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**Title: Quantification of vertical movement of low elevation topography combining a new compilation of global sea-level curves and scattered marine deposits (Armorican Massif, western France)**

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## Highlights

- A formalized method for quantifying low-amplitude vertical movements
- Re-assessment of the reliability of the published Cenozoic global sea level curves
- A low amplitude subsidence of the Armorican massif quantified from 30 to 3.6 Ma
- Cenozoic growth of Apulia-Eurasia convergence effects on Armorican deformations

1 **Abstract**

2 A wide range of methods are available to quantify Earth's surface vertical movements but most of  
3 these methods cannot track low amplitude (< 1 km, e.g. thermochronology) or old (> 5 Ma, e.g.  
4 cosmogenic isotope studies) vertical movements characteristic of plate interiors. The difference  
5 between the present-day elevation of ancient sea-level markers (deduced from well dated marine  
6 deposits corrected from their bathymetry of deposition) and a global sea-level curve are sometimes  
7 used to estimate these intraplate vertical movements. Here, we formalized this method by re-  
8 assessing the reliability of published global sea-level curves to build a composite curve that combines  
9 the most reliable ones at each stage, based on the potential bias and uncertainties inherent to each  
10 curve. We suggest i) that curves which reflect ocean basin volume changes are suitable for the ca.  
11 100 to 35 Ma "greenhouse" period ii) whereas curves that reflects ocean water volume changes are  
12 better suited for the ca. 35 to 0 Ma "icehouse" interval and iii) that, for these respective periods, the  
13 fit is best when using curves that accounts for both volume changes. We used this composite sea-  
14 level curve to investigate the poorly constrained Paleogene to Neogene vertical motions of the  
15 Armorican Massif (western France). It is characterized by a low elevation topography, a Variscan  
16 basement with numerous well dated Cenozoic marine deposits scattered upon it. Using our method,  
17 we identify low amplitude vertical movements ranging from 66 m of subsidence to 89 m of uplift  
18 over that time period. Their spatial distribution argues for a preferred scale of deformation at  
19 medium wavelengths (i.e., order 100 km), which we relate to the deformation history of  
20 northwestern European lithosphere in three distinct episodes. i) A phase of no deformation between  
21 38 and 34 Ma, that has been previously recognized at the scale of northwestern Europe, ii) a phase of  
22 low subsidence between 30 and 3.6 Ma, possibly related to buckling of the lithosphere and iii) a  
23 phase of more pronounced uplift between 2.6 Ma and present, which we relate to the acceleration  
24 of the Africa-Apulia convergence or to enhanced erosion in the rapidly cooling climate of the  
25 Pleistocene.

26

27 **Keywords:** Quantification of vertical movements; Global sea-level; Intraplate domains; Low  
28 amplitude deformation; Armorican Massif; Cenozoic

29 **1. Introduction**

30 Characterizing the Earth's surface vertical movements, i.e. its uplift and subsidence, by quantifying  
31 their amplitude and wavelength, and deciphering the nature of the processes responsible for these  
32 movements remain challenging questions for the geoscientist. These movements often result from  
33 lithospheric-scale deformations, which range in scale from short ( $\times 10^{-6}$  –  $\times 10^1$  km) to long  
34 wavelengths ( $\times 10^3$  –  $\times 10^5$  km; Şengör, 2009). Plate boundaries, where topography commonly takes  
35 the form of narrow mountain belts and rifts, are characterized by short-wavelength deformation  
36 processes (e.g. faulting) where the amplitude of deformation often exceeds its wavelength (Bishop,  
37 2011). Conversely, in plate interiors, where topography takes mostly the form of large plateaus  
38 surrounded by plains, hills and flat sedimentary basins, medium ( $\times 10^2$  km) to long ( $\times 10^3$  km)  
39 wavelength deformation processes dominate, where the amplitude of deformation is two to three  
40 orders of magnitude smaller than its wavelength (i.e.  $\times 10^2$  m). The processes thought to be  
41 responsible for medium wavelength deformation are lithospheric buckling, crustal loading or  
42 underplating (Watts, 2001; Anell et al., 2009) and, for the longest wavelength deformation, mantle-  
43 driven processes such as dynamic topography – the vertical deflection of the surface topography  
44 (gravitational stresses) required to balance the viscous stresses in the flowing mantle at the base of  
45 the lithosphere (Braun, 2010; Molnar et al., 2015).

46 A wide range of methods, such as low-temperature thermochronology (Apatite Fission Tracks  
47 Analysis, for example), cosmogenic isotopes and OSL (Optically Stimulated Luminescence) studies  
48 were developed and used to quantify denudation at plate boundaries where uplift leads to  
49 substantial erosion and rock cooling. Unfortunately, these methods cannot be readily used in most  
50 plate interiors, because the vertical movements and associated denudation are often i) too low to be  
51 quantified by thermochronology or ii) too old ( $> 5$  Ma) to be recorded by cosmogenic isotopes or OSL  
52 methods. Consequently, little attention has been paid to the non- or post-orogenic uplift and  
53 subsidence of plate interiors, characterized by low elevation plateaus and other small amplitude  
54 topographic features.

55 Recently, several studies have attempted to estimate the amplitude of vertical movements in  
56 continental interiors from the difference between the present-day elevation of dated ancient sea-  
57 level markers and their respective initial elevations (e.g. Bétard, 2010; Braga et al., 2003; Dorsey et  
58 al., 2011; Pederson et al., 2002; Peulvast and Bétard, 2015). The modern elevations of ancient sea-  
59 level markers are deduced from the elevations of well dated marine deposits corrected from their  
60 bathymetry at the time of deposition (corrected for sediment load and compaction effects in thick  
61 sedimentary series, e.g. Dorsey et al., 2011). The initial elevation of ancient sea-level markers can be  
62 inferred from any given global sea-level curve. Such a method is suitable for quantifying low  
63 amplitude vertical movements typical of intraplate domains and associated low elevation  
64 topography, but its results strongly depend on the assumed global sea-level curve used to infer the  
65 past sea-level elevation. This is especially true when the inferred amplitude of deformation is less  
66 than 200 meters, because of the large discrepancies that exist between several published sea-level  
67 curves (e.g. up to 200 m between the Haq et al. (1987) and Miller et al. (2005)'s curves during the  
68 Upper Cretaceous), which has led many to question their validity (e.g. Moucha et al., 2008; Müller et  
69 al., 2008; Miall, 2010).

70 The main purpose of this study is to improve and formalize this simple method of quantifying the  
71 timing and amplitude of vertical movements in low elevation areas where thin marine sedimentary  
72 veneers are preserved, by taking into account uncertainties on bathymetry estimates and global sea-  
73 level elevation at the time of deposition. As an accurate knowledge of global sea-level changes  
74 through times is a cornerstone in this method, we have re-assessed the reliability of many curves  
75 published since the pioneering work of Haq et al. (1987). We then constructed a composite sea-level  
76 curve by combining the most reliable intervals of several curves, taking into account the potential  
77 bias and uncertainties inherent to the methods used to build each curve. We then applied the  
78 method to compute improved estimates of uplift and subsidence of the Armorican Massif during the  
79 Cenozoic.

80 The Armorican Massif is one of numerous Paleozoic (Caledonian and Variscan) basement blocks of  
81 western Europe characterised by low to moderate elevation plateaus such as the Massif Central, the  
82 Rhenish Massif, the Bohemian Massif or the Scottish Highlands. Most of these basement blocks have  
83 experienced several episodes of burial and exhumation in the Mesozoic and/or Cenozoic (i.e. long  
84 after their post-orogenic planation; e.g. *Barbarand et al., 2013*). This is the case for the Armorican  
85 Massif (*Bessin et al., 2015*), that is part of the Variscan Belt (*Ballèvre et al., 2009*). This low relief and  
86 low elevation topographic feature was twice buried then exhumed between Jurassic and Paleocene  
87 times in response to relative movements between Iberia and Eurasia. However, as other western  
88 European basements, its Paleogene to Neogene uplift/subsidence and deformation history is still  
89 poorly constrained (*Bessin et al., 2015*) despite the presence of numerous well dated Cenozoic  
90 shallow marine deposits scattered upon it. The Armorican Massif is therefore an ideal place to use  
91 our method and derive from it improved estimates of the timing and amplitude of surface vertical  
92 movements. Using these estimates, we discuss the possible driving mechanisms responsible for these  
93 uplift/subsidence events in the framework of the recent tectonic history of western Europe.

94

## 95 **2. Calculation of vertical movements: methodology**

96

### 97 *2.1 Principles*

98 Plate interiors often preserved scattered remnants of marine sediments as thin sedimentary veneers  
99 (up to x 1 m to 10-20 m thick). Their occurrence and preservation are function of three parameters: i)  
100 global sea-level changes, ii) surface uplift and subsidence and iii) sedimentary flux that can drive  
101 further subsidence (*Posamentier et al., 1988*). In low preservation environments (e.g. most of  
102 western Europe Variscan massifs), this latter driver can be disregarded as sedimentary thicknesses  
103 are usually low (< 50 m) and the additional subsidence they can generate by isostasy or compaction  
104 is negligible (*Allen and Allen, 2013*). Areas where no section is missing should be privileged to limit  
105 the uncertainty associated with unknown amounts of erosion. In this case, the presence of a marine

106 sediment can be related to global sea-level changes and surface uplift or subsidence events only.  
107 Consequently, the difference between the modern elevation of a marine sediment and an estimate  
108 of its initial elevation, i.e. at time of deposition, can be regarded as an accurate estimate of the sum  
109 of all vertical movements it recorded until present (finite vertical movement). To compute this  
110 vertical movement, one needs to accurately measure:

- 111 - the age of the marine sediment, which is commonly obtained from its fossiliferous fauna and  
112 flora contents (biostratigraphic markers);
- 113 - the bathymetry under which this sediment was deposited, which is also constrained by its  
114 fossiliferous fauna and flora content (palaeo-environmental markers) together with  
115 sedimentary facies;
- 116 - the global sea-level at the time of sediment deposition, which we will obtain from a  
117 compilation of reliable global sea-level curves (as discussed in §3.3).

118 The uncertainty on the estimate of vertical movement can also be obtained from uncertainties on  
119 estimates of paleo-bathymetry and global sea-level height, together with the uncertainty on the age  
120 of the marine deposit.

121

## 122 *2.2 Finite vertical movement calculation*

123 For a given dated marine sedimentary remnant  $s_1$  deposited in a low preservation environment, the  
124 vertical displacement it recorded from its deposition at time  $t_1$  to present-day ( $t_t$ ), hereafter called  
125 finite vertical movement ( $fvm$ ), is

126 (1)  $fvm_{(1 \rightarrow t)} = z_{s_1 t_t} - z_{s_1 t_1}$                       thus,

127 (2)  $fvm_{(1 \rightarrow t)} = z_{s_1 t_t} - sl_1 + b_1$

128 where  $zs_{1t_t}$  is its present-day elevation,  $sl_1$  the global sea-level elevation at  $t_1$  (with respect to  
 129 present-day global sea-level) and  $b_1$  the bathymetry under which  $s_1$  was deposited (Fig.1). For each  
 130 location, we do not estimate a single value but a range of  $fvm_{(1\rightarrow t)}$  by considering a range of past  
 131 bathymetric estimates and a range of global sea-level values. The former is related to the uncertainty  
 132 in bathymetry inherent to using palaeo-environmental markers and sedimentary facies of  $s_1$ . The  
 133 latter is due to errors in global sea-level (amplitude) which have to incorporate the uncertainties in  
 134 the ages of the marine deposits (timing). These ranges of bathymetry and global sea-level lead to  
 135 estimates of  $fvm$  minimum ( $fvm_{(1\rightarrow t) \min}$ ), mean ( $fvm_{(1\rightarrow t) \text{mean}}$ ) and maximum ( $fvm_{(1\rightarrow t) \max}$ )  
 136 values, which are obtained from (uplift case)

$$137 \quad (3) \quad fvm_{(1\rightarrow t) \min} = zs_{1t_t} - sl_{1 \min} + b_{1 \max} \quad (\text{subsidence case: } fvm_{(1\rightarrow t) \max})$$

$$138 \quad (4) \quad fvm_{(1\rightarrow t) \text{mean}} = zs_{1t_t} - sl_{1 \text{mean}} + (b_{1 \max} + b_{1 \min})/2$$

$$139 \quad (5) \quad fvm_{(1\rightarrow t) \max} = zs_{1t_t} - sl_{1 \max} + b_{1 \min} \quad (\text{subsidence case: } fvm_{(1\rightarrow t) \max})$$

140 where  $sl_{1 \min}$  and  $b_{1 \min}$  are the minimum values of sea-level and bathymetry at  $t_1$ , respectively,  
 141  $sl_{1 \max}$  and  $b_{1 \max}$  correspond to the maximum value of sea-level and bathymetry at  $t_1$ , respectively,  
 142 and  $sl_{1 \text{mean}}$  is the mean value of global sea-level at  $t_1$  (Fig.1). These  $fvm$  computations were  
 143 performed using several global sea-level curves currently available (e.g. Haq et al., 1987; Miller et al.,  
 144 2005; Kominz et al., 2008; Müller et al., 2008; Rowley, 2013) in order to define a range of possible  
 145 sea-levels for any given time in the past and thus the related range of  $fvm$  values (Fig.1).

146

### 147 2.3 Quantification of successive vertical movements through times and surface elevation restoration

148 Each  $fvm$  quantified may integrate or "stack" several phases of uplift and subsidence which can be  
 149 dissociated in some places. Indeed, some topographic surfaces, such as basement flats or lows, may  
 150 have recorded several marine flooding events through time and low preservation marine deposits of  
 151 different ages can therefore be preserved on these surfaces. As a consequence, if two remnants  $s_1$

152 and  $s_2$  deposited at  $t_1$  and  $t_2$  ( $t_1$  being older than  $t_2$ ), respectively, are found close to one another  
 153 upon a same topographic surface (and in the absence of faulting or subsequent erosion), they both  
 154 underwent the same  $fvm$  from  $t_2$  to present-day ( $t$ ),  $fvm_{(2 \rightarrow t)}$ , implying that the  $fvm$  recorded by  $s_1$   
 155 from  $t_1$  to  $t_2$  is

$$156 \quad (6) \quad fvm_{(1 \rightarrow 2)} = fvm_{(1 \rightarrow t)} - fvm_{(2 \rightarrow t)}$$

157 Moreover,  $zs_{1t_2}$  the elevation of  $s_1$  (i.e. of the flooded topographic surface) at  $t_2$  can be restored  
 158 from the equations (1) and (6):

$$159 \quad (7) \quad zs_{1t_2} = (fvm_{(1 \rightarrow t)} - fvm_{(2 \rightarrow t)}) + zs_{1t_1} \quad \text{thus,}$$

$$160 \quad (8) \quad zs_{1t_2} = zs_{1t_t} - fvm_{(2 \rightarrow t)}$$

161 Its range are computed from  $fvm_{(2 \rightarrow t)}$  uncertainties, and therefore derived from the equations (3),  
 162 (4) and (5) :

$$163 \quad (9) \quad zs_{1t_2 \min} = zs_{1t_t} - fvm_{(2 \rightarrow t) \min} \quad (\text{subsidence: } zs_{1t_2 \max})$$

$$164 \quad (10) \quad zs_{1t_2 \text{ mean}} = zs_{1t_t} - fvm_{(2 \rightarrow t) \text{ mean}}$$

$$165 \quad (11) \quad zs_{1t_2 \max} = zs_{1t_t} - fvm_{(2 \rightarrow t) \max} \quad (\text{subsidence: } zs_{1t_2 \min})$$

166 where  $zs_{1t_2 \min}$ ,  $zs_{1t_2 \text{ mean}}$  and  $zs_{1t_2 \max}$  are the minimum, the mean and the maximum elevation of  
 167  $s_1$  at  $t_2$ , respectively. These computations were applied to a dataset from the Armorican Massif,  
 168 using several global sea-level curves as discussed hereafter (see §3.3) in order to restore ranges of  
 169 successive elevations of several topographic surfaces, and constrain the amplitude and rate of their  
 170 vertical movement through times.

171

### 172 3. A new compilation of available global sea-level curves for Mesozoic to Cenozoic times

173

174 3.1 Global sea-level change driving-factors

175 Several short-term ( $10^0 - 10^4$  yr) to long-term ( $10^6$ - $10^9$  yr) processes may lead to global sea-level  
176 fluctuations through times by changing i) ocean water volume or ii) ocean basin volume (Miller et al.,  
177 2005; Miller et al., 2011; Conrad, 2013).

178 Ocean water volume changes are mainly due to short-term ( $10^1 - 10^4$  yr, e.g. Milankovitch cycles)  
179 processes, chiefly ice sheet volume variations (up to 200 m of amplitude) between "icehouse" and  
180 "greenhouse" periods and ocean water thermal contraction or dilatation which together define  
181 climato-eustasy (Miller et al., 2011; Conrad, 2013). Such fluctuations trends can be sustained for  
182 several million years due to long-term climate change trends. Lower amplitude ( $\approx 5 - 10$  m  
183 amplitude) changes in ocean water volume can also be induced by variations in continental water  
184 storage (lakes and groundwater) and desiccation or flooding of marginal sea (Miller et al., 2011). On  
185 longer time scales ( $10^9$  yr), variations in global water distribution between the Earth's surface and the  
186 mantle also induce low amplitude global sea-level changes (20 – 40 m amplitude; Conrad, 2013).

187 Ocean basin volume changes are mainly related to long-term driving factors ( $10^6 - 10^8$  yr) and are  
188 chiefly induced by mid-ocean ridge volume variations (amplitude: 100-300 m), through variations in  
189 oceanic crust production and ridge length (Müller et al., 2008; Miller et al., 2011; Conrad, 2013)  
190 related to mantle convection and the dispersal and assembly of continents (Conrad, 2013). To a  
191 lesser degree, seafloor loading changes due to oceanic plateaus emplacement and removal or  
192 terrigenous sedimentary flux fluctuations can affect global sea-level (with amplitudes up to ca. 60  
193 m; Miller et al., 2005). On longer time scales, dynamic topography can induce extremely slow (up to 1  
194  $\text{m}\cdot\text{Ma}^{-1}$ ) global sea-level changes of relatively high amplitudes (up to 200 m; Spasojevic and Gurnis,  
195 2012). On shorter time scales ( $10^3 - 10^5$  yr), Glacial Isostatic Adjustment (GIA or postglacial rebound)  
196 also induces global sea-level changes ( $< 5$  m amplitude since 120 kyr; Conrad, 2013; Miller et al.,  
197 2011; Pedoja et al., 2011) but these can be neglected when considering global sea-level changes on  
198 longer time scales ( $> 1$  Ma; Miller et al., 2011).

199

200 *3.2 Compilation of available global sea-level curves*

201 Since the end of the 1970's, numerous and often conflicting global sea-level curves have been  
202 published, based on different assumptions and datasets, for Mesozoic to Cenozoic times. They are  
203 built using five main methods, including (data available in Supplementary Material S1 and plotted in  
204 S2):

205 (1) *Coastal onlap analysis* based on the recognition and measurement of coastal onlap  
206 constrained by correlations of stratigraphic sequences using boreholes, outcrops and seismic  
207 data. This method was used by Haq et al., (1987 ; global dataset) and Haq and Al-Qahtani  
208 (2005 ; Arabian platform regional dataset).

209 (2) *Continental flooding estimates based on using global hypsometric estimates combined with*  
210 *estimates of continental flooding through time derived from paleogeographic datasets.* This  
211 method was used by Rowley (2013), using data from four global paleogeographic datasets  
212 (Scotese and Golonka, 1992; Smith et al., 1994; Markwick, 2011; Blakey, 2012).

213 (3) *Backstripping* based on estimating the effects of sediment compaction, sediment loading and  
214 water-depth changes on sedimentary records at high biochronostratigraphic resolution  
215 located in presumably stable areas such as continental passive margins. These sea-level  
216 datasets mainly come from the eastern US margin (Miller et al., 2005; Kominz et al., 2008).

217 (4) A method based on *ocean floor age-area and depth-area distributions* where ocean basin  
218 volume changes are obtained from the distribution of ocean floor area with age and a  
219 relationship between age and bathymetric depth, derived from global geodynamic models.  
220 This method was used by Müller et al. (2008) and Spasojevic and Gurnis (2012).

221 (5) *Oxygen isotopes ( $\delta^{18}\text{O}$ ) proxies* based on  $\delta^{18}\text{O}/\delta^{16}\text{O}$  ratio measurements of marine  
222 carbonates (foraminifera) which provide indirect records of ice-volume and temperature  
223 changes since Late Neogene (ca. 9 Ma) times. This method was used by Miller et al. (2011).  
224 Backstripped curves were later corrected for the assumed effect of dynamic topography as  
225 published by Müller et al. (2008) from the Miller et al. (2005) dataset and by Kominz et al. (2008).  
226 The latest recalibration on the geologic time scale (Miller, 2013) and long-term filtering (Müller et  
227 al., 2008) were used for comparing the global sea-level curves. Mean, maximum and minimum  
228 sea-level values (including uncertainties) were computed for each geologic time scale stage since  
229 the Upper Cretaceous and at the dating resolution of studied marine sedimentary remnants  
230 (Supplementary Material S1; Fig.2).

231

### 232 3.3. Reliability and selection of compiled global sea-level curves

233 Our compilation of global sea-level curves shows large discrepancies between the available curves (of  
234 ca. 100 to 200 meters for some stage; Fig.2). For our calculations, we constructed a composite global  
235 sea-level curve (and associated uncertainty) by using different curves at different times, based on the  
236 nature of the data that was used to construct it and how reliable that data is for each time period  
237 considered (Fig.3; Supplementary Material S1).

238 Haq et al. (1987) and Haq and Al-Qahtani (2005) global sea-level curves based on coastal onlap  
239 measurements were discarded because of i) the lack of complete dataset publication, ii) the  
240 overestimation of sea-level amplitude due to insufficient correction for compaction, loading and  
241 tectonic subsidence and iii) the chronostratigraphic imprecision of correlated sequence boundaries  
242 (see Miall, 2010 for a review).

243 The latest curves from the backstripping method (Miller et al., 2005; Kominz et al., 2008) were  
244 selected for the "icehouse" times since ca. 35 Ma (Eocene-Oligocene transition) and the onset of

245 permanent Antarctic ice-sheet (Zachos et al., 2008). These curves are reliable for this period as they  
246 reflect well global sea-level changes driven by ocean water volume variations (Miller et al., 2011).  
247 They were however excluded for the preceding periods (pre-35 Ma) as the backstripping method  
248 used requires data from stable sedimentary basins and it is well known now that the eastern US  
249 margin used for constraining Miller et al. (2005) and Kominz et al. (2008)'s curves underwent  
250 dynamic topography due to North America's overriding upon the Farallon plate slab (Kominz et al.,  
251 2008; Moucha et al., 2008; Müller et al., 2008). Further back in time, the resulting global sea-level  
252 amplitudes are consequently downward shifted by about 50 m with respect to continental flooding  
253 studies (Miller et al., 2011). The backstripped curves corrected from dynamic topography were  
254 discarded as they require for Eocene times either i) an unrecognised deformation event of the  
255 eastern US margin (Kominz et al., 2008) or ii) an unrealistically high global sea-level which requires  
256 melting of an ice-sheet volume three times higher than the present-day one (Rowley, 2013). For  
257 Pliocene times, the curve from Miller et al. (2011) based on oxygen isotope proxies of ice-volume  
258 changes was selected as it reflects ocean water volume variations, which mainly drives Pliocene sea-  
259 level changes (Miller et al., 2011).

260 The global sea-level curve of Müller et al. (2008) was selected for the "greenhouse" times before ca.  
261 35 Ma (Zachos et al., 2008). This method is however discarded for estimates of global sea-level  
262 changes since ca. 35 Ma because ocean water volume changes, which became the main driving-  
263 factor of global sea-level changes ("icehouse" period; Miller et al., 2011), are not considered (Müller  
264 et al., 2008). The Spasojevic and Gurnis (2012) curve was not selected because sea-level amplitudes  
265 match the overestimated ones of the Haq et al. (1987) and Haq and Al-Qahtani (2005) curves for the  
266 pre-35 Ma times (Fig.2).

267 The global sea-level curve of Rowley (2013), based on global hypsometry and global paleogeographic  
268 maps was selected for the entire period (i.e. since Upper Cretaceous) as it encompasses the effects  
269 of both ocean water and ocean basin volume changes on sea-level (Rowley, 2013). The use of global

270 data (paleogeography and hypsometry) lowers the influence of not having a stable reference for  
271 global sea-level measurements and removes local and regional effects (as the integral of dynamic  
272 topography over the Earth's surface must be zero, assuming a constant radius for the Earth; Rowley,  
273 2013). The dispersion of estimates from each paleogeographic datasets used (Scotese and Golonka,  
274 1992; Smith et al., 1994; Markwick, 2011; Blakey, 2012) is responsible for the uncertainty on the  
275 resulting global sea-level curve ( $\pm 50$  m for 100 to 60 Ma and  $\pm 20$  m for 60 to 0 Ma (Rowley, 2013)).

276 It is worth pointing out that i) from c.a. 100 to 35 Ma, Müller et al. (2008)'s curve which reflects  
277 ocean basin volume changes and ii) from c.a. 35 to 0 Ma, (Miller et al., 2005) and (Kominz et al.,  
278 2008)'s curves which reflect ocean water volume changes, both agree for these respective periods  
279 with Rowley (2013)'s curve which integrates the effects of both driving factors.

280

#### 281 **4. Application to the example of the Armorican Massif for Cenozoic times**

282

##### 283 *4.1. Regional setting and available data*

284 The Armorican Massif, located in western France, is a basement that was strongly deformed from  
285 late Devonian to Carboniferous times as a part of the Variscan belt similar to many other western  
286 European massifs (e.g. Massif Central, Rhenish Massif, Ardennes Massif; Ballèvre et al., 2009). This  
287 basement is surrounded by three major sedimentary basins that started subsiding during Mesozoic  
288 times: i) the Western Approaches Basin to the north, ii) the starved South Armorican Margin to the  
289 west and south and iii) the intracratonic Paris Basin to the east (Fig.4).

290 Like many other western European Variscan domains, the Armorican Massif corresponds today to a  
291 region of low topographic elevation ranging from 150 m to 200 m (highest peaks: 417 m). Its present-  
292 day topography is made of three main upland plateaus or highs of elevation above 200 m: the  
293 Western Brittany Plateau to the west, the Vendée High to the South and the Lower Normandy  
294 Plateau to the North (Fig.5). These uplands plateaus are connected by low elevation plateaus

295 (ranging from 30 m to 100 m) such as the Eastern Brittany Low (Fig.5; *Bessin et al., 2015*). This  
296 collection of low relief plateaus is inherited from six stepped planation surfaces which have been  
297 dated using the marine sedimentary remnants scattered over them. Their analysis reveals that the  
298 Armorican Massif was (partly?) buried and exhumed twice in response to western European  
299 intraplate deformation events (*Bessin et al., 2015*):

300 (1) A first burial event of the massif beneath marine sediments took place at a time of overall  
301 subsidence across western Europe, i.e. during Jurassic times;

302 (2) The first exhumation event occurred during the early Cretaceous at the time of initiation and  
303 break-up of the rift between Iberia and Eurasia;

304 (3) A second burial episode of the massif beneath chalk deposits took place during a second  
305 overall western European subsidence phase in the Late Cretaceous;

306 (4) A second exhumation episode occurred during latest Cretaceous to early Eocene times,  
307 resulting from differential uplift of the Armorican Massif induced by the convergence  
308 between the African and Eurasian plates.

309 The maximum depth of burial during each subsidence episode is thought to be low (< 500 m) as  
310 indicated by the small amount of coeval siliciclastic sediments in the surrounding basins.

311 Previous sedimentological and geomorphological studies (*Bonnet et al., 2000; Brault et al., 2004*)  
312 found that the Armorican Massif low elevation topography was later incised during two successive  
313 episodes of river network development in response to the convergence between African and  
314 Eurasian plates. The first drainage network developed during Late Miocene times and the resulting  
315 valleys were later filled by Piacenzian to Gelasian marine (but also continental) deposits (*Brault et al.,*  
316 *2004*). The present-day river network developed around the early to middle Pleistocene boundary.  
317 Up to 90 m of Pleistocene uplift has been estimated from measurements of the resulting incision  
318 (*Bonnet et al., 2000*).

319 However, no constraint is currently available regarding vertical movements that may have affected  
320 the Armorican Massif between early Eocene and Late Miocene times. Our purpose here is to quantify  
321 surface vertical movements over that period and identifying the processes that may have caused  
322 them.

323 The data we will use for this come from marine sedimentary remnants deposited during the main  
324 Cenozoic marine flooding events of the massif (exhaustive reference list regarding dataset is  
325 provided in Supplementary Material S1). Marine sediments were dated using i) biostratigraphic data  
326 (benthic and pelagic foraminifera, ostracoda, charophytes, macrofauna, pollens, spores and  
327 dinocysts; see Guillocheau et al., 2003 for a review) and ii) Electron Spin Resonance data for Pliocene  
328 times (see Van Vliet-Lanoë et al., 2002 for a review). Respective bathymetric estimates at the time of  
329 marine sediment deposition were obtained from both paleo-ecological (fossiliferous fauna and flora)  
330 and sedimentological data (see Guillocheau et al., 2003 for a review). Four separate depositional  
331 environment types were defined:

- 332 (1) Brackish environments with water depth ranging from ca. 0 to 5 m;
- 333 (2) Foreshore environments with water depth ranging between sea-level at mean high tide and  
334 at mean low tide; these include bays and open lagoons (i.e. ca. 0-20 m) or inner estuaries (i.e.  
335 ca. 0-10 m);
- 336 (3) Shoreface environments with water depth ranging between sea-level at mean low-tide and  
337 fair-weather wave-base (i.e. ca. 20-60 m);
- 338 (4) Upper offshore (open marine) shelf environments with water depth ranging between fair-  
339 weather wave-base and storm wave-base (i.e. ca. 60-100 m).

340 Deposits corresponding to six marine flooding events during Cenozoic times are preserved on the  
341 Armorican Massif (Guillocheau et al., 2003). They correspond to a series of relative sea-level high  
342 stand during (1) the Ypresian (early Eocene; ca. 56-48 Ma), (2) the Bartonian (late Eocene; 41.0-38.0

343 Ma), (3) the Rupelian (early Oligocene; 33.9-28.1 Ma), (4) the Langhian-Serravallian (middle Miocene;  
344 16.0-11.6 Ma), (5) the uppermost Miocene (Messinian?; 7.3-5.3 Ma) and the Piacenzian (early  
345 Pleistocene; 3.6-2.6 Ma). Bartonian, Rupelian, Langhian-Serravallian and Piacenzian marine  
346 sediments were used for quantifying vertical movement as they match the following requirements: i)  
347 they are accurately dated (resolution around or lower than that of a Stage on the chronostratigraphic  
348 chart), ii) they are well distributed upon the massif and iii) their bathymetry at the time of deposition  
349 is well constrained. The Ypresian flooding, which was restricted along the South Armorican Margin  
350 (Guillocheau et al., 2003), and the Messinian flooding, which is exposed on too few outcrops (Brault  
351 et al., 2004), were not selected.

352 Almost all of these deposits correspond to a thin veneer of sediments over pre-existing topography,  
353 that must relate to marine flooding with little to no contribution from compaction or isostatic  
354 subsidence by sediment loading. However, some of the Bartonian to Rupelian deposits are preserved  
355 in small narrow grabens bounded by N150E faults (e.g. 2 to 4 km width for 400 m depth for the  
356 largest one, the Rennes Basin; Fig.4; Bauer et al., 2016). Because we focus here on estimating  
357 medium to long wavelength surface subsidence, these short wavelength deformation gradients were  
358 restored by assuming that the present-day elevation of the top of these basins can be used as a  
359 proxy for the elevation of the sediments at the time of their deposition.

360

## 361 *4.2. Vertical movement of the Armorican Massif*

362

### 363 *4.2.1. Finite vertical movements*

364 *fvm* calculations were performed for each global sea-level dataset available and for each timespan  
365 for which data match requirements for *fvm* computation (data available and plotted on  
366 Supplementary Material S1 and S3, respectively). Hereafter, we only present computations from the  
367 global sea-level curves that we selected for each time interval (see §3.2). The *fvm* values listed in the

368 text correspond to the mean *fvm* values computed (minimum and maximum values are plotted on  
369 Fig.6).

370 The *fvm* calculations point to an overall subsidence of the Armorican Massif from 41.0-38.0 Ma  
371 (Bartonian) to present-day, except in the Northern (Trégor area) and Eastern regions (Fyé Basin;  
372 Fig.6; uplift of ca. 50 m). This subsidence is of long wavelength and its magnitude ranges from i) -  
373 131.4 to -20.4 m using the Müller et al. (2008) global sea-level data to ii) -95.4 to 15.6 m using the  
374 Rowley (2013) global sea-level data. Both indicate a low differential subsidence (of ca. 100 m)  
375 between present-day offshore and onshore domains. This Armorican-scale differential subsidence  
376 characterized a deformation process with a wavelength of ca. 300 km, i.e. a medium wavelength  
377 deformation.

378 Since 33.9-28.1 Ma (Rupelian), the *fvm* values suggest an overall uplift with a differential component  
379 between the central (Eastern Brittany Low, Léon Platform) and the northern parts (Western  
380 Approaches Basin and Carentan Flats) of the massif (Fig.6). The magnitude of this uplift ranges from i)  
381 33.8 to 97.0 m using the Kominz et al. (2008) global sea-level data to ii) 39.0 and 109.4 m using the  
382 Rowley (2013) global sea-level data for the central part of the massif. Uplift is lower in the western  
383 part of the massif with values of 25.1 m using the Kominz et al. (2008) global sea-level data and 21.7  
384 m using the Rowley (2013) global sea-level data. Conversely, the northernmost part of the studied  
385 area (Western Approaches and Carentan Flats) exhibits stronger subsidence with a magnitude of i)  
386 78.2 m using the Kominz et al. (2008) global sea-level data and ii) 73.0 using the Rowley (2013) global  
387 sea-level data. These values may be underestimated because the Western Approaches Basin  
388 sediment thickness estimates were not decompacted. At a regional scale, these estimates suggest a  
389 doming of the Armorican massif with up to ca. 180 m (175.2 m and 182.4 m respectively using the  
390 Kominz et al. (2008) and the Rowley (2013) global sea-level data) of differential vertical movement  
391 between the dome apex and its edges.

392 A finite uplift is evidenced by our *fvm* computations (Fig.6) since 16.0-11.6 Ma (Langhian-  
393 Serravallian). Highest uplift magnitudes are located i) north of the central part of the Armorican  
394 Massif (North East of the Eastern Brittany Low) with up to 127.0 m to 130.7 m of uplift respectively  
395 computed using the Kominz et al. (2008) and the Rowley (2013) global sea-level data and ii) in a  
396 lesser degree, south of the central part of the Massif (East of Vendée Low) with magnitudes up to  
397 117.5 m from the (Kominz et al., 2008) global sea-level data and 121.2 m from the (Rowley, 2013)  
398 global sea-level data. Lower values are found to the north of the massif (Carentan Flats), to the west  
399 of the central part of the massif (West of the Eastern Brittany Low) and to the south of the massif  
400 (Vendée Low) with values ranging from ca. 30 m to 70 m using both Kominz et al. (2008) and Rowley  
401 (2013) global sea-level data.

402 Since 3.6-2.6 Ma (Piacenzian), an overall uplift of the Armorican Massif is suggested by the computed  
403 *fvm* values (Fig.6, S1). Higher magnitudes are located west of the Eastern Brittany Low and north of  
404 the southern branch of the SASZ, with up to 153,4 m according to the Miller et al. (2011) global sea-  
405 level data and 141.0 m when using the Rowley (2013) global sea-level data. Lower magnitudes of  
406 uplift ranging from ca. 55 m (Eastern Brittany Low) to ca. 70 m (northwestern and southwestern part  
407 of the massif) are found according to both Miller et al. (2011) and Rowley (2013) global sea-level  
408 data. Conversely, the northern part of the massif (Carentan Flats) is the only area of predicted  
409 subsidence since Piacenzian times, with subsidence values of ca. 25 m to 40 m using the Rowley  
410 (2013) and the Miller et al. (2011) global sea-level data, respectively.

411

#### 412 4.2.2. *Intra-Cenozoic vertical movements*

413 Three areas of the Armorican Massif contain well dated marine sediments of different ages  
414 (Bartonian, Rupelian, Langhian-Serravalian, Gelasian) that are close to each other and unaffected by  
415 faulting or post-depositional erosion. These are the Carentan Flat (in its northern part), the Eastern  
416 Brittany Low (in its central part) and the Vendée Low (in its southern part; Fig.5, Fig.7). Using the *fvm*

417 values computed above, the intra-Cenozoic vertical movement recorded by these domains were  
418 estimated using global sea-level curves as discussed above, i.e. Müller et al. (2008) and Rowley (2013)  
419 until Bartonian times, Miller et al. (2005), Kominz et al. (2008) and Rowley (2013) for Rupelian to  
420 Miocene times and Miller et al. (2011) and Rowley (2013) since Pliocene times. Finally, the successive  
421 elevations of the topographic surfaces at Bartonian, Rupelian, Langhian-Serravalian and Gelasian  
422 times were restored for each sea-level curve (data available in Supplementary Material S4). The  
423 amplitude of the vertical movements are low and three main results are obtained (values hereafter  
424 listed are means of estimated vertical movement from selected global sea-level curves calculations  
425 plotted on Fig.7):

426 - From 38 to 34 Ma (Priabonian), a phase of low subsidence is suggested for the northern  
427 (Carentan Flats) and southern (Vendée Low) parts of the massif. Values range between 7.5  
428 and 15.2 m of subsidence over that period (i.e. a subsidence rate of 1.3 to 2.1 m Ma<sup>-1</sup>). The  
429 38-34 Ma times are at the transition between the Müller et al. (2008)'s curve and the  
430 backstripped curves (Miller et al., 2005; Kominz et al., 2008) suitable periods which may  
431 introduce some bias. As no significant vertical movement is evidenced from the Rowley  
432 (2013)'s data (Fig.7), a phase of stability is therefore privileged. The central part of the massif  
433 (Eastern Brittany Low) underwent 65.9 m of subsidence, possibly overestimated due to fault  
434 gradients restoration.

435 - From 30 to 3.6 Ma (Rupelian to Pliocene times), the three domains record subsidence, which  
436 we infer as evidence for subsidence of the entire Armorican Massif. Between 30 and 16 Ma  
437 (Rupelian to Langhian-Serravalian times), vertical movement values ranges between 8.0 m of  
438 uplift (possibly overestimated due to fault gradients restoration) and 41.9 m (or rates of 0.4  
439 m Ma<sup>-1</sup> of uplift to 2.2 m Ma<sup>-1</sup> of subsidence). From 12 to 3.6 Ma (Serravalian to Piacenzian  
440 times), subsidence values range from 11.0 m (or a subsidence rate of 1.0 m Ma<sup>-1</sup>) in the

441 central part of the massif (Eastern Brittany Low) to 50.1 m (or subsidence rate of 4.7 m Ma<sup>-1</sup>)  
442 in its northern part (Carentan Flats).

443 - During the last 2.6 Ma (Piacenzian time to present), a late phase of uplift of the northern  
444 (Carentan Flats) and central (Eastern Brittany Low) parts of the massif is inferred with uplift  
445 values ranging from 48 to 89 m (or uplift rates of 15.5 to 28.8 m Ma<sup>-1</sup>). No data is available  
446 for the southern part of the massif (Vendée Low) but 38 m of uplift can be inferred over the  
447 past 12 Ma (Serravalian to present) which corresponds to an uplift rate of 2.7 m Ma<sup>-1</sup>.

448

#### 449 **5. Armorican Cenozoic vertical movements within the Western European tectonic framework**

450

451 From the uppermost Cretaceous to the early Cenozoic, the Armorican Massif is exhumed in response  
452 to medium wavelength ( $\times 10^2$  km) uplift which affected the overall NW European platform and  
453 marked the end of deposition and the deformation of the Upper Cretaceous chalk platform (Ziegler,  
454 1990; Anell et al., 2009).

455 During Bartonian times (41 – 38 Ma), homogeneous sedimentary facies preserved on the massif and  
456 in surrounding basins (Bauer et al., 2016) point out to a nearly flat and low Armorican topography,  
457 suggesting a phase of no deformation, which extended through to Priabonian times (ca. 34 Ma) as  
458 evidenced by our vertical movement estimates. This Bartonian to Priabonian phase (41 – 34 Ma) is  
459 coeval with the period of no deformation that affected most of north-western Europe during Eocene  
460 times, except for offshore Britain and the northern North Sea, which experienced anomalous  
461 subsidence possibly related to the development of the Iceland thermal anomaly (see Anell et al.,  
462 2009 for a review).

463 The Rupelian to Piacenzian (30 – 3.6 Ma) phase of slow subsidence of the Armorican Massif  
464 evidenced by our computations (Fig.7) is likely to be related to the growth of numerous small  
465 sedimentary basins during Oligocene to middle Miocene times (ca. 35 – 10 Ma) along the western

466 side of the British Isles (Cornwall, Wales, northern Ireland and the Hebrides Sea; Walsh, 1999) in a  
467 largely strike-slip regime, which led to local basin inversions (Williams et al., 2005). This low  
468 subsidence phase of the massif is also coeval with Oligocene (Eocene?) to Miocene short-wavelength  
469 deformation observed in surrounding basins and on the northwestern European platform (Anell et  
470 al., 2009) that includes i) strike-slip to compressive folding along the South Armorican Margin  
471 (Guillocheau et al., 2003), ii) major basin inversion (e.g. up to 700 m of reverse fault movement) in  
472 the Western Approaches Basin (Le Roy et al., 2011) and iii) NNE-SSW striking left-lateral  
473 transtensional wrenching of the European Cenozoic Rift System (Bourgeois et al., 2007), all of which  
474 are thought to be related to reactivation of pre-existing structures by in-plane stresses (e.g. Anell et  
475 al., 2009). Reactivation of these structures are superimposed on a medium wavelength deformation  
476 that initiated around 35 Ma and is thought to be related to lithospheric mantle buckling in response  
477 to the Apulia-Eurasia collision (Handy et al., 2010; Cloetingh et al., 2015). The paroxysm of this  
478 buckling is thought to have taken place around 17 Ma (Burdigalian) with the development of folds at  
479 a wavelength of ca. 225-275 km (Bonnet et al., 2000; Bourgeois et al., 2007). Evidence for this  
480 buckling includes uplift of the Bohemian Massif and the Vosges-Black Forest Arch and amplification  
481 of uplift of the Massif Central that initiated at the Oligocene-Miocene transition in response to  
482 thermal thinning of the lithosphere. Conversely, we propose here that the Armorican Massif is  
483 possibly located on a lithospheric-scale syncline which induced the subsidence of the massif from  
484 Rupelian to Piacenzian times (Fig.6 and Fig.7), i.e. from 30 to 3.6 Ma.

485 The Pleistocene uplift (2.6 – 0 Ma; of ca. 50 to 90 m; Fig.6) that we evidence is consistent with  
486 previous geomorphic studies of the Armorican Massif (Bonnet et al., 2000; Brault et al., 2004) which  
487 preclude a GIA origin from geomorphic data (Bonnet et al., 2000). This recent uplift is also observed  
488 in western Europe, e.g. in the Paris Basin (Antoine et al., 2007), the Ardennes Massif or the Rhenish  
489 Shield (Demoulin and Hallot, 2009). It is commonly thought to be related to either an enhanced  
490 convergence rate of the Apulia-Eurasia collision at the early–middle Pleistocene transition or a

491 climate-induced increase in erosion rate that led to topographic unloading and a change in stress  
492 regime (Cloetingh et al., 2015; Herman and Champagnac, 2016).

493

## 494 **6. Conclusion**

495

496 (1) We formalized a method to quantify low amplitude vertical movements which are difficult to  
497 document using low-temperature thermochronology, OSL dating or cosmogenic isotope methods.  
498 Our method is based on estimating the difference the present-day elevation of well dated marine  
499 sediments (corrected from their bathymetry of deposition) and selected global sea-level  
500 reconstructions at the time of sediment deposition.

501 (2) We compiled available global sea-level curves and re-assess their reliability to build a composite  
502 one. Considering the various processes that may have caused global sea-level changes through time,  
503 we are able to disregard some of the global sea-level curves because of a clear bias they introduce or  
504 because the method used to construct them is inapplicable for the time period considered. We  
505 concluded that the Müller et al (2008) 's curve is suitable for the ca. 100 to 35 Ma "greenhouse"  
506 period while the Miller et al. ( 2005) and the Kominz et al. (2008)'s curves better reflect global sea-  
507 level changes during the ca. 35 to 0 Ma "icehouse" period. We also note that both agree with Rowley  
508 (2013)'s curve in their respective period of optimum reliability.

509 (3) Based on our estimates of amplitude, wavelength and timing of the patterns of uplift/subsidence  
510 affecting the Armorican Massif, we suggest that lithospheric buckling related to the Apulia-Eurasia  
511 convergence is responsible for the medium-wavelength deformation we identify. More precisely, the  
512 Armorican Massif underwent i) a phase of tectonic quiescence characterizing most of NW Europe  
513 during Bartonian to Priabonian times (38 – 34 Ma) followed by ii) a phase of low subsidence during  
514 Rupelian to Piacenzian times (30 – 3.6 Ma) possibly due to the position of the massif within a  
515 downing limb of a lithospheric scale buckling instability driven by the Apulia-Eurasia convergence and

516 iii), most recently, a Pleistocene (2.6 – 0 Ma) phase of uplift related to either the intensification of the  
517 Africa-Apulia convergence or a climate-induced erosional enhancement of this long-term uplift.

518

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654

655 **Figure captions:**

656 **Fig.1:** Sketch illustrating our finite vertical movement quantification methodology based on i) the  
657 bathymetry of deposition and present-day elevation of well dated marine sediments versus ii)  
658 elevation of coeval global sea-level.

659 **Fig.2:** Compilation of global sea-level curves since the Upper Cretaceous; vertical bars indicate the  
660 range of acceptable sea-level elevations for each curve and stage.

661 **Fig.3:** Global sea-level curves since Upper Cretaceous times selected for their suitability in  
662 representing reliable proxy of global sea-level (see text for the details of the selection procedure).

663 **Fig.4:** Synthetic geological map of the Armorican Massif and Mesozoic to Cenozoic surrounding  
664 basins. Note the Cenozoic marine transgressions on the massif and the Eocene-Oligocene basins  
665 scattered around the massif (data from 1:1.000.000 Geological Map of France (Chantraine et al.,  
666 2003); Projection: RGF Lambert 1993). NASZ is North Armorican Shear Zone and SASZ is South  
667 Armorican Shear Zone.

668 **Fig.5:** Location map of marine sedimentary deposits used to quantify Cenozoic vertical movements of  
669 the Armorican Massif. Black lines: faults from 1:1.000.000 Geological Map of France (Chantraine et  
670 al., 2003; Projection: RGF Lambert 1993). White lines: Border of Basement outcrops. Red line:  
671 present-day coastline. Fy. : Fyé Basin, Tr.: Trégor Platform, C.F.: Carentan Flat, W.B.P.: Western  
672 Brittany Plateau, E.B.L.: Eastern Brittany Low, V.L.: Vendée Low.

673 **Fig.6:** Map illustrating computed finite vertical movement based on selected curves for global sea-  
674 level change, i.e. Müller et al. (2008) and Rowley (2013) for the Bartonian, Kominz et al. (2008) and  
675 Rowley (2013) for the Rupelian, Kominz et al. (2008) and Rowley (2013) for the Langhian-Serravallian  
676 and Miller et al. (2011) and Rowley (2013) for the Piacenzian-Gelasian Rupelian. Black lines: faults  
677 from 1:1.000.000 Geological Map of France (Chantraine et al., 2003; Projection: RGF Lambert 1993).  
678 White lines: Border of Basement outcrops. Red line: present-day coastline.

679 **Fig.7:** Predicted Cenozoic vertical movement and restored elevations for three Armorican lows,  
680 namely the Carentan Flats and the Eastern Brittany and Vendée Lows.

681 **Supplementary material:**

682 **S1:** Dataset (.xlsx file) used for i) our compilation of global sea-level curves and ii) quantify finite  
683 vertical movements of the Armorican Massif since Cenozoic times.

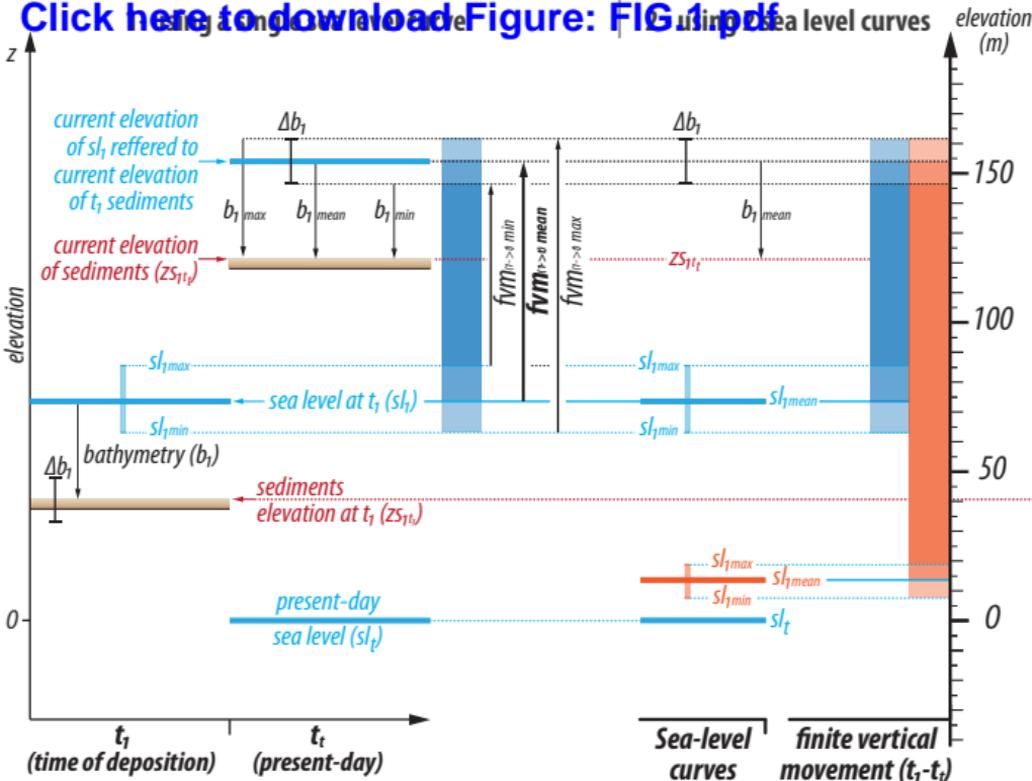
684 **S2:** Plot of global sea-level curves discussed in this study. Minimum, maximum and mean values were  
685 computed and plotted for each curve and each stage.

686 **S3:** Geographical plots of finite vertical movement (fvm) estimated for the Armorican Massif for each  
687 available curve.

688 **S4:** Dataset (.xlsx file) of vertical movement and restored elevations through Cenozoic times for three  
689 Armorican lows, namely the Carentan Flats and the Eastern Brittany and Vendée Lows.

# Figure 1 VERTICAL MOVEMENT (low sedimentary preservation)

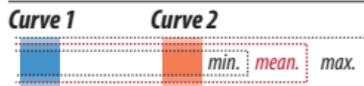
[Click here to download Figure: FIG 1.pdf](#) using sea level curves



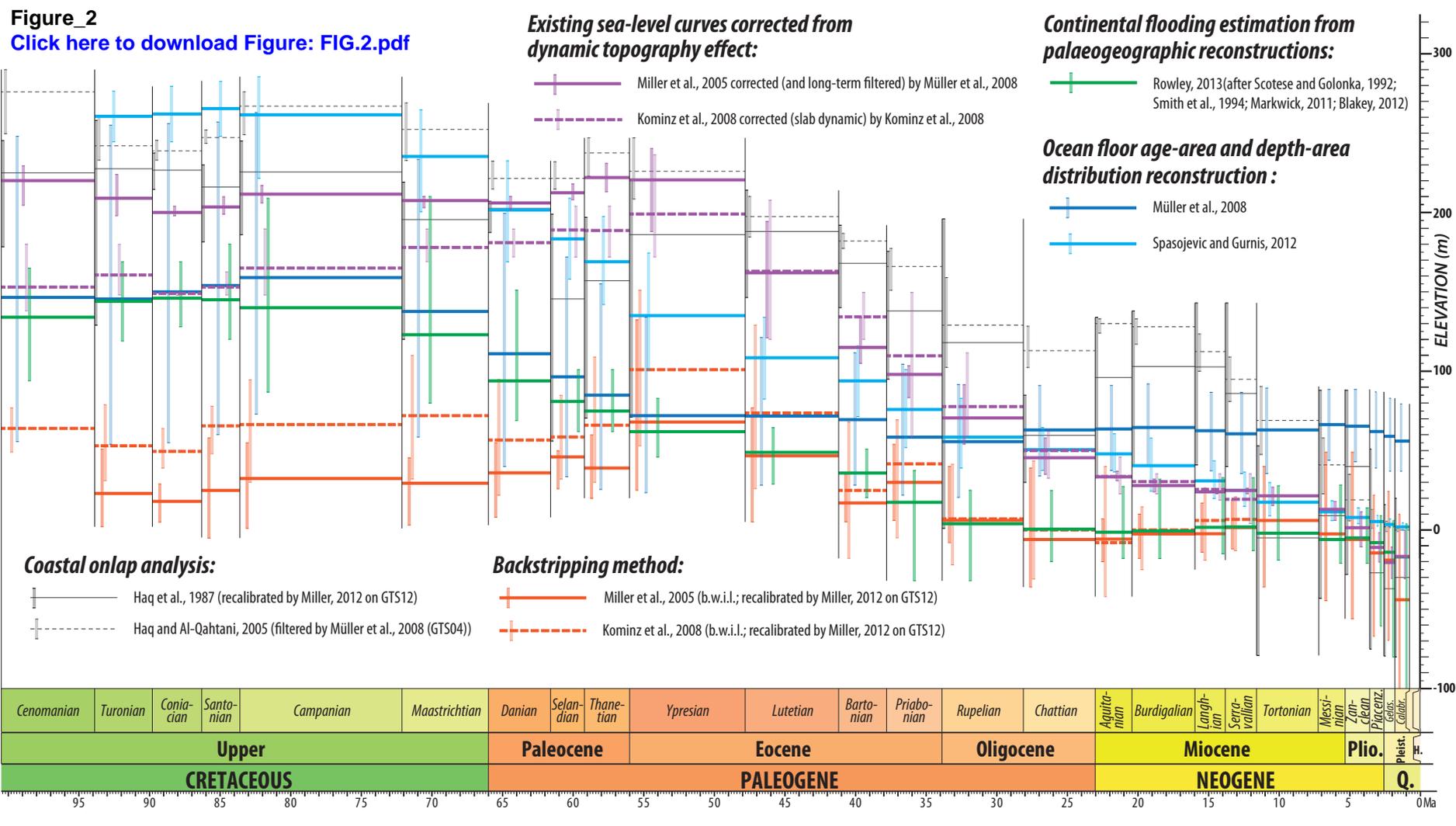
## Sea-level and paleo-elevation data



## Finite vertical movement (fvm)



**Figure\_2**  
[Click here to download Figure: FIG.2.pdf](#)



Figure\_3

[Click here to download Figure: FIG.3.pdf](#)

**Backstripping method:**

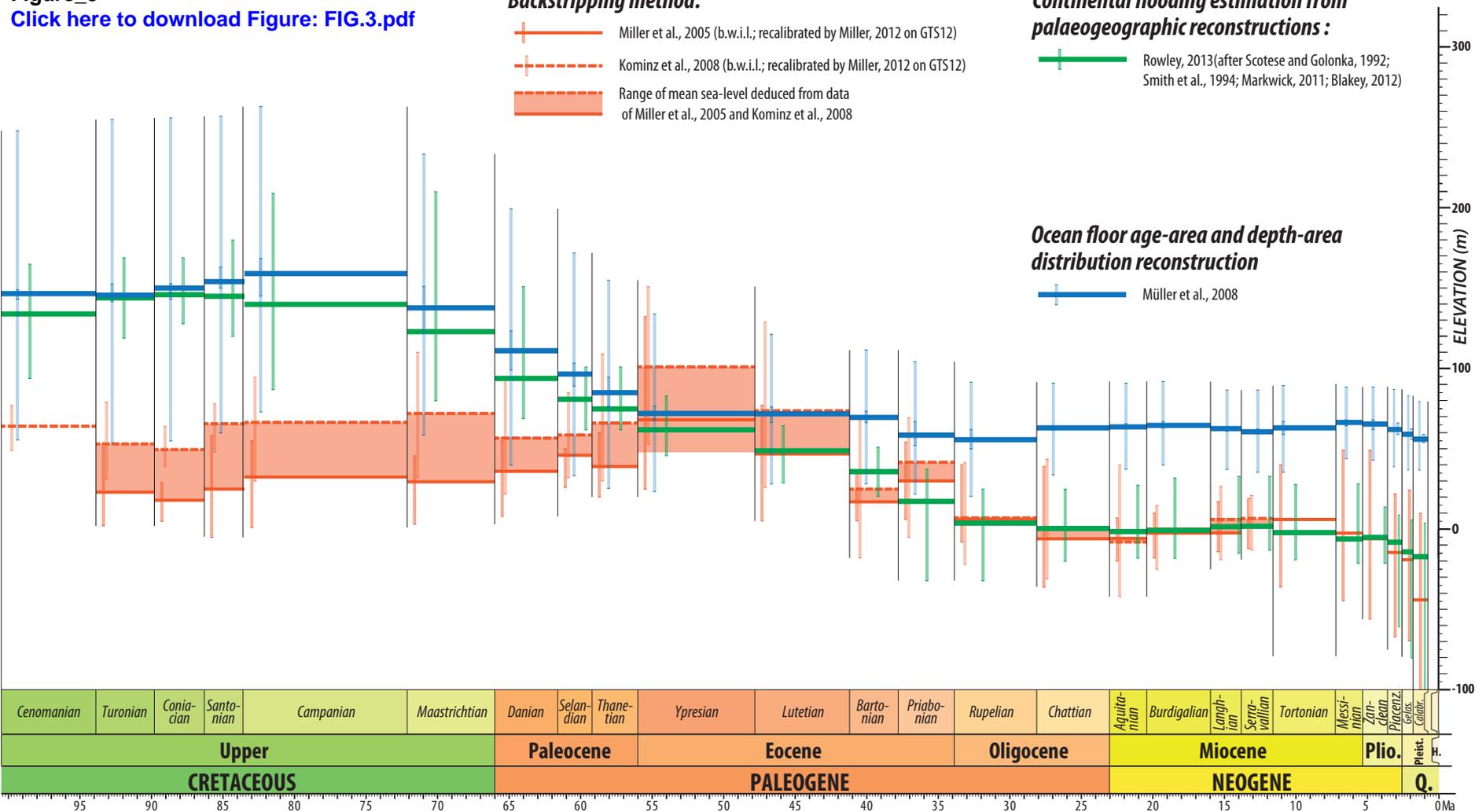
-  Miller et al., 2005 (b.w.i.l.; recalibrated by Miller, 2012 on GTS12)
-  Kominz et al., 2008 (b.w.i.l.; recalibrated by Miller, 2012 on GTS12)
-  Range of mean sea-level deduced from data of Miller et al., 2005 and Kominz et al., 2008

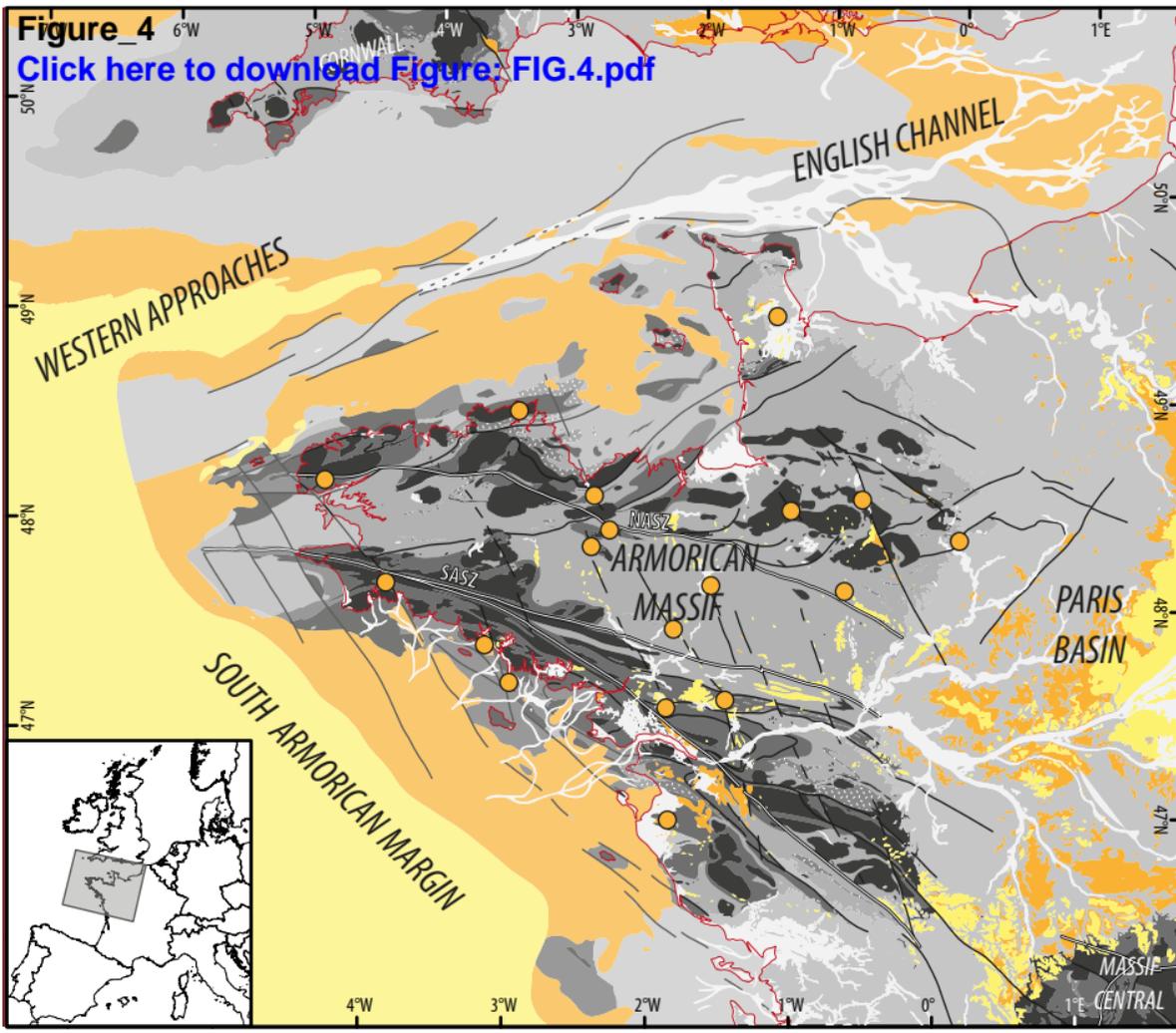
**Continental flooding estimation from palaeogeographic reconstructions :**

-  Rowley, 2013 (after Scotese and Golonka, 1992; Smith et al., 1994; Markwick, 2011; Blakey, 2012)

**Ocean floor age-area and depth-area distribution reconstruction**

-  Müller et al., 2008





**VARISCAN BASEMENT**

-  Plutonic rocks  
*granites, etc.*
-  Metamorphic rocks  
*micaschists, gneiss, etc.*
-  Volcanic rocks  
*basalts, etc.*
-  Sedimentary rocks  
*schists, sandstones, etc.*

**SURROUNDING BASINS**

-  Mesozoic series
-  Paleogene series  
*Eocene-Oligocene onshore basins*
-  Neogene series
-  Quaternary

**STRUCTURES**

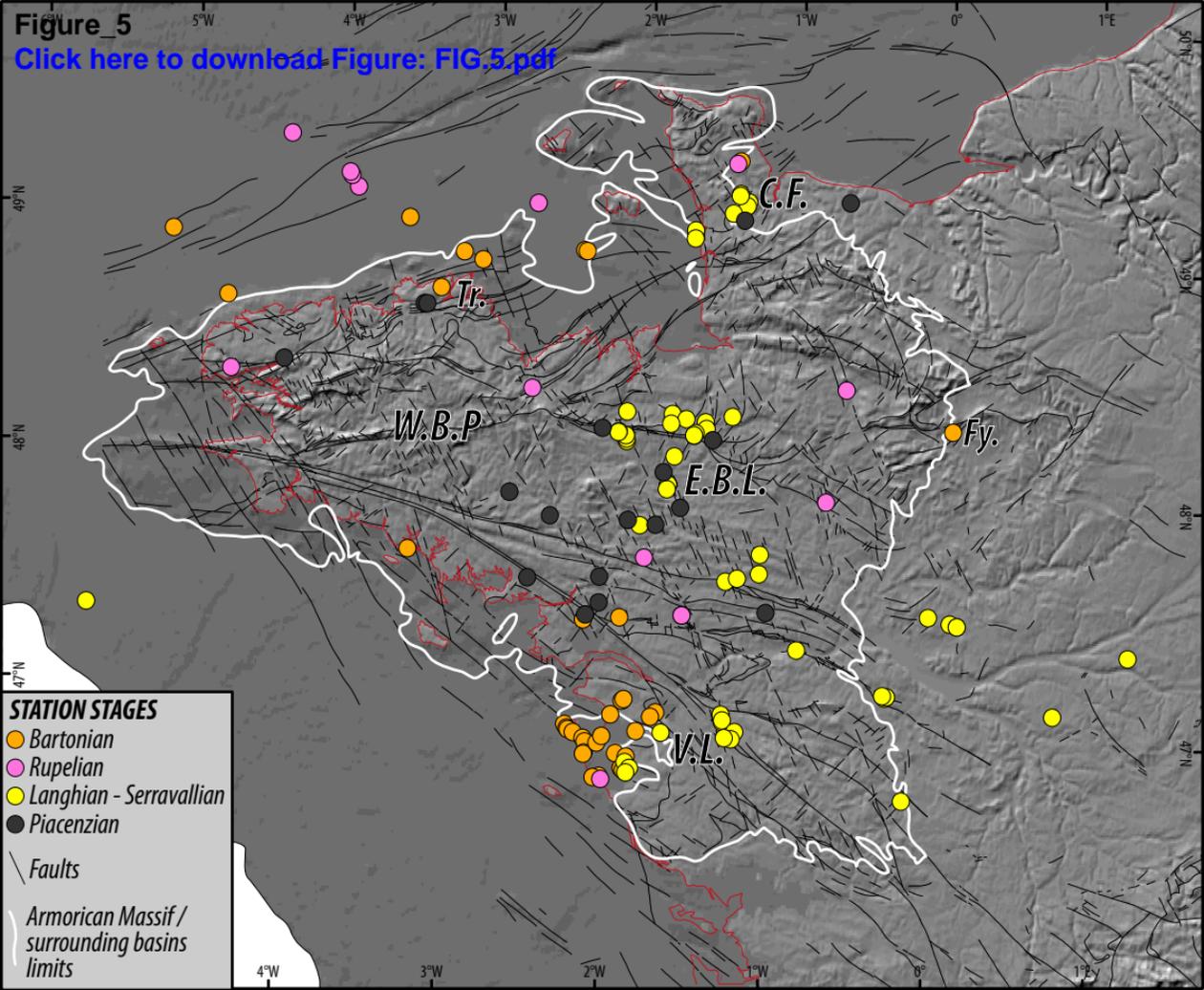
-  main fault  
SASZ: South Armoricain Shear Zone  
NASZ: North Armoricain Shear Zone
-  minor fault
-  hidden fault

**COASTLINE**

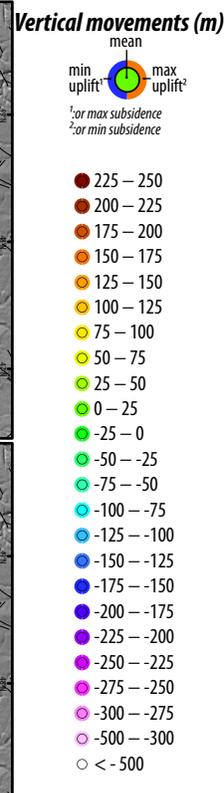
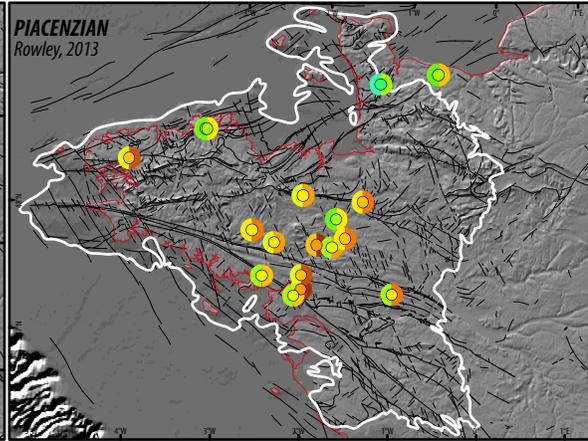
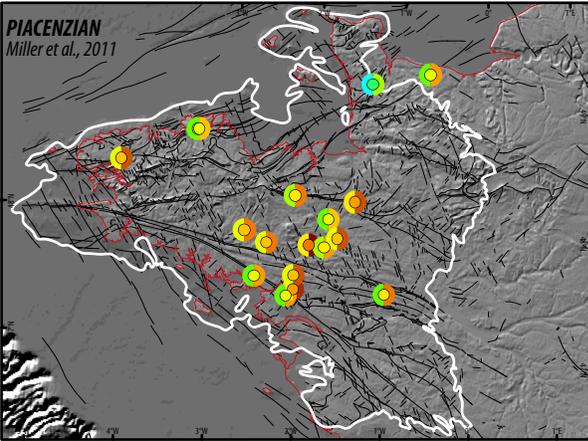
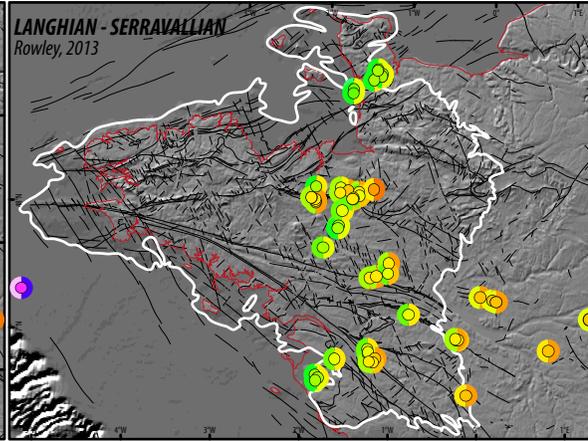
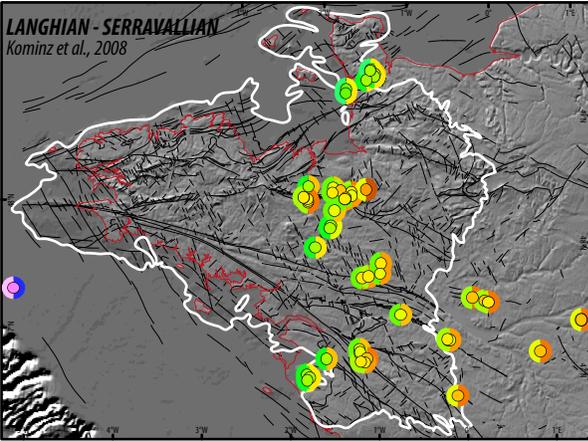
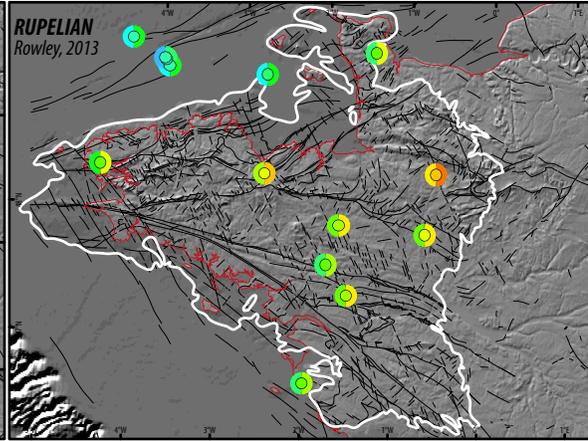
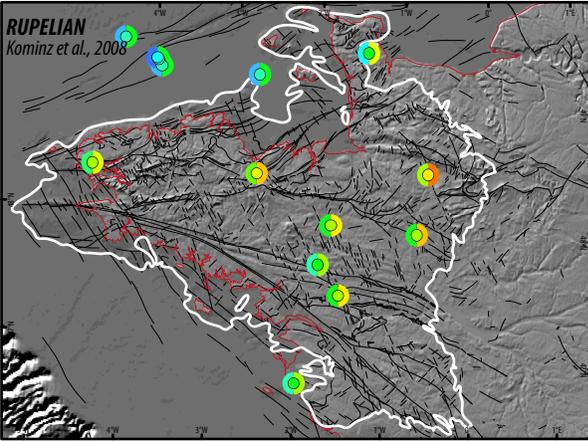
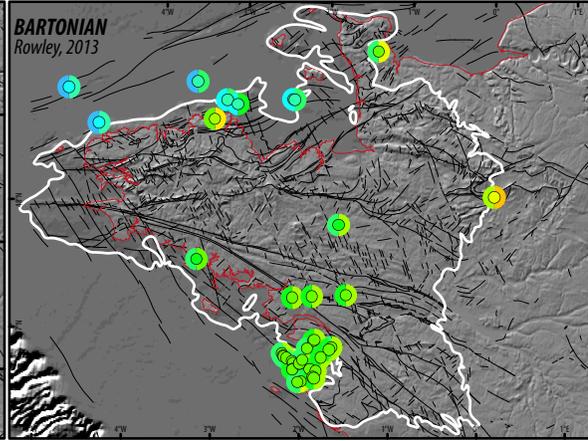
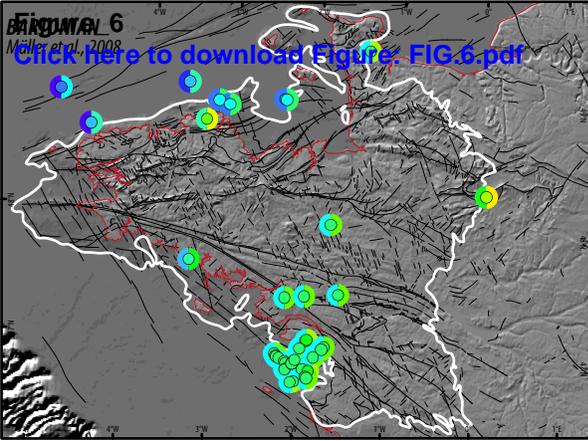
-  onshore
-  offshore
-  coastline

**Figure\_5**

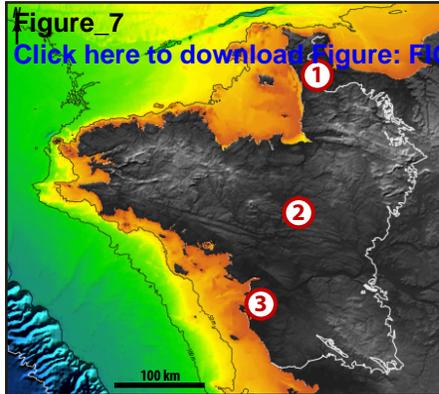
[Click here to download Figure: FIG.5.pdf](#)



**BARDOLIAN 6**  
 Müller et al., 2008  
[Click here to download Figure FIG. 6 pdf](#)



**Figure\_7**  
[Click here to download Figure: FIG.7.pdf](#)



**Location**

- ① Carentan Flats
- ② Eastern Brittany Low
- ③ Vendée Low

**Successive elevation of the topographic surfaces calculated from fvm calculation based on:**

Deep-sea benthic foraminifera  $\delta^{18}O$  proxy (only for Pliocene times):

— Miller et al., 2011

**Backstripping method:**

— Miller et al., 2005  
 (recalibrated by Miller, 2013 on GTS12)

- - - Kominz et al., 2008  
 (recalibrated by Miller, 2013 on GTS12)

**Ocean floor age-area and depth-area distribution reconstruction:**

+ Müller et al., 2008

**Continental flooding estimation from palaeogeographic reconstructions:**

+ Rowley, 2013

**Intra-Cenozoic vertical movements**



Amplitude (m) of vertical movement:  
 Rates ( $m Ma^{-1}$ ) of vertical movement:  
 (highest value; -lowest value)

★ Present-day elevations of the topographic surfaces

