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1 **Review of Iberia-Eurasia plate-boundary basins: Role of**
2 **sedimentary burial and salt tectonics during rifting and**
3 **continental breakup**

4 Nicolas Saspiturry^{1*}, Benoit Issautier², Philippe Razin¹, Thierry Baudin², Riccardo Asti³, Yves
5 Lagabrielle³, Cécile Allanic², Olivier Serrano² and Thibault Duret²

6 ¹ Université Bordeaux Montaigne / EA 4592 Géoressources & Environnement, 1 allée Fernand
7 Daguin, 33607 Pessac cedex, France

8 ² Université Rennes, CNRS, Géosciences Rennes - UMR 6118, F-35000, Rennes, France

9 ³ BRGM-French Geological Survey, 3 Avenue Claude Guillemin, 45100 Orléans, France

10 *Corresponding author (e-mail: saspiturry.nicolas@gmail.com)

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22 **Keywords**

23 Iberian-European boundary; Sedimentary burial; Salt tectonics; Depth-dependent thinning;
24 Ductile regime; HT/LP metamorphism

25 **Abstract**

26 We document the role of sedimentary burial and salt tectonics in controlling the deformation
27 style of continental crust during hyperextension. The Iberian-European boundary records a
28 complex history of Cretaceous continental extension, which has led to the development of so-
29 called smooth-slope type basins. Based on the review of the available geological constraints
30 (crustal-balanced cross sections, sedimentary profile evolution, RSCM thermometer, low-
31 temperature thermochronology) and geophysical data (Bouguer anomaly, Moho depth,
32 seismic reflection profiles, and Vp/Vs velocity models) on the Tartas, Arzacq, Cameros,
33 Parentis, Columbrets, Mauléon, Basque-Cantabrian and Internal Metamorphic Zone basins,
34 we shed light on the main characteristics of this type of basin. This synthesis indicates that
35 crustal thinning was influenced by two decoupling horizons: the middle crust and Triassic
36 pre-rift salt, initially located between the basement and pre-rift sedimentary cover. These two
37 horizons remained active throughout basin formation and were responsible for depth-
38 dependent thinning of the crust and syn-rift salt tectonics. We therefore identify several
39 successive deformation phases involving (1) pure shear dominated thinning, (2) simple shear
40 dominated thinning and (3) continental breakup. In the first phase, distributed deformation
41 resulted in the development of a symmetric basin. Field observations indicate that the middle
42 and lower crust were under dominantly ductile conditions at this stage. In the second phase,
43 deformation was localized along a crustal detachment rooted between the crust and the mantle
44 and connecting upwards with Triassic pre-rift salt. During continental breakup, basin
45 shoulders recorded the occurrence of brittle deformation while the hyperextended domain
46 remained under predominantly ductile thinning. The formation of smooth-slope type
47 extensional basins was intrinsically linked to the combined deposition of thick syn-rift and
48 breakup sequences, and regional salt tectonics. They induced significant burial and allowed

49 the continental crust and the pre-rift sequence to deform under high temperature conditions
50 from the rifting to continental breakup stages.

51 **1 Introduction**

52 Recent models and concepts related to hyperextended continental margins have been defined
53 on North Atlantic passive margins such as West Iberia/Newfoundland (Boillot et al., 1980,
54 1995; Reston et al., 1995; Whitmarsh et al., 2001; Péron-Pinvidic & Manatschal, 2009; Péron-
55 Pinvidic et al., 2015), Norway/Greenland (Osmundsen et al., 2002; Osmundsen & Ebbing,
56 2008; Osmundsen & Péron-Pinvidic, 2018), and the Alps (Lemoine et al., 1987; Froitzheim &
57 Manatschal, 1996; Manatschal & Nievergelt, 1997; Manatschal et al., 2006; Masini et al.,
58 2011; Mohn et al., 2014). They describe the mechanisms responsible for continental crust
59 thinning and subcontinental mantle exhumation at the ocean-continent transition, and have
60 shown that the hyperextended domain of continental margins is characterized by (1) sediment
61 starved conditions with very thin pre-rift and syn-rift sedimentary sequences, (2) asymmetric
62 margins, and (3) coupled deformation between the basement and sedimentary cover resulting
63 in the formation of extensional allochthons. The syn-rift and post-rift units have been
64 respectively defined as strata deposited during (1) the last extension episode leading to
65 continental breakup and (2) a period of relative tectonic quiescence above a regional breakup
66 unconformity (Driscoll et al., 1995; Wilson et al., 2001). DSDP and ODP drilling expeditions
67 along the West-Iberia margin have also contributed to a more complete understanding of how
68 continental breakup is recorded in syn-rift to post-rift sedimentary sequences (Alves & Cunha,
69 2018), and how it evolves through time along the strike of continental margins (Alves et al.,
70 2009). Indeed, such data allowed the identification of syn-rift, breakup and post-rift sequences
71 as correlating with major phases of crustal thinning, mantle denudation and oceanic spreading
72 (Alves et al., 2009; Soares et al., 2012; Soares, 2014; Alves & Cunha, 2018; Alves et al.,

73 2020). Continental breakup is thus defined as the deposits that are correlated with the phases
74 of hyperextension and mantle denudation on magma-poor continental margins (Soares et al.,
75 2012). Alves et al. (2020) have demonstrated that continental breakup is only possible along
76 magma-poor Type I margins (sensu Huisman & Beaumont, 2003, 2008, 2011, 2014). Indeed,
77 the upper lithosphere in Type I margins is much more coupled to the continental crust than in
78 Type II margins (Alves et al., 2020). Thus, along Type II margins, the upper lithosphere is
79 laterally subtracted and mantle denudation cannot occur. Syn-rift and breakup sequences are
80 delimited by major unconformities corresponding (from bottom to top) to the base syn-rift
81 unconformity (BSU; base of syn-rift sequence), the lithospheric breakup surface (LBS; base
82 of the breakup sequence) and the base post-rift unconformity (BPU; base of the post-rift
83 sequence) as defined by Soares et al. (2012) and Soares (2014). While Atlantic-type margins
84 develop under conditions of high mantle heat flow, as the product of asthenosphere upwelling
85 and subsequent denudation of subcontinental mantle (e.g., Huisman & Beaumont, 2003,
86 2008, 2011, 2014; Lavie & Manatschal, 2006; Brune et al., 2014, 2016), the North Pyrenean
87 hyperextended rift system differs from the Atlantic type by the presence of high-
88 temperature/low-pressure (HT/LP) metamorphism (Albarède & Michard-Vitrac, 1978a,
89 1978b; Golberg et al., 1986; Montigny et al., 1986; Golberg & Maluski, 1988; Golberg &
90 Leyreloup, 1990; Thiébaud et al., 1992). The North Pyrenean rift basins were affected by a
91 syn-rift stage throughout the Early Cretaceous that evolved to continental breakup and mantle
92 denudation during the Latest Albian to Early Cenomanian (Debroas et al., 2010; Saspiturry et
93 al., 2019a; Espurt et al., 2019; Labaume & Teixell, 2020). Thus, these margins can be defined
94 as Type I margins as mantle denudation occurred during continental breakup (sensu Huisman
95 & Beaumont, 2003, 2008, 2011, 2014). The hyperextended domain of these basins is
96 interpreted to have stretched under high-temperature conditions under a predominantly ductile
97 regime, resulting in the generation of large-scale boudins whose formation was controlled by

98 a strong thermal and structural inheritance (Clerc & Lagabrielle, 2014; Clerc et al., 2015,
99 2015b; Corre et al., 2016; de Saint Blanquat et al., 2016; Teixell et al., 2016; Lagabrielle et
100 al., 2016, 2019; Asti et al., 2019). The major difference between the hyperextended Atlantic-
101 type and Pyrenean Cretaceous rift margins is the rare occurrence of tilted crustal blocks and
102 related stepping fault scarps in the central parts of the basins, thus defining a dominantly
103 symmetrical basement profile with smooth slopes (see Lagabrielle et al., 2020 for a review).
104 Several other rift basins apart from the North Pyrenean rift basins exist along the plate
105 boundary separating Iberia from the Eurasia plate and share several features in common with
106 the Pyrenean system. They include the Tartas, Arzacq, Columbrets, Cameros and Parentis
107 syn-rift basins, which developed during Late Jurassic to Early Cretaceous rifting (more details
108 in the review section). The development of these basins is time-equivalent to the Late
109 Jurassic–Aptian rifting and Albian-Cenomanian breakup sequence of the V-shaped Bay of
110 Biscay margin (Montadert et al., 1979; Barbier et al., 1986; Thinon, 1999; Vergés & García-
111 Senz, 2001; Thinon et al., 2003; Tugend et al., 2014) and the NW-Iberia margin (Soares et al.,
112 2012; Alves & Cunha, 2018; Alves et al., 2020). All of these basins share similarities that
113 include (1) a small amount of brittle deformation affecting the upper crust, (2) a pre-rift cover
114 that is efficiently decoupled from its substratum thanks to the thick layer of Late Triassic to
115 earliest Jurassic evaporites, (3) hyperthinned continental crust resulting from a polyphase
116 rifting history, (4) a thick pre-rift to syn-rift sedimentary pile and (5) HT/LP syn-rift
117 metamorphism. This paper presents a structural and sedimentological review of these smooth-
118 slope type basins that sheds light on the major role of sedimentary burial and decoupling
119 levels in controlling the ductile crustal thinning of the Iberia-Eurasia plate boundary basins
120 during the syn-rift and breakup stages. We also compare the processes responsible for
121 continental crust thinning and discuss the timing of formation of the smooth-slope type basins

122 with respect to the evolution of the West Iberia and Bay of Biscay hyperextended continental
123 margins.

124 **2 Mesozoic basins of Iberia and Europe**

125 **2.1 Aquitaine basin and the Tartas and Arzacq basins**

126 The Aquitaine basin started to develop during the Late Permian in an extensional context
127 related to the breakup of Pangea (Burg et al., 1994). During the Triassic and earliest Jurassic
128 (Hettangian), extension led to an aborted rift basin filled with clastic rocks, carbonates, and an
129 evaporite sequence ~1–2 km thick (Curnelle et al., 1982; Curnelle, 1983; Curnelle & Dubois,
130 1986). The Jurassic was a tectonically stable period marked by the development of a
131 widespread carbonate platform (Peybernès, 1976). At the end of the Jurassic, the entire
132 platform was confined as suggested by the deposition of restricted dolomite and anhydrite
133 facies (BRGM et al., 1974; Serrano et al., 2006). The Jurassic pre-rift carbonate platform was
134 uplifted, weathered and partly eroded (Combes et al., 1998; James, 1998; Canérot et al., 1999)
135 in response to asthenosphere upwelling that preceded Early Cretaceous rifting (Saspiturry et
136 al., 2019a). The Early Cretaceous was a structurally active time in which the Aquitaine basin
137 evolved into different syn-rift basins characterised by very rapid subsidence (up to 130
138 m/Myr; Désaglaux & Brunet, 1990; Brunet, 1991), including the symmetric Arzacq and
139 Tartas basins. A ~1–3 km thick Early Cretaceous syn-rift sedimentary sequence
140 unconformably overlies pre-rift deposits of Triassic and Jurassic age totalling 2–3 km in
141 thickness (Serrano et al., 2006).

142 **2.1.1 Tartas basin**

143 The Tartas basin is located in the southern part of the Aquitaine domain (Fig. 1). It is bounded
144 to the north by the Celtaquitaine Flexure (BRGM et al., 1974), which separates the Tartas
145 Basin and the North Aquitaine Platform, and to the south by the Audignon-Pécorade-Antin

146 Maubourguet ridge. The Tartas basin is a relatively narrow (20 km) east-west elongated
147 trough (Serrano et al., 2006) that lies on slightly thinned continental crust (20–25 km; Wang
148 et al., 2016). It formed symmetrically during the Berriasian with the development of
149 widespread shallow-water facies and continued this pattern throughout Early Cretaceous
150 continental rifting. Sediments from this period range from marginal littoral facies (Berriasian
151 to Barremian; e.g., Issautier et al., 2018) to inner-platform facies (Aptian and Albian; e.g.,
152 Delfaud, 1969; Arnaud-Vanneau et al., 1979; Bouroullec et al., 1979; Peybernès, 1979, 1982).
153 Throughout the Early Cretaceous rifting, the balance between carbonate production and basin
154 subsidence in the Tartas basin led to the aggradation of a shallow-marine carbonate
155 succession 2500 m thick. The resulting palaeogeography was a homogeneous, flat-floored
156 basin with no major brittle deformation recorded within the basin. High subsidence rates are
157 consistent with ductile pure shear thinning of the lower/middle crust and formation of a
158 symmetric syn-rift basin (Issautier et al., 2020). This regime was probably enhanced by the
159 presence of a thick evaporite layer of Rhaetian-Hettangian age that favoured mechanical
160 decoupling between the pre-rift sediments and the basement. This salt blanket may account
161 for the lack of illustrated brittle structures in the syn-rift succession of the Tartas basin
162 (Issautier et al., 2020).

163 **2.1.2 Arzacq basin**

164 The Arzacq syn-rift basin lies to the south of the Tartas basin (Fig. 1). Similarly to the
165 Tartas basin, it is 40 km wide, elongated in the N110° direction and overlies slightly thinned
166 continental crust nearly 25 km thick (Wang et al., 2016). The Arzacq basin is bounded to the
167 north by the Audignon-Pécorade-Antin Maubourguet ridge (Mauriaud, 1987; Mediavilla,
168 1987; Serrano et al., 2006) and to the south by the North Pyrenean Frontal Thrust
169 (Choukroune & ECORS Team, 1989; Daignières et al., 1994; Bourrouilh et al., 1995). Its
170 southern flank coincides with north-dipping normal faults on the south side of the Grand Rieu

171 ridge (Fig. 2) (e.g., Serrano et al., 2006; Masini et al., 2014; Gómez-Romeu et al., 2019;
172 Saspiturry et al., 2019a). Early Mesozoic rifting (Triassic to Hettangian) was responsible for a
173 high subsidence rate and the deposition of a thick anhydrite sequence (Curnelle, 1983). The
174 basin's subsequent rifting history was characterised by two successive extensional regimes,
175 the first one symmetric (Berriasian to Aptian) and the second one slightly asymmetric
176 (Albian) (Issautier et al., 2020). The symmetric syn-rift stage, identical to that of the Tartas
177 basin, was accommodated by ductile flow of the lower/middle continental crust. Again, no
178 brittle structures are recorded, probably because deformation was decoupled by the thick salt
179 blanket (Canérot et al., 2005; Duretz et al., 2019). From Berriasian to Barremian times, the
180 palaeogeography was dominated by a shallow marginal carbonate platform (e.g., Bouroullec
181 & Deloffre, 1970; Peybernès & Combes, 1994; Biteau et al., 2006; Biteau & Canérot, 2007).
182 In late Aptian times, the basin differentiated with the growth of a median marl flanked by
183 carbonate-reef and inner-platform deposits (Delfaud & Gautier, 1967; Delfaud & Villanova,
184 1967). This second stage began during the Albian with a sudden change in both
185 palaeogeography and basin geometry, when the depositional profile became asymmetric and
186 salt tectonics affected its flanking ridges (Fig. 2; Issautier et al., 2020). Halokinesis consisted
187 of gliding displacement of the Grand Rieu sedimentary cover on the southern side of the basin
188 (Fig. 2), leading to the development of salt-detached ramp synclines as defined by Jackson
189 and Hudec, (2005) and Pichel et al. (2018). Salt diapirism in the Audignon district in the
190 northern side of the basin accompanied this stage (Mauriaud, 1987; Mediavilla, 1987;
191 Issautier et al., 2020). As the syn-rift stage continued, the gliding of the sedimentary cover
192 occurred. Motion resulted from simple shear strain localisation along an Albian detachment
193 fault that followed ductile pure shear thinning of the lower/middle crust during the Barremian-
194 Aptian interval (Issautier et al., 2020). The subsidence history of the Arzacq syn-rift basin
195 also shows that the second stage coincided with an unusually steep geothermal gradient that

196 ended around the initiation of the Pyrenean compression (Angrand et al., 2018). The post-rift
197 stage was marked by the deposition of Cenomanian shallow-water carbonates resting
198 unconformably on Albian strata. Although this basin underwent simple shear thinning during
199 the Albian, it did not reach the stage of continental breakup.

200 **2.2 Cameros basin**

201 The Cameros basin, on the northwestern edge of the Iberian mountain chain (Fig. 1), is a
202 WNW-ESE trending synclinorium 80 km wide and 120 km long resulting mostly from Early
203 Cretaceous rifting (Guimerà et al., 1995) (Fig. 3). The basin infill has been thrust onto
204 neighbouring Cenozoic basins, the Ebro basin to the north and the Duero basin to the south
205 (Fig. 3E). It underwent separate rifting stages from Permian to Triassic (Alvaro et al., 1979)
206 and Late Jurassic to Early Cretaceous times (Platt, 1990; Mas et al., 1993; Casas-Sainz & Gil-
207 Imaz, 1998; Salas et al., 2001). The Late Jurassic to Early Cretaceous rifting stage was
208 followed by Late Albian thermal subsidence recording the post-rift stage (Omodeo Salè et al.,
209 2014; Omodeo-Salé et al., 2014, 2017). The associated Late Jurassic to Early Cretaceous syn-
210 rift sequence ranges from 6.5 to 8 km in thickness (Fig. 3) (Casas-Sainz & Gil-Imaz, 1998;
211 Mas et al., 2011; Omodeo-Salé et al., 2014; García-Lasanta et al., 2017).

212 The base of the Mesozoic sedimentary pile of the Cameros basin is a 500 m thick Late
213 Triassic salt sequence (Fig. 3) (Casas Sainz, 1993; Casas-Sainz & Gil-Imaz, 1998). A pre-rift
214 Jurassic sequence reaching 800 m in thickness, composed of shallow marine carbonate
215 platform deposits, lies unconformably over both the Triassic salt and the Paleozoic basement
216 (Valladares, 1980; Platt, 1990; Aurell & Meléndez, 1993; Aurell et al., 2003). The syn-rift
217 sequence begins with 3–6 km of fluvial to lacustrine deposits and rare unconformable marine
218 layers of Late Jurassic to Barremian age (Guiraud & Séguret, 1985; Platt, 1990; Alonso &
219 Mas, 1993; Mas et al., 1993, 2011; Quijada et al., 2010; Suárez González et al., 2010). The
220 late Barremian to early Aptian depositional profile is characterized by lithologies typical of

221 coastal wetlands with both freshwater and marine influences, grading laterally to alluvial and
222 fluvio-lacustrine deposits (Platt, 1989, 1986). These syn-rift deposits show that the basin was
223 continuously shallow. The syn-rift sequence thins gradually toward the basin's flanks, where
224 it laps onto the pre-rift Jurassic sequence (Fig. 3). An apparent northward migration of the
225 depocenter suggests a slight asymmetry in the basin (Casas-Sainz & Gil-Imaz, 1998; Guimerà
226 et al., 1995; Mas et al., 1993; Omodeo-Salé et al., 2014, 2017). This northward migration of
227 the depocenter likely indicates that simple shear deformation was concentrated along a
228 detachment fault. This fault was reactivated as a reverse fault during the Pyrenean
229 compression, inducing the northward thrusting of the Cameros basin onto the Ebro flexural
230 basin (Fig. 3) (Omodeo-Salé et al., 2014).

231 The Cameros basin is characterised by (1) the perfect continuity of the pre-rift Jurassic
232 sequence, that is not fragmented along the reactivated detachment (Fig. 3), (2) the absence of
233 normal faulting affecting both basement and cover (Fig. 3) and (3) a lack of significant offset
234 of the top of the basement (Casas-Sainz & Simón-Gómez, 1992; Casas Sainz, 1993; Casas-
235 Sainz & Gil-Imaz, 1998; Casas et al., 2000, 2009; Omodeo-Salé et al., 2014). The basin has
236 therefore been interpreted as an extensional ramp syncline, formed above a décollement in the
237 Late Triassic evaporites rooting at depth on a blind south-dipping extensional ramp or crustal
238 detachment (Mas et al., 1993, 2011; Guimerà et al., 1995; Casas-Sainz & Gil-Imaz, 1998;
239 Casas et al., 2009; García-Lasanta et al., 2017; Omodeo-Salé et al., 2017). This interpretation
240 explains the northward migration of the syn-rift depocenter and associated edgeward onlap
241 onto the pre-rift deposits, and the development of a synformal rift basin (Fig. 3). The pre-rift
242 Jurassic deposits were stretched by this fault movement, but remained continuous (Casas et
243 al., 2009). The Triassic evaporite layer accommodated most of the shear strain during
244 extension, leaving both cover and basement well preserved. The resulting basinward gliding

245 of the pre-rift cover was accompanied by thinning of the Triassic evaporites and by salt
246 diapirism (Casas Sainz, 1993; Casas-Sainz & Gil-Imaz, 1998; Rat et al., 2019).

247 The Cameros syn-rift basin exhibits effects of HT/LP metamorphism in its deepest part, with
248 temperatures that reached around 350–400°C (Fig. 3E) (Guiraud & Séguret, 1985; Golberg et
249 al., 1988; Rat et al., 2019). The rifting stage developed under a high thermal gradient
250 estimated at around 70°C/km, assuming a sediment thickness of 8 km (Mata et al., 2001; Del
251 Río et al., 2009). Toward the basin's northern edge, the palaeogeothermal gradient decreases
252 to 41.5°C/km, along with the intensity of the HT/LP metamorphism (Fig. 3F) (Omodeo-Salé
253 et al., 2017). This lower thermal gradient is consistent with an estimated heat flow of
254 approximately 60–65 mW/m² (Omodeo-Salé et al., 2017). The HT/LP metamorphism reached
255 its peak temperature during the early post-rift stage and developed coevally with continental
256 breakup in the North Pyrenean Zone basins (Golberg et al., 1988; Casquet et al., 1992; Casas-
257 Sainz & Gil-Imaz, 1998; Mata et al., 2001).

258 **2.3 Parentis Basin**

259 The E-W elongated Parentis basin lies between the Landes High (Ferrer et al., 2009) to the
260 south and the Armorican Arc to the north (Lefort & Agarwal, 1999). This wedge-shaped basin
261 opens westward to the eastern edge of the Bay of Biscay continental margin characterized by
262 exhumed subcontinental mantle at the ocean-continent transition (Fig. 1) (Pinet et al., 1987;
263 Bois & ECORS Scientific team, 1990; Bois & Gariel, 1994; Jammes, 2009; Tugend et al.,
264 2014). The south edge of the basin is defined by the north-dipping Ibis fault (Fig. 4; Ferrer et
265 al., 2008). The continental crust of the future Parentis basin underwent several extensional
266 deformations during the Permian to Early Triassic period (Dardel & Rosset, 1971; Mathieu,
267 1986; Ferrer et al., 2009; Biteau et al., 2006). Extension was accompanied by the deposition
268 of a 1–3 km thick sequence of Late Triassic to Early Jurassic evaporites (Fig. 4D; Curnelle,

269 1983; Ferrer et al., 2009). The basin was subsequently filled with a 10 km thick sequence of
270 pre-rift Jurassic and Early Cretaceous syn-rift shallow platform carbonates and terrigenous
271 sediments that rests upon the Late Triassic evaporites (Fig. 4D; Montadert & Winnock, 1971;
272 Bourrouilh et al., 1995; Bois et al., 1997). Tectonic subsidence occurred in the Parentis basin
273 during a latest Jurassic to Early Albian syn-rift stage, followed by post-rift thermal subsidence
274 during a latest Albian to Late Cretaceous post-rift stage (Brunet, 1994; Ferrer et al., 2008).

275 The Parentis basin is characterised by a hyperextended continental crust with a Moho depth of
276 about 10 km in the thinnest portion (Bois & ECORS Scientific team, 1990; Bois, 1992; Ferrer
277 et al., 2008). The overall basin geometry shows gently dipping margins lacking major normal
278 fault scarps (Fig. 4D). The lower continental crust appears to be absent or very thin in the
279 hyperextended domain, while the proximal margins include both upper and lower crust (Fig.
280 4D; Pinet et al., 1987b; Marillier et al., 1988; Tomassino & Marillier, 1997; Ruiz, 2007). A
281 positive Bouguer gravity anomaly coincides with the hyperextended domain of the Parentis
282 rift basin (Pinet et al., 1987a), which originated during the latest Jurassic to Early Cretaceous
283 rifting affecting the peri-Pyrenean realm (Jammes et al., 2009; Tugend et al., 2015).

284 The ECORS deep seismic profiling project documented the symmetrical synclinal shape of
285 the Parentis basin and the paucity of normal faults in the stretched crust as well as in the
286 proximal rift margins (Fig. 4D) (Pinet et al., 1987b; Bois et al., 1997; Bois & Courtillot,
287 1988). Once the syn-rift stage was well established, the Parentis basin sedimentary profile
288 became slightly asymmetrical in response to simple shear localization along a crustal
289 detachment during Albian time (Fig. 4D; Pinet et al., 1987a; Jammes, 2009). This evolution
290 stage is comparable to the one identified in the Arzacq basin (Issautier et al., 2020). The
291 southern and northern margins of the rift have been interpreted by various workers as parts of
292 an asymmetric opening system (Jammes et al., 2010a, 2010b, 2010c; Masini et al., 2014;
293 Tugend et al., 2014). However, Pinet et al. (1987b) argued that the location and the geometry

294 of the thinned zone make it difficult to apply a classical simple shear model to the Parentis
295 basin. They proposed that mantle uplift induced stretching (i.e. active rifting) and ductile flow
296 in the lower crust and consequently a decoupling between the upper and lower crust. This
297 depth-dependent crustal thinning explains the discrepancy between the slight extension at the
298 surface and the substantial thinning of the lower/middle crust (thinning ratio greater than 6) at
299 depth (Fig. 4D). This interpretation implies that the crust beneath the Parentis basin reached
300 the ductile strain regime and was thinned under high-temperature conditions, as was the case
301 in the adjacent North Pyrenean Zone. Finally, according to various authors, the Parentis basin
302 first developed as a latest Jurassic–Aptian symmetrical rift that became asymmetrical when
303 thinning progressively occurred through simple shear concentrated along a ductile crustal
304 shear zone during Albian time. This suggests that the processes responsible for the Parentis
305 basin continental crust thinning bear similarities with the ones defined in the Arzacq and
306 Cameros syn-rift basins.

307 Jammes et al. (2010c) highlighted the major role played by the thick pre-rift salt sequence in
308 decoupling the deformation between the basement and the rest of the Mesozoic sedimentary
309 cover. The southern margin of the Parentis basin underwent gravity-driven cover gliding
310 followed by syn-rift thin-skinned extensional faulting along a décollement plane within the
311 salt (Fig. 4D; Tugend et al., 2014). This process induced the development of syn-rift salt
312 anticlines and welded diapirs affecting the Mesozoic sedimentary pile (Fig. 4D; Mathieu,
313 1986; Mediavilla, 1987; Ferrer et al., 2008). Moreover, Ferrer et al. (2012) reported that salt
314 structures are mainly localised on basin flanks (Fig. 2).

315 **2.4 Columbrets basin**

316 The Columbrets offshore basin is the southwestern part of the Valencia Trough between Spain
317 and the Balearic Islands (Figs. 1 & 5A). This ENE-WSW trending basin represents a mildly

318 inverted and thus exceptionally preserved hyperextended rift of Late Jurassic to Early
319 Cretaceous age (Fig. 5; e.g., Etheve et al., 2018). The pre-rift and syn-rift successions occupy
320 a large-scale synclinal basin with thinned borders, shaped by displacement along extensional
321 detachments (Figs. 5D-E). This domain underwent a polyphase rifting history that spanned
322 three major rifting events. During the Late Permian to Early Triassic, distributed deformation
323 formed an intracontinental rift basin filled by continental deposits (Arche & López-Gómez,
324 1996; Vargas et al., 2009). This first rifting stage was followed by a Late Triassic–Early
325 Jurassic rifting event related to the opening of the Alpine-Ligurian Tethys (Jiménez-Munt et
326 al., 2010; Frizon de Lamotte et al., 2011; Schettino & Turco, 2011). The climax of this rifting
327 event was marked by the deposition of a thick layer of evaporites that became a major
328 décollement during subsequent events (Ortí, 1974; Ortí et al., 2017). The Jurassic post-rift
329 sequence is mainly composed of shallow-water limestone (e.g., Roca, 1996). Partial crustal
330 thinning occurred during the first rifting stage (Salas et al., 2001; Nebot & Guimerà, 2016;
331 Etheve, 2016; Etheve et al., 2018; Roma et al., 2018).

332 The main rifting stage leading to the hyperthinning of the continental crust in the Columbrets
333 basin (Fig. 5D) occurred during the Late Jurassic to early Albian. The syn-rift succession
334 consists of platform carbonate deposits, which grade basinward into deep-water marl and give
335 way locally toward the basin's flanks to fluvial and deltaic deposits (Etheve et al., 2018 and
336 references therein). The platform limestones preserve the record of post-rift thermal
337 subsidence from the middle Albian to the Late Cretaceous (Salas et al., 2001; Nebot &
338 Guimerà, 2016; Etheve et al., 2018).

339 The Moho depth is 25–30 km under the margins of the Columbrets basin and only 8–10 km
340 under the central portion (Fig. 5A; Gallart et al., 1990; Banda & Santanach, 1992; Dañobeitia
341 et al., 1992; Torné et al., 1996; Vidal et al., 1997; Ayala et al., 2003, 2015; Gomez-Ortiz et
342 al., 2011; Etheve et al., 2018). The thickness of the continental crust reaches a minimum of

343 3.5 km in the hyperextended domain (Fig. 5D; Etheve et al., 2018), where it coincides with an
344 unusually strong Bouguer gravity anomaly ranging in amplitude between 60 and 80 mGal
345 (Fig. 5B; Ayala et al., 2015). Both features are consistent with the shallowing of the
346 lithosphere-asthenosphere boundary (60–65 km depth) indicated by geoid modelling and 3-D
347 gravity data consistent with extreme crustal thinning (Zeyen & Fernández, 1994; Ayala et al.,
348 1996, 2003; Carballo et al., 2015).

349 The Columbrets basin has been interpreted as a salt-detached syn-rift ramp-syncline basin
350 (Roma et al., 2018). The eastern side of the basin features an extensional detachment fault,
351 rooting at depth in the continental crust beneath the hyperextended rift domain (Figs. 5D-E;
352 Etheve et al., 2018) that coincides with the Triassic salt décollement. This detachment is
353 responsible for the overall thin-skinned extensional deformation of the pre-rift sedimentary
354 cover experiencing basinward halokinetic gliding (Etheve et al., 2018). Nevertheless, finite
355 motion along the detachment imaged in seismic reflection profiles is not enough to account
356 for the extreme crustal thinning identified in the basin core and the large discrepancy in
357 thinning ratios between the lower and upper crust (Etheve et al., 2018). In fact, the reflective
358 lower crust becomes thinner toward the axis of the Columbrets basin, where it appears to be
359 either absent or no thicker than 1–2 km (Gallart et al., 1990, 1994; Dañobeitia et al., 1992;
360 Torné et al., 1992; Sàbat et al., 1997; Vidal et al., 1997). These observations highlight the
361 differences in the rheological response of the upper and lower crust to crustal stretching as
362 deduced from the Parentis basin architecture. Etheve et al. (2018) suggested that the lower
363 crust underwent large-scale ductile deformation/boudinage during the Late Jurassic to Early
364 Cretaceous syn-rift stage followed by simple shear along a single crustal detachment at the
365 end of the syn-rift stage. In summary, the evolution of the Columbrets rift system was
366 controlled by shallow decoupling in the Triassic pre-rift evaporites and deep decoupling in the
367 middle crust (Etheve et al., 2018). Thus, its syn-rift evolution shares similarities with the

368 behaviour of the Arzacq, Cameros and Parentis basins, which first experienced homogeneous
369 ductile crustal thinning of the lower crustal levels passing to the activation of a shallower
370 crustal detachment.

371 The present-day surface heat flow in the southwestern Valencia Trough is about 65 to 100
372 mW/m² (e.g., Ayala et al., 2015; Carballo et al., 2015). That area also displays the thinnest
373 continental crust section (Banda & Santanach, 1992; Fernandez et al., 1995; Ayala et al.,
374 2015), demonstrating that a stable high thermal regime was inherited from the Late Jurassic to
375 Cretaceous rifting despite the fact that rifting ended in the Cenozoic.

376 **2.5 North Pyrenean Zone basins**

377 In the Pyrenees, late Hercynian (Permian) post-orogenic extension led to the development of
378 continental deposits in endorheic extensional basins (Bixel & Lucas, 1983, 1987; Bixel,
379 1984). At the same time, magmatic and granulitic rocks were exhumed all along the North
380 Pyrenean Zone (de Saint Blanquat, 1993; Olivier et al., 2004; Cochelin, 2016; Cochelin et al.,
381 2017; Saspiturry et al., 2019b). A consequence of the Hercynian collapse stage was major
382 thinning of previously hot lithosphere, an important structural inheritance regarding the
383 following Mesozoic stages. Recent studies have proposed that the continental crust was partly
384 thinned before the onset of Early Cretaceous hyperextension. Restorations of Iberian-
385 European crustal sections across the Mauléon basin (western Pyrenees) and the Ballongue
386 basin (central Pyrenees) show that the Moho was already very shallow (~20 km depth) by the
387 end of the Jurassic (Asti et al., 2019; Espurt et al., 2019; Saspiturry et al., 2020a). Thus, the
388 latest Paleozoic thinning stage, as well as the Late Triassic regional extensional stage, should
389 not be neglected when estimating the Cretaceous thinning in the North Pyrenean Zone.

390 The Triassic deposits of the Western Pyrenees are typical of the German-type succession,
391 ending with a thick evaporite and ophite complex (Curnelle, 1983; Lucas, 1985; Rossi et al.,

392 2003). The salt unit has played a major role in the Pyrenees, acting as a décollement layer at
393 the base of the Mesozoic sedimentary cover that controlled the deformational style during
394 Early Cretaceous hyperextension (Canérot, 1988, 1989; Canérot & Lenoble, 1993; James &
395 Canérot, 1999; Canérot et al., 2005; Jammes et al., 2010b; Lagabrielle et al., 2010, 2020;
396 Duretz et al., 2019). Unlike the previously described basins, the North Pyrenean Zone
397 typically lacks Berriasian and Valanginian deposits (Combes et al., 1998; James, 1998;
398 Canérot, 2008), a witness of the emersion of the area during the earliest Cretaceous. From
399 Barremian to Aptian times, the Iberian and European margins of the North Pyrenean Zone
400 basins were carbonate platforms grading to distal marls toward the basin axis. In summary,
401 the North Pyrenean Zone hyperextended basins are characterised by sedimentary sequences
402 consisting of (1) Late Triassic to Early Jurassic pre-rift evaporites 1–3 km thick (Curnelle,
403 1983), (2) Jurassic pre-rift carbonate platform rocks 0.5–2 km thick deposited in a relatively
404 stable tectonic context (Delfaud & Henry, 1967; Lenoble, 1992; James, 1998), (3) Barremian
405 to Aptian syn-rift carbonates and marls up to 1.6 km thick (Delfaud & Villanova, 1967,
406 Arnaud-Vanneau et al., 1979), (4) syn-rift Albian flysch-like deposits (the Black Flysch or
407 “Flysch Ardoisier”) 1–5 km thick (Debroas, 1978, 1987, 1990; Boirie, 1981; Boirie &
408 Souquet, 1982; Fixari, 1984; Roux, 1983; Souquet et al., 1985; Debroas et al., 2010) and (5)
409 post-rift turbidites 2–4 km thick (Casteras, 1971; Henry et al., 1987; Le Pochat et al., 1976;
410 Razin, 1989).

411 In this section we discuss the reconstructed architecture of the Mauléon and Basque-
412 Cantabrian basins, together with that of basins present within the Internal Metamorphic Zone.
413 We show that similarly to the previously described basins, the North Pyrenean Zone basins is
414 affected by a syn-rift stage characterised by pure shear ductile thinning of the lower middle
415 crust. The North Pyrenean Zone basins are characterised by the presence of positive Bouguer
416 gravity anomalies above their inverted hyperextended domain, as evidenced by the Basque-

417 Cantabrian, Mauléon and Saint-Gaudens anomalies (Fig. 6). These anomalies reflect the
418 presence of subcontinental mantle at shallow depth that was exhumed during the climax of the
419 Mesozoic extension. With the exception of the Columbrets basin, the North Pyrenean Zone
420 rift basins record continental breakup and mantle exhumation from the latest Albian to Early
421 Cenomanian (breakup sequence, sensu Soares et al., 2012).

422 **2.5.1 Mauléon basin**

423 The Mauléon basin, located south of the Arzacq basin (Figs. 1 & 7), coincides with a strong
424 positive gravity anomaly centred upon the basin axis (Figs. 6 & 7A; Gottis, 1972; Boillot et
425 al., 1973). First interpreted as lower crustal material (Grandjean, 1994; Vacher & Souriau,
426 2001; Pedreira et al., 2007), the anomaly is currently attributed to the presence at shallow
427 depth (~10 km) of a dome of subcontinental mantle (Fig. 7B; Casas et al., 1997; Jammes et
428 al., 2010a). This interpretation has gained support from recent work documenting P-wave
429 velocities of ~7.3 km/s in this deep material (Fig. 7B; Wang et al., 2016; Chevrot et al., 2018).
430 Above the mantle dome, the thickness of the crust is assumed to be roughly 5 km (Fig. 7B).
431 Mantle exhumation apparently occurred during Cretaceous hyperextension, when the
432 Mauléon basin developed as a hyperextended rift (Jammes et al., 2009; Lagabrielle et al.,
433 2010; Masini et al., 2014; Tugend et al., 2014; Corre et al., 2016; Teixell et al., 2016;
434 Lagabrielle et al., 2019a; Labaume & Teixell, 2020). The basin was inverted during Eocene
435 shortening and is a pop-up structure at present, bordered to the north and south by conjugate
436 thrusts (Fig. 7B; Saspiturry et al., 2020a).

437 The Mauléon basin began as a symmetric syn-rift basin that subsided in response to pure
438 shear ductile thinning of the lower/middle crust during the Early Cretaceous (Saspiturry et al.,
439 2019a) (Fig. 7E). Its structural style changed from Albian to early Cenomanian time, leading
440 to asymmetric basin morphology and sedimentary facies distribution. Gravity-flow
441 conglomerates accumulated at the foot of the Iberian margin slope, forming the Mendibelza

442 fan conglomerates (Boirie, 1981; Fixari, 1984; Souquet et al., 1985), in response to activity on
443 steep north-dipping normal faults (South Arbailles and North Arbailles faults; Saspiturry et
444 al., 2019a). These rocks originated in fan deltas reworking freshly uplifted Paleozoic
445 substratum. Restoration of syn-rift geometries indicates that the Iberian substratum was tilted
446 30° toward the north in Albian time (Saspiturry et al., 2019a). This implies a thickening of
447 syn-rift deposits to the north toward the steep south-dipping Saint-Palais fault, where the
448 conglomerates reach a maximum thickness of around 5 km (Fig. 7). This fault separated the
449 marls of the central basin from the European proximal margin to the north, where a carbonate
450 platform developed (Saspiturry et al., 2019a). The gentle southward slope of the European
451 margin contrasts with the steep northward slope of the Iberian margin. This geometry points
452 to the Saint-Palais fault as a major normal fault that was responsible for the change to
453 asymmetric basin margins during Albian to early Cenomanian time. In this scheme, the steep
454 north-dipping slope of the Iberian margin can be interpreted as a rollover structure in the
455 hanging-wall of the Saint-Palais fault. The rollover structure is also accommodated by minor
456 north-dipping normal faults that propagated toward the south.

457 Facies distribution significantly changed from the mid-Cenomanian to the late Santonian in
458 the Mauléon basin as shallow carbonate platforms developed on the Iberian and European
459 margins (Souquet, 1967; Alhamawi, 1992; Ternet et al., 2004; Serrano et al., 2006). On the
460 European side, the transition from platform to basin coincided with the steep south-dipping
461 South Grand Rieu fault (Fig. 7). The Iberian carbonate platform graded northward rather
462 abruptly to deep-sea calcareous breccias at the site of the Lakhoura normal fault, which
463 appears to have acted as a significant northward detachment during mid-Cenomanian times.
464 This detachment, responsible for a southward tilt of the Iberian basement, crosscuts the lower
465 part of the older Saint-Palais structure, which was inactive at that time. Thus, the Mauléon
466 basin was affected by ductile pure shear thinning of the lower/middle crust from the

467 Barremian to the Aptian, followed by Albian simple shear concentrated along two major
468 crustal detachments: (1) the south-verging Saint-Palais fault accommodating the thinning of
469 the European margin and (2) the north-verging Lakhoura detachment crosscutting the Saint-
470 Palais fault (Fig. 7C3). Along strike, the Mauléon basin was affected by continental breakup
471 from the latest Albian to the Early Cenomanian (Fig. 7D) as evidenced by the presence of
472 subcontinental mantle clasts into the latest Albian to Early Cenomanian Urdach breccias
473 (Roux, 1983; Duée et al., 1984; Fortané et al., 1986; Debroas et al., 2010) as well as by the
474 formation of ophicalcites at the surface of the denudated mantle (Jammes et al., 2009;
475 Lagabrielle et al., 2010, 2019a; Debroas et al., 2010).

476 Rifting and continental breakup in the Mauléon basin developed, from the Early Cretaceous to
477 the mid-Cenomanian (Fig. 7E), under a high geothermal gradient, as indicated by (1) Raman
478 spectroscopy of carbon materials (RSCM) showing peak temperatures consistent with a
479 gradient of 60–75°C/km (Corre, 2017; Saspiturry, 2019; Saspiturry et al., 2020b), (2) detrital
480 zircon fission-track data indicating a 80°C/km gradient (Vacherat et al., 2014) and (3) (U-Th-
481 Sm)/He thermochronology data indicating a 80–100°C/km gradient (Hart et al., 2017).
482 Numerical thermal models suggest that the base of the hyperextended domain had a mantle
483 heat flow of 100 mW/m² and a maximum temperature of 600°C during continental breakup
484 (Saspiturry, 2019; Saspiturry et al., 2020b). Vitrinite reflectance values from the Mauléon
485 basin also weigh in favour of HT/LP syn-rift metamorphism (Lescoutre, 2019; Lescoutre et
486 al., 2019). It should be noted that the peak temperature was reached in the lower part of the
487 basin during the post-rift period (Vacherat et al., 2014; Saspiturry, 2019), similar to the
488 Cameros basin. Due to Pyrenean thrusting, a detached slice of mantle rock crops out in the
489 eastern part of the basin (Urdach area). In that location, crustal material in contact with the
490 Urdach lherzolites shows ductile shearing, suggesting that the upper/middle crust was
491 extruded laterally from the basin axis at temperatures between 350°C and 450°C (Asti et al.,

492 2019). Hydrothermal fluid circulation sheds light on the ductile shearing of the Mauléon basin
493 pre-rift cover during continental breakup at ~94 Ma (Incerpi et al., 2020). The thermal regime
494 of the Mauléon basin from the end of the syn-rift stage (Albian, Fig. 7E) to the breakup
495 sequence (latest Albian to Early Cenomanian; Fig. 7E) attests a ductilely stretched
496 sedimentary cover and crystalline basement.

497 **2.5.2 Basque-Cantabrian basin**

498 The Basque-Cantabrian basin records a similar Mesozoic history as the North Pyrenean Zone
499 rift basins, although it developed to the eastern termination and south of the Pyrenean Axial
500 Zone (AZ in Fig. 1). The rift axis is characterised by a succession of Upper Jurassic and
501 Cretaceous sediments around 8–10 km thick with interlayered basic volcanic rocks of Aptian
502 to Santonian age (Figs. 8A-B; Azambre & Rossy, 1976; Rat et al., 1983; Rat, 1988; García
503 Mondéjar et al., 1996; Castañares et al., 1997; Castañares & Robles, 2004). The lithosphere is
504 extremely thin in the western central part of the basin, as suggested by a positive Bouguer
505 anomaly recorded in the Biscay Synclinorium (Fig. 6; Pedreira et al., 2007) that is interpreted
506 as the consequence of lithospheric mantle exhumed at shallow depth (Figs. 8A-B; Pedrera et
507 al., 2017, 2020; Garcia-Senz et al., 2019). Field observations and seismic interpretations
508 indicate that the Basque-Cantabrian basin has an overall symmetric shape characterised by
509 brittle deformation on its flanks and ductile deformation on its axis (DeFelipe et al., 2017;
510 Pedrera et al., 2017; Ducoux et al., 2019).

511 The basin continental crust was stretched during the Early Cretaceous rifting stage. Locally,
512 mantle denudation occurred during early Cenomanian continental breakup as revealed by the
513 presence of a strongly weathered mantle fragment near the inverted Leiza major detachment
514 fault (Mendia & Ibarra, 1991; DeFelipe et al., 2017; Ducoux, 2017; Ducoux et al., 2019).
515 Hyperextension was also recorded by the development of Cretaceous HT/LP metamorphism
516 (Golberg & Leyreloup, 1990; Cuevas & Tubia, 1999). Mineral assemblages and RSCM data

517 from the Nappes des Marbres, which represents the inverted eastern part of the Basque-
518 Cantabrian basin, indicate that temperature in the pre-rift sedimentary cover locally reached
519 500–600°C during hyperextension (Figs. 8C-D; Lamare, 1936; Martínez-Torres, 1989;
520 Mendia & Ibarra, 1991; Ducoux et al., 2019). Thus, the felsic crust and its sedimentary
521 cover underwent ductile stretching during the Albian syn-rift stage and Cenomanian
522 continental breakup. Interpretations of geophysical data have shown that decoupling between
523 basement and cover rocks occurred in low-strength Triassic evaporites and mudstones and
524 induced coeval formation of cover gliding, mini-basins, turtle salt anticlines, expulsion
525 rollovers, and salt diapirs in the cover strata (Fig. 8; Pedrera et al., 2017, Ducoux et al., 2019;
526 Cámara, 2020). The association of exhumed mantle along the Leiza fault with rift and post-
527 rift structural geometries suggests that a major south-dipping ramp-flat-ramp extensional
528 detachment was active from Albian to early Cenomanian time (Lagabriele et al., 2020).

529 **2.5.3 Internal Metamorphic Zone basins**

530 The central and eastern portions of the North Pyrenean Zone include a narrow Internal
531 Metamorphic Zone along their southern borders (Fig. 1). They consist of a east-west-trending
532 stretched zone of Variscan basement and pre-rift to syn-rift metamorphic rocks (Casteras,
533 1933; Mattauer, 1968; Choukroune, 1974) with several outcrops of subcontinental mantle
534 rocks (Monchoux, 1970; Choukroune & Mattauer, 1978; Fabriès et al., 1991, 1998;
535 Lagabriele et al., 2010). Recent studies have shown that the Internal Metamorphic Zone is an
536 inverted domain of continental crust that was hyperextended during Early Cretaceous rifting
537 (Lagabriele & Bodinier, 2008; Lagabriele et al., 2010; Clerc & Lagabriele, 2014; Clerc et
538 al., 2014, 2015; de Saint Blanquat et al., 2016; Teixell et al., 2018; Dielforder et al., 2019;
539 Espurt et al., 2019; Garcia-Senz et al., 2019). In the central Pyrenees, the Internal
540 Metamorphic Zone coincides with the Saint-Gaudens positive Bouguer anomaly (Fig. 9A)
541 and corresponds to the inverted Early Cenomanian continental breakup domain (Figs. 9B-C).

542 It has been shown that the whole continental crust of the North Pyrenean Zone was affected
543 by large-scale ductile boudinage during Early Cretaceous hyperextension, with E-W trending
544 rift basins exhibiting a relatively symmetrical profile (Clerc et al., 2015; Lagabrielle et al.,
545 2020). The presence of a thick pre-rift salt layer led to basinward gliding of the overlying
546 Jurassic cover during hyperextension (Lagabrielle et al., 2010; Clerc & Lagabrielle, 2014;
547 Clerc et al., 2015; Duretz et al., 2019; Lagabrielle et al., 2020).

548 Evidence of HT/LP metamorphism has been reported along the entire North Pyrenean Zone
549 (Ravier, 1957; Albarède & Michard-Vitrac, 1978a, 1978b; Golberg et al., 1988; Golberg &
550 Maluski, 1988; Golberg & Leyreloup, 1990; Clerc et al., 2015). This metamorphism resulted
551 from Early Cretaceous continental-crust thinning and an associated increase in thermal
552 gradient and burial (Choukroune & Mattauer, 1978; Vielzeuf & Kornprobst, 1984; Debroas,
553 1990; Clerc et al., 2015). RSCM peak temperatures reached 400–600°C in the marbles of the
554 Internal Metamorphic Zone, where some of the highest temperatures were recorded close to
555 exhumed mantle outcrops (Fig. 9D; Clerc, 2012; Clerc et al., 2015; Boulvais, 2016; Chelalou
556 et al., 2016; Lagabrielle et al., 2016; Ducoux, 2017). During the Albian syn-rift stage and the
557 Early Cenomanian breakup event, the crust was homogeneously and ductilely stretched in the
558 hyperextended domain, while detachment faults at the transition between the mantle and the
559 crust/sedimentary cover accommodated the thinning of the whole system (Lagabrielle et al.,
560 2019a). Likewise, mineral assemblages indicate that maximum temperatures of 550–650°C
561 and pressures of 3–4 kbar were reached locally (Bernus-Maury, 1984; Golberg & Leyreloup,
562 1990; Vauchez et al., 2013). Previous authors have established that this metamorphic event
563 was linked to high syn-rift geothermal gradients (Dauteuil & Ricou, 1989; Golberg &
564 Leyreloup, 1990; Clerc et al., 2015; Lagabrielle et al., 2016). Finally, these data collectively
565 indicate that the Jurassic to Early Cretaceous metacarbonate cover forming the current
566 Internal Metamorphic Zone corresponds to pre-rift to basal syn-rift sediments located in the

567 deepest part of the former North Pyrenean Zone basin, which was also characterised by a thin
568 continental crust.

569 The WNW-ESE trending Lourdes and Saint-Gaudens positive Bouguer anomalies coincide
570 with the maximum thickness of the Albian syn-rift turbidites (Figs. 9A-B; Casas et al., 1997).
571 Espurt et al. (2019) interpreted the Saint-Gaudens anomaly as a body of allochthonous mantle
572 pushed northward onto the European margin on the North Pyrenean Frontal Thrust. It
573 corresponds to an allochthonous body of subcontinental mantle that was previously exhumed
574 during Early Cretaceous time (Fig. 9B).

575 **3 Discussion: Tectono-sedimentary evolution of smooth-slope extensional** 576 **basins**

577 **3.1 Syn-rift**

578 **3.1.1. Pure-shear dominated thinning**

579 All the basins reviewed in this work are characterised by a strong heterogeneous structural
580 pattern inherited from the Late Carboniferous to Triassic rifting events related to the collapse
581 of the Variscan belt and the breakup of Pangea. These events ended with the deposition of
582 ~1–3 km thick Late Triassic to Early Jurassic salt deposits. Thus, Mesozoic hyperextension
583 initiated within a continental crust that was previously thinned to less than 30 km thick (Fig.
584 10A). Unfortunately, the precise thickness of the crust at the end of the Triassic remains
585 undetermined. The Late Jurassic to Early Cretaceous syn-rift stage was driven by distributed
586 deformation characterised by the lateral extraction of the lower/middle crust (Fig. 10B).
587 During this stage, thinning of the lower/middle crust triggered progressive subsidence. Our
588 review shows that homogeneous subsidence was partly balanced by the production of syn-rift
589 carbonates in most of the studied basins. The result is a relatively smooth basin floor profile
590 characterised by carbonate platform deposits with marls in a central trough. During this early

591 stage, the basin was symmetric and marked by edgeward onlap of the syn-rift deposits (Fig.
592 10B). The pre-rift cover was efficiently decoupled from its substratum thanks to the thick
593 layer of Keuper evaporites at its base. This rheological layering and the progressive sinking of
594 the central part of the rift system eventually led to the breakup of the pre-rift lid on the
595 external margins of the basin. This resulted in a large-scale pre-rift cover which remained in
596 the central part of the extensional system throughout the whole basin lifetime, while the
597 continental crust was progressively thinned below it (Fig. 10B and C). In contrast to all other
598 reviewed basins, the Tartas basin exemplifies this syn-rift stage since continental crust
599 thinning was not followed by the activation of extensional detachments, allowing the basin
600 floor profile to remain symmetric. During this first rift regime, the upper crust might have
601 been affected by very minor brittle deformation beneath the salt décollement level that led to
602 superficial salt diapirism affecting the pre-rift to syn-rift sedimentary cover (Fig. 10B), as
603 evidenced in the Arzacq (Issautier et al., 2020), Parentis (Ferrer et al., 2012), Columbrets
604 (Etheve et al., 2018), and Mauléon (Canérot et al., 2005) basins during Late Jurassic to Aptian
605 time. However, during this syn-rift stage, continental crustal thinning was mainly
606 accommodated by distributed ductile thinning within the lower/middle continental crust. This
607 process was first suggested in the case of the Parentis basin (Pinet et al., 1987a). It was then
608 put forward by Clerc and Lagabrielle (2014) in their interpretation of the architecture of the
609 eastern North Pyrenean Zone basins and used by Corre et al. (2016) in the reconstruction of
610 the eastern Mauléon basin border. It has more recently been applied to the central Mauléon
611 basin (Saspiturry et al., 2019a; Asti et al., 2019), Arzacq basin (Issautier et al., 2020) and
612 Columbrets basin (Etheve et al., 2018). It consists of a distributed thinning stage in which the
613 lower/middle crust was homogeneously and symmetrically thinned without major brittle
614 deformation of the upper crust. Interpretation of seismic profiles sheds light on the wide
615 discrepancy in thinning ratios between the lower and upper crust in the Columbrets (Etheve et

616 al., 2018) and Parentis (Pinet et al., 1987a) basins as they were slightly inverted. Such a
617 discrepancy in the amount of continental thinning has been numerically modelled and defined
618 as depth-dependent continental crust thinning (Huismans & Beaumont, 2008, 2011, 2014).
619 This pure shear dominated thinning stage is applicable to all the North Pyrenean Zone basins
620 since (1) their sedimentary profile was symmetric throughout the Barremian to Aptian
621 beginning of the syn-rift stage and (2) they do not record exhumation of the lower crust
622 during Mesozoic hyperextension in their central part (see section 2.5 for more details).

623 **3.1.2. Simple shear dominated thinning**

624 As thinning of the lower/middle crust continued, isostatic subsidence occurred in the centre of
625 the basin. The early smooth synclinal-shape basin progressively deepened, triggering
626 steepening of pre-to- syn-rift sequences. Deformation became localised and a simple shear
627 regime initiated along a crustal detachment connecting upward to the Late Triassic pre-rift salt
628 décollement. With the progression of extension, the pre-rift sequence was progressively
629 dissected in several large-scale boudins and turtle structures separated by intervening salt
630 diapirs (Fig. 10C and D). The top basement surface is steeper in the basin flanks than in the
631 basin core, therefore portions of the pre-rift cover may undergone gravity-driven gliding
632 leading to local syn-extensional thrusting (layer-parallel shortening). From a general point of
633 view, gliding was controlled by thickness variations of the sedimentary pile that cause
634 differential loading on the salt layer (Lundin, 1992; Liro & Coen, 1995; Rouby et al., 2002)
635 and basinward tilting of the proximal margin (Cobbold & Szatmari, 1991; Demercian et al.,
636 1993; Gaullier et al., 1993; Fort et al., 2004a, 2004b). The gravity gliding unroofed the
637 basement of the proximal rift margin. As the pre-rift and syn-rift sequences glided along the
638 décollement, salt was expelled both marginward and basinward as well as upward by
639 buoyancy. Thus, in the reviewed basins, halokinesis led to (1) the development of salt-
640 detachment synclines, salt rollovers or diapirs affecting and controlling syn-rift depocentres

641 and (2) the denudation of the proximal margin basement that was subsequently overlain by
642 syn-rift sediments. This scenario is reported from most of the studied basins.

643 (1) In the Arzacq basin, the southern margin recorded northward cover gliding (Fig. 2;
644 Issautier et al., 2020).

645 (2) In the Parentis basin, major diapirism (Pelican borehole, Fig. 4D; Ferrer et al., 2008, 2009,
646 2012) and denudation of its southern margin basement (Fig. 4D; Jammes, 2019) were
647 documented.

648 (3) In the Columbrets basin, the SE margin basement was denudated (Etheve et al., 2018).

649 (4) In the Mauléon basin, the proximal margin basement was locally denudated on both edges
650 of the rift (Fig. 7C3 & 7D; Teixell et al., 2016; Saspiturry et al., 2020a).

651 (5) In the Basque-Cantabrian basin, the prerift cover was removed from its proximal margins
652 (Fig 8A; Pedrera et al., 2017).

653 (6) In the Internal Metamorphic Zone, the pre-rift allochthonous cover remained in its central
654 part (Fig. 9C-D; Espurt et al., 2019).

655 Denudation of the basement proximal margin has also appeared in thermo-mechanical
656 numerical models of lithospheric-scale extension that integrate recent data collected along the
657 North Pyrenean Zone basins (Duretz et al., 2019). This process resulted in the formation of
658 syn-gliding wedge-shaped sedimentary geometries and syn-rift sequence depocenter
659 migration, as seen in (a) the Arzacq and Parentis basins, which display southward syn-rift
660 depocenter migration (Fig. 2 and 4D; Jammes, 2009; Issautier et al., 2020), and (b) the
661 Cameros, Columbrets and Mauléon basins which display northward, north-westward and
662 northward syn-rift depocenter migration, respectively (Figs. 3F, 4D and 7C2-C3; Omodeo-
663 Salé et al., 2014; Etheve et al., 2018; Saspiturry et al., 2020a). In the Mauléon basin, the

664 increasing northward slope-deepening of the Iberian margin is interpreted as a rollover effect
665 linked to the Saint-Palais detachment. Increasing tilting of the Iberian margin led to cover
666 gliding, immediately followed by the accumulation of deep basin gravity deposits. These
667 latter consist of reworked sediment from rafts of the pre-rift sedimentary cover and the freshly
668 exposed proximal margin basement (Teixell et al., 2016; Saspiturry et al., 2019a; Labaume &
669 Teixell, 2020). A similar evolution cannot be clearly reconstructed for the Basque-Cantabrian
670 and Internal Metamorphic Zone basins, which experienced severe inversion during the
671 compressional stages of the Pyrenean orogeny (e.g. Pedrera et al., 2017, Fig. 8A; Espurt et al.,
672 2019, Fig. 9C). In contrast to the Tartas basin, which did not reach the second syn-rift
673 extensional stage, all the other basins became asymmetric, as evidenced by shifts of the basin
674 depocenter.

675 Seismic profiles display clear images of crustal detachments in the Parentis (Fig. 4D; Jammes,
676 2009; Jammes et al., 2010c; Tugend et al., 2014) and Columbrets (Figs. 5D & 5E; Etheve et
677 al., 2018) basins. Both basins exhibit an asymmetric outline associated with a very thin crust
678 in their axial regions. This sheds light on the fact that important crustal thinning was partly
679 accommodated by simple shear deformation along the imaged detachments (Fig. 2B; Issautier
680 et al., 2020). In the Arzacq and Cameros basins, detachment faults have not been imaged but
681 only inferred, although both basins evolved with cover gliding and depocenter migration
682 during the syn-rift stage. For instance, the northern thrust edging the Cameros basin has been
683 interpreted as a reactivated syn-rift southward deepening ramp-flat structure corresponding to
684 an extensional detachment (Figs. 5E-F; Omodeo-Salé et al., 2014). Although the Arzacq,
685 Cameros, Parentis and Columbrets basins did not record continental breakup like the North
686 Pyrenean basins, crustal thinning was fairly advanced, developing under warm thermal
687 conditions during the simple shear stage. Indeed, mature mantle exhumation is evidenced by a
688 positive Bouguer anomaly and a current shallow Moho depth in the Parentis (Figs. 4A & 4D;

689 Jammes, 2009) and Columbrets basins (Figs. 5A-D; Ayala et al., 2015; Etheve et al., 2018).
690 Due to progressive burial and continental crust thinning, the sedimentary cover in the centre
691 of some smooth-slope basins experienced warm thermal regimes as typically shown by the
692 HT/LP metamorphism of the North Pyrenean Basins (more than 400°C). This is also
693 documented in the Cameros basin by mineralogical analysis, thermochronology and fluid
694 inclusion studies (Fig. 3F; Rat et al., 2019), as well as in the Columbrets basin by an elevated
695 mantle heat flow of around 100 mW/m² (Ayala et al., 2015; Carballo et al., 2015).

696 **3.2 Breakup stage**

697 The North Pyrenean Basins first underwent significant pure shear ductile thinning of the
698 lower/middle crust throughout the Barremian to Aptian, followed by simple shear localization
699 on detachment faults during the Albian (Saspiturry et al., 2019a). In concept, continental
700 breakup occurs once the sedimentary cover is removed from the proximal margins and the
701 lower crust fully withdrawn from the basin centre (Fig. 10D). This results in the development
702 of brittle deformation on the basin flanks and the formation of tilted basement blocks devoid
703 of sedimentary cover, while the central hyperextended domain records dominantly ductile
704 thinning and mantle exhumation (Fig. 10D). As previously shown by Soares et al. (2012),
705 while the upper continental crust deforms in a brittle manner in the hyperextended domain
706 during dominantly pure-shear thinning, the deformation regime switches during dominantly
707 simple-shear thinning and subsequent mantle exhumation. As the lower and middle crust is
708 removed from the basin centre during the syn-rift sequence, the isotherms rise under the most
709 highly extended parts of the rift. The 300°C to 500°C isotherms can be traced above the top of
710 the Variscan basement and overlying pre-rift and syn-rift sediments. This implies that large
711 volumes of the basement and cover were in a ductile regime during this stage (Fig. 10D).
712 Thus, during the breakup sequence, thinning of the crust is first controlled by the elevation of
713 the thermal conditions in relation with mantle exhumation. An additional parameter triggering

714 temperature elevation is the progressive burial of the continental crust under a thick
715 sedimentary cover that accumulates in the basin during the syn-rift stage. Both causes
716 significantly increase the temperature at the base of the basin. This process has also been
717 demonstrated by numerical modelling of the North Pyrenean Zone basins (Duretz et al.,
718 2019). In the hyperextended domain, extension is localised along a ductile shear zone at the
719 transition between mantle and continental crust/sedimentary cover.

720 Our review highlights important characteristics of the thinning processes in the most evolved
721 basins. We confirm that the conditions of HT/LP metamorphism evident along the
722 hyperextended domains of the North Pyrenean basins account for the ductile behaviour of the
723 crust. The climax of this thermal metamorphism occurred during continental breakup
724 (Albarède & Michard-Vitrac, 1978a; Montigny et al., 1986; Golberg et al., 1986; Golberg &
725 Maluski, 1988; Thiébaud et al., 1992). The peak temperature reached in the Jurassic to Albian
726 sedimentary cover varies between 500° and 600°C, as documented in various places: (1) the
727 Mauléon basin (Figs. 7C4-D; Corre, 2017; Saspiturry, 2019), (2) the Nappes des Marbres in
728 the Basque-Cantabrian basin (Fig. 8D; Lamare, 1936; Martínez-Torrez, 1989; Mendia &
729 Ibarra, 1991; Ducoux, 2017; Ducoux et al., 2019) and (3) the Internal Metamorphic Zone
730 basins in the central and eastern Pyrenees (Fig. 9D; Bernus-Maury, 1984; Golberg &
731 Leyreloup, 1990; Azambre et al., 1992; Clerc, 2012; Vauchez et al., 2013; Clerc et al., 2015;
732 Chelalou et al., 2016).

733 The Internal Metamorphic Zone and the Nappes des Marbres have been interpreted as the
734 inverted base of the North Pyrenean hyperextended rift domain (Clerc, 2012; Clerc &
735 Lagabrielle, 2014; Clerc et al., 2015; Lagabrielle et al., 2016; Ducoux, 2017). In these
736 settings, continental crust thinning resulted in a high geothermal gradient estimated at around
737 60–100°C/km in the Mauléon basin (Vacherat et al., 2014; Corre, 2017; Hart et al., 2017;
738 Saspiturry, 2019). The maximum temperature reached in the sedimentary cover increased

739 from the margins to the hyperextended domain. Thus, the thermal gradient increased together
740 with the thickness of the sedimentary pile (Saspiturry, 2019). Numerical thermal modelling
741 has shown that this thermal gradient was associated, in the Mauléon basin, with an elevated
742 mantle heat flow of around 100 mW/m^2 (Saspiturry, 2019). The basinward gliding of the pre-
743 rift cover contributed significantly to the burial and thus to the peak temperature increase at
744 the base of the hyperextended domain.

745 The coupled effects of heating by mantle exhumation and burial under the thick pre-rift (salt
746 tectonic) to syn-rift sedimentary sequence prevented crustal normal faults from propagating
747 into the hyperextended domain, as evidenced in the Nappes des Marbres (Fig. 8D; Ducoux et
748 al., 2019) and the Internal Metamorphic Zone (Fig. 9D; Espurt et al., 2019). Thinning of both
749 the Variscan basement and the allochthonous sedimentary pile occurred in the hyperextended
750 domain within a thick zone of dominantly ductile shear (Lagabrielle et al., 2019a). This
751 process led to the formation of large-scale boudins and lenses of continental crust and
752 strongly sheared metasedimentary rocks (Clerc & Lagabrielle, 2014; Clerc et al., 2015b;
753 Corre et al., 2016; Asti et al., 2019; Duretz et al., 2019; Lagabrielle et al., 2019a). These
754 mature basins then took the shape of typical pseudo-symmetric hyperextended rift basins;
755 their margins were affected by brittle normal faulting and their centres were dominated by
756 ductile stretching as observed along the most evolved smooth-slope basins of the North
757 Pyrenean Zone. As shown by the presence of clasts of mantle rocks in the Cretaceous
758 turbidite breakup sequence of the Urdach area, these basins underwent local denudation of
759 subcontinental mantle thanks to motion along a major detachment during the Late Albian to
760 Early Cenomanian continental breakup (Roux, 1983; Duée et al., 1984; Fortané et al., 1986;
761 Jammes et al., 2009; Debroas et al., 2010; Lagabrielle et al., 2010, 2019a, b). Although the
762 Mauléon basin displays clasts of mantle rocks reworked in its breakup sequence sensu Soares
763 et al. (2012), it underwent heterogeneous amounts of mantle denudation and seems to show a

764 less advanced stage of continental breakup than the Basque-Cantabrian and Internal
765 Metamorphic Zone basins. According to Saspiturry et al. (2019a, 2020a), the amount of
766 mantle denudation under the Mauléon basin varied along strike as a consequence of Permian
767 inheritance along N20° transverse structures (Fig. 7C3; Saspiturry et al., 2019b). It was
768 maximal in the eastern part (Urdach) but almost nonexistent in its western part (Fig. 7D;
769 Teixell et al., 2016; Labaume & Teixell, 2020). In addition, the Internal Metamorphic Zone
770 basin displayed a more advanced breakup sequence than the Basque-Cantabrian basin as
771 evidenced by the widths of exhumed mantle along two restored crustal-balanced cross-
772 sections, which are ~45 km (Fig. 9C; Espurt et al., 2019) and ~15 km (Fig. 8A; Pedrera et al.,
773 2017) long, respectively. The Bilbao, Mauléon and Saint-Gaudens positive Bouguer
774 anomalies (Fig. 6) represent the remains of continental breakup stages now buried at depth, as
775 they correspond to major pieces of mantle exhumed during the latest Albian to Early
776 Cenomanian breakup sequence and more recently inverted during the Pyrenean orogeny
777 (Figs. 7A-B, 8A-B & 9A-B).

778 Based on the observations listed in this section, we may conclude that the reviewed smooth-
779 slope extensional basins represent different degrees of hyperextension that occurred along the
780 Iberia-Eurasia plate boundary during the Cretaceous drift of Iberia. We thus propose to rank
781 these basins from least mature to most mature as follows: (1) Tartas, (2) Arzacq/Cameros, (3)
782 Parentis/Columbrets, (4) Mauléon, (5) Basque-Cantabrian and (6) Internal Metamorphic Zone
783 (Fig. 11). This ranking indicates that as the amount of extension increases in these basins, the
784 intensity of various fundamental processes also increases. These processes are lower/middle
785 crust lateral extraction, thermal gradient and heat flow, HT/LP metamorphism, ductile
786 thinning of the crust and its sedimentary cover, relative gliding of the pre-rift cover and
787 mantle exhumation.

788 A major consequence of all these processes is the formation of a succession of basins with
789 smooth basement slopes, which differ significantly from rift basins controlled by dominantly
790 normal faulting that affects both their borders and centres. These smooth-slope basins were
791 the locus of gentle, homogeneous subsidence that led to the deposition of syn-rift flysch-like
792 sediments a few kilometers thick. Therefore the response of the smooth-slope basins to
793 extensional stress was the accumulation of sediments that in turn increased the thermal burial
794 effect in the basin centres. Such burial appears to have been a key factor in the syn-rift
795 evolution of smooth-slope basins, along with parameters that were critical earlier at the
796 initiation of rifting, such as the Variscan-Triassic inheritance of thin crust (Asti et al., 2019)
797 and the presence of a very thick (average 2 km) layer of evaporites and shale of the Keuper
798 group that allowed the decoupling of the pre-rift cover and its stagnation within the centre of
799 the studied basins (Lagabrielle et al., 2020). Syn-rift deposits of mature smooth-slope basins
800 are well known in the North Pyrenean Zone as the “Flysch Noir” group (Souquet et al., 1985;
801 Debros et al., 1990). Pioneer authors such as Ravier (1959) have shown that this sequence
802 locally experienced HT/LP metamorphism and deduced from microstructural analysis that a
803 large part of this evolution was static and necessarily linked to passive burial in the Flysch
804 Noir basins. These authors clearly showed that the thermal pulse did not affect the
805 Cenomanian sediments, leading to the concept of a “phase ante-cénomaniennne” (e.g. Casteras,
806 1933). In contrast, further studies pointed to the synkinematic character of some HT/LP
807 mineral assemblages and claimed that Pyrenean metamorphism was linked to the orogenic
808 evolution of the belt (e.g. Choukroune, 1976). Later detailed studies revisited the link between
809 thermal metamorphism and the opening of the North Pyrenean basins (e.g. Golberg &
810 Leyreloup, 1990). However, these studies did not consider the role of sedimentary burial.
811 Therefore, in this review, we note the existence of pioneer works that despite the rudimentary

812 geological knowledge of the time deduced 60 years ago that sediment burial was a key factor
813 in the thermal evolution of the basins at the Iberia-Eurasia plate boundary.

814 **4 Comparison of smooth-slope type basins with Atlantic type margins: West** 815 **Iberia and Bay of Biscay margins**

816 **4.1. Review of the timing of the main Mesozoic events and unconformities**

817 In this section we address the timing of formation of the syn-rift, breakup, post-rift and
818 drifting sequences in the reviewed basins as well as in the adjacent Northwest Iberia and Bay
819 of Biscay margins (Fig. 12). As a reminder, the syn-rift, breakup, post-rift and drifting
820 sedimentary sequences record respectively crustal thinning, mantle denudation, post-rift
821 thermal cooling and oceanic spreading (e.g. Alves et al., 2009, 2020; Soares et al., 2012;
822 Alves & Cunha, 2018). These sequences are delimited by major unconformities
823 corresponding (from bottom to top) to the base syn-rift unconformity (BSU; base of syn-rift
824 sequence), the lithospheric breakup surface (LBS; base of the breakup sequence) and the base
825 post-rift unconformity (BPU; base of the post-rift sequence) as defined by Soares et al. (2012)
826 and Alves and Cunha (2018).

827 All the reviewed basins as well as the West Iberia and Bay of Biscay margins are
828 characterised by the onset of Mesozoic rifting between the Late Jurassic and earliest
829 Cretaceous as they are all related to Iberia plate motion associated with the North Atlantic
830 opening (e.g. Alves et al., 2009; Nirrengarten et al., 2017; Angrand et al., 2020). The
831 beginning of rifting is slightly diachronous among these basins as evidenced by the timing of
832 formation of their BSU (Fig. 12). Nearly all of the rifting sequences in the Tartas and Arzacq
833 (Désaglaux & Brunet, 1990; Brunet, 1991; Serrano et al., 2006; Issautier et al., 2020), Parentis
834 (Brunet, 1994; Ferrer et al., 2008; Jammes et al., 2009; Tugend et al., 2015) and Columbrets
835 (Salas et al., 2001; Nebot & Guimerà, 2016; Etheve, 2016; Etheve et al., 2018; Roma et al.,

836 2018) basins end at the same time, in the Early or Middle Cenomanian (Fig. 12). The North
837 Pyrenean Zone basins end their syn-rift sequence slightly earlier, as they are characterised by
838 a breakup sequence developing between the latest Albian and the Early Cenomanian (Fig. 12;
839 Masini et al., 2014; Teixell et al., 2016; Pedrera et al., 2017; Espurt et al., 2019; Labaume and
840 Teixell, 2020). Thus, the North Pyrenean Zone basins are characterised by the development of
841 a LBS surface (*sensu* Soares et al., 2012) separating the Early Cretaceous syn-rift sequence
842 and the Latest Albian to Early Cenomanian breakup sequence. Hence the syn-rift sequence of
843 the reviewed basins lasts around 35–45 Myr according to the timing of formation of the
844 different basins whereas the breakup sequence of the North Pyrenean Zone basins lasts about
845 5 Myr (Fig. 12). In contrast to the reviewed basins, the West Iberia margins have a relatively
846 short (~20 Myr) syn-rift sequence and a longer (~15 Myr) breakup sequence (Soares et al.,
847 2012; Pereira & Alves, 2012; Alves & Cunha, 2018; Alves et al., 2020), and the Bay of
848 Biscay margins have equally long syn-rift and breakup sequences (~20 Myr) (Fig. 12;
849 Montardet et al., 1979; Brunet, 1994; Thinon, 1999; Thinon et al., 2001, 2003; Tugend et al.,
850 2015). This is consistent with interpretations of the West Iberia margin as having a syn-rift
851 sequence getting longer to the north and a continental breakup sequence becoming shorter to
852 the north (Alves et al., 2002, 2006, 2009).

853 All the reviewed basins record a post-rift thermal cooling stage in which a post-rift sequence
854 overlaps the syn-rift sequence in the Tartas, Arzacq, Cameros, Parentis and Columbrets basins
855 and the breakup sequence in the North-Pyrenean Zone basins, where they record a more
856 advanced Mesozoic extension. The post-rift unconformity is clearly visible in the reviewed
857 basins as erosional truncation of the syn-rift/breakup sequences and onlap of the post-rift
858 sequence, as seen in a seismic profile in the Arzacq (Fig. 2; Issautier et al., 2020), Parentis
859 (Fig. 4D; Jammes, 2019), and Columbrets (Figs. 5C-E; Etheve et al., 2018) basins, and in
860 field observations in the Cameros (Omodeo-Salé et al., 2014, 2017), Mauléon (e.g. Saspiturry

861 et al., 2019a), Basque-Cantabrian (Rat, 1988) and Internal Metamorphic Zone (Debroas,
862 1978, 1987, 1990) basins. Although the North Pyrenean Zone basins underwent continental
863 breakup, it did not progress to oceanic spreading. It is conceivable that the Pyrenean
864 compression may have obliterated the stratigraphic evidence for a drifting stage in the North
865 Pyrenean Zone basins. However, (1) the entire mid-Cenomanian to Late Santonian post-rift
866 sequence is fully preserved above the North Pyrenean Zone hyperextended domain and
867 clearly records a post-rift thermal cooling stage, (2) there is no evidence of subducted oceanic
868 crust in a passive seismic Vp/Vs model of the Mauléon basin (Fig. 7B, Wang et al., 2016) and
869 Internal Metamorphic Zone basins (Fig. 9B, Chevrot et al., 2018) or in a crustal-scale 3D
870 gravity inversion of the Basque-Cantabrian basin (Pedrera et al., 2017). In contrast to the
871 North Pyrenean Zone basins, oceanic spreading occurred during mid-Cenomanian time in the
872 eastern Bay of Biscay margins (Fig. 12; Montadert et al., 1979; Brunet 1994; Thinon, 1999;
873 Thinon et al., 2001, 2003; Gong et al., 2008; Tugend et al., 2015) and at the Cenomanian-
874 Turonian boundary in the Northwest Iberia margin and western Bay of Biscay margins (Fig.
875 12; Gong et al., 2008; Soares et al., 2012; Pereira & Alves, 2012; Alves & Cunha, 2018;
876 Alves et al., 2020). Although the duration of the breakup sequence differs strongly between
877 the North Pyrenean Zone basins and the Bay of Biscay and West Iberia margins, the breakup
878 affected all of these regions simultaneously from the mid-Cenomanian to the earliest Turonian
879 (Fig. 12).

880 **4.2 Architectural contrasts and discrepancies in the modes of crustal thinning**

881 Atlantic-type margins such as West Iberia (Boillot et al., 1980, 1995; Reston et al., 1995;
882 Whitmarsh et al., 2001; Péron-Pinvidic & Manatschal, 2009; Sutra et al. 2013; Péron-Pinvidic
883 et al., 2015) or the Bay of Biscay margins (Jammes, 2009; Jammes et al., 2010a, 2010b,
884 2010c; Tugend et al., 2014) are characterized by five distinctive features: (1) deformation
885 coupling that occurs when the ductile layer has been removed and deformation in the strong

886 and brittle upper crust couples with deformation in the strong lower crust/upper mantle, (2)
887 detachment faults at the top of the basement that accommodate crustal extension through
888 tilting of blocks of the basement and their pre-rift cover showing a coupled deformation, (3)
889 formation of continental crust extensional allochthons, made up of upper crust and pre-rift
890 cover, tectonically placed over exhumed lower crust or serpentinized mantle, (4) a wide
891 domain of exposed subcontinental mantle at the ocean-continent transition and (5) large-scale
892 serpentinization of the exhumed mantle that was still active at ambient seawater temperatures
893 (Fig. 13A). Numerical models of such systems reproduce the palaeoarchitecture of the
894 continental margins and the detachment faults responsible for crustal thinning (Lavier &
895 Manatschal, 2006; Huisman & Beaumont, 2011). Numerical studies have also shown that
896 continental crust thinning develops under conditions of high heat flow from the mantle due to
897 asthenospheric upwelling. In these sediment-starved Atlantic-type margins, only small
898 volumes of syn-rift sedimentary cover can be found in the hyperextended domain (Masini et
899 al., 2011, 2012; Péron-Pinvidic et al., 2015; Ribes et al., 2019) (Fig. 13A).

900 Palaeogeographic reconstructions (Ziegler, 1982; Dercourt et al., 1986; Ortí et al., 2017; Soto
901 et al., 2017) show that the distribution of the Pyrenean and peri-Pyrenean smooth-slope
902 extensional basins corresponds closely to the distribution of Late Triassic evaporites and
903 claystones (Lagabrielle et al., 2020). Numerical modelling shows that these Triassic deposits
904 played a major role at the onset of continental rifting as zones of decoupling between the
905 Palaeozoic basement and the Jurassic to Albian sedimentary cover (Duretz et al., 2019). Thus
906 Triassic salt does not allow the coupling of the basement and its sedimentary cover during
907 crustal extension characteristic of Atlantic-type margins. Moreover, in smooth-slope basins,
908 the lower/middle crust is not exhumed during the breakup sequence, unlike Atlantic-type
909 margins, as it has been laterally extracted during initial rifting. In fact, the only lower crustal
910 rocks cropping out along the Cretaceous North Pyrenean Rift axis are quite old, having been

911 previously exhumed during Permian time (de Saint-Blanquat, 1993; Olivier et al., 2004;
912 Cochelin et al., 2018a, 2018b; Saspiturry et al., 2019b).

913 Finally, unlike Atlantic-type margins, the hyperextended domain of smooth-slope basins
914 deforms under ductile conditions and HT/LP conditions due to (1) the displaced cover
915 remaining preserved in the centre of the basin while the lower crust is thinned ductilely and
916 (2) the continental crust being buried under a very thick pre-rift to syn-rift sedimentary cover.
917 However, as in Atlantic-type margins, the proximal margins become subject to extensional
918 brittle faulting as the crust acquires a normal thermal gradient of $\sim 30^{\circ}\text{C}/\text{km}$ (Saspiturry, 2019)
919 and the proximal sedimentary cover is thinned or removed (Fig. 13B). Thus in smooth-slope
920 basin margins, the proximal margins undergo brittle deformation while the hyperextended
921 domain undergoes ductile thinning due to a complex interaction between salt beds,
922 sedimentary burial and changes in the syn-rift thermal gradient (Fig. 13B). This elevated
923 thermal regime is associated with intense metasomatism and fluid circulation affecting both
924 the continental basement and the sedimentary cover at temperatures of $500\text{--}600^{\circ}\text{C}$ (Corre et
925 al., 2016; Quesnel et al., 2019; Lagabrielle et al., 2019a, 2019b). The peak syn-rift
926 temperature is mainly controlled by burial but can be locally influenced by fluid circulation.
927 Indeed, adiabatic temperatures in the sedimentary cover have been interpreted as the presence
928 of local fluid generation and convective cells in the Mauléon (Saspiturry, 2019) and
929 Boucheville (Boulvais et al., 2016) basins.

930 An important difference between the two geological settings is the fact that Atlantic-type
931 margin are sediment-starved (less than 2 km of burial) while smooth-slope basins are
932 sediment-rich (syn-rift sequence is more than 5 km thick). In addition, in the reviewed basins,
933 the pre-rift cover remaining in the central part of the basin, thanks to the Late Triassic salt
934 décollement, contributes significantly to the increase in burial and thus the peak temperature
935 at the base of the hyperextended domain by adding around 2–3 km of pre-rift sediments to

936 preexisting syn-rift and breakup sedimentary sequences that are nearly 5 km thick. Therefore,
937 the thinned continental basement and the exhumed mantle may be buried under 7–10 km of
938 pre-rift to syn-breakup sediments. This allows the crust to deform in a ductile way. Finally,
939 the pre-rift salt, which is mostly absent along Atlantic-type margins, is also a major
940 contributor to sedimentary burial. Thus, it participates in the ductile thinning of the
941 continental crust and its sedimentary cover, as pointed out by Lagabrielle et al. (2020).

942 The review of the sequence of the main tectono-stratigraphic events presented in section 4.1
943 (Fig. 12) also sheds light on another possible controlling factor. Indeed, the duration of syn-
944 rift sequences in smooth-slope basins (~35–45 Myr) is significantly longer than those of the
945 West Iberia and Bay of Biscay margins (~15–20 Myr). Thus, the long lifetime of smooth-
946 slope extensional basins could favour the depth-dependent ductile thinning of the
947 lower/middle crust by pure shear at the beginning of the rifting stage and may prevent brittle
948 structures from forming within the upper crust.

949 **5 Conclusions**

950 We infer the evolution of smooth-slope type basins in the Iberian-Eurasian plate junction from
951 the rifting to the breakup stage. At the beginning of the syn-rift stage, depth-dependent crustal
952 thinning is dominantly controlled by distributed pure shear thinning within the lower/middle
953 crust due to the presence of two decoupling levels: (1) the middle crust, which allows the
954 lower crust to be extracted laterally without disturbing significantly the upper crust, and (2)
955 within pre-rift Triassic salt beds, which act as a décollement between the upper crust and the
956 overlying sedimentary cover. Then simple shear becomes localized along a crustal
957 detachment connecting upward with the Late Triassic décollement layer, inducing shearing in
958 the pre-rift salt. When continental breakup occurs, the basin flanks are affected by brittle
959 deformation while the hyperextended domain undergoes dominantly ductile thinning. The rise

960 of the 300°C to 500°C isotherms in the hyperextended domain, from the syn-rift to the
961 continental breakup stage, implies that the originally crystalline upper continental crust and
962 the overlying pre-rift and syn-rift sedimentary pile are affected by depth-dependent ductile
963 thinning. The deformation style during rifting and continental breakup is mainly controlled by
964 burial that results from the complex interaction between the syn-rift sedimentary sequence
965 and the pre-rift salt beds that indirectly contribute to the burial by preserving from disruption
966 the pre-rift sequence in the basin core.

967 To summarize, hyperextension in Atlantic-type margins leads to a progressive embrittlement
968 of the continental crust due to progressive extraction of the ductile middle crust (e.g. Pérez-
969 Gussinyé et al., 2001; Reston, 2009; Sutra et al., 2013; Mohn et al., 2015). In margins with
970 smooth-slope basins, in contrast, crustal thinning is mostly ductile. The latter is favoured by
971 the lateral extraction of the deep crust and the occurrence of thick sedimentary cover (Asti et
972 al., 2019; Duretz et al., 2019; Lagabrielle et al., 2020). This implies that in the distal domain
973 of smooth-slope margins deformation coupling never occurs, thanks to progressive upward
974 migration of the brittle/ductile transition during rifting. Finally, in the reviewed basins,
975 sedimentary burial coupled to the presence of pre-rift salt and a long-lived rifting sequence
976 allow the continental crust to stretch in a dominantly ductile regime from rifting until breakup.

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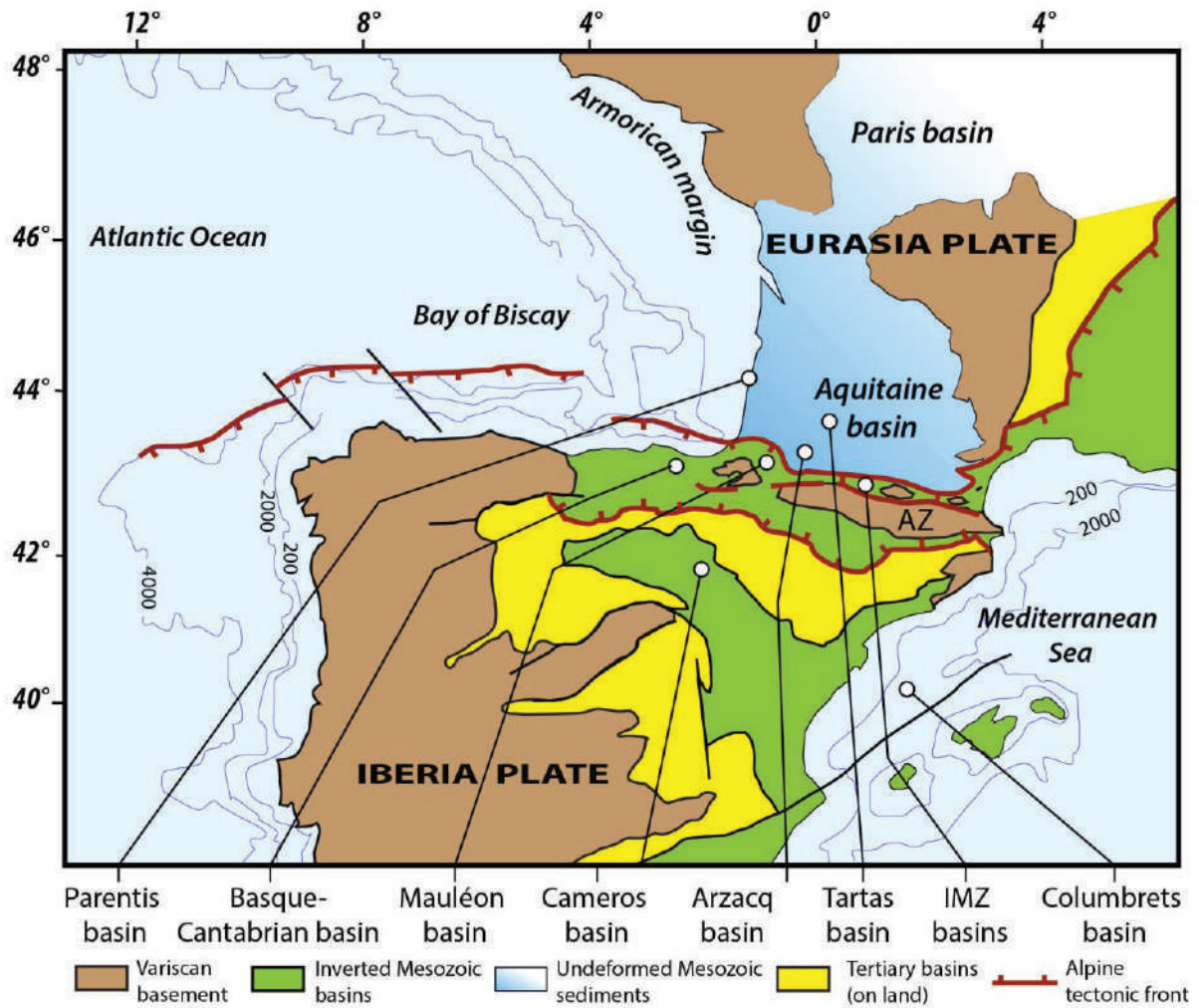
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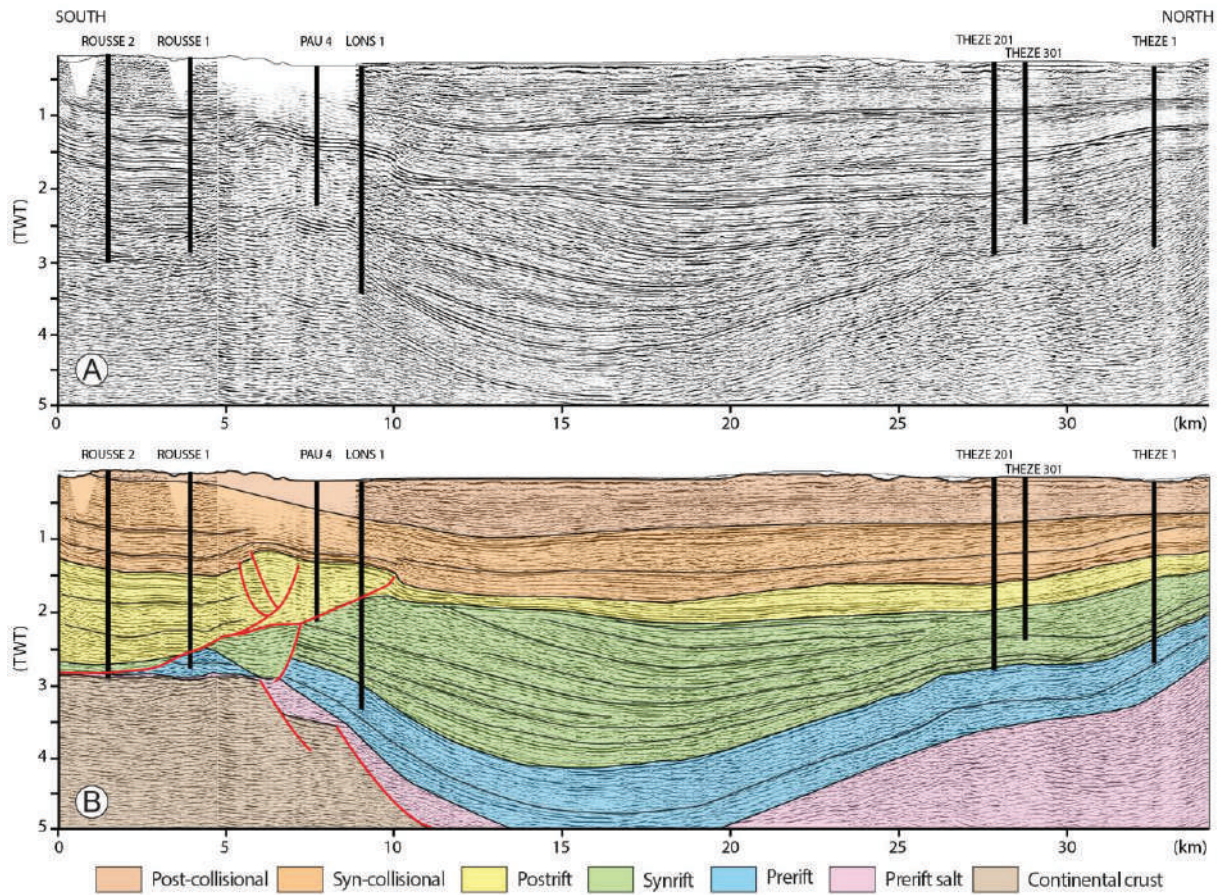
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- 1863

1864 **Figure.1**

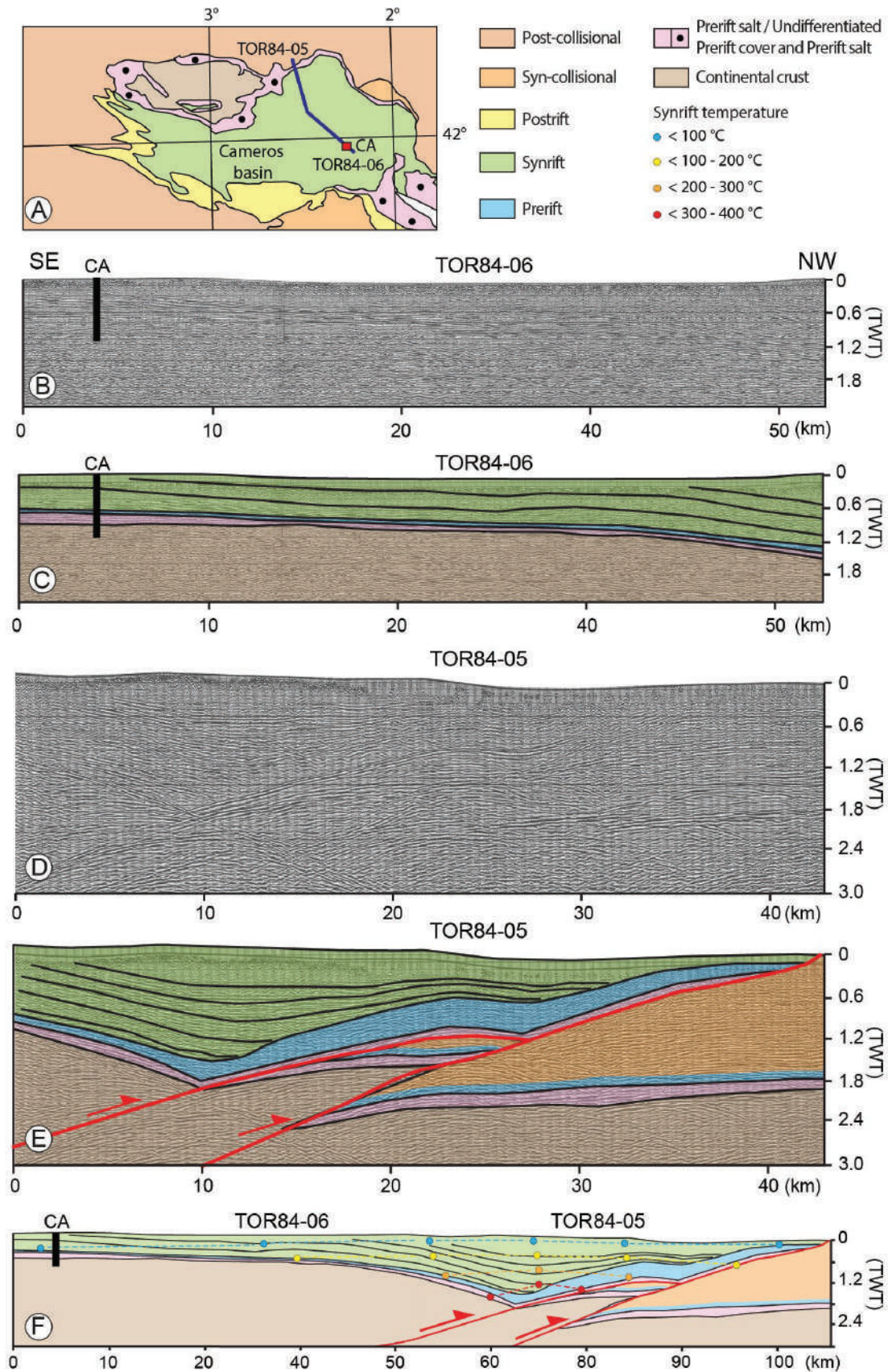


1866 **FIGURE 1** Simplified structural map of the Cantabrian-Pyrenean orogenic system and
1867 adjoining Iberia showing deformed and undeformed domains in the Eurasia plate and the
1868 locations of basins in this study (modified from Lagabrielle et al., 2020).

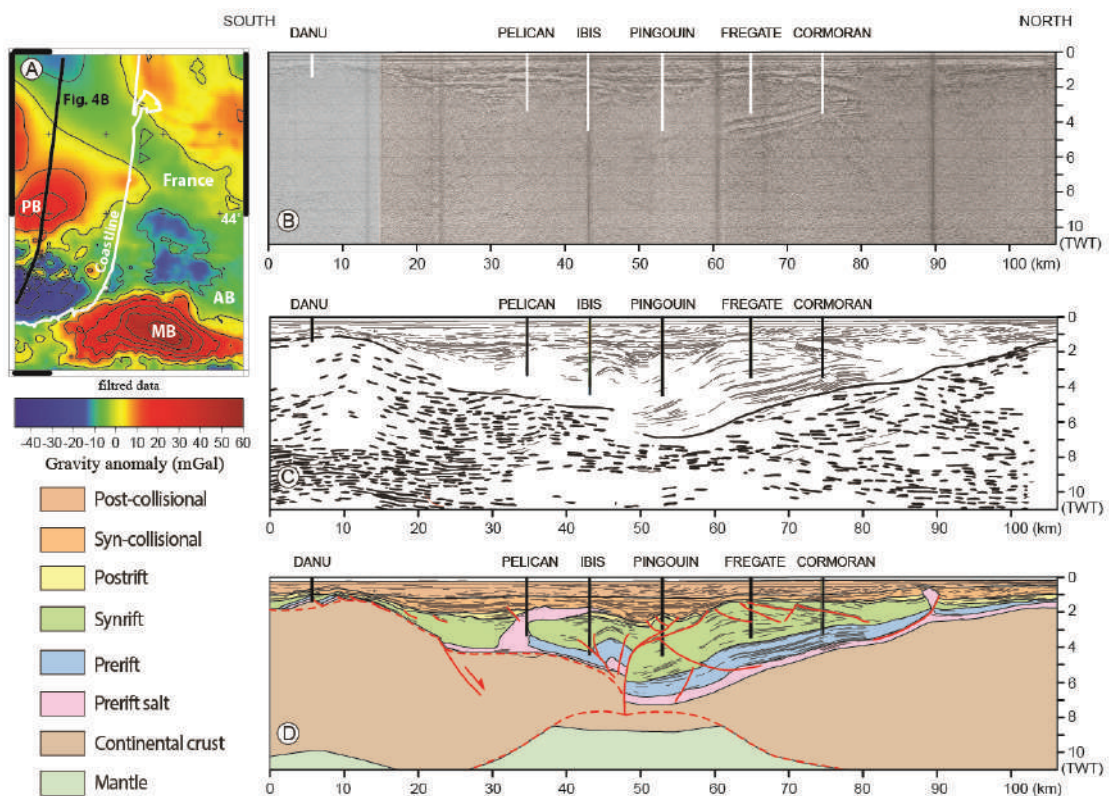


1870

1871 **FIGURE 2** Interpretation of a South-North seismic reflection profile across the Arzacq basin:
 1872 (A) Rouse-Thèze seismic reflection profile, calibrated using, from south to north, the
 1873 Rouse-2, Rouse-1, Pau-4, Lons-1, Thèze 201, Thèze 301 and Thèze 1 boreholes (Issautier et
 1874 al., 2020). (B) Interpreted section (modified from Issautier et al., 2020). The syn-rift sequence
 1875 is characterized by a maximum time thickness of around 2.5 TWT seconds corresponding to a
 1876 thickness of nearly 2000–3000 m. The depocenter of the syn-rift sequence in the Arzacq basin
 1877 migrates south from its position, indicating a northward salt-controlled cover gliding during
 1878 the rifting stage. The Arzacq syn-rift basin is characterized by a slight asymmetry of its
 1879 depocenter.

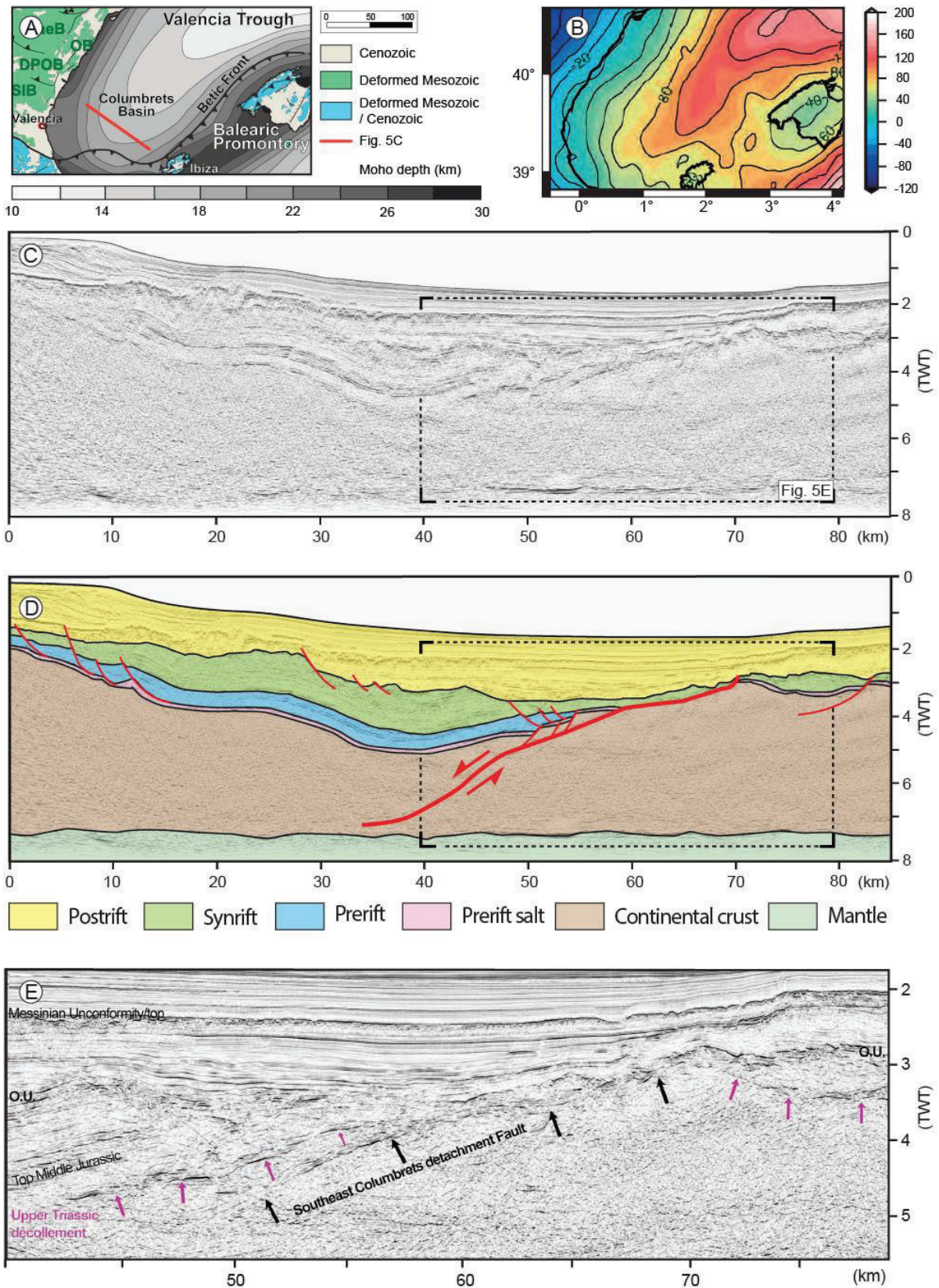


1882 **FIGURE 3** (A) Geological map of the Cameros basin (modified from Omodeo-Salé et al.,
1883 2014); (B-C) Interpretation of the SE-NW TOR84-06 seismic reflection profiles across the
1884 Cameros basin, location on Figure 3A (Omodeo-Salé et al., 2014), CA: Castelfrío 1 borehole;
1885 (D-E) Interpretation of the SE-NW TOR84-05 seismic reflection profiles across the Cameros
1886 basin, location on Figure 3A (Omodeo-Salé et al., 2014). (F) Line-drawing of the TOR84-05
1887 and TOR84-06 seismic reflection profiles. The syn-rift sequence displays an onlap geometry
1888 on the marine Jurassic substrate towards the north. The depocenter of the syn-rift sequence in
1889 the Cameros basin migrates north from its position, indicating southward salt-controlled cover
1890 gliding during the rifting stage rooting at depth on a crustal structure. The Cameros syn-rift
1891 basin is characterised by a slight asymmetry of its depocenter. The syn-rift sequence is
1892 characterised by a maximum time thickness of around 1.5 TWT seconds corresponding to a
1893 thickness of nearly 1500–2000 m. Available syn-rift paleotemperatures from Rat et al. (2019)
1894 indicate that the base of the syn-rift basin reached temperatures of around 300–400 °C.



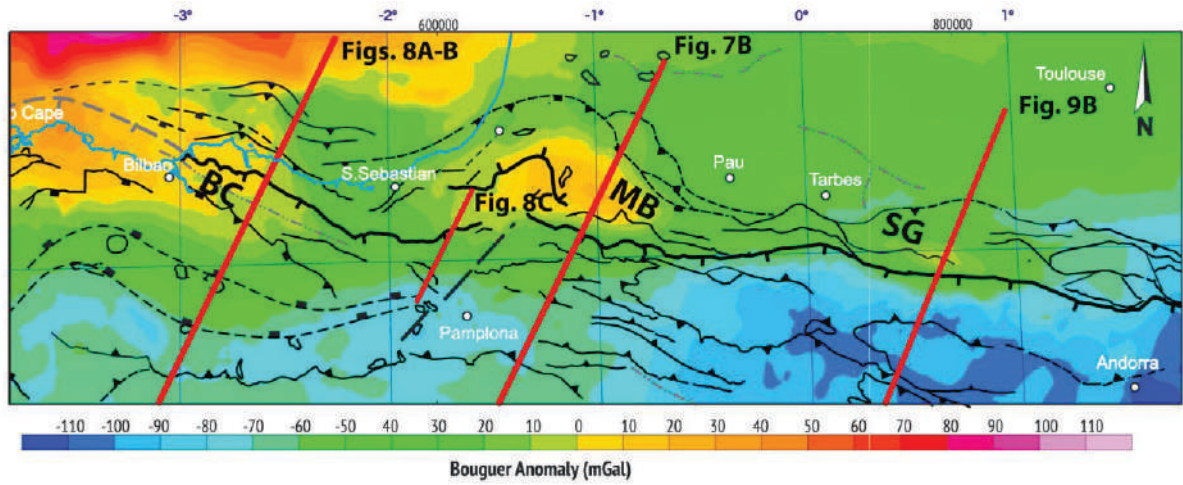
1896

1897 **FIGURE 4** Parentis basin architecture. (A) Filtered Bouguer gravity anomaly map (Jammes,
 1898 2009); PB: Parentis basin Bouguer Anomaly; MB: Mauléon basin Bouguer Anomaly; AB:
 1899 Arzacq basin. (B) Bay of Biscay ECORS profile, calibrated using, from south to north, the
 1900 Danu, Pelican, Ibis, Pingouin, Fregate and Cormoran boreholes (in Jammes, 2009). (C) Line
 1901 drawing of the sedimentary structures and deep crustal structures of the Bay of Biscay
 1902 ECORS profile (in Jammes, 2009). (D) Interpreted Bay of Biscay ECORS profile (modified
 1903 from Jammes, 2019). The depocenter of the syn-rift sequence in the Parentis basin migrates
 1904 south from its position indicating northward salt-controlled cover gliding during the rifting
 1905 stage, interpreted by Jammes (2009) and Tugend et al. (2014) as a northward-deepening
 1906 décollement rooting at depth on a crustal detachment fault. The Parentis syn-rift basin is
 1907 characterized by a mildly asymmetry of its depocenter. The syn-rift sequence is characterised
 1908 by a maximum time thickness of around 4 TWT seconds corresponding to a thickness of
 1909 nearly 3000–4500 m.



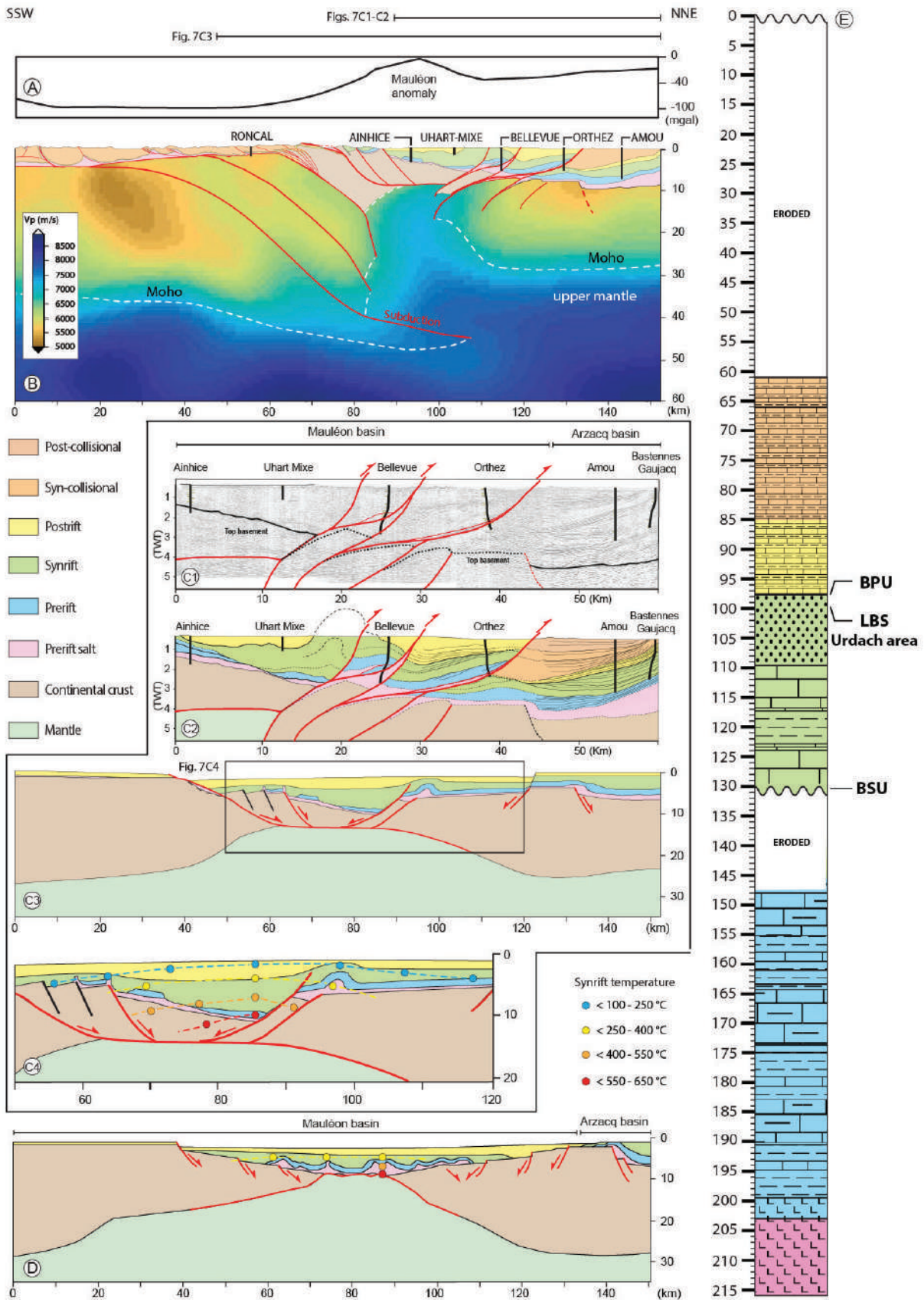
1912 **FIGURE 5** (A) Geological and structural map of the Columbrets basin showing the Moho
1913 depth varying from 10 to 30 km depth (modified from Etheve et al., 2018). (B) Map of the
1914 Bouguer anomaly (modified from et al., 2015). (C) NW-SE SGV01-113 seismic reflection
1915 profile; location on Figure 5A (Etheve et al., 2018). (D) Interpreted SGV01-113 seismic
1916 reflection profile (modified from Etheve et al., 2018). The syn-rift sequence is characterised
1917 by a maximum time thickness of around 1.5 TWT seconds corresponding to a thickness of
1918 nearly 1500–2000 m; however, its true thickness was greater as the top of this sequence is
1919 affected by a major Tertiary erosional unconformity. The depocenter of the syn-rift sequence
1920 in the Columbrets basin migrates south from its position, indicating northward salt-controlled
1921 cover gliding during the rifting stage interpreted by Etheve et al. (2018) as a northward-
1922 deepening salt décollement rooting at depth on a ductile crustal detachment fault. (E) Detail of
1923 the southern part of the SGV01-113 seismic reflection profile showing the northward-
1924 deepening ductile crustal detachment fault (Etheve et al., 2018).

1925 **Figure.6**

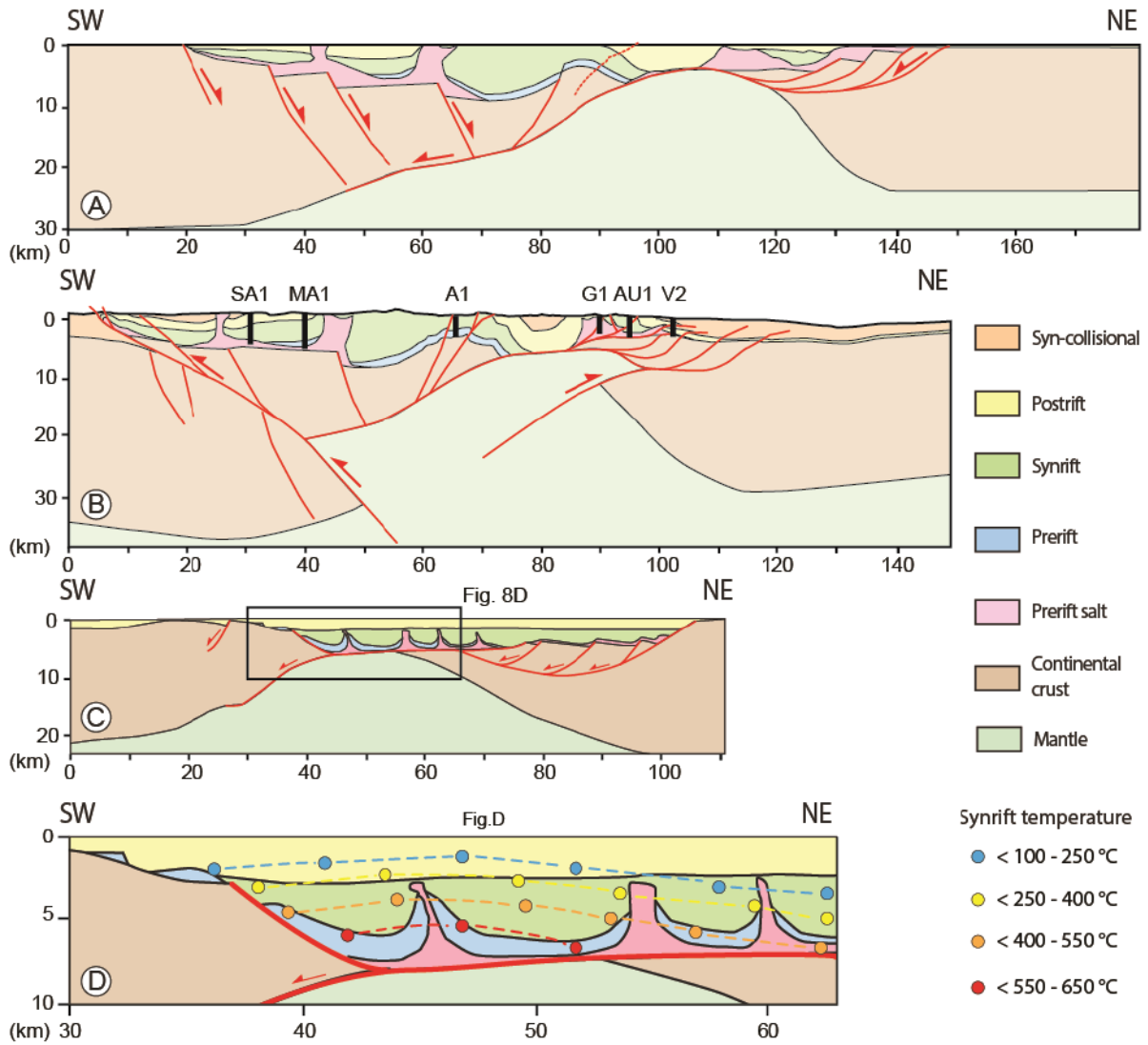


1926

1927 **FIGURE 6** Bouguer anomaly map from the Basque-Cantabrian basin to the Central Pyrenees
1928 (Pedrera et al., 2017). BC, Basque-Cantabrian Bouguer anomaly; MB, Mauléon basin
1929 Bouguer anomaly; SG, Saint-Gaudens Bouguer anomaly.

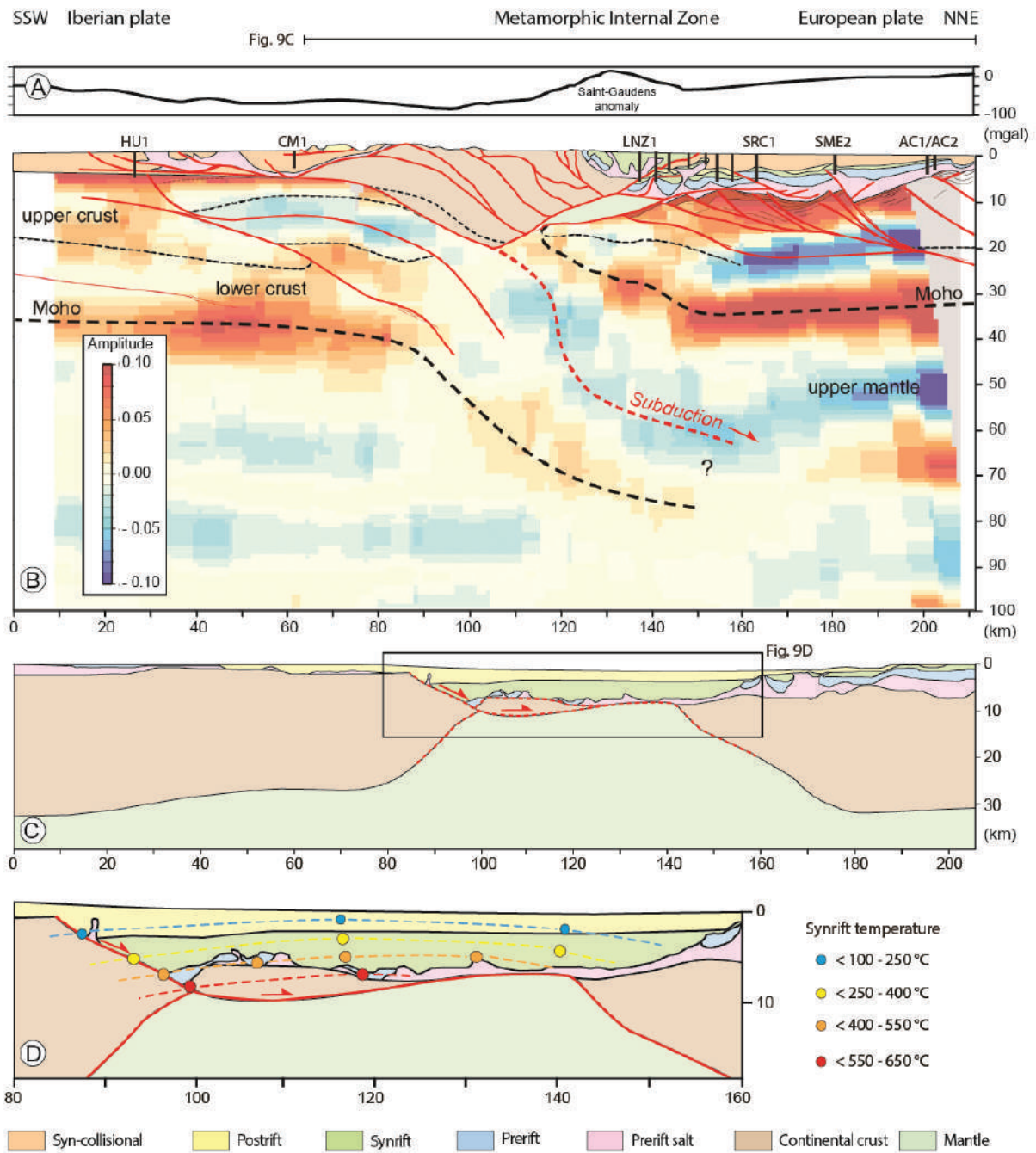


1932 **FIGURE 7** Mauléon basin overall architecture. (A) Bouguer anomaly (Casas et al., 1997).
1933 (B) Mauléon basin lithospheric section (see location on Figure 6). The lowermost part
1934 corresponds to the Wang et al. (2016) Vp model showing the presence at shallow depth of
1935 continental lithospheric mantle. The uppermost part is materialized by the Saspiturry et al.
1936 (2020a) crustal-scale balanced cross-section calibrated using (from south to north) the Ronca,
1937 Ainhice, Uhart-Mixe, Bellevue, Orthez, Amou and Bastennes Gaujacq boreholes. (C1, C2)
1938 Geological section through the present-day Mauléon basin based on interpreted seismic lines
1939 and field data (modified from Saspiturry et al., 2019a). The Mauléon basin is exposed within
1940 a pop-up structure formed during N-S Pyrenean compression. (C3) Palinspastic restoration, to
1941 Santonian time, of the Saspiturry et al. (2020a) crustal-scale balanced cross-section. (C4)
1942 Detail of the palinspastic restoration of the Saspiturry et al. (2020a) crustal-scale balanced
1943 cross-section, with RSCM syn-rift paleotemperature of Saspiturry (2019). (D) Palinspastic
1944 restoration, to Santonian time, of the Teixell et al. (2016) crustal-scale balanced cross-section,
1945 with RSCM syn-rift paleotemperature of Corre (2017). (E) Stratigraphic chart showing the
1946 position of the key markers of the Mauléon basin. BSU, basal syn-rift unconformity; LBS,
1947 lithospheric breakup surface; BPU, basal post-rift unconformity.



1949

1950 **FIGURE 8** (A) Basque-Cantabrian basin palinspastic restoration, to Santonian time, of the
 1951 Pedrera et al. (2017) crustal-scale balanced cross-section (see location on Fig. 6). (B) Basque-
 1952 Cantabrian basin crustal-scale balanced cross-section (Pedrera et al., 2017), calibrated using
 1953 (from south to north) the San Antonio 1 (SA-1), Marinda 1 (MA-1), Arratia 1 (A1), Gernika 1
 1954 (G1), Aulesti 1 (AU1) and Vizcaya C2 (V2) boreholes. (C) Schematic restoration of the
 1955 eastern part of the Basque-Cantabrian basin in the "Nappe des Marbres" (Ducoux et al.,
 1956 2019). (D) Detail of (C) with the RSCM syn-rift paleotemperatures measured by Ducoux et
 1957 al. (2019).

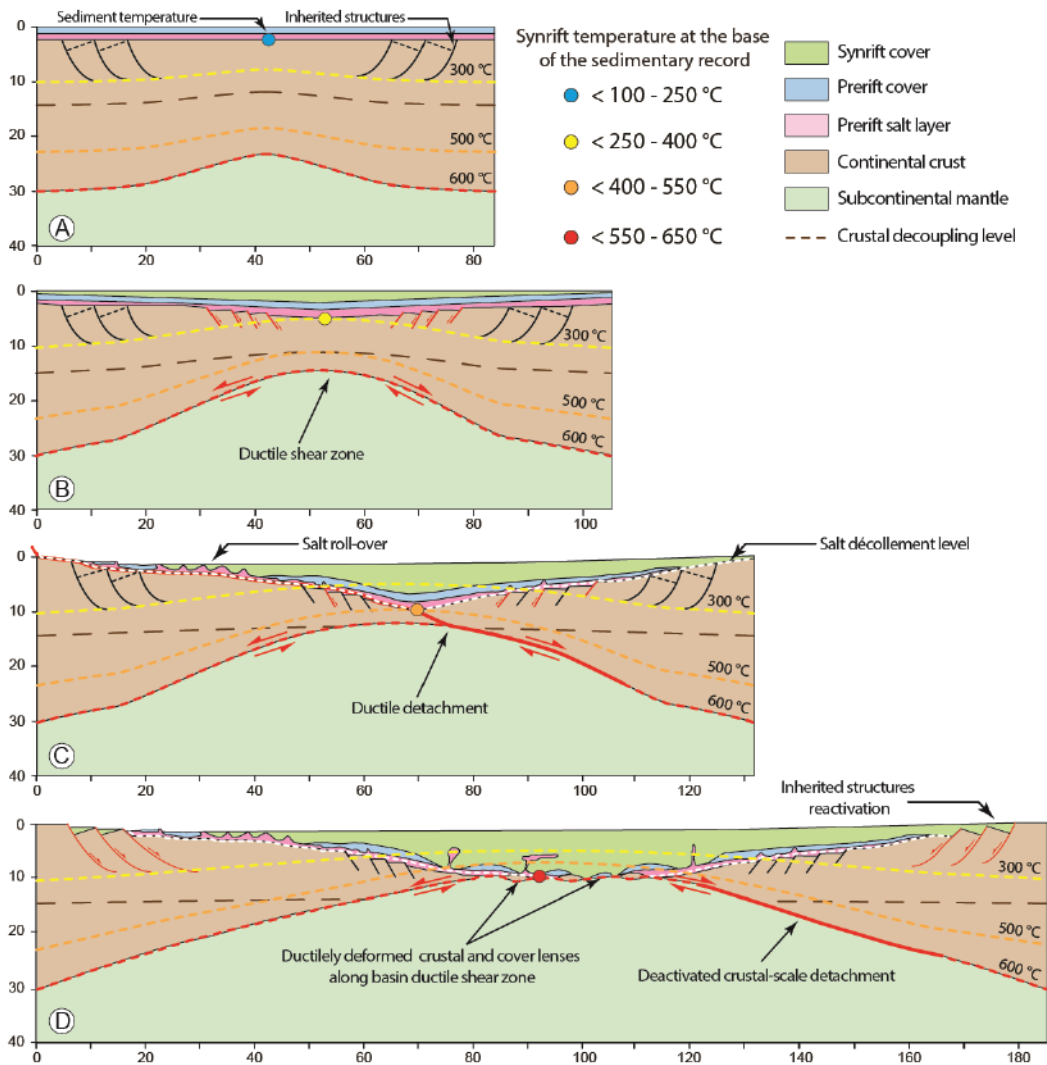


1959

1960 **FIGURE 9** Metamorphic Internal Zone overall architecture. (A) Bouguer anomaly (in Espurt
 1961 et al., 2019). (B) Metamorphic Internal Zone basin lithospheric section (see location on Figure
 1962 6). The lowermost part corresponds to the stack profile of receiver functions for the OROGEN
 1963 West profile across the Central Pyrenean belt (in Espurt et al., 2019) showing the presence at

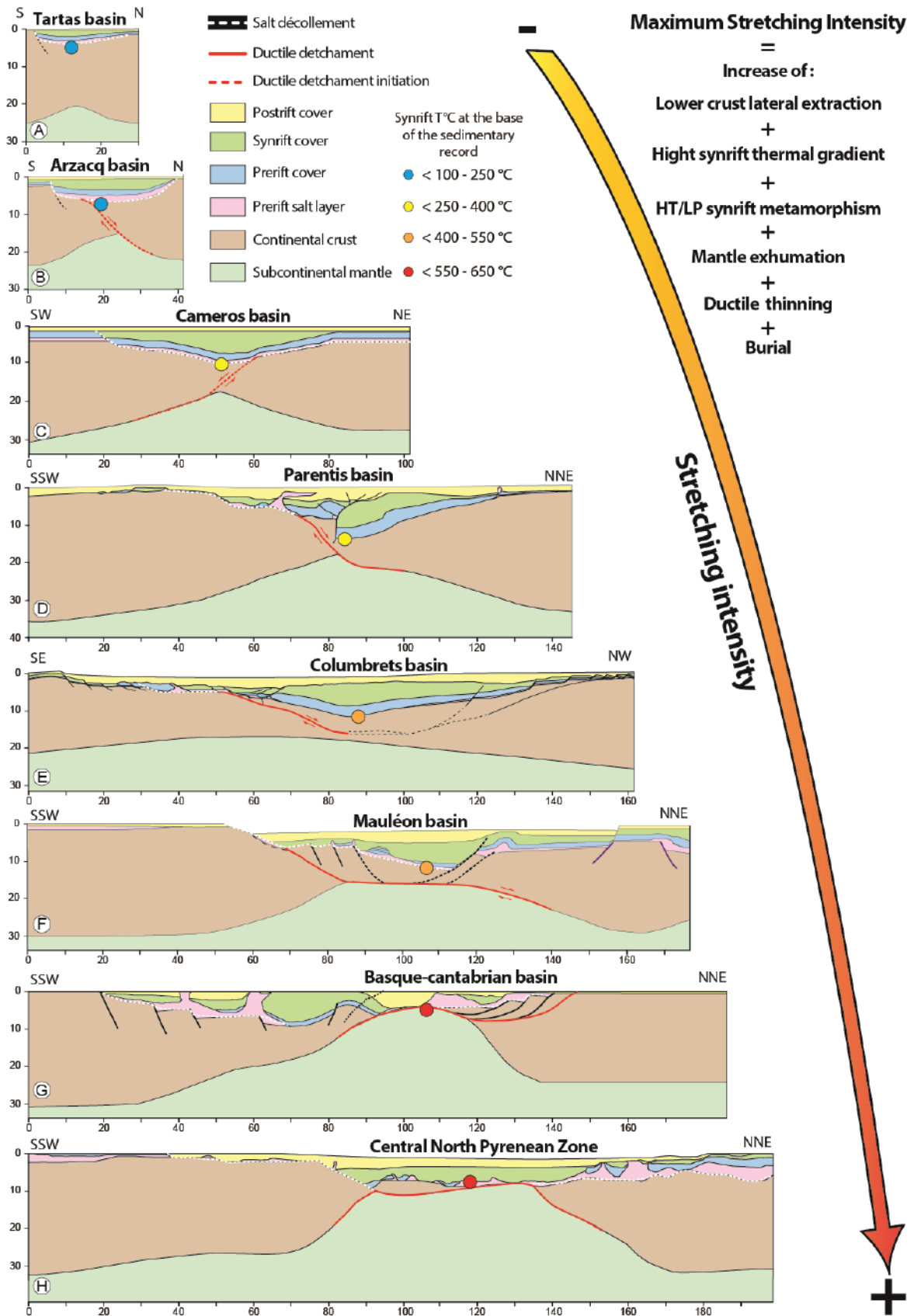
1964 shallow depth of continental lithospheric mantle. The uppermost part is based on the Espurt et
1965 al. (2019) crustal-scale balanced cross-section calibrated using (from south to north) the
1966 Huesca 1 (HU-1), Campanue-1 (CM-1), Lannemezan 1 (LNZ1), Sariae 1 (SRC-1), Saint
1967 Médard 2 (SME-2), Auch 1 (AC-1) and Auch 2 (AC-2) boreholes. (C) Palinspastic
1968 restoration, to Santonian time, of the Espurt et al. (2019) crustal-scale balanced cross-section.
1969 (D) Detail of the palinspastic restoration of the Espurt et al. (2019) crustal-scale balanced
1970 cross-section, with RSCM syn-rift paleotemperatures.

1971 **Figure.10**

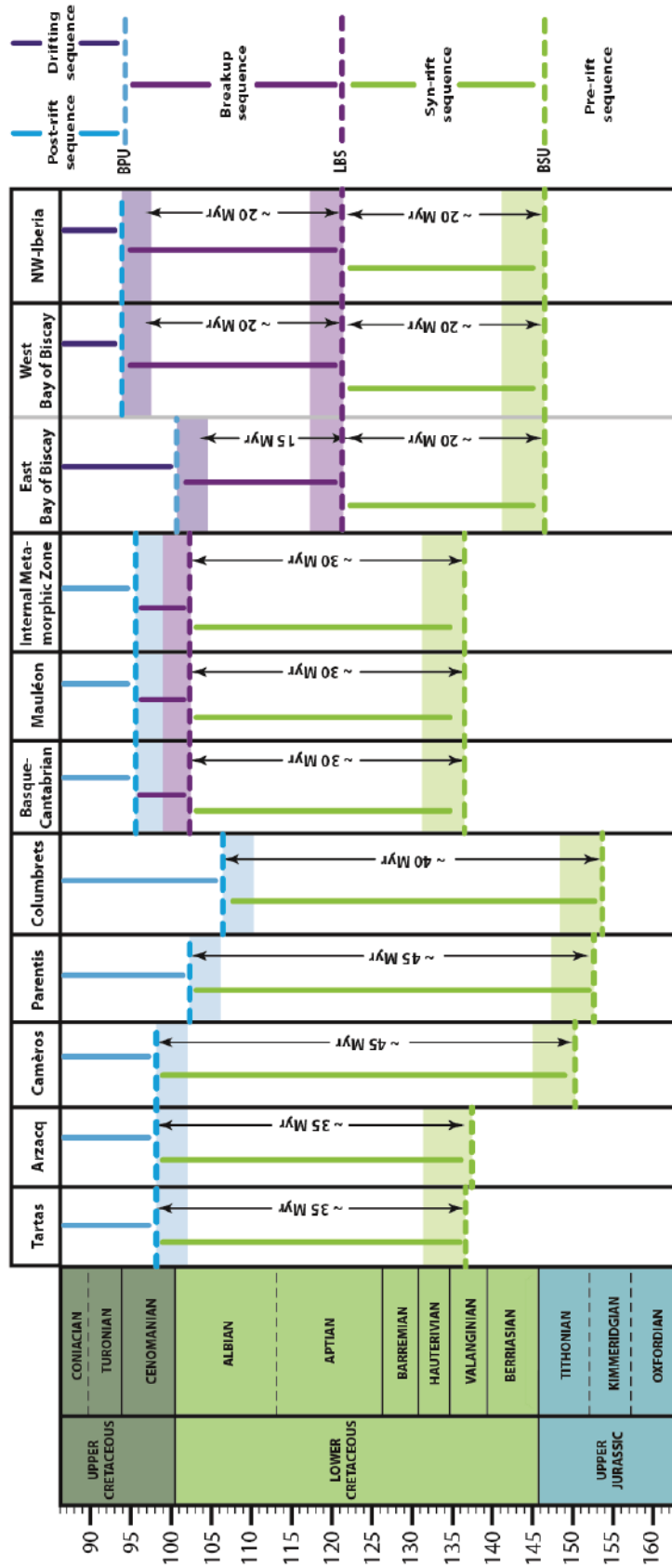


1972

1973 **FIGURE 10** Conceptual model of the evolution of a smooth-slope type basin. (A) Inherited
 1974 thin lithosphere. (B) Pure shear dominated thinning: formation of a symmetric basin
 1975 characterized by shallow-water sediments. (C) Simple shear dominated thinning: basinward
 1976 gliding of the pre-rift sedimentary cover along a major detachment connecting towards the
 1977 surface with the Late Triassic salt décollement, leading to an asymmetrical basin shape. (D)
 1978 Breakup stage resulting in the formation of a pseudo-symmetric basin undergoing brittle
 1979 deformation on the proximal margins and ductile deformation in the basin core. No vertical
 1980 exaggeration. Coloured dashed lines represent crustal isotherms; coloured circles indicate
 1981 maximum sediment temperatures during HT/LP metamorphism.

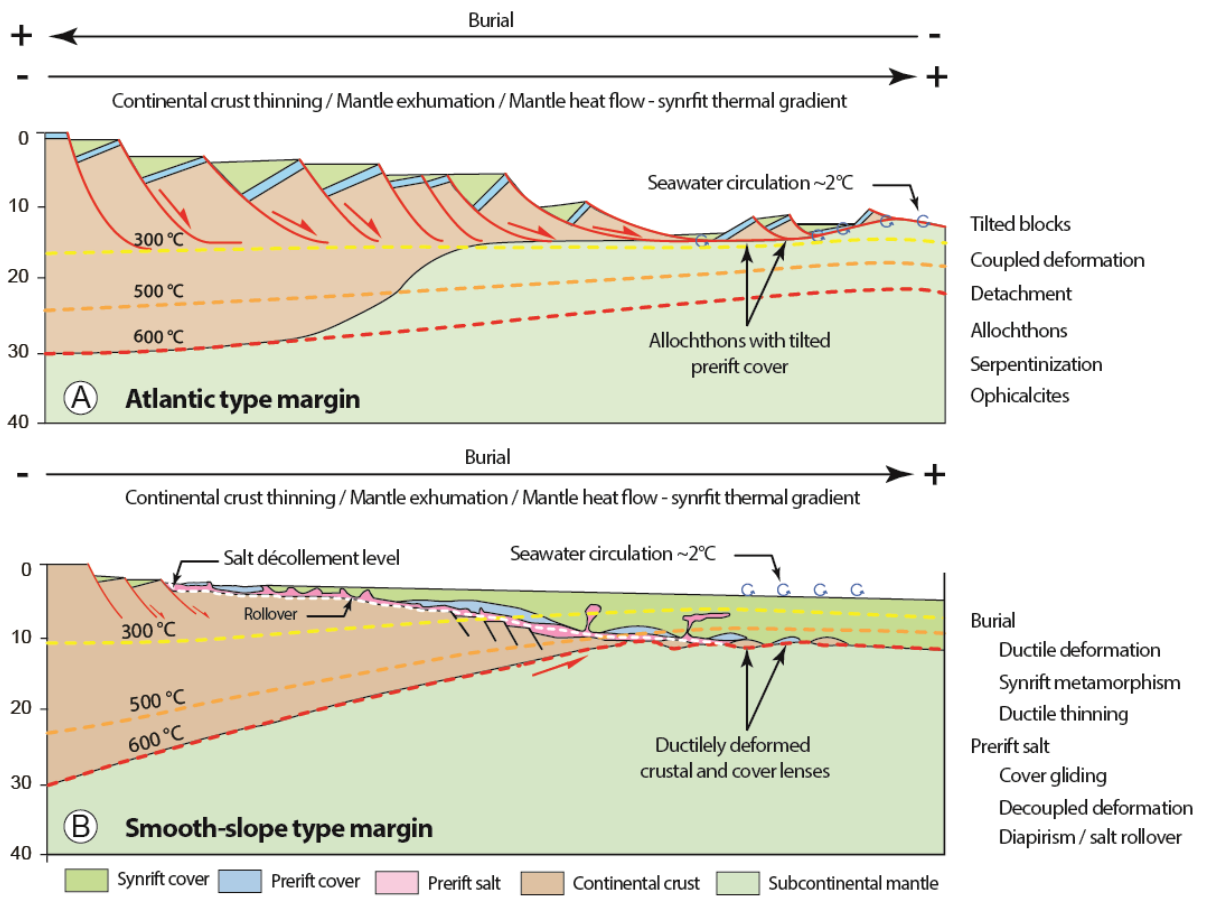


1984 **FIGURE 11** Smooth-slope type basins classified according to the degree of continental crust
1985 extension: (A) Tartas sub-basin (Issautier et al., 2020); (B) Arzacq sub-basin (Issautier et al.,
1986 2020); (C) Cameros basin (modified after Casas-Sainz & Gil-Imaz, 1998); (D) Parentis basin
1987 (Tugend et al., 2014); (E) Columbrets basin (Etheve et al., 2018); (F) Mauléon basin
1988 (Saspiturry et al., 2019a); (G) Basque-Cantabrian basin (Pedrera et al., 2017); (H) Central
1989 North Pyrenean Zone basins (Espurt et al., 2019); locations in Figure 1. No vertical
1990 exaggeration. The large arrow indicates the degree of extension and the rise in peak
1991 metamorphic temperatures throughout the basin evolutionary sequence. Coloured circles
1992 indicate maximum sediment temperatures during HT/LP metamorphism.



1995 **FIGURE 12** Chronological chart showing the timing of the syn-rift, late syn-rift, breakup
1996 (mantle exhumation), post-rift and oceanic spreading events of the reviewed basins, the West
1997 Iberia basins and the Bay of Biscay margins. References used to define the timing of the
1998 different events: Tartas and Arzacq basins (Désaglaux & Brunet, 1990; Brunet, 1991; Serrano
1999 et al., 2006; Issautier et al., 2020), Camèros basin (Platt, 1990; Mas et al., 1993; Casas-Sainz
2000 & Gil-Imaz, 1998; Salas et al., 2001; Omodeo-Salé et al., 2014, 2017), Parentis basin (Brunet,
2001 1994; Ferrer et al., 2008; Jammes et al., 2009; Tugend et al., 2015), Columbrets basin (Salas
2002 et al., 2001; Nebot & Guimerà, 2016; Etheve, 2016; Etheve et al., 2018; Roma et al., 2018),
2003 Basque-Cantabrian basin (Azambre & Rossy, 1976; Rat et al., 1983; Rat, 1988; García
2004 Mondéjar et al., 1996; Castañares et al., 1997; Castañares & Robles, 2004; Pedrera et al.,
2005 2017; Ducoux et al., 2019), Mauléon basin (Boirie, 1981; Fixari, 1984; Souquet et al., 1985;
2006 Jammes et al., 2009; Debroas et al., 2010; Masini et al., 2014; Teixell et al., 2016; Saspiturry
2007 et al., 2019a; Labaume and Teixell, 2020), Internal Metamorphic Zone basin (Lagabrielle &
2008 Bodinier, 2008; Lagabrielle et al., 2010, 2019; Clerc & Lagabrielle, 2014; Clerc et al., 2014,
2009 2015; de Saint Blanquat et al., 2016; Teixell et al., 2018; Espurt et al., 2019), Northwest-
2010 Iberia margin (Soares et al., 2012; Pereira & Alves, 2012, Alves & Cunha, 2018; Alves et al.,
2011 2020), and Bay of Biscay margins (Montardet et al., 1979; Brunet 1994; Thinon, 1999;
2012 Thinon et al., 2001, 2003; Gong et al., 2008; Tugend et al., 2015). BSU, basal syn-rift
2013 unconformity; LBS, lithospheric breakup surface; BPU, basal post-rift unconformity.
2014 Coloured shaded areas represent time uncertainty of the BSU, LBS and BPU unconformities.

2015 **Figure.13**



2016

2017 **FIGURE 13** Schematic diagrams showing (A) Atlantic-type margin architecture (section
 2018 modified from Péron-Pinvidic et al., 2015) and (B) smooth-slope type margin architecture.
 2019 Coloured dashed lines represent crustal isotherms; coloured circles indicate maximum
 2020 sediment temperatures during HT/LP metamorphism.