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H. Baghli, E. Mattioli, J.E. Spangenberg, M. Bensalah, F. Arnaud-Godet, B. Pittet, G. Suan

► **To cite this version:**

H. Baghli, E. Mattioli, J.E. Spangenberg, M. Bensalah, F. Arnaud-Godet, et al.. Early Jurassic climatic trends in the south-Tethyan margin. *Gondwana Research*, 2020, 77, pp.67 - 81. 10.1016/j.gr.2019.06.016 . hal-03487741

**HAL Id: hal-03487741**

**<https://hal.science/hal-03487741>**

Submitted on 20 Dec 2021

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## 1 **Early Jurassic climatic trends in the south-Tethyan margin**

2

3 Baghli, H.<sup>1,2</sup>, Mattioli, E.<sup>1,3</sup>, Spangenberg, J.E.<sup>4</sup>, Bensalah, M.<sup>2</sup>, Arnaud-Godet, F.<sup>1</sup>,  
4 Pittet, B.<sup>1</sup>, Suan, G.<sup>1</sup>

5

6 <sup>1</sup> Université de Lyon, UCBL, ENSL, CNRS, LGL-TPE, 69622 Villeurbanne, France.

7 <sup>2</sup> Laboratoire de Recherche n° 25, Université de Tlemcen, Algeria.

8 <sup>3</sup> Institut Universitaire de France, Paris, France.

9 <sup>4</sup> Institute of Earth Surface Dynamics (IDYST), University of Lausanne, Building  
10 Géopolis, CH-1015 Lausanne, Switzerland.

11

### 12 **Abstract**

13

14 The Early Jurassic period was characterized by extreme environmental changes, as  
15 reflected by major global carbon isotope anomalies and abrupt changes in oxygen  
16 isotope and elemental records of marine organisms. Available data suggest an overall  
17 warm Early Jurassic climate interrupted by periods of severe cooling, with a climatic  
18 optimum during the early Toarcian. Available geochemical studies, however, have  
19 mainly focused on the northern margin of the Tethys Ocean, so that the  
20 palaeogeographic extent of these environmental perturbations, latitudinal  
21 palaeotemperature gradients and climate belt boundaries remain poorly constrained.  
22 Here we report the first stable isotope records of brachiopod shells ( $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$   
23 values) from the Upper Sinemurian-Middle Toarcian interval in the southern margin of  
24 the Tethys Ocean (northwest Algeria). These data were used to better constrain the  
25 palaeoenvironmental evolution of the North Gondwana margin during the Early  
26 Jurassic, which likely played an important role on supra-regional climate. The  
27 diagenetic history of the analysed brachiopod shells was monitored using scanning  
28 electron microscopy, and elemental (manganese and strontium) compositions. The  
29 brachiopod  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  data show very similar trends as those reported for various  
30 Tethyan regions, and record negative carbon and oxygen isotope excursions near the  
31 Sinemurian–Pliensbachian and Pliensbachian–Toarcian transitions and during the  
32 Toarcian oceanic anoxic event (T-OAE). Despite these similarities, the carbon and  
33 oxygen isotope records are systematically offset towards more positive  $\delta^{13}\text{C}$  values

34 (average +0.5‰) and more negative  $\delta^{18}\text{O}$  values (-1.0‰) compared to those obtained  
35 from sites of higher palaeolatitudes in the northern Tethyan margin. These offsets  
36 suggest a spatial heterogeneity in the stable isotope composition of dissolved inorganic  
37 carbon in the Early Jurassic Ocean and a marked latitudinal temperature gradient  
38 between the southern and northern margins of the Tethys.

39

40 *Keywords:* Carbon and Oxygen stable isotopes, Early Jurassic, NW Algeria,  
41 Palaeoclimate

42

### 43 **1. Introduction**

44

45 The Early Jurassic period (~ 200-175 Ma) has been intensively studied because it was  
46 a key time interval for the environmental evolution of the western Tethys. Indeed, this  
47 time interval is characterized by extreme climate fluctuations, with possible glacial  
48 interludes occurring within an interval of general greenhouse conditions (Price, 1999).  
49 This interval is marked by major events in the large igneous provinces of the Central  
50 Atlantic and Karoo-Ferrar (e.g., Wignall, 2001), and the gradual disintegration of the  
51 Pangea supercontinent (Olsen, 1997). Severe perturbations of the carbon cycle are  
52 reflected by large fluctuations of the stable carbon isotope composition ( $\delta^{13}\text{C}$ ) of marine  
53 and terrestrial material. Severe biotic crises, such as the first-order extinction of the  
54 end-Triassic (Wignall, 2001) or the Toarcian oceanic anoxic event (T-OAE; Jenkyns,  
55 1988) produced marked floral and faunal turnovers (Harries and Little, 1999; Macchioni  
56 and Cecca, 2002; Mattioli et al., 2008; Suan et al., 2008a, 2010).

57 Numerous studies have focused on the palaeoenvironmental and biological  
58 perturbations in sites from the northwestern Tethys (e.g., Harries and Little, 1999;  
59 Hesselbo et al., 2000a; Jenkyns et al., 2002; Gomez et al. 2008; Harazim et al., 2013;  
60 Ruebsam et al., 2018). However, climate conditions during the Early Jurassic in the  
61 Tethyan southern margin are very poorly known (Bodin et al., 2010, 2016; Reolid et  
62 al., 2012; Krencker et al., 2014; Ait-Itto et al., 2017). The North Gondwana is  
63 considered to be characterized by a more arid climate with respect to the NW Tethys  
64 (Aberhan, 2001; van de Schootbrugge et al., 2005). The carbon and oxygen isotope  
65 trends have been documented for bulk carbonate, belemnites and brachiopods from  
66 uppermost Pliensbachian and Toarcian strata of Morocco (Bodin et al., 2010; Krencker  
67 et al., 2014; Ait-Itto et al., 2017). High-resolution carbon isotope records of bulk

68 carbonate and associated organic matter in micrite samples from the upper  
69 Pliensbachian and lower Toarcian intervals in three sections from the Central High  
70 Atlas Basin in Morocco were presented and discussed by Bodin et al. (2016). Apart  
71 from these records, there is a desperate lack of information about the climate and  
72 environmental evolution in North Gondwana during most of the Early Jurassic interval.  
73 The distribution of climate belts and the role played by North Africa in Early Jurassic  
74 climate evolution remain therefore poorly understood.

75 This study presents new carbon and oxygen isotope measurements of brachiopod  
76 shell calcite from sections in NW Algeria, which was a marginal part of the NW  
77 Gondwana. The data set spans the Early Jurassic interval, from the Upper Sinemurian  
78 to the Middle Toarcian. Well-preserved articulate brachiopod shells are considered as  
79 some of the most pertinent materials for geochemical analyses, as modern  
80 representatives are strictly sessile and benthic organisms that precipitate their calcite  
81 shell in close chemical and isotopic equilibrium with ambient seawater (e.g., Veizer et  
82 al., 1999). Therefore, their stable isotope record can safely be interpreted in terms of  
83 temperature and changes in the isotopic composition of the dissolved inorganic carbon  
84 (DIC) pool through time. The isotope data obtained in NW Algeria are here compared  
85 to the isotopic records from various NW Tethys settings in order to discriminate  
86 between local *versus* global fluctuations of environmental parameters during the early  
87 Jurassic.

88

## 89 **2. Geological background**

90

91 The samples studied in this paper were collected from five different sites, all located in  
92 NW Algeria, which in the Jurassic was part of the south-west part of the western Tethys  
93 (Fig. 1A). These sites are separated longitudinally by a distance of a few hundred  
94 kilometres, but were approximately situated at the same palaeolatitude (Fig. 1). Most  
95 of the brachiopod samples come from the Alméras collection (Almeras et al., 2007),  
96 hosted in the Collections de Géologie de Lyon. These specimens are stratigraphically  
97 well-constrained at the ammonite Zone to Subzone level, and come from 22 sections  
98 in the Traras, Rhar Roubane, Saïda, Ouarsenis and Nador Mountains (Fig. 1B). These  
99 regions are part of the Tello-Rifain chains, which constitute the western Mediterranean  
100 part of the Alpine chain and are separated from the Sahara craton by a narrow, tectonic  
101 corridor extending eastwards from Morocco (Agadir) to Tunisia (Gabes; Wildi, 1983).

102 Within the Tlemcen Domain, the Traras Mountains are bounded to the north by the  
103 Mediterranean Sea, to the east by the Mio-Pliocene and Quaternary basin of Tafna, to  
104 the west by the Beni Snassen Massifs (Eastern Morocco), and to the south by the  
105 Maghnia depression. The Traras mountains are subdivided into at least three  
106 subbasins (or “umbilici”) separated by structural highs in the Early Jurassic (Ameur,  
107 1999; Elmi et al., 2006). The slope-basin transition occurred in the eastern Traras  
108 Mountains (Benhamou, 1983) (Fig. 1B).

109 A field mission was realized in 2017 in the Mellala section (Traras Mountains) to collect  
110 brachiopod specimens from the Pliensbachian-Toarcian transition where brachiopod  
111 specimens were scarce in the Alméras collection (Table 1). This section exposes an  
112 exceptionally thick (~50 m) succession of interbedded argillaceous limestones and  
113 calcareous marls that are well-constrained biostratigraphically by assemblages of  
114 ammonites, brachiopods (Elmi et al., 2006), and calcareous nannofossils (Baghli et al.,  
115 in prep.).

116 The Rhar Roubane mountains are limited to the north by the Maghnia Plain, to the  
117 south by the Oran High Plains, to the east by the Tlemcen Mountains, and to the west  
118 by the Oujda Mountains. The Lower Jurassic facies associations are representative of  
119 protected carbonate ramp, slope and basin depositional environments (Mekahli, 1988)  
120 (Fig. 1B).

121 The Saïda Mountains are located in an intermediate region comprised between the  
122 Tell to the north, the Oran High Plains to the south, to the west by the transform fault  
123 of Ain-Sefra, and to the east by the high-plains of Sersou. Facies associations in Saïda  
124 Mountains are representative of deep-external platform environments, in contrast to  
125 the Traras Mountains and to the Rhar Roubane area, which represent shallower, more  
126 proximal environments. It was part of a protected transitional zone during most of the  
127 Early Jurassic, which opened towards the marine domain in the latest part of the middle  
128 Toarcian substage (Elmi and Almeras, 1984; Elmi et al., 1985).

129 In the southern part of the Tell, the Ouarsenis massif is bounded to the north by the  
130 Chelif valley, to the south by the Sersou plateau, to the west by the Beni-Chougrane  
131 and Mina Mountains, and to the east by the Bibans. During the Early Jurassic, this  
132 region corresponded to an internal platform evolving into an external platform,  
133 connected to the open sea and showing a deepening trend (Benhamou et al., 2000).

134 The Nador Mountains correspond to a SW-NE alignment of the reliefs located along  
135 the southern edge of the Sersou plateau. The Djebel Nador constitutes an intermediate

136 zone between the Tellian area to the north and the high-plains to the southwest,  
137 resulting from the collapse of a Sinemurian-Pliensbachian carbonate platform.  
138 Sedimentation in this area indicates a neritic environment to the west and an epi-  
139 oceanic environment to the east.

140

### 141 **3. Materials and methods**

142

#### 143 *3.1. Brachiopod shell material and preparation*

144

145 A total of 156 brachiopod shells from NW Algeria, ranging in age from the upper  
146 Sinemurian to the middle Toarcian, were analysed for their carbon and oxygen isotope  
147 composition. The isotopic data were obtained from a composite of different articulate  
148 brachiopod taxa, including Rhynchonellida (*Homoeorhynchia*, *Soaresirhynchia*; Table  
149 1) and Terebratulida (*Telothyris*, *Lobothyris*; Table 1). The sediment matrix was  
150 mechanically removed from brachiopod specimens using stainless dental tools, and  
151 the shell surface was thoroughly washed with deionized water in an ultrasonic bath  
152 (three times for 10-min with renewed water), rinsed with MQ water (obtained from a  
153 Direct-Q Millipore® purification system), and dried at 50°C for 48 h. The outermost layer  
154 of the brachiopod shells was carefully removed with a dental tool under a  
155 Stereomicroscope (Leica M125) at 80 X. Powders for stable isotope measurements  
156 were obtained by scraping the pristine long and thin fibrous textures lying in the  
157 innermost secondary layer of the anterior shell (Fig. 2).

158

#### 159 *3.2. Analyses*

160

161 The preservation of calcite fibres of all brachiopod shells was assessed under the  
162 optical microscope and scanning electron microscope (SEM) using a PHENOM G2  
163 PRO instrument in backscatter mode (Phenom-World B.V., Dillenburgstraat 9T, 5652  
164 AM, Eindhoven, The Netherlands) before stable isotope measurements.

165 The carbon and oxygen isotope measurements were performed using a Thermo Fisher  
166 Scientific (Bremen, Germany) Gas Bench II carbonate preparation device, connected  
167 to a Delta Plus XL isotope ratio mass spectrometer at the Institute of Earth Surface  
168 Dynamics of the University of Lausanne. The CO<sub>2</sub> extraction was done by reaction of  
169 100–200 µg of powdered carbonate sample with anhydrous phosphoric acid at 70°C.

170 The carbon and oxygen isotope ratios are reported in the delta ( $\delta$ ) notation as the per  
171 mil (‰) deviation relative to the Vienna Pee Dee belemnite standard (VPDB). The  
172 standardization of the  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values relative to VPDB was done by replicate  
173 analysis of the in-house working standard Carrara marble (UNIL-CM,  $\delta^{13}\text{C} = 2.05$  ‰,  
174  $\delta^{18}\text{O} = -1.70$  ‰) within the analytical sequences. The  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values of the  
175 reference  $\text{CO}_2$  gas and the UNIL-CM standard were obtained by normalization with the  
176 international reference materials NBS-19 limestone, NBS-18 carbonatite and LSVEC  
177 lithium carbonate using the values reported in Brand et al. (2014). Analytical  
178 uncertainty ( $1\sigma$ ), monitored by replicate analyses of NBS-19 and UNIL-CM was not  
179 greater than  $\pm 0.05$  ‰ for  $\delta^{13}\text{C}$  and  $\pm 0.1$  ‰ for  $\delta^{18}\text{O}$ .

180 Trace element analysis was performed at the Laboratoire de Géologie de Lyon, on the  
181 fibrous secondary layer of selected brachiopods. A total of 57 specimens were selected  
182 for element analysis, including those sample that had apparently anomalous  $\delta^{13}\text{C}$  and  
183  $\delta^{18}\text{O}$  values, and also some samples with isotopic ratios in the range of marine  
184 carbonates. For each brachiopod specimen, 1 mg of powdered sample was digested  
185 in 10 ml of  $\text{HNO}_3$  solution (5% v/v) for 24 h. An aliquot of 1 ml from each stock solution  
186 was then diluted in a solution of 9 ml of  $\text{HNO}_3$  (2% v/v) and Indium (2  $\mu\text{g/l}$ ). Each  
187 solution was analysed for Sr and Mn by an inductively plasma mass spectrometer (ICP  
188 MS), Agilen Technologies 7500 Series.

189

### 190 *3.3. Data treatment*

191

192 A numerical age was attributed to each analysed brachiopod, based on available  
193 stratigraphic information (i.e., ammonite zone and subzone). The ages of each biozone  
194 were derived from the timescale of Gradstein et al., (2012). The stable isotope data  
195 ( $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values) were smoothed to better highlight the long-term trends. Two  
196 methods for smoothing data by locally weighted regression were implemented, the  
197 LOESS smoothing function using the software PAST 3.18 (Hammer et al., 2001), and  
198 the lowess algorithm (Cleveland, 1981). Lowess is a linear polynomial smoother,  
199 whereas the LOESS procedure uses a quadratic polynomial to avoid over-fitting and  
200 excessive twisting and bending (e.g., Faber and Rajko, 2007). The smoothing factor  
201 applied to the whole data is 0.1. Such a small factor was chosen in order to highlight  
202 short-term fluctuations in the considered time interval.

203 The LOESS smoothing with a factor of 0.1 was also applied to the data available in the  
204 literature from different Tethyan regions. For each region, the smoothed isotopic data  
205 were computed, plotted against absolute age, and compared to the LOESS values  
206 obtained for NW Algeria. One-way ANOVA analyses were used to determine the  
207 significance (probability value  $P$ ) of the observed differences (offsets). The difference  
208 between two groups is considered to be statically significant at the 95% level when  $P$   
209  $\leq 0.05$ .

210

## 211 **4. Results**

212

### 213 *4.1. Optical microscopy and SEM images*

214

215 All the brachiopod specimens were screened under SEM and optical microscope  
216 before C and O isotope analyses. The textures observed vary from well-preserved,  
217 when calcite fibres are long and thin, without any clear trace of overgrowth (Fig. 2C),  
218 to mediumly-preserved material showing slightly overgrown fibres (Fig. 2B). Few  
219 brachiopod specimens have shown very poorly preserved calcite fibres, almost  
220 completely overgrown. The ultra-structure of the inner layers of the shell of all the  
221 brachiopod under SEM (Fig. 2) allowed us to discard specimens showing a variable  
222 but strong degree of post-depositional alteration that resulted in marked textural  
223 changes compared to that of modern specimens. These changes are expressed by a  
224 loss of pristine microstructures, ranging from slight dissolution of calcite fibres to almost  
225 complete obliteration of any fibrous structure (Bates and Brand, 1991). Such  
226 specimens were not processed for geochemical analyses. After this screening by  
227 optical microscopy and SEM analysis, 156 brachiopod shells were retained for  $\delta^{13}\text{C}$   
228 and  $\delta^{18}\text{O}$  measurements.

229

### 230 *4.2. Carbon and oxygen isotope data and trace element concentrations*

231

232 The  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values of well-preserved brachiopod shells are given in Appendix  
233 A (Supplementary data). The  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  scatterplot shows a large dispersion of data  
234 (Fig. 4), with  $\delta^{13}\text{C}$  values ranging from 0 to 5.0 ‰ and  $\delta^{18}\text{O}$  values from -5.0 to -1.0  
235 ‰. Few samples record anomalously negative  $\delta^{13}\text{C}$  (down to -4.8 ‰) and  $\delta^{18}\text{O}$  (down



236 to  $-7.6$  ‰) compared to the Mesozoic and modern brachiopod records, which are  
237 strongly suggestive of strong post-mortem diagenetic overprint. No clear  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$   
238 relationship was observed for 95% of the brachiopod specimens, even when  
239 considering the different localities separately (Fig. 4). The relative stratigraphic position  
240 of brachiopods in the different sections was respected and vertical bars on the isotope  
241 curves (Fig. 5) indicate the stratigraphic extension of brachiopod ages according to  
242 ammonite zones and sub-zones. The main differences in the isotopic composition of  
243 the brachiopod shells occur between localities and are likely due to the record of  
244 various time spans in the different study areas (Appendix B). Nonetheless, the  
245 relatively large variability of the composite specimen samples is certainly attributable  
246 to the presence of peculiar environmental conditions in a given area or to different  
247 water-depths that likely affect the isotope results (Fig. 5).

248 The  $\delta^{13}\text{C}$  trends reveal several events across the studied interval, which are more  
249 clearly visible in the LOESS smoothing curve (Fig. 5). The earliest event is recorded  
250 at the Sinemurian/Pliensbachian boundary, where a pronounced  $\sim 2.5$  ‰ negative  
251 excursion, with  $\delta^{13}\text{C}$  values as low as  $-4.0$  ‰ mark the *Raricostatum*-*Jamesoni*  
252 interval. According to the original study of Alm eras et al. (2007), the analysed  
253 brachiopods belong to an assemblage characteristic of the Sinemurian/Pliensbachian  
254 boundary. Importantly, the samples showing positive values were collected in the lower  
255 part of the section, the brachiopods showing negative values were collected in the  
256 upper part of the same section (see Supplementary material). In the *Ibex* zone and in  
257 the lower part of the *Davoei* zone,  $\delta^{13}\text{C}$  values are around  $1.0$  ‰. In the upper part of  
258 the *Davoei* zone, more positive values reach up to  $4.0$  ‰. The  $\delta^{13}\text{C}$  values return close  
259 to  $\sim 1.0$  ‰ near the transition between the *Lavinianum*-*Algovianum* Zones, and show a  
260  $\sim 1.0$  ‰ negative excursion in the uppermost Pliensbachian (*Emaciatum* Zone).  
261 Towards the lowermost part of *Tenuicostatum* zone, the  $\delta^{13}\text{C}$  values become more  
262 positive with a  $+2.0$  ‰ shift, and decrease upwards to  $-1.0$  ‰ in the basal part of the  
263 *Serpentinum* Zone. After this event, a large shift toward positive values (up to  $4.5$  ‰)  
264 characterizes the middle part of the *Serpentinum* zone, followed by some less  
265 pronounced fluctuations until the *Bifrons* zone with an average value of  $3.0$  ‰.

266 At the Sinemurian/Pliensbachian boundary, the  $\delta^{18}\text{O}$  values range between  $-5.2$  ‰  
267 and  $-3.0$  ‰ (on average  $-4.3 \pm 0.68$  ‰). The  $\delta^{18}\text{O}$  values increase considerably in the  
268 *Ibex* zone (lower Pliensbachian) and at the base of the *Davoei* zone, both with average  
269 values of  $-1.9$  ‰. In the upper part of the *Davoei* zone, the  $\delta^{18}\text{O}$  values decrease again

270 to an average of  $-4.0$  ‰. Up to the beginning of the Emaciatum zone, the oxygen-  
271 isotope values are rather stable with an average at  $-3.0$  ‰. Passing from the  
272 Emaciatum zone to the Pliensbachian/Toarcian boundary, the  $\delta^{18}\text{O}$  values record a  
273 negative excursion of almost  $2.0$  ‰ attaining an average of  $-4.0$  ‰ at the boundary. A  
274 positive  $2.0$  ‰ shift characterizes the base of the Toarcian with average values of  $-2.0$   
275 ‰ towards the middle part of the Tenuicostatum zone. This isotopic change is followed  
276 by a second negative excursion ending in the lower part of the Serpentinum zone, with  
277 average  $\delta^{18}\text{O}$  values of  $-3.8$  ‰. In the aftermath, small isotopic rebounds to less  
278 negative values (of  $\sim 0.8$  ‰) are recorded, after which  $\delta^{18}\text{O}$  values, although fluctuating,  
279 enter a  $\sim 2.0$  ‰ long term trend toward more negative values extending from the middle  
280 part of the Serpentinum zone to the middle part of the Bifrons zone and lasting around  
281 2 Myrs.

282 The trace element contents of the selected samples revealed Sr contents varying  
283 between 130 and 1000 ppm, with most samples exceeding 450 ppm (on average  $524 \pm 260$ )  
284 (Fig. 3). Mn contents range between 3 to about 80 ppm (on average  $64 \pm 58$ ).  
285 Half a dozen samples are depleted in Sr and enriched in Mn (Supplementary data).  
286 Ten samples show anomalously negative  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values, or anomalous  
287 contents in both Mn ( $< 80$  ppm) and Sr ( $> 450$  ppm) with respect to modern brachiopod  
288 shells secreted in equilibrium with ambient water (e.g., Brand and Veizer, 1980; Brand  
289 et al., 2003). These samples were hence excluded from the LOESS calculation, but  
290 are presented in the chemostratigraphic profiles in Figures 5 to 7.

291 The comparison of LOESS smoothing data from NW Algeria and from different  
292 Tethyan regions shows an offset of  $1$  ‰, which, according to the one-way ANOVA  
293 analyses, is statistically significant ( $P = 1\text{E-}05$ ; permutation  $n = 99999$ ).

294

## 295 **5. Discussion**

296

297 A certain degree of variability is observed both for carbon and oxygen isotopes (Fig.  
298 5). Such variability is higher than that observed by Suan et al. (2010) in brachiopods  
299 of comparable age from the Portuguese Peniche section in Portugal. However, this  
300 can be partly due to the different environmental origin of the brachiopods sampled in  
301 five different regions of the NW Algeria. The water depth differences between the  
302 studied regions in Algeria are not important, as samples belong to outer platform, slope  
303 or shallow-basin settings. Furthermore, these five regions were located almost at the

304 same palaeolatitude. Thus, apart from the data dispersion (Fig. 5) also reported from  
305 brachiopod isotope composition in other time intervals (e.g., Giraud et al., 2016), some  
306 prominent excursions are observed in the present work for  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values from  
307 NW Algeria (i.e., Sinemurian-Pliensbachian boundary, Pliensbachian-Toarcian  
308 boundary, early Toarcian Anoxic Event). These excursions can likely be related to  
309 palaeoclimatic and palaeoceanographic events, once proven the primary nature of the  
310 isotopic signal. The palaeoenvironmental significance and interpretations of these  
311 isotopic records can then be safely compared with analogous records from other  
312 Tethys regions.

313

### 314 *5.1. Preservation of brachiopods shells*

315

316 Present-day articulate brachiopods (Rhynchonellida, Terebratulida) are strictly benthic  
317 and sessile organisms that generally precipitate their secondary shell layer in isotopic  
318 equilibrium with the ambient waters (Brand et al., 2003; Brand et al., 2013), and  
319 incorporate trace elements and stable isotopes from the surrounding environments  
320 (Brand and Veizer, 1980). In the case of exposure of this carbonate phase to meteoric  
321 waters, it might undergo partial or total dissolution, exchange trace elements and  
322 stable isotopes with those of interstitial water, and finally re-precipitate as diagenetic  
323 calcite (Brand and Veizer, 1980). In modern brachiopod species, the outermost primary  
324 layer, which is characterized by a thickness of about 50  $\mu\text{m}$  and a granular texture, is  
325 most probably secreted out of isotopic equilibrium with sea water (Carpenter and  
326 Lohmann, 1995; Auclair et al., 2003; Brand et al., 2003). Specialized structures (mainly  
327 located in the posterior part of the shell) and the uppermost part of the secondary shell  
328 were excluded from analysis as, in modern brachiopod species, these parts may also  
329 have an isotope signature out of the expected values for calcite precipitated in isotopic  
330 equilibrium with environmental water (Carpenter and Lohmann, 1995; Curry and  
331 Fallick, 2002; Auclair et al., 2003). The stable isotope analyses were performed on  
332 powders obtained from the outermost layer of the brachiopod shells (see section 3.1).  
333 SEM images were used to evaluate the preservation of the microstructures of the  
334 secondary layer, the shell microstructures are considered to be well preserved when  
335 smooth fibrous surfaces of secondary layer occur, similarly to those observed in living  
336 specimens (Brand et al., 2003; Suan et al., 2008a). Diagenetically-altered structures  
337 of the secondary layer of brachiopod shells are identified by the poor individualization

338 of calcite fibres (Suan et al., 2008a; Giraud et al., 2016). In specimens showing a good  
339 preservation of the calcite fibres of the secondary layer under SEM, the primary  
340 character of the stable carbon and oxygen isotope composition of the brachiopod  
341 shells was tested using the concentrations of strontium and manganese, which are  
342 known to be sensitive to diagenetic alteration (Brand and Veizer, 1980; Brand et al.,  
343 2003). Although it has been shown that there were variations in the concentrations of  
344 these elements through the Phanerozoic in relation with episodes of aragonite and  
345 calcite seas (Steuber and Veizer, 2002), it is widely acknowledged that the Sr  
346 concentration of low-magnesium calcite in the well-preserved secondary brachiopod  
347 layer varies between 450 and 1900 ppm at middle-to-low latitudes, while Mn values  
348 vary between 0 ppm and about 80 ppm (Brand and Veizer, 1980; Korte et al., 2003),  
349 even in modern brachiopods (Brand et al., 2003).

350 The Sr and Mn contents of almost all the brachiopod shells are in the range of unaltered  
351 fossil brachiopod calcite mentioned above, with the exception of few specimens that  
352 are plotted with different symbols along non-altered samples in figure 3, but excluded  
353 from our isotopic interpretations and thus not discussed in the following. In some cases,  
354 the anomaly only concerns a single trace element (i.e., Mn or Sr), and the  $\delta^{13}\text{C}$  and  
355  $\delta^{18}\text{O}$  values are well within those expected for primary marine carbonates. Mn contents  
356 outside the range for low-Mg calcite are probably due to seasonal conditions that may  
357 influence the Mn contents of sea-water, with a decrease in winter and an increase in  
358 spring (Dromgoole and Walter, 1990). Alternatively, redox conditions may also affect  
359 the trace element concentrations (Vander Putten et al., 2000). The Sr content in calcite  
360 of modern bivalve species appears to be greatly controlled by kinetic and metabolic  
361 effects (Freitas et al., 2005; Vander Putten et al., 2000), a possible minor temperature  
362 and salinity influence may occur (Beck et al., 1992; Elderfield and Ganssen, 2000). In  
363 the brachiopods we analysed, no difference in Sr and Mn concentrations between  
364 specimens and species from the same stratigraphic level were observed. Our results  
365 show that the Mn content is similar between two analysed genera and varying Sr  
366 contents between genera, if any, could probably reflect particular metabolic effects  
367 (Table 1).

368 A  $\delta^{13}\text{C}_{\text{brachiopod}}$  *versus*  $\delta^{18}\text{O}_{\text{brachiopod}}$  scatterplot provided further information on the  
369 eventual diagenetic modification by meteoric waters (Fig. 4). A 95% ellipse data  
370 concentration shows that most of the values obtained in this paper are in the range of  
371 "normal" marine carbonates. Sedimentary diagenesis involves recrystallization

372 (dissolution and precipitation) and/or overgrowth on primary carbonate minerals,  
373 resulting in a depletion in  $^{13}\text{C}$  and  $^{18}\text{O}$  of the secondary carbonate phases (Burdett et  
374 al., 1990; Bates and Brand, 1991; McConnaughey et al., 1997; Auclair et al., 2003),  
375 and therefore generating a positive linear  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  correlation. The large dispersion  
376 of our results in the  $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$  scatterplot and the absence of clear correlation between  
377 values suggest that the isotopic composition of brachiopod shells did not suffered  
378 major diagenetic overprint. However, the  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values of samples from the  
379 different localities seem to form distinct clusters (Fig. 4). This is mainly related to the  
380 fact that the studied localities correspond to different stratigraphic intervals, recording  
381 distinct temporal changes of the  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$  values. Conversely, the differences in  
382 isotopic compositions between genera, species or specimens sampled within a given  
383 stratigraphic level are relatively small, ranging from 0 to about 1 ‰ for both  $\delta^{13}\text{C}$  and  
384  $\delta^{18}\text{O}$  values (Table 1), in line with the inter-specimen variability of modern brachiopod  
385 or bivalve specimens collected from the same localities (Carpenter and Lohmann  
386 (1995) and Vander Putten et al. (2000).

387

## 388 *5.2. Late Sinemurian - Early Pliensbachian interval*

389

390 In spite of the broad stratigraphic range of the brachiopod specimens analysed in this  
391 study that have been assigned to the Raricostatum to Jamesoni zones, the new data  
392 from NW Algeria sites show a prominent negative carbon isotope excursion (CIE) and  
393 a 1.0 ‰ negative shift in oxygen isotopes that might correspond to the  
394 Sinemurian/Pliensbachian interval. Previous studies have shown that the negative CIE  
395 at the Sinemurian/Pliensbachian interval has a supra-regional nature (Figs. 6 and 7).  
396 The CIE is recorded in Spain (Rosales et al., 2006), Portugal (Suan et al., 2010; Duarte  
397 et al., 2014), and the UK (Jenkyns et al., 2002; Korte and Hesselbo, 2011; Price et al.,  
398 2016). Some authors interpreted this negative  $\delta^{13}\text{C}$  excursion as reflecting the massive  
399 injection of isotopically light carbon into the ocean-atmosphere reservoirs (Korte and  
400 Hesselbo, 2011; Price et al., 2016). Several sources have been proposed for similar  
401 events recorded in the Phanerozoic, such as methane release from clathrates,  
402 wetlands, thermal metamorphism, kerogen maturation, or decomposition of fossil  
403 organic matter (e.g., Scholle and Arthur, 1980; Dickens et al., 1995; Kump and Arthur,  
404 1999; Higgins and Schrag, 2006). However, it is unclear which of these mechanisms  
405 is the most likely to have occurred at the Sinemurian-Pliensbachian interval. In the

406 case of NW Algeria, and in several other cases, the negative CIE at the  
407 Sinemurian/Pliensbachian boundary appear to be as a sharp  $\delta^{13}\text{C}$ -excursion. This can  
408 be explained either by the presence of a sedimentary condensation or a hiatus, or by  
409 the occurrence of a sudden event. The second hypothesis is the most plausible,  
410 considering the abruptness of the  $\delta^{13}\text{C}$  decrease in several localities, such as in the  
411 stratotype for the base of the Pliensbachian (Robin Hood's Bay, NE England), which  
412 is known to be only slightly or not affected by condensation or hiatuses (Hesselbo et  
413 al., 2000b).

414 Oxygen isotope data can help understanding the perturbations of the carbon cycle  
415 across the Sinemurian/Pliensbachian transition. The variations in the  $\delta^{18}\text{O}$  values of  
416 the brachiopod shell reflect mainly changes in ambient temperature and seawater  
417 oxygen isotope composition, which indirectly reflect changes in salinity (Brand et al.,  
418 2013). The oxygen isotope fractionation between calcite and water is temperature  
419 dependent, and warmer waters lead to  $^{18}\text{O}$  depleted calcite (Lowenstam, 1961; O'Neil  
420 et al, 1969). Although a certain dispersion is observed in Upper Sinemurian-Lower  
421 Pliensbachian  $\delta^{18}\text{O}$  values, a trend to more negative  $\delta^{18}\text{O}$  values in the brachiopod  
422 shells is observed. This trend has revealed to be statistically significant after an  
423 ANOVA test and most probably indicates a sudden temperature increase. This  
424 temperature increase would support the hypothesis of a massive injection of  
425 isotopically light carbon in the ocean-atmosphere system, because enhanced  
426 emissions of  $\text{CO}_2$  and  $\text{CH}_4$  might have produced an increased greenhouse effect.  
427 Furthermore, the oxygen isotope data of the studied areas do not show more positive  
428  $\delta^{18}\text{O}$  values than the northern localities, as it would be expected in case of higher  
429 seawater salinity (Epstein and Mayeda, 1953). High evaporation rates should have  
430 occurred in NW Algeria, which occupied the southern Tethys arid margin during the  
431 Early Jurassic (Turner and Sherif, 2007). A positive  $\delta^{18}\text{O}$  shift is conversely reported  
432 across the Sinemurian–Plienbachian boundary (Korte and Hesselbo, 2011). However,  
433 this shift is likely produced by a bottom water cooling related to a deepening phase in  
434 the Cleveland Basin (Korte and Hesselbo, 2011). This cooling might be the expression  
435 of local conditions rather than a climatic deterioration.

436 In the Davoei zone, another isotope event is recognized despite the relatively low data  
437 resolution in this interval. The  $\delta^{13}\text{C}$  values in the NW Algeria sections show a positive  
438 excursion similar to that observed in other regions (Silva et al., 2011; Franceschi et al.,  
439 2014; Gómez et al., 2016) (Fig. 6), suggesting that this event was a supra-regional

440 phenomenon; it is also marked by a drop in the ammonite species richness in the  
441 Mediterranean and NW European provinces (Dommergues et al., 2009). At the Davoei  
442 zone, also a prominent negative  $\delta^{18}\text{O}$ -excursion indicates warmer conditions (Fig. 5).  
443 This event was a relatively hot, humid phase, as shown in NW European sediments by  
444 kaolinite enrichment and a high  $^{87}\text{Sr}/^{86}\text{Sr}$  values caused by enhanced continental  
445 weathering (Jones and Jenkyns, 2001; Dera et al., 2009).

446

### 447 *5.3. Late Pliensbachian – Early Toarcian interval*

448

449 In the Late Pliensbachian, the  $\delta^{13}\text{C}$  values settle between 1 and 2‰ in the Lavinianum-  
450 Algovianum zones (time equivalent of the Margaritatus zone of the Boreal Realm). This  
451 trend has been reported in coeval strata from several localities in Europe (Jenkyns and  
452 Clayton, 1986; Bailey et al., 2003; Rosales et al., 2006; Suan et al., 2010; Korte and  
453 Hesselbo, 2011), and is accompanied in some sections, as those in Portugal and Spain  
454 by the occurrence of black shales in the Margaritatus Ammonite Zone (Rosales et al.,  
455 2006; Suan et al., 2010; Silva et al., 2015). A trend towards higher  $\delta^{18}\text{O}$  values in the  
456 Late Pliensbachian (Fig. 5) most probably indicates a trend to cooler seawater  
457 culminating in the Emaciatum (or Spinatum) zone, as it was suggested for similar  $\delta^{18}\text{O}$ -  
458 trends in coeval belemnites and brachiopods from France, Germany, England, Spain  
459 or Portugal (e.g., McArthur et al., 2000; Rosales et al., 2004a; van de Schootbrugge et  
460 al., 2005; Oliveira et al., 2006; Harazim et al., 2013). Several studies have discussed  
461 the possible causes of this cooling event. Guex et al. (2001) interpreted this cooling as  
462 the result of a decreased insolation following the repeated volcanogenic  $\text{SO}_2$  emissions  
463 in the Karoo-Ferrar Large Igneous Province. Similarly, Jourdan et al. (2008) inferred a  
464 temporal correspondence between the earliest pulses of the Karoo-Ferrar LIP and the  
465 beginning of the climate cooling in the upper Pliensbachian. However, this  
466 interpretation remains speculative given the short-term residence time in the  
467 atmosphere of sulphur aerosols (~10 years) (Wignall, 2001), while the cooling event  
468 lasted longer than 1 Myrs according to recent astronomical calibration of Upper  
469 Pliensbachian strata from the Mochras borehole in Wales (Ruhl et al., 2016). This  
470 cooling episode could also be ascribed to a global drawdown of atmospheric  $\text{CO}_2$   
471 (inverse greenhouse effect) caused by a widespread burial of isotopically light organic  
472 carbon ensuing the accumulation of organic matter-rich sediments in the Margaritatus  
473 Zone (Jenkyns and Clayton, 1997; Silva et al., 2015).

474 Across the Pliensbachian/Toarcian boundary, the negative  $\delta^{13}\text{C}$  excursion in Algeria  
475 correlates remarkably well with that recorded in coeval sites in NW Europe, which was  
476 interpreted as reflecting another pulse of massive injection of  $^{13}\text{C}$ -depleted carbon into  
477 the oceans and the atmosphere (Hesselbo et al., 2007; Suan et al., 2008a). This  
478 hypothesis is consistent with the marked  $\delta^{18}\text{O}$  negative excursion in biogenic  
479 carbonates at the same level in Algeria (Fig. 5), Spain and Portugal (Fig. 7), which has  
480 been interpreted in the latter localities as the evidence of a sharp increase in  
481 temperature (Suan et al., 2008b) more or less directly linked to a pulse of the activity  
482 in the Karoo-Ferrar LIP (Jourdan et al., 2008; Littler et al., 2009; Pálffy and Smith, 2000;  
483 Percival et al., 2015, 2016). At the end of this Pliensbachian/Toarcian event, a relatively  
484 small positive rebound of  $\delta^{13}\text{C}$  values is noticed in almost all the sites in the lower part  
485 of the Tenuicostatum Zone (Fig. 6), which correlates with a trend toward higher  $\delta^{18}\text{O}$   
486 values (Fig. 7) and can be interpreted as a temporary cooling event.

487

#### 488 *5.4. Early-Middle Toarcian environmental evolution*

489

490 A marked shift towards lower  $\delta^{13}\text{C}$  values is recorded in NW Algeria in the basal  
491 Serpentinum zone, followed by a gradual return to higher values. We interpret this shift  
492 as the expression of the well-documented negative CIE often used to define the base  
493 of the T-OAE interval. Again, this CIE is classically interpreted as being derived from  
494 the massive and rapid injection of  $^{13}\text{C}$ -depleted carbon to the ocean-atmosphere  
495 reservoirs, possibly as the result of the release of methane from continental margins  
496 (Hesselbo et al., 2000a; 2007; Kemp et al., 2005), of enhanced  $\text{CO}_2$  emissions during  
497 a renewed peak activity of the Karoo-Ferrar LIP (Suan et al., 2008a; Burgess et al.,  
498 2015), of thermogenic methane related to the LIP emplacement, or due to the  
499 combined effects of all these processes (Beerling and Brentnall, 2007). These  
500 hypotheses are consistent with the coeval negative  $\delta^{18}\text{O}$  values recorded in NW  
501 Algeria (Fig. 5) which, although less intense than observed in other areas (Fig. 7),  
502 suggests warm conditions during the T-OAE (e.g., Suan et al., 2010; Korte and  
503 Hesselbo, 2011; Svensen et al., 2007).

504 In the aftermath of the T-OAE, a marked 2.0 ‰ positive  $\delta^{13}\text{C}$  excursion at the  
505 Serpentinum zone is recorded in NW Algeria as well as in many other localities (e.g.,  
506 Jenkyns et al., 2002). This positive  $\delta^{13}\text{C}$  excursion is considered to reflect the global  
507 burial of excess organic carbon, which in turn likely caused a climate cooling through



508 CO<sub>2</sub> drawdown, as an inverse greenhouse effect (e.g., Jenkyns et al., 2002; Hermoso  
509 and Pellenard, 2014; Xu et al., 2017). This scenario is supported by the relatively large  
510 positive shift toward higher  $\delta^{18}\text{O}$  values recorded in the same interval and most likely  
511 reflecting a period of cooling (Fig. 7).

512

## 513 *5.5. Differences between C- and O-isotope records from NW Algeria and western* 514 *Tethys*

515

516 The new brachiopod C- and O-isotope profiles show some interesting similarities but  
517 also differences with previously published data from different NW Tethys regions  
518 based on measurements of belemnite or brachiopod calcite. The possible causes of  
519 these differences are discussed in the following sections.

520

### 521 *5.5.1. Trends in $\delta^{13}\text{C}$ values*

522

523 The  $\delta^{13}\text{C}$  curve of NW Algeria sections shows the main excursions recorded in other  
524 European settings, as well as a similar steady trend during the mid-Pliensbachian  
525 stage. In particular, the Sinemurian/Pliensbachian negative excursion, the Davoei  
526 positive excursion, the Pliensbachian/Toarcian negative excursion, the basal  
527 Toarcian positive shift and the Early Toarcian negative excursion are all recorded in  
528 NW Algeria. Importantly, the comparison of the different sections shows that the  
529 belemnite  $\delta^{13}\text{C}$  data from NW Europe are systematically offset towards more  
530 negative values than those of North Africa, with an overall offset of  $\sim 0.5\text{‰}$  (Fig. 6).  
531 This offset of the LOESS-smoothed  $\delta^{13}\text{C}$  values is statistically significant, and may  
532 suggest that the C-isotope composition of the dissolved inorganic carbon (DIC) in  
533 shallow shelf waters of the Western Tethys were spatially heterogeneous during the  
534 Early Jurassic.

535 This offset could be explained by differences in ecology, physiology and modes of life  
536 of brachiopods and belemnites. In fact,  $\delta^{13}\text{C}$  values of DIC decrease with increasing  
537 depth due to the remineralisation of  $^{13}\text{C}$ -depleted organic matter. The lower  $\delta^{13}\text{C}$   
538 values recorded in belemnites suggest that they secreted their calcite rostra in deep  
539 waters, more enriched in free  $^{13}\text{C}$ -depleted DIC (Wierzbowski and Joachimski, 2007;  
540 Dutton et al., 2007). Such differences could also explain why belemnite  $\delta^{13}\text{C}$  values  
541 from Peniche in Portugal (Jenkyns et al., 2002; Hesselbo et al., 2007) are more

542 negative than those obtained for brachiopods (Suan et al., 2010) from the same site  
543 (Fig. 6). The higher amplitude of the  $\delta^{13}\text{C}$ -excursions measured in NW Europe can be  
544 tentatively interpreted as the result of a greater supply of freshwater by rivers to the  
545 epi-continental N-Tethyan basins. In fact, the oceanic DIC is fed by freshwaters  
546 enriched in oxidized organic matter thus have a  $^{13}\text{C}$ -depleted DIC. This leads to the  
547 precipitation of biogenic carbonates with more negative  $\delta^{13}\text{C}$  values (Clayton and  
548 Degens, 1959). It has been also suggested that higher productivity at higher latitudes  
549 led to a  $^{13}\text{C}$  enrichment of the DIC in surface waters (e.g., the Cleveland Basin,  
550 Harazim et al., 2013). Episodes of higher productivity in western European sites or  
551 transient migrations of belemnites toward higher latitudes may have intensified  
552 changes in belemnite  $\delta^{13}\text{C}$  records.

553 The tectono-climatic regime of North Africa might also have influenced the C-isotope  
554 signals. Following the Triassic rift phase related to the opening of the North Atlantic  
555 and of the Neo-Tethys (Favre and Stampfli, 1992), restricted continental basins would  
556 have developed under arid climates in the Lower Jurassic northern part of Gondwana  
557 (Bracène and Frizon de Lamotte, 2002). High evaporation in such palaeo-lakes would  
558 have produced substantial  $^{13}\text{C}$  and  $^{18}\text{O}$  enrichment as the production of light isotopes  
559 is favoured in water vapour. The resulting covariance between  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  is hence  
560 generally considered as an evidence of hydrological closure in palaeo-lakes deposits.  
561 Accordingly, in the case of continental runoff of such palaeo-lake basins, the  $\delta^{13}\text{C}$   
562 signal in adjacent marine basins should be significantly shifted towards more positive  
563 values (Talbot, 1990; Deocampo and Jones, 2014). However, the measured  $\delta^{13}\text{C}$   
564 values in NW Algeria sections are in the range of “normal” marine carbonates, and do  
565 not show such very positive values. It seems therefore unlikely that the NW Algeria  
566 basins have been under the influence of continental runoff from surrounding arid lakes.

567

#### 568 *5.5.2. Trends in $\delta^{18}\text{O}$ values*

569

570 Although the carbon isotope records in NW Algeria show the main events reported in  
571 other Tethys settings, the oxygen isotope data show an important difference.  
572 Systematically, the  $\delta^{18}\text{O}$  values are 1.5 to 2.0 ‰ lower compared to those from NW  
573 Tethys sections (e.g., N Spain, France, England), where isotope data mainly derive  
574 from belemnite calcite. The offset of the LOESS-smoothed  $\delta^{18}\text{O}$  values is statistically  
575 significant. The relatively low  $\delta^{18}\text{O}$  values in NW Algeria samples are, however, in the

576 same range as those obtained by Krencker et al. (2014) on brachiopods from Toarcian  
577 shelf sediments in Morocco. This offset was partly expected, since these two regions  
578 were both located at lower palaeo-latitudes, and therefore were likely to have  
579 experienced warmer seawater temperatures (i.e., warm subtropical region; Aberhan,  
580 2001) than European sites. Interpreting the  $\delta^{18}\text{O}$  records solely in terms of temperature  
581 and considering available  $\delta^{18}\text{O}$ -temperature equations (Epstein et al., 1953; Anderson  
582 and Arthur, 1983), this 1.5 to 2.0 ‰ offset would suggest that seawater temperatures  
583 were about 6 to 8°C warmer in Northern Gondwana than in NW Europe, which appears  
584 unrealistic for the relatively low palaeo-latitudinal range implied (~5 to 20°; Fig. 1). In  
585 addition, the high belemnite  $\delta^{18}\text{O}$  values recently reported for the  
586 Pliensbachian/Toarcian boundary of Issouka in the Central Atlas of Morocco (0 to -1.0  
587 ‰; Ait-Itto et al., 2017) are almost identical to those obtained for most European  
588 belemnites of the same interval (Fig. 7). This similarity suggests that there was no  
589 difference in seawater temperature between these various areas, an interpretation that  
590 is at odds with the new data set.

591 As discussed above for  $\delta^{13}\text{C}$  records (section 4.5.1), however, the difference in  
592 material used for geochemical investigations, namely brachiopods in this study *versus*  
593 belemnites in most other published works, may introduce a substantial ecological and  
594 physiological bias in the comparison of the  $\delta^{18}\text{O}$  records (Ullmann et al., 2016). Indeed,  
595 several studies have suggested that belemnites precipitate their calcite in deeper  
596 waters than brachiopods. This hypothesis is supported by the  $\delta^{18}\text{O}$  record of  
597 belemnites from Portugal (Jenkyns et al., 2002; Hesselbo et al., 2007), which are ~0.5  
598 to 1.0 ‰ higher than stratigraphically equivalent brachiopods (Suan et al., 2010)  
599 sampled from the same site (Fig. 7). It is thus perhaps safer to compare  $\delta^{18}\text{O}$  records  
600 obtained on the biogenic calcite of organisms with similar ecology and physiology to  
601 avoid such palaeoecological complications. The new  $\delta^{18}\text{O}$  brachiopod data from  
602 Algeria are still 0.5 to 1.0 ‰ lower than brachiopod  $\delta^{18}\text{O}$  data from Portugal, suggesting  
603 a more reasonable difference in temperature of about 2 to 4°C between bottom water  
604 masses of these two areas.

605 It is very likely that the recorded spatial changes in  $\delta^{18}\text{O}$  values of biogenic carbonates  
606 also reflect spatial changes in seawater  $\delta^{18}\text{O}$  values, as a result of different  
607 hydrological regime between the N and S Tethys margins and thus indirectly related  
608 changes in seawater salinity. The northern margin of Gondwana was characterized  
609 from the Middle-Late Triassic to the Early Jurassic by the development of a number of

610 basins along the northern margin of the Saharan craton. As the opening of Tethys  
611 ocean proceeded and the relative sea level increased, these basins were flooded and  
612 turned to be evaporitic (Turner et al., 2001). Lithostratigraphic studies of continental  
613 basins in Eastern Algeria (e.g., Berkine Basin; Turner and Sherif, 2007) show a  
614 succession of lower Sinemurian-Pliensbachian halite deposits, and lower  
615 Pliensbachian-upper Toarcian sulphate/halite deposits. In Morocco, similar evaporitic  
616 basins are reported (Van Houten, 1977; Peretsman, 1988). This attests for the extent  
617 of evaporitic environments along the North Africa margin in the Lower Jurassic, with a  
618 vast assemblage of evaporitic basins (Le Roy and Piqué, 2001; Courel et al., 2003;  
619 Turner and Sherif, 2007), where Sabkha-type environments developed (Elmi and  
620 Benschilil, 1987). Given the proximity of these evaporitic basins to the study area, any  
621 continental water input to the marine environment would have drained these evaporitic  
622 basins. According to studies on current Sabkha environments, salinity may vary,  
623 according to the seasons, between 50 ‰ and 250 ‰ (Amarouyache et al., 2018).  
624 Since  $\delta^{18}\text{O}$  would change by about 0.2 ‰ per unit change in salinity (Sharp, 2017), a  
625 5.0 ‰ fluctuation in salinity would be accompanied by a 1.0 ‰ change in the  $\delta^{18}\text{O}$   
626 value in the ocean shallow-waters (Sharp, 2017). In the case of runoff of these waters  
627 into the ocean, a large  $\delta^{18}\text{O}$  positive offset would be expected. As we observe,  
628 conversely, a negative  $\delta^{18}\text{O}$  offset with respect to other Tethys settings, we can  
629 exclude the effects related to Sabkha environment runoff. Besides that, brachiopods  
630 are known to be intolerant to important salinity changes, which would result from a  
631 strong evaporation or from saline lake drainage (Ager, 1967). Higher evaporation at  
632 southern latitudes is thus very unlikely to be the main cause of the observed offset of  
633  $\delta^{18}\text{O}$  values of NW Algeria and the  $\delta^{18}\text{O}$  data from NW Europe. We cannot rule out,  
634 however, that the isotopic offset between the N and S margin of Tethys is somehow  
635 mitigated by such differences in evaporation. In the meantime, more intense  
636 precipitations, the combined impact of melting ice and the input of freshwaters by river  
637 systems should have produced lower  $\delta^{18}\text{O}$  values in the northernmost basins (Rosales  
638 et al., 2004b; Dera and Donnadiou, 2012), which may explain the anomalously low  
639 belemnite  $\delta^{18}\text{O}$  values in the UK compared to all other belemnites records (Fig. 7).  
640 The Pliensbachian-Toarcian record from England (Cleveland Basin) is also singular in  
641 that it shows an opposite trend for oxygen isotopes compared to NW Algeria, as well  
642 as with respect to other localities (Fig. 7). Indeed, the basal *Tenuicostatum* Zone is  
643 marked in England by a distinct positive shift toward higher values, while a wide

644 negative shift is recorded in NW Algeria (this work) and in Portugal (Suan et al., 2010).  
645 Some authors explain this trend by a deepening of sea-level near the  
646 Pliensbachian/Toarcian, which is indeed marked by a supra-regional marine  
647 transgression (Hardenbol et al., 1998; Pittet et al., 2014). Due to this sea-level rise,  
648 belemnites would have migrated in deeper and cooler waters (Korte and Hesselbo,  
649 2011). However, one may ask why the same effect was not recorded by belemnites  
650 during the T-OAE interval, which recorded an even more important sea-level rise, while  
651 belemnites display a negative excursion that can be interpreted as an evidence for  
652 warming. Korte and Hesselbo (2011) and Ullmann et al., (2014) attributed this variable  
653 response of belemnite  $\delta^{18}\text{O}$  record relative to sea-level changes to bottom water  
654 anoxia, which forced belemnites to secrete their calcite rostrum in shallower waters  
655 during the T-OAE. Similarly, substantial differences in spatial trends in  $\delta^{18}\text{O}$  recorded  
656 by belemnite rostra and bivalve shells have been documented for several other  
657 intervals, such as during the lower Bajocian, the middle Oxfordian, and Lower  
658 Tithonian cooling (Dera et al., 2011), and attributed to metabolic and ecological factors.  
659 In this regard, our new record emphasizes differences in life habitats, namely the depth  
660 in the water column belemnites thrived, as a crucial parameter to consider.  
661 Brachiopods conversely were sessile organisms, generally inhabiting shallow-water  
662 depth, making them more susceptible to record temperature fluctuations and changes  
663 in evaporation-precipitation budgets than belemnites.  
664 In summary, all these observations indicate that the  $\delta^{18}\text{O}$  trends in brachiopod shell in  
665 NW Algeria reflect mainly temperature fluctuations, and suggest that a latitudinal  
666 temperature gradient of around 2 to 4°C existed between bottom water masses of the  
667 central and southern part of the western Tethys during the Early Jurassic.

668

## 669 **6. Summary and conclusions**

670

671 We produced a continuous record of carbon and oxygen stable isotope for Upper  
672 Sinemurian-Middle Toarcian successions in NW Algeria based on brachiopod calcite.  
673 This fills an important gap in the palaeolatitudinal record of carbon and oxygen stable  
674 isotopes, which is essential for our understanding of Jurassic climatic belts. The  $\delta^{13}\text{C}$   
675 and  $\delta^{18}\text{O}$  excursions documented in NW Algeria confirm the main climate and  
676 environmental changes previously documented in the NW Tethys, although slight  
677 differences can be seen because of the peculiar paleogeographic context of the

678 studied area. The negative  $\delta^{13}\text{C}$  excursion near the Sinemurian/Pliensbachian  
679 boundary in NW Algeria, previously documented in England and Portugal, is  
680 accompanied by a decrease of  $\delta^{18}\text{O}$  values, supporting the hypothesis of a  
681 greenhouse-forced warming during this interval. The Pliensbachian/Toarcian boundary  
682 negative CIE is well marked on the S-Tethys margin and the coeval shift toward lower  
683  $\delta^{18}\text{O}$  values is also interpreted as reflecting a transient warming. Also, the well-known  
684 T-OAE CIE is documented in NW Algeria along with an intense warming phase, as  
685 shown by a coeval decrease of  $\delta^{18}\text{O}$  values.

686 The new NW Algerian records are also valuable in that they reveal a systematic  
687 positive offset of  $\sim 0.5\text{‰}$  for  $\delta^{13}\text{C}$  values and negative offset of  $\sim 1.5\text{‰}$  for  $\delta^{18}\text{O}$  values  
688 compared to European sections. We interpret these offsets as mainly reflecting higher  
689 seawater temperatures along the southern margin of Tethys. This was previously  
690 interpreted on the basis of the sedimentary record of widespread carbonate platforms  
691 and of evaporitic environments in coastal areas of N-Africa settings. We bring for the  
692 first-time quantitative data on the existing temperature gradient. This gradient can  
693 further explain the distribution of marine (Bucefalo Palliani and Riding, 2003; Mattioli  
694 et al., 2008) or continental organisms (Philippe and Thevenard, 1996), which show  
695 significant differences in terms of assemblage composition between the S and the N  
696 of the Tethys. These new data should be integrated in future climate models to  
697 constrain Early Jurassic climate belt distribution and their latitudinal fluctuations during  
698 times of environmental perturbations.

699

## 700 **Acknowledgments**

701

702 We acknowledge Stéphane Bodin and an anonymous reviewer who provided valuable  
703 comments to an early version of the manuscript. We wish to warmly thank Mohammed  
704 Adaci, Abbes Sebane, Sid-Ahmed Hammouda, Rachid Sidhoum, Abdelkader Mennad,  
705 Ayrton Nadin and Astrid Jonet for their help in fieldwork. Emmanuel Robert is  
706 acknowledged for his precious advice and help for fossil selection in the Collections de  
707 Géologie de Lyon. Financial support was provided by the Institut Universitaire de  
708 France. H.B. was funded by the Eiffel Excellence Scholarship by the French Ministry  
709 of Europe and Foreign Affairs, and by the Programme National Exceptionnel "PNE"  
710 (Algerian MESRS). This is a contribution of the IGCP 655 (IUGS-UNESCO).

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## 1158 **Legend of figures**

1159

1160 **Fig. 1.** A. Location of Western Algeria in the frame of an Early Jurassic  
1161 palaeogeographic map (modified after Thierry et al., 2000). The main current  
1162 circulation pattern is reported, as well as the main climatic modes (i.e., arid and wet)  
1163 for the Northern and Southern Tethys margin, respectively (after Dera and Donnadieu,  
1164 2012). B. Palaeogeographical map of western Algeria during the Toarcian (modified  
1165 after Elmi and Benshili, 1987).

1166

1167 **Fig. 2.** Scanning electron micrographs showing the ultra-structures of brachiopod  
1168 shells (A and C) *Cuersithyris provincialis*; (B) *Lobothyris hispanica*; (D)  
1169 *Soaresirhynchia tamazirta*.

1170

1171 **Fig. 3.** The relationships between  $\delta^{18}\text{O}$  values and Mn, Sr concentrations in brachiopod  
1172 shells. The arrows indicate the shift produced by diagenetic incorporation in Mn and  
1173 Sr in brachiopod calcite.

1174

1175 **Fig. 4.** Cross Plot of  $\delta^{13}\text{C}$  versus  $\delta^{18}\text{O}$  values of brachiopod shell calcite. Coloured  
1176 dots represent the different studied localities. The dashed circle corresponds to the  
1177 95% concentration of data points.

1178

1179 **Fig. 5.** C and O isotope compositions of brachiopod calcite from the Upper Sinemurian  
1180 to the Middle Toarcian in north-western Algeria. Numerical ages are derived from “The  
1181 Geologic Time Scale 2012” (Gradstein et al., 2012). The relative stratigraphic position  
1182 of brachiopods in the different sections was respected and vertical bars on the isotope  
1183 curves indicate the stratigraphic extension of brachiopod ages according to ammonite  
1184 zones and sub-zones. The colours of the symbols correspond to the region where the  
1185 samples come from. A LOESS Smoothing with a 0.1 factor was applied using Past

1186 3.18. Square symbols show samples recording anomalous Sr or Mn contents and  
1187 therefore not considered for the LOESS smoothing.

1188 **Fig. 6.** C-isotope comparisons and chemo-stratigraphy correlation between sections  
1189 studied in Algeria: Traras, Rhar Roubane, Ouarsenis, Saïda, and Nador Mountains  
1190 (this work); in England: Cleveland Basin (Hesselbo et al., 2000b; Mearthur et al., 2000;  
1191 Jenkyns et al., 2002; Bailey et al., 2003; Korte and Hesselbo, 2011; Li et al., 2012); in  
1192 France: Grands Causses Basin, Sud-Est Basin and Paris Basin (Dera et al., 2011;  
1193 Harazim et al., 2013); in Spain: Iberian Range, Asturias, and Basque-Cantabrian Basin  
1194 (Rosales et al., 2004a; Schootbrugge et al., 2005; Gómez et al., 2008; Armendáriz et  
1195 al., 2012); in Portugal: Lusitanian Basin (Jenkyns et al., 2002; Suan et al., 2010). All of  
1196 the data have been measured on belemnite calcite, except in Portugal (Suan et al.,  
1197 2010) and in NW Algeria (this work) where C isotopes have been measured on  
1198 brachiopod calcite. Light red and blue shaded areas respectively indicate warming and  
1199 cooling events.

1200

1201 **Fig. 7.** O-isotope comparisons and chemo-stratigraphy correlation between sections  
1202 studied in Algeria: Traras, Rhar Roubane, Ouarsenis, Saïda, and Nador Mountains  
1203 (this work); in England: Cleveland Basin (Hesselbo et al., 2000b; Mearthur et al., 2000;  
1204 Jenkyns et al., 2002; Bailey et al., 2003; Korte and Hesselbo, 2011; Li et al., 2012); in  
1205 France: Grands Causses Basin, Sud-Est Basin and Paris Basin (Dera et al., 2011;  
1206 Harazim et al., 2013); in Spain: Iberian Range, Asturias, and Basque-Cantabrian Basin  
1207 (Rosales et al., 2004a; Schootbrugge et al., 2005; Gómez et al., 2008; Armendáriz et  
1208 al., 2012); in Portugal: Lusitanian Basin (Jenkyns et al., 2002; Suan et al., 2010). All of  
1209 the data have been measured on belemnite calcite, except in Portugal (Suan et al.,  
1210 2010) and in NW Algeria (this work) where O isotopes have been measured on  
1211 brachiopod calcite. Light red and blue shaded areas indicate the warming or,  
1212 conversely, cooling events.

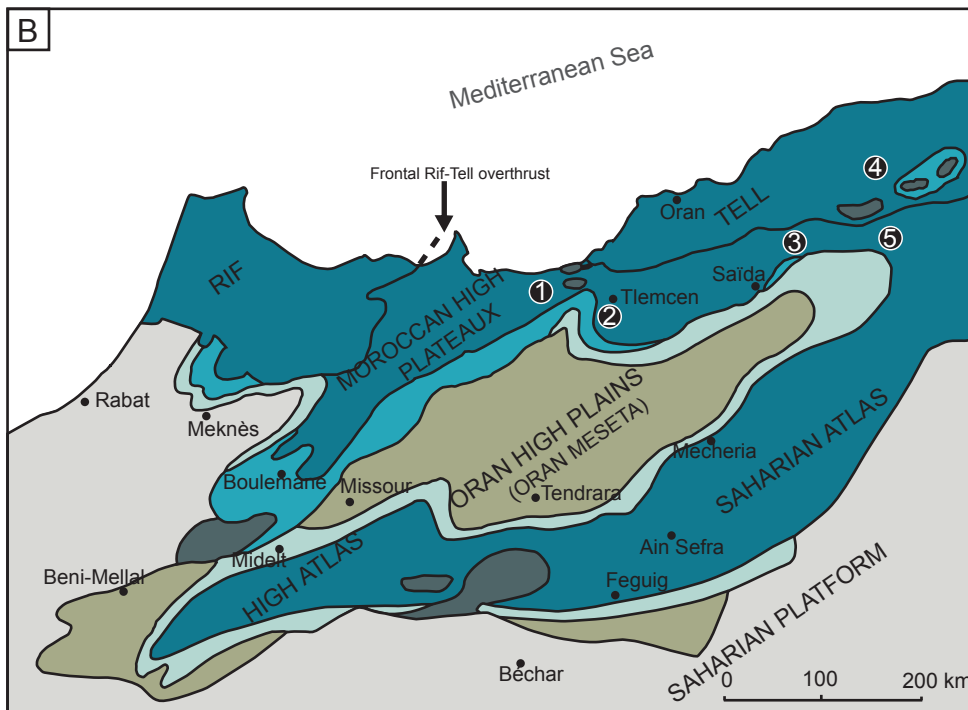
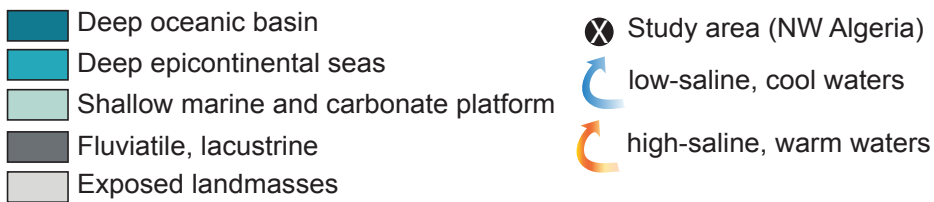
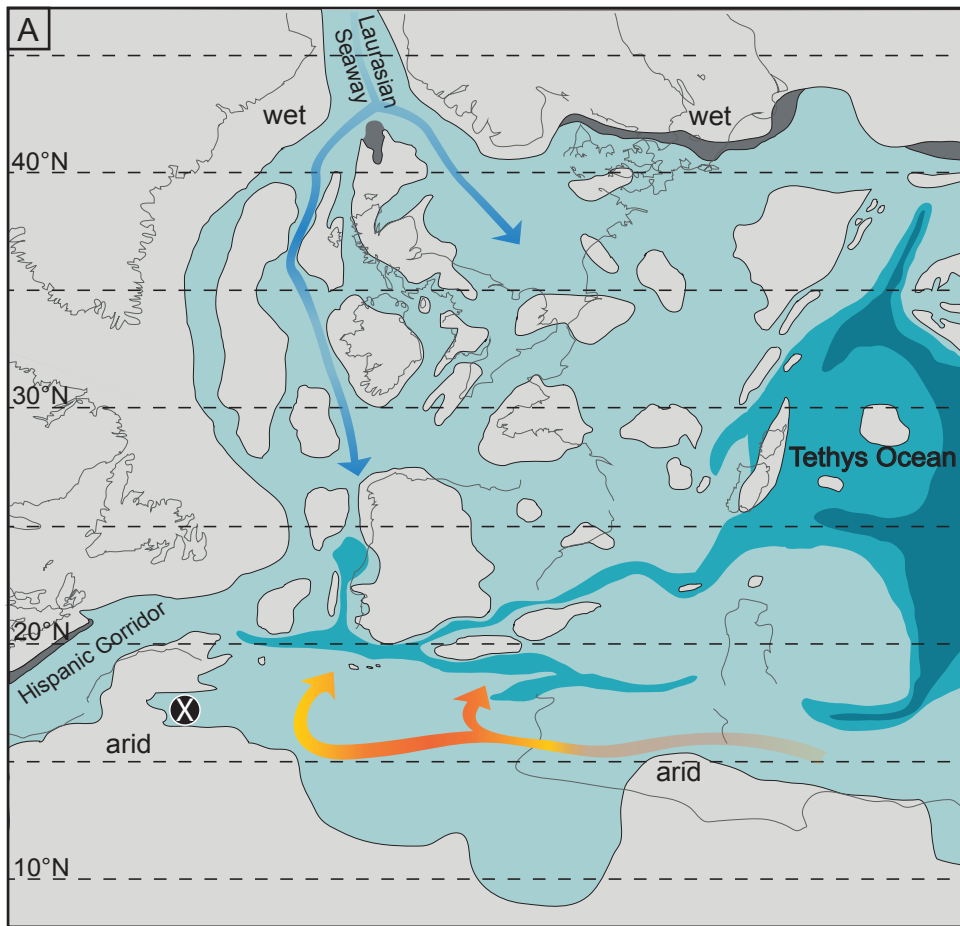
1213

1214 **Supplementary data (Appendix A)** List of all the brachiopod specimens analysed for  
1215 carbon and oxygen isotopic composition and strontium and manganese contents. Most  
1216 of these brachiopods come from Alméras collection (Alméras et al., 2007), which is  
1217 curated at the Collections de Géologie de Lyon. FSL numbers are provided along with  
1218 the section and the locality where they have been collected. Brachiopods from Melalla  
1219 section were sampled by the authors. A numerical age has been attributed to each

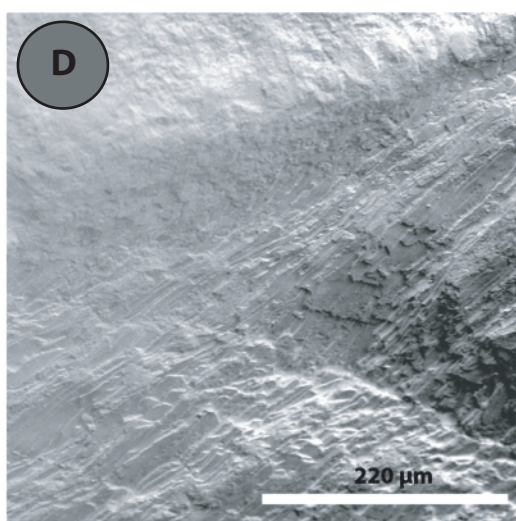
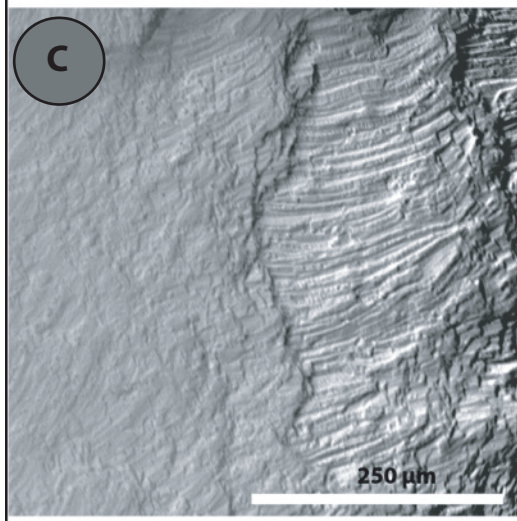
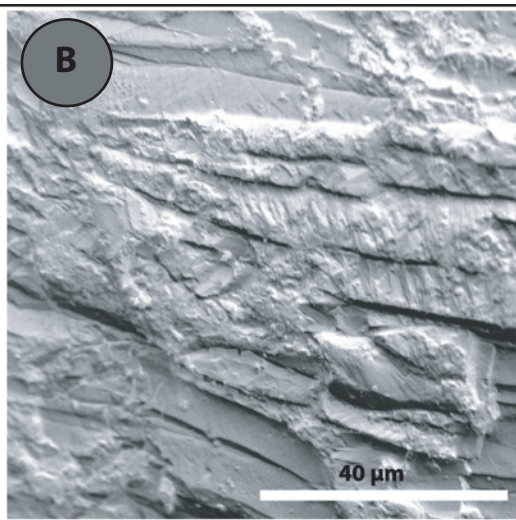
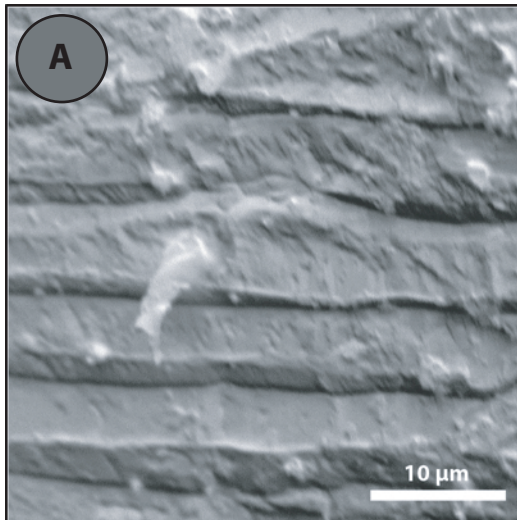
1220 specimen according to their stratigraphic position and corresponding ammonite  
1221 biozonation calibrated against the timescale of Gradstein et al. (2012).

1222

1223 **Supplementary data** (Appendix B) Synthetic stratigraphic logs with the position of the  
1224 brachiopod samples analysed for carbon and oxygen isotope composition. These logs  
1225 have been redrawn after Benhamou (1983, 1996); Elmi et al. (1985); Kharroubi (1987);  
1226 Mekahli (1988).

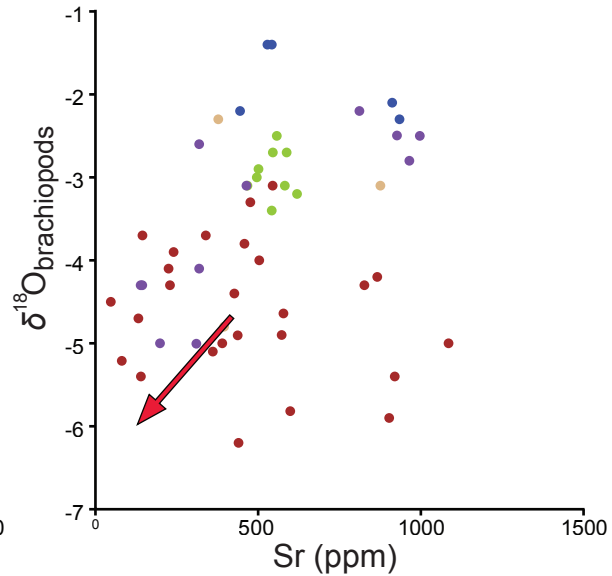
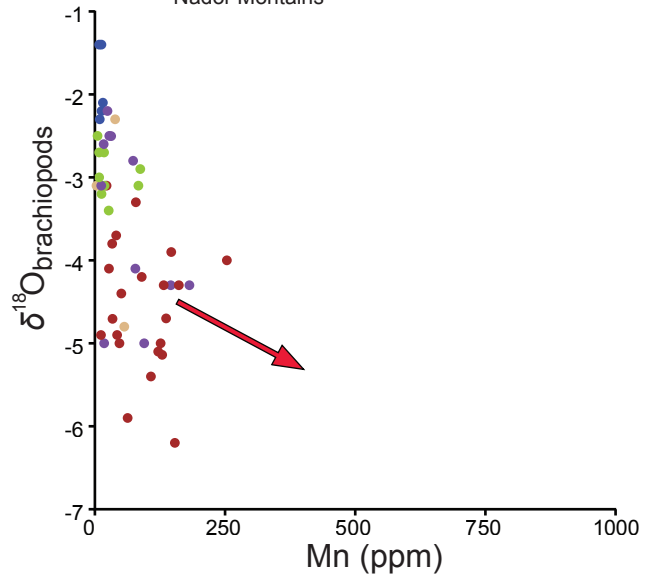


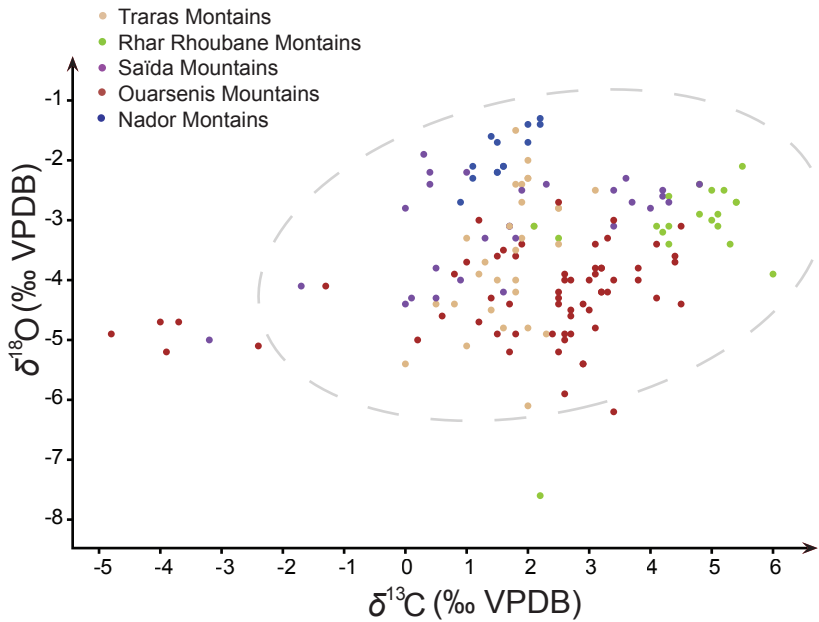
Baghli et al. Figure 1

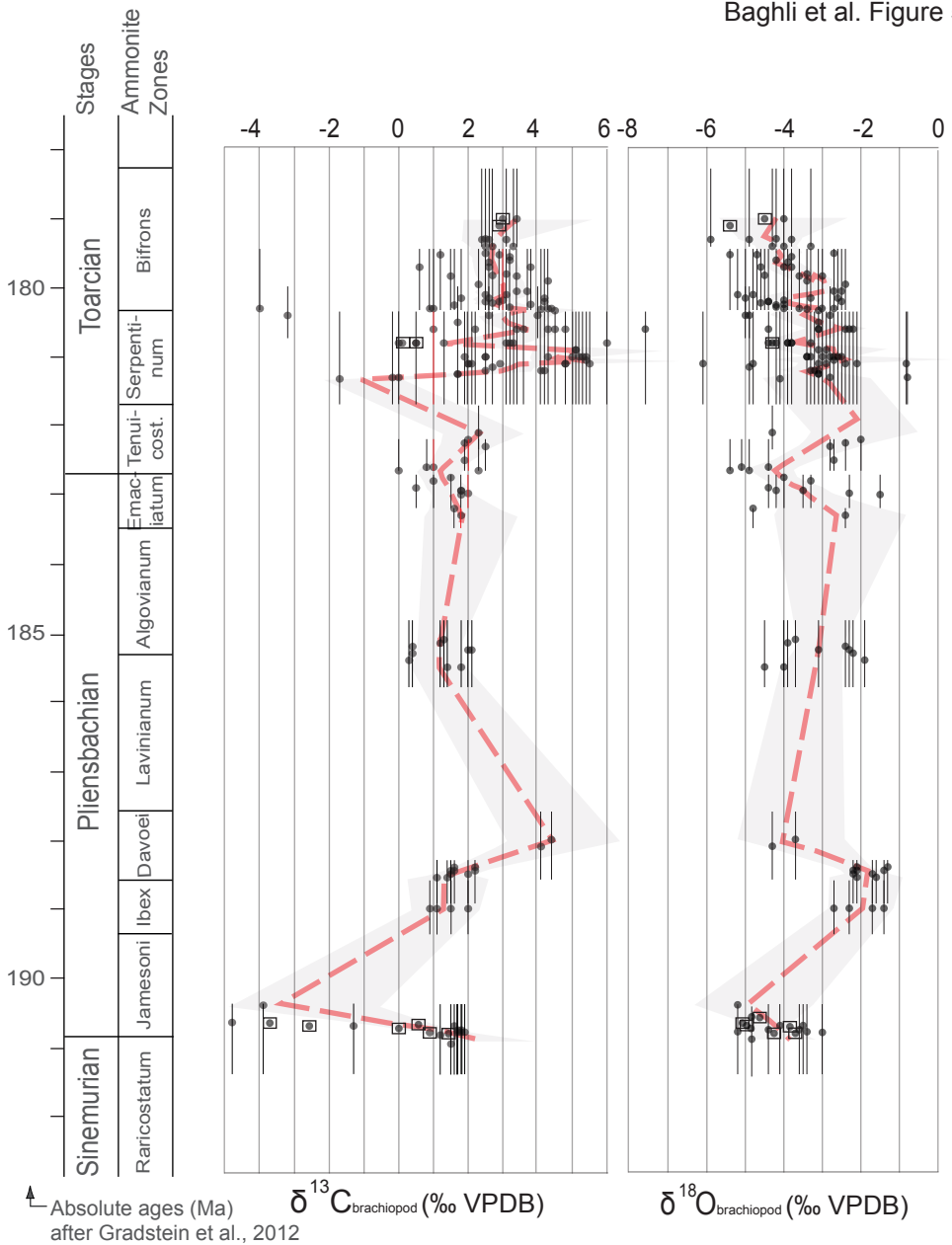


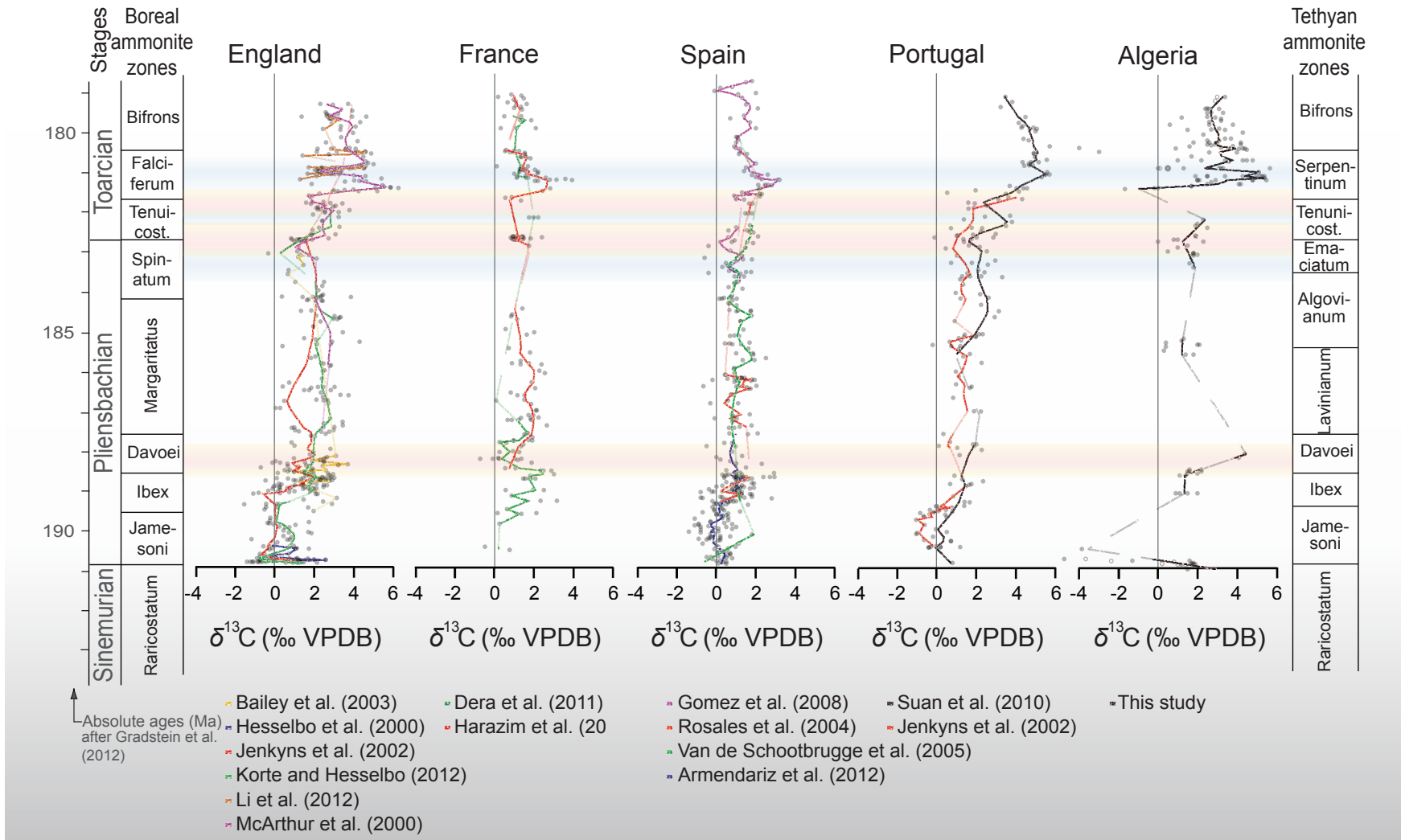


- Traras Montains
- Rhar Rhoubane Montains
- Saïda Mountains
- Ouarsenis Mountains
- Nador Montains

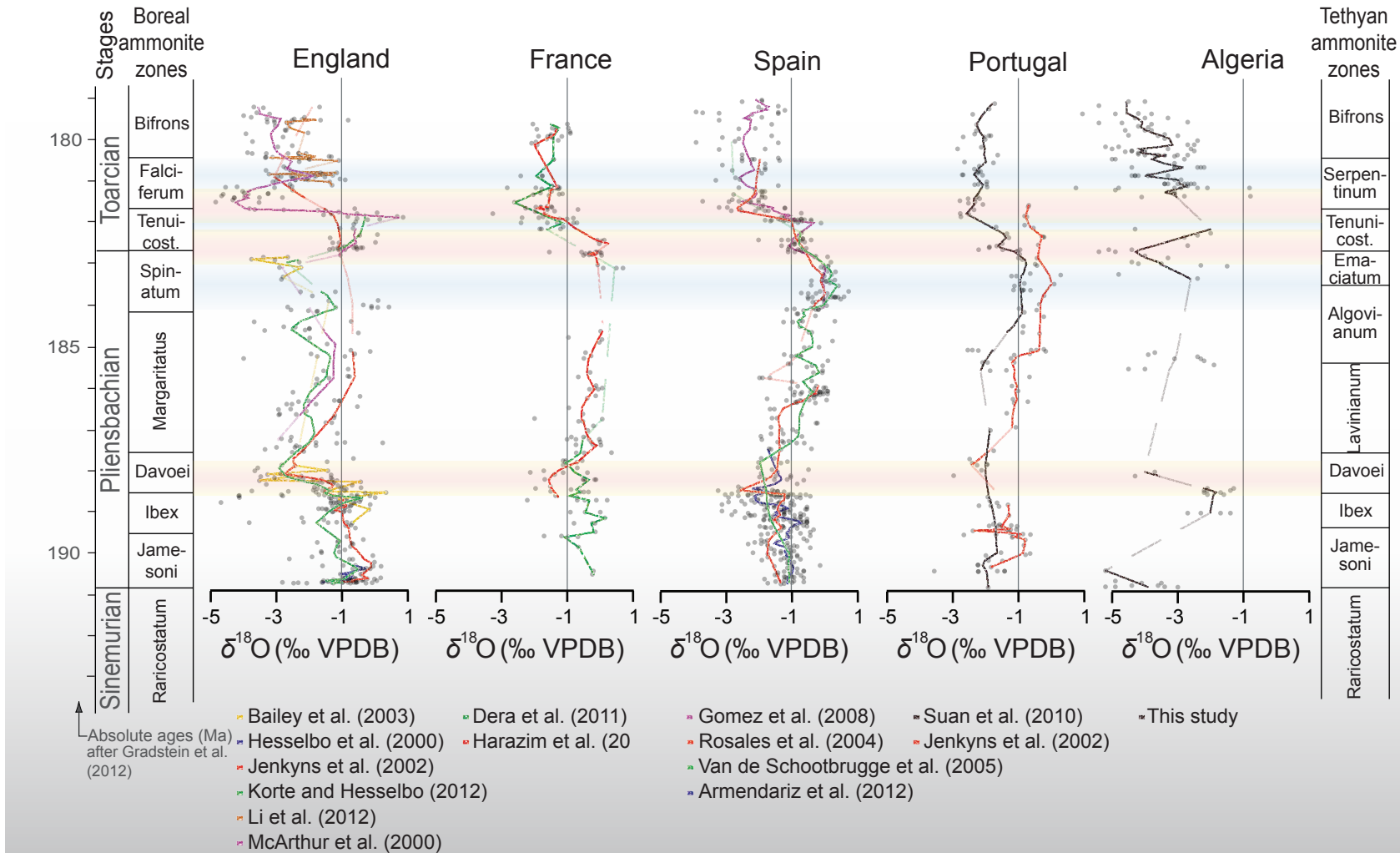




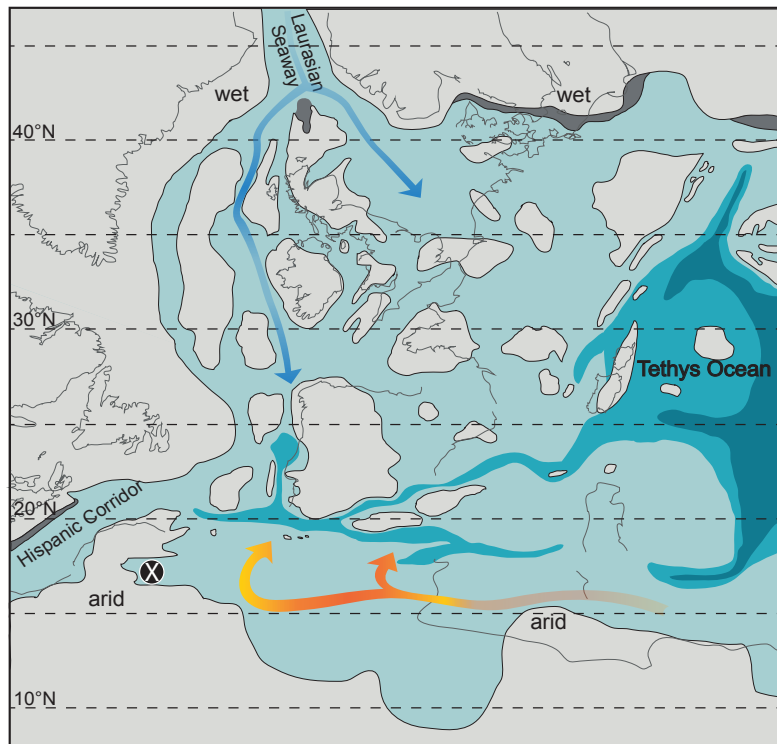




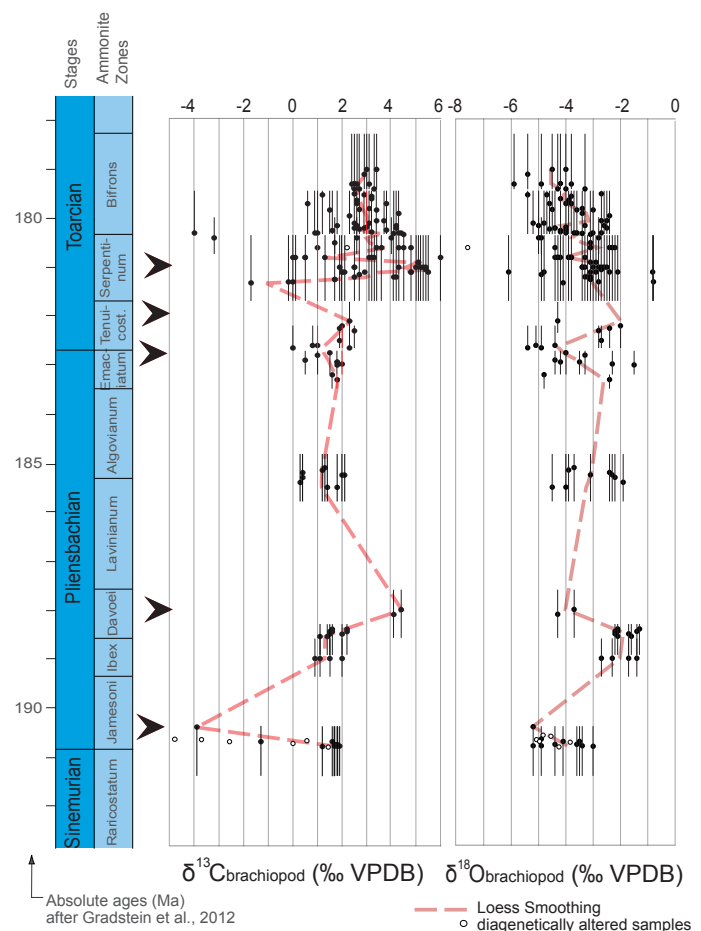
Baghli et al. Figure 6



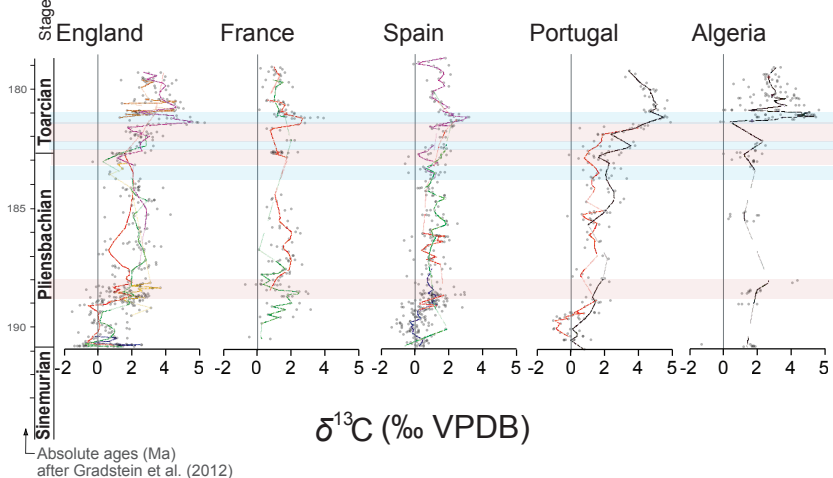
Baghli et al. Figure 7



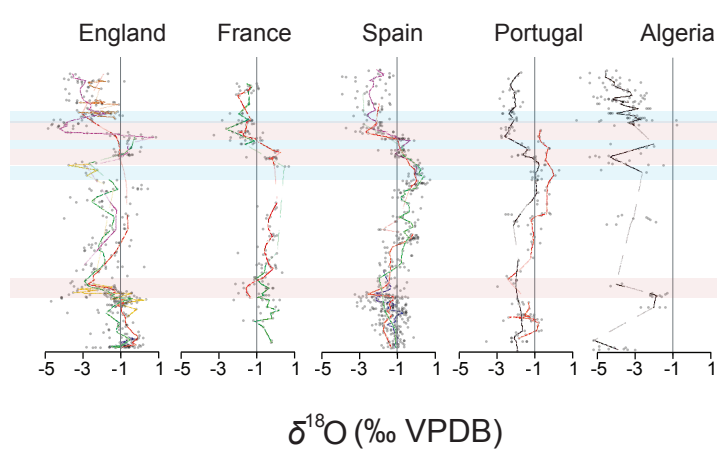
- Deep oceanic basin
- Deep epicontinental seas
- Shallow marine and carbonate platform
- Fluvialite, lacustrine
- Exposed landmasses
- X Study area (NW Algeria)
- low-saline, cool waters
- high-saline, warm waters



Carbon isotopes (+) offset



Oxygen isotopes (-) offset



Absolute ages (Ma) after Gradstein et al. (2012)