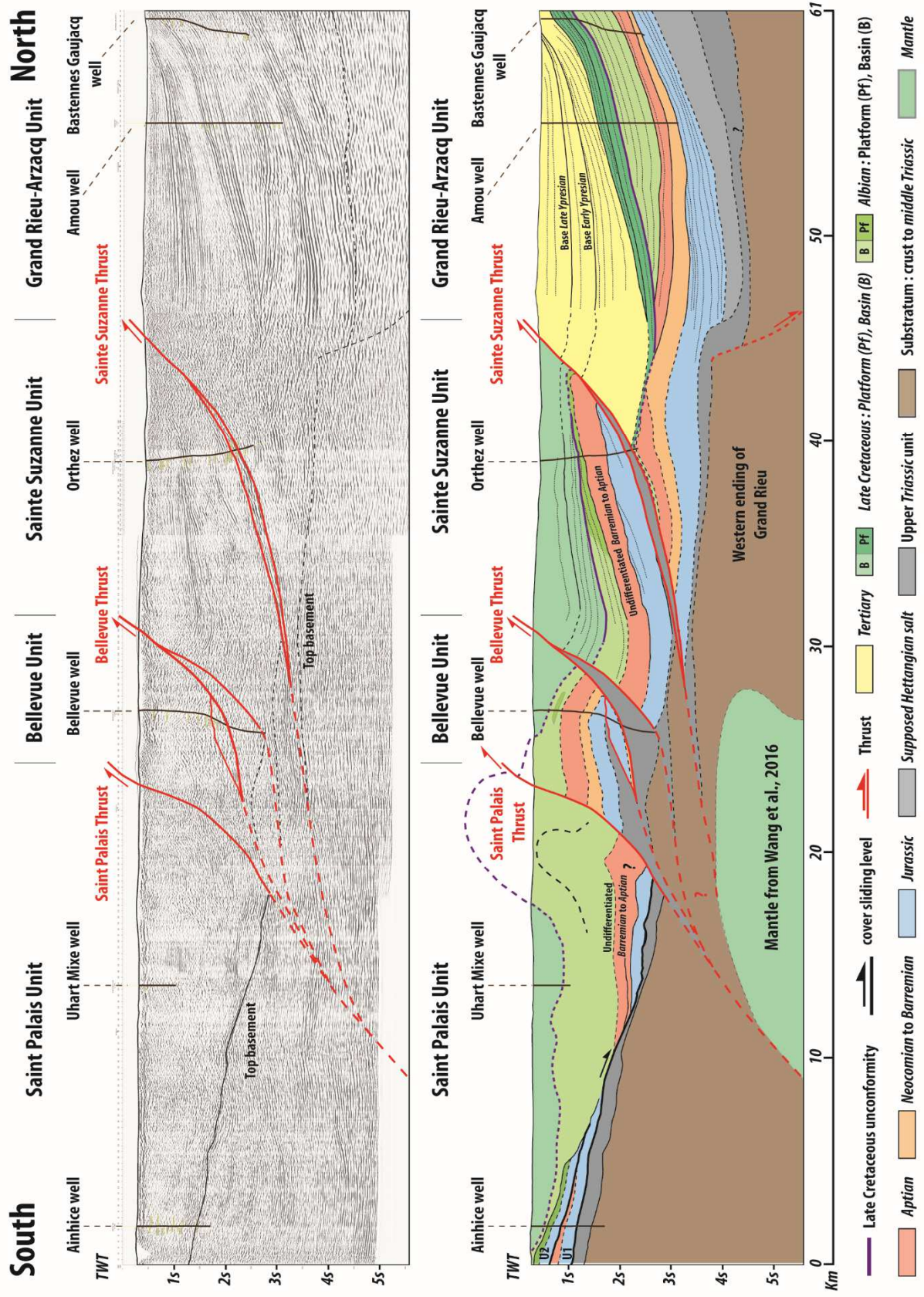
Figure. 9



1405

1406 Geological map of the *Arbailles* Unit and the southern part of the *Saint-Palais* Unit, using
 1407 updated BRGM 1/50 000 maps (Casteras et al., 1971; Boissonnas et al., 1974; Le Pochat et
 1408 al., 1976, 1978). The *Arbailles* Unit is separated from the *Saint-Palais* Unit by N120° normal
 1409 faults known as the North Arbailles Fault (NAF). These maps highlight the southward erosion
 1410 of the Jurassic cover and the southward back stepping of the Barremian to Aptian carbonate
 1411 platform, onlapping the previous eroded Jurassic deposits. In the North of the *Arbailles* Unit,
 1412 the Late Cretaceous deposits onlapped the Albian rift sequence towards the south from
 1413 Cenomanian to Santonian time.

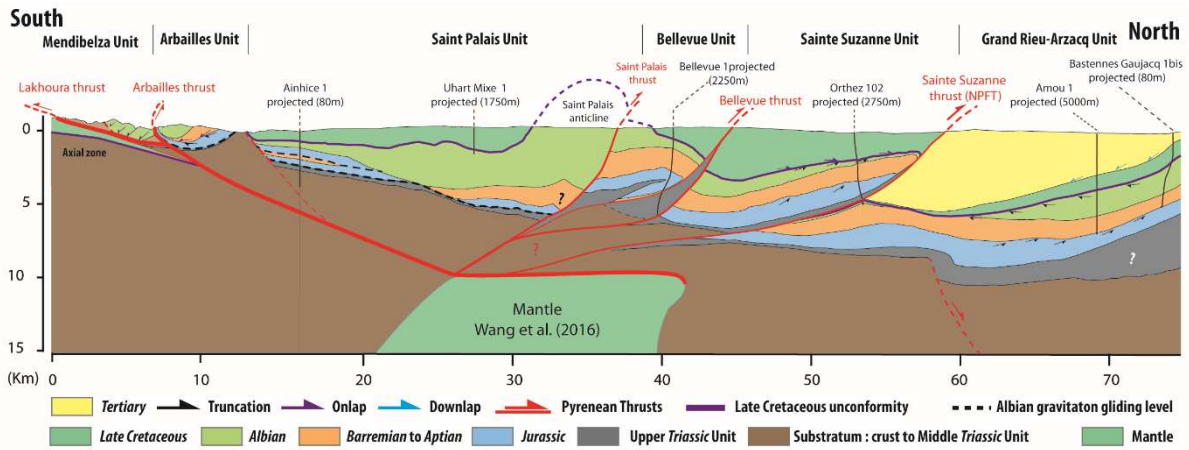
Figure. 10



1416 Interpretation of a N-S petroleum reflection composite seismic line, reprocessed by the
1417 BRGM in 2017 (MT104, MT112 and 83HBS02). This composite seismic line shows the deep
1418 geometry of the *Mauléon* Basin and its European margin. This seismic line through the
1419 *Mauléon* Basin cuts across four distinct structural units, separated from one another by
1420 thrusting overlaps. From south to north, the units are as follows: (1) *Saint-Palais*, (2)
1421 *Bellevue*, (3) *Sainte Suzanne*, and (4) *Grand-Rieul/Arzacq*. The *Saint-Palais* Unit corresponds
1422 to the Albian rift depocenter. The *Bellevue* and *Sainte Suzanne* Units materialized the
1423 northern conjugate margin of the Albian Cretaceous *Mauléon* rift basin, and the Barremian-
1424 Aptian depocenter.

1425

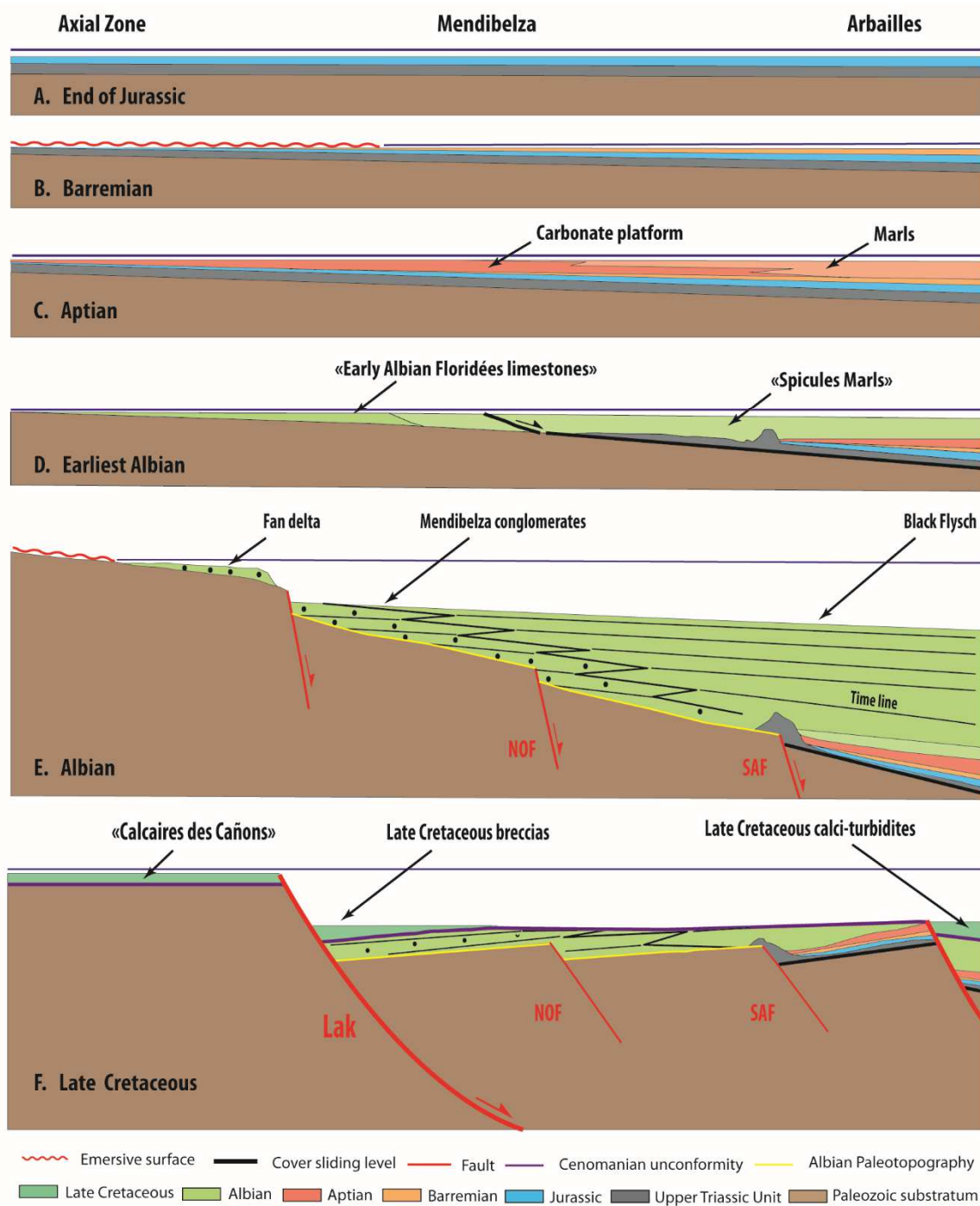
Figure. 11



1426

1427 N-S Composite cross-section of the *Mauléon* Basin. The southern part corresponds to the field
 1428 cross-section Fig. 4 and the northern part of the section is calibrated by the seismic line. The
 1429 *Mendibelza* and *Arbailles* Units correspond to the southern margin of the *Mauléon* Basin. The
 1430 *Saint-Palais* Unit represents the Albian rift depocenter. The *Bellevue* and *Sainte Suzanne*
 1431 Units materialized the northern rift margin of the *Mauléon* Basin and are northward overthrust
 1432 onto the *Arzacq* Basin.

Figure. 12

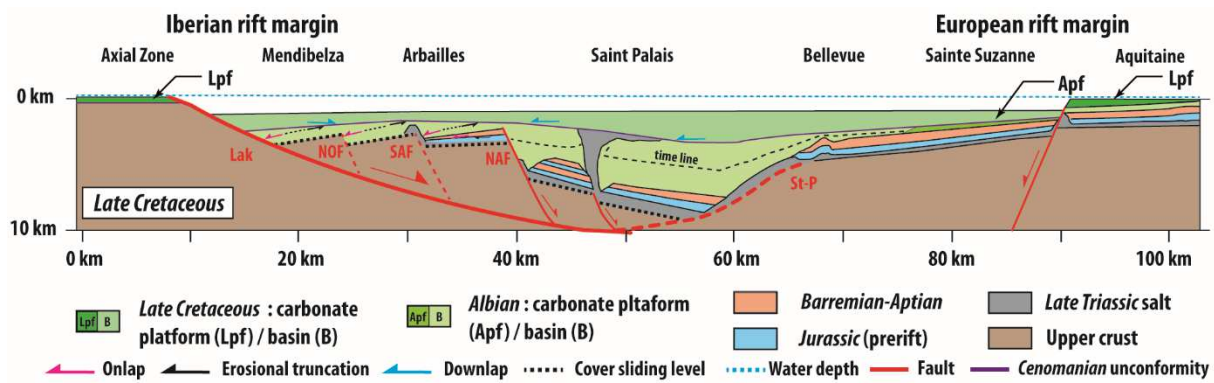


1435 Schematic tectono-sedimentary evolution of the southern margin of the *Mauléon* Basin from
 1436 the end of Jurassic to Late Cretaceous time. (A) During the Jurassic, this domain is
 1437 characterized by the development of a carbonate platform. (B) The Jurassic platform
 1438 underwent an emersion phase from the end of Jurassic to the base of Barremian time. This

1439 emersive phase was responsible for erosion of the previous carbonate platform, as shown by
1440 the southward erosional truncations affecting the Jurassic under the Early Cretaceous
1441 deposits. (C) The Early Cretaceous is characterized by a lack of Neocomian and the
1442 development of a flat carbonate platform during Barremian-Early Aptian time. During Late
1443 Aptian time, the previous carbonate platform distalized towards the north into more distal
1444 spicule marls, showing a slight tilt of the depositional profile. (D) At Early Albian time, the
1445 southern *Mauléon* margin was affected by the gravitational sliding of the Mesozoic cover
1446 towards the north leading to local diapirism. (E) The Albian time was characterized by
1447 differential vertical movements responsible for the uplift of the southern margin and tectonic
1448 subsidence towards the north. This stage is materialized by the sedimentation of the
1449 *Mendibelza* Fm. proximal turbiditic s.l. siliciclastic system, onlapping towards the south a
1450 Paleozoic basement inclined towards the north. The northward tilt of the Iberian margin is
1451 affected by southward propagating N120° normal faults. (F) At Cenomanian time, the
1452 *Mendibelza-Arbailles* Unit was affected by a tilt towards the south.

1453

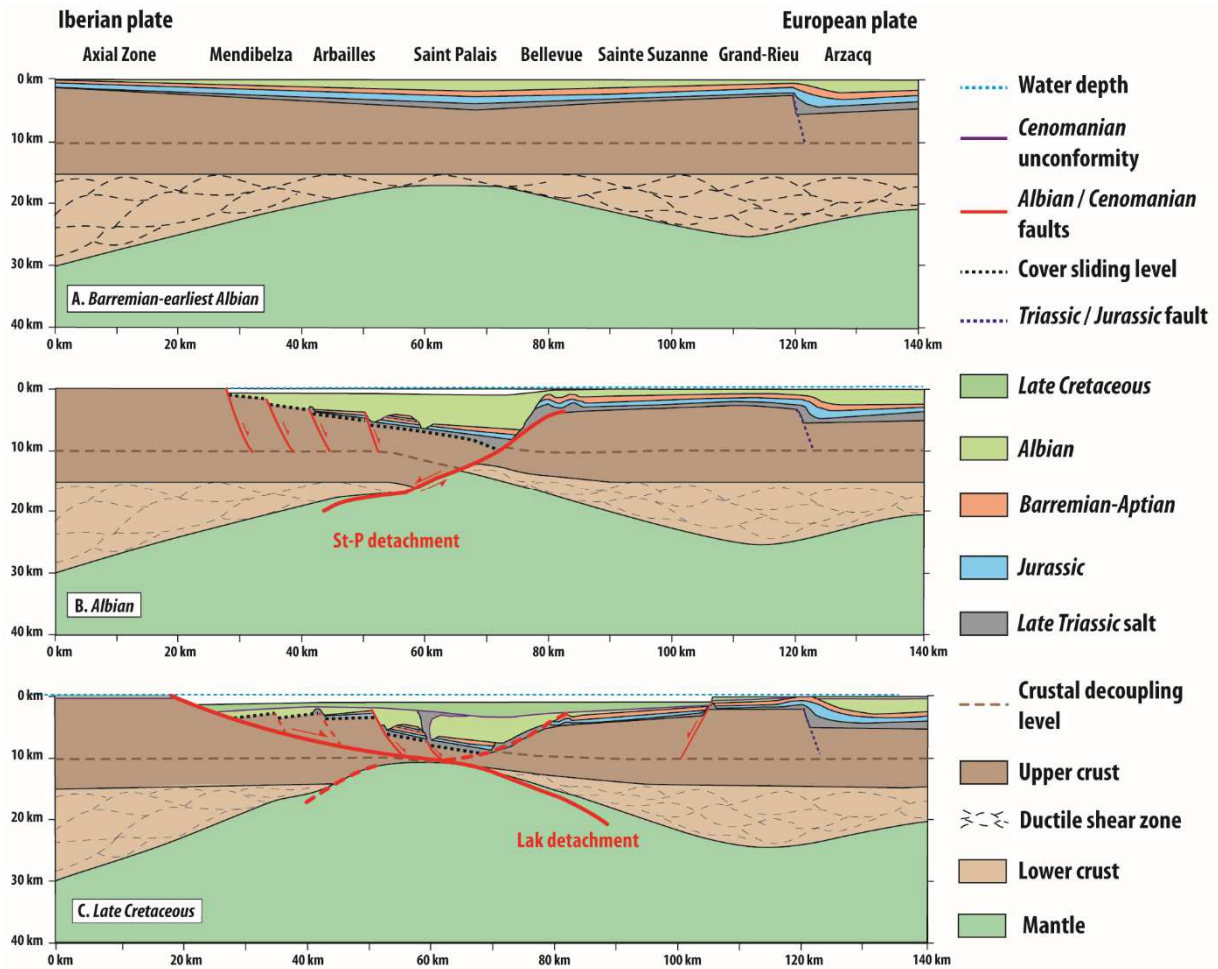
Figure. 13



1454

1455 Schematic cross-section of the *Mauléon* Basin during Late Cretaceous showing the
 1456 asymmetry of the rift basin margins and the geometry of the Cretaceous deposits. Lak:
 1457 Lakhoura detachment, NOF: North *Occabe* fault, SAF: South Arbailles fault, NAF: North
 1458 *Arbailles* fault, St-P: *Saint-Palais* detachment.

Figure. 14



1460

1461 Geodynamical model of crustal thinning of the *Mauléon* hyper-thinned rift basin. The genesis
 1462 of the *Mauléon* Basin comprises two lithospheric thinning stages. (A) Barremian to Aptian
 1463 rifting stage 1: "ductile pure-shear thinning phase" affecting mainly the lower crust, inducing
 1464 the formation of a symmetric synrift sag basin. (B) Albian rifting stage 2: simple-shear
 1465 thinning phase inducing the development of an asymmetric synrift basin, characterized by
 1466 proximal turbiditic system on the southern margin and carbonate platform deposits on the
 1467 northern one. The *Saint-Palais* southward detachment fault is responsible for the hyper-
 1468 thinning of the sub-continental crust during Albian time. (C) Early Cenomanian rifting stage
 1469 3: the southern margin was tilted towards the south along a northward dipping detachment

- 1470 fault responsible for local sub-continental mantle denudation, resulting in the formation of an
- 1471 apparent symmetric rift basin. St-P: *Saint-Palais*, Lak: *Lakhoura*.