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1 Modern drought conditions in western Sahel unprecedented in 2 the past 1600 years 3

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32 **Abstract**

33 As climate model uncertainties remain very large for future rainfall in the Sahel, a multi-
34 centennial perspective is required to assess the situation of current Sahel climate in the
35 context of global warming. We present here the first record of hydroclimatic variability over
36 the past 1600 years in Senegal, obtained from stable oxygen isotope analyses ($\delta^{18}\text{O}$) in
37 archaeological shell middens from the Saloum Delta. During the preindustrial period, the
38 region was relatively humid, with maximum humidity reached during the period from AD
39 1500 to AD 1800, referred to as the Little Ice Age. A significant negative link is observed at
40 the centennial scale between global temperature and humidity in the Sahel that is at odds with
41 the expected effects of latitudinal shifts of the intertropical convergence zone during the last
42 millennium. In the context of the past 1600 years, the Western Sahel appears to be
43 experiencing today unprecedented drought conditions. The rapid aridification that started *ca.*
44 AD 1800 and the recent emergence of Sahel drought from the natural variability point to an
45 anthropogenic forcing of Sahel drying trend. This new long-term perspective suggests that the
46 recovery of Sahel rainfall in the last decade may only result from short-term internal
47 variability, and supports climate models that predict an increase of Sahel drought under future
48 greenhouse climate.

49

50 **Keywords:** West African Monsoon, Climate Change, Paleoclimate, Shell middens

51 **1. Introduction**

52 Precipitation in the Sahel declined by ~30% in the 1960s, initiating a severe multi-
53 decadal drought that represents one of the most significant climate change episodes in the
54 global instrumental record (Christensen et al. 2013), with devastating socio-economic
55 impacts. The contributions of global warming and land use to this event, however, are still not
56 clearly understood, partly because of the underperformance of climate models in the
57 simulation of the West African Monsoon (WAM), and partly because of the shortness of the
58 instrumental record (Biasutti 2013). Rain gauge data in the Sahel start in AD 1851 in Saint
59 Louis, Senegal, but regional data is generally considered robust only after AD 1900
60 (Nicholson et al. 2012). In the instrumental record, the Sahel climate has been dominated by
61 multidecadal variability which may mask longer-term trends. Assessing current trends in
62 Sahel rainfall and determining the potential role of anthropogenic activities in this climatically
63 vulnerable region requires a multi-centennial perspective of natural variability, which is so far
64 lacking in this region (Nash et al. 2016). The closest continuous record documenting
65 hydroclimate change in the past two millennia comes from Lake Bosumtwi, Ghana (Shanahan
66 et al. 2009), which is not located in the Sahel but in the convection area of the WAM.

67 In this study, we use oxygen isotope ratios ($\delta^{18}\text{O}$) of *Senilia senilis* shells from the
68 Saloum Delta in Senegal, to construct the first record of centennial variability of the
69 hydrologic balance of the Western Sahel, spanning the last 1600 years. The people of the
70 region have long depended on *S. senilis* as a staple food, as evidenced by the massive
71 archaeological shell middens (up to 15 m high) that have been accumulated throughout the
72 Delta (Thilmans and Descamps 1982; Hardy et al. 2016). By virtue of their size and antiquity,
73 these middens present a unique opportunity to obtain diachronic records of hydrological
74 change.

75 The Saloum Delta is a mangrove estuary with a small catchment basin, disconnected
76 from the Senegal River or the Gambia River catchment basins. The extremely steep latitudinal
77 precipitation gradient at this location makes the Saloum estuary highly sensitive to changes in
78 the WAM (Fig. 1). Although a contrasted evolution of rainfall in central and western Sahel
79 has been described in the past decade, and projected in the future (Biasutti 2013), instrumental
80 observations show a remarkable coherence of precipitation for the whole of the Sahel
81 (Nicholson 2014) and in particular a strong correlation between precipitation in the Saloum
82 Delta and the Sahel (Wade et al. 2015) (Fig. 1). Assuming the stationarity of this spatial
83 coherence, we consider that our record is largely representative of conditions in at least the
84 western Sahel region and potentially the whole Sahel band.

85

86 **2. Material and methods**

87

88 **2.1. Bivalve shells**

89

90 Modern shells of *Senilia senilis* were collected live on intertidal mud flats in five
91 locations distributed across the mangrove delta close to the archaeological sites (Fig. 1).
92 Fossil shells from sites A10, A12, and A49 were collected on the surface of those shell
93 middens. In Dioron Boumak shell midden, the full stratigraphy of the shell accumulation
94 could be sampled in the outcrop produced by tidal erosion (Fig. 2). In Tioupane and
95 Diofandor, we sampled the middens from the surface to depths of 5.50 and 1.15m
96 respectively, in outcrops produced by ancient exploitation for lime production. The shell
97 middens were almost exclusively composed of *S. senilis* shells, with infilling composed of
98 sand, silts, and ashes, and had therefore a very poor cohesive structure (Fig. 2). After cleaning
99 the outcrops, the accumulations were divided into a continuous vertical sequence of

100 stratigraphic sections. About 30 shells were collected on the field in each section from the
101 surface to the bottom. Subsequently, subsamples of 1 to 7 shells per section were selected in
102 the lab based on shell size and preservation. Therefore, each shell comes from a depth interval
103 and as such, it represents a short ~2-10-yr window (~5 years in average) randomly drawn
104 from the time period when the section was accumulated. The depth, thickness and the number
105 of shells analyzed are indicated for each stratigraphic layer in the depth-age model plots (Fig.
106 3).

107 Shell preservation was tested by observing the aragonite microstructure in the hinge of a
108 subsample of shells from Dioron Boumak using a Scanning Electronic Microscope (SEM).
109 Images revealed a perfectly preserved cross-lamellar structure (Fig. 4C). To guarantee a well
110 preserved isotopic signal, aragonite powder samples for isotopic analyses were collected in
111 cross section to avoid surface alteration, in the hinge, which is the thickest and best preserved
112 part (Fig. 4A). For shells that were not observed by SEM, preservation was evaluated by
113 optical microscope observation of the polished cross sections. Aragonite opacity and the fact
114 that growth lines could be observed in every part of the shells are complementary evidence
115 that shells and their original isotopic composition were well preserved.

116

117 2.2. Chronology

118

119 For Dioron Boumak, Tioupane, and Diofandor, depth-age models were calculated from
120 radiocarbon dates obtained on charcoal fragments collected along the middens' stratigraphy
121 (Table 1). Radiocarbon date series were all stratigraphically consistent. Bayesian depth-age
122 models were calculated using the Bacon program (Blaauw and Christen 2011) with a broad
123 distribution of accumulation rate and a low memory strength parameter to allow for large
124 flexibility in the accumulation rate (Fig. 3). Default values of $t.a$ and $t.b$ (3 and 4 respectively)

125 parameters of the student's t-distribution were used. Based on these models, calibrated dates
126 and their 95% confidence interval were calculated for the depth boundaries of the
127 stratigraphic sections. Three shells (from A10, A12, and A49) were directly radiocarbon dated
128 from a shell hinge fragment. Calibration of these dates was performed using mixed terrestrial
129 and marine radiocarbon calibration curves (Reimer et al. 2013). We used the marine reservoir
130 age ($R=511\pm 50$ yrs) estimated for Senegal by Ndeye (2008), and a percentage of marine
131 carbon calculated from the shell $\delta^{13}\text{C}$ value, considering two end members of 0‰ for marine
132 carbon and -25‰ for terrestrial carbon (72% of marine carbon for A10, 64% for A12, and
133 92% for A49, see Table 1).

134

135 2.3. Shell analysis

136

137 45 modern shells and 120 archaeological shells were analyzed. Shells were embedded in
138 polyester resin and radially cut with a low-speed diamond saw to extract a 1mm thick lamina
139 that was then mounted on a microscope slide and polished. A sample of aragonite powder was
140 collected by drilling a thin groove across the hinge cross section using an automated
141 microsampler Micromill®. Each groove crossed the most part of the hinge growth, so that the
142 aragonite sample integrates most of the shell lifetime, ~5 years in average (Fig. 4A). One
143 fossil shell and 16 modern shells were serially microsampled in the outer shell layer with a
144 time resolution corresponding to approximately one month following the procedure described
145 by Azzoug et al. (2012a). Aragonite powder samples of typically 50-100 μg were analyzed for
146 their isotopic composition ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) using a Thermo Finnigan Delta V isotope ratio
147 mass spectrometer coupled to a Kiel IV carbonate device. Long-term reproducibility for $\delta^{13}\text{C}$
148 and $\delta^{18}\text{O}$ was better than 0.05‰ and 0.08‰ respectively based on repeated internal lab
149 standard analyses. Isotopic analyses of aragonite samples from Tioupane and Diofandor were

150 replicated 2 to 5 times. Shell isotopic ratios ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) were reported as relative
151 deviation from the V-PDB international reference. Since isotopic standards for calcium
152 carbonates are all calcitic and because aragonite and calcite have different acid fractionation
153 factors, a correction of -0.38‰ (at 70°C) was applied to aragonite isotopic values following
154 Kim et al. (2007).

155

156 2.4. Environmental monitoring and calibration

157

158 As a function of the region's highly seasonal rainfall regime, the salinity (and the water
159 oxygen isotopes) of the estuary varies seasonally from brackish in the monsoon season to
160 hypersaline in the dry season. The relationship between shell $\delta^{18}\text{O}$ and local climate was
161 determined by an *in situ* environmental monitoring in Toubakouta and isotopic analysis of
162 modern shells collected live in Toubakouta and in 4 other sites from across the estuary (Fig.
163 1). In Toubakouta, sea surface temperature was measured hourly with an Onset Tidbit.v2
164 datalogger since 2011. Rainfall was measured with an Onset RG-3 rain gauge datalogger
165 since 2011. Sea water samples were collected weekly for $\delta^{18}\text{O}_w$ since November 2010. The
166 oxygen isotopic composition of water samples was measured using a dual inlet Isoprime
167 IRMS with the conventional $\text{CO}_2\text{-H}_2\text{O}$ equilibration method (Cohn and Urey 1938). The
168 standard error estimated from repeated standard measurement was 0.07‰ . Water $\delta^{18}\text{O}$ values
169 were reported as relative deviations from the V-SMOW international standard. High-
170 resolution isotopic records of modern shells from Toubakouta allowed us to study seasonal
171 scale variability of shell isotopic ratios and their relationship with local climate (Fig. 5).

172 The first observation is that high resolution shell $\delta^{18}\text{O}$ records from Toubakouta show a
173 good reproducibility (Fig. 5A). Then, we calculated $\delta^{18}\text{O}_w$ from modern shell $\delta^{18}\text{O}$ using the
174 Grossman and Ku's third paleotemperature equation (Grossman and Ku 1986) and the local

175 mean SST annual cycle. Since calculated values faithfully reproduce measured water $\delta^{18}\text{O}_w$
176 (Fig. 5D), we conclude that *S. senilis* shells precipitate in isotopic equilibrium. At the seasonal
177 scale, $\delta^{18}\text{O}_w$ accounts for $\sim 3\%$ in aragonite $\delta^{18}\text{O}$ variations while the 8°C range of
178 temperature may account for 1.7% change in shell aragonite. However, because the
179 temperature cycle is not in phase with $\delta^{18}\text{O}_w$ ($\delta^{18}\text{O}_w$ lags by ~ 3 months), it contributes
180 positively to the shell $\delta^{18}\text{O}$ range only during the rainy season which represents a temperature
181 variation of $2\text{--}3^\circ\text{C}$. The temperature contribution to the shell isotopic variations is thus only
182 0.4 to 0.6% . $\delta^{18}\text{O}_w$ is the primary driver (85%) of shell $\delta^{18}\text{O}$ variations at the seasonal time
183 scale. On longer time scales, this predominance of $\delta^{18}\text{O}_w$ influence is expected to be even
184 larger since rainfall long-term variability is much larger than temperature variability.

185 The Toubakouta rainfall record (Fig. 5E) shows that the rainy seasons correspond exactly
186 to the periods of decreasing $\delta^{18}\text{O}_w$ values (reversed scale), while dry seasons correspond to
187 periods of increasing $\delta^{18}\text{O}_w$ values, a pattern that is also clear in shell $\delta^{18}\text{O}$ records. The
188 estuary water isotopic values, closely linked to salinity (Fig. 6), become lower than sea water
189 values in the rainy season and higher than the sea water values in the dry season (Fig 5C). The
190 sea water $\delta^{18}\text{O}_w$ measured at the mouth of the Delta in the dry season is close to 1% . This *in*
191 *situ* calibration dataset demonstrates that shell $\delta^{18}\text{O}$ faithfully reflect $\delta^{18}\text{O}_w$ which is closely
192 linked to precipitation.

193

194 2.5. Spatial variability across the Delta

195

196 Since fossil shells were collected in different sites within the Saloum Delta, we studied
197 the spatial variability of isotopic values in modern shells from across the Delta. First, modern
198 shell $\delta^{18}\text{O}$ values show no statistically significant difference between the five collection sites
199 (Fig. 7). This means that the variability within the fossil $\delta^{18}\text{O}$ record cannot be explained by

200 changes in their origin. In contrast, there is a clear negative trend for shell $\delta^{13}\text{C}$ toward the
201 continent (Fig. 7). This gradient can be explained by the mixing of two carbon sources:
202 dissolved inorganic carbon from the ocean ($\sim 0\text{‰}$) and terrestrial carbon, primarily from
203 mangrove litter ($\sim -25\text{‰}$). Since this gradient is also observed during the dry season when the
204 Saloum Delta is an inverse estuary, the mixing of carbon is not due to the mixing of marine vs
205 runoff water but to tidal mixing. In estuaries where DIC $\delta^{13}\text{C}$ has very large variations,
206 mollusk shell $\delta^{13}\text{C}$ variability primarily reflects DIC $\delta^{13}\text{C}$ (Gillikin et al. 2006). In this setting,
207 shell $\delta^{13}\text{C}$ can thus be used to estimate the ratio of terrestrial vs marine carbon 1) for
208 calibration of shell radiocarbon dates, and 2) to estimate past changes in the tidal mixing due
209 to the estuary geomorphology. $\delta^{13}\text{C}$ values of fossil shells from Diofandor and Tioupane fall
210 on the trend defined by modern shells (Fig. 7) suggesting that despite the strong dynamicity of
211 delta environments, the general morphology of the estuary remained similar to the current one
212 during the studied period. $\delta^{13}\text{C}$ values from Dioron Boumak shells are more positive than the
213 prediction of the modern trend, suggesting a stronger connection with the sea, but still within
214 the range of modern Saloum shells. The last marine transgression has been dated by Faure et
215 al. (1980) at 1545 ± 120 ^{14}C BP and was approximately 0.5m high. This would correspond to
216 the very beginning of our record and could partly explain Dioron Boumak $\delta^{13}\text{C}$ values, but
217 could not account for the more negative $\delta^{18}\text{O}$ values at Dioron Boumak compared to modern
218 values. The general consistency of past $\delta^{13}\text{C}$ values with the modern estuary geomorphology,
219 and the fact mean shell $\delta^{18}\text{O}$ values are constant across the estuary show that the estuary
220 morphology cannot account for the long-term oxygen isotope variability. This result confirms
221 geomorphological studies which suggested that the geometry of the Saloum Delta 1500 years
222 ago was already similar to the modern conditions (Barusseau et al. 1995; Diara and Barusseau
223 2006).

224

225 2.6. Accuracy of bulk isotopic analyses in *Senilia senilis* shells

226

227 Most shell isotopic values in this study were obtained from bulk sampling in the hinge
228 part of the shell. Isotopic values obtained from bulk samples could potentially be biased
229 because bivalve shell growth can stop or change because of physiological factors (age,
230 reproduction) or environmental factors (Goodwin et al. 2003). We discuss here the causes of
231 shell growth variations, their potential effect on bulk isotopic values, and eventually evaluate
232 the accuracy of bulk isotopic analyses in this study. *S. senilis* is a euryhaline species that has
233 been observed live in salinities ranging from 0 to 60 psu, on intertidal mudflats in coastal
234 lagoons from Mauritania to Angola (Elouard and Rosso 1977; Debenay et al. 1994; Azzoug
235 2012; Lavaud et al. 2013) . *S. senilis* can thus support and record large salinity changes.
236 Seasonal salinity variations in our study sites are clearly within this tolerance range.

237 The study of growth increment thickness and periodic growth lines (sclerochronology)
238 related to lunar tidal cycles observed in *S. senilis* shells (Fig. 4) allows for a detailed control
239 of growth rate changes or stops through the shell's life (Azzoug et al. 2012a; Debenay et al.
240 1994; Lavaud et al. 2013). In Mauritania, shell growth rate is lower in the cool season
241 (November-February) (Lavaud et al. 2013), while in the Saloum Delta, it is lower during the
242 rainy season (July-September) (Azzoug et al. 2012a). The number of periodic growth lines per
243 year observed in shells indicates that shells grow continuously throughout the year so that the
244 full annual cycle is recorded (Azzoug et al. 2012a). In some bivalve species, mainly from
245 mid-to high latitudes, the length of the seasonal slowdown or shutdown increases with age
246 which produces a progressive decrease in the seasonal isotopic range and an overall
247 systematic trend in high resolution isotopic shell records (Goodwin et al. 2003). Such a
248 systematic trend was not observed in modern or fossil shells of *S. senilis* (Azzoug 2012).

249 Nevertheless, growth slowdown in the rainy season may bias bulk analyses toward dry
250 season conditions characterized by less negative $\delta^{18}\text{O}$ values. We estimated the effect of bulk
251 sampling on isotopic values by comparing bulk values with average values obtained from
252 high-resolution isotopic records in 15 modern shells from 3 locations in the Saloum Delta.
253 Shell high-resolution isotopic records were resampled at a constant monthly resolution to
254 avoid biases due to variations in temporal resolution. The difference between bulk isotopic
255 values and averaged monthly values ranged from -0.57‰ to 0.83‰. The average difference is
256 $0.22 \pm 0.44\text{‰}$ (1σ). A student t-test shows that the bias due to bulk sampling is not different
257 from zero at the 0.05 significance level. This demonstrates that bulk sampling does not
258 significantly affect the accuracy of isotopic data in this study. It does introduce noise to the
259 signal ($\sigma \approx 0.44\text{‰}$), that we interpret as primarily related to variations in the relative
260 contribution of rainy and dry seasons in the shell life span. For serially sampled shells, we
261 used the average of the hinge bulk value and of the high-resolution record mean.

262

263 2.7. A composite record of shell $\delta^{18}\text{O}$ centennial variability

264

265 The record is composed of 164 isotopic values including 120 archaeological shell values
266 (Fig. 8). Except for the three shells that were directly radiocarbon dated, each of these
267 archaeological shell values is floating in a time interval defined by the ages of the upper and
268 lower depths of the stratigraphic section where the shell was collected (see section 2.1). Since
269 these shells do not have an assigned depth, their age cannot be directly estimated from the age
270 model. Only the upper and lower boundaries of stratigraphic sections can be dated using the
271 age model. From this floating discontinuous dataset, we built a continuous record of the
272 probability distribution of the average shell aragonite $\delta^{18}\text{O}$ value. This was achieved through a
273 Monte Carlo (MC) sampling where iterations represent random possibilities of the floating

274 chronology. In every iterations, the first step is a random sampling of the age of the
275 stratigraphic sections' limits from their respective probability density function determined by
276 the Bayesian depth-age model (Fig. 3). Once the time boundaries of stratigraphic sections
277 have been fixed, the second step assigns a random age for each shell within the time interval
278 represented by its stratigraphic section. In addition to the chronology uncertainty, the
279 uncertainty on $\delta^{18}\text{O}$ values related to bulk sampling was also sampled from a normal
280 distribution ($\mu=0$, $\sigma=0.44\%$). A 7 point moving average is then calculated and annually
281 interpolated. 10 000 iterations provide us with an ensemble of potential moving average
282 curves. Moving average curves smooth out the high frequency variability and most of the data
283 scattering related to shell sampling. This ensemble yields the probability distribution through
284 time of the centennial scale average shell $\delta^{18}\text{O}$ value, represented by the gray scale in Fig. 8, a
285 median value and a 95% confidence interval. The procedure was tested with running windows
286 of 5, 7, and 10 points. The running window size slightly affects the size of the confidence
287 interval but not the main features of our results, which shows that the method is robust to
288 different window widths. The statistical representativeness of the dataset through time was
289 estimated by the average number of shells per century in the MC experiment (Fig. 8B). The
290 least robust period is the 14th century with an average number of only 1.5 shells, while the
291 most robust (apart from the modern sample) is the 13th century with 12.8 shells. We have 7.1
292 shells per century in average (excluding the modern period). Considering that each shell
293 integrates in average ~ 5 years, and assuming no significant overlap, this means that the record
294 was built with ~ 35 years per century in average, which is statistically robust to estimate
295 centennial mean conditions ($\text{CI} = \pm 1.96 \cdot \sigma / \sqrt{N} = \pm 0.17\%$) even considering all possible
296 uncertainty sources related to mollusk shells (Carré et al. 2012). The raw dataset, the
297 reconstructed annual isotopic time series and the Matlab code of the MC procedure are
298 available in the electronic supplementary materials (Online Resource 1, 2, 3).

299

300 3. Results

301

302 Significant centennial variations of shell $\delta^{18}\text{O}$ show values depleted relative to the present
303 day during the whole preindustrial part of the record. The record can be divided into three
304 broad periods that roughly correspond to the Medieval Climate Anomaly (MCA; AD 800-
305 1250), the Little Ice Age (AD 1400-1850) and the industrial period (1850-today) (Fig. 8). The
306 MCA period shows a slight increasing trend of aridity until a maximum at *ca.* $-1.7\pm 0.2\text{‰}$ in
307 the 12th century. The transition into the LIA is characterized by decreasing isotopic values
308 from AD 1200 to AD 1500, to reach a minimum value of *ca.* $-2.6\pm 0.3\text{‰}$ from *ca.* AD 1500
309 to AD 1800, corresponding to the coolest period of the LIA. An abrupt increasing trend
310 started in the late 18th century and has continued through to the present day, which has clearly
311 the most positive isotopic values ($-1\pm 0.15\text{‰}$) of the last 1600 years. The values of the modern
312 period are significantly enriched compared to the LIA, the MCA, or the whole preindustrial
313 period (t-test, 0.01 significance level) (Fig. 8, Fig. 9). The difference between MCA and LIA
314 values is also statistically significant (t-test, 0.01 significance level).

315 The isotopic pattern in Senegal is strikingly similar to the pattern of temperature change
316 in the northern hemisphere (Fig. 10). We thus tested the correlation of the shell isotopic
317 record with a global temperature reconstruction (Pages2k 2013) and two north hemisphere
318 reconstructions (Moberg et al. 2005; Mann et al. 2009). To account for the degrees of freedom
319 of our dataset (163) and the chronology uncertainty, we calculated the Pearson correlation
320 coefficients using the 163 raw isotopic values coupled to the temperature reconstruction
321 values taken at the dates of the Monte Carlo chronologies, which provides an ensemble of
322 correlation coefficients with each temperature reconstruction. The average correlation
323 coefficients obtained with Pages2k, Moberg and Mann's reconstructions are $R=0.28$

324 (p=0.019), R=0.27 (p=0.022), and R=0.20 (p=0.092) respectively. The correlation between
325 the Saloum isotopic record and temperature reconstruction is thus statistically significant at
326 the 0.05 significance level for Pages2k global reconstruction and Moberg north hemisphere
327 reconstruction.

328 We compared the monthly resolved isotopic signals of a LIA shell from Diofandor and a
329 modern shell collected live in the same location. Both shells exhibit a perfectly regular
330 seasonal $\delta^{18}\text{O}$ signal, recording 4 monsoon seasons for the modern shell and 5 monsoon
331 seasons in the fossil shell (Fig. 11). The LIA shell $\delta^{18}\text{O}$ values are significantly more negative
332 than modern shell values by 1.18‰ on average (t-test, p-value<0.001), in agreement with
333 centennial scale variations. Rainy season conditions in the modern shell correspond to average
334 conditions in the fossil shells. Growth slowdowns observed in shell growth lines were similar
335 in the modern and the fossil shell. They occurred in both cases during the rainy season and
336 were similar in length. These observations combined with the consistency of high resolution
337 and bulk isotopic values confirm (1) that fossil shells are well preserved, (2) that growth was
338 continuous year-round and not significantly different in fossil shells, including those that lived
339 in the most different conditions from today. The amplitude of seasonal variations was 40%
340 larger in the LIA shell compared to the modern shell.

341

342 **4. Discussion**

343

344 4.1. Interpretation of the isotopic record

345

346 Since *S. senilis* aragonite was showed to precipitates in isotopic equilibrium with the
347 ambient water (Fig. 5), shell $\delta^{18}\text{O}$ values are determined by water temperature and $\delta^{18}\text{O}_w$ only
348 (Grossman and Ku 1986). We find that temperature cannot account for the shell $\delta^{18}\text{O}$

349 centennial variability, as shown by the sign and amplitude of the centennial isotopic signal: if
350 temperature were here the primary influence, the $\sim 1.5\text{‰}$ increase observed in aragonite $\delta^{18}\text{O}$
351 in the past 200 years would represent a 6.5°C *cooling*, which is unrealistic and opposed to
352 observations (Kuhnert and Mulitza 2011; Nicholson, 2013) (Fig. 10b). Hence, our signal must
353 result from centennial changes in the estuary $\delta^{18}\text{O}_w$. $\delta^{18}\text{O}_w$ in the Saloum Delta is determined
354 by 1) the relative contribution of sea water ($\delta^{18}\text{O}_w \sim 1\text{‰}$ in this area) *versus* monsoon rainfall
355 water ($\delta^{18}\text{O} \sim -4.9 \pm 0.4\text{‰}$), and 2) evaporation (Fig. 6). We showed that the relative
356 contribution of both end members was not significantly affected through time by changes in
357 the sea level or in the estuary connection to the sea (see section 2.5). On the other hand, the
358 strong increasing trend observed in the Saloum salinity from 1960 to 1990 shows the
359 sensitivity of water chemistry to changes in the P-E budget (Pagès and Citeau, 1990; Savenije
360 and Pagès, 1992; Mikhailov and Isupova, 2008). The estuary $\delta^{18}\text{O}_w$ could additionally be
361 affected by changes in the rain water isotopic composition, which could result from a change
362 in the moisture source or by a change in rainfall intensity (amount effect). Isotopic changes
363 due to the rainfall amount effect would strengthen this pattern. A change in the moisture
364 source cannot be ruled out, but we don't expect it to have a major quantitative effect since it
365 would only affect one of both end members. Therefore, we suggest that $\delta^{18}\text{O}$ variations in the
366 Saloum water and shells reflect changes in the P-E budget, with more positive (negative)
367 values indicating arid (humid) conditions.

368 We discuss now whether P-E changes observed in the past 1600 years are primarily due
369 to changes in evaporation or in the precipitation regime. Before AD 1850, the magnitude of
370 temperature changes in Senegal sea water (Kuhnert and Mulitza 2011) and in global
371 reconstructions (Moberg et al. 2005; Mann et al. 2009; Pages2k 2013) seems too small
372 ($< 0.5^\circ\text{C}$) to induce significant changes in evaporation (Fig. 10), which suggests that the pre-
373 industrial P-E centennial variability is likely primarily related to precipitation changes. Since

374 AD 1880, a warming of $\sim 1^{\circ}\text{C}$ has been measured in the ocean (ERSST 3b) (Smith et al. 2008)
375 and $\sim 1\text{-}2^{\circ}\text{C}$ on the continent (Nicholson 2013) that must have induced an increased
376 evaporation. An important indication, however, is given by the African Southwesterly Index
377 (ASWI), reconstructed from historical wind direction measurements, which indicates a clear
378 decreasing trend of rainfall in western Sahel since at least AD 1840 (Gallego et al. 2015). This
379 rainfall record, in agreement with our data, suggests that the aridification trend observed in
380 our isotopic record for the past 200 years is not only due to an increased evaporation but
381 largely to a rainfall deficit. Finally, more intense monsoon precipitation during the LIA is
382 indicated in the LIA shell isotopic record by $\sim 40\%$ larger seasonal peaks compared to the
383 modern shell (Fig. 11). During the monsoon season, calculated water isotopic values reached
384 $\sim -2.8\text{‰}$ (-3.1‰ when considering 1.5°C lower temperature in the LIA) which corresponds to
385 a salinity of ~ 13 psu (11 psu considering LIA temperature change). Such a low value close to
386 the estuary mouth requires a strong freshwater flow that is never observed today.

387 In conclusion, the Saloum isotopic variations reflect hydroclimate changes primarily due
388 to changes in local precipitation, which come almost entirely from the West African Monsoon
389 system. This means that (1) the AD 1500-1800 period in LIA was the most humid period in
390 the past 1600 years, (2) an abrupt aridification occurred in the past 200 years, (3) in the
391 context of the past 1600 years, drought conditions in western Sahel have emerged in the 20th
392 century from the range of natural climate variability.

393

394 4.2. The regional context

395

396 Records of hydroclimate in the past 1600 years are extremely scarce in the Sahel region.
397 Some documentary and archaeological evidence indicate moister conditions in the medieval
398 period. In Mauritania, the Ghana empire thrived in the 11th century suggesting a more

399 prosperous agriculture (El-Bekri 1859). Awdaghust was an active trade city in the 8th to 14th
400 century in what is today an extreme desert (Robert 1970). In the 16th century, Portuguese
401 explorer Fernandes reports that farmers did two crop yields per year in the Saloum region
402 while rainfall today only allows for one (Fernandes et al. 1951; Daveau 1969). Documentary
403 sources were used to reconstruct rainfall variability in Africa since AD 1800 (Nicholson et al.
404 2012). These sources indicate a very strong drought in the Sahel from AD 1800 to AD 1850,
405 also supported by evidence of a drop in the Lake Chad level approximately at this time
406 (Maley 1981). Considering the response time of the Saloum mean salinity, the drought
407 reported in historical documents is consistent with the onset of the abrupt isotopic trend
408 recorded by shells in the Senegal *ca.* AD 1800. West of Lake Chad, dust and ostracod $\delta^{18}\text{O}$
409 records were obtained from two oasis (Street-Perrott et al., 2000; Cockerton et al., 2014).
410 While dust flux increased through the past 2000 years possibly in relation to the changing
411 volume of Lake Chad and to human activities, ostracod $\delta^{18}\text{O}$ show slightly more negative
412 values in the LIA than in the MCA, suggesting more humid conditions in the LIA in
413 agreement with our record. No recent increase of $\delta^{18}\text{O}$ values was observed in these sites,
414 either because of different local conditions or because of a truncation of the core top.

415 A continuous record of WAM variability was provided by calcite $\delta^{18}\text{O}$ measured in Lake
416 Bosumtwi, Ghana (Shanahan et al. 2009). Interestingly, the Saloum hydrological changes are
417 very different from those recorded at Lake Bosumtwi (Fig. 10), which is located in the
418 convective zone of the monsoon. On the other hand, the Saloum isotopic record shows
419 similarity over the past 700 years with the first EOF of eastern Africa Lake records (Tierney
420 et al. 2013) and with a record of past precipitation located at the same latitude in the Gulf of
421 Aden (Tierney et al. 2015) (Fig. 10). These results suggest that the spatial coherence of
422 rainfall across the Sahel, as well as the weak correlation between the Sahel and the Gulf of

423 Guinea observed in the instrumental data at the decadal scale (Fig. 1) applies also at the
424 centennial time scale.

425

426 4.3. Sahel rainfall and global climate change

427

428 Explanations of millennial to centennial scale rainfall variability in the tropics have been
429 often related to latitudinal shifts of the mean position of the ITCZ in response to inter-
430 hemispheric temperature contrasts (Street-Perrott and Perrot, 1990; Mulitza and Rühlemann,
431 2000; Haug et al. 2001; Yancheva et al., 2007). Following this model, and according to
432 paleoclimate records from the Pacific (Sachs et al. 2009), South America (Haug et al. 2001,
433 Bird et al., 2011), Asia (Yan et al. 2015), and East Africa (Brown and Johnson 2005), the LIA
434 is characterized by a relatively southern position of the July ITCZ (Fig. 10), which is in
435 apparent contradiction with increased humidity over the Sahel during this period. The
436 centennial variability of conditions in the Sahel must therefore be governed by processes more
437 complex than the simple shift of the rain belt driven by the position of the July ITCZ.

438 Sea-surface temperature and the Atlantic multidecadal oscillation have been identified in
439 the instrumental record as one of the main drivers of Sahel rainfall (Folland et al. 1986;
440 Giannini et al. 2003; Zhang and Delworth 2006; Mohino et al. 2011). During positive AMO
441 phases, increased summer precipitation is observed in the Sahel in response to a strengthening
442 of the Sahara heat convection and increased moisture flux from the Mediterranean Sea
443 (Martin and Thorncroft 2014). However, past temperature reconstructions suggest that the
444 LIA period corresponds to a negative AMO-like phase (Mann et al. 2009) which, if the same
445 type of mechanism can be invoked at a longer time-scale, would therefore tend to induce dry
446 conditions in the Sahel, in opposition with our observations. Our multicentennial perspective
447 shows thus that western Sahel hydroclimate centennial variability responds to remote forcings

448 through processes that are different from those identified at multidecadal timescales (AMO)
449 or millennial time scales (ITCZ shifts).

450 The most prominent characteristics of the Saloum hydroclimate record is the significant
451 correlation with global or hemispheric temperature reconstructions, while no correlation is
452 observed with local SSTs (Fig. 10), suggesting that low frequency variability of Sahel rainfall
453 primarily depends on changes in the global atmospheric circulation. Although determining the
454 mechanisms responsible for the observed negative relationship between western Sahel rainfall
455 and global temperature is beyond the scope of this study, our results bring new constraints to
456 understand the processes that affect Sahel rainfall in the context of the on-going long-term
457 global warming trend.

458 First, the facts that (1) over the past 200 years, the strong aridification in the Western
459 Sahel coincides with rapid global warming, and (2) that drought conditions in western Sahel
460 are today beyond the range of the pre-industrial climatic variability of the past 1600 years,
461 point to an anthropogenic forcing of the Sahel drying trend. The last 200 years have also seen
462 the development and expansion of commercial agriculture in the Sahel, corresponding to an
463 increase of Saharan aeolian dust flux (Fig. 10) as a result of an increased erosion due to
464 increased agricultural land use (Mulitza et al. 2010). This human-induced change in
465 vegetation cover is likely to have contributed to the decrease in rainfall of the past 200 years
466 through the positive local vegetation-precipitation feedback. This local effect was probably
467 combined with the global effect of anthropogenic warming that modified the tropical
468 atmospheric circulation.

469 Second, the negative relationship between western Sahel rainfall and global temperature
470 revealed by our reconstruction is in agreement with model and instrumental data analyses that
471 showed a decreasing trend of rainfall in the margin of the convective regions (Allan and
472 Soden 2007) and a narrowing of the ITCZ as a result of global warming (Lau and Kim, 2015;

473 Byrne and Schneider 2016; Wodzicki and Rapp 2016; Su et al., 2017). The result is an
474 enhancement of the Precipitation-Evaporation pattern (Held and Soden, 2006). This trend has
475 been attributed to the “upped-ante mechanism” (Neelin et al. 2003), which describes changes
476 in horizontal moisture advection, and to a strengthening of the meridional gradient of moist
477 static energy with warming of the troposphere (Byrne and Schneider 2016). Our
478 reconstruction supports climate models predicting increasing drought in the Sahel with future
479 global warming (Held et al. 2005), although a majority of climate models tend now to predict
480 wetter future climate (Biasutti 2013). It also suggest that the rainfall “recovery” of the past 15
481 years in the Sahel is likely related to short-term multidecadal internal climate variability. In
482 the context of the past 1600 years, the modern conditions in western Sahel should be
483 considered as an extreme drought likely caused by human activities. Although the nature of
484 external forcings in the 21st century is different from the pre-industrial period, the long-term
485 relationship evidenced here underlines the risk of a major drought crisis in the 21st century.
486 Further paleoclimate reconstructions combined with paleoclimate modeling experiments are
487 required to confirm and understand centennial scale Sahel hydroclimate variability and
488 constrain greenhouse climate predictions in this region.

489

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695 **FIGURE CAPTIONS**

696 **Fig. 1.** (Left) Regional representativeness of summer precipitation in the Saloum study site.
697 Linear regression (color shades) of summer rainfall (JAS) decadal variability in the Saloum
698 Delta (green circle) with summer rainfall in the rest of the domain, using GPCP data
699 (Pendergrass et al. 2016). Mean summer (JAS) rainfall is shown by black contours (mm/d).
700 Yellow squares indicate the location, from west to east, of core GeoB9501 (Mulitza et al.
701 2010), Lake Bosumtwi (Shanahan et al. 2009), and core P178-15 (Tierney et al. 2015).
702 (Right) Map of the Saloum mangrove Delta showing the location of water and modern shell
703 sampling sites (orange circles) and fossil (blue circles) shell sampling.

704

705 **Fig. 2.** The Dioron Boumak shell midden. The shell midden is about 11m high, 300m long
706 and topped by a baobab forest (left). It was accumulated by ancient mollusk shell gatherers
707 during at least 600 years, and is one of the largest and best preserved archaeological shell
708 middens of the Saloum Delta. The beach in the front is exclusively composed of *S.senilis*
709 shells from the tidal erosion of the shell accumulation. (Center) Dioron Boumak stratigraphic
710 profile sampled for charcoal fragments and fossil shells. (Right) Close up view of the shell
711 midden showing the extreme density of the shell accumulation.

712

713 **Fig. 3.** Bayesian age-depth models of Dioron Boumak, Tioupane and Diofandor shell
714 middens. Models were calculated from charcoal radiocarbon dates using the Bacon R
715 software (34). Parameter values used for Bacon are indicated for each site. Two calendar
716 dates were assigned for Diofandor (gray): 1982±3 for the top, which corresponds to the site
717 abandonment estimated by local witnesses, and 1957±1 at 20 cm. This latter date is the only
718 possible calibrated date for this radiocarbon measurement (Table S1), after excluding the
719 1997-2002 interval which is inconsistent with the site abandonment date. The depth of the

720 stratigraphic layers and the number of shells analyzed are indicated along the vertical axis.
721 For Dioron Boumak, two columns are shown corresponding to two field campaigns in 2010
722 (left) and 2011 (right) (Azzoug et al. 2012b).

723

724 **Fig. 4.** *Senilia senilis* sclerochronology. A: polished cross section of a fossil shell from the
725 Saloum Delta. Dark bands can be seen in the outer layer and in the hinge (two of them are
726 indicated by black arrows), that correspond to monsoon season growth. The lifespan of the
727 shell can thus be estimated at about 8 years from the number of annual bands. Isotopic values
728 presented in this study were obtained from the analysis of aragonite powder collected in the
729 hinge along a transect that integrates most of the mollusk life. The sampling groove is here
730 indicated by the white arrow. B: Microscope view of tidal growth lines. Clusters of darker and
731 thicker tidal lines can be observed during spring tides every half lunar month (6). C: Scanning
732 electronic microscope image of a perfectly preserved aragonite cross lamellar structure
733 observed in a fossil shell hinge.

734

735 **Fig. 5.** Environmental monitoring and isotopic calibration in Toubakouta. (A) High resolution
736 profiles of aragonite $\delta^{18}\text{O}$ (reversed scale) measured along five modern shells of *S. senilis*
737 collected live in Toubakouta (three in 2011 and two in 2013). The chronology was determined
738 using tidal fortnightly growth lines and assigning minimum values to the average date of
739 minimum water $\delta^{18}\text{O}_w$. (B) Daily sea surface temperature (SST) measured in Toubakouta
740 (thin red line), and average monthly SST annual cycle (thick red line). (C) Weekly estuary
741 water $\delta^{18}\text{O}_w$ measured in Toubakouta from 2010 to 2015. Samples of early 2013 were lost.
742 (D) Water $\delta^{18}\text{O}_w$ calculated from shell aragonite $\delta^{18}\text{O}$ (individual shells in grey, averaged
743 shell profile in green) using Grossman and Ku third equation (Grossman and Ku 1986) and
744 the mean annual cycle of SST. Calculated $\delta^{18}\text{O}_w$ is compared with measured monthly $\delta^{18}\text{O}_w$

745 (black dotted line). (E) Monthly rainfall (mm) measured in Toubakouta. Light blue vertical
746 bands indicate the rainy seasons.

747

748 **Fig. 6.** Relationship between water isotopic ratio ($\delta^{18}\text{O}_w$) and salinity in the Saloum estuary.

749 Water samples were collected across the estuary from Dionewar to Bombugar Malik (Fig. 1)

750 at different seasons between 2011 and 2016. The least square linear regression is shown as

751 well as the determination coefficient R^2 . Sea water (~35 psu) corresponds to an isotopic ratio

752 of ~1‰. Higher values are observed in the dry season and indicate evaporated sea water,

753 while lower values correspond to a mix between sea water and rain water.

754

755 **Fig. 7.** Gradients of shell isotopic values across the Saloum Delta. Modern shell $\delta^{18}\text{O}$ (blue

756 triangles) and $\delta^{13}\text{C}$ values (red diamonds) from the 5 collection sites *versus* their distance to

757 the open ocean. The black curve shows the decreasing trend of $\delta^{13}\text{C}$ from the sea towards the

758 land, as a result of the increasing contribution of mangrove carbon. Fossil shell $\delta^{13}\text{C}$ (open red

759 squares) *versus* the distance to the open ocean. Distances were estimated from satellite

760 pictures.

761

762 **Fig. 8.** Variability of shell $\delta^{18}\text{O}$ values in the past 1600 years in the Saloum Delta. (a) Isotopic

763 values of individual shells (diamonds) are represented with 2σ analytical error. Modern shells

764 (black) were collected live in 5 locations across the estuary between 2011 and 2014. Fossil

765 shells from the stratified shell middens of Diofandor (red), Tioupane (green), Dioron Boumak

766 (blue) were dated using Bayesian age-depth models (Fig. 3). Fossil shells from sites A10,

767 A12, and A49 (pink) were directly ^{14}C dated (Table 1). Average $\delta^{18}\text{O}$ values for successive

768 stratigraphic levels are indicated by horizontal bars. The length of thick bars indicates the

769 most likely occupation period, and the thinner bars the 2σ intervals. Gray scale shading

770 represents the probability distribution of the shell $\delta^{18}\text{O}$ moving average considering
771 chronological uncertainty. The thick black lines represent the median probability and the
772 dashed lines the 95% confidence interval. (b) Histogram of the average number of shells per
773 century in the Monte Carlo sampling analysis. The average timespan integrated by a shell
774 value is approximately 5 years. (c) Box plots of isotopic values for modern shells, and fossil
775 shells of the MCA and LIA periods. MCA and LIA are here defined by the AD 800-1250 and
776 AD 1450-1850 periods respectively. Box plots represent the full range of values (whiskers)
777 excluding statistical outliers, the 25th and 75th percentiles (box edges), the median (central
778 mark), and the confidence interval (95% confidence level) of the median (box notch).

779

780 **Fig. 9.** Distributions of shell $\delta^{18}\text{O}$ values per time period. Modern= all modern values from
781 live-collected shells (N=44) (black). PI= all values with median date prior to A.D. 1850
782 (N=109) (green). LIA= all values with median dates between 1450 and 1850 (N=20) (red).
783 MCA= all values with median dates between 800 and 1250 (N=36) (blue). For each shell
784 sample, isotopic values are plotted as triangles at the bottom, and the corresponding normal
785 Kernel probability density curves are plotted (n=50, bandwidth=0.2). Mean values and the
786 95% confidence intervals of the estimated mean values are represented by dashed lines and
787 shaded areas.

788

789 **Fig. 10.** Comparison of the Saloum hydroclimate record with regional and global climate
790 records. The time scale is in calendar years AD. (a) Temperature variability: northern
791 Hemisphere temperature reconstructed by Mann et al. (2009) and smoothed with a 30-yr
792 triangular filter (orange curve and shading), low frequency northern hemisphere temperature
793 variability calculated by Moberg et al. (2005) (red curve and shading), global temperature
794 anomaly (SD unit) as the area-weighted average of continent scale reconstructions by Pages-

795 2k consortium (2013) (pink). Vertical red and blue shading was calculated from the
796 temperature anomaly in Moberg et al.'s (2005) record. MCA and LIA periods are indicated
797 above. (b) Records from marine sediment core GeoB9501: SST calculated from foraminifera
798 Mg/Ca (Kuhnert and Mulitza 2011) (green), and Aeolian dust flux (Mulitza et al. 2010)
799 (brown). (c) African Southwesterly Index (ASWI) of JAS, from 1840 to 2013, with 30-yr
800 LOESS smoothed curve (Gallego et al. 2015). (d) Shell $\delta^{18}\text{O}$ density of probability (gray
801 shading), median probability (black line) and confidence interval (dashed lines) in the Saloum
802 estuary as an indicator of the precipitation-evaporation budget in the Western Sahel (this
803 study). (e) Leaf wax hydrogen isotopic ratios ($\delta\text{D}_{\text{wax}}$) in the marine sediment core P178-15
804 from the Gulf of Aden (Tierney et al. 2015). The error bar is the ± 1 sigma analytical precision.
805 (f) Calcite $\delta^{18}\text{O}$ in Lake Bosumtwi, Ghana, and 30-yr triangular filter smooth (Shanahan et al.
806 2009). (g) Ti content in ODP1002 sediment core indicating continental freshwater runoff into
807 the Cariaco Basin, off Venezuela (Haug et al. 2001), and 10-point moving average curve
808 (dark blue). (h) Calcite $\delta^{18}\text{O}$ in Lake Pomacocha, Central Peruvian Andes (Bird et al. 2011),
809 and 30-yr moving average (purple).

810

811 **Fig. 11.** Monthly resolved $\delta^{18}\text{O}$ records of a modern and a fossil shell from Diofandor (Fig.
812 1). Average values are indicated by gray lines. Negative values reflect lower salinity. Time
813 axes represent the shell ontogenic age.

814

Figure 1

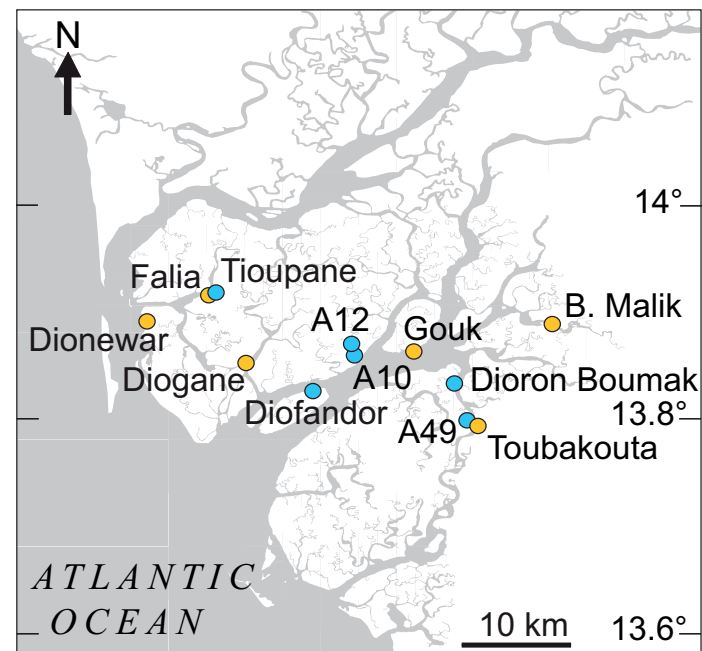
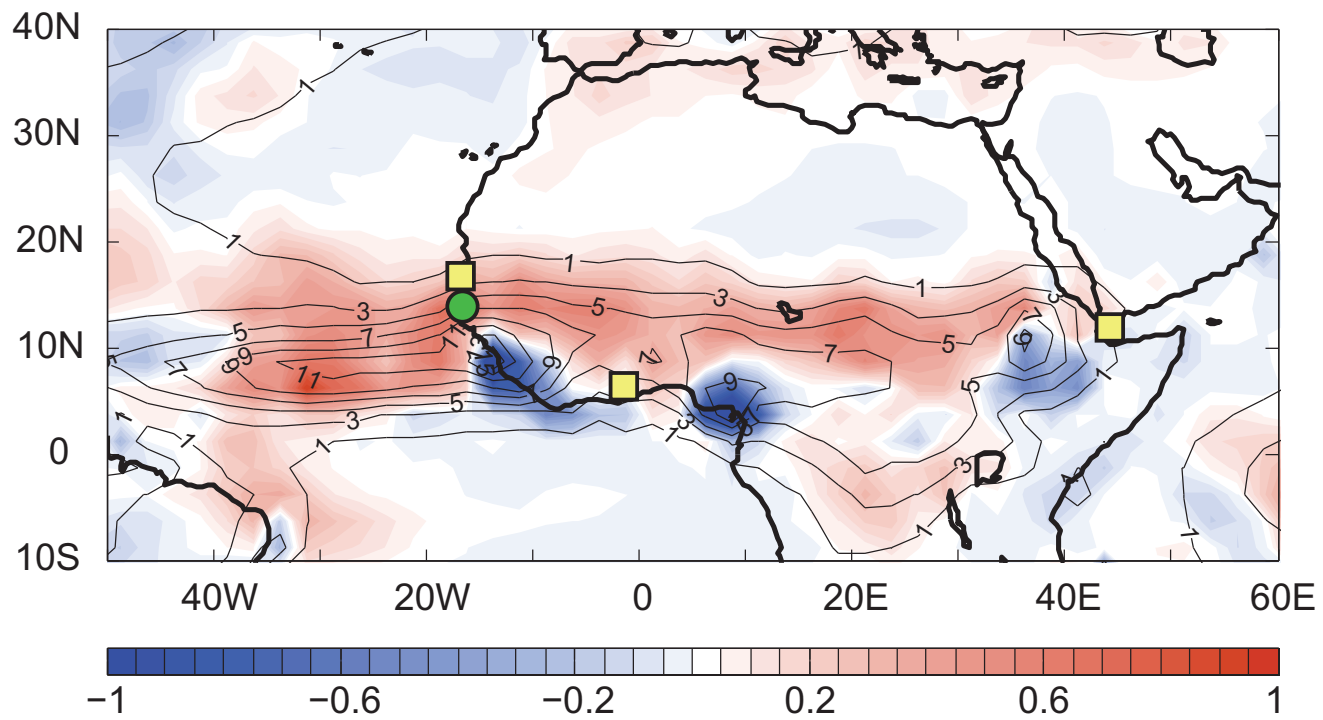




Figure 3

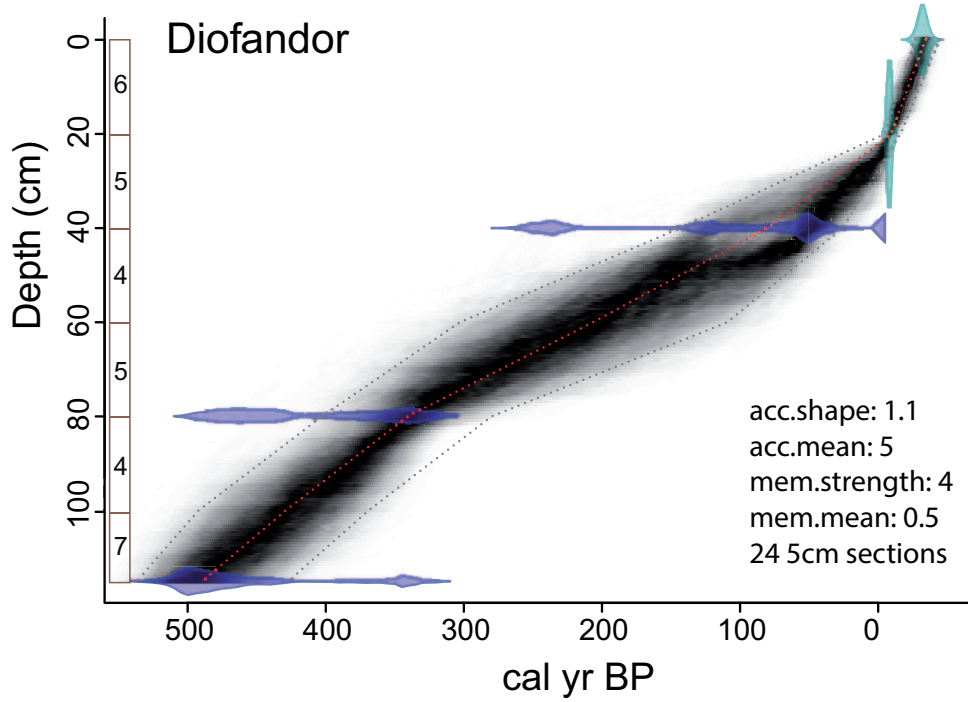
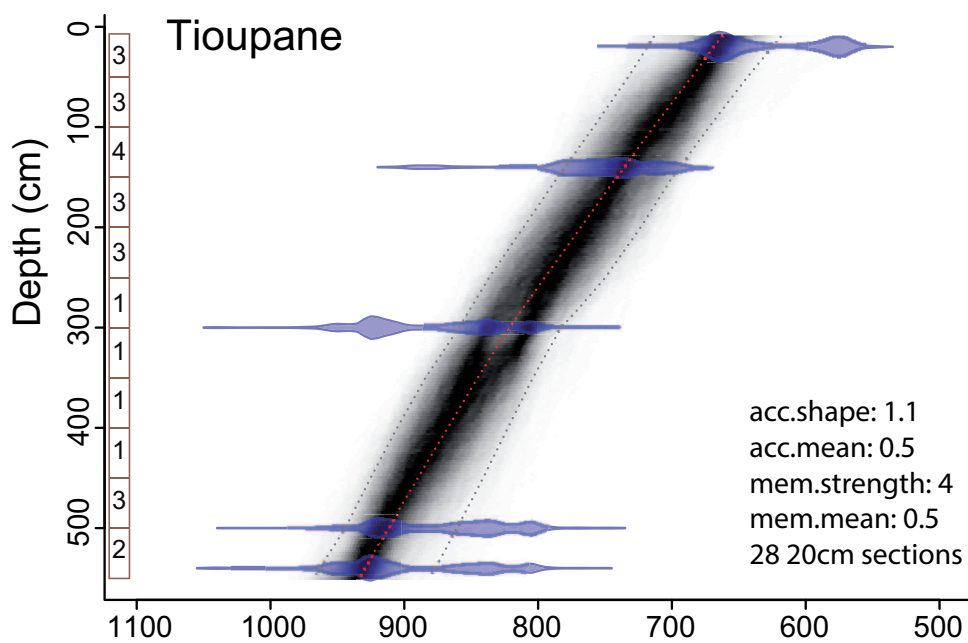
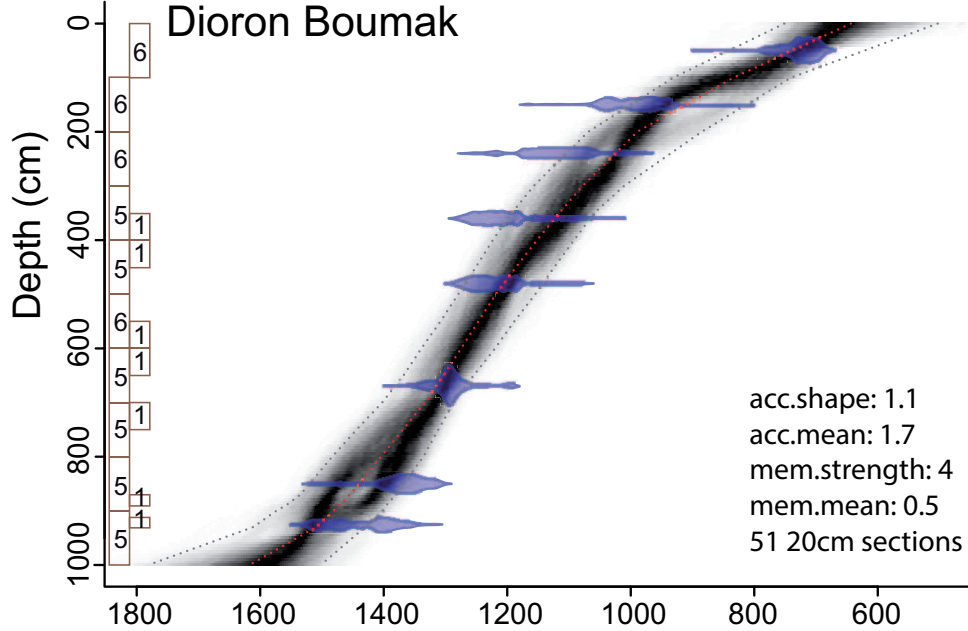
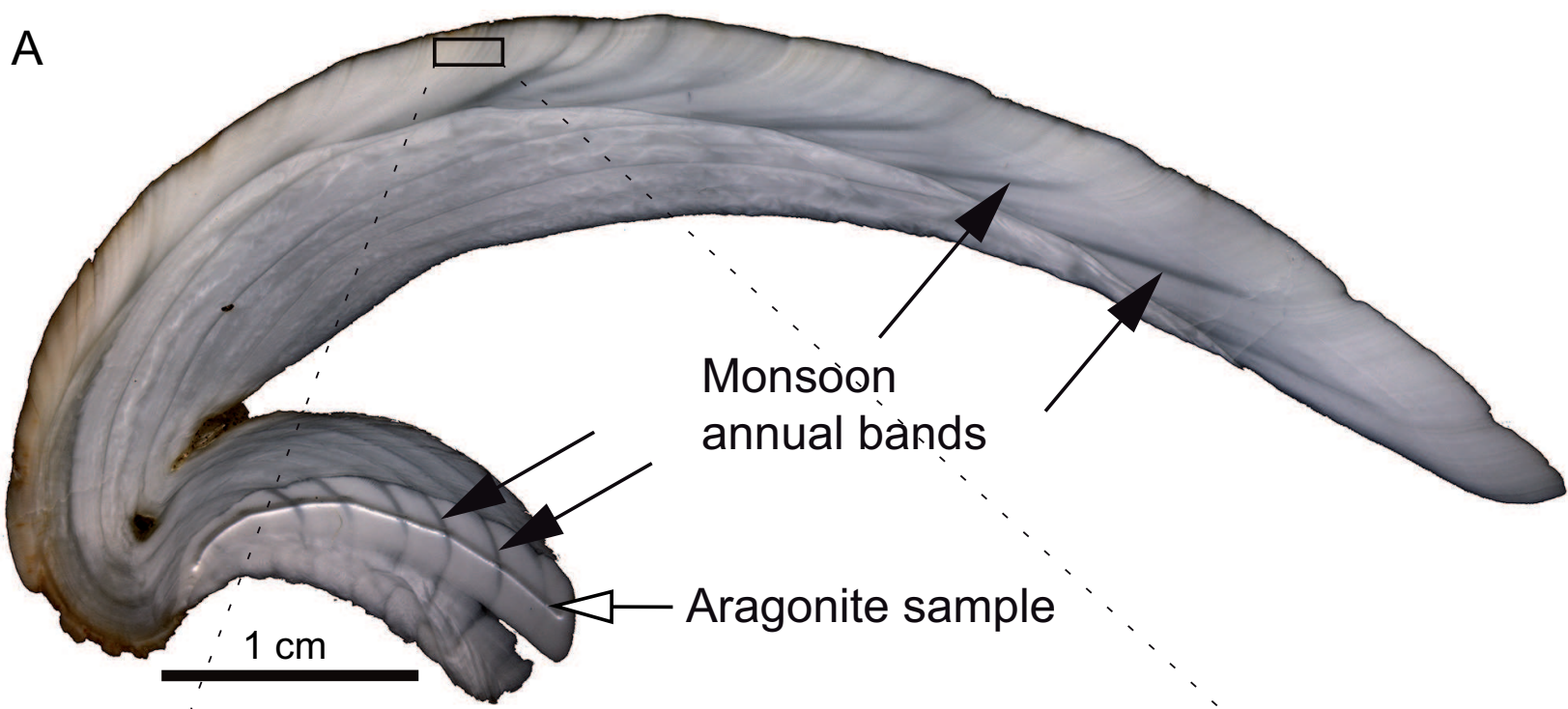
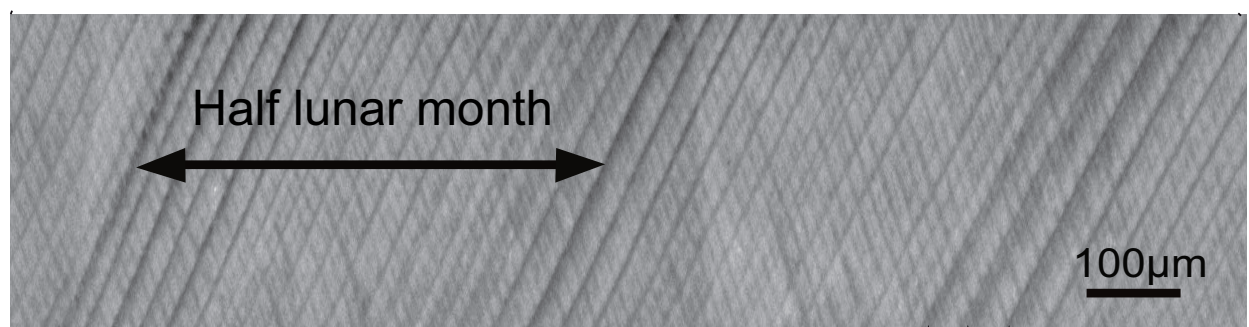


Figure 4

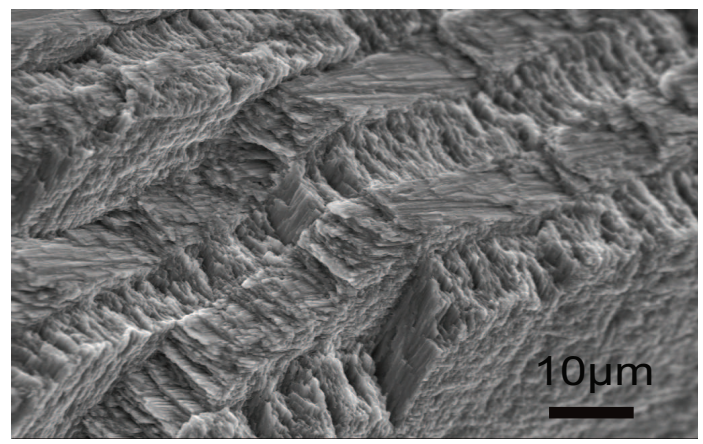
A



B



C



Daily tidal lines

Detailed description: Three vertical arrows pointing upwards towards the surface texture in panel C, with the label 'Daily tidal lines' below them.

Figure 5

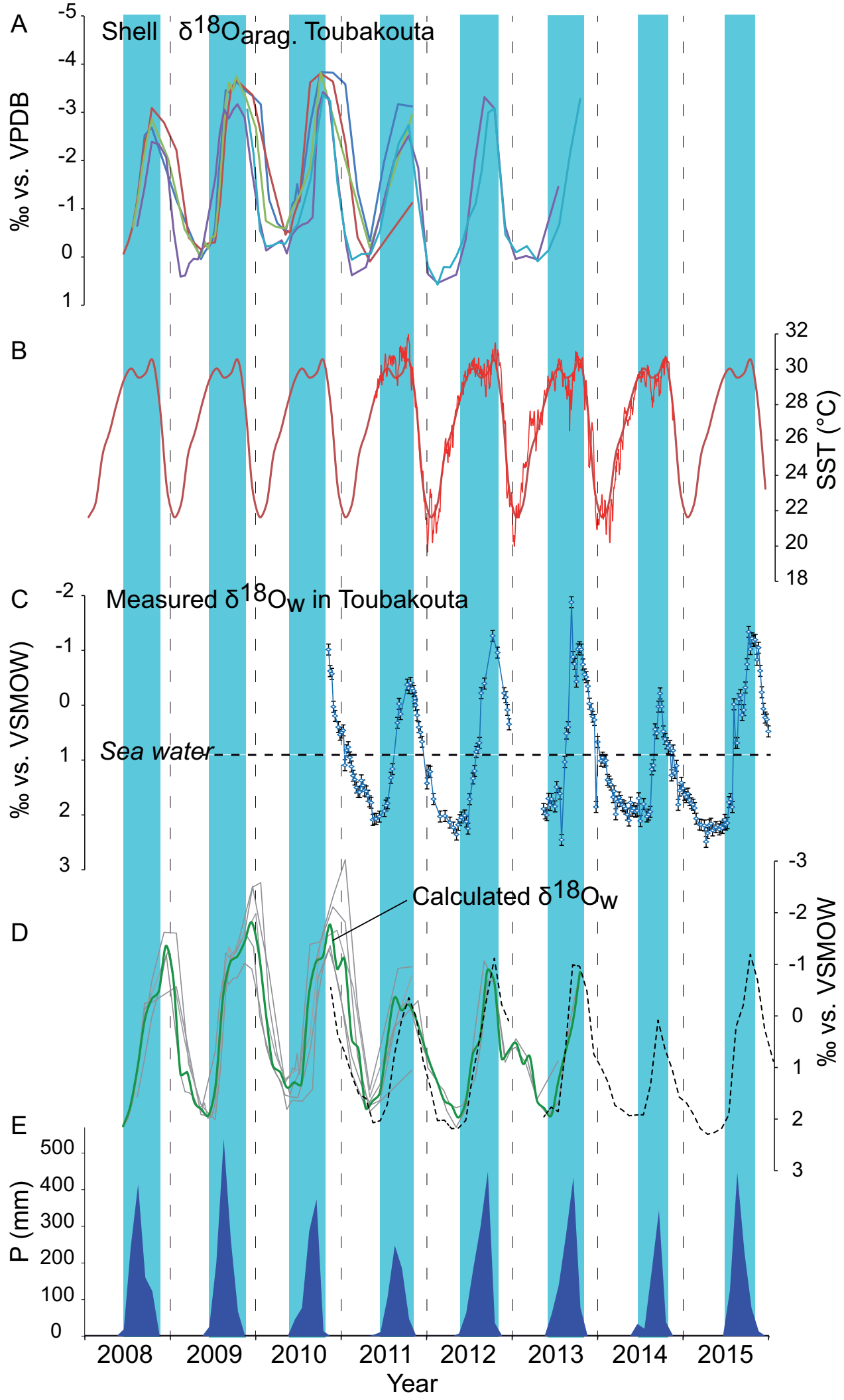


Figure 6

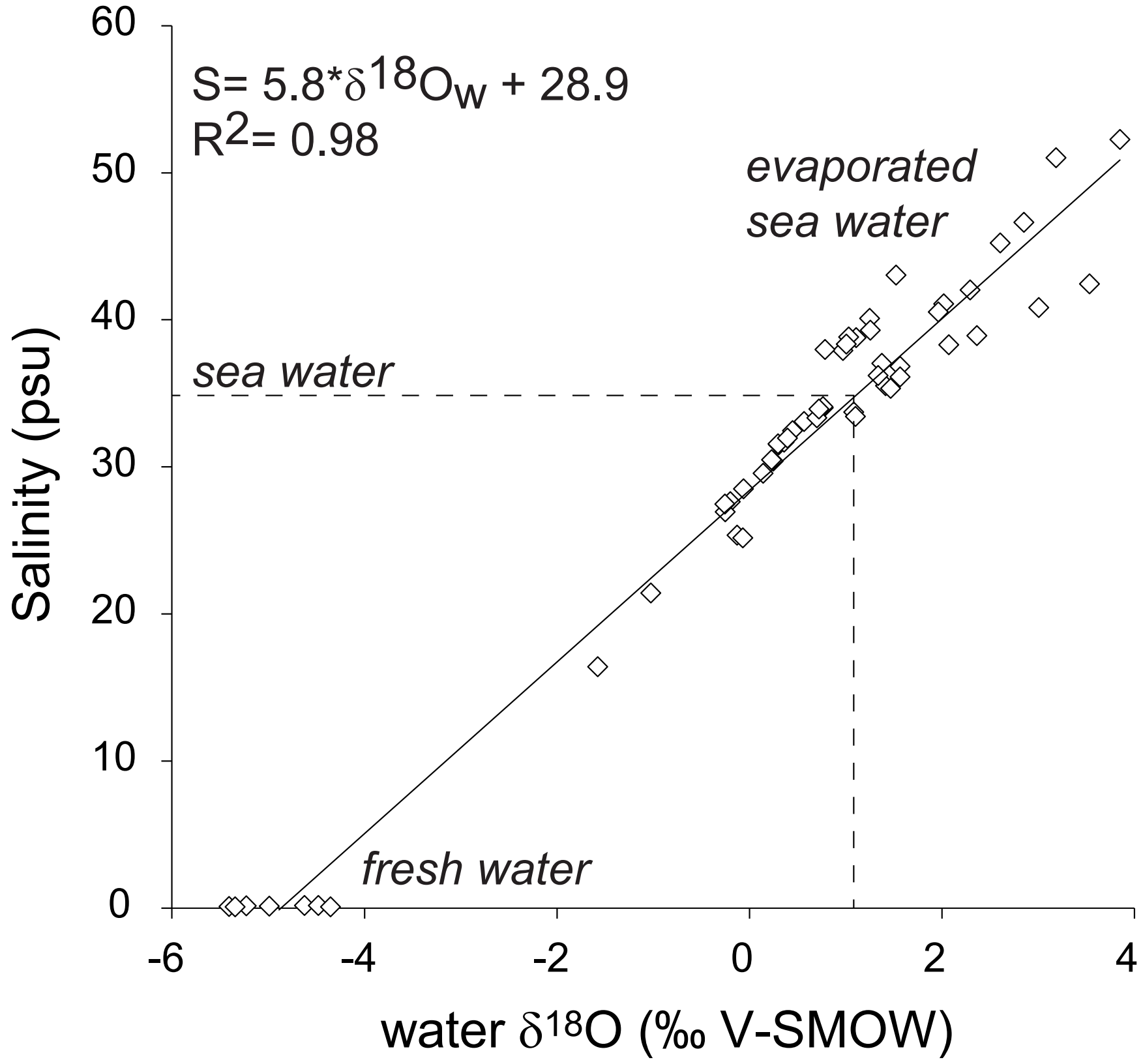


Figure 8

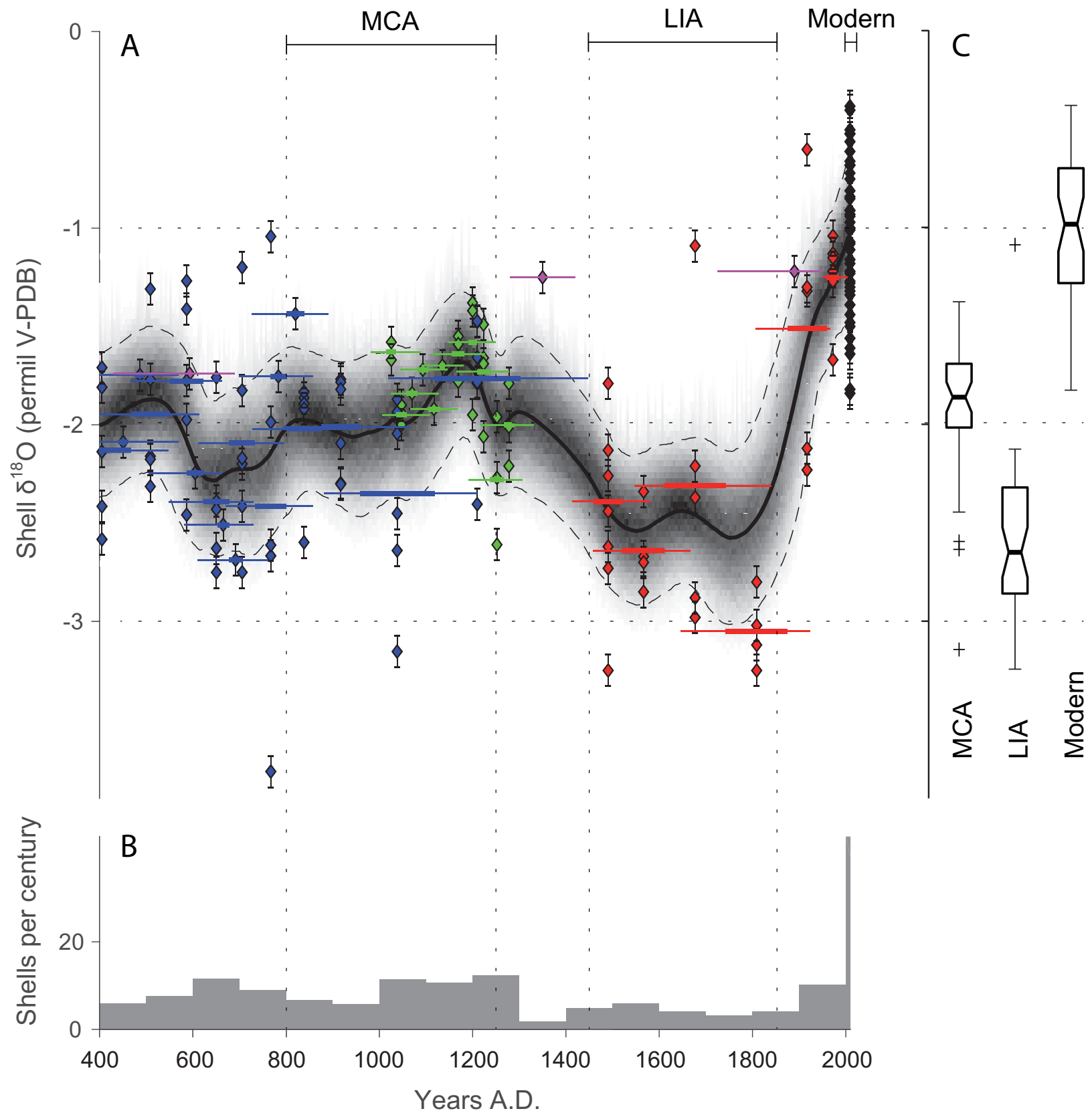


Figure 9

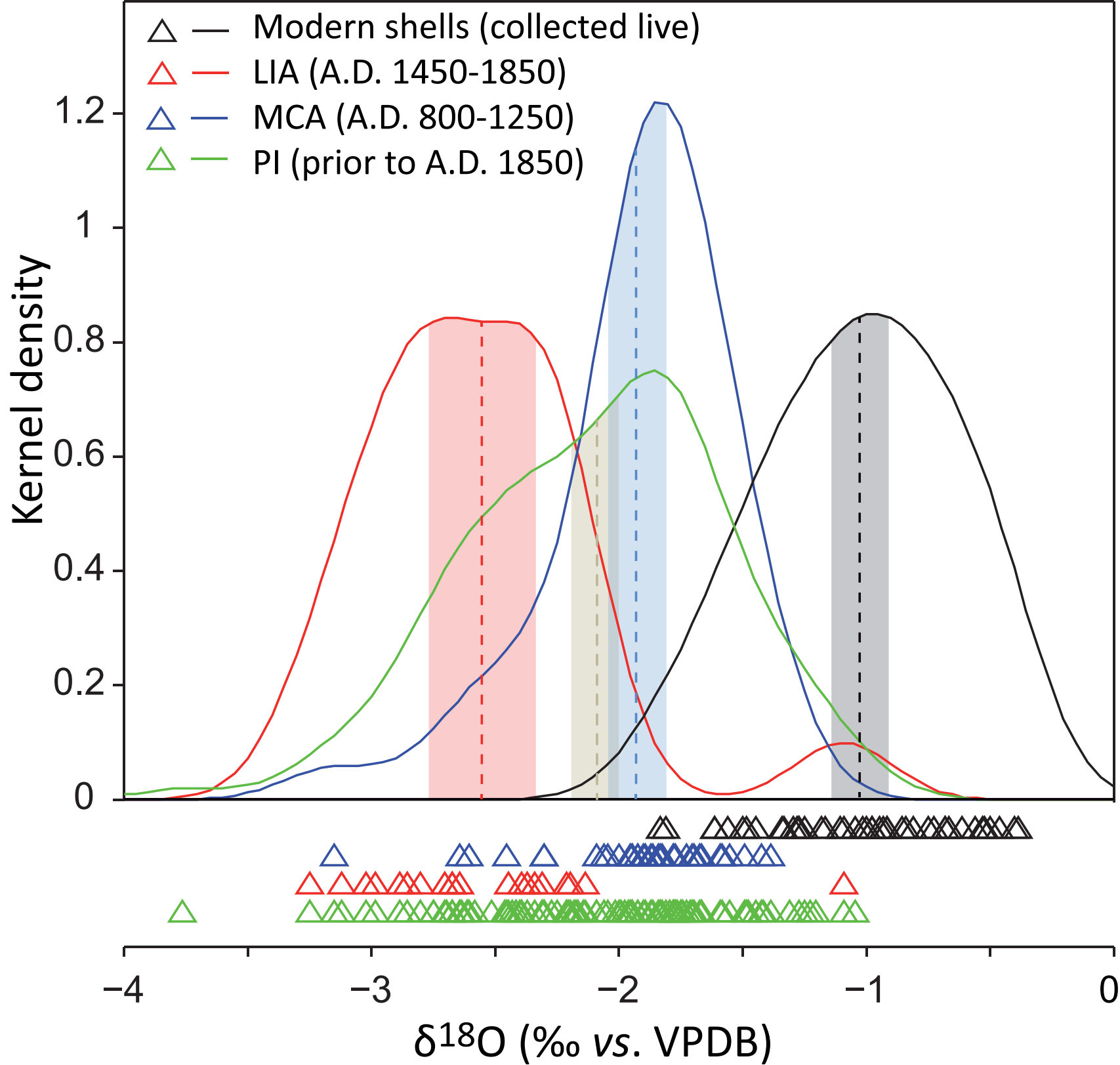


Figure 10

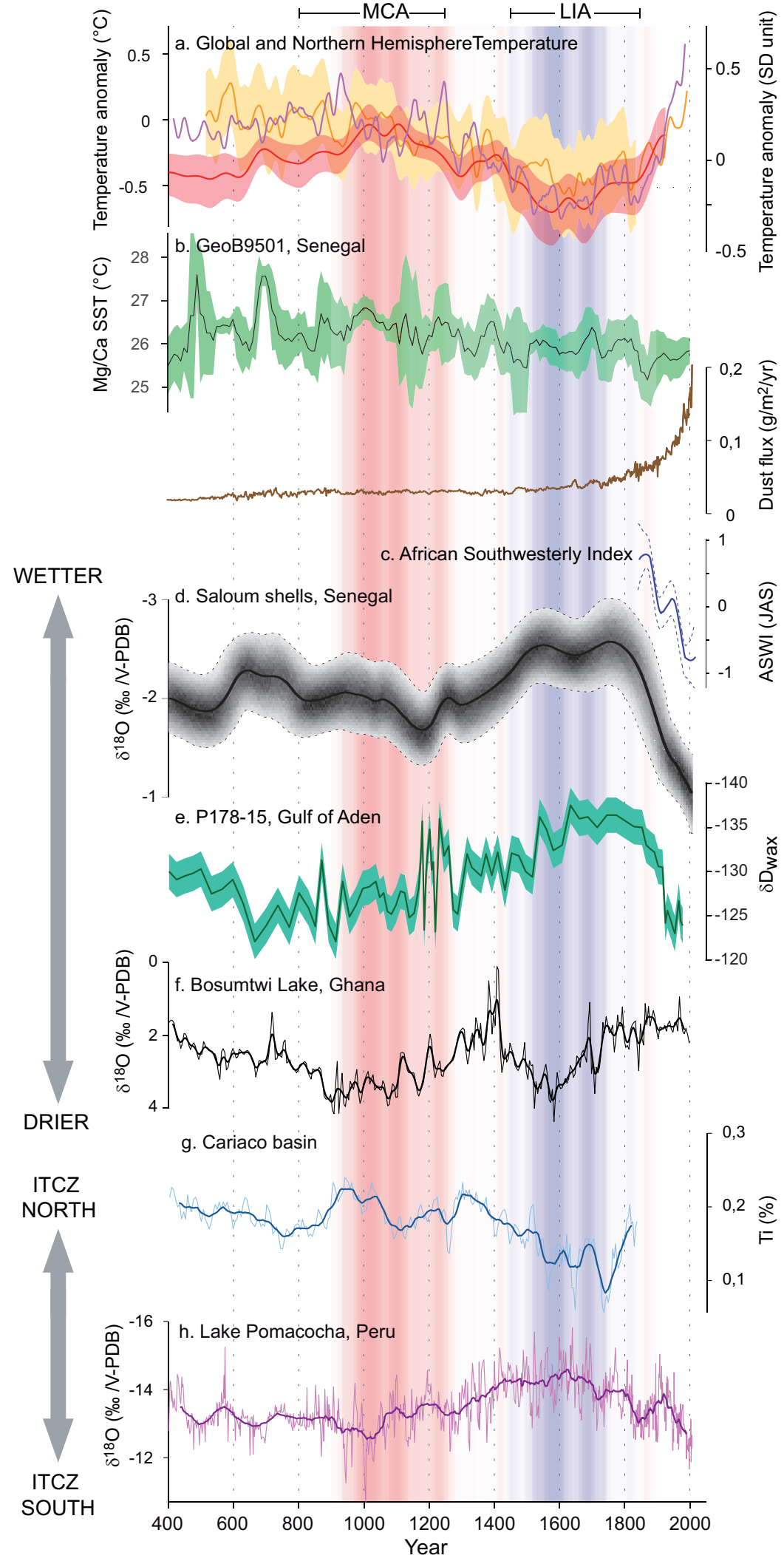


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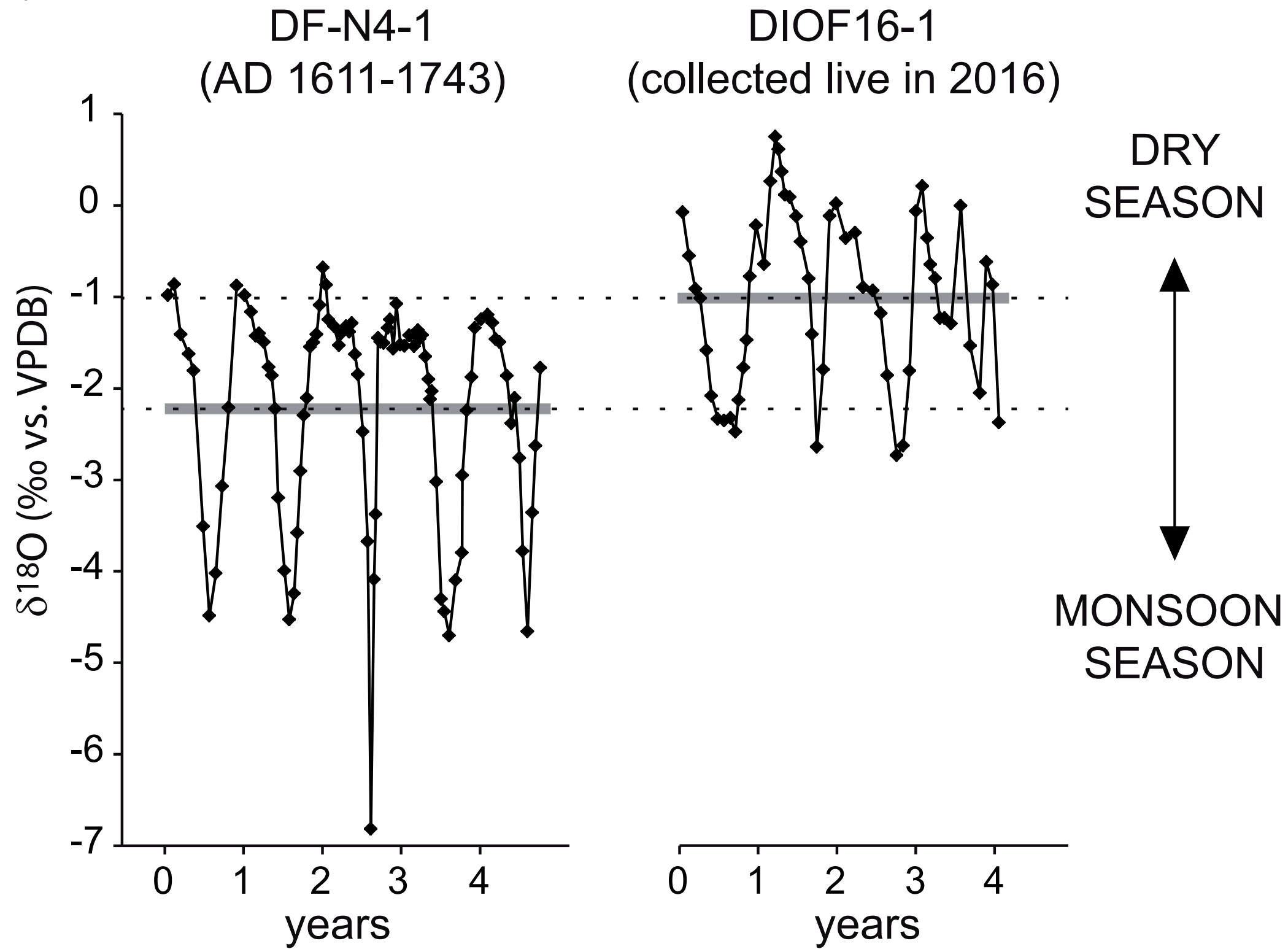


Table 1. Radiocarbon dates and calibration

Site	Material	Lab. Ref.	Depth (cm)	$\delta^{13}\text{C}$ (‰ VPDB)	^{14}C age / F^{14}C	Calibration dataset	%mar. carbon	Cal. median (AD)	2 σ range
Diofandour	charcoal	UBA-29457	20	-25*	1.0921 ± 0.0044	NHZ2		1957	1957-1958***
	charcoal	SacA 25602	40	-25.3	35 \pm 30	IntCal13		1893	1695-1955
	charcoal	UBA-29458	80	-25*	363 \pm 26	IntCal13		1519	1451-1633
	charcoal	SacA 25603	115	-25.9	415 \pm 30	IntCal13		1463	1429-1618
Tioupane	charcoal	SacA 25604	20	-31.4	690 \pm 35	IntCal13		1294	1262-1390
	charcoal	SacA37940	140	-26.6	840 \pm 30	IntCal13		1203	1059-1264
	charcoal	SacA25605	300	-24.1	985 \pm 30	IntCal13		1047	991-1154
	charcoal	SacA 25606	500	-24.4	970 \pm 30	IntCal13		1089	1017-1155
	charcoal	SacA 25607	540	-22.9	995 \pm 30	IntCal13		1031	986-1153
Dioron Boumak	charcoal	UBA 19943	50	-25*	817 \pm 23	IntCal13		1229	1179-1265
	charcoal	UBA 19944	150	-25*	1083 \pm 38	IntCal13		957	890-1019
	charcoal	SacA 25596	240	-24.2	1195 \pm 35	IntCal13		829	694-947
	charcoal	SacA 25597	360	-25.9	1260 \pm 30	IntCal13		736	669-862
	charcoal	SacA 25598	480	-19.8	1270 \pm 30	IntCal13		729	664-856
	charcoal	SacA 25599	670	-23	1370 \pm 30	IntCal13		655	608-688
	charcoal	SacA 25600	850	-24.8	1480 \pm 30	IntCal13		587	541-642
	charcoal	SacA 25601	925	-20.8	1530 \pm 30	IntCal13		537	432-600
A-10	shell	SacA37929	0	-7.1	340 \pm 30	Mixed **	72	1890	1725-1955
A-12	shell	SacA37930	0	-9.1	950 \pm 30	Mixed **	64	1350	1286-1415
A-49	shell	SacA37922	0	-2.1	1885 \pm 30	Mixed **	92	593	454-686

*assumed value

**mixed using IntCal13 and Marine13 with $\Delta R=0$. Marine carbon percentage was estimated from $\delta^{13}\text{C}$ value assuming a 0‰ value for the marine endmember and a -25‰ value for the terrestrial carbon endmember.

***the 1997-2002 interval was excluded from the calibration because the site was abandoned in 1982.