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Lagabrielle, Yves; Asti, Riccardo; Duretz, Thibault; [et al.]. «A review of cretaceous smooth-slopes extensional basins along the Iberia-Eurasia plate boundary : How pre-rift salt controls the modes of continental rifting and mantle exhumation». *Earth Science Reviews*, Vol. 201 (February 2020), art. 103071. DOI 10.1016/j.earscirev.2019.103071

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Revised version 2 December 17th 2019

A review of Cretaceous smooth-slopes extensional basins along the Iberia-Eurasia plate boundary: how pre-rift salt controls the modes of continental rifting and mantle exhumation

by :

Yves Lagabrie¹, Riccardo Asti¹, Thibault Duretz¹, Camille Clerc², Serge Fourcade¹, Antonio Teixell⁴, Pierre Labaume³, Benjamin Corre¹, Nicolas Saspiturry⁵

Addresses

1. Université de Rennes, CNRS, UMR 6118 Géosciences Rennes, Campus de Beaulieu, 35000 Rennes, France

2. ISEA, Université de la Nouvelle Calédonie, 98800 Nouméa, Nouvelle Calédonie

3. Géosciences Montpellier, CNRS-Université de Montpellier-Université des Antilles, 34095 Montpellier, France

4. Dept. de Geologia, Universitat Autònoma de Barcelona, 08193 Bellaterra, Spain

5. EA4592 Géoressources & Environnement, Bordeaux INP, Université Bordeaux Montaigne, 1 Allée Daguin, 33607 Pessac, France

Riccardo Asti riccardo.asti@univ-rennes1.fr, Duretz Thibault thibault.duretz@univ-rennes1.fr, Camille Clerc camille.clerc@univ-nc.nc, "Serge Fourcade; sg.fourcade@orange.fr" sg.fourcade@orange.fr, Antonio Teixell antonio.teixell@uab.cat, "Pierre Labaume" labaume@gm.univ-montp2.fr, "Benjamin Corre" benjcorre@hotmail.fr, Saspiturry Nicolas saspiturry.nicolas@gmail.com,

Corresponding author: Yves Lagabrie

yves.lagabrie@univ-rennes1.fr

35 **Key words** : Smooth-slopes basins; symmetrical profile; Iberia; Eurasia; Triassic evaporites;
36 décollement layer; thermal anomaly; sedimentary burial; dominating-ductile tectonic
37 regime.

38 **Short abstract (for submission)**

39

40 We enhance a striking correlation between the paleogeography of Upper Triassic deposits
41 and the mode of crustal stretching of the north Iberia plate during the Cretaceous
42 transtensional event. The basins which opened during the mid-Cretaceous times along the
43 Iberia-Eurasia plate boundary (like the emblematic Parentis basin) exhibit a peculiar
44 synclinal-shaped profile and are devoid of prominent block faulting. The top of the basement
45 is characterized by gentle slopes dipping symmetrically towards the basin center. Based on a
46 comparison with rifting models established from the North Pyrenean Zone, this architecture
47 appear to result from the thinning of the central basin continental crust under dominating-
48 ductile deformation in greenschist facies conditions. The common character shared by all
49 the pre-rift sequences of the studied basins is the presence of a thick low-strength Upper
50 Triassic evaporites and clays layer belonging to the Keuper group and forming a thick pre-rift
51 low-strength unit. Efficient décollement along this layer triggers mechanical decoupling and
52 gliding of the pre-rift cover remaining in the basin center as the continental crust is laterally
53 extracted. Using recent paleogeographic reconstructions, we show that the distribution of
54 the Keuper sediments remarkably matches the distribution of the Pyrenean and peri-
55 Pyrenean, smooth-slopes type basins. This allows for the first time to propose a genetic link
56 between the distribution of evaporite-bearing pre-rift sedimentary formations and the
57 development of smooth-slopes rift basins.

58

59 **Abstract (long version)**

60

61 This article points out for the first time a striking correlation between the paleogeography of
62 Upper Triassic deposits and the mode of crustal stretching around and inside the Northern
63 Iberia plate during the Cretaceous transtensional event. First, we propose a review of the
64 architecture of the basins, which opened during the mid-Cretaceous times along the Iberia-
65 Eurasia plate boundary. Similarly to the emblematic Parentis basin, these basins exhibit a
66 peculiar synclinal-shaped profile and are devoid of prominent block faulting. The top of the
67 basement is characterized by gentle slopes, which dip symmetrically towards the center of
68 the basins. As revealed by recent comparisons with geologically-constrained rifting models
69 established from the North Pyrenean Zone, this architecture results from the thinning of a
70 heterogeneous continental crust under greenschist facies conditions. Basement
71 deformations are thus dominantly ductile and are characterized by large-scale boudinage,
72 hyper-thinning and subsequent lateral extraction. Bulk deformation of the upper and
73 middle crust leads to the formation of anastomosed shear zones and the development of
74 mylonitic fabric. Tectonic lenses consisting of crustal material remain welded on top of the
75 exhuming mantle. The common character shared by all the pre-rift sequences of the studied
76 basins is the presence of a thick low-strength Upper Triassic evaporites and clays layer
77 belonging to the Keuper group (i.e. pre-rift salt and clay unit). In the studied basins, efficient
78 décollement along the pre-rift salt and clay unit triggers mechanical decoupling and gliding
79 of the pre-rift cover that remains in the center of the basin. Thus, during the early rifting
80 phase, the basement undergoes thinning while the pre-rift cover remains preserved in the
81 basin center. In response to hyper-thinning and horizontal extraction of the continental
82 crust, hot mantle is exhumed beneath the pre- and syn-rift cover. Subsequent thermal
83 exchange (i) promotes ductile deformation of the basement and (ii) induces the

84 development of HT-LP metamorphic conditions in the pre-rift sediments and at the base of
85 the syn-rift flysch levels. This thermal event is well recorded in the axial portion of the
86 Pyrenean realm (future North Pyrenean Zone) as well as in the pre-rift sediments of the
87 Cameros basin (northern Spain). The architecture of the smooth-slopes type basins thus
88 contrasts with the structure of Iberia-Newfoundland Atlantic margins which are
89 characterized by (i) top-basement detachment faults accommodating crustal extension
90 through rotation and translation of undeformed basement blocks, and (ii) the
91 individualization of continental extensional allochthons tectonically emplaced over exhumed
92 lower crust or mantle rocks. Finally, using recent paleogeographic reconstructions, we show
93 that the distribution of the pre-rift salt and clay unit remarkably matches the distribution of
94 the Pyrenean and peri-Pyrenean smooth-slopes type basins. This allows for the first time to
95 propose a genetic link between the distribution of evaporite-bearing pre-rift sedimentary
96 formations and the development of smooth-slopes basins.

97

98 Introduction

99

100 More than 30 years ago, important steps in our understanding of the mechanisms of
101 continental rifting were achieved through the acquisition and interpretation of ECORS
102 seismic reflection profiles (1983-1994) (Damotte et al., 1998) (fig. 1). New images of crustal
103 and Moho geometries beneath stretched continental crust were obtained, shedding light on
104 important discrepancies between structural patterns at the base of rift systems. In
105 particular, the ECORS profiles from the Rhine graben and the Parentis basin displayed
106 contrasting images of the thinned upper lithosphere. In the first case, the upper crust
107 appears clearly rifted and offset by stepping normal faults (Brun et al., 1991) whilst, despite
108 slight tectonic inversion, the second case exhibits a smooth basement top, with gentle
109 slopes dipping symmetrically towards the basin center (Bois et al., 1997). Because only few
110 cases of Parentis-type architecture were observed worldwide, little attention has been paid
111 to this symmetrical, smooth-slopes type continental rifts, which apparently lacks major
112 upper crustal faulting and block tilting. Rather, based on the emblematic case of the Iberia-
113 Newfoundland conjugate margins, most of the current models of rift-related crustal thinning
114 are characterized by tilted continental blocks and prominent detachment faulting (fig. 2).
115 Such architecture indicates that the upper crustal levels behave under a dominant brittle
116 mode in the proximal (or continent-ward) as well as in the distal (or ocean-ward) margin
117 domains. In such models, shallow detachment faults accommodate upper crustal extension
118 through the rotation and the translation of undeformed basement blocks. In the distal
119 margin, these blocks, referred to as extensional allochthons, are covered by syn-rift and
120 post-rift sediments and may lie tectonically over exhumed units, including subcontinental
121 mantle (Reston et al., 1995; Manatschal et al., 2001; Jammes et al., 2010c; Osmundsen and

122 Peron-Pinvidic, 2018, and references therein).

123 Recent geological investigations in the northern units of the Pyrenean belt forming the
124 North Pyrenean Zone (NPZ) as well as in the Basque-Cantabrian basin (fig. 1) show that
125 Parentis-type, smooth-slopes basins of mid-Cretaceous age were distributed all along the
126 boundary between the northern Iberia and southern Eurasia plates, thus introducing doubts
127 regarding the ubiquitous character of Iberia-Newfoundland-type margins (Lagabriele et al.,
128 2010; Clerc and Lagabriele, 2014; Teixell et al., 2016; 2018; Asti et al., 2019). In this article,
129 we first list the main characteristics of these smooth-slopes basins, based on the analysis of
130 detailed geological reconstructions from areas exposed all along the northern flank of the
131 Pyrenean belt. We then review the distribution of such basins at the scale of the Iberia and
132 Eurasia plates. We finally discuss some of the key-factors controlling the evolution of
133 smooth-slopes basins and we evaluate how such information increases our understanding of
134 the mechanisms of continental rifting and passive margin formation.

135

136 **I. Symmetrical, smooth-slopes basins of the north Iberia margin: insights from the North**
137 **Pyrenean Zone (NPZ) and the Basque-Cantabrian range**

138

139 The Pyrenees and the Cantabrian mountain (fig. 1) form a narrow, N110 trending fold-and-
140 thrust belt resulting from the collision of the northern edge of the Iberia plate (north Iberia
141 margin) with the southern edge of the Eurasia plate during the Late Cretaceous-Tertiary
142 (Choukroune and ECORS team, 1989; Muñoz, 1992; Deramond et al., 1993; Roure and
143 Choukroune, 1998; Teixell, 1998; Vergés and Garcia-Senz, 2001; Pedrera et al., 2017; Teixell
144 et al., 2018). Convergence initiated ca. 83 Ma, following an almost 40 Ma long period of

145 transtensional motion in relation with the counterclockwise rotation of Iberia relative to
146 Eurasia, also leading to oceanic spreading in the Bay of Biscay between Chron M0 and A33o
147 (ca. 125-83 Ma) (Le Pichon et al., 1971; Choukroune and Mattauer, 1978; Olivet, 1996;
148 Sibuet et al., 2004). Convergence led to the partial or complete tectonic inversion of
149 discontinuous Cretaceous rift basins opened along the Iberia-Eurasia plate boundary during
150 the transtensional episode (Puigdefàbregas and Souquet, 1986; Debroas, 1990). Rotation
151 was achieved just before the Albian according to paleomagnetic data collected inland (Gong
152 et al., 2008). Earlier Triassic and Jurassic rifting events preceded the development of the
153 Cretaceous rifts (Canérot, 2017, and references therein).

154 Along the northern flank of the Pyrenees, more than forty, up to km-sized exposures of
155 subcontinental lherzolites are widespread within the Mesozoic pre-rift and syn-rift
156 sediments forming the NPZ (Monchoux, 1970; Vielzeuf and Kornprobst, 1984; Fabriès *et al.*,
157 1991, 1998). The NPZ is bounded by two major thrusts, the North Pyrenean Fault (NPF) to
158 the South and the North Pyrenean Frontal Thrust (NPFT) to the North. The NPF represents
159 the tectonic boundary between the NPZ and the prominent axial zone of the belt (AZ)
160 constituted of a stack of Paleozoic basement units (Choukroune, 1976a; 1976b; Choukroune
161 et Mattauer, 1978).

162 Based on field and geophysical evidence from the central and western NPZ, exhumation of
163 sub-continental mantle is shown to have occurred coevaly with extreme thinning of the
164 continental crust in the Pyrenean realm during the mid-Cretaceous (Lagabrielle and
165 Bodinier, 2008; Jammes et al., 2009; Masini et al., 2014; Odlum and Stockly, 2019).
166 Therefore, mantle exhumation (locally followed by peridotite exposure up to the floor of the
167 Pyrenean basins) is now considered as a general mechanism accounting for the presence of
168 ultramafic material within the NPZ. It is established that the well-known regional high

169 temperature and low pressure (HT–LP) Pyrenean metamorphism (Ravier, 1957; Azambre &
170 Rossy, 1976; Bernus-Maury, 1984) developed along the southern NPZ in relation with
171 continental thinning during the major Cretaceous extensional event (Vielzeuf and
172 Kornprobst, 1984; Dauteuil and Ricou, 1989; Golberg & Leyreloup 1990; Clerc et al., 2015b;
173 2016). Following the early ECORS profiles (Choukroune and ECORS team, 1989), additional
174 information on the architecture of the paleo-margin of Northern Iberia in the Pyrenees is
175 provided by recent interpretation of tomographic data acquired during the temporary
176 PYROPE and IBERARRAY experiments across the Pyrenees (Chevrot et al., 2015; 2018; fig. 1).
177 Based on such data set, Wang et al. (2016) suggest the inversion of a northern Iberia margin
178 characterized by a short necking domain and a large distal domain made of strongly
179 attenuated crust (less than 10 km thick) overlying a large volume of subcontinental mantle.
180 As discussed further in this article, this domain can be compared to large sheets of hyper-
181 extended continental crust found in the distal portions of present-day passive continental
182 margins (see section III C)

183

184 Various models of continental crust thinning and associated mantle exhumation have been
185 proposed recently to account for geological constraints collected inside the metamorphic
186 NPZ. In figure 3, we present a selection of reconstructions extracted from recent literature,
187 which highlights numerous similarities between recently published models of Cretaceous
188 NPZ basins structure (Lagabriele et al., 2010; Clerc and Lagabriele, 2014; Masini et al., 2014;
189 Tugend et al., 2014; 2015; Clerc et al., 2016; Teixell et al., 2016, 2018; Corre et al., 2016;
190 Lagabriele et al., 2016; DeFelipe et al., 2017; Pedrera et al., 2017; Espurt et al., 2019;
191 Saspiturry et al., 2019; Asti et al., 2019; Ducoux et al., 2019; Odlum and Stockly, 2019). Most
192 of these architecture models stress the role played by a major cover décollement layer

193 during the Cretaceous crustal thinning. This weak layer corresponds to the Upper Triassic
194 Keuper evaporites, which contain clays and sands as well as minor carbonates and doleritic
195 MORB basalts (ophites). In the following, this décollement layer is referred to as the pre-rift
196 salt and clay unit. Its maximum thickness in the Pyrenean realm reached 2.7 km, as deduced
197 from field data in the southern Pyrenees coupled to well data in the Mauléon and Aquitaine
198 basins and the Bay of Biscay region (James & Canérot, 1988; McClay et al., 2004; Biteau et
199 al., 2006; Jammes et al., 2010a; 2010b; 2010c; Roca et al., 2011; Saura et al., 2016; Orti et
200 al., 2017; Saspiturry et al., 2019). In the décollement layer now exposed in the metamorphic
201 NPZ, Triassic clays were transformed into talc and chlorite-bearing rocks, and carbonates
202 often suffered intense tectonic brecciation involving talc, tremolite and dolomite
203 recrystallizations (Thiébaud et al., 1992; Lagabriele et al., 2019a, 2019b). Pre-rift to syn-rift
204 salt diapirism was also frequently observed in the non-metamorphic NPZ and in the
205 Southern Pyrenees (e.g. Canérot, 1988; 1989; Lenoble and Canérot, 1992; Canérot and
206 Lenoble, 1989; 1993; James and Canérot, 1999; Canérot et al., 2005; Jammes et al., 2009;
207 Jammes et al., 2010a; 2010b; Roca et al., 2011; Saura et al., 2016; Teixell et al., 2016).

208 As previously stated by Clerc and Lagabriele, (2014), the main consequence of the presence
209 of the pre-rift salt and clay unit along the north Iberia margin promoted mechanical
210 decoupling between the pre-rift Mesozoic cover and the Paleozoic basement during
211 Cretaceous rifting. As a result the pre-rift Mesozoic cover behaved as an allochthonous unit
212 which remained in the center of the developing rift while the underlying basement and
213 lithospheric mantle were accommodating stretching. Stretching amounts in the basement
214 and the pre-rift cover remains difficult to accurately estimate. This is mainly due to ductile
215 deformation of both the pre-rift Keuper layer and Mesozoic carbonates in the center of the
216 basin.

217 It must be noted that in the external parts of the Pyrenean rift, the borders of the subsiding
218 Cretaceous flysch basins remain at low temperature and display classical faulted and tilted
219 blocks (e.g. half-grabens of Quillan basin, Camarade basin, Gensac-Bonrepos basin, western
220 border of the Mauléon basin, edges of the Gran Rieu high and Lacq basin) (Debroas, 1978;
221 1990; Biteau et al., 2006; Lagabrielle et al., 2010; Masini et al., 2014; Grool et al., 2018;
222 Espurt et al., 2019).

223 To the west, the NPZ rift ends up against the Cinco Villas Paleozoic massif (CV in fig. 1) and
224 passes into the Basque-Cantabrian basin which develops along the current northern Iberia
225 Peninsula and terminates in turn at the Le Danois Bank (DB in fig. 1). Upper Jurassic-
226 Cretaceous sediments (12.5 km maximum thickness) with interlayered Aptian to Santonian
227 basic volcanic rocks (Azambre and Rossy, 1976; Rat et al., 1983; Rat, 1988; Castañares et al.,
228 2001; García-Mondéjar et al., 1996; 2004; Floquet, 2004) (fig. 3f-h). This basin was floored by
229 an extremely thinned lithosphere in its central parts (Biscay Synclinorium and Nappes des
230 Marbres) and was also affected by a Late Cretaceous thermal metamorphism (Golberg and
231 Leyreloup, 1990; Cuevas and Tubía, 1999; Pedrera et al., 2017). A peridotite outcrop is
232 exposed along the major Leiza fault (LF in fig. 1), representing the hyperthinned central
233 portion of the Basque-Cantabrian basin. This confirms that crustal thinning led to the
234 exhumation of the upper mantle close to the basin floor (Mendia and Gil-Ibarguchi, 1991;
235 DeFelipe et al., 2017). The basin architecture deduced from field investigations in the
236 eastern part of the Basque-Cantabrian basin (the “Nappe des Marbres” area) includes
237 smooth-slopes margins with normal faults and tilted blocks restricted to the external
238 domains (DeFelipe et al., 2017; Pedrera et al., 2017; Ducoux et al., 2019). These
239 reconstructed geometries bear affinities with basin architectures deduced from geological
240 observations in the NPZ (fig. 3f-h). Indeed, such architecture and the overall evolution

241 deduced for this rift system imply gliding of the pre-rift sequence over its basement during
242 crustal extension with ductile crustal thinning in its central part in a way similar to models
243 proposed for the NPZ (e.g. Clerc and Lagabrielle, 2014; Corre et al., 2016; Teixell et al.,
244 2016). The Leiza detachment system of DeFelipe et al. (2017) (fig. 3g) coincides with the pre-
245 rift salt and clay unit and promotes the allochthonous character of the overlying pre-rift
246 units. The presence of a high-density body beneath the Basque-Cantabrian basin has been
247 first interpreted as the result of intrusions of mafic magmatic rocks (Pedreira et al., 2007).
248 Using lithospheric-scale gravity inversion, this anomaly has now been interpreted as a high-
249 density mantle body emplaced at shallow level during the Cretaceous extensional event
250 (Pedrera et al., 2017). The association of this exhumed mantle body with rift and post-rift
251 structural geometries suggests the activation of a major south-dipping ramp-flat-ramp
252 extensional detachment between Valanginian and early Cenomanian times with horizontal
253 extension of ~48 km. Interpretation of geophysical data shows that low-strength pre-rift salt
254 and clay unit favored decoupling of the cover and triggered the formation of mini-basins,
255 rollovers, and diapirs (Pedrera et al., 2017).

256

257 Finally, the presence of a thick pre-rift salt layer underlying the Mesozoic carbonates
258 appears as an ubiquitous first-order parameter when reconstructing the evolution of the
259 Cantabrian-Pyrenean range. Recent models of rift development at the northern Iberia
260 margin show that Triassic lithology controls the three intrinsic characteristics of the
261 Pyrenean rifting, which can be summarized as follows:

- 262 i. Tectonic juxtaposition of exhumed peridotites and pre-rift sediments. This occurs
263 when the lateral extraction of the thinned continental crust is completed. In
264 response to plate separation, the stretched crust is removed horizontally from the

center of the rift and decoupling of the pre-rift cover from its basement occurs along the pre-rift salt and clay décollement. As a consequence a tectonic contact is established between the decoupled pre-rift sediments and the uplifted sub-continental mantle (Clerc and Lagabriele, 2014) (fig. 3e). In some locations, due to subsequent removal of the pre-rift cover, mantle rocks have been exposed to the seafloor as observed around the Lherz, Urdach and Bestiac Iherzolite bodies (Lagabriele et al. 2010; 2016; de Saint Blanquat et al., 2016).

- ii. Crustal stretching under dominantly ductile conditions. The geometry of the crustal units located in the distal domain of the rift margins does not correspond to that of extensional allochthons (succession of isolated, triangular-shaped undeformed blocks) as described along the Iberia-Newfoundland conjugate margins and along the reconstructed alpine paleo-margins (Manatschal, 2001; Manatschal et al., 2001; 2006; Peron-Pinvidic and Manatschal, 2009; Mohn et al., 2010; 2012; 2015) (fig. 2). In contrast, the continental material appears as an assemblage of very thin lenses (10-50 m thickness) consisting of ductilely deformed pre-Mesozoic basement rocks, mainly originating from the middle crust. These lenses are separated by anastomosed shear zones, which formed in greenschist facies conditions at low pressure (e.g. Corre et al., 2016; Teixell et al., 2016; Asti et al., 2019; Espurt et al., 2019) (fig. 3b-d). Detailed study of crustal material welded on the Urdach Iherzolites suggests that the middle crust was extracted laterally from the rift axis and that deformation occurred at temperatures ranging between 450°C and 350°C (Asti et al., 2019). Large strains in the greenschist facies affect ortho- and para-derived mylonites. They are testified by strongly elongate quartz ribbons with bulging recrystallization along with brittle fracturing of feldspar in cataclastic flows (fig. 4a-b).

In general, high temperature is promoted by the presence of the overlying allochthonous pre-rift cover, which mostly preserves its original thickness during the initial stages of rifting.

- iii. Dominantly ductile deformation of the pre-rift and syn-rift sediments under HT-LP conditions. Throughout the rifting phase, the decoupled pre-rift cover remains in the center of the rift. At this location, the temperatures rises in reponse to the ascent of the underlying mantle. The local preservation of an overlying pre-rift cover and the deposition of syn-rift deposits contributes to the high temperatures recorded regionally. As a result, the detached pre-rift cover locally undergoes syn-metamorphic ductile thinning and boudinage during continental breakup (fig. 5a-d). Such mechanical behaviour is outlined in all published rifting models (i.e. base of Nappe des Marbres basins, Leiza detachment system, base of Mauléon and Chaînons Béarnais basin infills, base of Baronnies and Boucheville basins infill: see references in fig. 3). Progressive rifting triggers the upward propagation of the brittle-ductile transition which may reach syn-rift sediments deposited at the early stage of the basin opening (Clerc et al., 2016). Brittle deformation dominated by cataclastic brecciation overprints ductile shearing and flattening in the sedimentary units accompanying final exposure of mantle rocks to the seafloor, as proposed from studies in the Lherz area (Lagabrielle et al., 2016). The ductile-brittle transition is frequently observed at the mesoscopic and microscopic scale as sets of normal faults offsetting the extensional HT foliation (fig. 5e-f, 5h). Finally, at the scale of the entire rift, extensional deformation in the lower margin is accompanied by tectonic

denudation of the cover in the upper margin (Lagabrielle et al., 2010; Teixell et al., 2016, 2018; Saspitury et al., 2019).

To summarize, figure 6 presents the intrinsic characteristics of the Pyrenean rifting listed above, compiled along an idealized column of the NPZ lithologies with photographs illustrating the most relevant deformed levels exposed along the NPZ.

II. A review of smooth-slopes basins in the surroundings of the Pyrenees-Cantabrian range

Seismic images of oceanic margins and intracontinental rifts in the proximity of the Pyrenees and Cantabrian ranges bear crucial information on the mode of crustal thinning along the northern Iberia margin and adjacent areas during the Cretaceous period.

(1) Parentis basin. First interpretations of the Parentis ECORS profile point to a symmetrical, syncline-shaped basin, with only few normal faults in the stretched crust, even in the proximal domain (Pinet et al., 1987; Bois et al., 1997; Jammes et al. 2009) (fig. 1 and 7a).

Beneath the Parentis basin fill, the crust is less than 10 km thick and decreases westward from 7 km (along the ECORS Bay of Biscay profile, fig. 1), to 6–5 km (along the MARCONI 3 profile, fig. 1) (Tomassino and Marillier, 1997; Gallart et al., 2004; Ruiz, 2007). More recently, Jammes et al. (2010a), proposed that the southern Parentis basin represents a lower plate sag basin floored by a top-basement detachment system with an asymmetrical mode of opening. These authors emphasize the presence of a thick pre-rift salt layer in the area undergoing extreme crustal thinning, forcing sub- and suprasalt layers to deform differently. Whatever the processes of crustal thinning are favored, both older and recent models of

336 Parentis basin evolution highlight three major features: (1) the occurrence of symmetrical
337 smooth-slopes gently dipping basinward; (2) the presence of a crust thinning regularly
338 towards the basin axis, without any observed steeply dipping faults, and (3) the presence of
339 a thick pre-rift salt layer allowing décollement of the pre-rift cover from its basement
340 (Jammes et al., 2010b, 2010c).

341 (2) South Bay of Biscay margin. Both the northern and southern margins of the Bay of Biscay
342 have been explored seismically. North-south transects of the Armorican margin (Norgasis
343 profiles, fig. 1: Thinon et al., 2003; Tugend et al., 2014) reveal a short necking domain that
344 concentrates most of the crustal deformation. Crustal thickness decreases from 35 km at the
345 shelf break to less than 10 km at the foot of the slope. Steep rise of mantle implies the
346 disappearance of the lower crust beneath the slope. Based on results of gravity inversion
347 combined with seismic interpretations, Tugend et al. (2014) map a continuous domain of
348 exhumed mantle from the Armorican basin toward the hyperthinned Parentis basin where
349 minimum crustal thickness occurs (fig. 7a) (Pinet et al., 1987, Bois et al., 1996, Jammes et al.,
350 2010a). According to Roca et al. (2011), the Bay of Biscay Abyssal Plain itself consists of a
351 transitional zone formed by a thin (4–9 km) crust with riders of Mesozoic pre-rift and syn-rift
352 sediments and continental crustal rocks that are extensionally detached over an exhumed
353 sub-continental mantle with seismic velocities comprised between 7.2 and 8 km/s. In the
354 distal domain, the Bay of Biscay Abyssal Plain bounds to the north the North Iberian margin,
355 an extended continental margin with Cretaceous basins (e.g. the Asturian basin, up to 10 km
356 thick, fig. 1) and basement highs such as Le Danois Bank (Cadenas and Fernández Viejo,
357 2016; Teixell et al., 2018), where granulites have been dredged (Capdevila et al., 1980;
358 Fügenschuh et al., 2003) (fig. 1).

359 (3) North-eastern Iberia intra-crustal basins (Iberian Chain and Valencia trough). Helpful
360 additional information regarding the thinning modes of the northern Iberia crust can be
361 obtained from seismic images of the Cameros, Maestrat and Columbrets basins now partly
362 inverted in the Iberian Chain (fig. 1 and fig. 7b-d). These basins result from the distributed
363 extension of the northern Iberia plate synchronously with the opening of the Bay of Biscay-
364 Pyrenees in the mid-Cretaceous (Verges and Garcia-Senz, 2001; Mas et al., 2011). They
365 represent a well-developed Mesozoic rift sharing similarities with the North Atlantic margins
366 (Salas and Casas, 1993; Salas et al., 2001). In their internal parts, reconstructed Iberian Chain
367 basin geometries point to simple troughs exhibiting gentle slopes devoid of marked fault
368 stepping, suggesting the absence of tilted blocks and a smooth basement top. The Moho
369 generally shows an arched outline with a progressive shallowing toward the basin center
370 where the crustal thickness is reduced to a few kilometers only (e.g. Guimerà et al., 1995;
371 Casas-Sainz and Gil-Imaz, 1998; Omodeo et al., 2014). Triassic evaporites play an important
372 role during the Albian rifting in the basins of northeast Iberia. This role was recently well
373 illustrated by interpretation of seismic reflection profiles in the Valencia trough (Etheve et
374 al., 2018) (fig. 7b). These profiles reveal the presence of a large Albian basin, the Columbrets
375 basin (fig. 1), filled with up to 10 km thick Mesozoic sediments over a highly extended
376 continental basement (locally only 3.5 km thick). The pre-rift and syn-rift successions form a
377 large-scale syncline with thinned borders. Whole Basin deformation is interpreted to result
378 from interaction between decoupling promoted by a pre-rift salt unit and dominantly ductile
379 thinning of the basement (Etheve et al. 2018) leading to the development of an abnormally
380 thin continental crust (Gallart et al., 1990; Dañobeitia et al., 1992; Ayala et al., 2015). In the
381 Cameros basin (fig. 7c-d), the pre-rift cover is decoupled from the basement along the
382 Keuper evaporites and is smeared all over the stretched domain. The syn-rift record does

383 not document any major offset of the top basement (Casas-Sainz and Gil-Imaz, 1998; Casas-
384 Sainz et al., 2000). In a similar fashion than in the NPZ, syn-thinning HT-LP metamorphism of
385 the **pre-** and syn-rift sediments is reported in the Cameros basin fill (Guiraud and Séguret,
386 1985; Goldberg et al., 1988; Rat et al., 2019).

387 **III. Evolution of smooth-slopes basins: some rheological considerations**

388

389 **A. Symmetrical smooth-slopes basins: simple or pure shear-dominated structures?**

390

391 A common characteristic of smooth-slopes basins is the lack of tilted crustal blocks and
392 related stepping fault scarps in their central part, thus defining a dominant symmetrical
393 smooth-slopes profile of basement tops (figs. 2 and 7). Based on field data from the NPZ, we
394 have shown that stretching of the crustal basement occurs in a dominant ductile mode
395 under greenschist facies conditions, since the central part of the basin remains overlain by a
396 permanent cover of detached pre- and syn-rift sediments (Asti et al., 2019) and underlain by
397 exhuming mantle.

398 The architecture of extended crustal systems depends on the geometrical and temporal
399 associations between simple shear and pure shear regimes. Following the pioneer works of
400 McKenzie (1978), Wernicke (1981, 1985), Lister and Davis (1989) and Lister et al. (1991), the
401 symmetry and asymmetry of lithosphere stretching processes have been largely debated
402 over the last 30 years (i.e. Buck et al., 1988; Allemand et al., 1989; Buck, 1991; Brun, 1999,
403 Huisman and Beaumont, 2003). In the same way, the symmetrical character of the final
404 architecture of passive margins following continental breakup has been discussed by many
405 authors (i.e. Reston et al., 1995; Michon and Merle, 2003; Huisman and Beaumont, 2007;

406 Reston and Perez-Gussinye, 2007; Sutra et al., 2013; Brune et al., 2014). It follows that
407 apparent architecture symmetry does not imply dominant pure shear thinning mechanisms
408 but may result from asymmetrical tectonic processes involving large-scale discrete
409 extensional shear zones (simple shear) as discussed by Brun and Beslier (1996) and Nagel
410 and Buck (2004) (fig. 8).

411 In their four-layer laboratory model, Brun and Beslier (1996) combine simple and pure shear
412 deformation in order to account for the exhumation of mantle rocks at ocean-continent
413 boundaries (fig. 8b). Model layers are composed of sand and silicone putty, regarded as
414 analogues of the brittle and ductile layers of both crust and mantle. The lower crust deforms
415 ductilely whereas the upper mantle remains strong. The Brun and Beslier (1996) model
416 shares characteristics with models derived from the geology of the Parentis and NPZ basins
417 which involve homogeneous thinning of the crust during its lateral extraction from the rift
418 axis (e.g. Corre et al., 2016; Teixell et al., 2016, 2018) (fig. 3). It does not favor the activation
419 of a restricted number of crustal detachment faults. Therefore, necking of the whole
420 lithosphere model appears nearly symmetrical (pure shear). However, asymmetrical
421 structures (simple shear) develop internally, due to boudinage and/or faulting of brittle
422 layers as deduced from field observations at the outcrop scale. The Brun and Beslier (1996)
423 model also accounts for the symmetrical ductile deformation of the NPZ bulk crust as shown
424 by Asti et al. (2019) and explains the occurrence of localized shear zones in the lithospheric
425 mantle as described by Vissers et al. (1995) in the Pyrenean mantle.

426

427 **B. Smooth-slopes basins: crustal shear zones and lenticular fabrics at the mesoscale.**

428

429 Petrological studies of continental units exposed around the Urdach and Saraillé lherzolite

bodies (western NPZ) provide information on the deformation mode associated with crustal thinning and mantle exhumation of northern Iberia (Corre et al., 2016; Asti et al., 2019). Reconstruction of sections across the NPZ Cretaceous basins by Clerc et al. (2015b), Teixell et al. (2016), Corre et al. (2016) and Asti et al. (2019) highlight the role of high-temperature ductile deformation of the basement induced by regional-scale, pure shear extension. It was shown that extension in the Paleozoic basement was achieved through lenticular deformation and pervasive ductile flattening of the crust with anastomosing extensional mylonitic shear zones developing at temperatures of 350-450°C. Stretching led to the development of undulated shear contacts separating tectonic lenses of crustal rocks (phacoid-shaped) as described in figure 10. At the final step of continental breakup, very thin continental crustal lenses (10-50 m thickness) remained welded to the exhumed mantle.

A similar lenticular mode of deformation derives from investigations in the Basin and Range province. Hamilton (1987) describes tectonic lenses of middle crustal rocks that normally lie at separate levels in the crust separated by undulating shear contacts (fig. 8c). This deformation mode allows the juxtaposition of different lithologies by uplifting deeper lenses during extensional deformation. In a different way, Gartrell (1997) propose a large-scale crustal boudinage involving successive necking regions where the ductile middle crust is extremely sheared (fig. 8d). The resulting architecture is a succession of tectonic lenses that may evolve toward a large-scale lenticular geometry as proposed by Espurt et al. (2019) for the evolution of the North Pyrenean massifs (fig. 3d).

In their recent detailed study of the tectonic and metamorphic evolution of the Urdach and Sarailé mantle bodies and associated units, Lagabrielle et al. (2019a; 2019b) describe two types of low-angle shear zones that accommodated part of extension of the distal domain of

454 the Iberia passive margin during the mid-Cretaceous (fig. 10a, b). The deepest shear zone is
455 the crust-mantle detachment. It separates the ultramafic mantle rocks from strongly thinned
456 basement Paleozoic rocks. It is composed of a basal 20-50 m thick lenticular layer of sheared
457 serpentinites overlain by a 10 m thick damage zone. The lenticular layer consists of
458 ultramafic symmetrical tectonic lenses (few meters long) separated by anastomosed
459 serpentine-rich shear zones. The damage zone consists of an assemblage of centimeter-sized
460 symmetrical lenses of a soft, talc-rich, sheared material, separated by conjugate shear zones.
461 The shallowest shear zone is the basement-cover décollement. It corresponds to a tectonic
462 boundary separating the base of detached pre-rift Mesozoic metasedimentary cover from
463 either mantle lherzolites or continental basement rocks. It consists of a thick deformation
464 zone (some meters to tens of meters) that was the locus of important metasomatic
465 crystallizations involving notably fluids of Triassic origin (Corre et al., 2016). Detailed
466 structural study of the basement and mantle rocks shows that it is not easy to discriminate
467 between dominant pure shear and dominant simple shear processes at the outcrop and
468 regional scales (Lagabrielle et al. 2019a; 2019b). Indeed, a major detachment fault zone
469 (typically related to regional simple shear) may contain abundant symmetrical lenses
470 suggesting locally dominant pure shear deformation.

471 Finally, in the studied smooth-slopes basins, dominant pure shear mechanisms concentrate
472 into the strongly thinned continental tectonic lenses whereas simple shear mechanism occur
473 along the main detachments. Pure shear deformation, causing the overall flattening of the
474 syn-rift and pre-rift sedimentary pile, progressively develop into the basin center as
475 represented in figure 10a. Chronological constraints need to be integrated in this proposed
476 temporal sequence in order to establish a possible succession from simple shear-dominated
477 to pure shear-dominated deformation at the scale of the entire system.

478

479 **C. Contrast between Iberian-Alpine and Pyrenean rifts: role of a basement-cover**
480 **décollement**

481

482 In contrast with the NPZ symmetrical model, recent models of margin evolution based on
483 the Iberian or Alpine examples have put forward asymmetric architectures resulting from
484 the development of few major detachment faults, and promoted the use of “lower-plate”
485 and “upper-plate” terminology (Manatschal, 2004; Mohn et al., 2010, 2012, 2015; Sutra et
486 al., 2013). Mohn et al. (2012) propose a model of three-layer continental crust where the
487 brittle upper and lower crusts are strongly decoupled by a ductile middle crust (fig. 3b).
488 Crustal thinning, accommodated through a so-called necking zone, is the result of the
489 interplay between detachment faulting in the brittle layers and decoupling in the ductile
490 quartzo-feldspatic mid-crustal levels along localized ductile décollements. The excision of
491 ductile mid-crustal layers and the progressive embrittlement of the crust by coupling the
492 lower and upper crusts enable major detachment faults to cut into the underlying mantle,
493 exhuming it to the seafloor.

494 By studying the evolution of the western Betics, including the exhumation of the Ronda
495 subcontinental mantle, Frasca et al. (2016) identify three successive steps: (i) ductile crust
496 thinning and ascent of subcontinental mantle thanks to mid-crustal shear zone and crust-
497 mantle shear zones acting synchronously; (ii) disappearance of the ductile crust bringing the
498 upper crust in contact with the subcontinental mantle, (iii) complete exhumation of the
499 mantle in the zone of localized stretching and high-angle normal faulting cutting through the
500 Moho, with related block tilting. These steps do not fully apply to the Pyrenean case, notably
501 because field and geophysical studies of the metamorphic NPZ never document the

502 occurrence of brittle faulting across the Moho during the Cretaceous rifting.

503

504 Finally, based on these examples, we may stress that both Alpine and Betic examples of
505 reconstructed passive margins do not involve the activation of a prominent décollement at
506 the base of the pre-rift cover. They rather promote evolutionary models with pre-rift
507 sediments welded to the basement blocks, in contrast with the examples detailed in section
508 I and II. In addition, both Alpine and Betic models refer to a progressive embrittlement of the
509 thinning crust resulting in the complete elision of ductile crustal layers. Again, this contrasts
510 with the NPZ examples where thin ductile crustal layers were extracted in the distal domain
511 and remained welded to the exhumed mantle.

512

513 **IV. Evolution of smooth-slopes basins : a comparison with current rift and margin models**

514

515 **A. Smooth-slopes basins formation, insights for the evolution of passive continental** 516 **margins.**

517

518 From the above section III, we interpret that dominant pure shear deformation is
519 responsible for the formation of the strongly stretched continental basement (anastomosed
520 tectonic lenses) in the central region of smooth-slopes basins. In the following, we review
521 examples of comparable uniform modes of ductile deformation in passive margins
522 worldwide.

523 A lenticular mode of deformation devoid of any steep normal fault was proposed at the
524 scale of an entire passive margin by Gernigon et al., (2014) to account for the symmetrical
525 stretching of the continental crust during the formation of the Barents margin (fig. 11a). This

526 lens-shaped lozenges geometry recalls the structures shown by Gartrell (1997) from
527 laboratory models (fig. 8d) and by Reston (1988) from lower crust reflection profiles.
528 Lenticular fabric was also suggested for deep crustal units connected to tilted blocks through
529 listric faults along the Norway margin (Osmundsen and Ebbing, 2008; fig. 11b). These
530 structures accommodate crustal thinning to only a few kilometer thicknesses through
531 dominant ductile deformation. Symmetrical stretching, implying ductile thinning or
532 boudinage of some crustal layers, can be compared to processes of depth-dependent
533 stretching or thinning (DDT and DDS) envisioned by Reston and McDermott (2014) in order
534 to account for extensional discrepancies at some passive margins.

535 Several distal domains of North Atlantic passive margins display geometries that suggest the
536 presence of lens-shaped units of thinned to hyper-thinned continental crust detached along
537 anastomosed shear zones and separated domains of exhumed mantle (e.g. Labrador and
538 West Greenland margins (Chalmers and Pulvertaft, 2001; Reston and Perez-Gussinyé, 2007)
539 (fig. 11c, d). These units do not resemble extensional allochthons of the West Iberia-type
540 margins (figs. 2 and 11e) and show geometrical affinities with crustal boudins extracted
541 during the Pyrenean extension in the center of the Cretaceous rift (e.g. the Baronnies and
542 Agly crustal boudins; Espurt et al., 2019; Clerc et al., 2016) (fig. 3). Such large areas of hyper-
543 thinned continental crust, composed of heterogeneous boudins, can be viewed as sheets
544 representing considerable volumes of sheared and flattened continental material (thickness
545 less than 10 km, width of 100 km and length of few 1000 km along the margin), formed
546 through processes of uniform pure shear at a crustal scale. We infer that the modes of
547 deformation exhibited by the Pyrenean crustal units welded to the exhumed mantle
548 (although at a much smaller scale) can apply to the formation of these crustal sheets made
549 up of greenschists facies mylonites. Similar crustal sheets underlying sag basins were

550 modeled in recent numerical simulations of margin evolution (Brune et al., 2014 ; Huismans
551 and Beaumont, 2011; 2014) as shown in figure 12a, b. Crustal sheets are present along the
552 Angola margin (fig. 12d), they may be present in the very distal domain of the Gulf of Lion
553 margin where they may originate by extraction of lower crustal material (Jolivet et al., 2017)
554 (fig. 11f). Similar long and thin sheets are typically imaged by Wang et al. (2016) at the base
555 of the reconstructed Iberia margin of the Mauléon basin, and by Roca et al. (2011) in their
556 reconstruction of the Iberia margin north of the Cantabrian coast (fig. 9d).

557 In their compilation of high-quality and deep penetration seismic profiles of several passive
558 margins (Uruguay, Southern Namibia, Gabon, South China Sea and Barents Sea), Clerc et al.
559 (2015a; 2018) suggest that the lower crust of some margins is weaker than assumed and
560 accommodates a large part of extension by ductile shearing (fig. 8e). Boudinage appears as a
561 recurrent deformation process accounting for the thinning of the continental crust at
562 variable scales. This observation led authors to an unorthodox vision of some types of
563 passive margins where: (i) the lower crust depicts heterogenous ductile deformation
564 patterns, (ii) boudinage and associated low-angle shear zones controls a large part of the
565 deformation, and (iii) normal faults often dip toward the continent. This study highlights a
566 crustal behavior dominated by boudinage and lenticulation, implying interplay between
567 ductile shear zones (boudin edges) and more competent crustal volumes (boudin cores). As
568 discussed above in section B, this deformation mode may apply to the thinned crustal levels
569 located in the axis of the Cretaceous Pyrenean rifts (Teixell et al. 2016, 2018; Asti et al.,
570 2019) (fig. 10) and is supported by recent numerical models of lithospheric rifting
571 incorporating macroscale anisotropy (Duretz et al., 2016).

572 In their interpretation of deep seismic profiles of the Gulf of Lion margin, Jolivet et al. (2015)
573 point to an intense stretching of the distal margin and reveal a 80 km-wide ocean-continent

574 transition zone that may consist of thin lower continental crust (the “Gulf of Lion
575 metamorphic core complex”) and exhumed mantle (fig. 11e). They infer an overall hot
576 geodynamic environment with a shallow lithosphere-asthenosphere boundary able to
577 thermally weaken the upper mantle and the lower crust enough to make them flow south-
578 eastward. In this example, the lower crust bears an important role. Such process is not fully
579 documented in the NPZ since evidence of lower crust exhumation during Cretaceous rifting
580 has not yet been reliably reported. Moreover, in sections of figures 3a, e, f, h and 7, the
581 lower crust is considered as a high-strength layer that does not deform ductilely but rather
582 tends to form large scale boudin, which remain at depth during rifting.

583

584 **B. Comparison with thermo-mechanical models of crustal hyper-extension.**

585

586 The examples discussed above lead us to emphasize the abundance of lenticular fabrics at
587 various scales reported from different studies in both the upper mantle and the crust. The
588 formation of lenticular fabrics, necking and lateral extraction during continental rifting have
589 been addressed in mechanical and thermo-mechanical numerical models (Duretz and
590 Schmalholz, 2015; Duretz et al., 2016). These models emphasize the role of a pre-existing
591 macroscopic mechanical anisotropy on the development of continental rifts. They illustrate
592 the interplay between necking and lateral extraction of strong layers along weak
593 décollements, thus defining a lenticular fabric and anastomosed shear zone networks at the
594 regional scale as envisioned in the NPZ case.

595 Models of metamorphic core complexes (MCCs) formation generally involve a thick and hot
596 continental crust (Brun and van den Driessche, 1994). This does not apply to the Pyrenean
597 case but constructive inputs can be expected from a confrontation with the rheological

598 parameters used for MCCs modeling. For instance, Tirel et al. (2008) use initial Moho
599 temperatures of 800°C or higher, with crustal thicknesses of 45 km or greater. This contrast
600 with the post-Variscan crust in the Pyrenees, whose thickness ranges between 30 and 20 km
601 (Teixell et al., 2018, and references therein) and Moho temperature was likely lower than
602 800°C. In the Tirel et al. (2008) experiment, exhumation process in metamorphic domes
603 result in the progressive development of a detachment zone. Because both the lower crust
604 and the mantle lithosphere have a low strength, no strain localization occurs in the mantle.
605 As a result, the Moho remains flat. With Moho temperatures lower than 800°C, the sub-
606 Moho mantle has higher strength, which can result in a Moho deflections and lithosphere-
607 scale necking. Such conditions (relatively cold mantle and thin crust) may have prevailed in
608 the Pyrenean region, explaining why the Pyrenean mantle reached the surface during rifting.
609 A former numerical model that applies to the formation of passive continental margins
610 suggests that the crust may also be thinned by permanent pure shear both at the proximal
611 and distal margin (Huisman and Beaumont, 2011) (fig. 12a). This scenario can be applied to
612 the Pyrenean case where the ductile behaviour of the middle crust is recognized (Asti et al.,
613 2019). The model of Huisman and Beaumont (2011) produced symmetric margins
614 associated with distal domain characterized by large sheets of thinned crustal material, as
615 discussed above. The symmetrical outline is well imaged by current reconstructions of the
616 Pyrenean basins from the North Pyrenean Zone and associated examples (Parentis, Cameros
617 and Columbrets basins, fig. 1, 2 and 7).

618 Brune et al. (2014) produced a model that emphasizes rift migration accomplished by
619 sequential upper crustal faults and balanced through lower crustal flow (fig. 12b). The
620 authors provided the concept of 'exhumation channel', which defines a region where the
621 crust and the uppermost mantle are actively deformed and extremely thinned during their

622 transfer from lower to shallower levels. This high strain volume is not formally a detachment
623 fault and may be at the origin of the crustal lenses exhumed within the NPZ (Asti et al.,
624 2019). As discussed in section C, the resulting geometry is a set of sheets of drastically
625 thinned crust forming the distal margin domain lying over a cooled and strengthened
626 mantle. This mantle may become exposed at the rift axis depending on the extension rate.
627 The final sketch derived from this model, a dome of strong mantle rimmed in its upper part
628 by a thin layer of mylonitic crust, is a reliable image for the geometry resulting from the
629 Pyrenean rifting and associated basins at a lithospheric scale.

630 Jammes et al. (2015) and Jammes and Lavier (2016), introduced compositional complexities
631 in the lithosphere by using an explicit bimineralic assemblage which results in the
632 development of anastomosing shear zone. In their models, the deformation appears
633 localized in the middle/lower crust and the upper lithospheric mantle and leads to the
634 preservation of almost undeformed lenses of material surrounded by localized shear zones
635 concentrating most of the deformation. Such a lenticular final geometry is also evocative of
636 the one observed in the North Pyrenean Zone as discussed in detail by Asti et al. (2019) and
637 illustrated in fig. 10.

638 To unravel the dynamic evolution of the Cretaceous Pyrenean rift, Duretz et al. (2019)
639 carried out a set of thermo-mechanical numerical models of lithosphere-scale extension
640 based on the available geological constraints listed above in section I. The models were used
641 to explore the role of a km-thick basement-cover décollement layer at the base of the pre-
642 rift sediments. These numerical experiments highlight the first order role of the décollement
643 layer that can alone explain collectively: (i) salt tectonics deformation style and cover
644 décollement, (ii) high temperature metamorphism of the pre-rift cover, and (iii) widespread
645 ductile deformation of the basement. In the axis of the synclinal-shaped basin ("sag" basin in

646 the margin literature), extreme pure shear leads to the development of a very thin
647 basement layer, overlain by slightly thinned pre-rift and syn-rift sediments and underlain by
648 exhuming mantle. These models are in good agreement with the current knowledge of the
649 architecture of the Cretaceous Pyrenean basins as exemplified by reconstructions of figs. 2
650 and 7, as well as with the presence of large sheets of hyper thinned crustal material (crustal
651 sheets) in the distal part of numerous magma poor passive margins.

652

653 **C. The pre-rift salt and clay décollement layer: a key factor for the evolution of smooth-**
654 **slopes basins. Establishing a new link between Triassic paleogeography and rifting**
655 **mechanisms.**

656

657 As reported in section I and II, the common character shared between all pre-rift sequences
658 of the aforementioned smooth-slopes basins is the presence of the thick pre-rift salt and
659 clay unit. All related geological and geophysical studies highlight the importance of this
660 décollement layer in the evolution of these rift basins. As detailed above, efficient
661 décollement along such décollement triggers mechanical decoupling and gliding of the pre-
662 rift cover that remains in the center of the basin while parts of the basement are laterally
663 extracted from the rift axis. In response to hyper-thinning and horizontal extraction of the
664 basement, hot mantle material approaches the detached pre-rift cover. As a consequence,
665 HT-LP metamorphism developed in the pre-rift sediments and at the base of the syn-rift
666 flysch levels as recorded in the NPZ and in the pre-rift sediments of the Cameros basin. The
667 presence of abundant alkaline volcanic products in the Cantabrian basin and in the NPZ also
668 argues for partial melting of the exhuming mantle. Subsequent deposition of syn-rift
669 sediments allows for maintaining high temperatures in the underlying units with major

670 consequences on the deformation style of the pre-rift sediments and basement.
671 Temperature increase in the center NPZ basins progressively led to the rise of the
672 brittle/ductile transition avoiding the development of prominent crustal normal faults and
673 leading to the dominantly ductile thinning of the Paleozoic basement and parts of the pre-
674 rift and syn-rift sediments (Clerc and Lagabriele, 2014; Clerc et al., 2015b; Asti et al., 2019;
675 Duretz et al., 2019).

676 We may now question the paleogeographic distribution of the Keuper group sediments at
677 the Europa-Iberia scale and evaluate the possibility of a genetic link between the occurrence
678 of a thick pre-rift décollement and the style of rifting. Several extensional systems interacted
679 in the Iberia platform during the Triassic, resulting in the creation of intraplate basins or
680 troughs including the Valencian, Basque-Cantabrian, and Pyrenean basins (figs. 1 and 13).
681 The sedimentary infill of these platform basins continued throughout the Mesozoic. Seismic,
682 well and field data from the Bay of Biscay region, the Pyrenees and the Aquitanian Basin,
683 suggest initial thickness of Upper Triassic formations ranging from 1000 to 2700 m (James
684 and Canérot, 1999; Biteau et al., 2006; Jammes et al., 2010a; Roca et al., 2011; Rowan, 2014;
685 Lopez-Mir et al., 2014; Saura et al., 2016; Soto et al., 2017; Zamora et al., 2017; López-
686 Gómez et al., 2019). The salt-rich layers generally consist of shales and evaporites including
687 dominant gypsum and minor halite and anhydrite (figs. 13 and 14). Paleogeographic
688 reconstructions are available for the Triassic period at the scale of the Iberia-western Europa
689 region (Dercourt et al., 1986; 1993; Ziegler, 1988; Ortí et al., 2017; Soto et al., 2017). This
690 paleogeography is confirmed by a compilation of data collected independently by D. Frizon
691 de Lamotte (fig. 13c). Evaporites are well developed along the eastern edge of Iberia (Tethys
692 side) and in the rifts located at the future NPZ, the Basque-Cantabrian basin, the Bay of
693 Biscay basin, the Asturian-Cantabrian Mountains and the southern part of the Armorican

694 margin. At the location of the future North Atlantic rift system, evaporites are restricted to
695 the Peniche, Lusitanian, Alentejo and Algarve basins along the southern half of the Portugal
696 margin. They are lacking along the northern half of the Iberia Atlantic margin. Along the
697 conjugate north-American margin, evaporites are known only at the base of the Jeanne-
698 d'Arc basin and are of restricted extent compared to the Keuper group exposed in Central
699 Europe (fig. 13b, c).

700 Finally, along the western half of the Iberia-Newfoundland transect, the presence of a
701 potential evaporitic formation is not reported. As outlined by figures 13 and 14, this
702 paleogeography matches the distribution of the two opposite types of basins discussed in
703 this article (smooth-slopes type vs. Iberia-Newfoundland type). Thus, we suggest a link
704 between the presence of a pre-rift salt layer and the style of rifting. Indeed, pre-rift salt and
705 clay décollement allows for the preservation of a 5-6 km thick allochthonous pre-rift cover
706 and overlying syn-rift sediments in the central part of the rift system. The exhumation of hot
707 mantle and the presence of a thick sedimentary pile lead to abnormally high temperature in
708 the thinning basement. It thus favors the occurrence of ductile deformation of the pre-rift
709 cover and the underlying basement as well as HT-LP metamorphism of the sedimentary
710 cover.

711 The lack of a major décollement level at the base of the pre-rift sequence may explain why
712 pre-rift sediments remained welded and coupled to the basement on the top of tilted blocks
713 in the Iberia-Newfoundland-type margins (figure 3a, b). Indeed, in the Iberia as well as in
714 Alpine margin-types, only syn-rift sediments are deposited over exhumed lower crustal
715 levels and subcontinental mantle (Péron-Pinvidic et al., 2007; Péron-Pinvidic and
716 Manatschal, 2009; Mohn et al., 2012), which contrasts with the evolution of the smooth-
717 slopes type basins. This is a type of pre-rift inheritance often considered in the history of

718 continental breakup during passive margins formation (i.e., Andersen et al., 2012; Chenin et
719 al., 2014; Asti et al., 2019) as well as during the formation of the Pyrenean rift (Munoz, 1992;
720 2002; Clerc and Lagabriele, 2014; Teixell et al., 2018; Asti et al., 2019). Recently, Fernández-
721 Lozano et al. (2019) conclude that differences in the nature of the lithosphere as a result of
722 the Cantabrian orocline formation may be also responsible for the distribution and shape of
723 the Cretaceous extensional basins north of the Iberia margin.

724

725 Based on examples clustering along the Iberia-Eurasia plate boundaries, we emphasize the
726 major role played by the Upper Triassic evaporitic layer during extensional processes. In the
727 reported smooth-slopes basin examples, cover gliding occurred on a pre-rift salt and clay
728 layer, which with cases involving syn- to post-rift salt tectonics. The latter cases have been
729 largely documented by studies of passive margins displaying thick syn-rift salt formations
730 such as the Angola margin where the post-salt sedimentary units have glided gravitationally
731 after margin formation (e.g. Brun and Fort, 2011; Jammes et al., 2010c and references
732 therein). To summarize, the specificities emphasized in this review are two-fold: (i) peri-
733 Pyrenean salt is pre-rift and allows for the conservation of the pre-rift cover over the high-
734 strain axial rift. Crustal faulting has not disrupted the continuity of the pre-rift salt and clay
735 unit. This décollement promoted the detachment of pre-rift sequences, which were then
736 preserved in the distal part of the margins. (ii) Consequently, the exhumation of hot mantle
737 underneath thick overlying sedimentary cover promoted the occurrence of high
738 temperature deformation and metamorphism. Dominant ductile deformation of the
739 basement deformation prevented the development of fault-related scarps and led to
740 formation of smooth-slopes basin edges.

741

742 **D. Time-dependent rheology during the evolution of smooth-slopes basins**

743

744 From the statements above (sections I, II, III), we stress that the models of NPZ Pyrenean
745 Cretaceous rifting differ significantly from the classical models of passive margin formation
746 based on the Iberia-Newfoundland margins example involving a dominantly brittle crust with
747 tilted fault blocks (fig. 3a) (Peron-Pinvidic and Manatschal, 2009; Sutra et al., 2013;
748 Osmundsen and Peron-Pinvidic, 2018, and references therein).

749 The peculiar evolution of the NPZ basins is depicted on figure 15 based on an original model
750 by Clerc et al., (2016). This model is conceptual and was designed to account for geological
751 constraints gathered from various sites along the NPZ. The conceptual model includes
752 subcontinental mantle, continental basement, a first decollement level in Triassic evaporites
753 and a cover of pre-rift carbonates and syn-rift flysch. Corresponding lithologies are
754 illustrated and briefly described in the NPZ lithostratigraphical column of figure 5.

755 In order to better assess the time-dependent rheological changes that necessarily affect
756 each geological layer involved during this three steps evolution, we provide synthetic
757 rheological profiles and geotherms for selected parts of the basin: in the external portion
758 representing the initial pre-rift model (fig. 15a) and in the center of the basin for the
759 following two steps (fig. 15b and c). The data used to construct these profiles are derived
760 from the reference model of Duretz et al. (2019), discussed in section D.

761 The three steps of this conceptual evolutionary model can be described as follows:

762 (1) At an early rifting stage (fig. 15a) moderate extension leads to crustal thinning
763 accommodated through normal faults in the upper crust. The rheological profile consists of a
764 15 km thick, cold and brittle upper crust ($T < 300^{\circ}\text{C}$) overlying a 15 km thick ductile lower

765 crust with Moho temperature around 550°C. The uppermost mantle is a strong 15 km thick
766 layer. In the inner part of the system, normal faults may change downward to ductile shear
767 zones dipping toward the external side thus delineating a small central horst. The pre-rift
768 salt and clay unit act as a décollement layer that allows the pre-rift carbonates to remain in
769 developing basins on both sides of the central horst while the syn-rift flysch is deposited.
770 Sliding of pre-rift carbonates towards distal domain leads to tectonic denudation of the
771 margins where carbonates remnants form isolated rafts.

772 (2) At the mid-rifting stage (fig. 15b), dominantly ductile basement thinning occurs in
773 response to stretching and heating due to mantle rise. Due to blanketing effect under the
774 syn-rift sediments, HT pre-rift carbonates undergo syn-metamorphic ductile deformation.
775 Rheological profile in the center of the basin underline the pre-rift salt and clay unit at the
776 base of the pre-rift cover and the basement that deforms at temperatures between 300°C
777 and 500°C. The lower crust has been extracted laterally and temperature increases from
778 400°C (step 1) to 1000°C at only 20 km depth in the strong lithospheric mantle.

779 (3) At the final rifting stage (fig. 15c), extreme thinning and boudinage of the crust leads to
780 local denudation of exhumed mantle, which is locally emplaced in tectonic contact with the
781 pre- or syn-rift sediments. The crust in the center of the basin has been split ~~cut~~ into few
782 lenses that move independently. The crust at both edges of the proximal domain is laterally
783 extracted from the basin axis (Clerc and Lagabriele, 2014). The pre-rift salt and clay
784 décollement layer undergoes thickness variations (e.g. diapirism) and is the locus of fluid-
785 assisted tectonic brecciation with metasomatic dissolution and crystallization as observed in
786 the Urdach and Sarailé massifs in the western NPZ (Lagabriele et al., 2019a, 2019b). HT
787 marbles of the pre-rift cover progressively accommodate a large part of the deformation at

788 the base of the basin, involving calcite creep and recrystallization, pinch-and-swell, drag
789 folding and low angle normal shear bands. In turn, the lower levels of the syn-rift flysch
790 sequence are progressively affected by HT metamorphism and ductile deformation induces
791 the development of bedding-parallel foliation and pinch-and-swell structures. Continuous
792 extension of the basin floor leads also to the progressive exhumation of the metamorphic
793 pre-rift sediments, originating from below the syn-rift cover (see complete description of
794 this process in Clerc et al., 2016). In the thinnest crustal portion, the rheological profile bears
795 similarities with that of step 2. The crustal thickness has now reduced to less than one km
796 and the brittle/ductile transition has moved upward. The pre-rift cover, pre-rift salt and clay
797 décollement as well as the thinned basement thus deform under dominant ductile
798 deformation.

799 **V. Conclusions**

800

801 This review highlights the affinities between the architecture of the smooth-slopes
802 extensional basins that opened during the mid-Cretaceous times around the Iberia-Eurasia
803 plate boundary, now variously inverted in the Pyrenean orogeny (Parentis, Basque-
804 Cantabrian, NPZ and Cameros to Columbrets basins). Taking the Parentis basin profile as a
805 reference and using geological reconstructions of NPZ rift architecture, we elaborate the
806 concept of smooth-slopes type basins shown in figure 16. The dominant features of this
807 conceptual model are the absence of prominent normal faults and large-scale tilted crustal
808 blocks in center of the basin. Smooth-slopes type basins involve the activation of
809 décollements (e.g. basin-cover décollement) that promotes sedimentary cover gliding and
810 lateral extraction of basement units. Basement and cover thinning occurs in a high
811 temperature regime and thus lead to the widespread development of ductile deformation

812 structures as well as metamorphism. The morphology of the margins of smooth-slopes
813 basins is rather symmetrical and thus drastically contrasts with that of Iberia-Newfoundland
814 type margins.

815

816 The common character between all pre-rift sequences of the studied smooth-slopes basins is
817 the presence of the thick low-strength Late Triassic evaporitic layer (Keuper facies). As
818 established by geological studies in the NPZ, efficient décollement along this pre-rift salt and
819 clay unit triggers mechanical decoupling and gliding of the pre-rift cover that remains in the
820 center of the basin as the crust is laterally extracted. Subsequent deposition of syn-rift
821 sediments allows for the preservation of the initial thermal anomaly with a major
822 consequence on the deformation regime in the pre-rift sediments and crustal basement. The
823 ubiquitous character of the ductilely deformed marbles in the metamorphic NPZ relates to a
824 dominant-ductile deformation regime in the pre-rift cover during the Cretaceous extension.
825 In these smooth-slopes basins, the ductilely stretched crust behaves homogeneously at the
826 regional scale and extensional allochthons are not recognized. A lenticular mode of
827 homogeneous deformation is thus defined implying interplay between hectometer-sized
828 lenses of ductile crustal material separated by anastomosed shear zones.

829

830 Both laboratory and thermo-mechanical numerical models reproduce remarkably the mode
831 of deformation deduced from geophysical and geological constraints. It appears that the
832 occurrence of a pre-rift salt and clay unit and the activation of syn-rift salt tectonics is a key
833 factor that affects the overall rifting style. This is in contrast with Atlantic margins where salt
834 is either syn-rift or post-rift. For the first time, we evidence a strong link between the
835 occurrence of a sedimentary layer covering the future rifted region (here Keuper salt and

836 clays deposits) and the style of rifting. Décollement along the pre-rift salt and clays units
837 finally favors the formation of smooth-slopes type basins. This new mode of crustal
838 deformation might not be restricted to the Pyrenean region, but may apply to other margins
839 that host thick pre-rift décollement series. This rifting mode may indeed have affected distal
840 portions of continental margins that are devoid of tilted blocks and extensional allochthons
841 and where large volumes of extremely thinned continental crust are present.

842

843 **Acknowledgements**

844 This work is the result of a 10 years field and laboratory research conducted in the Pyrenean
845 range with funding from various programs and institutions. We benefited of grants from the
846 French ANR Pyramid program, the Bureau de Recherche Géologique et Minière (BRGM),
847 Référentiel Géologique de la France (RGF) program and the Total/INSU/BRGM OROGEN
848 program with contributions from the CNRS and from Géosciences Montpellier and Rennes
849 research units. We thank Thierry Baudin (BRGM) and Sylvain Calassou (Total) for program
850 management and encouraging discussions. We also thank Th. Nalpas, B. Azambre, P.
851 Boulvais, M. Poujol, and many others for stimulating exchanges that helped improve our
852 ideas. We are indebted to one anonymous referee, to Maria Laura Balestrieri and to Gabriel
853 Gutiérrez-Alonso for their useful comments and remarks on a first version of this
854 manuscript. We also benefited of comments by Editor Carlo Doglioni who is acknowledged
855 here.

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 1528

1529 **Figure captions**

1530

1531 **Figure 1. Location of the studied basins and their paleogeographic position during the**
1532 **Cretaceous at the onset of the Iberia drift.**

1533 (a) Simplified structural map of the Cantabrian-Pyrenean orogenic system and adjoining
1534 Iberia showing Eurasia deformed and undeformed domain (modified from Verges and
1535 Garcia-Senz, 2001 and Teixell et al., 2018). (b) Hypothetical reconstruction at the onset of the
1536 Iberia drift (modified after Tugend et al., 2014).

1537

1538 **Figure 2. Structure and evolution of Iberia-Newfoundland-type and Alpine-type passive**
1539 **margins** (modified from Péron-Pinvidic and Manatschal, 2009 and Mohn et al., 2012).

1540 (a): two sketches showing the main concepts linked to Iberia-Newfoundland-type margin
1541 evolution, namely: (i) strong final asymmetry with upper and lower plates separated by a
1542 single detachment fault (HHD, Hobby High detachment), (ii) emplacement of extensional
1543 allochthons as rigid crustal blocks over the exhumed mantle. (b): strain distribution and
1544 strain partitioning during lithospheric thinning at magma-poor rifted margin, with example
1545 from the fossil Alpine Tethys margin. In this model, the pre-rift cover remains welded on the
1546 tilted crustal blocks; the middle crust is thinned to zero and the upper crust and upper
1547 mantle are juxtaposed at the break up stage.

1548

1549 **Figure 3. A compilation of Cretaceous basins architecture from the Cantabrian-Pyrenean**
1550 **belt.**

1551 Reconstructions from field and geophysical data collected by various authors in the Basque-
1552 Cantabrian basin (a, b, c) and in the North Pyrenean Zone (NPZ): Mauléon basin (d),
1553 Chainons Béarnais (e, f), Baronnies basin (g) and Agly massif-Boucheville basin (h).

1554

1555 **Figure 4. The geological record of the Cretaceous extension in the Paleozoic basement and**
1556 **exhumed mantle of the North Pyrenean Zone (NPZ).**

1557 The map shows the location of mantle bodies and crustal units illustrated in photographs a
1558 to k. (a): dated crustal mylonites associated with the Urdach Iherzolites; thin section
1559 microphotograph (natural light) of the leucocratic gneissic mylonite exposed at Col d'Urdach

1560 and containing numerous micafishes (dated by Ar/Ar method at 105 Ma; after Asti et al.,
1561 2019). (b): thin section of typical ultramylonite from lenses of Paleozoic material welded on
1562 the exhumed mantle rocks of the Saraillé lherzolite (Asti et al., 2019). (c): phacoidal fabric
1563 defined by anastomosed shear zones in the mantle body of Bestiac. This fabric is typical of
1564 the lenticular layer as defined by Lagabrielle et al. (2019a, 2019b). (d): phacoidal fabric in the
1565 lenticular layer of the lherzolite body of Moncaup. (e): phacoidal fabric in the lenticular layer
1566 of the lherzolite body of Saraillé (Lagabrielle et al., 2019b). (f): curved shear zones and
1567 elongated tectonic lenses in serpentinized lherzolites of the lenticular layer in the Moncaut
1568 peridotite body. (g and h): phacoidal fabric in the lenticular layer of the lherzolite body of
1569 Urdach: h shows pervasive carbonation (Lagabrielle et al., 2019a). (i and j): thin section and
1570 outcrop of anastomosed serpentinized shear bands in the lherzolites of Etang de Lers (Lherz).
1571 (k): anastomosed serpentinized shear bands in Avezac lherzolites.

1572

1573 **Figure 5. The geological record of the Cretaceous extension in the pre-rift cover of the**
1574 **metamorphic North Pyrenean Zone (NPZ). Some field view of outcrops showing the layer**
1575 **perpendicular flattening and the S0/S1 syn-metamorphic foliation.**

1576 (a): layer-parallel boudinage in the Calce quarry (Jurassic dolostones of the Agly massif
1577 cover, Eastern NPZ). (b): layer-parallel ductile stretching of the meta-laterite and carbonate
1578 breccia in the Benou quarry near Turon de la Tecouère lherzolite body (Chainons Béarnais,
1579 Western NPZ). (c): flattened fossils in Jurassic meta-dolostones of the Saleix valley (Aulus
1580 basin, Central NPZ). (d): extreme stretching of a rudist-rich Urgonian marbles at Sarrance
1581 (Chainons Béarnais, Western NPZ) (see also fig. 6c). (e): tight normal faults affecting the
1582 early S0/S1 syn-metamorphic foliation in pre-rift cover marbles of the Agly massif. These
1583 features characterize the ductile-brittle transition that occurred at the end of the rifting
1584 history. (f): same features as (e) but in marbles of the detached Lherz body cover (southern
1585 side). (g): recumbent folds associated with the early ductile foliation in marbles from the
1586 detached cover of the Pays de Sault Paleozoic basement (Eastern NPZ). (h): tectonic
1587 brecciation with calcite veining marking the ductile-brittle transition in the marbles of the
1588 Lherz body cover (western side).

1589

1590 **Figure 6 . A theoretical log of the lithological succession in the internal domain of the**
1591 **Cretaceous NPZ rift basins.**

1592 The photographs illustrate the various rock-types forming the basin basement (crust and
1593 mantle) and the pre-rift and syn-rift series. (a): Chaînons Béarnais (Saraillé massif, western
1594 NPZ). (b): Boucheville basin (eastern NPZ). (c): Urgonian at Sarrance (western NPZ) (see also
1595 fig. 5d). (d): Jurassic dolomites at Calce (eastern NPZ). (e): base of pre-rift series (Bestiac,
1596 eastern NPZ). (f): base of pre-rift series (Moncaup, central NPZ). (g): crustal lenses of Saraillé
1597 massif (western NPZ). (h): lenticular layer (Urdach mantle body, western NPZ).

1598

1599 **Figure 7. Interpreted and reconstructed profiles of peri-Pyrenean Cretaceous basins**
1600 **architecture.**

1601 (a): Parentis basin. (b): Columbrets basin. (c and d): Cameros basin. See location of basins in
1602 fig. 1.

1603

1604 **Figure 8. A compilation of model results and conceptual representations of extended to**
1605 **hyper-extended continental crust.**

1606 This compilation aims enhancing the main mechanical concepts involved in the processes of
1607 crustal extension and how they apply or not apply to the genesis and evolution of the
1608 smooth-slopes basins defined in this article (see text for discussion).

1609

1610 **Figure 9. A compilation of reconstructed architecture of Pyrenean Cretaceous basins and a**
1611 **Basque-Parentis transect.**

1612 All represented sections are based on the activation of a restricted number of detachment
1613 faults. As discussed in text, such representations do not match the newly defined smooth-
1614 slopes architecture that characterize the Pyrenean and peri-Pyrenean Cretaceous basins.

1615

1616 **Figure 10. Deformation regimes of the various units composing a typical smooth-slopes**
1617 **basin.**

1618 (a): distribution of pure shear and simple shear regimes in a simplified smooth-slopes basin
1619 system. (b): Detail of the very distal part of the hyper-extended crust (area shown in a). (b1):
1620 simplified log showing the association of metric to hectometric crustal lenses separated
1621 from the mantle rocks by the crust-mantle detachment and from the detached pre-rift cover
1622 by the cover décollement (see definition in Lagabrielle et al., 2019a, 2019b). (b2): field view
1623 of crustal sheets from the base of the Saraillé massif (western NPZ). (b3): field view of

1624 anastomosed shear zones cutting through the serpentized peridotite of the Sarailé body
1625 and forming the lenticular layer of the crust-mantle detachment (see also fig. 4c to k).

1626

1627 **Figure 11. A compilation of schematic architecture of selected Atlantic and Mediterranean**
1628 **passive margins.**

1629 These margin profiles are selected because they offer architectures which contrast with the
1630 Iberia-Newfoundland-type margin (see fig. 2). In particular, they show large scale crustal
1631 boudinage and lenticulation that are consistent with a ductile regime of extensional
1632 deformation. Sheets of hyper thinned crustal material is indicated by the orange arrow (see
1633 comments in text). Note that scale is similar in all profiles.

1634

1635 **Figure 12. Three numerical models of rift development compared to the Angola-Brazil and**
1636 **Iberia transects.**

1637 All models highlight a mode of deformation that leads to the development of very thin and
1638 long sheets of crustal material also observed in the Angola-Campos transect but not in the
1639 Iberia transect. Such deformation necessarily implies a dominantly ductile behaviour of the
1640 crust consistent with processes acting in the central part of the smooth-slopes basins
1641 studied in this paper (see text for further comments).

1642

1643 **Figure 13. Paleogeography of Triassic deposits and Cretaceous rifting around the Iberia**
1644 **plate.**

1645 (a): paleogeographic maps for the Triassic period (modified from Orti et al., 2017) and
1646 location of some further Cretaceous rifted regions. Note that by contrast to the area where
1647 Cretaceous smooth-slopes basins will open, the area corresponding to the future Iberia-
1648 Newfoundland conjugate margins are devoid of thick evaporitic series. (b): paleogeographic
1649 maps for the Ladinian and Carnian (Middle-early Late Triassic times, 242-227 Ma) modified
1650 after Scotese and Schettino (2017). (c): paleogeography of Upper Triassic deposits prepared
1651 after a compilation of unpublished data by D. Frizon de Lamotte (pers. com.) superimposed
1652 on a plate reconstruction by Olivet (1996).

1653

1654 **Figure 14. Correlation between the paleogeography of Triassic deposits and the mode of**

1655 **rifting around the Iberia plate.**

1656 (a): cartoons (a1 and a2) illustrating the contrasted rifting modes between the Iberia-
1657 Newfoundland-type and the Parentis-type margins (modified from Clerc and Lagabriele,
1658 2014). (b): paleogeography of Triassic (Late Norian) deposits according to Marcoux et al. in
1659 the Dercourt et al. (1993) map atlas. As paleogeographic maps in fig. 13, this reconstruction
1660 points to the lack of thick evaporites deposits in the future Iberia-Newfoundland rifting
1661 domain (see text for discussion).

1662

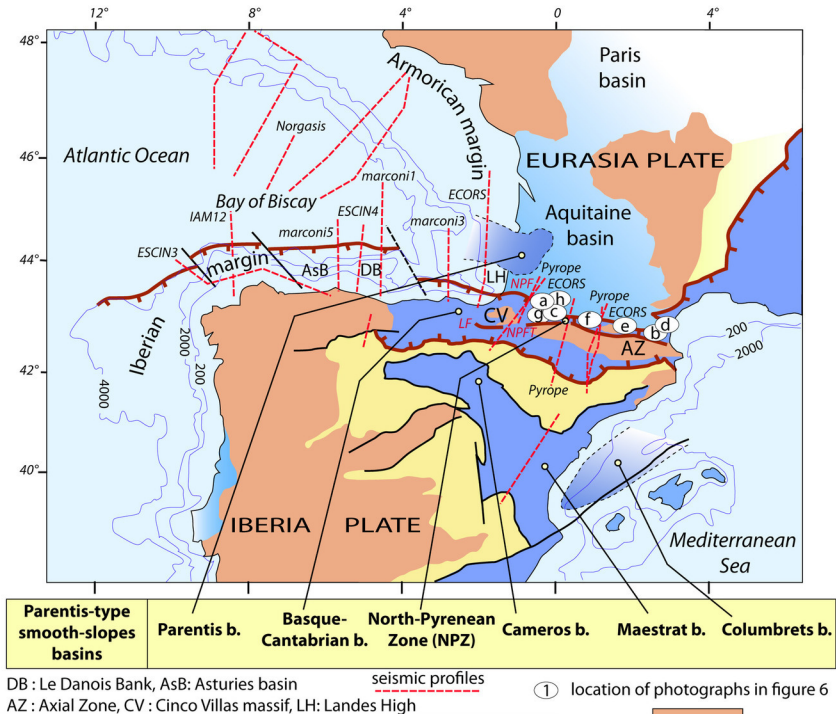
1663 **Figure 15. Time-dependent rheological evolution of the Pyrenean rifting based on**
1664 **geological constraints from the North Pyrenean Zone and numerical results from a thermo-**
1665 **mechanical numerical modeling.**

1666 Sketches depicting the geological evolution are extracted from the Clerc et al. (2016) model.
1667 Rheological profiles are derived from the Duretz et al., (2019) model. They are placed at
1668 critical locations (1, 2 and 3) of the rift in order to emphasize the drastic changes in the
1669 mechanical behavior during its evolution from limited crustal extension to local mantle
1670 exhumation (see detailed description in text).

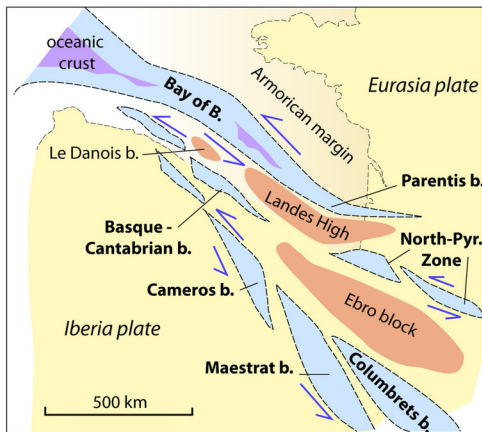
1671

1672 **Figure 16. A theoretical structural model for the Cantabrian, Pyrenean and Iberian**
1673 **symmetrical smooth-slopes basins based on the features and concepts discussed in this**
1674 **article** (see comments in the Conclusion section).

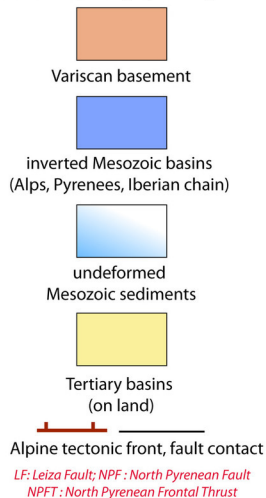
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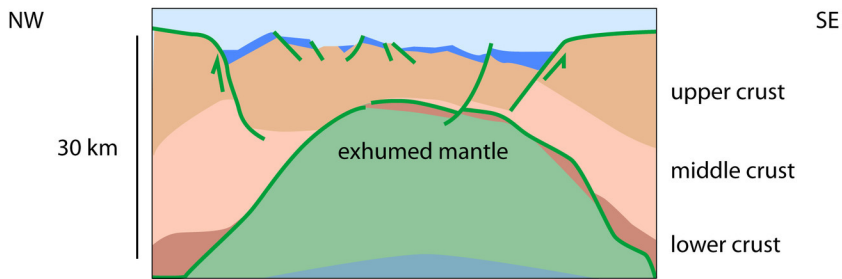
a



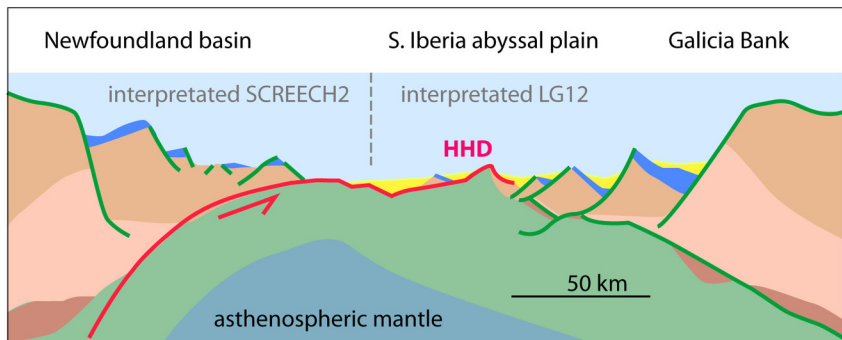
b



The Iberia-Newfoundland-, Alpine- types passive margin



145 Ma, end Tithonian



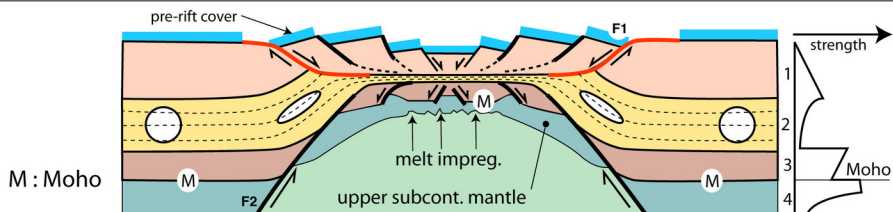
130 Ma, Hauterivian/Barremian

syn-rift sediments

pre-rift sediments

modified from Péron-Pinvidic and Manatschal (2009)

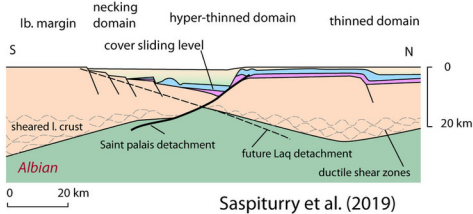
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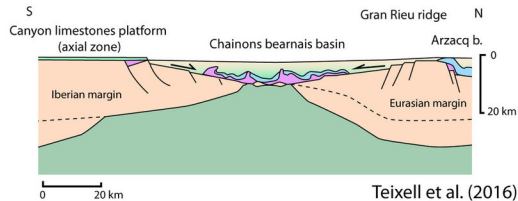
1: upper crust, 2: ductile middle crust, 3: strong lower mafic crust, 4: strong upper mantle

modified from Mohn et al. (2012)

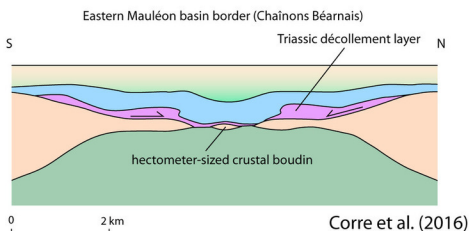
b



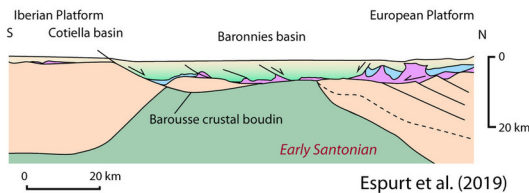
a. Western North Pyrenean Zone (1)



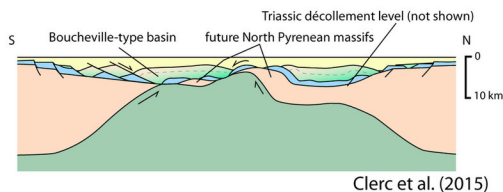
b. Western North Pyrenean Zone (2)



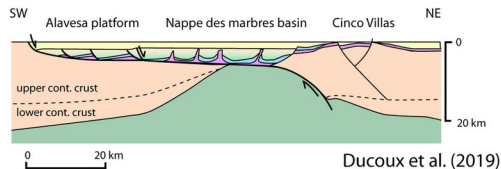
c. Western North Pyrenean Zone (3)



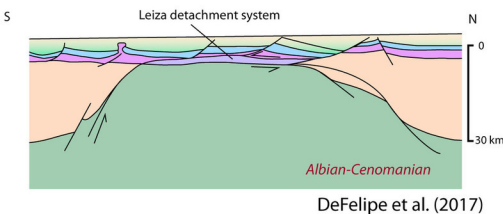
d. Central North Pyrenean Zone



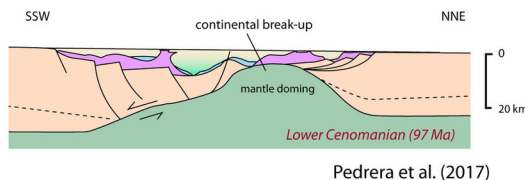
e. Eastern North Pyrenean Zone



f. Basque-Cantabrian basin (1)

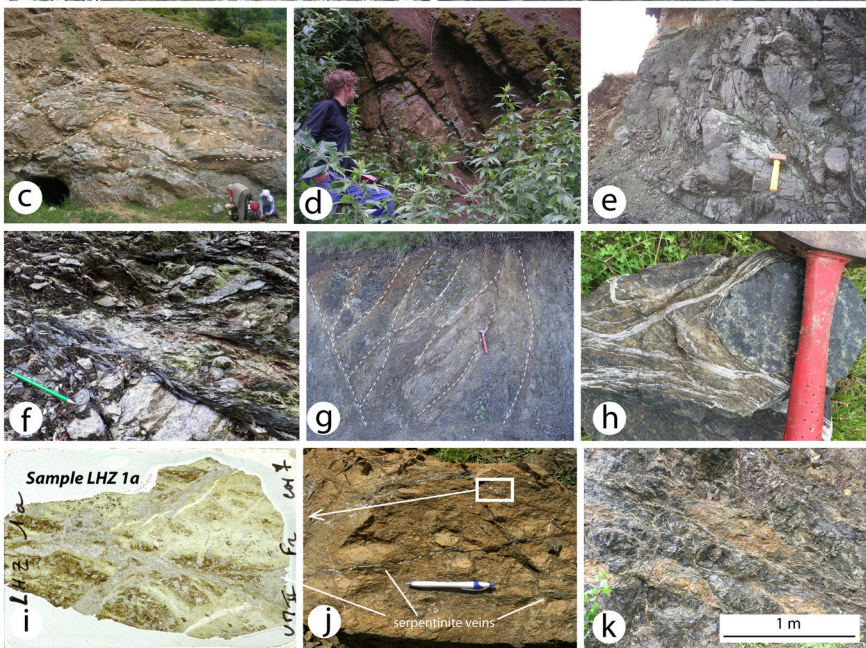
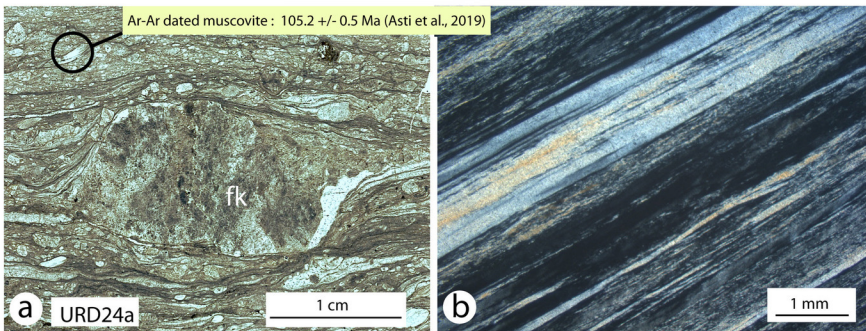
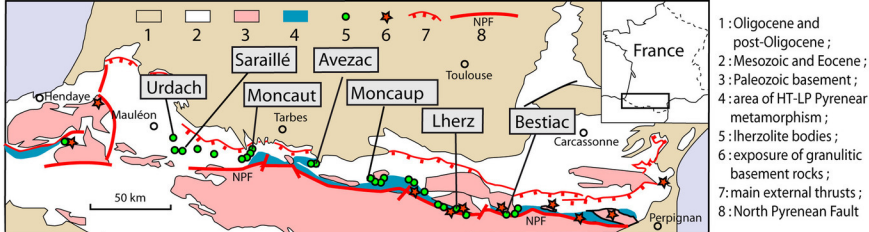


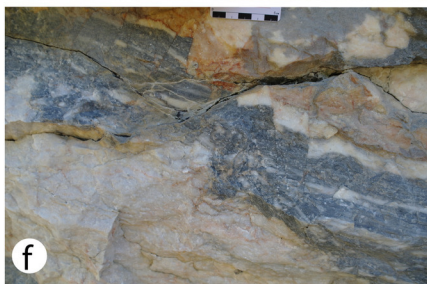
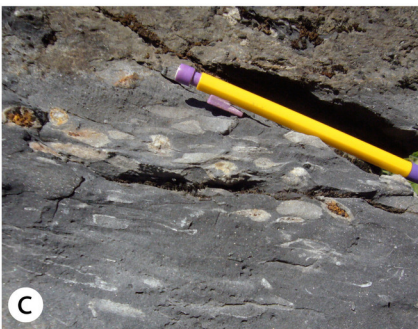
g. Basque-Cantabrian basin (2)

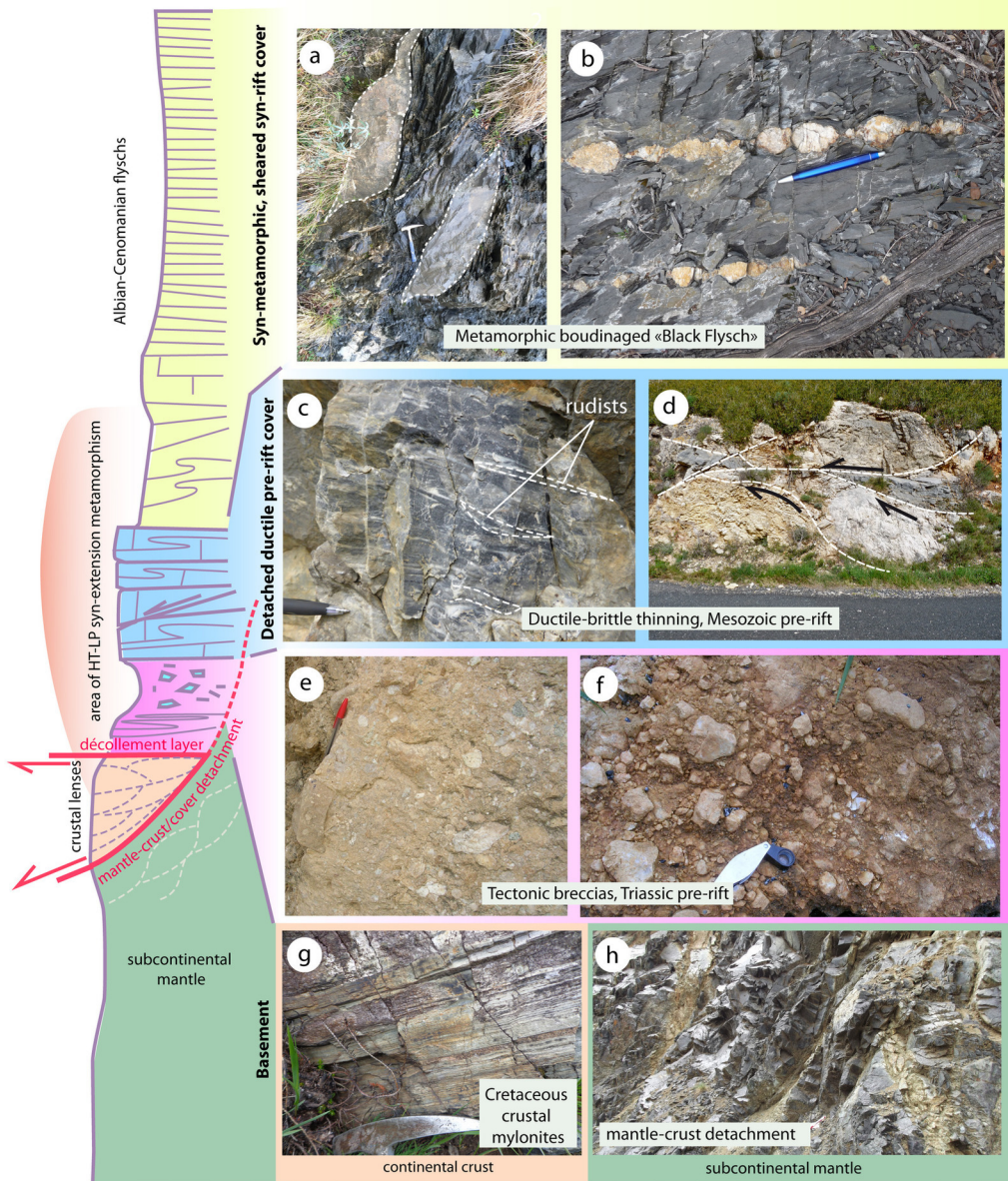


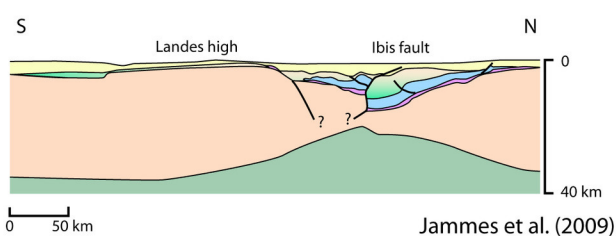
h. Basque-Cantabrian basin (3)



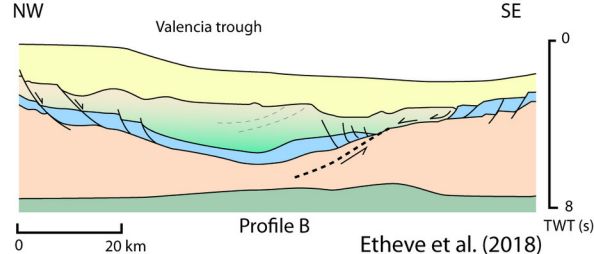




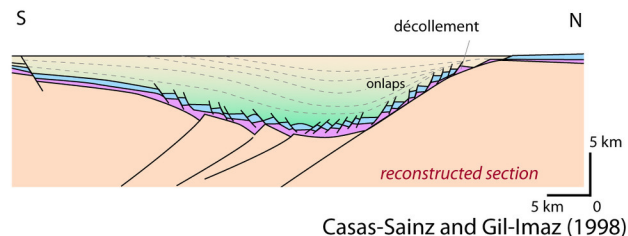




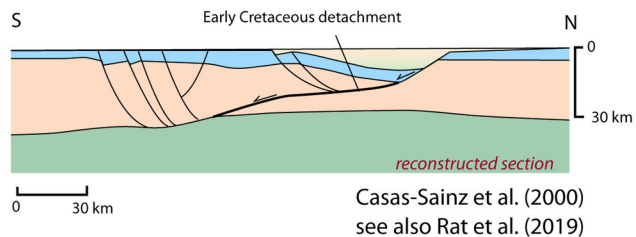
a. Parentis basin



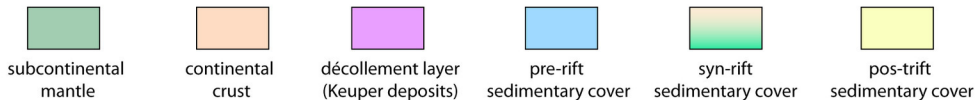
b. Columbrets basin

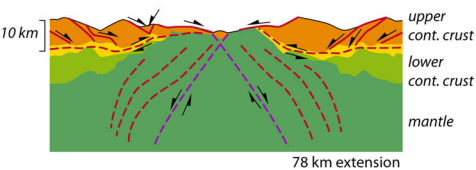


c. Cameros basin (1)

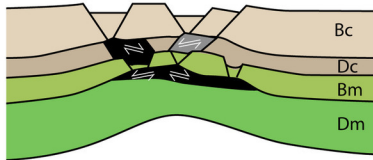


d. Cameros basin (2)





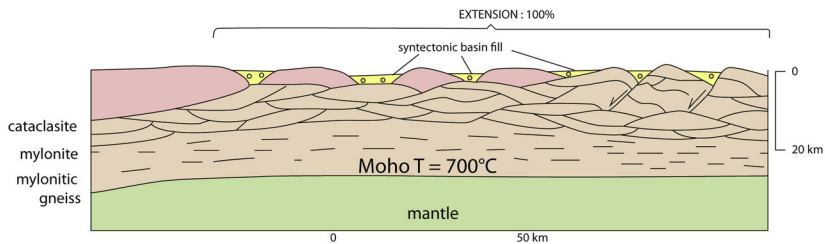
- brittle deformation (abandoned faults)
- - - ductile deformation (abandoned shear zones)
- - - ductile deformation (active shear zones)



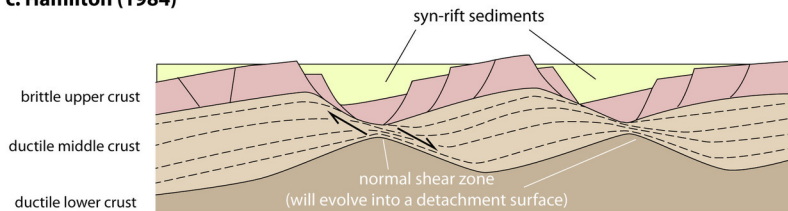
Bc: brittle crust Bm: brittle mantle
Dc: ductile crust Dm: ductile mantle

b. Brun and Beslier (1996)

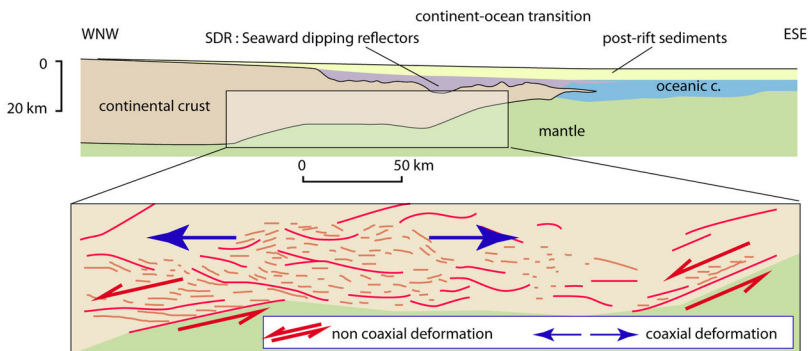
a. Nagel and Buck (2004)



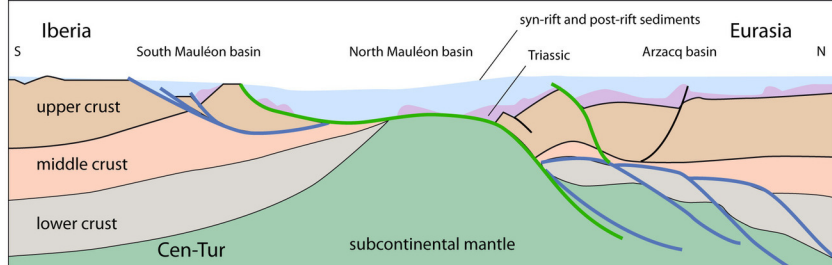
c. Hamilton (1984)



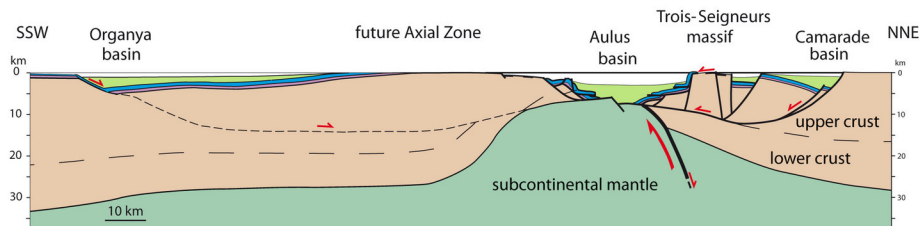
d. Gartrell (1997)



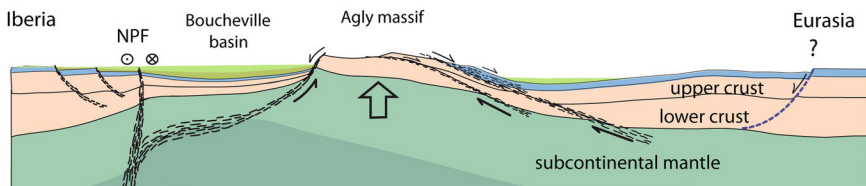
e. Clerc et al. (2018)



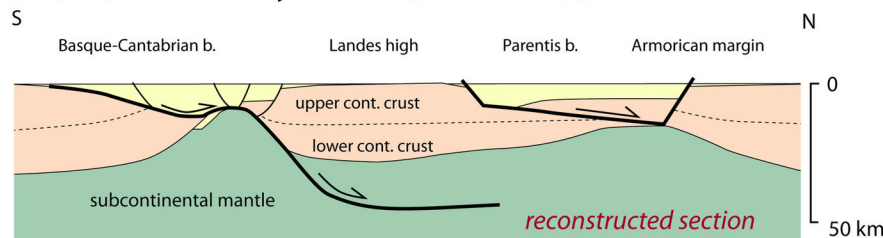
a. Masini et al. (2014): Mauléon basin



b. Lagabrielle et al. (2010): Central North Pyrenean Zone (Aulus basin, Etang de Lers)



c. Vauchez et al. (2013): Eastern North Pyrenean Zone (Boucheville basin)



d. Roca et al. (2011) : Basque-Parentis transect

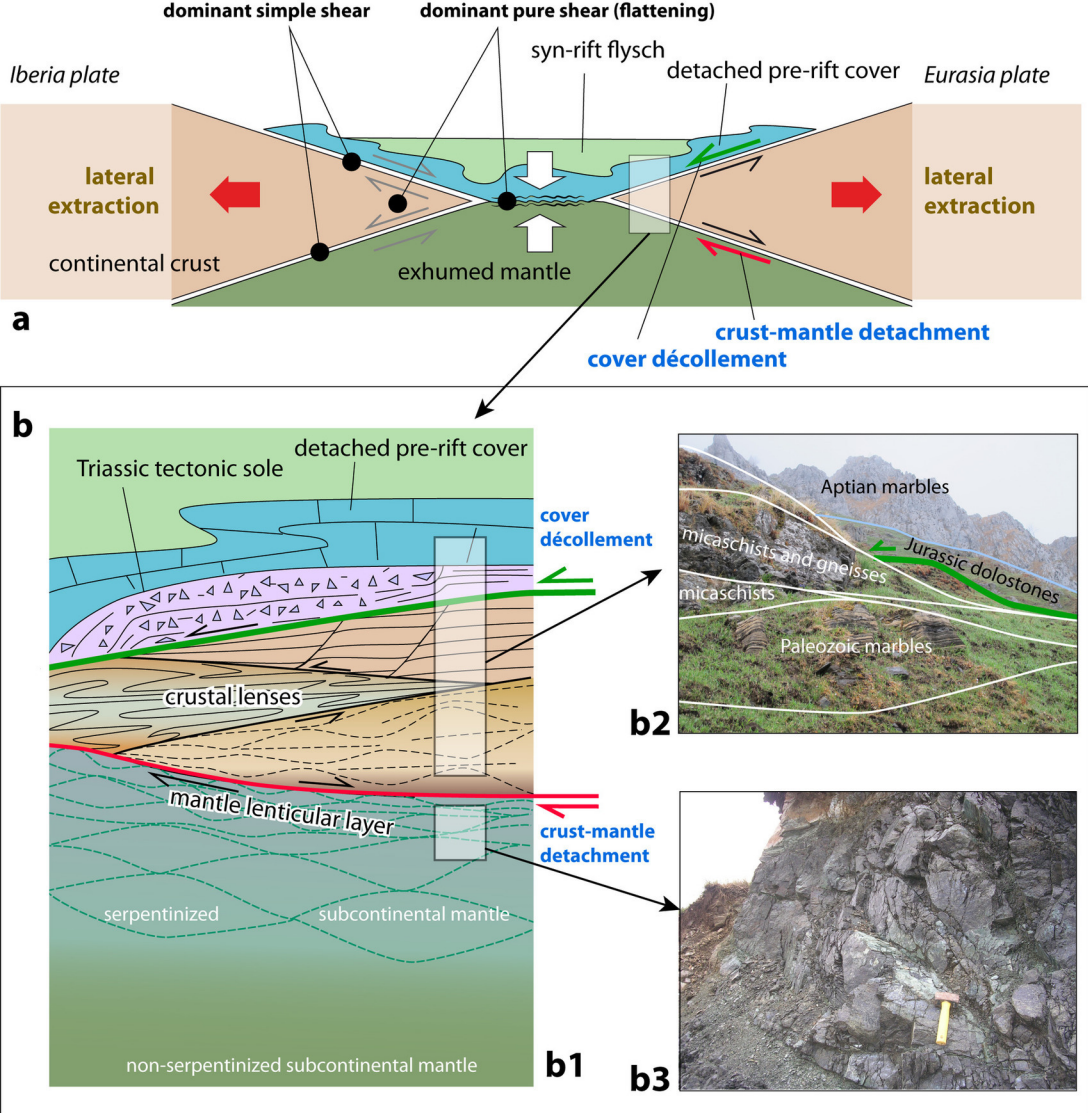
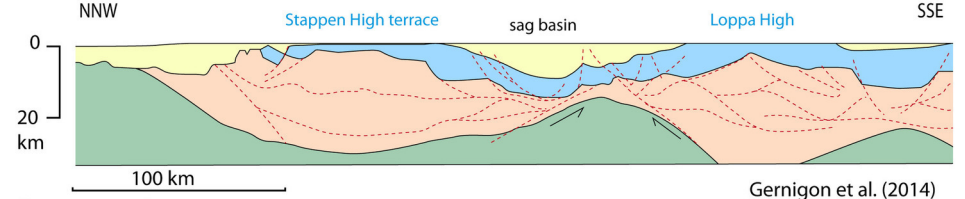
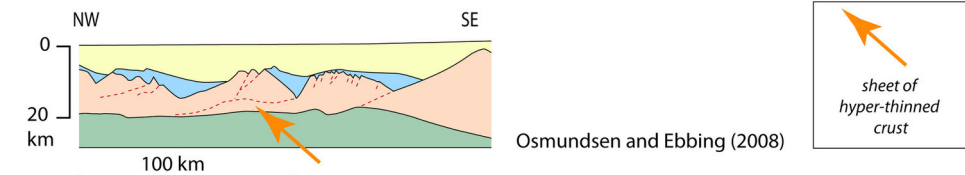


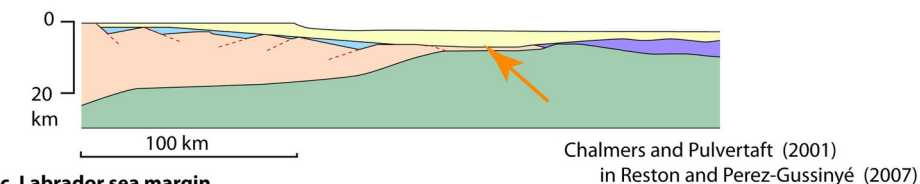
fig. 10



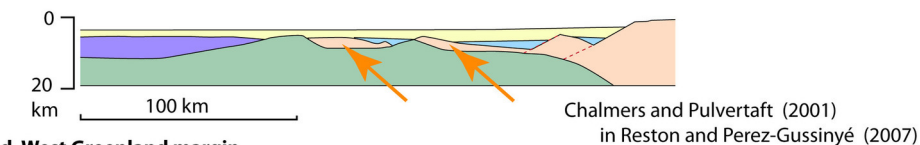
a. Barents margin



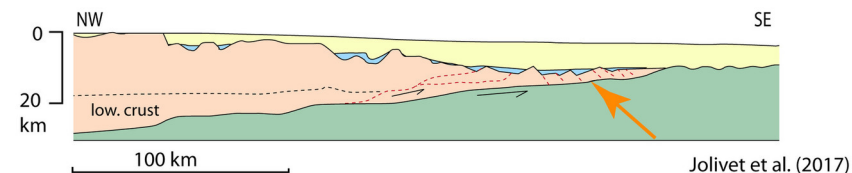
b. Northern More basin, Norway margin



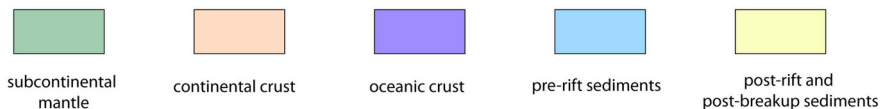
c. Labrador sea margin

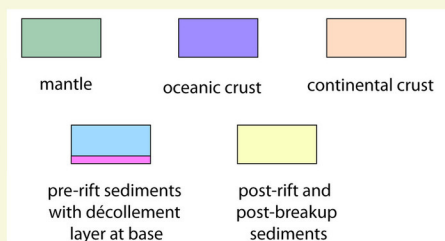
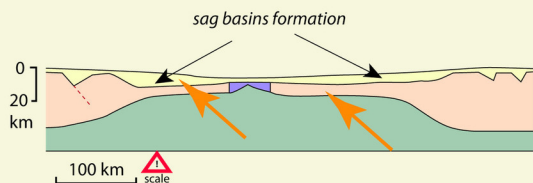


d. West Greenland margin

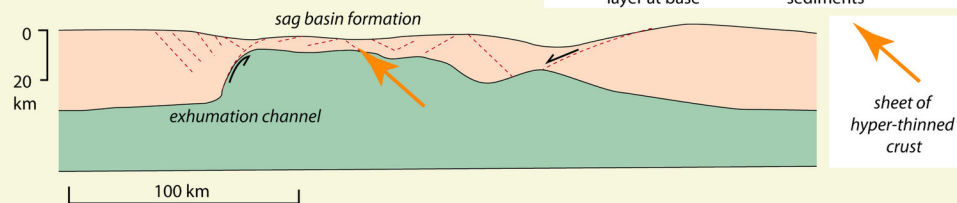


e. Western Mediterranean, Gulf of Lion

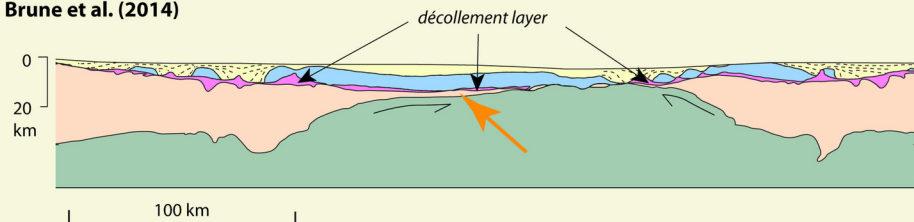




a. Type II Rifted Margins, Huismans and Beaumont (2011)

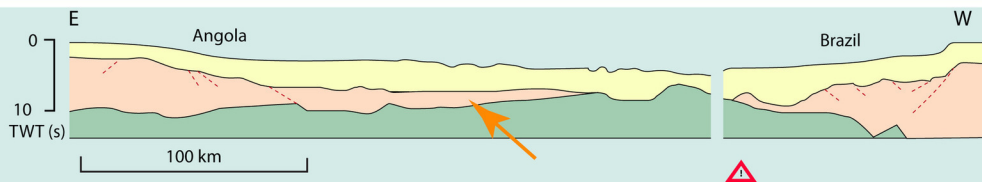


b. Brune et al. (2014)

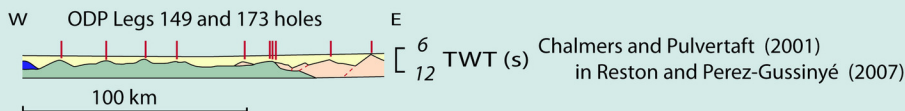


c. Duretz et al. (in revision)

MODELS

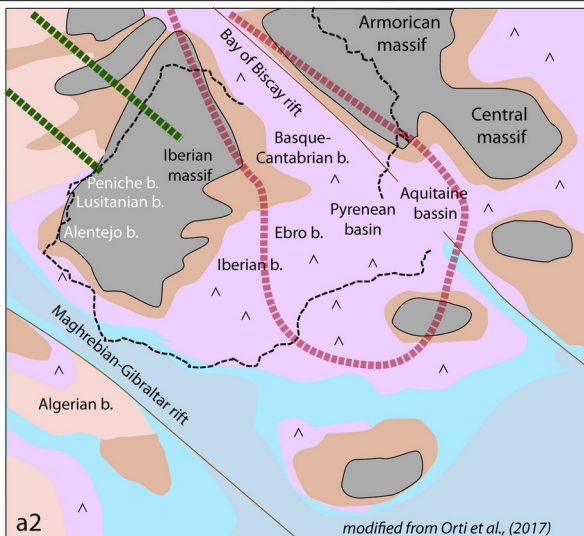
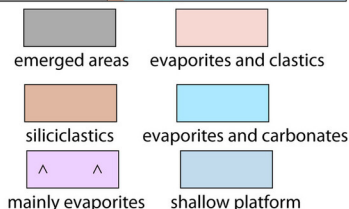
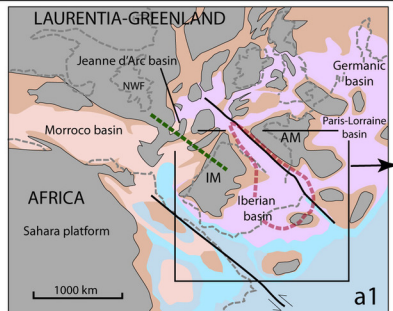


d. Conjugate Angola-Campos basin margins (modified after Brune et al. (2014) : mirror image)



e. West Iberia

PASSIVE MARGINS

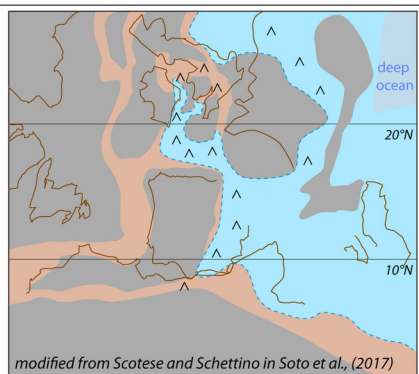


IM: Iberian massif, AM: Armoric massif, NWF: Newfoundland

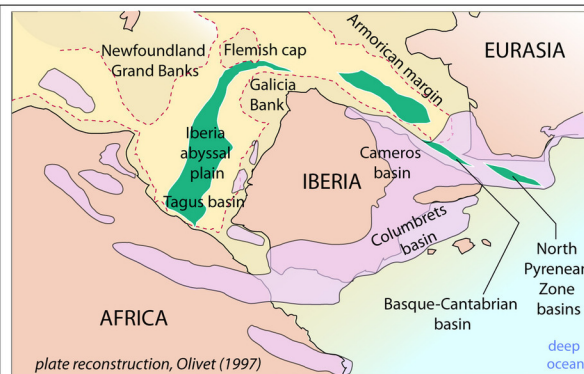
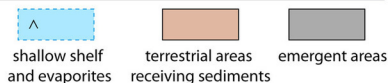
Regions characterized by further development of Cretaceous smooth slopes rift basins

Western part of the Iberia-Newfoundland transect

a



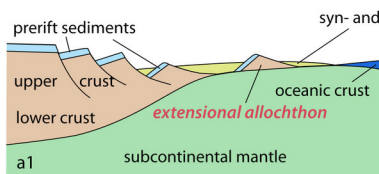
b



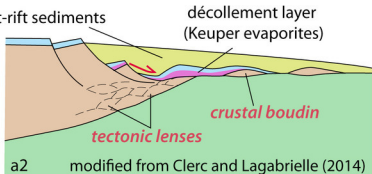
c



Iberia-Newfoundland-type margins



Smooth slopes, Parentis-type basins



a lack of pre-rift décollement layer

with a pre-rift décollement layer

