

Early Jurassic climatic trends in the south-Tethyan margin

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

Pittet, G. Suan

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     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>,
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     Pittet, B.<sup>1</sup>, Suan, G.<sup>1</sup>
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     <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.
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12
     Abstract
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     The Early Jurassic period was characterized by extreme environmental changes, as
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     reflected by major global carbon isotope anomalies and abrupt changes in oxygen
     isotope and elemental records of marine organisms. Available data suggest an overall
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     warm Early Jurassic climate interrupted by periods of severe cooling, with a climatic
     optimum during the early Toarcian. Available geochemical studies, however, have
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     mainly focused on the northern margin of the Tethys Ocean, so that the
20
                          extent of these
                                                 environmental
                                                                  perturbations,
     palaeogeographic
                                                                                   latitudinal
21
     palaeotemperature gradients and climate belt boundaries remain poorly constrained.
     Here we report the first stable isotope records of brachiopod shells (\delta^{13}C and \delta^{18}O
22
     values) from the Upper Sinemurian-Middle Toarcian interval in the southern margin of
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     the Tethys Ocean (northwest Algeria). These data were used to better constrain the
     palaeoenvironmental evolution of the North Gondwana margin during the Early
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26
     Jurassic, which likely played an important role on supra-regional climate. The
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     diagenetic history of the analysed brachiopod shells was monitored using scanning
     electron microscopy, and elemental (manganese and strontium) compositions. The
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     brachiopod \delta^{13}C and \delta^{18}O data show very similar trends as those reported for various
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     Tethyan regions, and record negative carbon and oxygen isotope excursions near the
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     Sinemurian-Pliensbachian and Pliensbachian-Toarcian transitions and during the
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32 Toarcian oceanic anoxic event (T-OAE). Despite these similarities, the carbon and

33 oxygen isotope records are systematically offset towards more positive δ^{13} C values

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

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

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43 **1. Introduction**

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The Early Jurassic period (~ 200-175 Ma) has been intensively studied because it was 45 a key time interval for the environmental evolution of the western Tethys. Indeed, this 46 time interval is characterized by extreme climate fluctuations, with possible glacial 47 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 (δ^{13} 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 considered to be characterized by a more arid climate with respect to the NW Tethys 63 (Aberhan, 2001; van de Schootbrugge et al., 2005). The carbon and oxygen isotope 64 trends have been documented for bulk carbonate, belemnites and brachiopods from 65 uppermost Pliensbachian and Toarcian strata of Morocco (Bodin et al., 2010; Krencker 66 67 et al., 2014; Ait-Itto et al., 2017). High-resolution carbon isotope records of bulk carbonate and associated organic matter in micrite samples from the upper Pliensbachian and lower Toarcian intervals in three sections from the Central High Atlas Basin in Morocco were presented and discussed by Bodin et al. (2016). Apart from these records, there is a desperate lack of information about the climate and environmental evolution in North Gondwana during most of the Early Jurassic interval. The distribution of climate belts and the role played by North Africa in Early Jurassic 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 between local *versus* global fluctuations of environmental parameters during the early 86 87 Jurassic.

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89 2. Geological background

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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 well-constrained at the ammonite Zone to Subzone level, and come from 22 sections 97 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).

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

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

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

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

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

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141 **3. Materials and methods**

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143 3.1. Brachiopod shell material and preparation

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A total of 156 brachiopod shells from NW Algeria, ranging in age from the upper 145 146 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 Direct-Q Millipore[®] purification system), and dried at 50°C for 48 h. The outermost layer 153 154 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).

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159 *3.2. Analyses*

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

The carbon and oxygen isotope measurements were performed using a Thermo Fisher Scientific (Bremen, Germany) Gas Bench II carbonate preparation device, connected to a Delta Plus XL isotope ratio mass spectrometer at the Institute of Earth Surface Dynamics of the University of Lausanne. The CO₂ extraction was done by reaction of 100–200 µg of powdered carbonate sample with anhydrous phosphoric acid at 70°C.

The carbon and oxygen isotope ratios are reported in the delta (δ) notation as the per 170 171 mil (‰) deviation relative to the Vienna Pee Dee belemnite standard (VPDB). The standardization of the δ^{13} C and δ^{18} O values relative to VPDB was done by replicate 172 173 analysis of the in-house working standard Carrara marble (UNIL-CM, δ^{13} C = 2.05 ‰, 174 $\delta^{18}O = -1.70$ ‰) within the analytical sequences. The $\delta^{13}C$ and $\delta^{18}O$ values of the reference CO₂ gas and the UNIL-CM standard were obtained by normalization with the 175 176 international reference materials NBS-19 limestone, NBS-18 carbonatite and LSVEC 177 lithium carbonate using the values reported in Brand et al. (2014). Analytical uncertainty (1σ) , monitored by replicate analyses of NBS-19 and UNIL-CM was not 178 179 greater than ± 0.05 ‰ for δ^{13} C and ± 0.1 ‰ for δ^{18} 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 for element analysis, including those sample that had apparently anomalous δ^{13} C and 182 δ^{18} O values, and also some samples with isotopic ratios in the range of marine 183 184 carbonates. For each brachiopod specimen, 1 mg of powdered sample was digested in 10 ml of HNO₃ solution (5% v/v) for 24 h. An aliquot of 1 ml from each stock solution 185 186 was then diluted in a solution of 9 ml of HNO₃ (2% v/v) and Indium (2 μ g/l). Each 187 solution was analysed for Sr and Mn by an inductively plasma mass spectrometer (ICP 188 MS), Agilen Technologies 7500 Series.

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190 3.3. Data treatment

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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 (δ^{13} C and δ^{18} O values) were smoothed to better highlight the long-term trends. Two 195 196 methods for smoothing data by locally weighted regression were implemented, the LOESS smoothing function using the software PAST 3.18 (Hammer et al., 2001), and 197 198 the lowess algorithm (Cleveland, 1981). Lowess is a linear polynomial smoother, 199 whereas the LOESS procedure uses a guadratic 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.

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

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211 **4. Results**

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213 4.1. Optical microscopy and SEM images

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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 δ^{13} C and δ^{18} O measurements. 228

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30 4.2. Carbon and oxygen isotope data and trace element concentrations

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

to -7.6 ‰) compared to the Mesozoic and modern brachiopod records, which are 236 strongly suggestive of strong post-mortem diagenetic overprint. No clear $\delta^{13}C-\delta^{18}O$ 237 relationship was observed for 95% of the brachiopod specimens, even when 238 239 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 curves (Fig. 5) indicate the stratigraphic extension of brachiopod ages according to 241 242 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 δ^{13} C trends reveal several events across the studied interval, which are more clearly visible in the LOESS smoothing curve (Fig. 5). The earliest event is recorded 249 250 at the Sinemurian/Pliensbachian boundary, where a pronounced ~2.5 ‰ negative 251 excursion, with δ^{13} C values as low as -4.0 % mark the Raricostatum-Jamesoni 252 interval. According to the original study of Alméras et al. (2007), the analysed brachiopods belong to an assemblage characteristic of the Sinemurian/Pliensbachian 253 254 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 lbex zone and in the lower part of the Davoei zone, δ^{13} C values are around 1.0 ‰. In the upper part of 257 258 the Davoei zone, more positive values reach up to 4.0 %. The δ^{13} C values return close 259 to ~1.0 ‰ near the transition between the Lavinianum-Algovianum Zones, and show a 260 ~1.0 ‰ negative excursion in the uppermost Pliensbachian (Emaciatum Zone). Towards the lowermost part of Tenuicostatum zone, the δ^{13} C values become more 261 262 positive with a +2.0 % shift, and decrease upwards to -1.0 % in the basal part of the Serpentinum Zone. After this event, a large shift toward positive values (up to 4.5 ‰) 263 264 characterizes the middle part of the Serpentinum zone, followed by some less pronounced fluctuations until the Bifrons zone with an average value of 3.0 ‰. 265

At the Sinemurian/Pliensbachian boundary, the δ^{18} O values range between -5.2 %and -3.0 % (on average $-4.3 \pm 0.68 \%$). The δ^{18} O values increase considerably in the lbex zone (lower Pliensbachian) and at the base of the Davoei zone, both with average values of -1.9 %. In the upper part of the Davoei zone, the δ^{18} O values decrease again 270 to an average of -4.0 ‰. Up to the beginning of the Emaciatum zone, the oxygenisotope values are rather stable with an average at -3.0 ‰. Passing from the 271 Emaciatum zone to the Pliensbachian/Toarcian boundary, the δ^{18} O values record a 272 273 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 δ^{18} O values of -3.8 ‰. In the aftermath, small isotopic rebounds to less negative values (of ~0.8 ‰) are recorded, after which δ^{18} O values, although fluctuating, 278 enter a ~2.0 ‰ long term trend toward more negative values extending from the middle 279 280 part of the Serpentinum zone to the middle part of the Bifrons zone and lasting around 281 2 Myrs.

The trace element contents of the selected samples revealed Sr contents varying between 130 and 1000 ppm, with most samples exceeding 450 ppm (on average 524 \pm 260) (Fig. 3). Mn contents range between 3 to about 80 ppm (on average 64 \pm 58). Half a dozen samples are depleted in Sr and enriched in Mn (Supplementary data).

Ten samples show anomalously negative δ^{13} C and δ^{18} O values, or anomalous contents in both Mn (< 80 ppm) and Sr (> 450 ppm) with respect to modern brachiopod shells secreted in equilibrium with ambient water (e.g., Brand and Veizer, 1980; Brand et al., 2003). These samples were hence excluded from the LOESS calculation, but are presented in the chemostratigraphic profiles in Figures 5 to 7.

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

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295 **5. Discussion**

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A certain degree of variability is observed both for carbon and oxygen isotopes (Fig. 5). Such variability is higher than that observed by Suan et al. (2010) in brachiopods of comparable age from the Portuguese Peniche section in Portugal. However, this can be partly due to the different environmental origin of the brachiopods sampled in five different regions of the NW Algeria. The water depth differences between the studied regions in Algeria are not important, as samples belong to outer platform, slope 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 brachiopod isotope composition in other time intervals (e.g., Giraud et al., 2016), some 305 prominent excursions are observed in the present work for δ^{13} C and δ^{18} O values from 306 307 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.

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314 5.1. Preservation of brachiopods shells

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Present-day articulate brachiopods (Rhynchonellida, Terebratulida) are strictly benthic 316 317 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 µm 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 these elements through the Phanerozoic in relation with episodes of aragonite and 344 345 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 layer varies between 450 and 1900 ppm at middle-to-low latitudes, while Mn values 347 348 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 δ^{13} C and 355 δ^{18} 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 effects (Freitas et al., 2005; Vander Putten et al., 2000), a possible minor temperature 361 362 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 δ^{13} Cbrachiopod versus δ^{18} Obrachiopod 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, resulting in a depletion in ¹³C and ¹⁸O of the secondary carbonate phases (Burdett et 373 al., 1990; Bates and Brand, 1991; McConnaughey et al., 1997; Auclair et al., 2003), 374 375 and therefore generating a positive linear δ^{13} C- δ^{18} O correlation. The large dispersion of our results in the δ^{13} C- δ^{18} O scatterplot and the absence of clear correlation between 376 values suggest that the isotopic composition of brachiopod shells did not suffered 377 378 major diagenetic overprint. However, the δ^{13} C and δ^{18} 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 δ^{13} C and δ^{18} 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 δ^{13} C and 384 δ^{18} 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).

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5.2. Late Sinemurian - Early Pliensbachian interval

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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., 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., 2016). Some authors interpreted this negative δ^{13} C excursion as reflecting the massive 398 399 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

case of NW Algeria, and in several other cases, the negative CIE at the 406 407 Sinemurian/Pliensbachian boundary appear to be as a sharp δ^{13} C-excursion. This can be explained either by the presence of a sedimentary condensation or a hiatus, or by 408 409 the occurrence of a sudden event. The second hypothesis is the most plausible, considering the abruptness of the δ^{13} C decrease in several localities, such as in the 410 411 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 across the Sinemurian/Pliensbachian transition. The variations in the δ^{18} O values of 415 416 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 dependent, and warmer waters lead to ¹⁸O depleted calcite (Lowenstam, 1961; O'Neil 419 420 et al, 1969). Although a certain dispersion is observed in Upper Sinemurian-Lower Pliensbachian δ^{18} O values, a trend to more negative δ^{18} O values in the brachiopod 421 422 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 CO₂ and CH₄ might have produced an increased greenhouse effect. 427 Furthermore, the oxygen isotope data of the studied areas do not show more positive 428 δ^{18} 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 Early Jurassic (Turner and Sherif, 2007). A positive δ^{18} O shift is conversely reported 431 432 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.

In the Davoei zone, another isotope event is recognized despite the relatively low data resolution in this interval. The δ^{13} C values in the NW Algeria sections show a positive excursion similar to that observed in other regions (Silva et al., 2011; Franceschi et al., 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 δ^{18} 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 ⁸⁷Sr/⁸⁶Sr values caused by enhanced continental 445 weathering (Jones and Jenkyns, 2001; Dera et al., 2009).

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447 5.3. Late Pliensbachian – Early Toarcian interval

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449 In the Late Pliensbachian, the δ^{13} 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 Hesselbo, 2011), and is accompanied in some sections, as those in Portugal and Spain 453 454 by the occurrence of black shales in the Margaritatus Ammonite Zone (Rosales et al., 2006; Suan et al., 2010; Silva et al., 2015). A trend towards higher δ^{18} O values in the 455 456 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 δ^{18} 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 SO₂ emissions in the Karoo-Ferrar Large Igneous Province. Similarly, Jourdan et al. (2008) inferred a 463 464 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 Pliensbachian strata from the Mochras borehole in Wales (Ruhl et al., 2016). This 469 470 cooling episode could also be ascribed to a global drawdown of atmospheric CO₂ 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).

Across the Pliensbachian/Toarcian boundary, the negative δ^{13} C excursion in Algeria 474 correlates remarkably well with that recorded in coeval sites in NW Europe, which was 475 interpreted as reflecting another pulse of massive injection of ¹³C-depleted carbon into 476 477 the oceans and the atmosphere (Hesselbo et al., 2007; Suan et al., 2008a). This hypothesis is consistent with the marked δ^{18} O negative excursion in biogenic 478 479 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; Pálfy and Smith, 2000; Percival et al., 2015, 2016). At the end of this Pliensbachian/Toarcian event, a relatively 483 small positive rebound of δ^{13} C values is noticed in almost all the sites in the lower part 484 485 of the Tenuicostatum Zone (Fig. 6), which correlates with a trend toward higher δ^{18} O 486 values (Fig. 7) and can be interpreted as a temporary cooling event.

487

488 5.4. Early-Middle Toarcian environmental evolution

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490 A marked shift towards lower δ^{13} 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 the massive and rapid injection of ¹³C-depleted carbon to the ocean-atmosphere 494 495 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 CO₂ emissions during 497 a renewed peak activity of the Karoo-Ferrar LIP (Suan et al., 2008a; Burgess et al., 2015), of thermogenic methane related to the LIP emplacement, or due to the 498 499 combined effects of all these processes (Beerling and Brentnall, 2007). These 500 hypotheses are consistent with the coeval negative δ^{18} O values recorded in NW 501 Algeria (Fig. 5) which, although less intense that observed in 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).

In the aftermath of the T-OAE, a marked 2.0 ‰ positive δ^{13} C excursion at the Serpentinum zone is recorded in NW Algeria as well as in many other localities (e.g., Jenkyns et al., 2002). This positive δ^{13} C excursion is considered to reflect the global burial of excess organic carbon, which in turn likely caused a climate cooling through 508 CO₂ 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 δ^{18} O values recorded in the same interval and most likely 511 reflecting a period of cooling (Fig. 7).

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513 5.5. Differences between C- and O-isotope records from NW Algeria and western

514 Tethys

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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.

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

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523 The δ^{13} 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 δ^{13} C data from NW Europe are systematically offset towards more 530 negative values than those of North Africa, with an overall offset of ~0.5 ‰ (Fig. 6). 531 This offset of the LOESS-smoothed δ^{13} 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, δ^{13} C values of DIC decrease with increasing 537 depth due to the remineralisation of ¹³C-depleted organic matter. The lower δ^{13} C 538 values recorded in belemnites suggest that they secreted their calcite rostra in deep waters, more enriched in free ¹³C-depleted DIC (Wierzbowski and Joachimski, 2007; 539 540 Dutton et al., 2007). Such differences could also explain why belemnite δ^{13} 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 (Fig. 6). The higher amplitude of the δ^{13} C-excursions measured in NW Europe can be 543 tentatively interpreted as the result of a greater supply of freshwater by rivers to the 544 545 epi-continental N-Tethyan basins. In fact, the oceanic DIC is fed by freshwaters enriched in oxidized organic matter thus have a ¹³C-depleted DIC. This leads to the 546 547 precipitation of biogenic carbonates with more negative δ^{13} C values (Clavton and Degens, 1959). It has been also suggested that higher productivity at higher latitudes 548 549 led to a ¹³C 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 changes in belemnite δ^{13} C records. 552

The tectono-climatic regime of North Africa might also have influenced the C-isotope 553 554 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 ¹³C and ¹⁸O enrichment as the production of light isotopes 559 is favoured in water vapour. The resulting covariance between δ^{18} O and δ^{13} C is hence 560 generally considered as an evidence of hydrological closure in palaeo-lakes deposits. Accordingly, in the case of continental runoff of such palaeo-lake basins, the δ^{13} C 561 562 signal in adjacent marine basins should be significantly shifted towards more positive 563 values (Talbot, 1990; Deocampo and Jones, 2014). However, the measured δ^{13} C 564 values in NW Algeria sections are in the range of "normal" marine carbonates, and do not show such very positive values. It seems therefore unlikely that the NW Algeria 565 566 basins have been under the influence of continental runoff from surrounding arid lakes.

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

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Although the carbon isotope records in NW Algeria show the main events reported in other Tethys settings, the oxygen isotope data show an important difference. Systematically, the δ^{18} O values are 1.5 to 2.0 ‰ lower compared to those from NW Tethys sections (e.g., N Spain, France, England), where isotope data mainly derive from belemnite calcite. The offset of the LOESS-smoothed δ^{18} O values is statistically significant. The relatively low δ^{18} 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 were both located at lower palaeo-latitudes, and therefore were likely to have 578 579 experienced warmer seawater temperatures (i.e., warm subtropical region; Aberhan, 2001) than European sites. Interpreting the δ^{18} O records solely in terms of temperature 580 581 and considering available δ^{18} 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 δ¹⁸O values addition, the high belemnite recently reported for the 586 Pliensbachian/Toarcian boundary of Issouka in the Central Atlas of Morocco (0 to -1.0 %; Aït-Itto et al., 2017) are almost identical to those obtained for most European 587 588 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 δ^{13} 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 δ^{18} 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 δ^{18} 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 δ^{18} O records 600 obtained on the biogenic calcite of organisms with similar ecology and physiology to avoid such palaeoecological complications. The new δ^{18} O brachiopod data from 601 Algeria are still 0.5 to 1.0 % lower than brachiopod δ^{18} O data from Portugal, suggesting 602 603 a more reasonable difference in temperature of about 2 to 4°C between bottom water 604 masses of these two areas.

It is very likely that the recorded spatial changes in δ^{18} O values of biogenic carbonates also reflect spatial changes in seawater δ^{18} O values, as a result of different hydrological regime between the N and S Tethys margins and thus indirectly related changes in seawater salinity. The northern margin of Gondwana was characterized 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 turned to be evaporitic (Turner et al., 2001). Lithostratigraphic studies of continental 612 613 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 Benshilil, 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, according to the seasons, between 50 ‰ and 250 ‰ (Amarouayache et al., 2018). 623 Since δ^{18} O would change by about 0.2 % per unit change in salinity (Sharp, 2017), a 624 5.0 % fluctuation in salinity would be accompanied by a 1.0 % change in the δ^{18} O 625 626 value in the ocean shallow-waters (Sharp, 2017). In the case of runoff of these waters into the ocean, a large δ^{18} O positive offset would be expected. As we observe, 627 628 conversely, a negative δ^{18} 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 δ^{18} O values of NW Algeria and the δ^{18} 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 systems should have produced lower δ^{18} O values in the northernmost basins (Rosales 637 638 et al., 2004b; Dera and Donnadieu, 2012), which may explain the anomalously low belemnite δ^{18} O values in the UK compared to all other belemnites records (Fig. 7). 639 640 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

as with respect to other localities (Fig. 7). Indeed, the basal Tenuicostatum Zone is
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). Some authors explain this trend by a deepening of sea-level near the 645 Pliensbachian/Toarcian, which is indeed marked by a supra-regional marine 646 647 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 δ^{18} 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 δ^{18} 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. Brachiopods conversely were sessile organisms, generally inhabiting shallow-water 661 662 depth, making them more susceptible to record temperature fluctuations and changes 663 in evaporation-precipitation budgets than belemnites.

In summary, all these observations indicate that the δ^{18} O trends in brachiopod shell in NW Algeria reflect mainly temperature fluctuations, and suggest that a latitudinal temperature gradient of around 2 to 4°C existed between bottom water masses of the central and southern part of the western Tethys during the Early Jurassic.

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669 **6. Summary and conclusions**

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

studied area. The negative δ^{13} C excursion near the Sinemurian/Pliensbachian 678 679 boundary in NW Algeria, previously documented in England and Portugal, is accompanied by a decrease of δ^{18} O values, supporting the hypothesis of a 680 681 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 δ^{18} 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 δ^{18} O values.

- 686 The new NW Algerian records are also valuable in that they reveal a systematic 687 positive offset of ~0.5 ‰ for δ^{13} C values and negative offset of ~1.5 ‰ for δ^{18} O values compared to European sections. We interpret these offsets as mainly reflecting higher 688 689 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 significant differences in terms of assemblage composition between the S and the N 695 696 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

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712 **References**

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1157

1158 Legend of figures

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

1166

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

1170

1171 **Fig. 3.** The relationships between δ^{18} 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.

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1175 **Fig. 4.** Cross Plot of δ^{13} C *versus* δ^{18} 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.

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Fig. 5. C and O isotope compositions of brachiopod calcite from the Upper Sinemurian to the Middle Toarcian in north-western Algeria. Numerical ages are derived from "The Geologic Time Scale 2012" (Gradstein et al., 2012). The relative stratigraphic position of brachiopods in the different sections was respected and vertical bars on the isotope curves indicate the stratigraphic extension of brachiopod ages according to ammonite zones and sub-zones. The colours of the symbols correspond to the region where the samples come from. A LOESS Smoothing with a 0.1 factor was applied using Past 3.18. Square symbols show samples recording anomalous Sr or Mn contents andtherefore not considered for the LOESS smoothing.

1188 **Fig. 6.** C-isotope comparisons and chemo-stratigraphy correlation between sections studied in Algeria: Traras, Rhar Roubane, Ouarsenis, Saïda, and Nador Mountains 1189 1190 (this work); in England: Cleveland Basin (Hesselbo et al., 2000b; Mcarthur et al., 2000; Jenkyns et al., 2002; Bailey et al., 2003; Korte and Hesselbo, 2011; Li et al., 2012); in 1191 1192 France: Grands Causses Basin, Sud-Est Basin and Paris Basin (Dera et al., 2011; Harazim et al., 2013); in Spain: Iberian Range, Asturias, and Basque-Cantabrian Basin 1193 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.

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Fig. 7. O-isotope comparisons and chemo-stratigraphy correlation between sections 1201 1202 studied in Algeria: Traras, Rhar Roubane, Ouarsenis, Saïda, and Nador Mountains 1203 (this work); in England: Cleveland Basin (Hesselbo et al., 2000b; Mcarthur 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 al., 2012); in Portugal: Lusitanian Basin (Jenkyns et al., 2002; Suan et al., 2010). All of 1208 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.

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

- specimen according to their stratigraphic position and corresponding ammonitebiozonation 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
- have been redrawn after Benhamou (1983, 1996); Elmi et al. (1985); Kharroubi (1987);
- 1226 Mekahli (1988).





Carbonate ramps

Shoals and importants gaps

Sabkha confined facies

Probably emergent areas

- Rhar Roubane Mountains
- **3** Saïda Mountains
- **4** Ouarsenis Mountains
- **6** Nador Mountains





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



Baghli et al. Figure 3



Baghli et al. Figure 4













