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1 **SILICICLASTIC SEDIMENT VOLUMES AND RATES OF THE NORTH PYRENEAN RETRO-**
2 **FORELAND BASIN**

3

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33 **Abstract**

34 Sediment accumulation rates in foreland basins results from a complex interplay between
35 surface and deep processes in both the exhumed relief domain and sedimentary basins. The
36 growth and decay of a mountain belt during orogenic and post-orogenic phases have been
37 largely studied thanks to thermochronological and structural studies. The sedimentary
38 response of the orogenic phases in the preserved sediments of the surrounding basins is well
39 known in terms of sedimentary filling patterns and architecture, but much less better
40 quantified.

41 Here, we performed a measurement of the siliciclastic sediment volumes of the Pyrenean
42 retro-foreland basin – the Aquitaine Basin and the Bay of Biscay during Cenozoic times – for a
43 better understanding of the erosion and the sediment transfer and deposition during the
44 convergence (syn-orogenic) to post-convergence (post-orogenic) periods of the Pyrenees
45 Mountain. The measurement of the compacted siliciclastic sediment is based on sediment

46 thickness (isopach) maps of known lithologies, derived from the interpretation of 40 000 kms
47 of seismic profiles.

48 Thanks to the siliciclastic sediment volumes quantification and a well-known retro-foreland
49 basin tectono-sedimentary evolution, we bring quantitative results as:

50 (1) The amount of preserved sediments is of $51\,500 \pm 16\,800 \text{ km}^3$ for the Cenozoic.

51 (2) The siliciclastic sediment rate curve during Cenozoic shows two major increase around
52 26.0 Ma and 2.5 Ma. The 26.0 Ma increase is clearly related to the erosion of the
53 Pyrenees of tectonic origin. The major 2.5 Ma one would be mainly related to a climatic
54 forcing.

55 (3) The mass balance between the Aquitaine Platform and the deepest domains change
56 through time in favour of the deep domain. This might be explained by the ratio
57 between subsidence that created accommodation space and the sediments feed by
58 the mountain belt and stored in the Aquitaine Platform.

59

60 **Keywords:** accumulation rates, retro-foreland, Pyrenees, Aquitaine Basin, Bay of Biscay,
61 Cenozoic

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74 **1. Introduction**

75

76 The growth and decay of mountain belts reliefs result from feedbacks between
77 convergence rates, surface processes (Beaumont et al., 1992; Willett, 1999) and isostatic
78 response (Molnar and Lyon-Caen, 1988; Watts, 2001). The topography of a mountain belt is
79 first built by continental crust thickening during convergence, generating loading on the
80 lithosphere and its downward deflection at the origin of the surrounding foreland basins and
81 their vertical motion (Beaumont, 1981; Allen and Allen, 2013). Topographic evolution of the
82 mountain belt is also influenced by climate which can modify the erosion by surface processes
83 (e.g. Whipple, 1999). Finally, the base level variations in the foreland basins may play a
84 dominant role, on piedmont dynamics, in preserving reliefs (base level rise with emplacement
85 of endorheism, Babault et al., 2005) or enhancing its incision (base level fall by major capture
86 for instance, see Burbank and Anderson, 2009)

87

88 The estimation of denudation, i.e. the amount of eroded rocks, through time and its
89 controls by lithospheric deformation and/or climate (precipitation) are of primary importance
90 for understanding (1) the topographic growth and decay of mountain belts, but also (2) the
91 volumes and lithologies of the sediments produced and transferred to the surrounding basins.
92 The quantification of the denudation of a mountain belt is mainly based on low temperature
93 chronology (e.g. fission track and (U-Th)/He on apatite and zircons – ZFT, AFT, ZHe, AHe) that

94 provides a cooling history classically converted into exhumation with using thermal numerical
95 models (Braun, 2003 for instance, Ketcham, 2005; Gallagher et al., 2009). The sediment
96 production (volumes, petrography and grain-size) was performed on several foreland basins
97 (Molnar and England, 1990; Einsele, 2000; Najman and Garzanti, 2000; Goodbred and Kuelh,
98 2000; Schlunegger et al., 2001; Garzanti, 2019). In the Pyrenean domain, the P. Allens's group
99 has carried out numerous studies on these thematic, focus on the southern side (e.g. Michael
100 et al., 2013, 2014 a,b; Armitage et al., 2015).

101 The quantification of the siliciclastic sediment volume (sink) in foreland basins was first
102 performed at mountain-scale in the Alps (England, 1981, Kuhlemann, 2000; Kuhlemann et al.,
103 2002), but most of them concerns the latest deposition stages found in the ultimate sink: the
104 passive margin laterally connected to the foreland basins. This is the case for Asian mountains
105 (Métivier et al., 1999; Clift et al., 2006) or Rocky Mountains and the margins of the Gulf of
106 Mexico (Galloway and Williams, 1991; Galloway et al., 2011). They all show an increase of the
107 sedimentation rates up to today, a critical point discussed by P. Molnar (Molnar and England,
108 1990; Molnar, 2004) that might be related to global climate cooling (e.g. Herman and
109 Champagnac, 2016) and/or large erosion/recycling of the foreland deposits (Schlunegger and
110 Mosar, 2011; Leroux et al., 2017), linked frontal accretion, climate variations and drainage
111 network reorganization. Thus, the volumes distribution (storage, transfer, recycling) between
112 the foreland and the final sink at different stages of mountain belt evolution have not been
113 fully explored since the synthesis of Hinderer, 2012.

114

115 We propose in this study to measure the siliciclastic sediment volumes in a retro-
116 foreland basin where limited frontal accretion allows a more complete preservation of syn-
117 tectonic deposits – the Aquitaine Basin and the Bay of Biscay during Cenozoic times for a

118 better understanding of the erosion, the sediment transfer and deposition during the
119 convergence to post-convergence periods of the Pyrenees Mountains and discuss at the first
120 order the tectonic and climatic control factors. The measurement of the compacted siliciclastic
121 sediment volumes (for a solid rocks equivalence) are based on sediment thickness (isopach)
122 maps of known lithologies, built from an extensive dataset of seismic lines interpreted using
123 the principles of seismic stratigraphy and dated on wells (Ortiz et al., 2020).

124

125 **2. Geological setting**

126

127 The North Pyrenean retro-foreland and its lateral deep area of ultimate sediment
128 deposition – the deep Bay of Biscay Basin – is segmented into three geographical-geological
129 units (Fig. 1): (1) upstream, bounded westward by the Pamplona transfer zone, the Aquitaine
130 Platform (i.e. the modern onshore Aquitaine Basin and its shelf), (2) the Landes Plateau,
131 bounded westward by the Santander “soft” transfer zone, is a step (1000-1800 m deep)
132 bounded by slopes and canyons (Capbreton, Cap-Ferret) and (3) the deep Bay of Biscay (4000-
133 4500 m deep).

134

135 **2.1. Tectonic, relief and exhumation evolution of the Pyrenees**

136

137 The Pyrenees Mountains are a collisional belt bringing Eurasian lithosphere over the
138 Iberian lithosphere (e.g. Roure et al., 1989; Muñoz, 1992). Both Pyrenees and Basque-
139 Cantabrian Mountains result from the inversion and compression of the Albian hyperextended
140 Eurasian lithosphere (e.g. Fabriès et al., 1991, 1998; Lagabrielle et Bodinier, 2008; Jammes et
141 al., 2009; Lagabrielle et al., 2010; Masini et al., 2014; Clerc et al., 2012, 2016; Ducoux et al.,

142 2019; Saspiturry et al., 2019; Ducoux et al., 2021, Tugend et al., 2014). The relative motion of
143 Eurasia and Africa plates controlled the orogeny: the compression started at the time of
144 convergence of the two plates, i.e. at the end of the Santonian (83.6 Ma, Schettino and Turco,
145 2011) and ended around the Oligocene-Miocene boundary (chron 6c, Roest and Srivastava,
146 1991, Macchiavelli et al., 2017).

147 Thermochronological data also document the early mountain building with a cooling
148 phase at around 70 Ma (Whitchurch et al., 2011; Beamud et al., 2011; Mouthereau et al.,
149 2014; Ternois et al., 2019; Waldner et al., 2021). During Cenozoic, several studies resolve the
150 increase of exhumation during late Eocene-Oligocene times (Fitzgerald et al., 1999; Sinclair et
151 al., 2005; Gibson et al., 2007; Jolivet et al., 2007; Gunnel et al., 2009; Metcalf et al., 2009;
152 Whitchurch et al., 2011; Fillon and van der Beek, 2012; Bosch et al., 2016) following an earlier
153 uplift of the eastern range during Middle Eocene times (Maurel et al., 2008; Beamud et al.,
154 2011; Mouthereau et al., 2014; Vacherat et al., 2014; 2016; Ternois et al., 2019; Whitchurch
155 et al., 2011; Waldner et al., 2021). Finally, a late Miocene exhumation phase is recorded in the
156 Western Pyrenean Range by several authors (Jolivet et al., 2007; Bosch et al., 2016; Fillon et
157 al., 2021)

158

159 **2.2. Stratigraphic and tectonic evolution of the Aquitaine Basin and Bay of Biscay deep Basin**

160 Several studies have been carried out on the tectono-stratigraphic evolution of the
161 Aquitaine Basin since the late Cretaceous (Brunet, 1984; Desegaulx and Brunet, 1990; Brunet,
162 1994; Sztrákos et al., 1998; Serrano et al., 2001; Ford et al., 2016; Sztrákos and Steurbaut,
163 2017; Angrand et al., 2018). They identified two subsiding phases (Latest Santonian - Danian
164 and Thanetian – Oligocene) separated by the so called “Palaeocene quiet period”
165 characterized by a low to near zero subsidence phase (Ford et al., 2016). The first subsidence

166 phase has been largely influenced by the post rift thermal subsidence because of the short
167 time between the end of the rifting and the beginning of the convergence (around 10 Ma,
168 Angrand et al., 2018). Serrano et al., (2001) differentiate two distinct phases in the Western
169 Aquitaine Basin (1) from Palaeocene to Middle Eocene - compressional basin due to
170 lithospheric buckling (mixed system – East-West oriented turbidites and carbonates platform)
171 and (2) from Middle Eocene to Oligocene – foreland stage (progressive continentalization on
172 the eastern part and deltaic progradation toward the West).

173 Recently, through an extensive bio- and seismic stratigraphy study of both basins, Ortiz
174 et al. (2020) identified a set of several discontinuities of tectonic origins (fig. 2). The orogenic
175 period is subdivided into two steps (fig. 2): (1) the dominant lithosphere flexure with the
176 formation of foredeeps up to the uppermost Lutetian and (2) the propagation of the
177 deformation from the orogenic wedge through the basin mainly along salt-controlled thrusts
178 and anticlines locating subbasins. From the basin record point of view, the transition between
179 orogenic and post-orogenic periods is Chattian in age, from 27.1 to 25.2 Ma. This time interval
180 corresponds to the end of thrusting/inversions and a modification of the sediment
181 preservation pattern. The post-orogenic period is also subdivided into two intervals (fig. 2):
182 (1) a sharp decrease of the subsidence rates over the Aquitaine Platform, mainly filled by
183 continental sediments during early Miocene times (25.2-16.4 Ma) and (2) an uplift of the
184 Aquitaine Platform from 16.4 Ma to today characterized by a low-preservation domain up to
185 10.6 Ma. The main uplift phase occurred at 10.6 Ma and caused a generalized by-pass on the
186 Aquitaine Platform (Ortiz et al., 2020).

187

188 **3. Measurement of siliciclastic sediment volumes and rates**

189

190 The measurement of siliciclastic sediment (sink) volumes is based on four sediment
191 thickness (isopach of preserved sediments) maps: Palaeocene-Eocene (66-33.9 Ma - fig. 3),
192 Oligocene (33.9-23.03 Ma - fig. 4), Miocene (23.03-5.3 Ma - fig. 5) and Pliocene-Pleistocene
193 (5.3-0 Ma - fig. 6). These isopach maps are the products of a seismic stratigraphic analysis of
194 an extensive seismic line dataset (TotalEnergies, BRGM, BSS, fig. 1). This quite low time-
195 resolution is due to the absence of wells in the deep Bay of Biscay and the difficulty to extend
196 timelines from the Aquitaine Platform where several dated wells are available (see Ortiz et al.,
197 2020). For each interval the main lithologies proportion is known in the Aquitaine platform
198 and in the Landes Plateau (see Ortiz et al., 2020) and for the distal part of the Bay of Biscay,
199 we used the studies available in this domain about the Cenozoic sedimentary record (Cremer,
200 1983; Iglesias, 2009; Brocheray et al., 2014).

201 The sediment volumes measured from the isopach maps were later decompacted and
202 the amount of sediments produced in the basin (here mainly carbonates) was removed to get
203 solid rock volumes. We used the protocol and code published by Guillocheau et al. (2012) that
204 consider uncertainties or ranges for (1) time-depth conversion parameters, (2) absolute ages
205 of the different horizons, (3) carbonate content and (4) surface porosity and e-folding depth
206 for porosity decrease. During Cenozoic times the deep-sea Bay of Biscay fan fed through the
207 Capbreton and Cap-Ferret Canyon by the sediments coming from the Aquitaine Basin has
208 progressively migrated westward (Cremer, 1983; Iglesias, 2009; Brocheray et al., 2014), our
209 calculation area is evolving accordingly (dashed line on figs, 3, 4, 5, 6). In order to compare the
210 volumes, they will be presented through the evolution of the ratio between Aquitaine Basin
211 volume (V_{On}) and Bay of Biscay volume (V_{Off}) as follow: $R_{OnOff} = V_{On}/V_{Off}$.

212

213 **4. Isopach maps: results and interpretation**

214

215 **4.1. Paleocene-Eocene (66-33.9 Ma)**

216

217 The *Paleocene-Eocene* (66-33.9 Ma - Fig. 3) time interval is characterized by sediment
218 preservation along the North Pyrenean Thrust on the Aquitaine Platform, with three subsiding
219 domains - 10-50 km large - filled by 3500 m of sediments accumulated in 32 Ma. Two less
220 subsiding domains were located north of the previous ones and bounded by two ridges with
221 very low accumulation. Few sediments were preserved along the Landes Plateau. In the deep
222 Bay of Biscay, two depocenters occurred south of the South Armorican Margin and north of
223 the Biscay Wedge Front.

224 *Interpretation:* The three subsiding domains correspond to a segmented foredeep
225 (Ortiz et al., 2020) controlled by the inherited structures from the Early Cretaceous rifts history
226 (Angrand et al., 2018). These domains are mainly filled during the first orogenic phase (up to
227 uppermost Lutetian) and the small depocenter in the northern part of the ridges correspond
228 mainly to the second orogenic phase, when the deformation is propagated basinward. The
229 two deep depocenters of the Bay of Biscay are disconnected from the Aquitaine Platform and
230 bounded by the inverted extensional structures of the Gasconne Dome Bulge (Thinon, 1999;
231 Thinon et al., 2001). The first depocenter located in the Armorican subbasin, was fed by
232 sediments coming from the Loire River (fig. 1) or from the Armorican Massif (Guillocheau et
233 al., 2003).

234

235 **4.2. Oligocene (33.9-23.03 Ma)**

236

237 The *Oligocene* (33.9-23.03 Ma - fig. 4) was a period of more widely distributed
238 sediments accumulation along the three domains. In the Aquitaine Platform depocenters are
239 located (i) on both sides of salt-related controlled thrusts forming anticlines or ridges
240 (Maubourguet-Antin, Audignon) or (ii) along N140° faults (e.g. Téthieu Fault, location on
241 figures 1 and 4). In the south Landes Plateau, north of the Basque-Cantabrian Mountains, a 60
242 km-large depocenters occurred. In the deep Bay of Biscay, a NW-SE trending depocenter is
243 located south-west of the inverted structures of the Gascogne Dome Bulge and north of the
244 Biscay Wedge Front.

245 *Interpretation:* Oligocene is a period during which (1) segmented foredeep was no
246 longer subsident and (2) compressive deformation propagated from the orogenic wedge to
247 the retro-foreland basin along salt decollement levels localizing little subbasins (Serrano,
248 2001; Ortiz et al., 2020). The depocenters of the south Landes Plateau may be due to the
249 loading effect of the inverted and thrustured Basque Cantabrian Margin starting at 37 Ma and
250 ending at 28.5 Ma (Gomez et al., 2002). The depocenter of the deep Bay of Biscay is
251 disconnected from the Aquitaine Platform and moderately fed by the erosional product
252 coming from the inversion and deformation of the Cantabrian Margin and the uplift of the
253 Cantabrian Mountains from Late Eocene to Oligocene times (Gallastegui et al., 2002; Pedreira
254 et al., 2015; Fillon et al., 2016).

255

256 **4.3. Miocene (23.03-5.3 Ma)**

257

258 The *Miocene* (23.03 to 5.3 Ma – fig. 5) recorded a major change in the sediment
259 distribution. The Aquitaine platform was henceforth a place of low sediment preservation
260 (less than 200 m for 17.5 Ma). Most of the sediments were transferred and stored along the

261 Landes Plateau and the deep Bay of Biscay. Two major sediment accumulations were active
262 in the Landes Plateau, (1) to the north, a progradational wedge (Ortiz et al., 2020) located at
263 the transition between the southeastern part of the South Armorican shelf and the Landes
264 High and (2) to the south, an E-W trending depocenter located north of the Capbreton canyon.
265 The preserved sediments along the deep Bay of Biscay are organized as an E-W trending body
266 more than 100 km-large located on the Biscay Wedge Front. It is upstream in connection with
267 the two depocenters of the Landes Plateau.

268 *Interpretation:* The Aquitaine Platform was no longer a significant subsiding domain.
269 Most of the sediments bypassed to deepest areas: (1) through deltaic progradational wedges
270 on the Landes Plateau, (2) through canyons and at the end of the system (3) as deep-sea fans.
271 The depocenters located north to the Capbreton canyon correspond to the northern levee of
272 the canyon already active at this period (Cremer, 1983; Iglesias, 2009, Ortiz et al., 2020).
273 Progradational deltaic wedge accumulate at the head of the Cap-Ferret canyon (western
274 extension of the north Landes Plateau depocenter – Cremer, 1983) and downstream (East of
275 the deep Bay of Biscay depocenter), an erosion domain where the canyon is active. These two
276 canyons domains fed downward, the Cap-Ferret deep-sea fan (Cremer, 1983; Iglesias, 2009)
277 in the deep Bay of Biscay.

278

279 **4.4. Plio-Pleistocene (5.3-0 Ma)**

280

281 The *Pliocene-Pleistocene* (5.3-0 Ma – fig. 6) time-interval shows an accentuation of the
282 sediment distribution pattern established during Miocene. (1) No to very few sediments were
283 preserved on the Aquitaine Platform. (2) Two depocenters occurred on the Landes Plateau: (i)
284 a N-S trending one north to the Landes High (progradational wedges – Ortiz et al., 2020) and

285 (ii) a N-S trending one south of the Landes High. (3) The depocenters of the deep Bay of Biscay
286 are located (i) at the base of slope of the South Armorican shelf and (ii) along the Biscay Wedge
287 Front in connection with the northern depocenter of the Landes Plateau.

288 *Interpretation:* The Aquitaine Plateau was upstream uplifted and incised by rivers
289 (Mouchené et al., 2017; Ortiz et al., 2020). All the products of erosion are transferred to the
290 deepest domains with by-pass to very low sedimentation along the modern Landes Forest
291 (Dubreuilh et al., 1995). The prograding sedimentary wedge is now located at the border of
292 the Aquitaine shelf and the two canyons (Capbreton and Cap-Ferret) alternated periods of
293 erosion and deposition (Cremer, 1983; Iglesias, 2009).

294

295 **5. Measurement of the siliciclastic sediment budget: results (fig. 7)**

296

297 In the deep-sea plain, siliciclastic volume measurements were only performed in the
298 south deep Bay of Biscay Basin, south of the Gascogne Dome Bulge, the sediment located
299 northward (Armorican Subbasin) being fed from the Loire River (fig. 1) and the Armorican
300 Massif (see 4.1). In addition, during Cenozoic times the deep-sea Bay of Biscay fan fed through
301 the Capbreton and Cap-Ferret Canyon by the sediments coming from the Aquitaine Basin is
302 progressively migrating westward (Cremer, 1983; Iglesias, 2009; Brocheray et al., 2014). This
303 is the reason why, our calculation area is evolving according to the deep-sea fan location (Figs,
304 3, 4, 5, 6).

305

306 The total amount of siliciclastic sediments (Tab. 1, Fig. 7) deposited from the Aquitaine
307 Platform to the south deep Bay of Biscay during Cenozoic times is 51 500 km³ with an
308 uncertainty of 16 800 km³, i.e. 33%.

309

310 The four-time intervals here defined, have not the same duration and therefore
311 siliciclastic sedimentation rates had to be used for a comparison between time intervals
312 instead of volumes (Tab. 1, fig. 7). At first order the sedimentation rates are increasing from
313 535 km³/Myr (between 66 and 33.9 Ma) to 3150 km³/Myr (from 5.3 to 0 Ma). Oligocene and
314 Pliocene-Pleistocene seems to be significant instants of siliciclastic sediment rates increase.

315

316 The siliciclastic sediment budget comparison (Tab. 1, fig. 7) between the onshore
317 (Aquitaine Basin) and offshore (Aquitaine shelf, Landes Plateau, south Bay of Biscay) shows a
318 shift, trough Cenozoic times, from onshore to offshore preferential accumulation since
319 Oligocene-Miocene boundary. The sediment ratio between onshore and offshore (R_{OnOff}) is (i)
320 25.3 from 66 to 33.9 Ma, (ii) 7.5 from 33.9 to 23.03 Ma, (iii) 0.18 from 23.03 to 5.3 Ma and (iv)
321 0.01 from 5.3 to 0 Ma.

322

323 **6. Discussion on the siliciclastic sediment budget**

324

325 *Volume comparison with South Pyrenean Foreland and others orogenic adjacent basins* - The
326 amount of siliciclastic sediments preserved in northern foreland is $51\,500 \pm 16\,800$ km³ for
327 Cenozoic times. A comparison with the sediment volumes of other orogenic systems
328 supposes to get a sediment budget for both sides of the Pyrenees, i.e. including the South
329 Pyrenean Foreland basin. Several studies have been conducted for quantifying the preserved
330 sediment volumes in the South Pyrenean sink (i.e. Ebro and Valencia Basins)(Nelson, 1990;
331 Garcia-Castellanos et al., 2003; Babault et al., 2006; Filleaudeau, 2011; Arche et al., 2010;
332 Watts and Torné, 1992). The Ebro Basin has been under endoreic conditions since 36.2 Ma

333 (Costa et al., 2010) and then opened towards the Mediterranean Sea between 12 Ma and
334 7.5 Ma (Garcia-Castellanos et al., 2003; Fillon et al., 2012; Garcia-Castellanos and
335 Larrasoana, 2015). The total amount of preserved sediments in the Ebro and Valencia Basin
336 is 260 300 km³ since 66 Myr. For a better sediment budget on both sides of the Pyrenees,
337 only the sediments coming from the mountain belt should be considered, i.e. the total
338 Cenozoic preserved sediments volumes in the Ebro basin and the Neogene for the Valencia
339 Basin. This corresponds to a range between 78 000 km³ (Filleaudeau, 2011) and 110 000 km³
340 ± 5000 km³ (Garcia-Castellanos et al., 2003) in the Ebro Basin and 92 000 km³ in the Valencia
341 Basin (Garcia-Castellanos et al., 2003). This gives a total of preserved sediments between
342 170 000 km³ and 202 000 km³. At the first order, this result shows a factor of four between
343 the northern retroforeland (our study) and the South Pyrenean proforeland.

344 The first-order estimation of total volume preserved in the Pyrenean adjacent basins
345 (i.e. Aquitaine Basin, Bay of Biscay, Ebro Basin, Valencia Basin) is between 220 000 km³ and
346 250 000 km³. The total volume of siliciclastic sediments produced by the Alps since Oligocene
347 times is of 890 000 km³ (220 000 km³ from the Eastern Alps and 670 000 km³ from the western
348 ones- Kuhlemann et al., 2002) i.e. four times more than the Pyrenees. However, for larger
349 orogens the order of magnitude changes with 3.35 10⁶ km³ of sediment produced by the Rocky
350 Mountains on its continental part and stored in the Gulf of Mexico (Galloway et al., 2011) and
351 the 4.95 10⁶ km³ by the western side of the Himalaya and accumulated along of the Indus
352 Delta (Clift et al., 2006).

353

354 *Volume comparison with Pyrenean thermochronological data* – We present here some
355 volumes estimation for the Cenozoic period based on published themrochronology data in the
356 Pyrenees.

357 - Too few data (AFT and AHe) provide Palaeocene to early Eocene ages to obtain a
358 consistent exhumation trend across the entire Pyrenean range.

359 - During the late Eocene/lower Oligocene (from 37 to 30 Ma), orogenic exhumation (>
360 0.5 km/Myr) is recorded in Central and Eastern Pyrenean Massif (Maladetta, Noguères,
361 Canigou) according to studies from Fitzgerald et al., (1999), Gibson et al., (2007), Fillon and
362 Van der Beek, (2012) and Gunnell et al., (2009). These authors modeled an exhumation rate of
363 1.5 to 4 km/Myr. Considering an average of 2 km/Myr integrated over the whole area
364 (inducing a strong assumption of a continuous constant exhumation) during the period of 35-
365 30 Ma, the eroded volume should be of 50 000 km³. This value is very high compared to what
366 has been measured in the Aquitaine Basin during the same period (16 500 km³ from base
367 Cenozoic to base Oligocene). This volume seems therefore to supply mainly the Ebro basin
368 with a significant increase observed at the base of Oligocene (i.e. 33.9 Ma) (Filleaudeau, 2011).

369 - Since middle Oligocene times, a moderate exhumation (between 0.1 and 0.5 km/Myr)
370 is proposed by several authors in Western, Central and Eastern massifs of the Pyrenean range
371 (DeFelipe et al., 2019; Bosch et al., 2016; Fillon et al., 2021; Fitzgerald et al., 1999; Gibson et
372 al., 2007). An average exhumation rate of 0.25 km/Myr across the entire range provides an
373 estimation of 17 500 km³ for the Middle to Late Oligocene period. This volume is, like during
374 Late Eocene/Lower Oligocene, one order of magnitude higher than what we measure in the
375 Aquitaine Basin. This volume supply mainly the Ebro basin where the Oligocene volumes are
376 very large (around 45 000 km³ in Filleaudeau, 2011).

377 - Finally, the important increase of exhumation rate recorded in the Western Pyrenees
378 (see Fillon et al., (2021) for the data synthesis and modelling) infers a volume estimation of
379 2 000 km³ for the late Miocene period. This quantification is low compared to the volumes of
380 preserved sediments.

381 We draw attention to the fact that these estimation are produces from an average of
382 the results of modelling data done in 1D and extrapolated on a surface. These volumes based
383 on thermochronological data are probably overestimated.

384

385

386 **6.1. Age estimation of siliciclastic sediment flux increase from interpreted seismic data**

387

388 As we mentioned earlier, the low stratigraphic resolution (Palaeocene-Eocene,
389 Oligocene, Miocene and Pliocene-Pleistocene) is due to difficulties to propagate the high
390 resolution age model (established in shallow marine setting in the Aquitaine Platform and
391 Landes Plateau – Ortiz et al., 2020) across the whole seismic survey in the deep Bay of Biscay
392 Basin and continental Aquitaine series. In order to be more precise about the age of the
393 siliciclastic rate increase during the two key periods (Oligocene and Plio-Pleistocene) and their
394 possible links to climatic or geodynamic event, we performed a 2D quantification on the cross
395 section published in Ortiz et al., (2020). This quantification aims at comparing on three
396 different locations (proximal Aquitaine platform – deep Bay of Biscay Basin – progradational
397 wedges in between) the percentage of 2D surface (comprising sediment) normalized on their
398 duration between each high-resolution timeline.

399 The figure 8 present the results and shows a refined estimate of increase at 26.0 ± 2.0
400 Ma for Oligocene period and 2.5 ± 0.5 Ma for Pliocene-Pleistocene period.

401

402 **6.2. The 26.0 ± 2.0 Ma increase**

403

404 The measured Late Oligocene (26.0 ± 2.0 Ma) siliciclastic rate increase could be
405 explained by both a tectonic factor – the accretion of Pyrenees basal units – and a climatic
406 factor – the icehouse-greenhouse transition and the late Oligocene warming (Zachos et al.,
407 2001).

408 *Tectonic* - A Late Eocene to Oligocene/Miocene paroxysm of denudation (37-29 Ma
409 central Pyrenees and 37-20 Ma western Axial Zone) is measured by the numerous
410 thermochronological studies (see references in section 2.1) and by the dated active tectonics
411 mainly in the Southern Pyrenees (Montsec and Sierras exteriores thrust systems associated to
412 Orri and Rialp thrust sheets, see Munoz, 1992, Teixell, 1998, Vergés et al., 2002 and
413 Mouthereau et al., 2014), or in the Jaca Basin cover thrust and associated Gavarnie and more
414 frontal thrusts (Labaume et al., 2016). During late Oligocene time, the retro-foreland is
415 affected by several truncation/erosional surfaces from 27.1 to 25.2 Ma (Ortiz et al., 2020).
416 During this event, some syn-orogenic deposits are recycled and can be transferred to more
417 distal sink. The observed increasing volumes and increasing export of sediment to the Bay of
418 Biscay is thus contemporaneous with a (long) denudation paroxysm, but also with a more
419 temporally resolved drop of preservation in the foreland which becomes less subsident and
420 might be locally eroded and recycled. The observed modification of source to sink pattern
421 might record combination of two processes, denudation increasing phase and cessation of
422 subsidence in the foreland could explain the increase export toward the final sink and
423 sedimentation rate increase observed in the western Aquitaine Platform part (fig. 4).

424 *Climate* - Few palaeoclimatological data dealing with the palaeoprecipitation record
425 are available in the Aquitaine Basin. A palaeobotanical synthesis (Dupéron-Laudoueneix and
426 Pons, 1985) indicate a cooling expressed by a decrease of the tropical taxon from the Eocene

427 to the Oligocene that corresponds to the Eocene-Oligocene global cooling event (Zachos et
428 al., 2011). This event takes place at 33.9 Ma, thus well before the observed 26 Ma event.

429 The 26.0 ± 2.0 Ma increase event is contemporary with the Late Oligocene Warming,
430 that correspond to a sustained *ca.* 6°C global warming (Zachos et al., 2001). Warming events
431 (such as PETM) can cause sediment pulses (Foreman et al., 2012). Schlunegger and Norton,
432 (2015) have tested this increase hypothesis in the Alpine Foreland Basin. They focus on the
433 Napf megafan sedimentary evolution in the Alpine foothills during this period of sudden
434 warming. As in our study the period is also a time of increasing exhumation phase in the Alpine
435 mountain (Schmid et al., 1996). They didn't observe remarkable shift in the water discharge,
436 and they ruled out the hypothesis of sediment rate increase due to the Late Oligocene
437 warming but linked it to tectonic process.

438 In our case, the Late Oligocene (26.0 ± 2.0 Ma) siliciclastic rate increase is linked to the
439 high erosion rates and lower preservation in the Aquitaine Platform controlled by the orogenic
440 tectonic processes. This increase is also registered in the Ebro basin (Filleaudeau, 2011).

441

442 **6.3. The 2.5 ± 0.5 Ma increase and the long-term volume increase**

443

444 The change in the rate of siliciclastic sediment supply during Plio-Pleistocene (at $2.5 \pm$
445 0.5 Ma) is significant (fig. 8) and sharply accentuates the overall increasing rate trend through
446 Cenozoic times. This confirms for this area, the Hay's curve (Hay et al., 1989, 1990; Wold et
447 Hay, 1990) showing a world-scale increase of the siliciclastic fluxes up to today, interpreted by
448 Molnar (Molnar and England, 1990; Molnar, 2004) as the consequence of the global Earth
449 cooling during Plio-Pleistocene times. Herman et al., (2013) further suggest a Plio-Pleistocene
450 (at around 6 Ma) increase in erosion rates and with an acceleration at around 2 Ma that they

451 attribute to the global cooling Pliocene and Pleistocene episodes that enhanced erosion in
452 mountainous glaciated area.

453 Onshore, the most sensible low-temperature thermochronological data, ((U-TH)/He
454 and Fission tracks on apatites) do not show any Pliocene increase of denudation either in the
455 Pyrenees Mountains (e.g. Bosch et al., 2016) or in the French Massif central (T. François,
456 personal communications). The youngest cooling episode is recorded at around 10 Ma (Late
457 Miocene) in the western axial zone and no basin erosion > 1.5 km could be retrieved in the
458 Aquitaine Basin for the last 20 Myrs (Fillon et al., 2021), except in easternmost Aquitaine Basin
459 (Corbières area) Al Reda et al., (2021) although this denudation event is temporally poorly
460 resolved between 20 and 5 Ma, this event is attributed by the authors to early Miocene and
461 opening of the Gulf o Lion.

462 With these observations, several hypothesis have yet to be tested: (1) a climate effect
463 on the intensity and pattern of erosion, with the effect of glaciations and change of the pattern
464 of erosion, from widely distributed before 5.3 Ma to localized in valleys after with a greater
465 climate effect from 2 Ma, which could be associated to (2) an “underestimated” significant re-
466 erosion of previously deposited sediments in the Aquitaine Basin (“cannibalization”) and (3)
467 lateral supply of sediments from other sources in the deep-sea domain under the action of
468 deep-sea currents (Le Danois Contourite Depositional System due to Mediterranean Outflow
469 Water, Van Rooij et al., 2010). Moreover, Pliocene-Pleistocene evolution of the Aquitaine
470 Platform records a two steps evolution, the emplacement of megafans with the most striking
471 one i.e. Lannemezan megafan starting during Messinian up to Uppermost Pliocene (Ortiz et
472 al., 2020). This megafan mainly source by the Neste River is dissected during Late Quaternary
473 (at or before 300 ka, Mouchéné et al., 2017). This indicates a shift from a sedimentation phase

474 to an incision phase, that takes place between 2.6 Ma and 300 ka. This supports the hypothesis
475 of recycling by erosion of previous deposits sediments in the Aquitaine Platform.

476 Although glaciations is well documented during Late Quaternary in the Pyrenees
477 (Calvet et al., 2011; Delmas et al., 2008), their importance in terms of sedimentary budget is
478 still not quantified. Identified glacial remnants are restricted to the upper part of the valleys
479 and mostly confined in the mountains belts (Delmas et al., 2011). The preservation over large
480 areas of preglacial flat surfaces in the high chain (Bosch et al., 2016) claim for erosion localized
481 mainly in glacial valleys. This supports the hypothesis of a more localized erosion in the
482 mountain valleys.

483

484

485 **6.4. Sediment budget evolution between onshore and offshore**

486

487 The change of the sediment ratio between onshore and offshore (R_{OnOff}) from 25.3 (66
488 to 33.9 Ma) to 0.01 (5.3 to 0 Ma) has to be related to the general evolution of the two basins
489 as summarized in item 2.2. This drastic change of ratio in favour of the “ultimate sink” is in
490 line with the synthesis of Hinderer, 2012.

- 491 • The first interval (66.0-33.9 Ma) spans a long period of time and several steps of the
492 orogeny. A first phase (66-56 Ma) of relatively quiescent with very low relief in the
493 mountain belt and siliciclastic input, a second phase (56-33.9 Ma) that correspond to
494 progressive mountain building with increasing relief and erosion. At around 40 Ma,
495 deformation propagate in the foredeep and segmented the foredeep. During this entire
496 period, accommodation space created by the flexure was large enough to trap siliciclastic

497 sediments coming from the incipient mountain belt. This may explain the general trend
498 between 66 and 33.9 Ma.

- 499 • The second period (33.9-23.03 Ma) also spans different periods, paroxysm of denudation
500 in the Mountain belt, while in the Aquitaine Platform, two periods are deciphered, a first
501 period of accumulation and deformation (up to 25.2 Ma) and a second with a decrease of
502 subsidence and increase of export to the distal sink (since 25.2 Ma). As a consequence,
503 since 25.2 Ma, siliciclastic supply is too important compare to the accommodation space
504 creation and the sediments are transferred to the progradational wedges in the Aquitaine
505 platform. This explains the trend during Oligocene and particularly since 25.2.
- 506 • During the post-orogenic period, first the Aquitaine platform subsidence drastically
507 decreased (25.2-10.6 Ma) and second this domain was partly uplifted (10.6-0 Ma). This
508 favored the sediment export to the deep-sea domain, first poorly balanced (R_{OnOff} at 0.18
509 – map 23.03-5.3 Ma) and then unbalanced (R_{OnOff} at 0.01 - map 5.3-0 Ma).

510

511 7. Conclusion

512

513 Our objective was to measure the siliciclastic sediment volumes and rates of a retro-
514 foreland basin and its lateral deep equivalent, during the end tectonic and topographic
515 evolution of the Pyrenees mountain belt.

516

517 (1) The amount of sediments is of $51\,500 \pm 16\,800 \text{ km}^3$ for the Cenozoic.

518 (2) The siliciclastic sediment rate curve during Cenozoic shows two major increase around
519 26.0 and 2.5 Ma. This last one is a major increase. This is in agreement with the world-
520 scale trend of Hay et al. (1989), expected of climatic origin by Molnar and England (1990).

521 The 26.0 Ma increase is clearly related to the exhumation of the Pyrenees of tectonic
522 origin. The discussion on the major 2.5 Ma episode is still open.

523 (3) The mass balance between the Aquitaine Platform and the deepest domains (Landes
524 Plateau, Bay of Biscay) change through time in favour to the deep domain. This might be
525 explained by the ratio between subsidence created accommodation space and the
526 produced siliciclastic sediments by the mountain belt on the platform. This confirms and
527 quantifies the stratigraphic model of sink preservation proposed by Ortiz et al. (2020). This
528 also suggests the sediment accumulation in the deep part of the margin is only controlled
529 by the accommodation/sedimentation balance on the platform.

530

531 **Data Availability Statement**

532 Research data are not shared

533

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882 **Table 1 caption:** Siliciclastic sediment volumes and rates values, with the uncertainties

Accumulated solid volumes, rates and associated variance for the Aquitaine Basin

Time interval	(Ma)	Volume ($\times 1 \text{ km}^3$)	σ	Rate ($\times 1 \text{ km}^3 / \text{Myr}$)	σ
Pliocene - Pleistocene	5.33 - 0	228.6	± 10.8	43.1	± 2.0
Aquitanian - Messinian	23.03 - 5.33	1 370.3	± 385.2	77.2	± 21.7
Rupelian - Chattian	33.9 - 23.03	7 894.7	$\pm 1 949.9$	726.2	± 179.3
Danian - Priabonian	66 - 33.9	16 545.6	$\pm 5 563.1$	515.4	± 173.3

Accumulated solid volumes, rates and associated variance for the southern Bay of Biscay

Time interval	(Ma)	Volume ($\times 1 \text{ km}^3$)	σ	Rate ($\times 1 \text{ km}^3 / \text{Myr}$)	σ
Pliocene - Pleistocene	5.33 - 0	16 445.0	$\pm 5 929.1$	3 084.2	$\pm 1 111.9$
Aquitanian - Messinian	23.03 - 5.33	7 301.3	$\pm 2 231.5$	411.8	± 125.9
Rupelian - Chattian	33.9 - 23.03	1 043.7	± 351.8	96.0	± 32.4
Danian - Priabonian	66 - 33.9	652.7	± 352.0	20.6	± 11.1

Accumulated solid volumes, rates and associated variance for the whole basins

Time interval	(Ma)	Volume ($\times 1 \text{ km}^3$)	σ	Rate ($\times 1 \text{ km}^3 / \text{Myr}$)	σ
Pliocene - Pleistocene	5.33 - 0	16 673.6	$\pm 5 940.0$	3 145.9	$\pm 1 120.8$
Aquitanian - Messinian	23.03 - 5.33	8 671.7	$\pm 2 616.8$	489.1	± 147.6
Rupelian - Chattian	33.9 - 23.03	8 938.4	$\pm 2 301.7$	822.3	± 211.7
Danian - Priabonian	66 - 33.9	17 198.3	$\pm 5 915.2$	535.8	± 184.3

883 **Table 1**

Carbonate fraction for the Bay of Biscay

Time interval	(Ma)	Carbonate fraction (%)	σ
Pliocene - Pleistocene	5.33 - 0	7.5	± 7.5
Messinian	7.24 - 5.33	14	± 10
Aquitainian - Tortonian	23.03 - 7.24	18	± 10
Rupelian - Chattian	33.9 - 23.03	44	± 15
Danian - Priabonian	66 - 33.9	60	± 20

Carbonate fraction for the «Façade Atlantique» Basin

Time interval	(Ma)	Carbonate fraction (%)	σ
Pliocene - Pleistocene	5.33 - 0	5	± 5
Tortonian - Messinian	11.63 - 5.33	10	± 10
Langhian - Serravallian	15.97 - 11.63	20	± 20
Burdigalian	20.44 - 15.97	50	± 20
Aquitainian	23.03 - 20.44	60	± 30
Rupelian - Chattian	33.9 - 23.03	40	± 20
Priabonian	37.8 - 33.9	50	± 20
Danian - Bartonian	66 - 37.8	40	± 20

Carbonate fraction for the Arzacq Basin

Time interval	(Ma)	Carbonate fraction (%)	σ
Pliocene - Pleistocene	5.33 - 0	5	± 5
Tortonian - Messinian	11.63 - 5.33	10	± 10
Langhian - Serravallian	15.97 - 11.63	20	± 20
Burdigalian	20.44 - 15.97	50	± 20
Aquitainian	23.03 - 20.44	60	± 30
Rupelian - Chattian	33.9 - 23.03	20	± 10
Priabonian	37.8 - 33.9	20	± 10
Danian - Bartonian	66 - 37.8	40	± 20

Carbonate fraction for the Carcassonne Basin

Time interval	(Ma)	Carbonate fraction (%)	σ
Pliocene - Pleistocene	5.33 - 0	5	± 5
Tortonian - Messinian	11.63 - 5.33	10	± 10
Langhian - Serravallian	15.97 - 11.63	20	± 20
Burdigalian	20.44 - 15.97	50	± 20
Aquitainian	23.03 - 20.44	60	± 30
Rupelian - Chattian	33.9 - 23.03	25	± 20
Priabonian	37.8 - 33.9	20	± 20
Danian - Bartonian	66 - 37.8	40	± 20

Carbonate fraction for the Tarbes Basin

Time interval	(Ma)	Carbonate fraction (%)	σ
Pliocene - Pleistocene	5.33 - 0	5	± 5
Tortonian - Messinian	11.63 - 5.33	10	± 10
Langhian - Serravallian	15.97 - 11.63	20	± 20
Burdigalian	20.44 - 15.97	50	± 20
Aquitainian	23.03 - 20.44	60	± 30
Rupelian - Chattian	33.9 - 23.03	20	± 10
Priabonian	37.8 - 33.9	15	± 10
Danian - Bartonian	66 - 37.8	40	± 20

Carbonate fraction for the Aquitaine platform

Time interval	(Ma)	Carbonate fraction (%)	σ
Pliocene - Pleistocene	5.33 - 0	5	± 5
Tortonian - Messinian	11.63 - 5.33	10	± 10
Langhian - Serravallian	15.97 - 11.63	20	± 20
Burdigalian	20.44 - 15.97	50	± 20
Aquitainian	23.03 - 20.44	60	± 30
Rupelian - Chattian	33.9 - 23.03	20	± 10
Priabonian	37.8 - 33.9	15	± 10
Danian - Bartonian	66 - 37.8	40	± 20

Carbonate fraction for the Castres Basin

Time interval	(Ma)	Carbonate fraction (%)	σ
Pliocene - Pleistocene	5.33 - 0	5	± 5
Tortonian - Messinian	11.63 - 5.33	10	± 10
Langhian - Serravallian	15.97 - 11.63	20	± 20
Burdigalian	20.44 - 15.97	50	± 20
Aquitainian	23.03 - 20.44	60	± 30
Rupelian - Chattian	33.9 - 23.03	40	± 20
Priabonian	37.8 - 33.9	40	± 20
Danian - Bartonian	66 - 37.8	40	± 20

Carbonate fraction for the Tartas Mirande Basin

Time interval	(Ma)	Carbonate fraction (%)	σ
Pliocene - Pleistocene	5.33 - 0	5	± 5
Tortonian - Messinian	11.63 - 5.33	10	± 10
Langhian - Serravallian	15.97 - 11.63	20	± 20
Burdigalian	20.44 - 15.97	50	± 20
Aquitainian	23.03 - 20.44	60	± 30
Rupelian - Chattian	33.9 - 23.03	30	± 20
Priabonian	37.8 - 33.9	50	± 10
Danian - Bartonian	66 - 37.8	40	± 20

888 **Figure 1:** A: Main physiographic and structural features and available dataset of seismic
889 reflection lines in the Aquitaine Basin, Landes Plateau and Biscay Bay deep-basin. B: Location
890 of the studied area in Europe.

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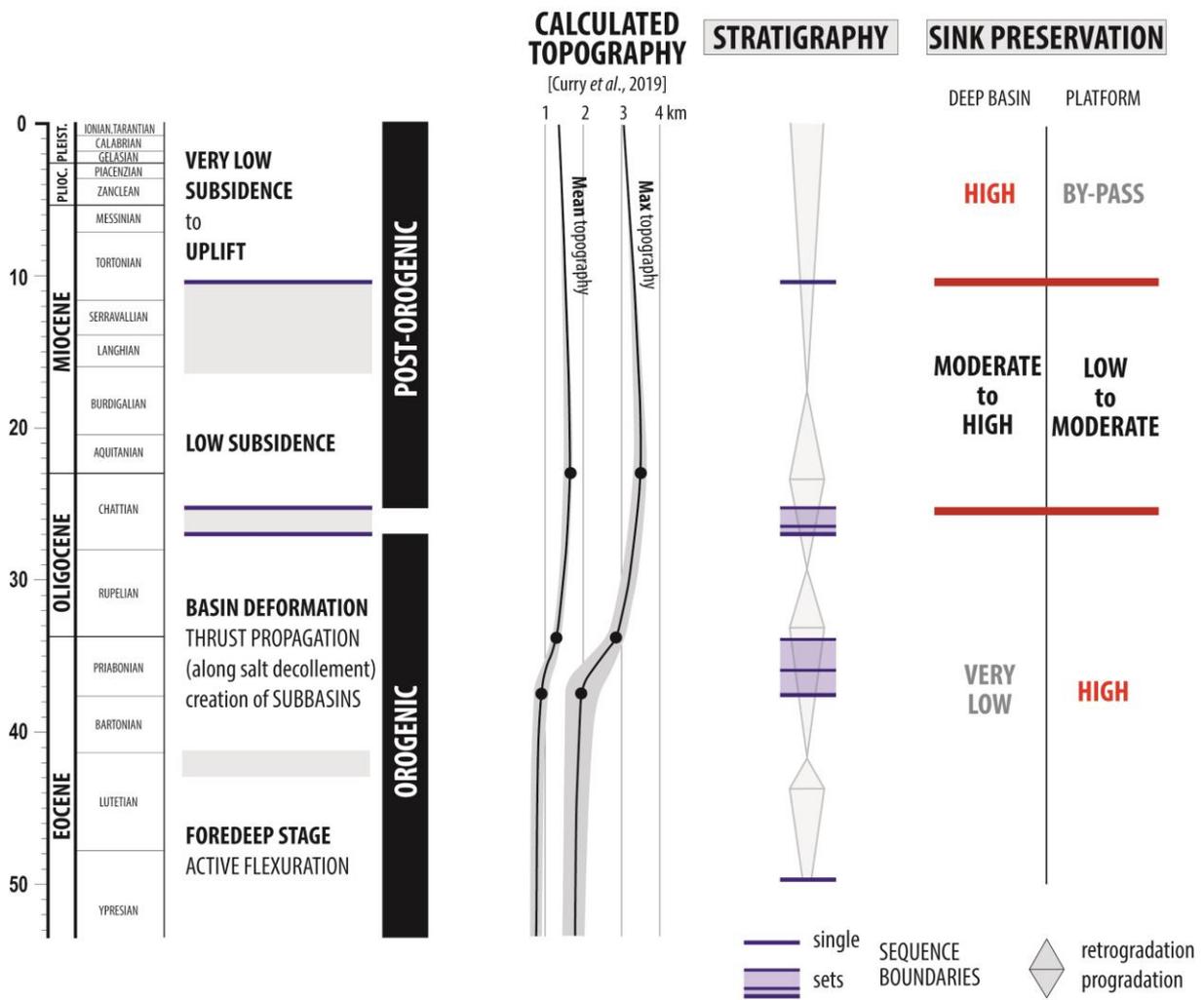
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912 **Figure 2**

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914 **Figure 2:** Synthetic chart of the main events (deformation, topography, sediment routing) of
915 the Aquitaine Basin to deep Biscay Bay Basin sedimentary system

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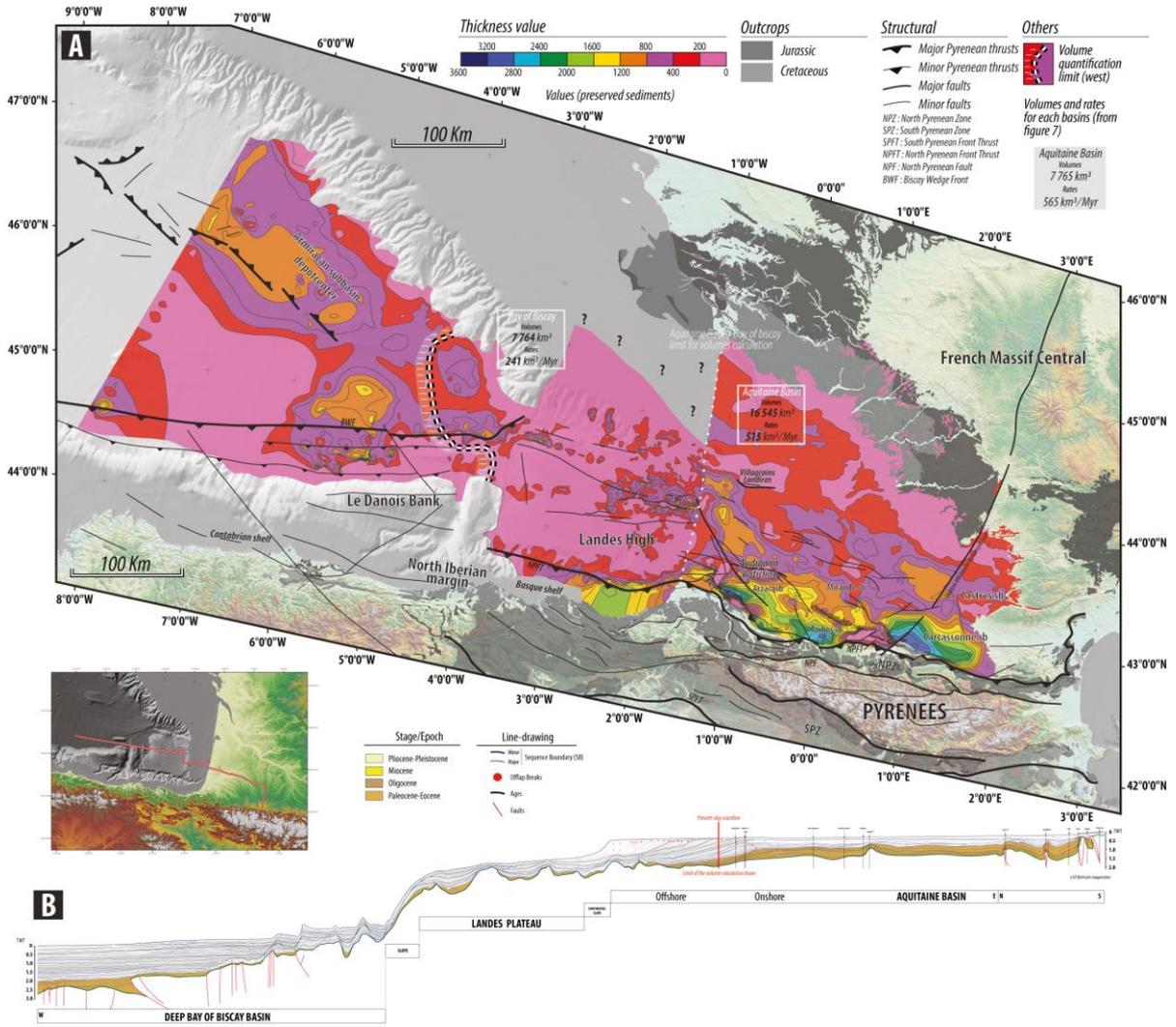
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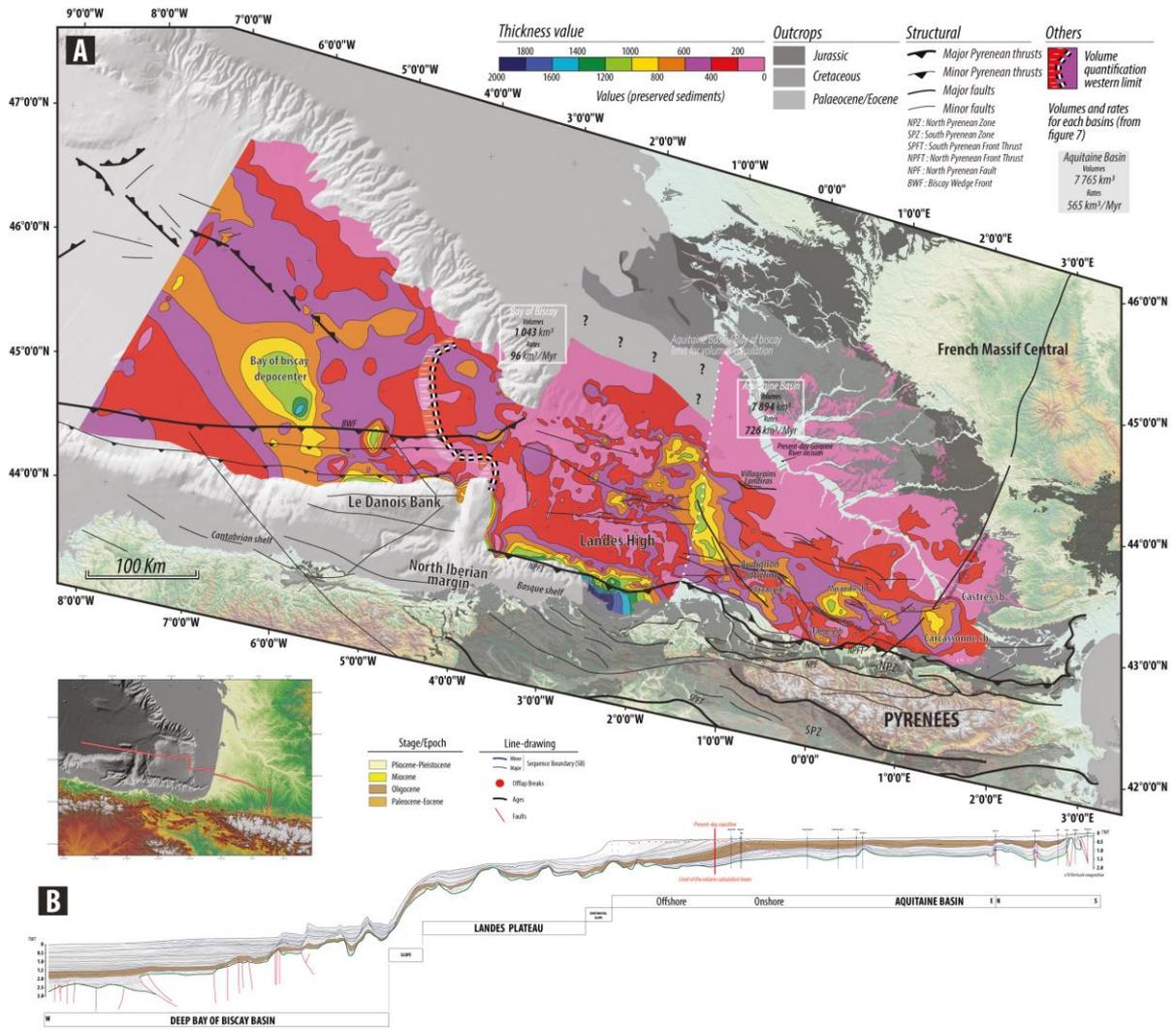
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937 **Figure 3**



938 **Figure 4**

939 **Figure 3:** A: Palaeocene-Eocene preserved sediment thickness (isopach, meters) map. The
940 volume quantification western limit is according to the deep sea fan location (Cremer, 1983;
941 Iglesias, 2009). B: interpreted cross-section from the proximal Aquitaine Basin part to the deep
942 Bay of Biscay Basin, the considered period (isopach map) is highlighted by colour in the cross-
943 section.

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966 **Figure 4:** A: Oligocene preserved sediment thickness (isopach, meters) map. The volume
967 quantification western limit is according to the deep sea fan location (Cremer, 1983; Iglesias,
968 2009). B: interpreted cross-section from the proximal Aquitaine Basin part to the deep Bay of
969 Biscay Basin, the considered period (isopach map) is highlighted by colour in the cross-section.

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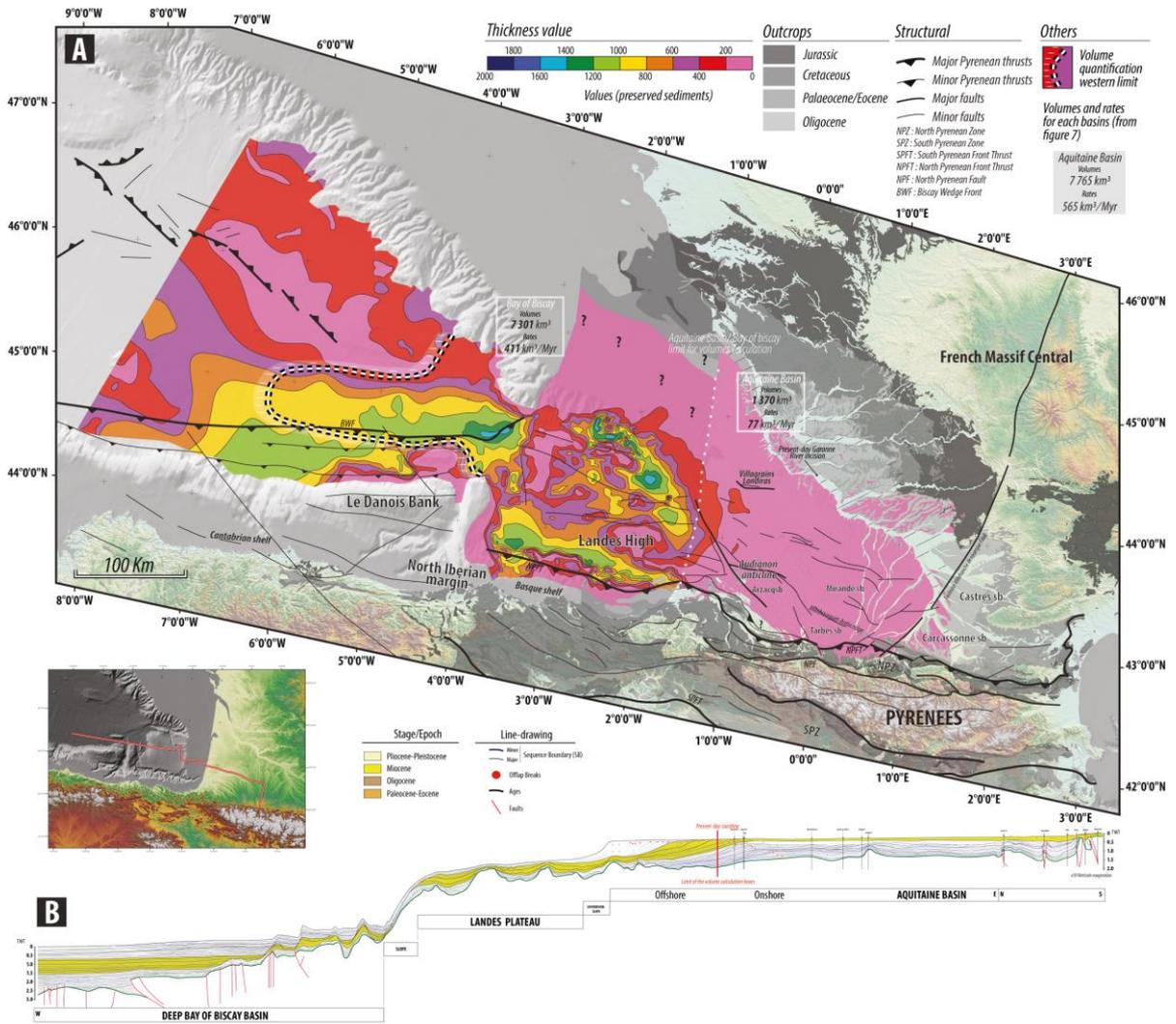
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991 **Figure 5**

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993 **Figure 5:** A: Miocene preserved sediment thickness (isopach, meters) map. The volume
994 quantification western limit is according to the deep sea fan location (Cremer, 1983; Iglesias,
995 2009). B: interpreted cross-section from the proximal Aquitaine Basin part to the deep Bay of
996 Biscay Basin, the considered period (isopach map) is highlighted by colour in the cross-section.

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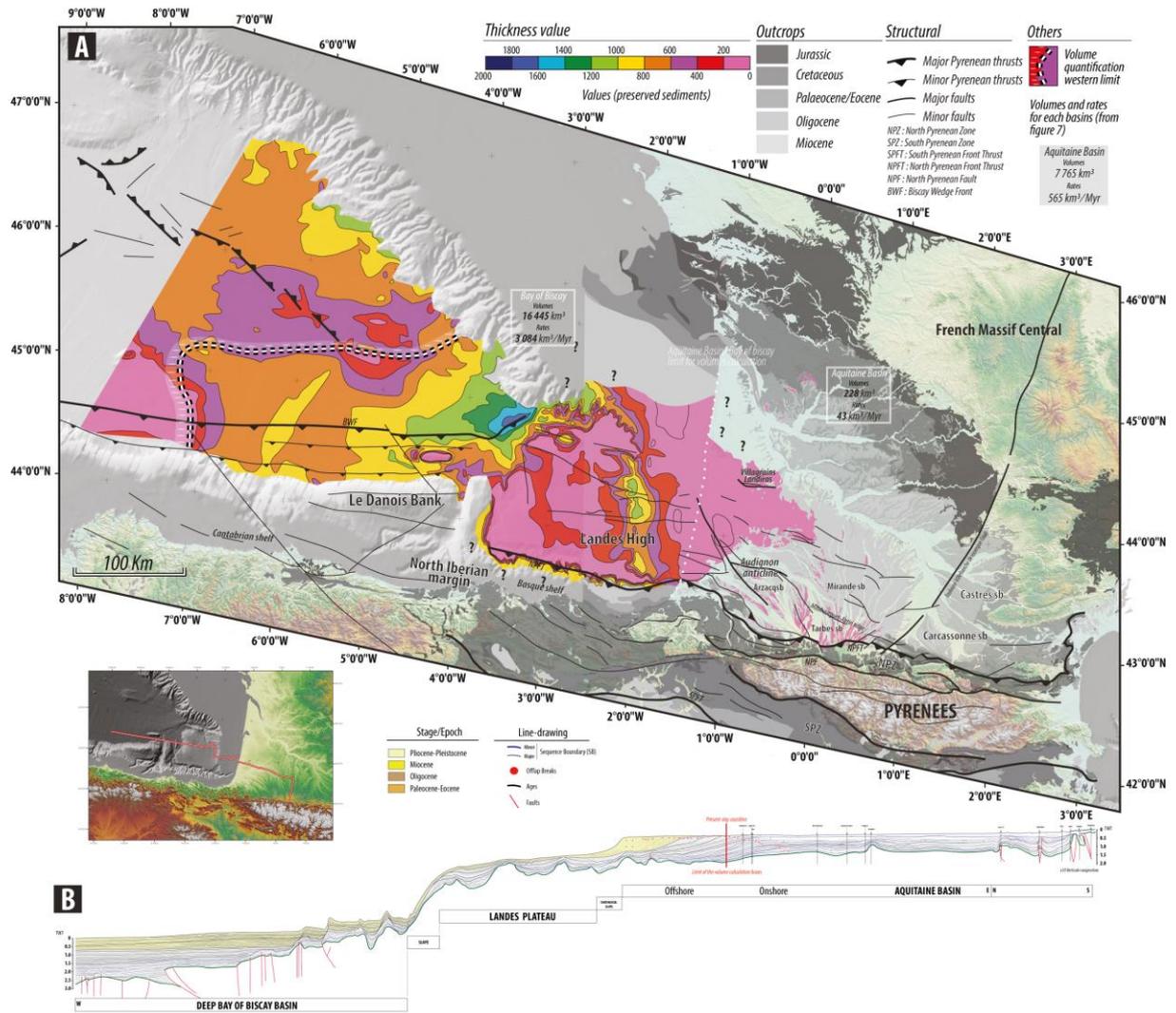
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1019 **Figure 6**

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1021 **Figure 6:** A: Pliocene-Pleistocene preserved sediment thickness (isopach, meters) map. The
1022 volume quantification western limit is according to the deep sea fan location (Cremer, 1983;
1023 Iglesias, 2009, Brocheray et al., 2014). B: interpreted cross-section from the proximal
1024 Aquitaine Basin part to the deep Bay of Biscay Basin, the considered period (isopach map) is
1025 highlighted by colour in the cross-section.

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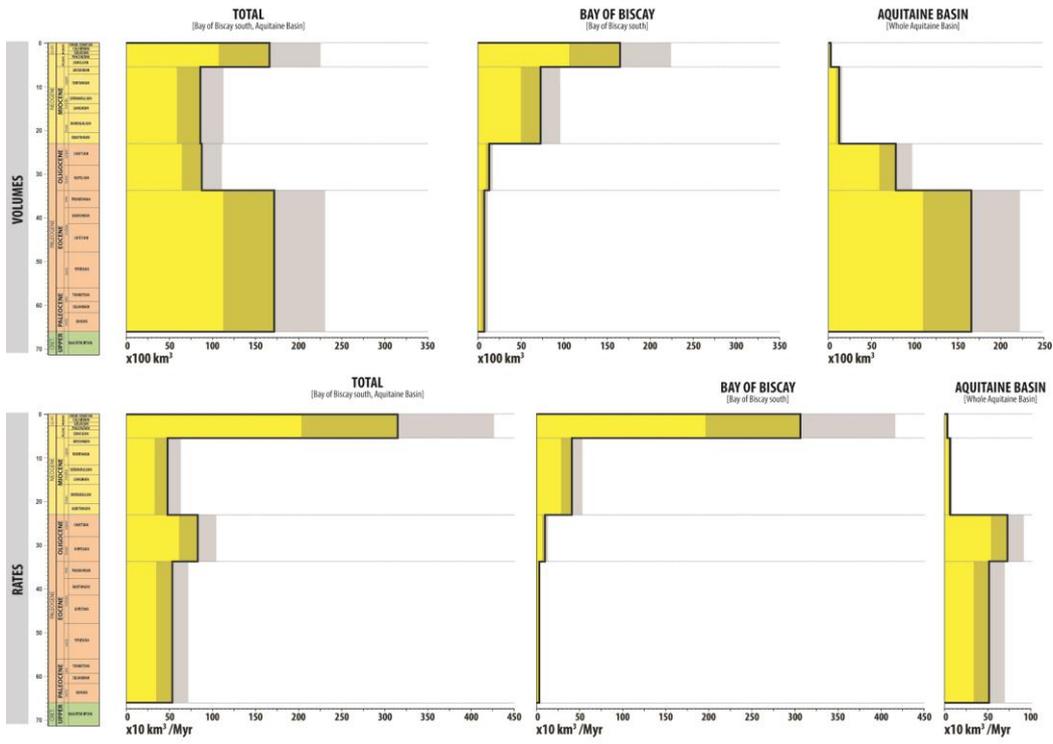


Figure 7

1049 **Figure 7:** Siliciclastic sedimentation volumes and rates of the Aquitaine Basin and Bay of
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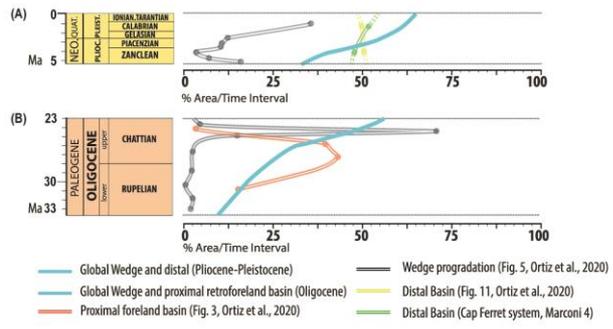


Figure 8

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1075 **Figure 8:** In the cross sections published in Ortiz et al., (2020), some parts of the interpretation
1076 are of higher resolution than the resolution of the calculations in this paper (i.e. Palaeocene-
1077 Eocene, Oligocene, Miocene, Pliocene-Pleistocene). For example, the figure 5 in Ortiz et al.,
1078 (2020) shows several high resolutions surfaces (thanks to orbitostratigraphy) at 32.2, 32.0,
1079 31.3, 29.4, 27.1, 26.4, 25.2, 24.5, 24.4 Ma for the Oligocene period. This resolution on different
1080 parts of the cross section allows us to make a precise quantification shown in the figure. This
1081 figure is a compilation of the percentage of 2D area normalized to their duration thanks to
1082 high resolution timeline on different cross section published in Ortiz et al., (2020)

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