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Sediment routing system and sink preservation during the post-orogenic evolution of a retro-foreland basin: The case example of the North Pyrenean (Aquitaine, Bay of Biscay) Basins

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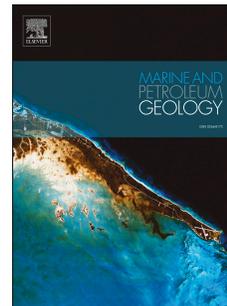
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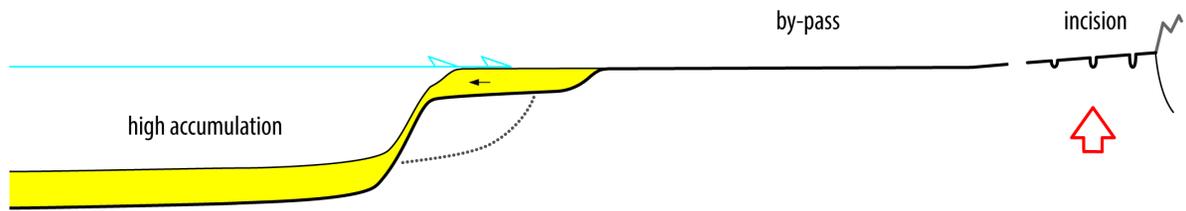
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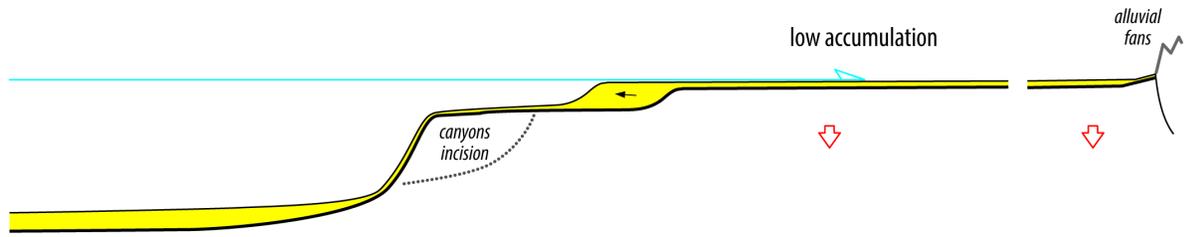
3. POST-FORELAND - 2

$$\Delta A_{sub} \ll \Delta S_{sc} [\Delta A_{sub} \leq 0]$$



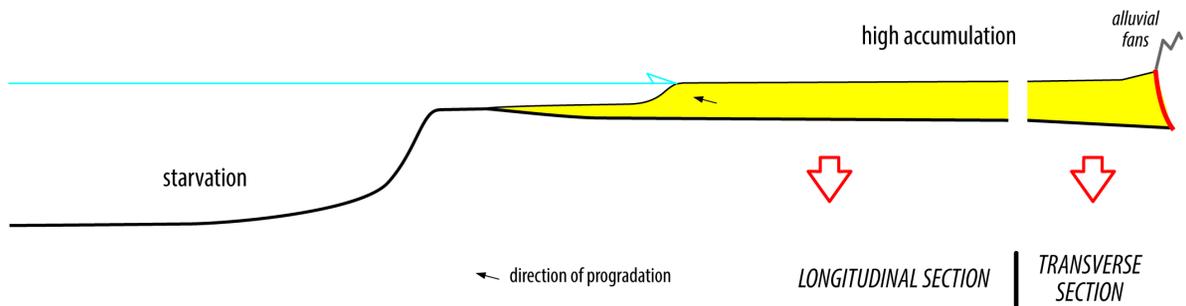
2. POST-FORELAND - 1

$$\Delta A_{sub} < \Delta S_{sc} [\Delta A_{sub} > 0]$$



1. FORELAND (Foredeep)

$$\Delta A_{sub} \leq \Delta S_{sc}$$



1 **Sediment routing system and sink preservation during the post-orogenic evolution of a**
2 **retro-foreland basin: the case example of the North Pyrenean (Aquitaine, Bay of Biscay)**

3 **Basins**

4

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14

15 **Abstract**

16

17 We investigated here the evolution of the sediment routing system, i.e. the sediment

18 transport and deposition evolution along successive depositional topographies and

19 environments, and the sink (i.e. deposited sediments) preservation in a foreland basin from

20 the period of mountain belt shortening to its post-orogenic stage. The studied system is the

21 North Pyrenean retro-foreland basin from 50 Ma to today which is, composed of a subsiding
22 platform (the Aquitaine Basin) fed by the erosion of the Pyrenees passing laterally to a slope
23 and a deep-sea plain (the Bay of Biscay deep basin), the ultimate area of deposition. This
24 study is based on a double seismic stratigraphic and structural analysis of an extensive
25 seismic dataset and on an age model of the sediments combining biostratigraphy,
26 orbitostratigraphy and sequence stratigraphy with a time resolution of 0.1 Ma.

27 Four major periods of deformation corresponding to a single or a set of stratigraphic
28 sequence boundaries were characterized. The Pyrenean shortening decrease may be
29 recorded by a basin-scale uplift at 49.8 Ma (Late Ypresian). The paroxysm of the piggy-back
30 shortening and related uplift is dated at 35.8 Ma (Priabonian). The end of the Pyrenees
31 shortening (transition to post-orogenic conditions) is well dated between 27.1 and 25.2 Ma
32 (Chattian). A major West European scale deformation of possible mantle origin uplifted the
33 basin from 16.4 to 10.4 Ma (Late Burdigalian to Early Tortonian).

34 These major periods of deformations controlled the sink preservation at the first order
35 through the ratio between the accommodation space created by subsidence on the
36 Aquitaine platform (A_{sub}) and the siliciclastic sediment supply coming from the erosion of the
37 Pyrenees (S_{sc}). A general model is proposed. At the time of the foreland basin when $\Delta A_{sub} \leq$
38 ΔS_{sc} , most of the sediments are preserved on the platform as a progradational-aggradational
39 wedge (up to 25.2 Ma here). At the time of the post-foreland evolution when $\Delta A_{sub} < \Delta S_{sc}$,

40 most of the sediments are transferred to the deep-sea plain with few preservations on the
41 platform and when $\Delta A_{\text{sub}} \ll \Delta S_{\text{sc}}$ with $\Delta A_{\text{sub}} \leq 0$, all of the sediments are transferred to the
42 deep-sea plain as deep-sea fans and fluvial by-pass or erosion is the dominant process on
43 the platform.

44 The continental sediment routing system is mainly provided along nearly flat alluvial plains
45 with extensive lakes and/or humid zones, i.e. the local base levels of the alluvial fans. The
46 Aquitaine retro-foreland was never overfilled.

47

48

49 **Keywords:** Foreland, Post-orogenic, Sediment routing, Sink, Pyrenees, Aquitaine Basin, Bay
50 of Biscay Basin

51

52

53 **1. Introduction**

54

55 Plate flexure that results from tectonic loading by collisional orogens creates
56 accommodation space that is filled by sediments thereby creating foreland basins
57 (Dickinson, 1974; Beaumont, 1981; Allen and Allen, 2013). In doubly-vergent belts, there are
58 two types of foreland according to their position with regards to the orogenic wedge

59 (Johnson and Beaumont, 1995; Naylor and Sinclair, 2008): (i) pro-foreland on the lower
60 (underthrust) lithosphere and (ii) retro-foreland on the upper (overriding) lithosphere.
61 The dynamic topography due to mantle flow generated by slab subduction may add an
62 additional component of the subsidence in foreland basins (Mitrovica et al., 1989; Gurnis,
63 1992; DeCelles and Giles, 1996).

64

65 The first-order *stratigraphic architecture* of foreland basins is well understood since the
66 work of Fleming and Jordan (1989). The main parameters controlling the basin infill are :
67 thrust loading, flexural rigidity of the continental lithosphere, erosion rates of the mountain
68 belt, 'depositional styles' (e.g. Sinclair et al., 1991; Sinclair and Allen, 1992). They control
69 whether foreland basins are underfilled, balanced or, overfilled according to the balance
70 with subsidence and sediment flux (Covey, 1988). Some authors replaced foreland infilling in
71 a sequence stratigraphy framework (Posamentier and Allen, 1993). Lastly, some others
72 integrated the effect of both flexural and dynamic topography-induced subsidence
73 (Catuneanu et al., 1997) with the concept of reciprocal stratigraphy.

74

75 The *sediment routing* system in foreland basins (Allen, 2017) is highly dependent on the
76 basin physiography. Most forelands are exoreic systems connected to the sea. In that case,
77 the connection of the drainage system to the ocean (e.g. Miall, 1981) may be (1) a lateral

78 evolution of the foreland basin to a passive margin in a single subsiding domain (e.g. the
79 Western Interior Basin, the foreland basin of the Rocky Mountains passing southward to the
80 passive margin of the Gulf of Mexico) or (2) two disconnected basins, the foreland and
81 passive margin basins, with a non subsiding domain in between (e.g. Amazon foreland and
82 the Fos de Amazonas passive margin in Brazil). During the last stage of the foreland
83 evolution when part of the basin can be transported (piggy-back), the effect of the growing
84 thrusts (lateral and frontal ramps) on the drainage system is well understood in the South
85 Pyrenean foreland (e.g. Vergés and Garcia-Senz, 2001). Recently some studies have focused
86 on the quantification of the sediment routing system and the sediment mass balance
87 ('source-to-sink' approach) in the South Pyrenean foreland basin (Michael et al, 2013,
88 2014ab; Armitage et al., 2015), examining the role of catchments uplift and/or surface runoff
89 variations, in addition to the effect of relative sea level variations on sediment infilling.

90

91 Little attention has been paid to the post-orogenic evolution of the foreland basins. The
92 best documented example is the Alpine 'Molassic' Basin in Switzerland (Schlunegger and
93 Mosar, 2011; Willett and Schlunegger, 2010) for which the transition to post-foreland
94 conditions is an overall uplift of the basin at the time of the Jura wedge main activity. These
95 authors emphasized the importance of the boundary conditions in controlling the end of the

96 foreland subsidence period: the mechanical properties of the lithosphere, the existence of a
97 decollement level and the importance of emerging relief around.

98

99 By further investigating how the sink is preserved in the North-Pyrenean foreland basin
100 we propose a model of sediment response to the syn-/post- orogenic transition. The studied
101 system is the retro-foreland of the Pyrenees (Fig. 1), the Aquitaine Basin and its lateral
102 equivalent, the deep Bay of Biscay Basin – the ultimate area of deposition on intermediate
103 to oceanic crust.

104 Here, we present a seismic stratigraphic analysis of an extensive 2D seismic dataset
105 supplemented by wells (petroleum and water resources). This analysis is based on (1) an age
106 model of the sediments using a new method of dating that combine biostratigraphy
107 (published and new data), orbitostratigraphy and sequences stratigraphy, (2) a geometrical
108 reconstruction of the basin based on seismic stratigraphy and structural analysis (2D sections
109 and isopach maps), (3) a reconstruction of the successive depositional profiles using facies
110 sedimentology and (4) a characterization of the tectonic structures.

111

112 **2. Geological setting**

113

114 *2.1. Main topographic and structural features (Fig. 1)*

115

116 The studied area is subdivided into three main physiographic units. Eastward, the
117 *Aquitaine Basin*, bounded to the south by the Pyrenees and to the north by the French
118 Massif Central (exhumed Variscan basement) extends offshore up to the shelf-break.
119 Westward, the *Bay of Biscay deep basin*, with a mean water depth of 4000 – 4500 m and
120 bounded to the north by the South Armorican Margin and to the south by the Cantabrian
121 (North Iberian) Margin, is located on both an hyperextended continental crust and an
122 oceanic crust. In between, the *Landes Plateau* is a step (more than 100 km wide), that is
123 bounded by quite steep slopes (maximum values: 13-15.5°) eastward and westward, by, the
124 Cap Ferret Canyon northward and the Cap Breton Canyon southward.

125 The continental basement of the Aquitaine Basin is made of late orogenic Variscan
126 structures (Carboniferous-Permian) and little deformed Early Paleozoic rocks (Le Pochat,
127 1984; Paris and Le Pochat, 1994). The key structural feature is the occurrence of a Triassic
128 deposits south of the so-called Celtaquitaine flexure' or hinge line, which is in fact onlap limit
129 of the Late Triassic salt sediments (Bourouilh et al., 1995, Fig. 1).Evaporitic deposits
130 controlled Cretaceous to Cenozoic salt tectonic and diapiric features both in the southern
131 Aquitaine Basin and on the Landes Plateau. The most remarkable ones are the Arzacq,
132 Tartas, Tarbes and Mirande Subbasins bounded by blind thrusts (Audignon and
133 Maubourguet Ridges) or transcurrent zones. The Parentis Subbasin, a rifted basins aborbed

134 in the Early Cretaceous (Mathieu, 1986; Ferrer et al., 2012; Tugend et al., 2015), is located in
135 the middle part of the Aquitaine Basin, mainly offshore. The Bay of Biscay deep basin is
136 characterized by highs (or banks) resulting from the inversion of extensional blocks both to
137 the north (e.g. Gasconne Dome and Trevelyan High - Thinon, 1999; Thinon et al., 2001, 2002)
138 and to the south with the Le Danois Bank, inversion of the Asturian Basin (Cadenas et al.,
139 2017) bounded by the Biscay wedge front (Fernández-Viejo et al., 2012). The North Iberian
140 Biscay wedge front and the North Pyrenean front are connected along a faulted zone, the
141 Santander “soft” transfer zone (Roca et al., 2011) corresponding to the sharp westward limit
142 of the Landes Plateau and controlling the location of the north-south trending Torrelavega
143 and Santander Canyons.

144

145 *2.2. Evolution of the Pyrenees and its foreland basins*

146

147 The Pyrenees mountain range and its westward equivalents the Basque-Cantabrian
148 Mountains result from the compression and inversion of the hyperextended Eurasian
149 lithosphere during Albian times since 85 Ma (e.g. Lagabrielle et al., 2010; Masini et al., 2014;
150 Clerc et al., 2016; Saspiturry et al., 2019). Even if the Pyrenean Belt is not cylindrical (Chevrot
151 et al., 2018), its structure can be described as a wedge of Eurasian lithosphere over the
152 Iberian lithosphere plunging to the north (e.g. Roure et al. 1989; Teixell et al., 2018).

153

154 The inversion tectonics started at the time of the Africa-Eurasia convergence with the
155 Iberian microplate at the end of the Santonian (83.6 Ma - e.g. Schettino and Turco, 2011).
156 Plate kinematic constraints (Roest and Srivastava, 1991) impose an end of convergence
157 between Iberia and Eurasia not no later than chron 6c, i.e. around the Oligocene-Miocene
158 boundary (22.6-24.1 Ma, Gradstein et al., 2012). The total amount of shortening varies
159 along-strike: 82 km to the East (Verges et al., 1995) , 142 km to 165 km in the center
160 (Beaumont et al., 2000, Mouthereau et al., 2014) and 114 km to the West (Teixell et al.,
161 2016). The measurements of the shortening rates through time along different segments of
162 the mountain belt (Mouthereau et al., 2014; Teixell et al., 2016 –see synthetic Fig. 12) shows
163 maximum rates from 66 Ma (base Palaeocene) to 48 Ma (base Middle Eocene) (32 km of
164 shortening for Mouthereau et al. vs. 54 km for Teixell et al.) followed by a decrease up to 20
165 – 15 Ma. The collision occurred synchronously along strike. However, the exhumation was
166 delayed toward the west due to the progressive closure of a larger Early Cretaceous domain
167 to the west (Vacherat et al., 2017). Numerical modelling (Curry et al., 2019 - see Fig. 12) of
168 the lithospheric flexure due to loading by the Pyrenees suggests a sharp topographic growth
169 of the Pyrenees during the Priabonian (38-34 Ma – up to 2 to 3 km of maximum elevation)
170 reaching its maximum (3.5 km) around the Oligocene-Miocene boundary (23 Ma). Isotopic
171 studies (Huyghe et al., 2012) suggest an earlier uplift of the eastern mountain belt during

172 Middle Eocene times. Curry's model (2019) is in agreement with thermochronological data
173 (e.g. Fitzgerlad et al., 1999; Sinclair et al., 2005; Fillon and van der Beek, 2012; Bosch et al.,
174 2016) which showed an acceleration of the exhumation during late Eocene-Oligocene times.

175 This convergence results in the formation of a major pro-foreland basin to the south, the
176 South Pyrenean Basin and a retro-foreland - to the north, the Aquitaine (Carcassonne) Basin.
177 The South Pyrenean Basin was initiated during the Late Santonian (around 84 Ma –
178 Puigdefabregas and Souquet, 1986) and is transported as a piggy-back basin at the base of
179 the Ypresian (Puigdefabregas and Souquet, 1986; Vergés et al., 2002). This basin opens up
180 toward the Atlantic and became an endoreic system at time of the uplift of the Basque-
181 Cantabrian Mountains, i.e. during the Late Eocene (37 Ma – Gomez et al., 2002).

182 Along the Bay of Biscay the former Lower Cretaceous extensional blocks of the South
183 Armorican Margin are inverted during early Upper Cretaceous, Palaeocene and Upper
184 Eocene times, this last being the major one with a significant dextral strike-slip component
185 (Thinon, 1999; Thinon et al., 2001, 2002). The southern part, the Asturian Basin (Le Danois
186 Bank), is inverted and thrust from the Upper Eocene to the Eocene-Oligocene boundary
187 (paroxysm of the deformation - Gallastegui et al., 2002).

188

189 Many studies focussing on the Pyrenees mountain belt and surrounding domains
190 (Desegaulx et al., 1991; Angrand et al., 2018; Cochelin et al., 2018; Espurt et al., 2019)

191 conclude on the importance of structural inheritance of previous deformation events, e.g.
192 the late Variscan (Carbonifereous to Permian) orogeny and Lower Cretaceous extension.
193 The most important event is the Albian lithospheric thinning that controlled the
194 segmentation of the foredeep into numerous subbasins during the Palaeogene (Angrand et
195 al., 2018). The second inheritance effect of the Albian rifting is the occurrence of a rigid
196 block located between the Parentis and Arzacq-Mauleon rifts, the Landes High, which
197 extends from the Landes Plateau to the southwestern part of the Aquitaine Basin (Tugend et
198 al., 2014). For the late Variscan orogenic deformations, although the role of the inherited
199 structure is clearly demonstrated on the 2D sections (Espurt et al., 2019), no clear plan-view
200 data (maps) are available and the meaning of the N20° faults, such as the Pamplona and
201 Toulouse Faults (with possible other ones in between), is still unclear.

202

203 *2.3. Cenozoic stratigraphy, palaeogeography and deformation of the retroforeland of the*
204 *Aquitaine Basin*

205

206 We mainly focused on the Aquitaine Basin in the present study. Little is known on the
207 stratigraphy of the Bay of Biscay due to (i) the absence of accurately located deep-sea-
208 drillings (e.g. DSDP site 128 leg 12 drilled north of Galicia on the Cantabria Seamount) and by

209 (ii) the few studies available (e.g. Cremer, 1983) that extrapolate ages of the deep sediments
210 from the shelf wells.

211 At the first order the Cenozoic infilling of the Aquitaine Basin result from an overall
212 progradation from east (Carcassonne- Corbières area) to west (Atlantic Margin) due to deltas
213 or carbonate platforms prograding wedges with clinoforms that are hundreds of meters tall
214 (Winnock et al., 1973; Dubreuilh et al., 1995). Despite numerous stratigraphic studies (e.g.
215 Cavelier et al., 1997; Sztrákos et al., 1997, 1998, 2010; Sztrákos and Steurbaut, 2017), the
216 lithostratigraphic nomenclature and dating are still quite contradictory. Few 3D
217 reconstructions based on sequence stratigraphic analysis are available (Serrano, 2001;
218 Serrano et al., 2001 for the northern part of the foredeep). Nevertheless, the history of the
219 Cenozoic basin infilling can be summarized in five steps, following a Late Cretaceous
220 (Campanian-Maastrichtian) early phase of flexuration (e.g. Ford et al., 2016).

- 221 • From the *Danian* to *Thanetian* (66-56 Ma), large shallow-water carbonate platforms
222 covered the southern part of the Aquitaine Basin (Sztrákos et al., 1997; Serrano, 2001)
223 passing northward to laterites (Gourdon-Platel et al., 2000). Southward thick deep-
224 water deposits infilled the foredeep. They are made up of gravity carbonate deposits for
225 the Danian and siliciclastic deep-sea-fans for the Thanetian (Dubarry, 1988).
- 226 • After a major Early Ypresian retrogradation and a maximum marine flooding of Middle
227 Ypresian age, the *Upper Ypresian* to *Early Lutetian* (52-44 Ma) time interval corresponds

228 to a sharp deltaic progradation (Cavelier et al., 1997) driven by an increase in siliciclastic
229 sediment supply coeval with the Pyrenean shortening (Serrano, 2001; Serrano et al.,
230 2001).

- 231 • The *Late Lutetian to Oligocene* (44-23 Ma) was characterized by carbonate platforms
232 passing upstream to an alternation of evaporitic lagoonal (clays with gypsum) to
233 continental (lacustrine carbonate, fine-grained fluvial deposits) environments, called
234 'Molasse' by French authors. This time range is the end of the infilling of the foredeep
235 and the beginning of the propagation of the thrusts in the Aquitaine Basin along the
236 Triassic salt decollement level and the growth of active ridges. The timing of this change
237 of deformation pattern (end of the foredeep and beginning of the thrusting) is poorly
238 constrained. The sediment thickness maps of Serrano (2001) provide age constrains for
239 the end of the foredeep subsidence and the initiation of the propagation of the thrusts
240 in the Aquitaine Basin with the growth of the Audignon thrust and salt ridge (Fig. 1):
241 after a transition period (Nousse Fm, here dated as Lutetian in age, 47.8-43.5 Ma), the
242 growth of the Audignon ridge was active just after the Nousse Fm (here dated of Late
243 Lutetian age, 43.5 Ma). Rocher et al. (2000) measured the shortening of the Landes-de-
244 Siougos Anticline located in the Tartas Subbasin in front of the Audignon thrust (Figs. 1
245 and 2) that reached a shortening paroxysm during Priabonian times (38-34 Ma).

- 246 • The *Early Miocene* (23-16 Ma) was characterized by marine flooding (the so-called
247 'Faluns' by French authors) of a large embayment located in the central part of the
248 Aquitaine Basin passing eastward (below the present-day Ger and Lannemezan plateaus)
249 to continental environments dominated by lacustrine limestones (Crouzel, 1957; Antoine
250 et al., 1997). Rocher et al. (2000) measured paleostress magnitudes from calcite twin
251 data in the Arzacq and Tartas Subbasins and suggested a late NW-SE shortening in the
252 Mio-Pliocene.
- 253 • The *Middle Miocene to Present-Day* (16-0 Ma) corresponds to the major
254 continentalization of the Aquitaine Basin with the deposition of thin (10-40 m) coarse-
255 grained alluvial deposits (Middle Miocene, 16-11.5 Ma) flooded by the sea during the
256 Langhian and base Serravallian (Gardere et al., 2002; Gardere, 2005). The Pliocene is
257 characterized by sandy alluvial deposits (maximum thickness of 100 m) interstratified
258 with several lacustrine (clays) and marsh (lignites) sediments (Dubreuilh et al., 1995).
259 During the Calabrian (1.8-0.8 Ma) the drainage of the Aquitaine Basin is reorganized from
260 rivers flowing to the Parentis Subbasin to the modern one with a single river conduit, the
261 Garonne-Gironde system (Dubreuilh et al., 1995).

262

263 3. Available data and methods

264

265 *3.1. Available data (Fig. 2)*

266

267 The Aquitaine Basin has been extensively studied since discovery of gas (Lacq structure) in
268 the 1950s (Biteau et al., 2006). Around 40 000 km of seismic reflection lines and 1 600
269 industrial wells were available for this study. In order to date the sediments we focused on
270 some key wells, offshore, Ibis 2 and Pingouin 1 wells (cuttings), onshore Laborde 1D
271 (cuttings) and Landes-de-Siougos (cores) wells (see Fig. 2 for location). We also used the
272 water and geotechnical shallow drillings collected by the French geological survey (BRGM)
273 available in the French drillings database (BSS).

274 In the deep Bay of Biscay and Landes Plateau 35 000 km of industrial and regional seismic
275 lines shot as part of the MARCONI Spanish project, were studied. The only deep wells drilled
276 in that area are DSDP wells (sites 118 and 119 – leg 12 and some leg 48 sites – see Fig. 2 for
277 location), located on top of sea mounts or inverted tilted blocks that make them difficult to
278 use for calibrating the seismic lines of the Bay of Biscay in terms of lithology and ages.

279

280 *3.2. Seismic and wells interpretation: sequence stratigraphy*

281

282 We here performed a seismic stratigraphic analysis in order to define depositional
283 sequences. Two approaches of seismic stratigraphy, standardized by Catuneanu et al. (2009),
284 were applied.

285 The first approach (Brown and Fisher, 1977; Mitchum et al., 1977) is based on the analysis
286 of *seismic reflectors terminations* (onlap, downlap, toplap, truncations). The second one
287 (Helland-Hansen and Gjelberg, 1994; Helland-Hansen and Martinsen, 1996; Helland-Hansen
288 and Hampson, 2009) is based on the *offlap break* (shoreline or shelf-edge break) *trajectory*
289 over time by defining stratal patterns: forced (descending) regressive, normal (ascending)
290 regressive and transgressive.

291 A depositional sequence is defined here as follows (see Ponte et al. 2019 for a summary
292 of the different approaches):

- 293 • the *sequence boundary*, which corresponds to the first correlative conformity (CC,
294 Catuneanu et al., 2009) and the onset in continental domain to the subaerial
295 unconformity, an erosion surface overlain by onlapping strata;
- 296 • the forced regressive (FR) deposits (Catuneanu et al., 2009) (equivalent of the forced
297 regressive wedge system tract of Hunt and Tucker (1992) or the falling-stage system tract
298 (FSST) of Plint and Nummedal, 2000), which correspond to descending regressive
299 shorelines (i.e forced progradation) passing toward the deep-sea plain to the basin floor
300 fan;

- 301 • the lowstand normal regressive (LNR) deposits (Catuneanu et al., 2009) which
302 correspond to the lowstand system tract (LST) of Posamentier and Vail (1988) or the
303 ascending regressive shorelines (i.e progradation-aggradation) of Helland-Hansen (e.g.
304 Helland-Hansen and Hampson, 2009);
- 305 • the *maximum regressive surface* (MRS), which corresponds to the former transgressive
306 or flooding surface Posamentier and Vail (1988) above the toplapping strata;
- 307 • the transgressive deposits (T; Catuneanu et al., 2009) which correspond to the
308 transgressive system tract (TST) of Posamentier and Vail (1988) or transgressive
309 shorelines (i.e retrogradation) of Helland-Hansen and Hampson, (2009);
- 310 • the *maximum flooding surface* (MFS) which lies below downlapping strata;
- 311 • the highstand normal regressive (HNR) deposits (Catuneanu et al., 2009) which
312 correspond to the highstand system tract (HST) of Posamentier and Vail (1988) or
313 ascending regressive shorelines (i.e progradation-aggradation) of Helland-Hansen and
314 Hampson (2009).

315

316 Depending on their cause - (eustasy or lithosphere deformation), several orders of
317 depositional sequences of different durations have shaped the past geological periods
318 (Graciansky et al., 1998). As a consequence, the stratigraphic record is a stacking of different
319 orders of nested sequences. Vail et al. (1977) and Vail et al. (1991) defined the orders of

320 sequences based on their duration: first order sequences have a duration around 100 - 200
321 Myrs, second order sequences around several 10 Myrs, third order sequences around
322 several 1 Myrs, fourth order around several 100 kys. Determining the hierarchy of
323 depositional sequences (between two sequence boundaries) or stratigraphic cycles
324 (between two MFS) supposed to date the sequences for establishing their order, and
325 possibly their causes.

326 The nature of the control of the depositional sequences – eustasy or tectonic
327 (lithosphere deformation) is based on (1) the accommodation space measurement using the
328 shoreline wedges on both sides of the sequence boundaries (SB) and (2) the geometrical
329 relationships between the SB and underlying sediments based on the principles (e.g. Robin
330 et al., 1998) that (i) *eustasy* is only a function of space and must have an equal sea level
331 variation value (for a given time interval) over the whole basin and (ii) a *lithosphere*
332 *deformation* is a function of both space and time, i.e. it may change in amplitude along the
333 basin with possible truncations of tectonic structures with different wavelengths.

334 The accommodation space is the vertical displacement of the shoreline (Jervey, 1988) and
335 is measured as the vertical displacement of the last shoreline below the SB and the first one
336 preserved above. This distance measured in travel double times are later converted in depth
337 (m) using the velocity law presented in the supplementary materials. Sediment

338 decompactions was not performed: the measurements are therefore the minimum relative
339 sea level variations values.

340 The eustatic curves used are Haq's curve (Haq et al., 1987, 1988; Hardenbol et al., 1998)
341 based on a compilation of coastal onlap curves in different basins of the world and Miller's
342 one (Miller et al., 2005, 2008) based on the accommodation space measurement along the
343 well-known New Jersey margin (New York). The amplitudes of both curves are different and
344 recent studies or compilations (e.g. Bessin et al., 2017 for a review) suggest that the
345 amplitudes measured by Miller's group are more realistic.

346

347

348 Some well-logs correlations using the principles of sequence stratigraphy (Posamentier
349 et al., 1988) and the 'stacking pattern' technique (see van Wagoner et al., 1990; Catunenau,
350 2006) were used for this work.

351

352 A space-time stratigraphic diagram (known as a Wheeler diagram) was compiled (see Fig.
353 8 and supplementary materials) based on the age model (see 3.3), and indicating (1) the
354 amount of time preserved as volumes of sediment, as condensation (no deposition by
355 downlap or onlap) and as eroded sediments or by-pass, (2) the location and nature of the

356 offlap break (see 3.4), (3) some remarkable environments (alluvial fans and braided alluvial
357 plains) and (4) the lithostratigraphy.

358

359 *3.3. Sediment dating*

360

361 The applied dating is a combination of biostratigraphy, orbitostratigraphy and sequence
362 stratigraphy in order to define a high-resolution (0.1 Ma) age model for the Aquitaine Basin
363 both onshore and offshore. This method is developed in supplementary material 1.

364 The used lithostratigraphic nomenclature is taken from Sztrákos et al. (1997, 1998) and
365 Serrano et al. (2001) for the Eocene, from Sztrákos and Steurbaut (2017) for the Oligocene,
366 of Cahuzac (1980) for the Lower and Middle Miocene and from Dubreuih et al. (1995) from
367 the Upper Miocene to the present-day.

368

369 *3.4. Reconstruction of the depositional profile*

370 For a given time interval, the reconstruction of successive depositional profiles is based
371 on the lateral evolution of the sedimentary environments from the most proximal preserved
372 continental facies to the deepest marine deposits located on the oceanic crust. Specific
373 attention was paid to the continental environments, major constrain for the reconstruction
374 and the evolution of the sediment routing systems. The characterization of the sedimentary

375 environments is based on (i) the facies sedimentology of the cores and outcrops, (ii) the
376 palaeoecology provided by the faunas and floras fossilized in the cuttings and (iii) the seismic
377 geometries.

378

379 The location of the *shoreline* is based on the identification of offlap breaks. There are at
380 least two types of offlap breaks (1) shorelines or (2) shelf breaks (e.g. Helland-Hansen and
381 Hampson, 2009). They also can result from subaqueous shoals or reef breaks in carbonate
382 platforms. In order to discern shorelines from shelf breaks a well control is required to
383 check, in the case of shorelines, that upstream facies are clearly continental, based on the
384 lithology (e.g. coal) and palaeoecology provided by the cuttings or the well-logs signatures
385 (see van Wagoner et al., 1990; Catuneanu, 2006).

386

387 The reconstruction of the *marine environments* is based on a seismic analysis of the
388 clinoforms and certain specific seismic geometries. In marine environments, the high and
389 slope of the clinoforms provide indications regarding their origin (delta front, slope
390 downstream of the shelfbreak – see Patruno and Helland-Hansen, 2018, for a review). We
391 did not study deep-sea deposits, e.g. deep-sea fans, contouritic ridges, sand-waves as it was
392 outside the scope of this study.

393

394 3.5. *Isopach maps*

395

396 Two isopach maps were built in this study based on the seismic interpretation of the
397 available seismic lines, a first one from the Palaeocene to Oligocene (66-23 Ma) and a
398 second one from the Miocene to today (23-0 Ma).

399 The base Palaeocene (base Cenozoic) and base Miocene (base Neogene), as well as other
400 key surfaces, were propagated from the present-day Pyrenean piedmont to the distal part of
401 the Bay of Biscay using the sequence stratigraphy principles, based on seismic lines (90% of
402 the area), shallow wells (BSS database) and 1:50 000 geological maps at for the onshore
403 outcropping areas. The main concern was the conversion of seismic lines from two-way
404 travel time (TWT) in seconds into depth in meters, using velocity laws (see supplementary
405 materials) available from the industrial wells that were compiled as part of this study. The
406 extrapolation of the 2D (seismic lines) and 1D (wells) data, both irregularly distributed, was
407 based on the Natural Neighbour method (GIS).

408

409

410 **4. Results**

411

412 *4.1. Main steps of the Aquitaine Basin infilling: sequence stratigraphy*

413

414 4.1.1. Main sedimentary environments

415

416 The *continental environments* of the Aquitaine Basin can be grouped into three facies

417 associations characteristic of different depositional profiles and slopes: (1) extensive

418 lacustrine to palustrine facies alternating with fluvial systems, (2) megafans or braided

419 alluvial plains and (3) alluvial fans.

420 • The *lacustrine to palustrine* environments – which can be correlated over a distance of421 several tens of kilometers up to one hundred kilometers – that alternate with *fluvial*422 *channels (alluvial to coastal plains)* indicate nearly flat domains with almost no slope. For

423 large coastal plains, the cuttings and cores contain dinocysts (e.g. Mudie et al., 2017) and

424 no marine microfaunas (absence of benthic foraminifers and calcareous nannofossils).

425 The lithology is either claystones with evaporites (gypsum) in the case of evaporitic

426 coastal plains (e.g. Warren, 2010) or claystones with coals (lignites) in the case of

427 marshes (palustrine environments – McCabe, 1984). For extensive lakes (e.g. Gierlowski-

428 Kordesch, 2010), the cuttings and cores are made up of correlatable micritic to bioclastic

429 limestones with charophytes (e.g. Anadon et al., 2002) and fresh water gastropods

430 (limnees, planorbis, etc). The interstratified channels are composed of medium to fine-

431 grained sandstones without any marine fossils. They correspond to suspended-load to
432 mixed-load fluvial channels.

433 • The *megafans to braided alluvial plains* are homolithic coarse-grained sandy (sometimes
434 with pebbles) deposits continuous over a distance of several tens to one hundred of
435 kilometers, without any marine fossil (Singh et al., 1993; Shukla et al., 2001). They
436 correspond to a stacking of bedload fluvial channels. The main difference between
437 megafans and braided alluvial plains is the source of sediments deduced from
438 palaeogeographic maps: point source for the megafans and multiple sources for the
439 alluvial braided plains.

440 • The *alluvial fans* are mostly homolithic conglomerates continuous over a distance of
441 several kilometers to ten kilometers, without any marine fossils (e.g. Stanistreet and
442 McCarthy, 1993; Blair and McPherson, 2009), with some evidences of ephemeral lakes
443 (non correlatable “multi-coloured” clays) or subaqueous soils (calcretes s.l.).

444

445 The *shallow marine environments* of the Aquitaine Basin are mainly mixed siliciclastic –
446 carbonate platforms. They became siliciclastic - similar to the modern environments, in
447 Middle Miocene. Previously based on the seismic geometries or on facies on cores, there are
448 two types of depositional profiles (i) rimmed carbonate platforms or (ii) ramps (e.g.
449 Handford and Loucks, 1993). Rimmed carbonate platforms – the most common profile – are

450 highly variable. The barrier may be large reefal build-ups, reef patches or bioclastic shoals. At
451 the back of these barriers, the inner platform may be (i) more or less large lagoons or bays
452 passing upstream to the above described coastal plain or (2) large tide-dominated epeiric
453 seas (called the 'faluns' by French authors).

454

455 *4.1.2. Description: second order sequences*

456 In the Aquitaine Basin major sequences boundaries (SB) are defined using three criteria:
457 (1) a removal of at least several tens of meters of accommodation space, (2) a change in the
458 sedimentary system and/or (3) the importance of erosional truncations of the underlying
459 sediments. They may be distributed in the Aquitaine stratigraphic record as single SB or as
460 sets of several SB due to a long term (second order) decrease or removal of the
461 accommodation space upon which the shorter (third order) accommodation variations were
462 superimposed. In this second case the base of the second order sequence is defined here as
463 the first SB.

464

465 Four unconformity-bounded second order sequences were identified along the Aquitaine
466 platform, from the uppermost Ypresian until today.

467

468 *YP's second order sequence (Late Ypresian - lowermost Priabonian – 49.7-37.6 Ma)*

469 The *age* of the base sequence boundary (SByp) is Late Ypresian (49.7 Ma - base of the
470 Lussagnet Fm – see supplementary material 1 for age constrains). The age of the MFS (YP-
471 MFS-4) is at the base of the Bartonian (41.6 Ma – within the lower member of the
472 Brassempouy Fm). At least five main third order sequences were identified.

473 The *base sequence boundary* (Fig. 5 and supplementary material 2) separates a highstand
474 normal regressive (HNR) wedge (possible pure progradation) from a lowstand normal
475 regressive (LNR) wedge with no forced regressive deposits in between. The amplitude of the
476 relative sea level of fall is 40 to 80 ms, i.e. 51 to 104 m (velocity law: 2540 to 2590 m/s). No
477 tectonic structures are truncated by this sequence boundary (Figs. 3 and 6).

478 The YP sequence is characterized by downlap terminations, with time condensation in the
479 deepest part of the Aquitaine platform (Fig. 8).

480 The *depositional profile* changed through times with three main periods.

- 481 • The Late Ypresian depositional profile (Table 1) is characterized onshore by megafans
482 deposits (Lussagnet Fm) mainly fed from the French Massif Central (Schoeffler, 1971)
483 supplying a large deltaic system (Donzacq Fm , E. Lasseur and J. Briais, work in progress).
- 484 • The Early Lutetian and Late Lutetian depositional profiles (Table 1) are very similar
485 starting with an alluvial fan (Member 1 of the Palassou Fm - eroded during Chattian
486 times along section LR6 - Fig. 3) passing to nearly flat alluvial to coastal plains (Cavalante
487 Fm and Tartas Fm). The main difference concerns the carbonate platforms that are both

488 reef rimmed carbonate platforms and flat (Nousse Fm) for the Early Lutetian and with
489 seismic-scale build-ups (Brassempouy Fm) for the Late Lutetian/Priabonian.

490

491 *PC's Second order sequence (lowermost Priabonian – Chattian – 37.6-27.1 Ma)*

492 The *age* of the base sequence boundary (SBpc) is at the base of the Priabonian (37.6 Ma -
493 base of the Campagne Fm, just above the top of the Brassempouy Fm – see supplementary
494 material 1 for age constrains). The age of the MFS (PC-MFS-9) is Late Rupelian (29.4 Ma –
495 base of the upper member of the Mugron Fm). At least five main third order sequences were
496 identified.

497 The *base sequence boundary* (Figs. 4 and 5, supplementary material 2) separates
498 highstand normal regressive (HNR) wedge from a lowstand normal regressive (LNR) wedge
499 with no forced regressive deposits in between. The amplitude of the relative sea level of fall
500 is difficult to quantify due to the poor offlap break record and doubts regarding their nature
501 (shorelines or shoal highs). It ranges between 40 ms and 90-80 ms, i.e. 52 to 123 m (velocity
502 law: 2580 to 2620 m/s). No tectonic structures of any wavelength are truncated by this
503 sequence boundary. Along the Mirande Subbasin (Fig. 3) the depocenters migrated
504 northward of the salt-controlled faulted anticlines (St-Medard, Auch).

505 The third order sequence boundaries show a well-recorded forced regressive (FR) wedge
506 for the intra-Priabonian (PC-SB-7) one and no forced regressive wedge but with a marked

507 onlap for the uppermost Priabonian (PC-SB-8) one. The amplitude of the intra-Priabonian
508 relative sea level fall is 150 ms, i.e. 188 to 191 m (velocity law: 2500-2550 m/s) and the
509 amplitude of the uppermost Priabonian is 60 ms (minimum estimation) i.e. 74 to 75 m
510 (velocity law: 2470-2500 m/s).

511 The Priabonian to Rupelian *depositional profile* (Table 1) is very similar to the Lutetian
512 profiles, with upstream alluvial fans (member 2 and 3 of the Palassou Fm and its westward
513 equivalent, the Jurançon Fm – Fig. 3 and 8) with, as a local base level, nearly flat alluvial to
514 coastal plains (Campagne and Agenais Fms). The carbonate platforms are made up of bio-
515 constructed mounds (or shoals – Siest Fm - Priabonian) or patchy reefs along a mixed
516 siliciclastic-carbonate ramp (Gaas and Mugron Fms – Rupelian).

517

518 *CT's Second order sequence (Chattian-base Tortonian – 27.1-10.6 Ma)*

519 The *age* of the base sequence boundary (SBct) is Early Chattian (27.1 Ma - top of the
520 Mugron Fm or base of the Escornebéou Fm – see supplementary material 1 for age
521 constrains). The age of the MFS (CT-MFS-15) is Late Burdigalian (17.5 Ma – intra Pontonx
522 Fm). At least eight main third order sequences were identified.

523 The *base sequence boundary* is topped by a stacking of third order sequences organized
524 in a large second order forced regression (FR) wedge, comprising at least two third order
525 sequence boundaries (CT-SB-12a – 26.4 Ma and CT-SB-12b – 25.2 Ma). The amplitude of the

526 relative sea level fall is difficult to estimate, once again due to the nature of the offlap break
527 (shoreline or shelfbreak). The maximum value from the shoreline (upward offlap break) to a
528 possible shelfbreak (downward offlap break) is 160 ms, i.e. 192 to 198 m (velocity law: 2400-
529 2470 m/s). The water depth of the shelfbreak can be deduced from the height of the shelf
530 clinofolds (topped by the shoreline), i.e. 60 to 62 meters (50 ms). The relative sea level fall
531 is therefore around 132 to 136 m. Along the Northern Pyrenean Front (line LR6, Fig. 3 and
532 'offshore-shore-parallel' line, Fig. 6), the structures (North-Pyrenean Front, downstream
533 thrusts and folds and salt-related anticlines – e.g. St-Médard, Auch, Landes High diapirs, etc.)
534 are truncated by the latest Chattian SB (CT-SB-12b – 25.2 Ma).

535 The Chattian to base Tortonian *depositional profiles* (Table 1) changed over time. Their
536 common characteristic is the occurrence of large marine embayments, corresponding to
537 mixed bioclastic-siliciclastic tide-dominated deposits ('faluns' as per French authors).

538 • Onshore the Early Miocene depositional profile (Table 1) is a large coastal to nearly flat
539 alluvial plains with carbonate lacustrine deposits ('Calcaires blancs de l'Agenais' Fm)
540 onlapping the SB southward (Crouzel, 1957 – Fig. 8).

541 • The Middle Miocene depositional profile (Table 1) is characterized onshore by braided
542 alluvial deposits ('Sables fauves' Fm - Gardere et al., 2002; Gardere, 2005) passing
543 upstream to alluvial fans. These fans are located downstream from the pre-Chattian
544 ones, south of the North Pyrenean Front thrust between these two generations of

545 fans (Fig. 8). Carbonate lakes (e.g. Auch and Astarac Fms) occurred in between or in front
546 of these alluvial fans (Crouzel, 1957).

547 The major environmental change of the 'Sables fauves' Fm corresponds to a significant
548 third order SB (CT-SB-16, 16.4 Ma), followed during uppermost Langhian times (CT-SB-17,
549 14.7 Ma) by a second one coeval with the fluvial valley incisions (Gardere et al., 2002;
550 Gardere, 2005).

551 The two offshore canyons (Fig. 1) are initiated (Cap Breton Canyon) or became active (Cap
552 Ferret Canyon) just after the last Chattian SB (CT-SB-12b, 12.5 Ma, i.e. at the end of the
553 forced regressive wedge (Fig. 6 and 7) feeding the so-called Cap Ferret deep-sea fan.

554

555 *TT's Second order sequence (base Tortonian-today – 10.6-0 Ma)*

556 The *age* of the base sequence boundary is well dated onshore at 10.6 Ma (base of the
557 'Argiles à galets' Fm at the boundary with the Montréjeau ('Molasse') Fm – see
558 supplementary material 1 for age constraints). Only the first progradational trend of the
559 sequence is preserved. At least six main third order sequences were identified.

560 The *base sequence boundary* (SBtt) is onshore (Fig. 3) an aerial unconformity and offshore
561 (Fig. 5) a SB topped by a lowstand normal regressive (LNR) wedge. The uncertain nature of
562 the offlap break (reasonably the shelf break) means that no relative sea level measurements
563 can be taken.

564 The base Tortonian to today *depositional profiles* (Tab. 1) changed over time. They are
565 characterized onshore by (i) low sediment preservation or aerial erosion and (ii) by a coastal
566 plain corresponding to the present-day Landes Forest and offshore by pure siliciclastic shelf
567 deposits. Along the alluvial plain the Tortonian to Early Pleistocene times correspond to the
568 growth of low preservation megafans (pebbly coarse-grained sands of the 'Argiles à galets'
569 Fm) with numerous evidences of by-pass periods (Fig. 8). The coastal plain (see 2.3) is a
570 stacking of low preservation fluvial sediments (coarse to medium-grained sands) and
571 marshes deposits (lignites). Although poorly dated, the Early to Middle Pleistocene is a
572 period of major incision of the alluvial systems with numerous incised valleys (Dubreuilh et
573 al., 1995).

574

575 *4.1.3. Interpretation: tectonic or eustatic controls of the second order sequences*

576

577 The *second order sequence boundary (SByp) of Late Ypresian age* (49.7 Ma - base of the
578 Lussagnet Fm) recorded (1) a relative sea level fall of 50 to 105 m, (2) no deformations with a
579 wavelength shorter than the size of the Aquitaine Basin and (3) a major change in the
580 sediment routing system with the brief growth of a megafan (Lussagnet Fm, Fig. 8).
581 According to Haq et al. (1987, 1988), the end Ypresian is characterized by a major short
582 eustatic fall (130-140 m), whereas Miller et al. (2005, 2008) quantified a quite minor one

583 around 40 m. Haq's value is too high compared to what was measured here and is therefore
584 questionable (see the comments on Haq's curve in 3.1). The Miller's accommodation
585 measurements provide a maximum value for this eustatic fall. Our values are higher than 40
586 m and therefore a tectonically-enhanced eustatic fall in response to basin-scale
587 deformations is the most reasonable explanation for this unconformity. This is supported by
588 the short perturbation (in time) of the sediment routing system with the growth of
589 megafans.

590 The *second order sequence boundary (SBpc) of the base of the Priabonian (37.6 Ma - base*
591 *of the Campagne Fm)* recorded (1) a relative sea level fall of 50 to 125 m and (2) a
592 reorganization of the subsiding areas. Haq's studies (1987, 1988) indicate a major eustatic
593 sea level fall of around 90 m, when Miller's work confirmed a major one of 70-80 m. This last
594 value falls within the range of our measurements and therefore a eustatic origin may be
595 inferred for this sequence boundary (SBpc). Nevertheless, the reorganization of the
596 subsidence pattern and relative sea level variations ranging up to 125 m, may once again
597 suggest tectonic forcing.

598 The *third order sequence boundaries of intra-Priabonian age (PC-SB-7)* recorded a relative
599 sea level fall of 190 m. This value is much higher than the eustatic variations of 30-40 m
600 proposed by Haq (1987) and a tectonic origin is proposed here.

601 The *third order sequence boundaries of uppermost Priabonian age (PC-SB-8)* recorded a
602 relative sea level fall of 75 m. In terms of age, this event occurred before the major sea level
603 fall of the Eocene-Oligocene boundary of around 100 m (Miller et al., 2008) due to the onset
604 of the Antarctic glaciation (transition to icehouse conditions). A tectonic origin is therefore
605 proposed for this SB.

606 The *second order sequence boundary (SBct) of Early Chattian age (27.1 Ma - top of the*
607 *Mugron Fm or base of the Escornebéou Fm)* recorded (1) a relative sea level fall of 190 to
608 200 m and (2) the late deformation stages of the North Pyrenean Front and related tectonic
609 structures which are no longer active beyond 25.2 Ma. This second point indicates a tectonic
610 control of this sequence boundary confirmed by the amplitude of the relative sea level fall
611 (190-200m) which is much higher than the eustatic variations occurring during Chattian
612 times (around 30 m for Haq and around 40 m for Miller). The truncation of all of the
613 structures, both major (North Pyrenean Front) and minor (salt-controlled thrusts and folds)
614 by the third Chattian SB (CT-SB-12b, 25.2 Ma) support a very long deformation period
615 initiated at the base of the forced regression wedge (SBct) i.e. from 27.1 Ma to 25.2 Ma.

616 The *second order sequence boundary (SBtt) of base Tortonian age (10.6 Ma - base of the*
617 *“Argiles à galets” Fm* is much younger than the major eustatic fall of Late Serravallian age
618 due to an increase in the volume of the Antarctic ice sheet (Zachos et al., 2001; Miller et al.,
619 2011); therefore, a tectonic origin can be assumed even though it was not possible to take

620 any measurements of the relative sea level fall here. The SBtt probably records the
621 paroxysm of a period of uplift (see 5.1) initiated during the uppermost Burdigalian (CT-SB-16,
622 16.4 Ma) at the time of the onset of the braided alluvial plain deposits of the 'Sables Fauves'
623 Fm.

624

625 *4.2. Sediment distribution through space and time: isopach maps*

626

627 *4.2.1. Description (Figs. 9 and 10)*

628

629 The *Palaeocene to Oligocene* (66-23 Ma – up to CTf7) isopach map (Fig. 9) shows a clear
630 difference between the Aquitaine Basin and the Bay of Biscay deep basin with a low
631 accumulation zone in between corresponding to the Landes Plateau offshore and the Landes
632 High onshore. In the Bay of Biscay deep basin, two main depocentres can be defined; (i) a
633 first one located north of the Le Danois Bank (up to 2500 m of sediments in 23 Ma) with few
634 sediments eastward toward the present-day Cap Ferret Canyon and (ii) a second one located
635 in the Armorican Subbasin northwest of the Gascogne Dome (up to 2000 m in 23 Ma). A
636 little patch of sediment is preserved in front of the present-day Cap Ferret Canyon. Seismic
637 data suggest a Late Eocene to Oligocene age for this patch. In the Aquitaine Basin/Landes
638 Plateau area two main domains can be defined according to the Pamplona Transfer Zone

639 and its poorly known possible northward prolongation toward the Aquitaine Basin.
640 Westward toward the Landes Plateau, three units can be recognized, from south to north, (i)
641 a major depocenter (up to 3500 m of sediments in 23 Ma) located south of the Cap Breton
642 Canyon, (ii) a low accumulation zone corresponding to the Landes Plateau and Landes High
643 and (iii) a medium accumulation zone (up to 2500 m in 23 Ma) around and north of the
644 Parentis Basin. Eastward along the present-day onshore Aquitaine Basin patchy main
645 depocentres occurred in the Arzacq, Tarbes and Carcassonne Subbasins. Low sediment
646 accumulations characterized the Audignon and Maubourguet Ridges. The amount of
647 sedimentation is very low north of the Aquitaine Basin, along a line that more or less
648 corresponds to the Celtaquitaine 'flexure', the former onlap of the Triassic salt deposits. In
649 between the two domains (Landes Plateau and High and present-day onshore Aquitaine
650 Basin) a north-south trending depocentre crossed the Thétieu Fault along the possible
651 prolongation of the Pamplona Fracture Zone.

652

653 The *Miocene to today* (23-0 Ma – from CTf7) isopach map (Fig. 10) again shows a major
654 difference between the Aquitaine Basin and the Bay of Biscay deep basin. In the Bay of
655 Biscay deep basin, a single depocentre (up to 3000 m in 23 Ma) is located at the intersection
656 between the mouths of the present-day Cap Ferret Canyon eastward and Torrelavega and
657 Santander Canyons southward. This depocenter extends as an east-west ribbon bounded by

658 two low sediment accumulations domains, the Gascogne Dome to the north and the
659 Cantabria Mountains to the west. In the Aquitaine Basin/Landes Plateau area two main
660 domains can be defined according to the Pamplona Transfer Zone and the Thétieu Fault.
661 Westward the Landes Plateau is made up of several patches of depositional and very low
662 depositional zones with sediment accumulations (up to 2000 m in 23 Ma) along (i) the
663 Cantabrian Margin, (ii) the axis of the Cap Breton Canyon, (iii) the Parentis Basin and (iv) the
664 axis of the Cap Ferret Canyon. A low sediment accumulation axis is located north of the front
665 the Basque-Cantabrian Basin. The south-eastern part of the South Armorican shelf was a low
666 accumulation area forming a southward thickening wedge of sediments. The present-day
667 onshore Aquitaine Basin was a quite low sediment accumulation domain (20 to 500 m in 23
668 Ma) with three main depocentres: (i) a wedge at the transition with the Landes Plateau in
669 the continuity with the one of the South Armorican Shelf, (ii) in the Tartas Subbasin north of
670 the Audignon Ridge and (iii) in the Tarbes Subbasin (up to 200 m in 23 Ma).

671

672 *4.2.2. Interpretation : sink preservation and lithosphere deformation*

673

674 From *Palaeocene* to *Oligocene* (66-23 Ma) times, the sediment thick accumulations
675 located at the front of the Pyrenees and Basque-Cantabrian folds and thrusts belts
676 correspond to foredeeps, i.e. the north Basque-Cantabrian, Arzacq, Tarbes and Carcassonne

677 depocenters. The occurrence of three north Pyrenean foredeeps (Arzacq, Tarbes and
678 Carcassonne) and the major role of the Pamplona Transfer Zone suggest a strong
679 segmentation of the foreland basin in good agreement with the numerical modelling of
680 Angrand et al. (2018) taking the Albian rifting inheritances into account. The meaning of the
681 Basque Cantabrian depocenter is debatable: does it represent the western end of the South
682 Pyrenean pro-foreland Basin or the merging of both forelands, i.e. the pro- and retro-
683 forelands? The palaeocurrent pattern (towards the S-SW) of the Jaizkibel Ypresian turbidites
684 (Kruit et al., 1972) located east of the Basques Massif (Fig. 9), may support this second
685 scenario with the deflection of the turbidity currents coming from the Arzacq foredeep along
686 the emerged Basques Massif, in good agreement with the palaeogeographic reconstructions
687 of Vacherat et al. (2017).

688 During Cenozoic times, the Landes High behaved as a rigid zone perturbing the flexural
689 response of the lithosphere due to mountain loading.

690 At shorter wavelengths, salt-related thrusts and associated anticlines or ridges (Audignon,
691 Maubourguet) started to be active, in good agreement with the Lutetian age of the onset of
692 these structures (Serrano, 2001 – see 2.3).

693 In the Bay of Biscay deep basin, the Santander transfer zone is active controlling the limit
694 between accumulating and non-accumulating domains. The deposits forming the patch of
695 sediments located northward of the transfer zone and westward of the present-day Cap

696 Ferret Canyon may be fed either by a proto-Cap Ferret Canyon or by proto-Torrelavega-
697 Santander Canyons during Late Palaeogene times. The depocentre located north of the Le
698 Danois Bank is interpreted here as the products of the erosion of the inverted Asturian Basin
699 (Le Danois Bank) during the Late Eocene to Oligocene (Gallastegui et al., 2002). The
700 depocenter of the Armorican Subbasin is controlled by the inversion of the lower Cretaceous
701 extensional blocks during Palaeocene and Upper Eocene times (Thinon, 1999; Thinon et al.,
702 2001, 2002) and fed by rivers coming from the uplifted Armorican Massif (Guillocheau et al.,
703 2003).

704
705 From *Miocene* to *today* (23-0 Ma) times, the thick sediment accumulation zones located
706 in front of the Pyrenean Belt during the previous time interval no longer existed, confirming
707 the end of the foredeep before 23 Ma (during the Lutetian, according to Serrano, 2001).
708 Nevertheless, some salt-controlled blind thrusts (Audignon Ridge) are still controlling a low
709 differential subsidence between the Arzacq and Tartas Subbasins (Fig. 9), which is discussed
710 later (5.2.1). The contour lines of the Tarbes Subbasin may be partly residual due to the
711 erosion and growth of the Lannemezan Plateau in Tortonian times (SBtt - see 4.1).

712 In the Bay of Biscay deep basin, the main depocentres corresponds to a major deep-sea
713 fan fed by both the Cap Ferret Canyon and the Torrelavega and Santander Canyons the last
714 canyon is supplied by the Cap Breton Canyon (initiated at the end of the Chattian, post CT-

715 SB-12b, i.e. post 25.2 Ma). The inverted structures of the Gascogne Dome and its westward
716 prolongation (Thinon, 1999, Thinon et al., 2001, 2002) have controlled the location of the
717 deep-sea fan. At that time, the Cantabrian Seamount is a submarine relief.

718

719 *4.3. First order evolution of the sediment preservation during Cenozoic times: a regional 2D*
720 *section*

721

722 *4.3.1. Description (Fig. 11)*

723

724 A regional E-W seismic line (Fig. 11) has been compiled from the onshore Aquitaine Basin,
725 north of the Lannemezan Plateau, to the Bay of Biscay deep basin. This section crosses the
726 Aquitaine platform north of the foredeep, the Landes Plateau and the Torrelavega-
727 Santander Canyons and reaches the deep-sea abyssal plain of the Biscay Basin.

728 At the first order the *Aquitaine platform* was prograding from Ypresian times (Serrano
729 et al., 2001) with 1400 to 700 m high clinoforms. This section confirmed (see 4.2) the
730 specific nature of the *Landes Plateau* that behaved as a quite low accumulation zone during
731 the Cenozoic. From the offshore Aquitaine Basin to the Landes Plateau, several pre-existing
732 salt diapirs of Cretaceous age were reactivated during Cenozoic times. The *Bay of Biscay*
733 *deep-sea plain* – not studied in details here – is composed of deep-sea fans and oceanic

734 currents deposits (contouritic mounds and ridges, sand-waves). The deep-sea fans initiated
735 during the end of the Palaeogene and active since the last Chattian SB (CT-SB-12b, 25.2 Ma),
736 are characterized by few channel deposits (rare occurrences on strike seismic lines) and are
737 mainly made up of stacked turbiditic lobes. Oceanic currents deposits were preserved from
738 the Late Chattian and became dominant along the slope from Messinian times.

739

740 This section can be subdivided into three main units bounded by major discontinuities
741 that correspond, in the continental record, to the subaerial unconformities (sequences
742 boundaries) of Chattian (27.1-25.2 Ma) and Early Tortonian (10.6 Ma) age (4.1) (see captions
743 in Fig. 11 for an age discussion).

744 • From *base Cenozoic to Chattian* times (66 to 27.1-25.2 Ma), most of the sediments were
745 stored in the Aquitaine platform with some in the Bay of Biscay deep-sea plain. At the
746 first order, the progradational wedge was mixed progradational-aggradational. The
747 offlap break was the shoreline. The clinoforms (slope: 1 to 1.5°; height up to 1400 m)
748 correspond to mixed carbonate-siliciclastic slopes to ramps and during the Late Ypresian
749 to deltas. The map of the Palaeogene sediment thickness (Fig. 9) suggests that the
750 sediments of the Bay of Biscay deep-sea plain were fed from the inverted and eroded
751 Asturian Basin (4.2.2). A high preservation on the Aquitaine platform and its
752 consequence, a low export toward the continental slope and the deep-sea plain, are in

753 good agreement with the high amount of condensation by downlap measured during

754 this time interval (see the Wheeler diagram, Fig. 8).

755 • From *Chattian to Early Tortonian* times (27.1-25.2 to 10.6 Ma), sediments were
756 distributed all along the depositional profile, with less sediment on the platform than in
757 the deep-sea basin. At the first order, the progradational wedge while progradational-
758 aggradational, was dominated by the progradation. The offlap break was either the
759 shoreline or shelf break. The clinoforms (slope: 1.5 to 2°; height: 1200-1400 m)
760 correspond to mixed carbonate-siliciclastic slopes. Since 25.2 Ma both Cap Ferret and
761 Cap Breton Canyons (see Fig. 1 for location) were actively transferring (laterally to the
762 section) sediments toward the Cap Ferret deep-sea fan.

763 • From *Early Tortonian to present-day* times (10.6 to 0 Ma), most of the sediments were
764 stored in the Bay of Biscay deep basin, with a very thin layer of continental sediments
765 preserved on the Aquitaine platform. At the first order, the progradational wedge is a
766 purely progradational one. The offlap break is the shelfbreak. The clinoforms (slope: 2.5
767 to 5.5°; height: 700-1000 m) correspond to siliciclastic slopes.

768

769 *4.3.2. Interpretation*

770

771 From the base Cenozoic to today, the sediment preservation, i.e. sink preservation,
772 changed through time with a first period when most of the sediments are preserved on the
773 platform, a second one when the sediment distribution is more well-balanced between the
774 platform and the deep basin with a slight imbalance in favour of the deep basin and a third
775 one when most of the sediments are preserved in the deep basin.

776 The sink preservation evolution observed here may be explained by the balance between
777 the accommodation space created by the subsidence [A_{sub}] and siliciclastic sediment supply
778 [S_{sc}] over tens of million years (Posamentier and Allen, 1993; Catuneanu, 2006). The first
779 order measurement of tectonic subsidence (Desegaulx and Brunet, 1990) indicates an
780 increasing rate from the Palaeocene (maximum rates: 2 to 9 m/Ma) to the Eocene
781 (maximum rates: 21 to 83 m/Ma) and then a decrease until the present day. Many
782 thermochronological studies document the main exhumation phase of the Axial Zone during
783 late Eocene-early Oligocene (Fitzgerald et al., 1999, Gibson et al., 2007, Fillon and van der
784 Beek, 2012, Dinclair et al., 2005) suggesting an increase of the erosion rate and of the
785 siliciclastic sediment volume feeding the basin.

- 786 • When $\Delta A_{sub} \leq \Delta S_{sc}$, most of the sediment are preserved on the platform and few
787 siliciclastic sediments are transferred and preserved in the deep-sea plain. The
788 stratigraphic pattern is aggradational ($\Delta A_{sub} = \Delta S_{sc}$) or progradational-aggradational
789 ($\Delta A_{sub} \leq \Delta S_{sc}$ with low differences between ΔA_{sub} and ΔS_{sc}).

- 790 • When $\Delta A_{\text{sub}} < \Delta S_{\text{sc}}$ (with significant differences between ΔA_{sub} and ΔS_{sc}), most of the
791 produced siliciclastic sediment are transferred to the deep-sea plain. Some sediments
792 may be preserved on the platform.
- 793 • When $\Delta A_{\text{sub}} \ll \Delta S_{\text{sc}}$, most of the produced siliciclastic sediments are crossing through the
794 platform (by-pass to low preservation) and are preserved in the deep-sea plain.

795

796 5. Discussion

797

798 5.1. Deformation causes sediment routing and sink preservation changes

799

800 The Aquitaine Basin recorded at least two wavelengths of deformation: (1) thrusts and
801 anticlines (ridges and domes) related to the Triassic salt decollement level, with a
802 wavelength of several tens to one hundred kilometers (medium wavelength) and (2) at the
803 basin-scale, i.e. with a wavelength of at least several hundreds of kilometers (long
804 wavelength).

805

806 5.2.1. Medium wavelength deformations: salt-related thrust and ridges

807

808 Thrusts and related salt diapirs and anticlines (ridges) were initiated quite early in the
809 retro-foreland evolution during the Lutetian (Serrano, 2001 – see 2.3) with a paroxysm of
810 shortening during the Priabonian (Rocher et al., 2000 – see 2.3) that might correspond to the
811 35.8 Ma SB (PC-SB-7, intra-Campagne Fm, intra-Priabonian) and the related uplift occurring
812 north of the Audignon Ridge. These deformations are truncated and sealed at 25.2 Ma (CT-
813 SB-12b).

814 Some seismic lines show evidence along salt-controlled anticlines of pure vertical
815 movements after 25.2 Ma (Fig. 3, e.g. St-Medard Anticline), i.e. after the end of the
816 shortening. Some authors (Rocher et al., 2000 onshore Aquitaine; Ferrer et al., 2012 offshore
817 Aquitaine) interpreted these structures as an indicator of the latest Pyrenean compression.
818 We interpret these structures as a result of differential sediment loading in response to high
819 sediment supply.

820

821 *5.2.2. Long to very long wavelength deformations*

822

823 Basin-scaled deformations – mainly regional uplifts - may have occurred during the Late
824 Ypresian (SByp) and occurred during the Chattian (SBct to CT-SB-12b) and Early Tortonian
825 (SBtt).

826 The possible *Late Ypresian* (49.8 Ma) basin-scale uplift may be related to the end of the
827 period of increase of mountain shortening and the incorporation of thicker portions of crust
828 in the collision belt, ranging from the Palaeocene to Early Eocene (Ypresian) as proposed by
829 Mouthereau et al. (2014) and Teixell et al. (2016) (see 2.2 and Fig. 12).

830 The *Chattian* forced regression wedge and its related SB (SBct to CT-SB-12b) record a
831 quite long lasting (27.1-25.2 Ma) basin-scale uplift. The last Chattian SB (CT-SB-12b, 25.2 Ma)
832 eroded and fossilized the North Pyrenean Front and the medium wavelength salt-related
833 thrusts and anticlines, and thus it records the end of the compression and then the evolution
834 from the syn-orogenic to post-orogenic period. This is in good agreement with the plate
835 kinematics data (Roest and Srivastava, 1991) indicating a stop of the convergence between
836 Eurasia and Iberia around the Oligocene-Miocene boundary (see 2.2).

837 The origin of the *Early Tortonian* deformations— an uplift with truncations of the Pyrenean
838 piedmont and a stop of the subsidence in the area of the present-day Landes, which is
839 clearly post-orogenic as indicated by the absence of compressive structures truncated by
840 this the Early Tortonian SB (SBtt), is probably at a longer wavelength than the Aquitaine
841 Basin. This unconformity is announced by the facies changes and the SB at the base of the
842 'Sables fauves' Fm deposited in a large braided alluvial plain, of base Langhian age (around
843 16 Ma). In Western Europe, this time interval corresponds to major uplifts (e.g. Ziegler,
844 1990; Ziegler and Dèzes, 2007; Carminati et al., 2009). In the French Massif Central, the

845 Middle to Late Miocene corresponds to (1) the emplacement of the Cantal strato-volcano at
846 13 Ma (paroxysm at 7.2 Ma, Nehlig et al., 2001) on top of a mantle anomaly (Granet et al.,
847 1995a,b; Barruol and Granet, 2002) and (2) to the first incision and then uplift of the Upper
848 Tarn (13 Ma, Ambert and Ambert, 1995) and Upper Loire (8.2 Ma, Defive et al., 2007). In the
849 Armorican Massif, the incision of river drainage filled by Late Tortonian to Messinian
850 sediments (Red Sands) recorded a massif-scale uplift (Guillocheau et al., 2003; Brault et al.,
851 2004). In southern Britain (Weald Basin), a major denudation occurred during Mio-Pliocene
852 times in response to a southern Britain-scale uplift (e.g. Jones, 1980). In southern Germany,
853 in the area located between the Rhine Graben and Bohemian Massif north of the Alpine
854 Foreland Basin, geomorphological studies of the stepped planation surfaces and related
855 scarps (Bremer, 1989) indicate a major low amplitude uplift of this area during Miocene
856 times (poorly dated). In conclusion, this brief but not exhaustive review of Western Europe
857 uplifts, suggest a major Western Europe-scale uplift during Middle and Late Miocene times.
858 Because of this very long wavelength (more than 1 000 km), this deformation might be
859 related to mantle dynamics coeval with the Alps formation.

860

861 *5.2.3 Sink preservation and sediment routing system in the Aquitaine/Bay of Biscay Basins*
862 *during Cenozoic times.*

863

864 Two alluvial systems composed the aerial part of the sediment routing system of the
865 Aquitaine Basin (see 4.1.1): (1) a nearly flat fluvial system (suspended-load and mixed
866 channels) with widespread lakes and marshes and (2) alluvial (mega)fans or braided alluvial
867 plains. The alluvial fans may be small to medium sized (several kilometers to several tens of
868 kilometers long from the upstream source to the downstream ultimate deposition –
869 Palassou and Montréjeau ‘Molasse’ Fms) or large ones (several tens to hundred kilometers-
870 the so-called megafans – Lussagnet Fm). Nearly flat fluvial to lacustrine systems, megafans
871 and large braided alluvial plains (‘Sables fauves’ and ‘Argiles à galets’ Fms) are connected to
872 the sea level, while small to medium-size alluvial fans are connected to local base levels
873 corresponding to the nearly flat fluvial to lacustrine systems.

874 The most intriguing unexpected result is the occurrence of nearly flat alluvial plains in a
875 foreland basin at the feet of growing up mountain belts. This raises the corollary question of
876 the existence of similar flat depositional topographies in other foreland basins. The South
877 Pyrenean pro-foreland basin evolved differently from its twin North Pyrenean (Aquitaine)
878 retro-foreland. One of the major differences is the closure and disconnection of the basin
879 from the sea at the time of uplift of the Basque-Cantabrian Mountains at 37Ma (Gomez et
880 al., 2002). Unfortunately, no or few widespread lacustrine systems have been described
881 during the exoreic phase of the foreland. The Swiss Molassic Basin (Homewood et al., 1986;
882 Berger et al., 2005) began as a deep basin with turbidites filled by deltaic progradations

883 (Lower Marine Molasse – Rupelian to Early Chattian). They passed upward into fluvial and
884 lakes deposits (Lower Freshwater Molasse – Early Chattian to Early Aquitanian), the local
885 base level of the large alluvial fans active up to the Middle Miocene. After a marine flooding
886 (Upper Marine Molasse – Late Aquitanian to Burdigalian) the basin is filled by lacustrine,
887 fluvial and alluvial fans deposits (Upper Freshwater Molasse – Middle Miocene). This
888 example also indicates the occurrence of lacustrine deposits as well as widespread marine
889 flooding, both suggesting quite low slope alluvial plains for the Swiss Molassic foreland
890 basin. This might suggest that the Aquitaine retro-foreland basin is not a unique case
891 example. Nevertheless, more sedimentological studies focussing on the palaeotopography
892 of alluvial plains are required for other foreland basins.

893

894 The Aquitaine retro-foreland basin from 50 to 16.4 Ma suggests an equilibrium between
895 accommodation space and sediment influx : nearly flat fluvial to lacustrine systems behave
896 as a local base level for the alluvial fans. This time span covered both the foreland stage and
897 first post-orogenic period.

898 This retro-foreland was never an overfilled basin (sensu Covey, 1986) filled by large
899 subsiding alluvial fans as expected by some stratigraphic models. As already mentioned, the
900 megafans described here ('Argiles à galets' Fm) initiated during the Late Miocene (10.6 Ma)
901 up to today, (1) resulted from a large uplift in response to a West European-scale

902 deformation and (2) recorded an overall sediment by-pass of the continental domain along
903 steeper slopes generated by the uplift, feeding the deep-sea plain of the Bay of Biscay. The
904 by-pass megafans do not represent the last 'overfilled' stage of the foreland evolution.

905

906 *5.3. Building a sink preservation model in the foreland basin from active to post-foreland*
907 *periods (Fig. 13)*

908

909 Based on the Aquitaine retro-foreland example and its outlet to the Bay of Biscay deep-
910 sea plain, we proposed a model for the evolution of the sink preservation in the foreland
911 basins connected to a passive margin, from their subsiding period to post-orogenic uplifts. In
912 this model, the foreland and upstream part of the margin (shelf and coastal plain) belong to
913 same subsiding domain. The depositional profile, parallel to the mountain belt, is a platform
914 on a continental crust passing to a continental slope and a deep-sea plain on oceanic crust.

915 The post-subsidence evolution of each foreland basin seems to be different (see the
916 Introduction). This is mainly due to the inheritance (structure of the upper crust, existence of
917 a decollement level(s), etc.) and the rate and amount of shortening. In the case of the Swiss
918 Molassic Basin, the post-foreland evolution (Schlunegger and Mosar, 2010; Willett and
919 Schlunegger, 2010) was characterized by the thrusting and uplift of the basin and the
920 formation of a new orogenic wedge (the Jura Mountains) in front of the former foreland. In

921 the northern Alps, the end of the foreland did not coincide with the end of the shortening as
922 in the case of the Aquitaine retro-foreland basin.

923 The key control factor is the balance between accommodation space created by the
924 flexural subsidence [A_{sub}] and the siliciclastic sediment supply [S_{sc}]. Possible effects of the
925 dynamic topography (e.g. Catuneanu, 2006) and therefore possible delays between the
926 flexural and dynamic subsidence responses were not taken into account. Similarly, the
927 effect of carbonate production was not considered here in the sediment budget. Three
928 stages are defined.

- 929 • Stage 1: *foreland period* (both foredeep/forebulge and basin propagation of salt-
930 controlled thrusts). When $\Delta A_{sub} \leq \Delta S_{sc}$ with low differences between ΔA_{sub} and ΔS_{sc} , the
931 sediments are stored on the platform and no deposition occurred from the distal
932 platform (condensation by downlap) to the deep-sea plain. Due to a slight imbalance in
933 favour of ΔS_{sc} , the first order platform geometry is progradational-aggradational.
- 934 • Stage 2: *post-foreland period 1 – subsiding platform*. When $\Delta A_{sub} < \Delta S_{sc}$ with significant
935 differences between ΔA_{sub} and ΔS_{sc} , most of the sediments are transferred to the deep-
936 sea plain with few preservations on the platform. The first order platform geometry is
937 progradational with a low aggradational component.
- 938 • Stage 3: *post-foreland period 2 – by-pass and/or uplift of the platform*. When $\Delta A_{sub} \ll$
939 ΔS_{sc} with $\Delta A_{sub} \leq 0$, all of the sediments are transferred to the deep-sea plain as deep-sea

940 fans. If $\Delta A_{\text{sub}} = 0$, the overall fluvial by-pass occurs on the platform and feeds pure
941 progradational wedges. If $\Delta A_{\text{sub}} < 0$, uplift and overall fluvial erosion occurs on the
942 platform and feeds forced progradational wedges.

943

944 In some foreland basins (e.g. Swiss Molassic Basin), stage 2 may be missing, with a direct
945 transition from subsiding foreland (stage 1) to uplifted basin (stage 3). This model does not
946 prejudice of the evolution of the shortening that may stop between stage 1 and 2 (case of
947 the Aquitaine retro-foreland) or can still be happening during stage 3 (case of the Swiss
948 Molassic Basin).

949

950 **6. Conclusion**

951

952 (1) *a new chronostratigraphic framework*: Four second order depositional sequences and at
953 least 24 third order cycles have been identified, and an age model based on a combination
954 of biostratigraphy, orbitostratigraphy and sequence stratigraphy with a time resolution of
955 0.1 Ma is proposed. From 50 Ma to today the duration of deposition, no deposition and
956 erosion periods were quantified.

957

958 (2) *dating and wavelength assessment of the main phases of deformation of the retro-*

959 *foreland from syn-orogenic to post-orogenic stages:*

960 • The end of the retroforeland activity and therefore the transition to a post-orogenic

961 setting has been dated to the Chattian, ranging from 27.1 to 25.2 Ma.

962 • During the orogenic period, the transition from a foredeep/forebulge system to

963 transported piggy-back basins occurred during Lutetian times. The shortening paroxysm

964 of this medium wavelength deformation occurred during Priabonian times around 35.8

965 Ma.

966 • The post-orogenic period is marked by a major uplift of the Aquitaine Basin from Late

967 Burdigalian (16.4 Ma) to Early Tortonian (10.6 Ma) in response to a possible mantle-

968 controlled West European-scale uplift.

969

970 (3) *a reconstruction of the successive depositional profiles and related depositional*

971 *topographies:* The type depositional profile up to the middle Miocene is a nearly flat coastal

972 to alluvial plain characterized by an alternation of laterally extensive lakes and marshes with

973 fine-grained fluvial channels. These nearly flat plains extended from the shorelines to the

974 feet of the Pyrenees where they played the role of local base levels for alluvial fans. Since

975 the Middle Miocene braided alluvial plains and low-preservation ('by-passing') megafans

976 replaced these nearly flat plains.

977

978 *(4) a three step evolution of the Cenozoic sedimentation of the Aquitaine Basin - proposal of*979 *a sink preservation model:*

- 980 • During the foreland period (foredeep then piggy-back) – here up to 25.2 Ma – when the
981 accommodation space created by the subsidence was balanced or slightly lower by/than
982 the siliciclastic sediment supply, most of the sediments are stored on the platform (here
983 the Aquitaine Basin). No sediments reached the deep-sea plain.
- 984 • During the post-foreland period (i.e. here at the end of the mountain belt shortening)
985 when the accommodation space created by the subsidence was lower than the
986 siliciclastic sediment supply and when the mountain belt reached its highest elevation
987 and erosion rate, most sediments are transferred and stored in the deep-sea plain of the
988 margin. Few sediments are preserved on the platform. In the case of the Pyrenees retro-
989 foreland, the Middle to Late Miocene West European-scale uplift enhanced this trend.

990

991

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993

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999

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1001

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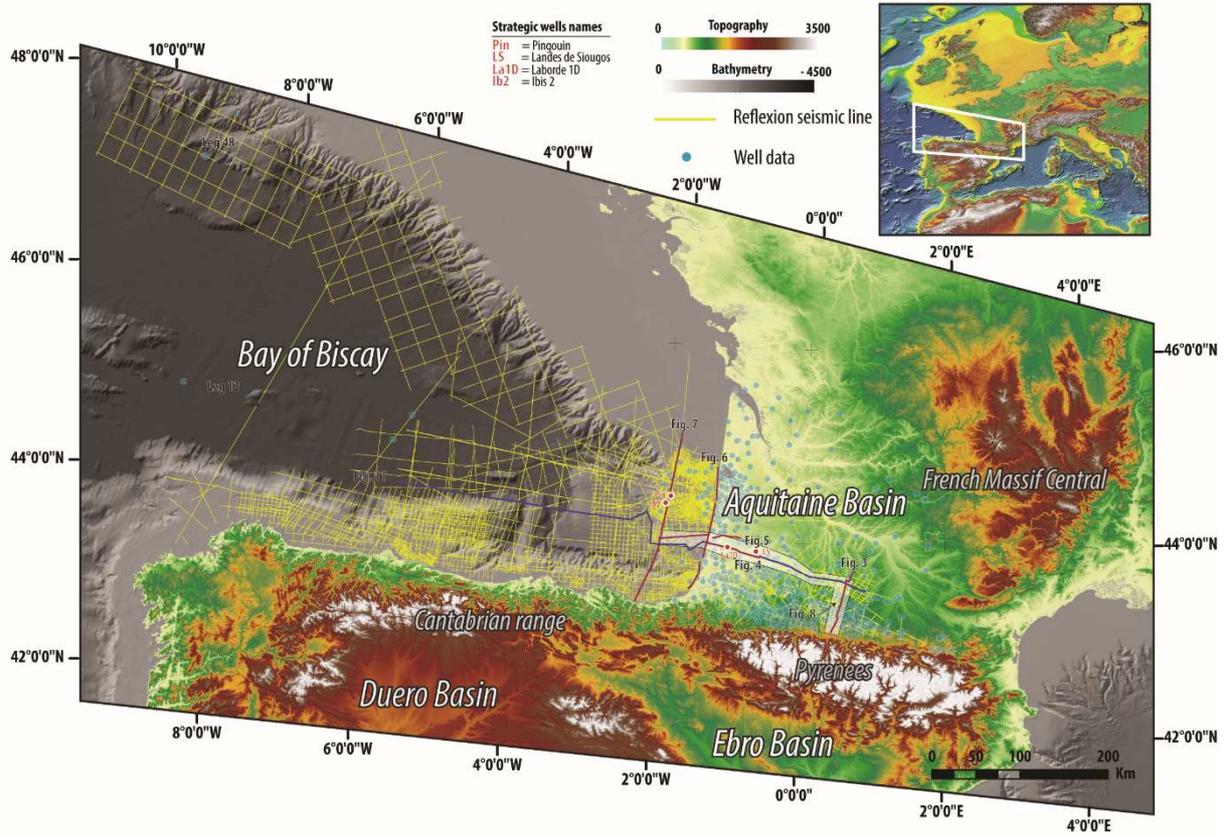
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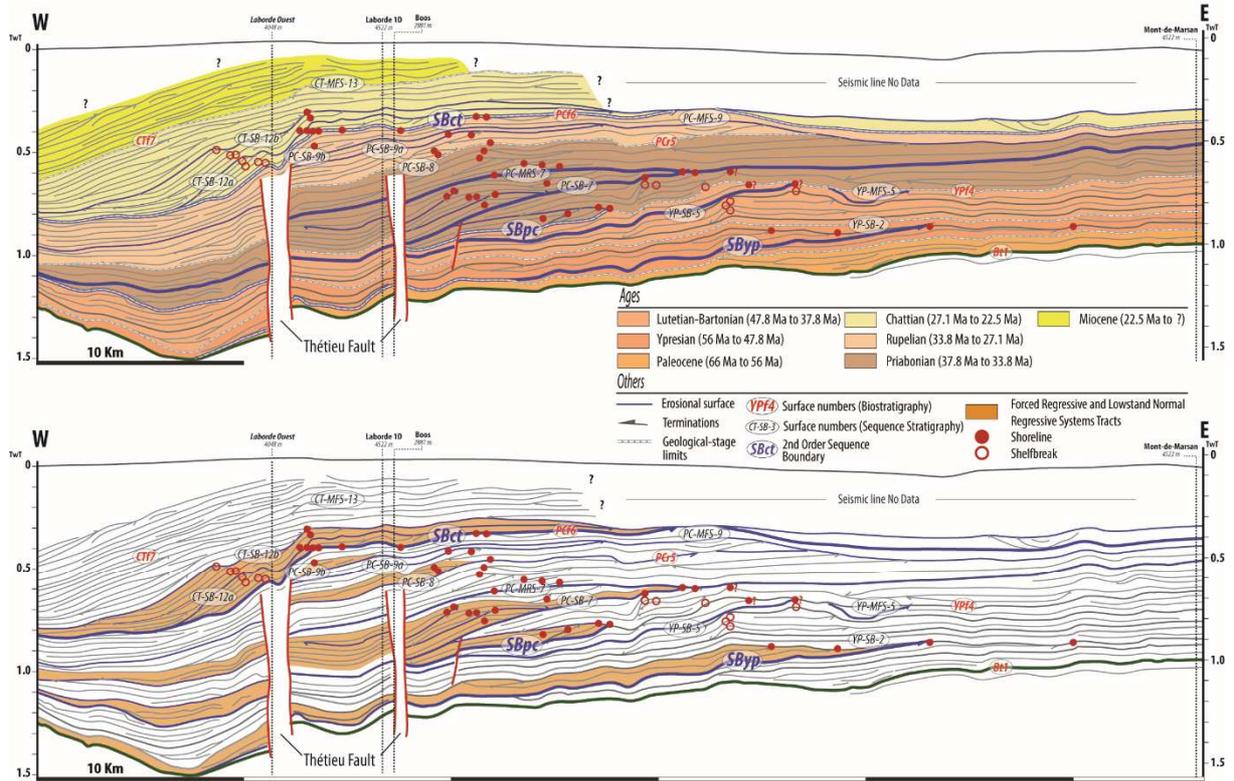
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- 1378

1381 **Fig. 1.** A: Location of the studied area in Europe. B: Main physiographic and structural
1382 features of the Aquitaine Basin, Landes Plateau and Bay of Biscay deep-basin.

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1383 **Figure 2**



1390 **Figure 4**

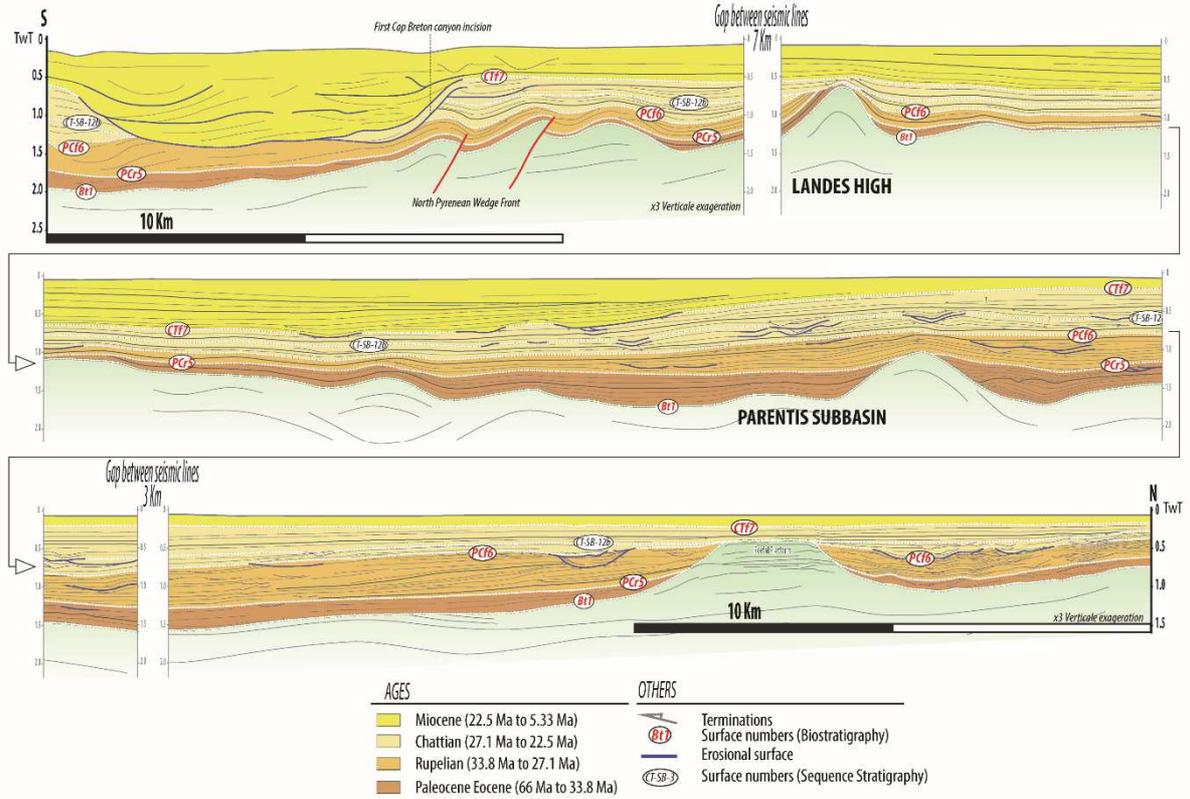
1391 **Fig. 4.** Sequence stratigraphic and structural interpretation of the onshore seismic line LR11

1392 (see Fig. 2 for location)

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1394 **Fig. 5.** Sequence stratigraphic and structural interpretation of the ‘offshore Mimizan Lake’

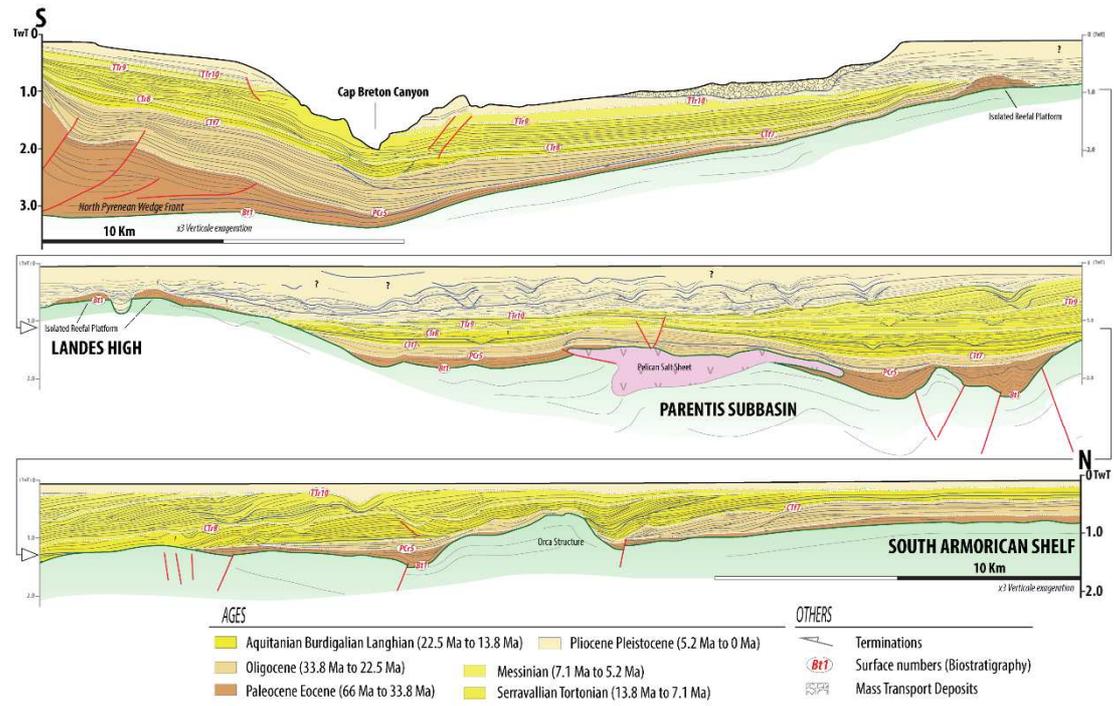
1395 seismic line (see Fig. 2 for location).



1396 Figure 6

1397 **Fig. 6.** Sequence stratigraphic and structural interpretation of the ‘offshore shoreline-

1398 parallel’ seismic line (see Fig. 2 for location).



1399 **Figure 7**

1400 **Fig. 7.** Sequence stratigraphic and structural interpretation of the ECORS offshore seismic

1401 line (see Fig. 2 for location).

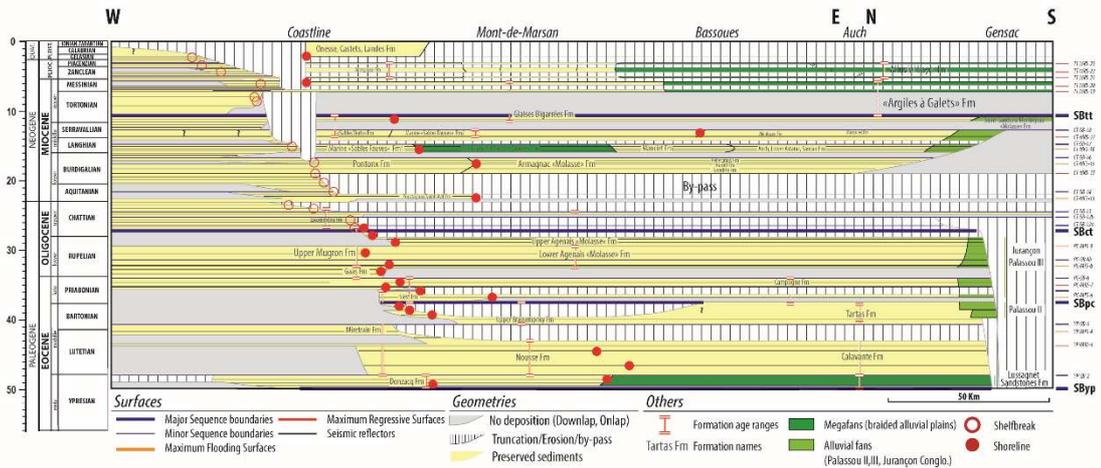
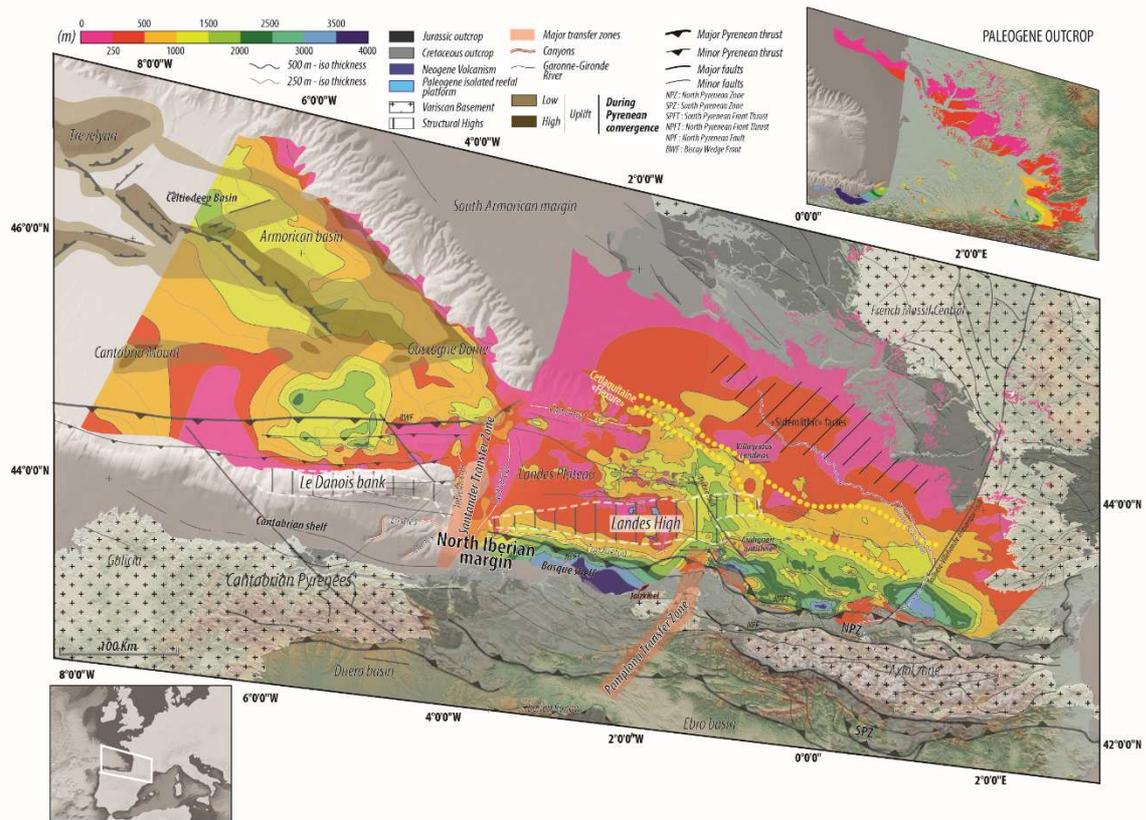


Figure 8

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1403 **Fig. 8.** Space-time stratigraphic (Wheeler) diagram of the Aquitaine Basin along a W-E-S

1404 transect from the near offshore to the Lannemezan Plateau.



1405 **Figure 9**

1406 **Fig. 9.** Sediment thickness (isopach) map of the Palaeogene (66-23 Ma)

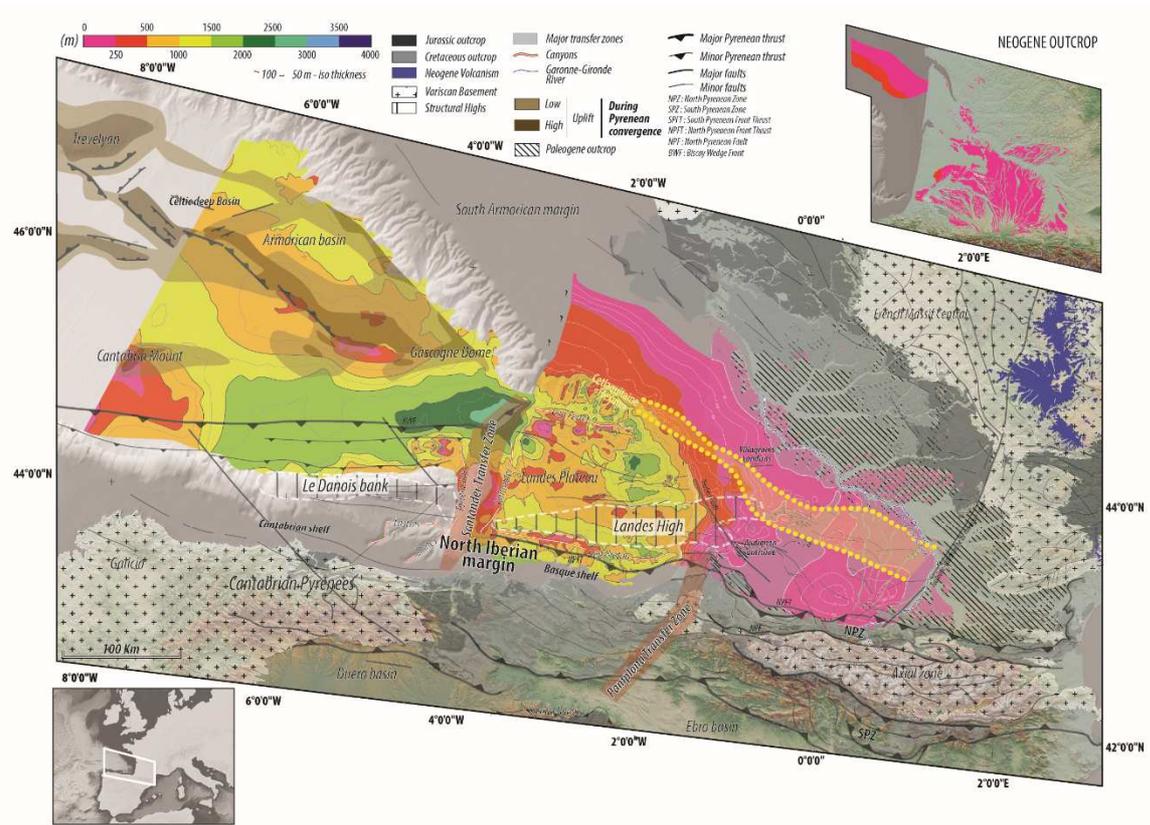


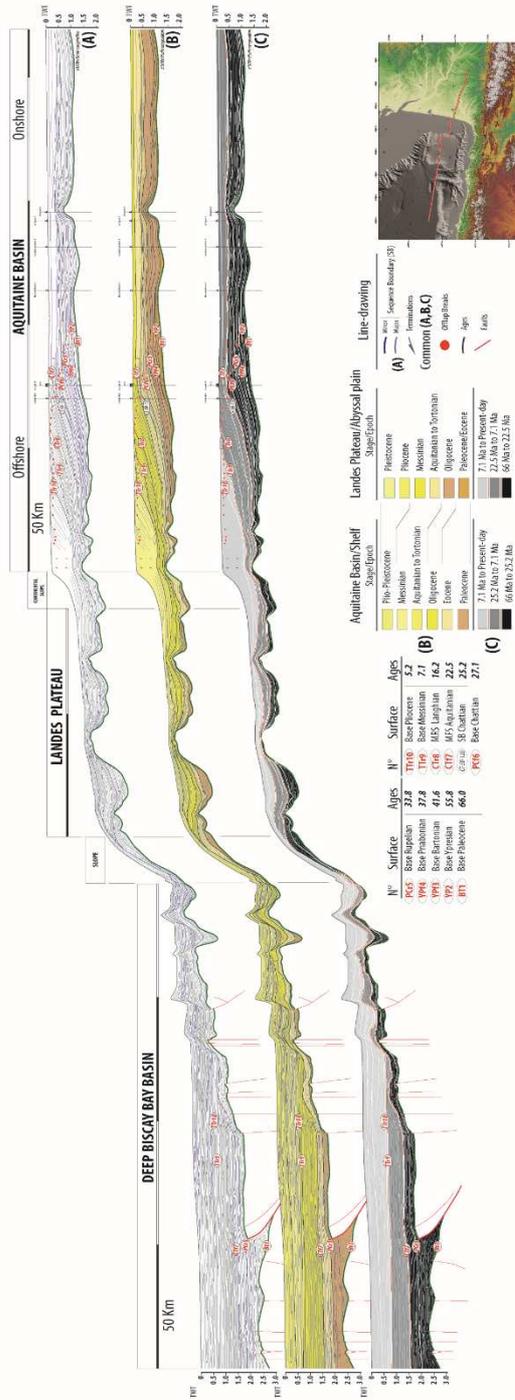
Figure 10

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1408 **Fig. 10.** Sediment thickness (isopach) map of the Neogene (23-0 Ma)

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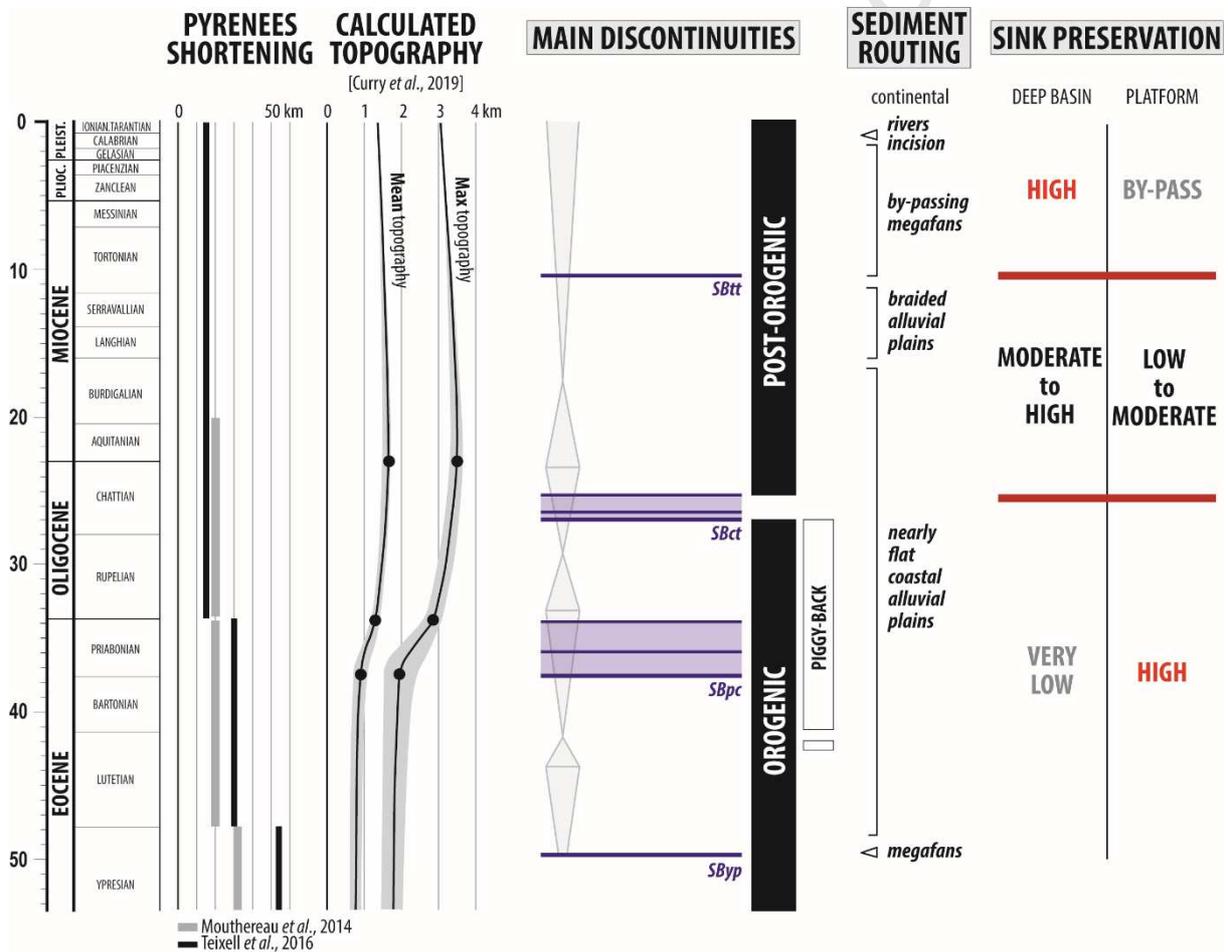


1411 **Figure 11**

1412 **Fig. 11.** East-West onshore-offshore regional seismic line from the Aquitaine Basin to the Bay

1413 of Biscay deep basin (see Fig. 2 for location). Due to the superimposition and truncation of

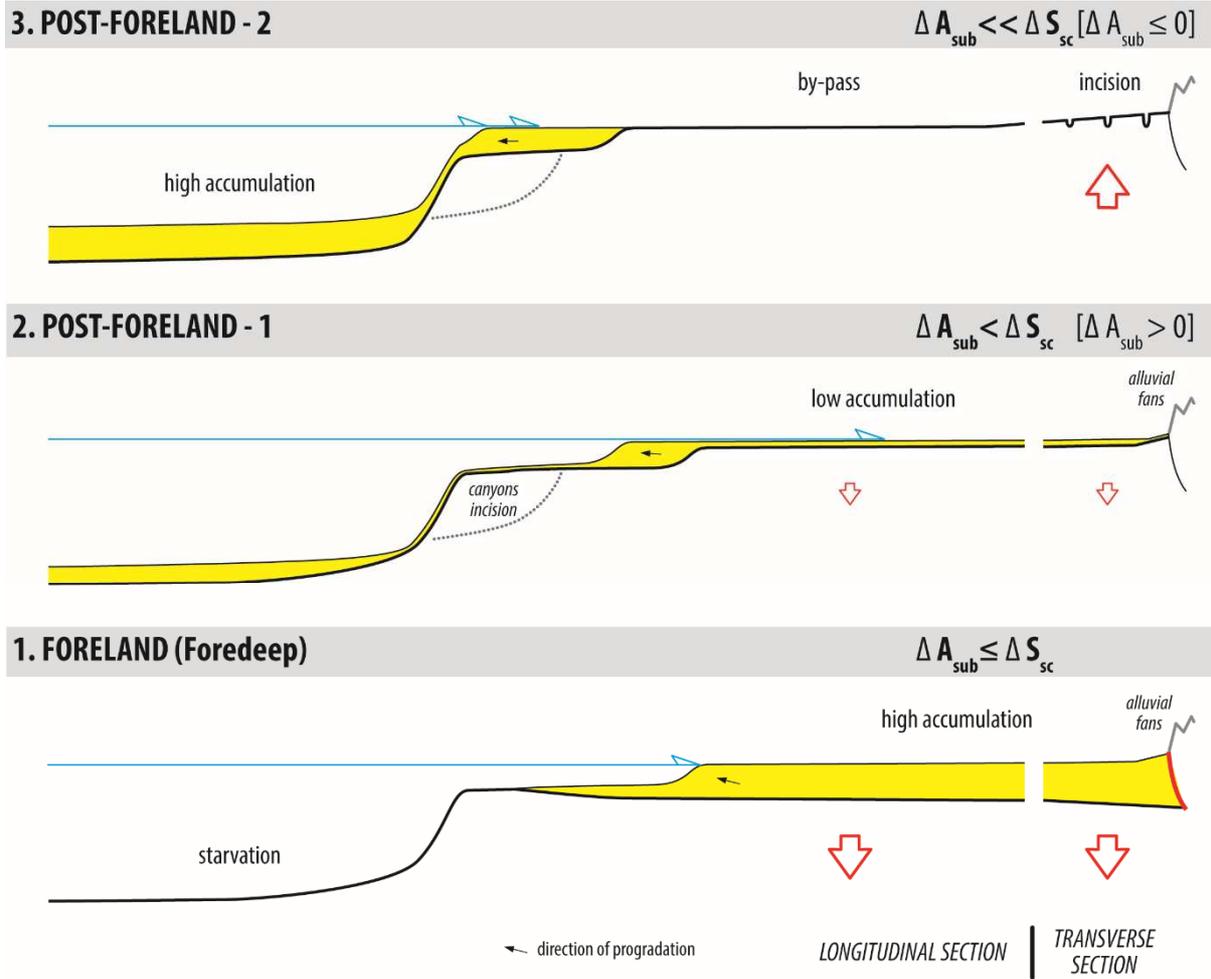
1414 successive erosional surfaces on the shelf and slope or poor time-resolution in the deep
 1415 basin, the two main discontinuities traced here do not strictly correspond to SBct and SBtt.
 1416 The first main discontinuity (equivalent to SBct) is the last Chattian SB (CT-MFS-12b)
 1417 truncating SBct on the shelf and slope and the MFS of the Aquitanian (CTf7) in the deep
 1418 basin. The second main discontinuity (equivalent to SBtt) is the base Messinian MRS (TTr9)
 1419 truncating the base Tortonian sequence boundary SBtt.



1420 **Figure 12**

1421 **Fig. 12.** Synthetic chart of the main events (deformation, topography, sediment routing) of

1422 the Aquitaine Basin to Bay of Biscay deep basin sedimentary system.



1423 **Figure 13**

1424 **Fig. 13.** A model of the sediment preservation and sediment routing system of the retro-
 1425 foreland basin passing laterally to a passive margin.

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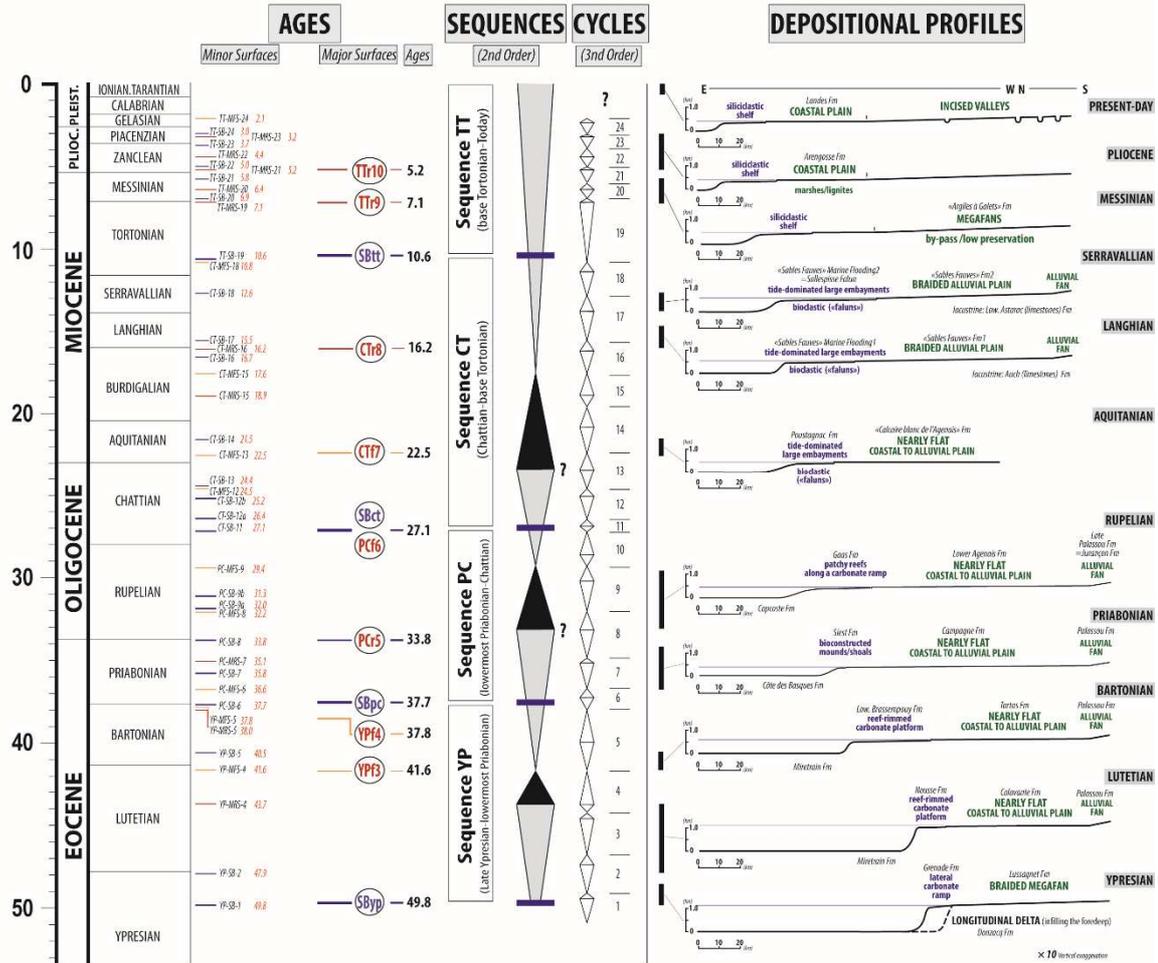


Table 1

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1429 **Table1.** Chronostratigraphic framework of the different orders of sequences and related

1430 surfaces (see supplementary material 1 for age constraints) and evolution of the

1431 depositional profiles on the Aquitaine platform.

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- A new chronostratigraphic and sequence stratigraphic framework
- Characterisation of successive deformations: age, wavelengths, causes
- Evolution of the sediment routing system: depositional profiles and topographies
- A sink preservation model based on the ratio vertical movements / sediment supply

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Declaration of interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: