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1 **Last Interglacial sea surface warming during the sea-level highstand in the Canary**
2 **Islands : Implications for the Canary Current and the upwelling off African coast**

3
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21
22 Keywords : LIG; MIS5e; Sea-level; Sea surface temperature; Stable isotopes; Molluscs;
23 Canary current; Upwelling; North Atlantic Gyre ; Climate change

24

25 **Abstract** – The Canary Islands, east of the North Atlantic Ocean (27°N-29°N), are under the
26 influence of the Canary Current, the descending branch of the North Atlantic Gyre, which is
27 modulated by coastal upwelling off North-West Africa. They constitute strategic sites for
28 palaeoclimatic reconstructions, especially for the Last Interglacial (LIG, 129 to 116 ky BP)
29 estimated to be warmer than present. Seventy-four carbon and oxygen isotope bulk analyses
30 and time series measurements were performed on 32 aragonitic mollusc shells from the LIG
31 marine deposits on Lanzarote and Fuerteventura islands during a period of sea-level highstand
32 that we estimated to occur between ≈ 125 and 119-116 ky BP. Our SST calculations, inferred
33 from shell $\delta^{18}\text{O}$ values using available isotopic fractionation equations, provide a seasonal
34 SST amplitude ranging from 3.5°C to 6.0°C, in agreement with the modern seasonal
35 amplitude, along with a mean SST comprised between $20.4 \pm 1.3^\circ\text{C}$ and $22.2 \pm 1.2^\circ\text{C}$. With
36 respect to the pre-industrial times, we deduce a positive SST anomaly in the range of
37 $+1.0 \pm 1.4^\circ\text{C}$ to $+2.8 \pm 1.3^\circ\text{C}$, consistent with the presence of the species *Persististrombus*
38 *latus*, typical of warm SSTs. Although this finding does not match with the zonal negative
39 anomaly of a reconstructed SST at low latitudes of the North Atlantic, it is nevertheless
40 corroborated by other climate reconstructions in the northeastern tropical Atlantic region. We
41 attribute this trend to an excess of summer insolation during the LIG which warmed the
42 Canary Current, enhanced by a weakening of the North African upwelling. The entire North
43 Atlantic Gyre was probably warmer during the LIG.

44

45 Abbreviations used:

46 LIG : Last Interglacial period; GMST : Global Mean Surface Temperature; SST : Sea Surface
47 Temperature; SL : Sea-level.

48

49 1. Introduction

50

51 Since pre-industrial times (middle of the 19th century) the global mean surface
52 temperature (GMST) has increased by about 1°C due to the emission of anthropogenic
53 greenhouse gases (IPCC, 2013). As these emissions will continue during the 21th century,
54 global warming of between +2°C and +5°C is projected by the end of the century (IPCC,
55 2013). In such scenarios of large climate change, simulations of global warming must be
56 performed with the highest precision possible. The confidence in the future climate
57 simulations by models is tested through the ability of the same models to simulate past
58 climates correctly, particularly climates warmer than the present. This calls for the most
59 precise description of such past warm periods. Even though the cause of past warming differs
60 from that of the present (i.e., not due to the greenhouse effect), it is of paramount importance
61 that the state of the world in a warmer climate be accurately simulated.

62 Such findings are obtained from reconstitutions of regional and local climates, in
63 which the temperature remains one of the key parameters. In the reconstitutions three criteria
64 should be kept in mind : (1) a warmer than present context, (2) a geographical region, which,
65 as far as possible, is consistent with the globally relevant climate mechanisms, and (3) a
66 chronological framework that possesses sufficiently high precision for the data from the
67 proxies used by the models.

68 The present study focuses on the sea surface temperature reconstruction during the
69 Last Interglacial (LIG) in the Canary Islands (northeast tropical Atlantic) (Figure 1) from
70 marine sediment deposits (Figure 2). It satisfies the above criteria in the following way :

71 (1) During the Quaternary glacial/interglacial oscillations, it is known that two interglacials
72 were warmer than the present interglacial : the Marine Isotope Stage MIS11c, from \approx 245 to
73 \approx 235 thousand years before present (ka BP), and the Last Interglacial (LIG) which covers
74 approximately the time interval between \approx 129 and \approx 116 ka BP (Masson Delmotte et al.,

75 2013; Capron et al., 2017a). The LIG, i.e. the most recent, gives rise to a more detailed
76 reconstitution. Moreover, of the last five Quaternary climate cycles, it appears to be the
77 warmest interglacial (Jouzel et al., 2007; Past Interglacials Working Group of PAGES, 2016)
78 by about +2°C with respect to the present interglacial.

79 (2) Temperature reconstructions during the LIG have recently been the subject of intense
80 interest. Whereas many publications focus on the polar and mid latitude regions, temperature
81 data from the tropics, particularly seawater surface temperature (SST) data, are much scarcer.
82 The Canary Islands, located in the tropical northeast Atlantic along the west African coast, are
83 exposed mainly to the southward marine Canary Current, the descending branch of the North
84 Atlantic Gyre, and are also influenced by the cold coastal upwelling current off North-West
85 Africa (Aristegui et al., 2009). The northeast tropical Atlantic region is particularly important
86 from the point of view of ocean circulation because the North Atlantic Gyre, an essential
87 component of the global circulation, participates directly in the climate equilibrium of the
88 planet. At present, few LIG data exist for this region and those that do exist appear to be
89 contradictory, some suggesting warming and others cooling. Reconstruction of the SST in this
90 region is therefore of major importance.

91 (3) Over the last decades, the geology, sedimentology and chronology of the Canary Islands
92 have been thoroughly investigated, as well as the composition of faunal taxa found in the
93 marine fossil deposits and in paleosols (Meco and Stearns, 1981 ; Meco et al., 2002 ; Zazo et
94 al., 2002 ; Meco et al., 2006 ; Meco et al., 2011 ; Muhs et al., 2014 ; Montesinos et al., 2014 ;
95 Meco et al., 2018). This rich source of information is relevant to establishing the chronology
96 of the deposits under study.

97 In this investigation, we used the oxygen and carbon isotope compositions of
98 carbonate shells dating of the LIG in order to estimate not only the SST, but also the
99 magnitude of seasonal temperature variations and the seasonal activity of the coastal

100 upwelling. This method has not been applied in the region yet. In order to obtain qualitative
101 indications of local warming/cooling of the SST, we also reconstituted the living environment
102 of the faunal fossil record. These new data complement other SST results from the tropical
103 northeast tropical Atlantic and afford a deeper understanding of regional ocean currents: the
104 upwelling off the western African coast and the Canary Current.

105

106 2. Background

107

108 2.1. Geological setting

109

110 The Canary Islands comprise seven main volcanic islands and several islets. They are
111 located in the Eastern Atlantic Ocean between 27°N and 30°N, forming a chain that extends
112 about 450 km in latitude, with the easternmost point just over 100 km off the north-western
113 African coast (Figure 1a). The islands have a complex geological history, with volcanic
114 formations over 20 million years old (Carracedo et al., 2002), but they also include extensive
115 sedimentary deposits (Meco and Stearns, 1981). As part of this study, two specific sites with
116 sedimentary deposits were sampled in the easternmost Canary Islands, namely Fuerteventura
117 and Lanzarote. While the eastern Canary Islands experienced some uplift during the last
118 million years, discontinuous vertical movements with a reverse trend over the last hundreds of
119 thousands of years have been observed (Zazo et al., 2002).

120 In the Canary Islands, the sedimentary deposits are related to the glacial–interglacial
121 cycles and are characterized by a succession of four distinct climatic phases (Meco et al.,
122 2011): (i) calcareous dune deposits, corresponding to a cold and arid glacial climate, (ii)
123 paleosols, corresponding to a climate that became progressively warm and wet at the
124 beginning of the interglacial period, (iii) marine sediments deposited during the sea-level

125 highstand of the interglacial period, and (iv) pedogenic calcretes, marking the return to the
126 next glaciation. These different phases were identified in the Eastern Canary Islands for the
127 MIS11.3 interglacial period as well as for the MIS10–MIS9.3, MIS6–MIS5.5 (LIG) (Figure
128 2a), MIS2–MIS1 (Holocene) glacial–interglacial oscillations. In the Canary Islands, the warm
129 and wet climate of phase (ii) that generated paleosol deposits is considered to be
130 contemporaneous with the wet phase in West Africa ("African Humid Period"). The latter,
131 evaluated through many proxies in different areas of North Africa, stems from the maximum
132 summer insolation in the tropics (Rossignol-Strick, 1985; deMenocal et al., 2000).

133 LIG marine deposits are found in the Eastern Canary Islands. In Lanzarote, these
134 deposits have outcrops mainly at the localities of La Santa, Matagorda and Punta Penedo
135 (Meco et al., 2006; Muhs et al., 2014), whereas in Fuerteventura they occur at the localities of
136 Matas Blancas, Las Playitas and La Guirra (Meco et al., 2002; Zazo et al., 2002; Meco et al.,
137 2006; Montesinos et al., 2014; Meco et al., 2018) (Figure 1b). LIG marine deposits also occur
138 in the central island of the archipelago in Las Palmas, Gran Canaria (Meco et al., 2002). In the
139 present work two LIG marine deposits were studied: La Santa (Lanzarote) and Matas Blancas
140 (Fuerteventura). The La Santa site (29°06'48,43''N, 13°38'57,50''W), which is located along
141 the northwest coast of Lanzarote, is composed of fossiliferous marine sands overlying Middle
142 Pleistocene basaltic lava flows at ~8 m above the present mean sea-level (Muhs et al., 2014).
143 This sandy deposit developed in a foreshore environment. The same layer crops out at the
144 Punta Penedo site, a few kilometers northeast of La Santa and shows the most complete
145 section in the area (Figure 2a). The Matas Blancas site (28°10'26,60"N, 14°11'46,91"W) is
146 located along the eastern coast of the Jandia Isthmus, in Southern Fuerteventura. The marine
147 sedimentary deposit consists of an extensive fossiliferous marine conglomerate at -2 m to +3
148 m above the present medium sea-level, identified as a marine terrace (Meco et al., 2002). It is
149 composed of coarse basaltic fragments and fossils with a calcareous sandy matrix and

150 includes abundant specimens of *Persististrombus latus* (also named *Strombus bubonius*)
151 (Figure 2b). It originally developed in a high energetic foreshore environment, forming part of
152 a gravel berm or a beach of pebbles and sand. The deposit appears naturally cemented and can
153 be considered as a beachrock.

154

155 2.2. Modern oceanic features

156

157 The eastern part of the North Atlantic Gyre is driven by the eastward Azores Current
158 which partly feeds the Canary Current, a southward-flowing current along the Northwest
159 African coast (Figures 3a and 1a). As it passes the Canary Islands archipelago, the current
160 splits in two, with branches to the east and west of Lanzarote. The surface current is
161 associated with a strong coastal upwelling regime along the north-west African coast, caused
162 by the northeast trade winds (Figure 3b). Strengthening of the alongshore wind enhances
163 upwelling and results in lower SST over the shelf (Figure 3b). In the latitude range 22°N to
164 30°N, where the Canary Islands are located, the coastal upwelling is permanent all year
165 round, with the strongest activity occurring in summer and early fall (Pardo et al., 2011;
166 Navarro-Pérez and Barton, 2001; Mittelstaedt, 1991).

167 The recent mean annual temperature of the eastern Canary Islands coastal waters has
168 been estimated by different methods (Table 1). In the framework of this study we retain the
169 combined satellite data and *in situ* SST datasets from the eastern Canary Islands, which give a
170 sea surface temperature (SST) of 20.4°C from 2007 to 2017 (Meco et al., 2018). This finding
171 is consistent with the other satellite measurements.

172

173 3. Materials

174

175 3.1. Sample collection

176

177 Thirty-two fossil marine mollusc shells belonging to 6 different species were collected
178 in the eastern Canary Islands. The largest number of fossils was collected at La Santa site
179 (Lanzarote) with 28 fossil mollusc shells belonging to 5 different species, whereas 4 fossil
180 mollusc shells belonging to two different species were collected at Matas Blancas site
181 (Fuerteventura). The fossils are either gastropods (*Cerithium vulgatum* Bruguière, 1792;
182 *Stramonita haemastoma* Linnaeus, 1767; *Luria lurida* Linnaeus, 1758; *Persististrombus latus*,
183 Gmelin 1791, usually referred to as *Strombus bubonius* Lamarck, 1822) or bivalves (*Loripes*
184 *lacteus* Linnaeus, 1758 – *nomen nudum* ; *Cardium edule* Linnaeus, 1758, also named
185 *Cerastoderma edule* Linnaeus, 1758) (Figure 4). Quite prevalent in the LIG sample deposits,
186 they are representative of the sites, gastropods being more frequent than bivalves. They all
187 lived in nearshore environments and shallow sea beds (Cuerda, 1987; Niclès, 1950; Rolán,
188 2011; Meco, 1977; Cosel von and Gofas, 2019; Cosel von and Gofas, 2019, respectively).

189 Among the 6 studied species, one lives in warm waters: *Persististrombus latus*
190 (*Strombus bubonius*). This gastropod is currently found on the western African coasts from
191 Dakar in Senegal to Angola, including the Cape Verde Islands. It belongs to the tropical
192 Guinean/Senegalese faunal province (Muhs et al., 2014). In the Canary Islands, it is present
193 only in the form of fossil species in the marine deposits of previous interglacials (LIG and
194 MIS11) (Meco, 1977; Meco et al., 2002; Meco et al., 2007; Muhs et al., 2014). *Luria lurida*
195 and *Stramonita haemastoma* live in an environment of warm to cooler waters : *Luria lurida*
196 currently lives on the North-western African coasts from Morocco to Angola (Rolan et al.,
197 2011), whereas *Stramonita haemastoma* currently lives on the western African coasts from
198 Morocco to Angola, including the Canary Islands (Rolán, 2011). Finally, *Cardium edule*,
199 *Loripes lacteus* and *Cerithium vulgatum* live in cooler marine waters. *Cardium edule*

200 (*Cerastoderma edule*) is present in the north-western Atlantic and in the north-eastern
201 Atlantic, from Iceland and North Europe to Senegal, but rarely in the Canary Islands (Rolán,
202 2011; Nicklès, 1950). *Loripes lacteus* currently lives in the Eastern Atlantic from England and
203 the North Sea to Senegal (Rolán, 2011; Nicklès, 1950). *Cerithium vulgatum* currently lives in
204 the Atlantic from Portugal and Morocco to the Canary archipelago (Rolán, 2011). All the
205 species except *Persististrombus latus* (*Strombus bubonius*), which at present lives far south of
206 the Canary Islands, are now present at the latitude of the Canary archipelago in the eastern
207 Atlantic (Figure 5).

208 The shells were directly sampled from the sedimentary marine beds at the sites where
209 they outcrop, then labelled and put into plastic bags.

210

211 3.2. Age of the sampled deposits

212

213 The age of the marine deposits containing the mollusc shells has been estimated by
214 three different approaches : the various LIG fossil datings in the eastern Canary archipelago,
215 the sea-level history during interglacials, and the sedimentary successions during glacial-
216 interglacial cycles in the Canary Islands.

217 U-series dating of two fragments of coral *Siderastraea radians*, Pallas 1766 coming
218 from San Cristobal, Gran Canaria, and La Santa marine deposits, and which are ascribed to
219 the LIG sea-level highstand, yields the ages 120.5 ± 0.8 ka and 130.2 ± 0.8 ka, respectively
220 (Muhs et al., 2014). Five marine terraces cropping out in Lanzarote, which are considered to
221 be contemporaneous with those of La Santa on the basis of stratigraphic arguments, provided
222 U-series ages from the aragonitic shells of the gastropod *Patella* of 121 ± 17 ka BP (n = 11)
223 (Zazo et al., 2002). At the Matas Blancas site, the gastropods *Persististrombus latus*
224 (*Strombus bubonius*) has yielded numerous radiometric ages based on the U/Th method,

225 namely 106 ± 7 and 112 ± 7 ka BP according to Meco et al. (1992); 104 ± 2 and 178 ± 4 ka BP
226 (Zazo et al., 2002); 115, 125 and 135 ka BP (Bard, personal communication in Meco et al.,
227 2002). Dating of several mollusc shells from six other contemporaneous sites in Fuerteventura
228 give mean U/Th ages of 118 ± 11 ka BP ($n = 22$) (Zazo et al., 2002). Taken altogether, these
229 results indicate that the La Santa and Matas Blancas marine deposits can be attributed to the
230 LIG. Finally, the numerous U/Th datings at LIG sites in Lanzarote and Fuerteventura from
231 Zazo et al. (2002) exhibit a first order normal distribution, thereby confirming the validity of
232 the mean ages, despite the large standard deviation. According to these data, the marine
233 sediments, aged 121 ± 17 ka BP on Lanzarote and 118 ± 11 ka BP on Fuerteventura, were
234 probably deposited during the second part of the LIG.

235 As the shells analyzed in the present study were sampled from a marine deposit, the
236 average age of the analyzed molluscan assemblage therefore corresponds to the average age
237 of the deposit. Sedimentation took place during a steady period of sea-level highstand when
238 sedimentary particles and shells could accumulate over a period of time long enough to
239 generate substantial sedimentary formation, i.e., at least a few thousand years. To understand
240 the general context in which our samples were deposited, we first present here the different
241 reconstructions of sea-level evolution during the LIG. In order to reconstruct past sea-levels,
242 many studies have relied on the oxygen isotope compositions of foraminifera tests trapped in
243 marine sediments (Shackleton et al., 1987; McManus et al., 1999; Shackleton et al., 2000;
244 Waelbroeck et al., 2002; Lisiecki and Raymo, 2005; Rohling et al., 2009; Sosdian and
245 Rosenthal, 2009; Elderfield et al., 2012; Rohling et al., 2014; Shakun et al., 2015). In
246 successive reviews the issue of sea-level and its relative variations during the LIG has
247 received much attention in the scientific community. Different chronologies, deduced from
248 proxy reconstructions and synthesis compilations, of the sea-level highstand during the LIG
249 were compiled in Table 2. It should be noted that the diversity of chronological methods used

250 to quantify paleo-sea-levels has sometimes led to substantial differences in the acquisition and
251 appreciation of ages, a fact that may contribute to uncertainty in the chronology of sea-level
252 changes. Dutton et al. (2015a) specified that sea-level during the LIG varied from +6 m to +9
253 m, with a peak occurring after 125 ka BP, probably between 122 and 119 ka BP. Capron et al.
254 (2017a,b), on the basis of the work made by Dutton et al. (2015a), emphasized that most
255 studies consider a two-phase sea-level change during the LIG, including a late event that
256 peaked between 122 and 119 ka BP. We conclude from all these works that during the LIG
257 the global sea-level rose by between +6 m and +9 m, with a maximum rise that very likely
258 took place in the second part of the LIG, after ≈ 125 ka BP, and extended for a few thousand
259 years until 119-116 ka BP. Secondly, the age estimation of the onset of sea-level highstand
260 during the LIG is corroborated by the history of sea-level and that of global surface
261 temperature during the last deglaciation and the Holocene. The timing of the changes in the
262 global mean surface temperature can be documented by that of the evolution of the Antarctic
263 temperature (see § 6.2.2) (Figure 6b,c). During the Holocene, the mean surface temperature
264 has remained maximum and relatively steady since ≈ 12 ka BP (Marcott et al., 2013) (Figure
265 6b). From ≈ 18 ka to ≈ 7 ka BP, the sea-level has risen by more than 120 metres (Waelbroeck
266 et al., 2002; Lambeck et al., 2014) (Figure 6d). Dutton et al. (2015a) note that at the beginning
267 of the Holocene the global mean sea-level was still ≈ 60 m lower than today. The sea-level
268 highstand has been steady only since ≈ 7 ka BP. The delay of about 5 ka between the
269 establishment of the maximum of the mean temperature and the establishment of the
270 maximum mean sea-level stemmed mainly from the melting dynamics of ice sheets from the
271 northern hemisphere and Antarctica. It follows that a time lag of a few thousand years
272 between the beginning of the LIG and the beginning of the global mean sea-level highstand
273 during the LIG is consistent with the history of the Holocene global mean sea-level. Since
274 during the LIG the mean surface temperature reached its maximum at ≈ 130 ka BP (Figure

275 6c), which marks the beginning of the interglacial stage, we infer that sea-level highstand
276 began at ≈ 125 ka BP (Figure 6e). Therefore, based on the history of the last deglaciation and
277 that of the Holocene it is very likely that the LIG marine deposits took place during the long
278 time period of the sea-level highstand starting from ≈ 125 ka BP to 119-116 ky BP.

279 Finally, the chronology of the sea-level highstand during the LIG and the resulting
280 estimate of the age of our samples is corroborated by the stratigraphic analysis performed by
281 Meco et al. (2011) in the Canary Islands (see §2.1). The beginning of interglacials is
282 characterized by a humid phase ("African Humid Period") with the development of a soil
283 (labelled (ii)); as sea-levels rise, the soil is covered by seawater and marine sediments begin to
284 form (labelled (iii)). During MIS1 or the Holocene, the two phases (ii) and (iii) are clearly
285 identified: (ii) paleosols were mainly established between ≈ 10 ka and ≈ 8 ka BP, (iii) marine
286 deposits between ≈ 5 ka and ≈ 1 ka BP, as dated from different samplings (Meco et al., 2011)
287 (Figure 6a). The genesis of paleosols thus preceded that of marine sediments by a few
288 thousand years. Such observations are consistent with the timing of the sapropel deposits in
289 the Mediterranean (Rossignol-Strick, 1985) known to be correlated with the maximum
290 intensity of the African wet phase between ≈ 10 ka and ≈ 8 ka BP (deMenocal et al., 2000;
291 Lézine et al., 2011). As the sea-level has remained steady since ≈ 7 ka BP (Lambeck et al.,
292 2014), the marine deposits are contemporaneous with the sea-level highstand. During the
293 LIG, these two phases are well documented, the paleosol formation (ii) preceding the marine
294 deposit (iii) (Figure 2a). Therefore, as pointed out by Meco et al. (2011), the stratigraphy of
295 the paleosols and marine deposits indicate that the paleosol formation occurred during the
296 first part of the LIG, and was followed by the sea-level highstand few thousand years after the
297 beginning of the LIG, in the second part of the interglacial. It was during this sea-level
298 highstand that the marine deposits were set down.

299 Finally, the three different approaches (deposit dating, the sea-level highstand history,
300 and succession of paleosols and marine deposits) led to the same conclusion : the marine
301 deposits from La Santa, Lanzarote, and Matas Blancas, Fuerteventura, date from the second
302 part of the LIG, i.e. from ≈ 125 ka to $\approx 119-116$ ka BP, at a time of sea-level highstand.

303

304 4. Methods

305

306 4.1. Analytical techniques

307

308 4.1.1. Raman spectroscopy and mineralogy of the sample shells

309

310 We used Raman spectroscopy in order to characterize the mineralogy of all the studied
311 skeletal carbonates. The Raman Spectrometer X'plora, hosted by the Laboratoire de Géologie
312 de Lyon at the University Claude Bernard Lyon 1, has been operated by using an objective x
313 1000, an optical network of 1800 lines per mm, a monochromatic laser (wavelength of 532
314 nm) filtered by 10% with ten acquisitions (15 s per acquisition) performed between 100 and
315 1600 cm^{-1} .

316 Aragonite was systematically identified from the low frequency part of the Raman
317 spectra (Figure 7), and especially from the position and splitting of the symmetric bending
318 mode which occurs as a doublet at $701-704\text{ cm}^{-1}$ according to Unvros et al. (1991) and Gillet
319 et al. (1993). These analyses attest that the fossils shells studied consist of aragonite. It is
320 known that the calcite to aragonite transition takes place at high pressure/high temperature
321 (Gillet et al., 1987). As the mollusc shells precipitate at low pressure/low temperature, any
322 polymorphism transformation from calcite to aragonite is *a priori* prevented during the shell
323 growth. Thus the biogenic carbonates are considered to be originally aragonitic and the fossils

324 pristine. This is confirmed by Raman spectroscopy on the modern bivalve *Cerastoderma*
325 *edule* (or *Cardium edule*) species which forms an entirely aragonitic shell (Füllenbach et al.,
326 2015). Also, Cornu et al. (1993) acknowledged that the inorganic part of the modern gastropod
327 *Persististrombus latus* (*Strombus bubonius*) is composed exclusively of aragonite. Those two
328 species are studied in the present work. Therefore, the results obtained by Raman
329 spectroscopy, which indicate that the shell carbonate of the fossil specimens is aragonitic, are
330 considered to be a reliable indicator of isotopic signal (O and C) preservation.

331

332 4.1.2. Oxygen and carbon isotope analysis of aragonite mollusc shells

333

334 Stable carbon and oxygen isotope compositions were measured on a total of 32
335 mollusc shells of 6 species. Selected samples were cleaned with double-deionized water, then
336 placed in an ultrasonic bath to remove any trace of sediment, and finally dried in a dry-oven.
337 Two types of sampling were performed : (1) bulk sampling, which delivers an average signal
338 over the entire life of the specimen, and (2) incremental sampling, which delivers discrete
339 signals during the ontogenic evolution of the specimen, i.e., sampling of different periods of
340 the shell growth (younger to older periods). (1) For bulk shells, the entire sample was crushed
341 with a hammer in a steel mortar to obtain a coarse powder, which we reduced to a fine powder
342 (50 to 100 μm particle size) with a pestle and agate mortar. The aliquot size was about 300 μg
343 of CaCO_3 . Each shell represents one bulk sample. 30 shells from La Santa and Matas Blancas
344 sites were selected for the bulk sampling, including 4 gastropods and 2 bivalves species. (2)
345 For incremental sampling of the shells, first, as the surface of the outer shell layer is most
346 often easily altered, the outermost part of the shell was gently removed using a diamond
347 micro-drill (DremelTM). Second, after cleaning the shell with double-deionized water, the
348 layer underneath the outermost surface of the shell was drilled at regular intervals along the

349 growth direction. Two shells from La Santa (Lanzarote) and Matas Blancas (Fuerteventura)
350 sites were selected for the incremental sampling: one bivalve specimen and one gastropod
351 specimen, respectively. For the bivalve *Cardium edule* (*Cerastoderma edule*), 16 incremental
352 samplings were performed along its main growth axis, one sampling every ≈ 3 mm (Figure
353 S1a,b in the Supplementary materials). As the average growth rate of the *Cardium edule*
354 shells ranges from ≈ 1.8 mm.month⁻¹ to ≈ 0.8 mm.month⁻¹ (Richardson et al., 1980; Eisma et
355 al., 1965), and as the growth rates decrease with increasing age (Bourget and Brock, 1990),
356 we estimated that we performed one sampling every ≈ 2 to ≈ 4 months. For the gastropod
357 *Stramonita haemastoma*, 29 incremental samplings were carried out, one sampling every ≈ 5
358 mm, the sampling scheme precisely following the growth spiral of the shell according to the
359 protocol of Cornu et al. (1993) (Figure S1c,d).

360 Stable isotope ratios were determined by using an auto sampler MultiPrepTM system
361 coupled to a dual-inlet GV IsoprimeTM isotope ratio mass spectrometer (IRMS). For bulk
362 shells, aliquot size was about 300 μ g of calcium carbonate while they were about 10–15 μ g in
363 the case of an incremental sampling strategy. All aliquots were reacted with anhydrous
364 oversaturated phosphoric acid at 90°C during 20 minutes. By default, oxygen isotope ratios of
365 calcium carbonate are computed assuming an acid fractionation factor $1000\ln\alpha(\text{CO}_2\text{-CaCO}_3)$
366 of 8.1 between carbon dioxide and calcite according to Swart et al. (1991) experimental data.
367 However, acid fractionation factors differ between aragonite and calcite as shown by Kim and
368 O'Neil (1997) and Kim et al. (2007a). At a temperature of 90°C, the mean difference in acid
369 fractionation factors between calcite and aragonite is -0.41‰ (Kim et al., 2007a).
370 Consequently, such offset value was applied to all mollusc oxygen isotope measurements as
371 all the samples are made of aragonite. In addition, all sample measurements were duplicated
372 and adjusted to the international reference NIST NBS19 ($\delta^{18}\text{O}_{\text{V-PDB}} = -2.20\text{‰}$; $\delta^{13}\text{C}_{\text{V-PDB}} =$
373 $+1.95\text{‰}$) and NIST NBS18 ($\delta^{18}\text{O}_{\text{V-PDB}} = -23.2\text{‰}$; $\delta^{13}\text{C}_{\text{V-PDB}} = -5.01\text{‰}$). External

374 reproducibility (2σ) is lower than $\pm 0.10\text{‰}$ for $\delta^{18}\text{O}$ and $\pm 0.05\text{‰}$ for $\delta^{13}\text{C}$. $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of
375 aragonite mollusc shells are expressed with respect to V-PDB.

376

377 4.2. Isotope fractionation equations for the aragonite-water system

378

379 The Raman spectra analyses indicate that the shell carbonate of the fossil specimens is
380 aragonitic. As modern biogenic aragonites are precipitated close to oxygen isotope
381 equilibrium with ambient water (Kim et al., 2007b), it is reasonable to assume that this
382 condition was also met for past biogenic aragonites. Indeed, Grossman and Ku (1986) and
383 Kim et al. (2007b) demonstrated that oxygen isotope fractionation between water and
384 aragonite can be described by a well-constrained relationship with temperature, which is a
385 necessary and sufficient condition for reconstructing marine paleotemperatures from aragonite
386 fossils.

387 The sea surface temperatures were calculated using two oxygen isotope fractionation
388 equations for the aragonite-water system. The first of these (Grossman and Ku, 1986) was
389 determined by analyzing modern aragonitic molluscs, foraminifera and ambient water in the
390 temperature range ≈ 3 to 22°C with $\delta^{18}\text{O}_{\text{aragonite}}$ ($\delta^{18}\text{O}_{\text{ar}}$) and $\delta^{18}\text{O}_{\text{seawater}}$ ($\delta^{18}\text{O}_{\text{sw}}$) relative to V-
391 PDB:

392

$$393 \quad T = 20.60 - 4.34(\delta^{18}\text{O}_{\text{ar}} - \delta^{18}\text{O}_{\text{sw}}) \quad (1)$$

394

395 where T ($^\circ\text{C}$) is the temperature in which the shell grew. The second oxygen isotope
396 fractionation equation (Kim et al., 2007b) was obtained from laboratory experiments with
397 $\delta^{18}\text{O}_{\text{ar}}$ and $\delta^{18}\text{O}_{\text{sw}}$ relative to V-SMOW:

398

399 $1000 \ln \alpha_{ar-sw} = 17.88 \left(\frac{10^3}{T} \right) - 31.14$ (2)

400

401 T being the temperature in Kelvin, α the oxygen fractionation coefficient between synthetic
402 aragonite and water:

403

404 $\alpha_{ar-sw} = \frac{\left(\frac{^{18}\text{O}}{^{16}\text{O}} \right)_{ar}}{\left(\frac{^{18}\text{O}}{^{16}\text{O}} \right)_{sw}} = \frac{(\delta^{18}\text{O}_{ar} + 1000)}{(\delta^{18}\text{O}_{sw} + 1000)}$ (3)

405

406 As we need the $\delta^{18}\text{O}$ of the aragonitic shell expressed relative to the V-SMOW (Eq.(2)) we
407 use the following relationship between the V-PDB isotope scale and the V-SMOW scale
408 established at 25°C by Coplen et al. (1983):

409

410 $\delta^{18}\text{O}_{(\text{V-SMOW})} = 1.03091 \delta^{18}\text{O}_{(\text{V-PDB})} + 30.91$ (4)

411

412 The two temperature calibrations for the aragonite-water oxygen isotope fractionation
413 have been tested on 50 modern shells, both gastropods and bivalves, in Crete, Greece, where
414 present-day prevailing mean SSTs are similar to those of the Canary Islands (about 20°C)
415 (Lécuyer et al., 2018). The SST calculated from the $\delta^{18}\text{O}$ values of those shells and $\delta^{18}\text{O}$
416 values of ambient seawater in Crete match the SST available from oceanic databases available
417 for the Mediterranean Sea on using either the Kim et al. (2007b) equation or the Grossman
418 and Ku (1986) equation. The best match between the present SST and calculated values was
419 obtained by applying the Kim et al. (2007b) equation. Therefore it seems more appropriate to
420 privilege the use of the Kim et al. (2007b) equation for calculating SST of Lanzarote and
421 Fuerteventura coastal waters during the LIG. However, as Grossman and Ku (1986)'s

422 equation was established from an *in situ* calibration and is predominantly used in the literature
423 for SST reconstructions based on mollusc shells, we also present the results of this equation in
424 order to allow comparisons of the results of this study with other publications.

425

426 4.3. $\delta^{18}\text{O}$ reconstruction of coastal seawater during the LIG

427

428 Reconstitution of marine paleotemperatures based on the $\delta^{18}\text{O}$ values of aragonite
429 mollusc shells requires an estimate of the oxygen isotope composition of contemporaneous
430 seawater of the living shells (Equations (1) and (2)). In this study, we therefore estimated the
431 $\delta^{18}\text{O}$ of seawater during the LIG sea-level highstand in the coastal environment of sampling
432 sites. Firstly, we considered the present-day coastal marine waters of the Canary Islands.
433 Seawater sampled at 1 m to 5 m depth from Lanzarote and Fuerteventura (Figure 1b) has
434 comparable $\delta^{18}\text{O}$ V-SMOW values, $1.09 \pm 0.04\text{‰}$ and $1.08 \pm 0.07\text{‰}$, respectively (Clauzel et
435 al., 2019). Secondly, we calculated the oxygen isotope composition of the LIG global ocean
436 by taking into account the sea-level change relative to present-days. Dutton et al. (2015a)
437 assumed a mean seawater elevation of 7.6 m with respect to the Holocene level that was due
438 to the melting of ice caps with a contribution of +2.0 m from Greenland + 4.6 m from
439 Antarctica, +0.6 m from the glaciers, and +0.4 m from the thermal expansion of the oceans.
440 Goelzer et al. (2016) proposed similar values of +1.4 m for Greenland, +4.4 m for the
441 Antarctica, and +0.4 m for the thermal expansion of the oceans, the contribution of glaciers
442 not being taken into account in this study. The $\delta^{18}\text{O}$ of ice in the Greenland Ice Core Project
443 (GRIP) and North Greenland Ice Core Project (NGRIP) ice cores (North Greenland Ice Core
444 Project members, 2004), located on the ice cap plateau, enable us to estimate for the
445 Greenland ice cap an average $\delta^{18}\text{O}$ value of about $-40.0 \pm 2\text{‰}$. From the data of EPICA Dome
446 C (EDC) and Dome F (Fuji) ice cores (e.g., Landais et al., 2007, and Dome Fuji Ice Core

447 Project Members, 2017, respectively), located on the ice cap plateau, we evaluate to a first
448 approximation the mean $\delta^{18}\text{O}$ of the Antarctic ice sheet at about $-57.5\pm 2\%$. Consequently,
449 considering a sea-level higher by 7.6 m (Dutton et al., 2015a) during the highstand period of
450 the LIG (± 1.5 m), the present-day average ocean depth of 3682 ± 44 m (Charette and Smith,
451 2010), and neglecting the glacier contribution, a mass balance calculation indicates that the
452 contemporaneous seawater of the molluscs investigated was $0.09\pm 0.02\%$ lower than at
453 present. Finally, as we consider that the melting of the ice caps is the main source of $\delta^{18}\text{O}$
454 variation in coastal seawater in the Canary Archipelago between the LIG and now, the $\delta^{18}\text{O}$
455 (V-SMOW) of Lanzarote and Fuerteventura coastal seawaters were likely close to
456 $1.00\pm 0.06\%$ during the LIG sea-level highstand. In using Equation (1), a correction of -
457 0.27% was applied to convert $\delta^{18}\text{O}$ of seawater from the V-SMOW to V-PDB scales
458 according to Bemis et al. (1998) and established by Hut et al. (1987).

459

460 4.4. Estimation of the uncertainties

461

462 There are different sources of uncertainty associated with the calculation of the SST
463 values. We shall consider first the uncertainties in the two equation variables, $\delta^{18}\text{O}_{\text{ar}}$ and $\delta^{18}\text{O}_{\text{sw}}$
464 (see Equations (1), (2) and (3)). The external reproducibility of $\delta^{18}\text{O}_{\text{ar}}$ measured from the
465 mollusc shells ($\pm 0.10\%$) gives an error of $\pm 0.5^\circ\text{C}$. The error associated with $\delta^{18}\text{O}_{\text{sw}}$ of coastal
466 seawater in the Canary archipelago during the LIG ($\approx \pm 0.06\%$) is $\pm 0.3^\circ\text{C}$. Consequently, the
467 uncertainty range in the SST reconstruction is $\pm 0.6^\circ\text{C}$. This is the uncertainty stated in the
468 results. It is calculated as follows:

469

$$470 \quad \sigma_{\text{total}} = \sqrt{(\sigma_{\text{ar}})^2 + (\sigma_{\text{sw}})^2} = \sqrt{(0.5)^2 + (0.3)^2} = 0.6 \quad (5)$$

471

472 We now consider the uncertainty associated with the linear regression of the empirical
473 data ($y = mx+b$), which establishes the isotopic fractionation equation for oxygen in the
474 aragonite-water system (Equation (1) and Equation (2)). This uncertainty is not available in
475 the original publications. We used the “lm()” and “pred()” functions of R language (Chambers
476 and Hastie, 1992) and obtained $\pm 0.8^\circ\text{C}$ for the Grossman and Ku (1986) equation (Eq.(1)) and
477 $\pm 0.3^\circ\text{C}$ for the Kim et al. (2007b) equation (Eq.(2)).

478

479 5. Results

480

481 5.1. Oxygen isotope compositions of the mollusc shells

482

483 The bulk sampling method delivers an average signal over the entire life of the
484 specimen. The bulk oxygen isotope data for the two sites of La Santa, Lanzarote ($n = 27$) and
485 Matas Blancas, Fuerteventura ($n = 3$) range from -0.14‰ to 0.97‰ (V-PDB) with a mean
486 value of $0.36 \pm 0.27\text{‰}$ (Table S1, in the Supplementary materials).

487 Incremental sampling yields discrete signals of different periods of the shell growth
488 (younger to older periods). Both the bivalve specimen (*Cardium edule* or *Cerastoderma*
489 *edule*) from La Santa ($n = 1$) and the gastropod specimen (*Stramonita haemastoma*) from
490 Matas Blancas ($n = 1$) show pseudo-cyclic variations of their oxygen isotope compositions
491 recorded during shell growth. The *Cardium edule* (*Cerastoderma edule*) fossil shell has
492 oxygen isotope ratios ranging from -0.11‰ to 2.54‰ with a mean value of $1.18 \pm 0.72\text{‰}$
493 while the *Stramonita haemastoma* fossil shell has oxygen isotope ratios ranging from -0.52‰
494 to 0.66‰ with a mean value of $0.07 \pm 0.36\text{‰}$ (Table S2). A signal drift is observed during
495 the ontogenetic evolution of *Cardium edule* (*Cerastoderma edule*) with decreasing isotopic
496 compositions from the younger ($\approx 2\text{‰}$) to the older ($\approx 0.5\text{‰}$) parts of the shell (Figure 8). The

497 particularly high $\delta^{18}\text{O}$ values for the younger part of the record lies outside the range of
498 variation of the other recordings, i.e., the 30 bulk samples and the *Stramonita haemastoma*
499 incremental sample. No such drift is observed during the ontogenetic evolution of *Stramonita*
500 *haemastoma*.

501

502 5.2. Carbon isotope compositions of the mollusc shells

503

504 The bulk carbon isotope data for La Santa and Matas Blancas sites ($n = 30$) range from
505 1.48 ‰ to 3.40 ‰ (V-PDB) with a mean value of 2.46 ± 0.57 ‰ (Table S1).

506 For the incrementally-sampled specimens, $\delta^{13}\text{C}$ of *Cardium edule* (*Cerastoderma*
507 *edule*) range from 0.51 ‰ to 3.02 ‰ with a mean value of 2.04 ± 0.74 ‰ ($n = 1$), while $\delta^{13}\text{C}$
508 of *Stramonita haemastoma* range from 1.19 ‰ to 3.06 ‰ with a mean value of 1.98 ± 0.50 ‰
509 ($n = 1$) (Table S2). For the mollusc *Stramonita haemastoma*, the two-tailed probability P
510 calculated at a 95% level of confidence is 0.976 ($n = 28$; $r = 0.0068$) thus rejecting the null
511 hypothesis that states there is a correlation between the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ of shell aragonite while
512 in the case of *Cardium edule* (*Cerastoderma edule*), the two-tailed probability P calculated at
513 a 95% level of confidence is 0.0014 ($n = 16$; $r = 0.7275$) confirming that both variables are
514 significantly correlated (Figure 9).

515

516 5.3. Paleo-temperature reconstructions of coastal seawater

517

518 5.3.1. Mean annual SST during the LIG sea-level highstand

519

520 Marine molluscs living in tropical environments grow all year round and record
521 variations in temperature or salinity at the daily scale (e.g. Frank, 1969). Lifespan for most

522 marine molluscs, and especially bivalves, is lower than 11 years. The data compiled by Moss
523 et al. (2016) reveal that populations of tropical (less than 30° in latitude) bivalves record a
524 mean Maximum reported LifeSPan (MLSP) of 7.9 years, whereas those living under higher
525 latitudes have a mean MLSP of 24.7 years. Therefore, the bulk sampling of mollusc shells
526 delivers an average isotopic record integrated over the entire growth period of the specimen
527 and whose duration is commonly several years. Finally, temperatures calculated from the $\delta^{18}\text{O}$
528 of bulk shells are considered to represent the mean annual SST. The data are listed in Table
529 S1, in the Supplementary materials.

530 For La Santa and Matas Blanca sites taken altogether (30 samples), we obtain an
531 average SST of $22.2 \pm 1.2^\circ\text{C}$ according to the Grossman and Ku (1986) equation (Eq. (1)) and
532 $20.4 \pm 1.3^\circ\text{C}$ according to the Kim et al. (2007b) equation (Eq.(2)) (Figure 10). The standard
533 deviation is $\leq \pm 1.3^\circ\text{C}$ and the SST follow a normal distribution according to the use of
534 normal probability plots with regression coefficients of 0.985. This indicates the sampling of
535 a homogenous assemblage of molluscs, representative of their living environment during the
536 sampling interval, which corresponds to the duration of the LIG sea-level highstand. In
537 summary, values of $22.2 \pm 1.2^\circ\text{C}$ (Eq. (1)) and $20.4 \pm 1.3^\circ\text{C}$ (Eq. (2)) represent an average SST
538 coming from 30 molluscs whose deposit extends over several thousand years (≈ 125 to 119-
539 116 ky BP), each sample providing a mean annual SST value over a few years of life.

540

541 5.3.2. Seasonal SST during the LIG sea-level highstand

542

543 As the incremental sampling gives access to different periods recorded during the
544 growth of the mollusc shell, it could potentially allow seasonal temperature variations to be
545 reconstructed. Sclerochronological studies performed on modern bivalve molluscs show that
546 they form patterns of periodic annual banding in their shell (Butler and Schöne, 2016).

547 Moreover, image processing by Scanning Electron Microscopy (SEM) on the outer surface of
548 modern bivalve *Cerastoderma edule* shells from high latitudes (North Sea, 55°N) show that
549 winter and summer growth lines are consistent with $\delta^{18}\text{O}$ variations from shell incremental
550 measurements, higher $\delta^{18}\text{O}$ values corresponding to cold water conditions ($\approx 10^\circ\text{C}$) and lower
551 $\delta^{18}\text{O}$ values to warm water conditions ($\approx 22^\circ\text{C}$) (Milano et al., 2015). Data from *Stramonita*
552 *haemastoma* (Matas Blanca) (Figure 11) and *Cardium edule* (*Cerastoderma edule*) (La Santa)
553 (Figure 12) (2 samples) show a pseudo-cyclical SST record, particularly obvious for
554 *Stramonita haemastoma*. The variations, of sinusoidal type, are distorted in the ontogenic
555 profiles since the exoskeleton growth mode of many molluscs follows a logarithmic law
556 (Cornu et al., 1993; Roman-Roman et al., 2010). Mean temperatures reconstituted from the
557 $\delta^{18}\text{O}$ of shell increments from the *Stramonita haemastoma* specimen collected at Matas
558 Blanca, Fuerteventura, and the *Cardium edule* (*Cerastoderma edule*) specimen collected at La
559 Santa, Lanzarote, are between $23.5 \pm 1.6^\circ\text{C}$ and $21.8 \pm 1.8^\circ\text{C}$ and between $18.7 \pm 3.1^\circ\text{C}$ and
560 $16.6 \pm 3.3^\circ\text{C}$, respectively, according to Equation (1) and Equation (2). The first set of
561 temperatures is close to those obtained for the 30 bulk samples, while the second set is much
562 lower and displays a wide distribution (standard deviation of more than 3°C) due to the $\delta^{18}\text{O}$
563 signal drift (§5.1).

564 For the gastropod *Stramonita haemastoma*, about 4 seasonal SST cycles are observed
565 over a total length to apex of 16 cm (Figure 11) with variations in SST comprised between
566 $\approx 3.5^\circ\text{C}$ and $\approx 5.5^\circ\text{C}$. For the bivalve *Cardium edule* (*Cerastoderma edule*), we observe SST
567 fluctuations that mimic a seasonal-like signal recorded over two or more cycles, over a total
568 length to umbo of 4.5 cm (Figure 12). Due to a signal drift (§5.1), we focus here on the oldest
569 biological age of the mollusc shell. The average SST comprised between $21.1 \pm 1.6^\circ\text{C}$ (Eq.
570 (1)) and $19.2 \pm 1.8^\circ\text{C}$ (Eq. (2)) is consistent with the average temperature reconstructed from
571 the 30 bulk samples. The seasonal SST fluctuation is approximately 6.0°C . We note that this

572 SST amplitude is also observed for the youngest part of the mollusc shell. The low SST
573 values recorded by the youngest part of the *Cardium edule* (*Cerastoderma edule*) shell,
574 caused by the $\delta^{18}\text{O}$ signal drift, could result from an inter-annual fluctuation of the SST, with
575 particularly cooler marine waters during the first period of life of the mollusc sample, a period
576 that may last for 1 or 2 years. Finally, seasonal SST values during the LIG sea-level
577 highstand, established from two samples with a lifespan of a few years, range from $\approx 3.5^\circ\text{C}$ to
578 $\approx 6.0^\circ\text{C}$.

579

580 6. Discussion

581

582 6.1. Interpretation of the fossil faunal record

583

584 The ecology of the species present in the LIG marine deposits indicates that they lived
585 in different types of environments (§3.1) that can be designated as (a) warm water
586 environments, (b) warm and cooler water environments and (c) cooler water environments
587 (Figure 5).

588 The co-occurrence of these different faunal groups in the LIG fossil record can be
589 understood as follows. The marine deposits accumulated throughout the second part of the
590 LIG, a period when the warm climate became progressively cooler, similar to that observed
591 during the second part of the Holocene (Marcott et al., 2013). The whole LIG marine deposit
592 therefore consists of both “warm fauna”, corresponding to the warm phase of the second part
593 of the LIG, and “cooler fauna”, corresponding to the end of the LIG, before the sea-level fell
594 and the sediment accumulation stopped. Firstly, during the LIG warming, the “warm species”
595 or extralimital southern species (Muhs et al., 2014) migrated from low latitudes up to latitudes
596 as far as the Mediterranean sea (Muhs et al., 2014). Secondly, when the SST decreased,

597 extralimital southern species disappeared in the Canary Islands. This scenario of faunal shifts
598 during the LIG in the Canary Islands was also documented by Muhs et al. (2014), but whereas
599 they deal with the whole of the LIG period, here, we consider only the sequence of sediment
600 deposits, i.e., the second part of the LIG. It follows that in principle the first arrival of the
601 “warm species” is not recorded in our marine deposits.

602 These faunal changes are illustrated by the presence of *Persististrombus latus*
603 (*Strombus bubonius*), known as a Guinean-Senegalese taxon, in the samples we collected in
604 the Matas Blancas deposit, i.e., north of its present geographical distribution. This observation
605 indicates that during the LIG the SST was higher than in the present interglacial, as already
606 noted by Muhs et al. (2014). This finding is consistent with the particularly high and
607 reproducible LIG SST values recorded for the 3 specimens of *Persististrombus latus*, namely
608 $23.4 \pm 0.5^\circ\text{C}$ (Eq.(1)) and $21.7 \pm 0.6^\circ\text{C}$ (Eq.(2)) (see Table 2). These values are clearly higher
609 by about $+1.2^\circ\text{C}$ than the mean temperature found for the total bulk sampling (30 samples,
610 §5.3.1), which includes the cooling period at the end of the LIG.

611

612 6.2. Interpretation of the mean annual paleo-SST reconstructions

613

614 6.2.1. LIG SST anomaly with respect to pre-industrial times

615

616 As in the first step, where the mean annual SST was estimated from mollusc shell $\delta^{18}\text{O}$
617 values in the eastern Canary Islands during the LIG, in a second step we estimated the change
618 in SST with respect to pre-industrial times, i.e., the middle of the nineteenth century.

619 We recall the SST value of 20.4°C in the eastern islands of Canary archipelago for the
620 period from 2007 to 2017 (§2.2).

621 The recent evolution of air temperature has been evaluated by Martín et al. (2012) from
622 meteorological measurements on Tenerife Island, Canary archipelago. A warming trend of

623 $0.09\pm 0.04^{\circ}\text{C}\cdot\text{decade}^{-1}$ between 1944 and 2010 was observed, which increased to
624 $0.17\pm 0.04^{\circ}\text{C}\cdot\text{decade}^{-1}$ in the period 1970-2010. At Gran Canaria, Canary archipelago, Luque
625 et al. (2014) observed similar trends over the same periods, respectively. Such acceleration in
626 the temperature rise is consistent with the SST trends from Advanced Very High Resolution
627 Radiometer (AVHRR) and *in situ* data in the eastern Canaries published by Gómez-Letona et
628 al. (2017) who estimated this trend to be about $0.22^{\circ}\text{C}\cdot\text{decade}^{-1}$ during the period 1993–2014.

629 From the estimated SST warming of about $0.2\pm 0.04^{\circ}\text{C}\cdot\text{decade}^{-1}$ since the seventies
630 (Gómez-Letona et al., 2017; Martín et al., 2012; Luque et al., 2014), and the increase in SST
631 of about $0.2\pm 0.05^{\circ}\text{C}$ between the beginning of the twentieth century to the nineteen seventies
632 based on the GMST trend (IPCC, 2013), we assume that the SST increased by about
633 $1.0\pm 0.1^{\circ}\text{C}$ in the eastern Canaries between the pre-industrial times and the 2010s. This
634 estimate is based on the fact that that global temperature increase observed from the
635 beginning of the pre-industrial until the beginning of the 20th century (IPCC, 2013) is
636 negligible. We therefore assume a pre-industrial SST of the order of $19.4\pm 0.1^{\circ}\text{C}$. This result
637 is consistent with the calculated SST value of 19.8°C in 1986 (Borges et al., 2004), which
638 implies a pre-industrial SST of about 19.3°C .

639 Consequently, with respect to pre-industrial times, our results show that the mean SST
640 anomaly that prevailed during the LIG sea-level highstand is comprised between at least
641 $+1.0\pm 1.4^{\circ}\text{C}$ (Equation (2)) and $+2.8\pm 1.3^{\circ}\text{C}$ (Equation (1)) at the two coastal sites of La Santa
642 and Matas Blancas.

643

644 6.2.2. Global and regional LIG temperature distribution

645

646 A full understanding of the context in which the local SST warming took place in the
647 eastern Canary Islands calls for a broad view of the temperature change both on a regional

648 scale and on a global scale. Owing to coupling between the regional and local scale pattern
649 and the global climate, it is possible to estimate the consistency of our results with the
650 temperature pattern at the LIG. Two strong qualitative pieces of evidence show that the LIG
651 was globally warmer than the Holocene : the amply documented Antarctic polar warming,
652 and the sea-level rise and high-stand above that of the present (+6 to +9 m) (e.g., Table 2,
653 Figure 6). How much warmer the LIG was on Earth at the regional and global scale remains
654 an intensely discussed topic within the paleoclimate community. Nevertheless, some major
655 aspects of this warming can be summarized as follows.

656 The Past Interglacials Working Group of PAGES (2016) recently published an
657 overview of the climate patterns of the LIG. The different interglacial periods over the last
658 800 ka were described. The characteristics of the successive interglacials depend on seasonal
659 and latitudinal insolation, which are determined by astronomical parameters (Milankovitch,
660 1941; Berger, 1988). The time dependence of these parameters explains the markedly warm
661 climate of some interglacial periods, in particular the LIG (e.g., Mélières and Maréchal,
662 2015). The LIG appears to be the warmest of the last eleven interglacial stages: compared to
663 the Holocene, it was warmer by between 0°C and +4°C depending on the latitude (Past
664 Interglacials Working Group of PAGES, 2016). At the global scale, the LIG global mean
665 surface temperature (GMST) can first be inferred independently from the $\delta^{18}\text{O}$ and δD
666 temporal series measurements of ice cores located in the plateau of Antarctica. These ice core
667 records allow us to reconstruct changes in air temperatures above the ice cap. We focus here
668 on the Dome C ice core, which shows an air temperature record consistent with those of three
669 other coring sites at Vostok, Dronning Maud Land and Dome F (see for example Capron et
670 al., 2014). During the LIG, they indicate that, with respect to the mean Holocene temperature,
671 the Antarctic air surface temperature record divides into two periods, in which a peak
672 anomaly of about +4°C around ≈ 130 ka BP is followed by a plateau type signal of about

673 +2°C between ≈ 126 ka and ≈ 119 ka BP (Figure 6c). These local temperatures reflect the
674 regional temperature because the Antarctic plateau is characterized by spatially-equalized
675 climatic conditions. In different studies a linear relationship of GMST change for every 1°C
676 change in Antarctic temperatures across glacial-interglacial cycles is inferred : between 0.5°C
677 (Masson-Delmotte et al., 2010; Chylek and Lohmann, 2008; Hansen et al., 2008) and 0.6°C
678 (Snyder et al., 2016). We can therefore estimate a maximum GMST anomaly of at least
679 $\approx +2^\circ\text{C}$ around ≈ 130 ka BP, followed by a plateau type temperature signal of at least $\approx +1^\circ\text{C}$
680 between ≈ 126 ka and ≈ 119 ka BP. Our samples are therefore contemporaneous with a period
681 defined by a GMST warmer than present by about $+1^\circ\text{C}$ relative to the mean Holocene
682 temperature. These GMST anomalies are consistent with the estimates of Snyder et al. (2016)
683 and Hansen et al. (2013), which are based on a broader time scale. Snyder et al. (2016) used a
684 multi-proxy database compilation to reconstruct the GMST over the last 2 Ma: they found
685 that during the LIG the GMST was $+2^\circ\text{C}$ warmer than in the late Holocene (0-5 ka BP).
686 Finally, Hansen et al. (2013) reconstructed the GMST over the last 65 Ma and calculated that
687 the LIG was $+1.4^\circ\text{C}$ higher than during the Holocene. Regarding exclusively the global sea
688 surface temperature, a warming of $+0.5^\circ\text{C} \pm 0.3^\circ\text{C}$ at 125 ka with respect to the pre-industrial
689 times has been reported by Hoffman et al. (2017).

690 Focusing now on regional changes, during the LIG the Arctic region was consistently
691 warmer than today. The CAPE-Last Interglacial Project Members (2006) offered quantitative
692 estimates of circum-Arctic LIG summer air and sea surface temperatures (SST), reconstructed
693 from proxies inferred from terrestrial and marine records. These reconstructions indicate that
694 Arctic summer air temperatures were on average about $+4$ to $+5^\circ\text{C}$ higher than at present for
695 most of the Arctic, well above the planetary LIG average. A synthesis of temperature data
696 obtained at different climatic time slices at 125 ka BP and 120 ka BP indicates that the highest
697 temperature was about $+4^\circ\text{C}$ to $+5^\circ\text{C}$ warmer than present in central Greenland (Capron et al.,

698 2015, 2014). Regarding continental data obtained at mid latitudes in the northern hemisphere,
699 Kaspar et al. (2005) compared the anomalies between reconstructed January and July
700 temperatures for a 125 ka time slice and present-day observed temperatures in Europe. Their
701 reconstructions, based on pollen and plant macrofossils, indicate warming by a few degrees.
702 Brewer et al. (2008) report climate reconstructions from across the European continent. Their
703 results show a traditional three-part LIG, with an early optimum, followed by a slight cooling
704 stage and finally a sharp drop in temperature. The climate anomalies were mapped for four
705 periods (127 ka, 123 ka, 117 ka, 109 ka BP). Maximum warming took place at 127 ka BP, a
706 period during which the climate anomalies ranged from +4°C in northern Europe to +1°C in
707 the south of France.

708 Concerning the sea surface temperature, Capron et al. (2017a) published a critical
709 evaluation of data compilation at high latitudes. These authors emphasise that all high-latitude
710 regions experienced warmer conditions at the 125 ka time slice relative to the present-day,
711 with temperature anomalies reaching $+1.6\pm 0.5^{\circ}\text{C}$ and $+0.8\pm 0.5^{\circ}\text{C}$ in the North Atlantic and
712 in the Southern Ocean, respectively. In the last decades, global syntheses have been
713 performed, first by Turney and Jones (2010) and by McKay et al. (2011). McKay et al. (2011)
714 suggest a peak in the LIG global annual SST warming of $+0.7\pm 0.6^{\circ}\text{C}$ with respect to Late
715 Holocene (5 to 0 ka BP). Both compilations indicate a general warming at high and middle
716 latitudes but, for the tropics, a less consistent pattern emerges in which SST cooling is
717 identified in the tropical Atlantic Ocean. However, Capron et al. (2017b) suggest that these
718 two LIG data syntheses may have major methodological limitations related to the chronology.
719 Recently, Hoffman et al. (2017) published a new LIG SST compilation of global extent with a
720 consistent temporal framework. The temperature anomalies with respect to 1870-1889 are
721 estimated for three different time slices during the LIG (129, 125 and 120 ka BP). Three
722 world maps of proxy-based mean annual SST anomalies at the LIG, based on 83 marine

723 sediment core sites, were established using temperature estimates from Mg/Ca in planktonic
724 foraminifera, $U^{K'}_{37}$ from alkenones, and microfossil reconstructions. For each time window,
725 significant cooling by a few degrees at the LIG took place in the tropical north Atlantic Ocean
726 and along the west African coast. At the global scale, as already noted, SSTs reached
727 $+0.5^{\circ}\text{C}\pm 0.3^{\circ}\text{C}$ at 125 ka.

728 All the above estimates concur to show that the GMST at LIG was on average
729 higher than during the Holocene. This warming was not uniform, but increased towards the
730 poles. Some oceanic regions, however, did not follow this rule, as suggested by SST
731 reconstitutions performed in the North Tropical Atlantic Ocean by Hoffman et al. (2017). In
732 particular, the waters of the tropical Atlantic gyre, whose descending branch, the Canary
733 Current, bathes the West African coast, appeared to be cooler.

734 Turning now to the regional scale, in the northeast tropical Atlantic area where the
735 Canary Islands are located, a few studies aim to reconstruct the SST during the LIG (Figure
736 3b, Table 3). Sicre et al. (2000) analysed terrigenous and marine biomarkers from a core
737 (SU94-20bK) that recorded the last 150 ka in the northeast Atlantic off Northwest Africa at
738 $25^{\circ}00.60'\text{N}$, $16^{\circ}23.40'\text{W}$. Their results from $U^{K'}_{37}$ analyses revealed an annual SST of 23.3°C
739 during the LIG (from ≈ 128 to ≈ 120 ka BP) and 21.8°C during the beginning of the
740 Holocene, thereby indicating an anomaly of $+1.5\pm 0.3^{\circ}\text{C}$ during LIG with respect to the core
741 top (dated at ≈ 9 ka BP). Matsuzaki et al. (2011) studied a core (MD03-2705) retrieved off
742 the Cape Verde Islands at $18^{\circ}05.81'\text{N}$, $21^{\circ}09.19'\text{W}$, spanning the last 220 ka. The annual SST
743 anomaly estimated from planktonic foraminifera assemblages between LIG (from ≈ 122 to
744 ≈ 118 ka BP) and the core top (dated at ≈ 10 -8 ka BP) is $+0.5\pm 0.6^{\circ}\text{C}$. Castañeda et al. (2009)
745 analysed a sediment core (GeoB9528-3) retrieved from the Guinea Plateau Margin at
746 $09^{\circ}09.96'\text{N}$, $17^{\circ}39.81'\text{W}$, spanning the last 192 ka. The $U^{K'}_{37}$ -based annual SST gives a LIG
747 anomaly value of $+1.0\pm 0.5^{\circ}\text{C}$ (from ≈ 125 to ≈ 120 ka) with respect to the core top (dated at

748 ≈ 9 ka). The SST anomalies from these studies (Sicre et al. (2000), Matsuzaki et al. (2011),
749 Castañeda et al. (2009)) with respect to the core top values are displayed in Figure 13a. These
750 values can be expressed relative to pre-industrial times by taking into account the results of
751 the two following investigations. Fischer et al. (2018) indicate a cooling of the global surface
752 air temperature by $\approx 0.7 \pm 0.3^\circ\text{C}$ between ≈ 9 ka BP and the period 1850-1900. The same
753 order of magnitude is obtained from Arbuszewski et al. (2013). In order to obtain the SST
754 anomalies with respect to the pre-industrial times, the value of 0.7°C must therefore be added
755 to the core top values (Figure 13b). Turning now to the evidence from coastal molluscs, Muhs
756 et al. (2014) noted a dramatic warming of marine waters around the Canary Islands during the
757 LIG, on the basis of on the presence of 16 exclusively extralimital southern species found in
758 their LIG deposits. At present, these species live between the Cape Verde Islands and south
759 Angola and are members of the tropical Senegalese-Guinean faunal province. During the LIG
760 one of the most celebrated Senegalese taxa, *Persististrombus latus* (*Strombus bubonius*)
761 (Figure 4), was distributed between Cape Verde Islands and the Mediterranean region. The
762 geographic extent of these extralimital southern species during the LIG indicates that these
763 numerous mollusc species migrated northwards during the LIG, a situation that requires
764 warmer waters than at present. Montesinos et al. (2014) estimated the SST anomaly around
765 the Canary Islands by comparing the present geographical distribution of an extralimital
766 southern warm-water species, *Harpa doris* Röding, 1798 (*Harpa rosea*), with that prevailing
767 during the LIG. This estimate is based on the presence of *Harpa doris* in marine deposits
768 from the Canary Islands (Gran Canaria, Lanzarote and Fuerteventura) during the LIG. This
769 mollusc disappeared from the Canary Islands after the LIG and has not reappeared since.
770 Today, *Harpa doris* lives far to the south in the Gulf of Guinea, from Angola to Senegal.
771 Comparison between the annual SST of the present habitat of *Harpa doris* and the annual
772 SST of the Canary Island coasts enabled Montesinos et al. (2014) to evaluate a surface

773 temperature anomaly at LIG of at least $+3.3^{\circ}\text{C}$ higher than today. To relate this value to the
774 pre-industrial period, we applied the correction method described in § 6.2.1. Since
775 Montesinos et al. (2014) took the 2000s as the reference date for the present period, we
776 assume the increase in SST in the Canary Islands from the pre-industrial to the 2000s to be
777 about 0.8°C . The annual SST anomaly is therefore at least $+4.1^{\circ}\text{C}$ higher than in pre-
778 industrial times. This value spans the LIG marine deposit period that we estimated to be
779 between ≈ 125 and 119-116 ka BP (§ 3.2) (Figure 13b). Finally, the compilation work of
780 Hoffman et al. (2017) includes three cores located in the tropical Northeast Atlantic region:
781 GIK15637 ($27^{\circ}00.00'\text{W}$, $18^{\circ}58.80'\text{W}$), M12392-1 ($25^{\circ}09.60'\text{W}$, $16^{\circ}51.00'\text{W}$), and V22-196
782 ($13^{\circ}49.80'\text{N}$, $18^{\circ}58.20'\text{W}$), where foraminiferal assemblage-based annual SST anomalies
783 with respect to the pre-industrial times (1870-1889) for three different time slices during the
784 LIG (129, 125 and 120 ka BP) are deduced from the three maps shown. The values range
785 from $-4.8 \pm 0.5^{\circ}\text{C}$ to $+0.3 \pm 0.5^{\circ}\text{C}$ (estimated from the color code of the maps).

786

787 6.2.3. LIG SST data comparisons

788

789 Our results for the coastal surface waters during the second part of the LIG recorded in
790 Lanzarote and Fuerteventura indicate a positive temperature anomaly comprised between
791 $+1.0 \pm 1.4^{\circ}\text{C}$ and $+2.8 \pm 1.3^{\circ}\text{C}$ with respect to pre-industrial times. This positive anomaly,
792 obtained from geochemical analyses of mollusc shells in LIG marine deposits, is congruent
793 with data obtained from the northeast tropical Atlantic by two other methods, namely analyses
794 of microfossils and alkenones in deep-sea cores (Sicre et al., 2000 ; Matsuzaki et al., 2011 ;
795 Castañeda et al., 2009) and analyses of mollusc assemblages in LIG marine deposits (Muhs et
796 al., 2014 ; Montesinos et al., 2014) (Figure 13b). The relatively strong SST anomaly in the
797 Canary Islands reported by Montesinos et al. (2014) is consistent with the qualitative study of

798 Muhs et al. (2014), which covers a wider geographical area stretching as far as the
799 Mediterranean and includes a larger number of taxa. All of these studies clearly indicate a
800 warming in the second part of the LIG with respect to pre-industrial times.

801 However, such positive temperature anomalies appear to conflict with the compilation
802 of Hoffman et al. (2017) for the North Atlantic tropical SST, where two of the three cores
803 reported indicate colder surface waters during the LIG with respect to pre-industrial times.
804 The amplitude of these negative temperature anomalies differs according to the LIG time
805 slices studied. We exclude the beginning of the LIG (129 ka BP), which could be affected by
806 glacial-interglacial transition mechanisms (e.g., Capron et al., 2014). At 125 ka BP, only one
807 core (M12392) exhibits significant cooling of $-3.0 \pm 0.5^\circ\text{C}$. At 120 ka BP, two cores (M12392,
808 V22-196) indicate negative SST anomalies of $-3.0 \pm 0.5^\circ\text{C}$ and $-1.5 \pm 0.5^\circ\text{C}$, respectively. The
809 SST anomalies from these cores are less convincing when the core-top SST value
810 reconstructed for the Holocene is taken into account (Hoffman et al., 2017; Hessler et al.,
811 2014) : the absence of original SST data in the core-top (CLIMAP, 1984) prevents accurate
812 comparison of the SST with respect to the Holocene. For this reason, the few data points that
813 display negative LIG temperature anomalies in the compilation by Hoffman et al. (2017) may
814 not be reliable. Furthermore, they are inconsistent with the warmer surface waters
815 documented by the other investigations in the region, as well as with the results of our present
816 study.

817 Our evidence therefore points to a warming of the SST off Northwest Africa from ≈ 125
818 ka to ≈ 119 -116 ka BP with respect to pre-industrial times. This warming, which only
819 concerns the second part of the LIG, is in principle weaker than that of the first part of the
820 interglacial, where the insolation was maximum. It therefore indicates a lower value of the
821 temperature anomaly for the LIG (129 ka to 116 ka). One possible cause of this warming
822 could have been weakening of the North African upwelling, which would increase the SST.

823 The warming could also be due to the high insolation during the LIG, thereby warming the
824 Canary Current from the North Atlantic gyre.

825

826 6.2.4. Upwelling impact and insolation effect

827

828 To understand the potential impact of climate change on the coastal upwelling activity,
829 further information is needed about how it changed during the different climatic alternations
830 between recorded warming and cooling events in the past (interglacial and glacial periods,
831 respectively). For the Holocene period, the summer insolation in the northern hemisphere
832 slowly decreased, resulting in a slow diminishing of the GMST and a reduction in summer
833 monsoon activity in the northern tropical regions of Africa (deMenocal et al., 2000; Shanahan
834 et al., 2015; Collins et al., 2017), Asia and North America (Wang et al., 2009). The global
835 Earth climate, which was initially warm and humid, became cooler and drier, with a transition
836 at around 5 ka BP. Re-interpretation of two SST proxies at ODP Hole 658C, off Cape Blanc
837 (20°45'N, 18°35'W, south of the Canary Islands, see Figure 1) combined with new data,
838 provided a better understanding of the history of upwelling of North Africa (Adkins et al.,
839 2006). These authors concluded that weaker upwelling activity coupled with stronger African
840 monsoon activity occurred during the African Humid Period in the first part of the Holocene.
841 A similar observation was made by Abrantes (1991) who documented a warming period in
842 the region of the Canary Islands that spanned several millennia in the first part of the
843 Holocene, accompanied by a reduction in the intensity of the northwest African upwelling.
844 During the LIG, such a weakening of the upwelling is also suggested by Muhs et al. (2014)
845 from the faunal record in the Canary Islands. The LIG may have been characterized by early
846 insolation-forced warming along with northward migration of the intertropical convergence
847 zone (ITCZ), and accompanied by weakened trade winds and diminished upwelling. This

848 allowed the arrival of extralimital southern taxa from the tropical Senegalese faunal province.
849 During the latter part of LIG, decreased insolation may have resulted in southward migration
850 of the ITCZ, strengthened trade winds, and re-establishment of the upwelling. Such
851 conditions may have brought about the local extinction of the Senegalese fauna. The change
852 of the upwelling strength can also be documented in the framework of larger climate changes,
853 such as glacial and interglacial oscillations. Those oscillations extend over millennia, the
854 cyclic climate alternation driven by variations in the summer insolation at middle to high
855 latitudes in the northern hemisphere. Although the boundary conditions of the climate in the
856 LGM and the LIG were different, they were both the result of insolation changes, and these
857 modulate atmospheric circulation and hence the upwelling strength. The global warming that
858 took place from the LGM to the Holocene was of the order of +5°C (Masson-Delmotte et al.,
859 2013; Snyder, 2016). Henderiks et al. (2002) estimated the changes in the Canary seasonal
860 coastal upwelling from LGM to the Holocene : they documented stronger trade winds and
861 enhanced productivity during the LGM, in good agreement with previous studies of the
862 northwest African upwelling and trade wind-system (e.g., Sarinthein et al., 1982, 1988;
863 Abrantes, 1991; Rognon and Coude-Gaussen, 1996). Similar observations, but spanning a
864 larger time scale, were made by Freudenthal et al. (2002), who reconstructed changes in trade
865 wind intensities over the last 250 ka at 4 sites close to the Canary Islands. Furthermore, from
866 a sediment core in the Canary Basin, Nave et al. (2003) indicated that the glacial stages MIS4
867 and MIS2 were more productive than the interglacial intervals (LIG and Holocene). Moreno
868 et al. (2002) analysed the primary productivity record from sediment cores over the last 250
869 kyr in the North Canary Basin (30°N). Their work demonstrates that productivity is lower
870 during interglacials than during glacial periods. These observations corroborate the fact that in
871 glacial times the northwest Africa coastal upwelling activity increased. All of the above
872 observations indicate a weakening of upwelling activity during periods of global warming.

873 We therefore conclude that the North African upwelling was weaker during the LIG than
874 during the Holocene. This finding implies warmer coastal seawater during the LIG relative to
875 the Holocene.

876 The maximum increase in SST attributable to a weakened upwelling can be estimated
877 from the present-day isotherms of the eastern Atlantic tropical gyre. Upwelling only decreases
878 the SST along the coast (Figure 3b). According to recent data, in the eastern Canary Islands
879 this cooling ranges from 0.75°C to 1°C (Gómez-Letona et al., 2017 ; Sousa et al., 2017 ;
880 Santos et al., 2012 ; Arístegui et al., 2009 ; Muhs et al., 2014, after Conkright et al., 2002). It
881 follows therefore that a weakening of the upwelling will lead to a warming by between
882 0.75°C and 1°C at the very most, since the cold deep waters never completely stop rising.
883 However, the increase in SST with respect to pre-industrial times, indicated by proxies from
884 the upwelling zone off northwest Africa (excluding the values of Hoffman et al., 2017) is
885 significantly greater. It lies in the range +1°C to +4°C (Figure 13b). As such a temperature
886 rise is too large to be attributed to a weaker upwelling alone, an alternative and additional
887 mechanism, namely the insolation at the LIG, must therefore be examined.

888 *Summer insolation* in the northern hemisphere, the main driving force of the glacial-
889 interglacial oscillations, was more pronounced during the LIG than in the Holocene, leading
890 to an interglacial that was on average warmer (CAPE-Last Interglacial Project Members,
891 2006; Masson-Delmotte et al., 2013; Capron et al., 2014). Between 30°N and 90°N the
892 insolation anomaly 125 ka ago exhibited a maximum in July between $\approx+50$ W/m² and $\approx+70$
893 W/m² with respect to pre-industrial times (Pedersen et al., 2016). These results are in
894 agreement with those of Capron et al. (2017b). Not only the *summer insolation* but also the
895 *annual insolation* anomaly was positive (between 0 and +2.5 W/m²) (Pedersen et al., 2016).
896 By contrast, the *annual insolation* anomaly observed in the tropical latitude belt between
897 30°N and 30°S displayed a slight decrease comprised between 0 and -0.2 W/m² (Pedersen et

898 al., 2016). It is to this *annual insolation* decrease in the tropical latitudes that Hoffmann et al.
899 (2017) attributed their estimated cooling of the SST in the tropical Atlantic. According to the
900 latter authors, the simulations of Pedersen et al. (2016) at 125 ka “successfully simulate the
901 weak cooling found in our tropical SST stack in response to reduced mean *annual insolation*
902 at those latitudes”. However the simulations performed by Pedersen et al. (2016) revealed
903 *annual cooling* of the air temperature only in the continental tropical zones of North Africa
904 (5°N-20°N) and Asia (India). The tropical north Atlantic not only does not exhibit cooling but
905 rather a *warming* of between +0.5°C and +3°C, and in particular a warming of between +1
906 and +3°C in its north-eastern part from 22°N to 30°N with respect to pre-industrial times.
907 Pedersen et al. (2016) attribute the continental cooling to the impact of *summer warming* in
908 the Northern hemisphere, which amplified the North African and Indian monsoons. Owing to
909 this amplification, evaporation and spread of the cloud cover became more pronounced,
910 consequently cooling the continental surface. The continental cooling was thus mainly caused
911 by the increase in seasonal *summer insolation*, rather than being due to a reduction of the
912 *annual insolation* in the tropics. Accordingly, in spite of the reduced *annual insolation* that
913 took place in the tropics during the LIG, the simulation of Pedersen et al. (2016) did not
914 reproduce the cooling indicated by Hoffman et al. (2017) in the north-east tropical Atlantic.
915 On the contrary, it reveals warming of the surface water in the northern tropical Atlantic, in
916 agreement with our own measurements as well as with the majority of compilations
917 conducted in this zone. These results lend support to the impact of summer insolation in the
918 Northern hemisphere on the SST of the Canary current water masses during the LIG. The
919 warming by about +1°C to +4°C (Figure 13b) of the SST in the tropical northeast Atlantic,
920 reported by all the data sets, apart from those reconstituted by Hoffman et al. (2017), thus
921 stemmed principally from the excess *summer insolation* during the LIG and, to a lesser
922 degree, from a weakening of the upwelling. Therefore this indicates a warming of the

923 descending branch of the North Atlantic Gyre (Canary current), mainly engendered by the
924 larger summer insolation. As the descending branch is part of the gyre, this implies that the
925 warming extended over the whole of the gyre.

926

927 6.3 Seasonal variations in paleo-SST and paleo-Dissolved Inorganic Carbon (DIC) $\delta^{13}\text{C}$

928

929 6.3.1. Seasonal SST during the LIG sea-level highstand

930

931 According to the SST inferred from incremental sampling of two specimens with a
932 lifespan of a few years, which are the gastropod *Stramonita haemastoma* sampled from Matas
933 Blancas, Fuerteventura, and the bivalve *Cardium edule* (*Cerastoderma edule*) sampled from
934 La Santa, Lanzarote, the pseudo-cyclic SST pattern observed along the growth direction of the
935 shells can be attributed to the seasonality of the SST. Those two mollusc shells recorded
936 seasonal SST variations ranging from $\approx 3.5^\circ\text{C}$ to $\approx 6.0^\circ\text{C}$, with a mean value of about
937 $4.8 \pm 1.2^\circ\text{C}$, during the LIG sea-level highstand (≈ 125 to ≈ 119 - 116 ka BP) in the eastern
938 Canary archipelago. Other LIG seasonal SST values for the northeast Atlantic region may be
939 deduced from the difference between the winter mean SST (JFM) and the summer mean SST
940 (JAS) reconstructed from foraminifera assemblages, by a method that is complementary to
941 our approach based on stable oxygen isotope compositions of carbonates. Indeed, Matsuzaki
942 et al. (2011) estimated seasonal SST variations to be about $5.0 \pm 1.2^\circ\text{C}$ during the second part
943 of the LIG, from ≈ 122 to ≈ 118 ka BP, from the core MD03-2705 (Figure 3b) located in the
944 northeast of Cape Verde Islands ($18^\circ 05.81' \text{N}$, $21^\circ 09.19' \text{W}$). From the cores M12392-1 south
945 of the Canary archipelago ($25^\circ 09.60' \text{W}$, $16^\circ 51.00' \text{W}$) and V22-196 between the Cape Verde
946 Islands and Senegal ($13^\circ 49.80' \text{N}$, $18^\circ 58.20' \text{W}$) (Figure 3b) (CLIMAP Project Members,
947 1984), seasonal SST variations were estimated to be $\approx 5.0 \pm 2.4^\circ\text{C}$ and $\approx 5.5 \pm 2.6^\circ\text{C}$ during the

948 LIG, respectively. The core top of those two cores, which is not available, is not necessary for
949 these calculations. All these LIG seasonal SST estimates in the northeast Atlantic region
950 match our values.

951 Meco et al. (2018) reported a seasonal SST amplitude along the eastern coast of
952 Fuerteventura ranging from 3.5°C to 6.7°C between 2007 and 2016. The seasonal temperature
953 variations from $\approx 3.5^\circ\text{C}$ to $\approx 6.0^\circ\text{C}$ recorded by the two mollusc shells, *Stramonita*
954 *haemastoma* and *Cardium edule* (*Cerastoderma edule*), during the LIG sea-level highstand in
955 the eastern Canary Islands are therefore comparable to those of modern times. Finally, our
956 data obtained from mollusc samples and those coming from the cores MD03-2705, M12392-1
957 and V22-196 indicate that the SST seasonality of the northeast Atlantic waters was of the
958 same order during the LIG as that of present-days.

959

960 6.3.2. Mean annual and seasonal DIC $\delta^{13}\text{C}$ during the LIG sea-level highstand

961

962 Coastal upwelling areas are important dynamic parts of the ocean carbon cycle. The
963 bulk carbon isotope data for La Santa and Matas Blancas sites provide a mean $\delta^{13}\text{C}$ value of
964 $2.46 \pm 0.57 \text{ ‰}$ (Table S1, in the Supplementary materials). This value represents an average
965 of 30 mollusc shells from a LIG sedimentary deposit that extends over several thousand years,
966 each sample being characterized by a multiannual $\delta^{13}\text{C}$ value. Several authors (e.g.
967 Mackensen et al., 1993; Rohling and Cooke, 1999; Maslin and Swann, 2005) have shown that
968 the $\delta^{13}\text{C}$ value of biologically precipitated carbonate in seawater is very close to that of
969 Dissolved Inorganic Carbon (DIC). Therefore the $\delta^{13}\text{C}$ values recorded in the aragonite
970 mollusc shells should represent that of seawater DIC. As emphasized by Maslin and Swan
971 (2005), the $\delta^{13}\text{C}$ values of planktonic carbonate foraminifera increase with higher biological
972 productivity. This is due to isotope fractionation during the photosynthesis process where the

973 lighter $^{12}\text{CO}_2$ is preferentially selected by the organic matter, thereby leaving the surface water
974 DIC enriched in ^{13}C . Thus, the positive mean annual $\delta^{13}\text{C}$ values of carbonate shells from the
975 eastern Canary archipelago reflect a high marine productivity of surface waters during the
976 sea-level highstand, i.e., from ≈ 125 ky to 119-116 ky BP (Maslin and Swan, 2005).

977 The mollusc shells selected for isotopic ontogenetic profiles provide complementary
978 information at the seasonal scale of the $\delta^{13}\text{C}$ values of DIC. We consider that the $\delta^{13}\text{C}$
979 fluctuations recorded in the aragonite mollusc shell represent the $\delta^{13}\text{C}$ fluctuations of seawater
980 DIC. $\delta^{13}\text{C}$ values of the incrementally sampled shell of *Cardium edule* (*Cerastoderma edule*)
981 sampled from La Santa sedimentary sequence show a pseudo-cyclic signal recorded over a
982 few years, with $\delta^{13}\text{C}$ values ranging from 0.51 ‰ to 3.02 ‰ (Table S2, in the Supplementary
983 materials). These values, obtained from one sample, range around a mean value
984 (2.04 ± 0.74 ‰) similar to the mean annual $\delta^{13}\text{C}$ averaged over several thousands of years
985 during the LIG (2.46 ± 0.57 ‰), which indicates highly productive marine waters. The $\delta^{13}\text{C}$
986 values of the incremental samples from the *Cardium edule* (*Cerastoderma edule*) shell are
987 correlated with those of $\delta^{18}\text{O}$, the regression line having a positive slope of 0.73 and a
988 regression coefficient R^2 of 0.51 (Figure 9). At the seasonal scale, these observations mean
989 that the lowest shell $\delta^{13}\text{C}$ values correspond to the higher SST values during LIG (Figure S2,
990 in the Supplementary materials). In upwelling areas, the superficial upwelled seawater is
991 characterized by ^{13}C -depleted DIC generated by oxidation of organic matter. This is due to the
992 fact that after the death of phytoplankton, the ^{12}C -enriched organic matter sinks into the deep
993 ocean and is degraded and remineralized. As a result, upwelled deep waters are characterised
994 by ^{13}C -depleted DIC (Peeters and al., 2002). The low values of shell $\delta^{13}\text{C}$ therefore indicate an
995 enhanced seasonal upwelling activity. Owing to the recorded correlation between the shell
996 $\delta^{13}\text{C}$ values and the SST, the LIG enhanced upwelling activity corresponded to high SST
997 periods, i.e., to summer seasons (Meco et al., 2018). A more active upwelling in summer

998 during the LIG sea-level highstand is consistent with the present enhanced upwelling during
999 summer (Pardo et al., 2011; Navarro-Pérez and Barton, 2001; Mittelstaedt, 1991) (see §2.2).
1000 In one of the most active upwelling areas, the Peruvian upwelling, Sadler et al. (2012) noted
1001 the same relationship between $\delta^{13}\text{C}$ DIC recorded by fossil shell $\delta^{13}\text{C}$ values and the upwelled
1002 water season, i.e., low shell $\delta^{13}\text{C}$ values corresponding to maximum upwelling intensity.
1003 However, unlike the upwelling off western Africa, that off Peru is strengthened in winter, and
1004 accordingly, lower values of shell $\delta^{13}\text{C}$ correspond to lower SSTs. We note that as the Canary
1005 Current SSTs are highest during summer (Meco et al., 2018), the main factor controlling the
1006 seasonal variation of SST is the insolation. The thermal impact of upwelling, which decreases
1007 the SST particularly when it is active in summer, however, is only of second order.

1008 The seasonal influence of the Canarian upwelling seems to be only recorded in La
1009 Santa, Lanzarote, and not in Matas Blancas, Fuerteventura, a difference that may be explained
1010 by the geographical location of these two sites. Lanzarote and Fuerteventura are both situated
1011 at the eastern end of the archipelago where the global influence of the upwelling is the
1012 strongest. However, La Santa lies on the western side of the island of Lanzarote while Matas
1013 Blancas is located east of the island of Fuerteventura. Local upwelling always takes place on
1014 the western sides of lands due to the westerly direction of the Trade Winds. The reconstitution
1015 of the pigment content of the surface seawater from satellite recordings (Nave et al., 2001,
1016 2003; Borges et al. 2004), which is a signature of the primary oceanic productivity, indeed
1017 exhibits a local annual coastal upwelling on the western coasts of Lanzarote and
1018 Fuerteventura, and not along the eastern coasts.

1019 Our results suggest seasonal variations in the oceanic surface water parameters of
1020 the Canary archipelago during the LIG related to the SST and to the North African coastal
1021 upwelling activity. These seasonal fluctuations resemble those of the present time.

1022

1023 7. Conclusion

1024

1025 The stable isotope study of aragonitic mollusc shells sampled from the sites of La
1026 Santa, Lanzarote, and Matas Blancas, Fuerteventura, in the eastern part of the Canary
1027 archipelago reveals several important findings about the climate during the Last Interglacial,
1028 the LIG :

1029 (i) The sample collection investigated is composed of mollusc shells deposited
1030 during the sea-level highstand that occurred during the second part of LIG. Our sample
1031 collection is therefore contemporaneous with the second part of the LIG, and is dated between
1032 ≈ 125 and $\approx 119-116$ ky BP.

1033 (ii) From 30 mollusc shell $\delta^{18}\text{O}$ values, SST comprised between $20.4\pm 1.3^\circ\text{C}$ and
1034 $22.2\pm 1.2^\circ\text{C}$ is estimated during the LIG record in the eastern Canary marine sedimentary
1035 sequences. Two mollusc shells, sampled from La Santa and at Matas Blancas in order to
1036 obtain ontogenetic profiles, provide a seasonal SST amplitude comprised between about
1037 3.5°C and 6°C , similar to that of the present. The $\delta^{13}\text{C}$ variations recorded within one of these
1038 two mollusc shells, reflecting those of seawater DIC, are correlated with the SST and indicate
1039 seasonal variations in the intensity of the upwelling off the African coast similar to those of
1040 the present.

1041 (iii) According to the modern SST documented in the eastern part of the Canary
1042 archipelago and the warming gradient established in this zone, we deduce an average SST
1043 value of about $19.4\pm 0.1^\circ\text{C}$ during the pre-industrial times. This conclusion enables us to
1044 propose a SST anomaly ranging from $+1.0\pm 1.4^\circ\text{C}$ to $+2.8\pm 1.3^\circ\text{C}$ during the second half of
1045 the LIG with respect to pre-industrial times. This warming, which concerns only the second
1046 part of the LIG, is in principle weaker than that of the first part of the interglacial (≈ 129 ka to

1047 ≈ 125 ka), where the insolation was maximum and greater than in the Holocene. It therefore
1048 indicates a lower value of the temperature anomaly for the LIG (≈ 129 ka to ≈ 116 ka).

1049 (iv) This positive anomaly is consistent with the composition of the faunal species
1050 recorded in the marine highstand deposit with a warm extra-limital southern species
1051 (*Persististrombus latus* or *Strombus bubonius*), typical of warm sea water, which shows that
1052 at the time of the LIG marine sediment deposits it was warmer than in the present interglacial,
1053 as already documented by Muhs et al. (2014). In fact, the SST recorded by the specimens of
1054 *Persististrombus latus* is definitely higher ($\approx +1.2^\circ\text{C}$) than that recorded by all 30 mollusc
1055 shells. The co-occurrence in the deposit of cooler species, nowadays present in the Canary
1056 Islands, illustrates the SST cooling that set in at the end of the LIG.

1057 (v) This positive anomaly is also consistent with a global mean surface temperature
1058 during the LIG that was warmer than in the Holocene (a warming that we estimate to be
1059 $\approx +2^\circ\text{C}$ in the first part of the LIG (≈ 130 ka BP), followed by $\approx +1^\circ\text{C}$ in the second part
1060 (≈ 126 to ≈ 119 ky BP). By contrast, it does not accommodate with the zonal negative
1061 anomaly of reconstructed SST (cooler temperature) at low latitudes in the North Atlantic
1062 (Hoffman et al., 2017). The positive anomaly is nonetheless fully consistent with all the other
1063 existing regional reconstructions (Sicre et al., 2000; Matsuzaki et al., 2011; Castañeda et al.,
1064 2009; Muhs et al., 2014; Montesinos et al., 2014), as well as with the regional simulations of
1065 air temperature (Pedersen et al., 2016). This finding calls into question the reconstructed zonal
1066 cooling at low latitude in the North Atlantic proposed by Hoffman et al. (2017) as well as data
1067 in other studies that may be based on this work (e.g., Fisher et al., 2018).

1068 (vi) We interpret the positive SST anomaly as resulting principally from warming of
1069 the descending branch of the North Atlantic Gyre (Canary current) due to the excess summer
1070 insolation during the LIG with respect to the Holocene and, to a lesser degree, from a

1071 weakening of the upwelling off the African coast. As the Canary current is part of the North
1072 Atlantic Gyre, this implies a warming that extended over the whole of the gyre.

1073

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1078

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1467

1468 **Tables captions**

1469 Table 1 : Present sea surface temperature (SST in °C) in the eastern Canary Islands. (*)

1470 Values and standard errors estimated from color codes indicated on the figures.

1471

1472 Table 2: Period of sea-level (SL) highstand(s) during the LIG, together with the sea-level
1473 anomaly relative to the modern period.

1474

1475 Table 3 : Different studies reconstructing the SST during the LIG in the northeast tropical

1476 Atlantic.

1477

1478 **Figure captions**

1479 Figure 1: (a) Location of the Canary Islands and of the Canary Current (blue arrows). (b) Map
1480 of eastern Canary Islands and sites of the marine deposits from the Last Interglacial period
1481 (red circles) including the LIG marine terrace deposits studied here at La Santa, Lanzarote,
1482 and Matas Blancas, Fuerteventura (yellow squares). Sites of coastal water samplings in the
1483 eastern Canary Islands are indicated by blue stars (Clauzel et al., 2019). LPG = Las Palmas de
1484 Gran Canaria, MB = Matas Blancas, LP = Las Playitas, LG = La Guirra, LS = La Santa, PP =
1485 Punta Penedo, M = Matagorda, NGC = North Gran Canaria, NFV = North Fuerteventura,
1486 WLZ = West Lanzarote.

1487

1488 Figure 2 : (a) Stratigraphy of cliff sections at Punta Penedo site, a few kilometers north of the
1489 site at La Santa, Lanzarote, studied here, showing the most complete section in the area
1490 (adapted from Meco et al., 2006). The sedimentary deposits are characterized by a succession
1491 of four distinct climatic phases: (i) calcareous dune deposits, (ii) paleosols (with locust egg
1492 pods), (iii) marine sediments, and (iv) pedogenic calcretes (see text). (b) Marine deposits at
1493 the studied site at Matas Blancas (Fuerteventura) (Muhs et al., 2014) with well-preserved
1494 fossil specimens of *Persististrombus latus* (*Strombus bubonius*).

1495

1496 Figure 3 : (a) Annual mean sea surface temperature (SST in °C) in the North Atlantic Ocean
1497 from Pathfinder satellite (adapted from Darfeuil et al., 2016, data from Armstrong and
1498 Vazquez-Cuervo, 2001). The surface circulation patterns are indicated by black arrows. AzC
1499 = Azores Current, CC = Canary Current, PC = Portugal Current. (b) SST (°C) in the northeast
1500 tropical Atlantic Ocean on 25 July 2007 from OSTIA satellite (adapted from Aristegui et al.,
1501 2009, data from Stark et al., 2007). Studies reporting a LIG SST record and discussed in the
1502 present work are indicated by filled circles (sediment cores) and by empty circles (marine
1503 deposits). Black arrows correspond to Trade Winds (from Nave et al., 2003).

1504

1505 Figure 4: Photographs of a selection of shells of the 6 studied molluscan species : gastropods
1506 (a) *Cerithium vulgatum*, (b) *Stramonita haemastoma*, (c) *Luria Lurida*, (d) *Persististrombus*
1507 *latus* (*Strombus bubonius*) and bivalves (e) *Loripes lacteus*, (f) *Cardium edule* (*Cerastoderma*
1508 *edule*). The length of a rectangle corresponds to 1 cm.

1509

1510 Figure 5: Modern geographical distribution of the mollusc species studied in the la Santa,
1511 Lanzarote, and Matas Blancas, Fuerteventura, marine LIG deposits of the Canary Islands,

1512 from south (red) to north (blue) of the archipelago (orange). The solid lines indicate the
1513 observed presence of the molluscs, the dashed lines their presumed presence.

1514

1515 Figure 6: (a) Sediments in the Canary Islands from the paleosols, corresponding to the
1516 African Humid Period, and from the marine deposits during the Holocene (Meco et al., 2011);
1517 (b) Antarctic temperature anomaly records ($^{\circ}\text{C}$) from δD measurements of the ice at EPICA
1518 Dome C during the last 30 ky and (c) in the period 140-100 ky BP (Jouzel et al., 2007); (d)
1519 Sea-level variation (m) reconstructed from far-field coral and sediment data during the last 30
1520 ky (grey circles) (Lambeck et al., 2014); (e) Sea-level variation (m) in the period 140-100 ky
1521 BP, reconstructed from combined field observations and glacial isostatic adjustment modeling
1522 (orange bar) (Dutton et al., 2015a) and from benthic foraminifera oxygen isotope ratios (grey
1523 curves) (adapted from Waelbroeck et al., 2002). The red and blue vertical lines correspond to
1524 the moment when the maximum temperature and the maximum sea-level, respectively, were
1525 reached during the interglacials. The green arrows indicate the lapse of several thousand years
1526 between these two moments. The blue dashed vertical line indicating the beginning of the
1527 highstand and the blue dashed curves describing the sea-level correspond to our interpretation
1528 (see text).

1529

1530 Figure 7: Raman spectra of selected mollusc shells sampled from La Santa, Lanzarote ((a)
1531 gastropod *Cerithium vulgatum* LAS-BULK-3 and (b) bivalve *Cardium edule* (*Cerastoderma*
1532 *edule*) LAS-INCR-14) and from Matas Blancas, Fuerteventura ((c) gastropod
1533 *Persististrombus latus* (*Strombus bubonius*) MAB-BULK-1 and (d) bivalve *Stramonita*
1534 *haemastoma* MAB-INCR-4). All these shells are made of aragonite that is characterized by
1535 the presence of relative intense peaks in the 150–250 cm^{-1} range, and a splitting of the

1536 carbonate group bending mode yielding a doublet at 700–701 cm⁻¹. The peak located at 1080–
1537 1082 cm⁻¹ is characteristic of the carbonate group stretching mode.

1538

1539 Figure 8: $\delta^{18}\text{O}$ (‰ V-PDB) of aragonitic shell increments of *Cardium edule* sampled from La
1540 Santa, Lanzarote, as a function of the distance from the umbo (cm) along the growth
1541 direction.

1542

1543 Figure 9: $\delta^{13}\text{C}$ (‰ V-PDB) as a function of $\delta^{18}\text{O}$ (‰ V-PDB) of aragonitic shell increments of
1544 (a) *Cardium edule* sampled from La Santa, Lanzarote and (b) *Stramonita haemastoma*
1545 sampled from Matas Blancas, Fuerteventura. (a) The regression line (red) is shown with its
1546 regression equation and coefficient ($R^2 = 0.51$).

1547

1548 Figure 10: Frequency histogram of reconstructed LIG SST from bulk mollusc shells sampled
1549 from La Santa, Lanzarote, and Matas Blancas, Fuerteventura, according to Equation (1)
1550 determined by Grossman and Ku (1986) (blue, on the left-hand side) and to Equation (2)
1551 determined by Kim et al. (2007b) (green, on the right-hand side).

1552

1553 Figure 11: Reconstructed LIG SST during the ontogenetic evolution of *Stramonita*
1554 *haemastoma* (Matas Blancas, Fuerteventura) according to Equation (1) determined by
1555 Grossman and Ku (1986) (blue) and to Equation (2) determined by Kim et al. (2007b)
1556 (green). The total length to apex is 16.0 cm.

1557

1558 Figure 12: Reconstructed LIG SST during the ontogenetic evolution of *Cardium edule* or
1559 *Cerastoderma edule* (La Santa, Lanzarote) according to Equation (1) determined by

1560 Grossman and Ku (1986) (blue) and to Equation (2) determined by Kim et al. (2007b)
1561 (green). The total length to umbo is 4.5 cm.

1562

1563 Figure 13 : SST (°C) anomalies in the northeast tropical Atlantic relative (a) to core top values
1564 and (b) to the pre-industrial times. (1): Sicre et al. (2000), (2): Matsuzaki et al. (2011), (3):
1565 Castañeda et al. (2009), (1*) (2*) (3*): The same studies with SST anomalies estimated for
1566 the pre-industrial times (see text), (4): This study, (5): Adapted from Montesinos et al. (2014)
1567 (see text), (7), (8), (9): Hoffman et al. (2017) with cores GIK15637-1 (7), M12392-1 (8),
1568 V22-196 (9). See Figure 3(b) for the sites location.

1569

1570 **Supplementary material captions:**

1571

1572 **Table captions**

1573

1574 Table S1: Oxygen and carbon isotope compositions of bulk marine mollusc shells, listed
1575 together with calculated sea surface temperatures (SST) according to the oxygen isotope
1576 fractionation equations of Grossman and Ku (1986) (Eq.(1)) and Kim et al. (2007b) (Eq.(2)).
1577 Samples collected from La Santa, Lanzarote (LAS-BULK-1 to LAS-BULK-27) and from
1578 Matas Blancas, Fuerteventura (MAB-BULK-1 to MAB-BULK-3).

1579

1580 Table S2: Oxygen and carbon isotope compositions of marine mollusc shell increments and
1581 calculated sea surface temperatures (SST) according to the oxygen isotope fractionation
1582 equations of Grossman and Ku (1986) (Eq.(1)) and Kim et al. (2007b) (Eq.(2)). The
1583 *Cardium edule* specimen was collected from La Santa, Lanzarote (LAS-INCR-1 to LAS-

1584 INCR-16) and the *Stramonita haemastoma* specimen from Matas Blancas, Fuerteventura
1585 (MAB-INCR-1 to MAB-INCR-29). Note: MAB-INCR-20 is missing.

1586

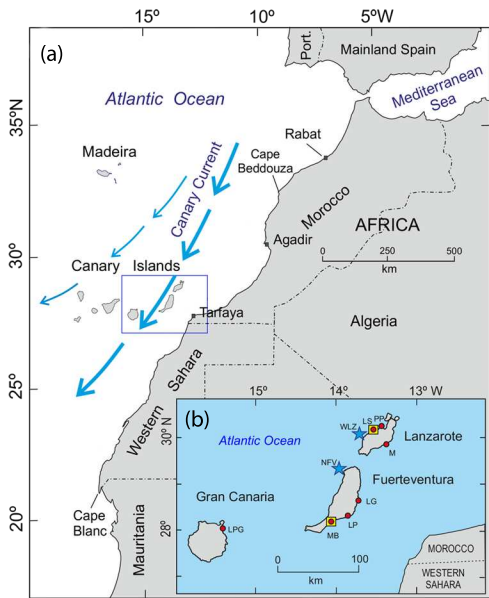
1587 **Figure captions**

1588

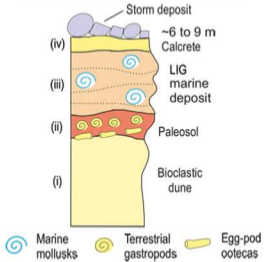
1589 Figure S1: (a) Cross-section of a bivalve shell showing the growth direction of the shell from
1590 the umbo (adapted from Füllenbach et al., 2015). (b) Photomicrograph of the bivalve *Cardium*
1591 *edule* (*Cerastoderma edule*) with the incremental samplings carried out along the growth axis
1592 (drilling holes). (c) Diagram of a gastropod shell showing the growth spiral of the shell from
1593 the apex (frontal view). (d) Photomicrograph of the gastropod *Stramonita haemastoma* with
1594 the incremental samplings, represented by grey circles, made along the growth spiral. The
1595 length of a rectangle corresponds to 1 cm.

1596

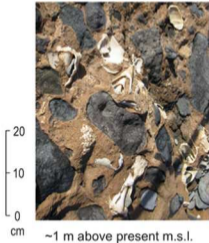
1597 Figure S2: $\delta^{13}\text{C}$ (‰ V-PDB) as a function of reconstructed SST (°C) of aragonitic shell
1598 increments of *Cardium edule* sampled from La Santa, Lanzarote. The regression line (red) is
1599 shown with its regression equation and coefficient ($R^2 = 0.51$). High SSTs occur in summer
1600 (Meco et al., 2018) a season where the upwelling activity is the strongest (Pardo et al., 2011;
1601 Navarro-Pérez and Barton, 2001; Mittelstaedt, 1991).

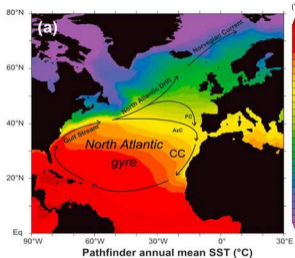


(a) Punta Penedo (close to La Santa)



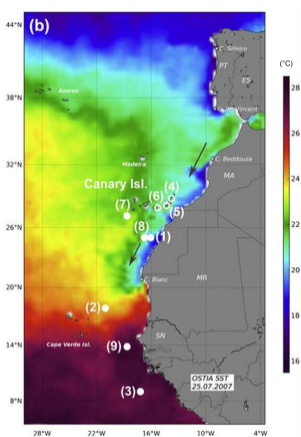
(b) Matas Blancas





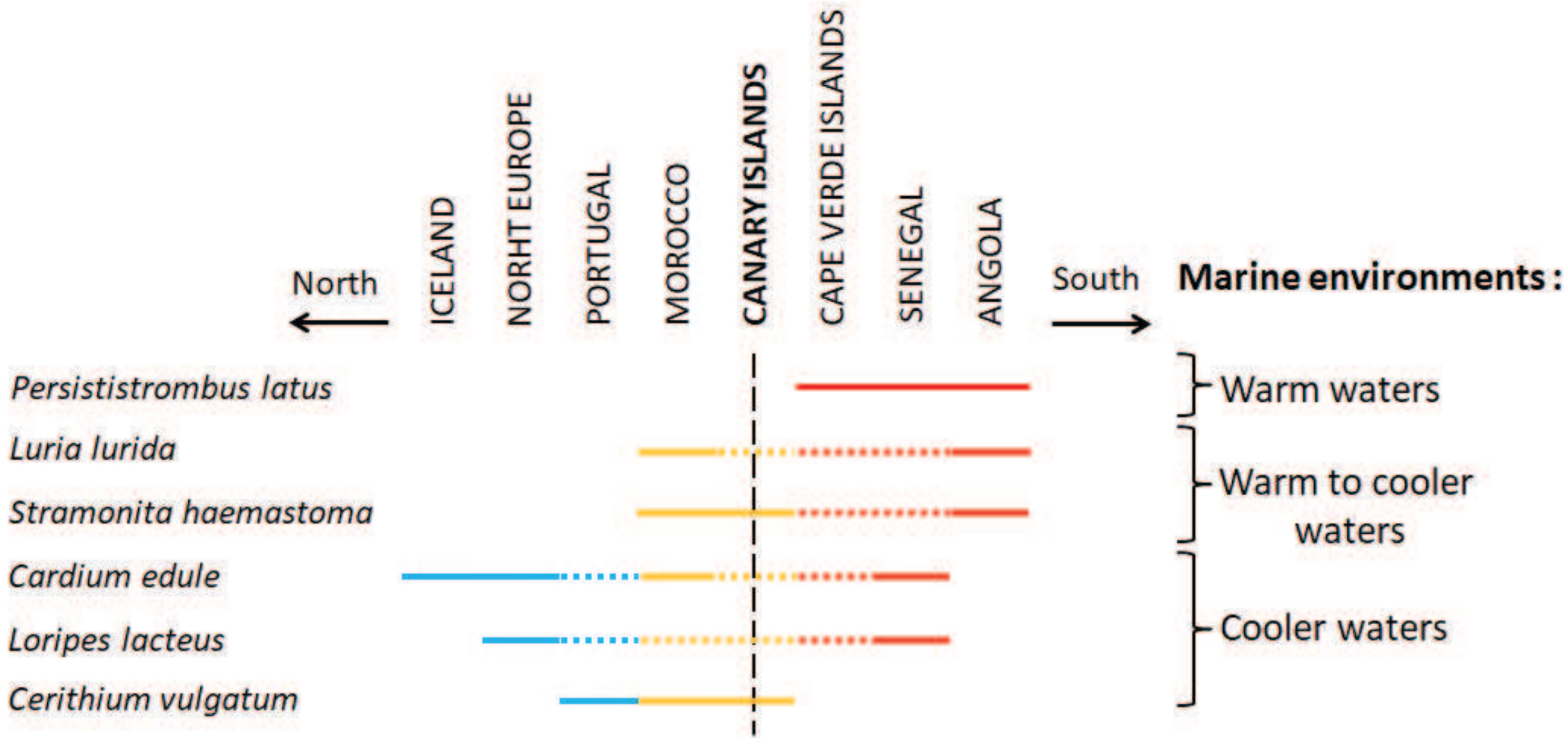
(b)

- | | |
|--|-----------------|
| (1) Sicre et al. (2000) | (7) GIK15637-1, |
| (2) Matsuzaki et al. (2011) | (8) M12392-1, |
| (3) Castaneda et al. (2009) | (9) V22-196, |
| (4) This study | |
| (5) Montesinos et al. (2014) | |
| (6) Muhs et al. (2014) | |
| (7), (8), (9) in Hoffman et al. (2017) | |





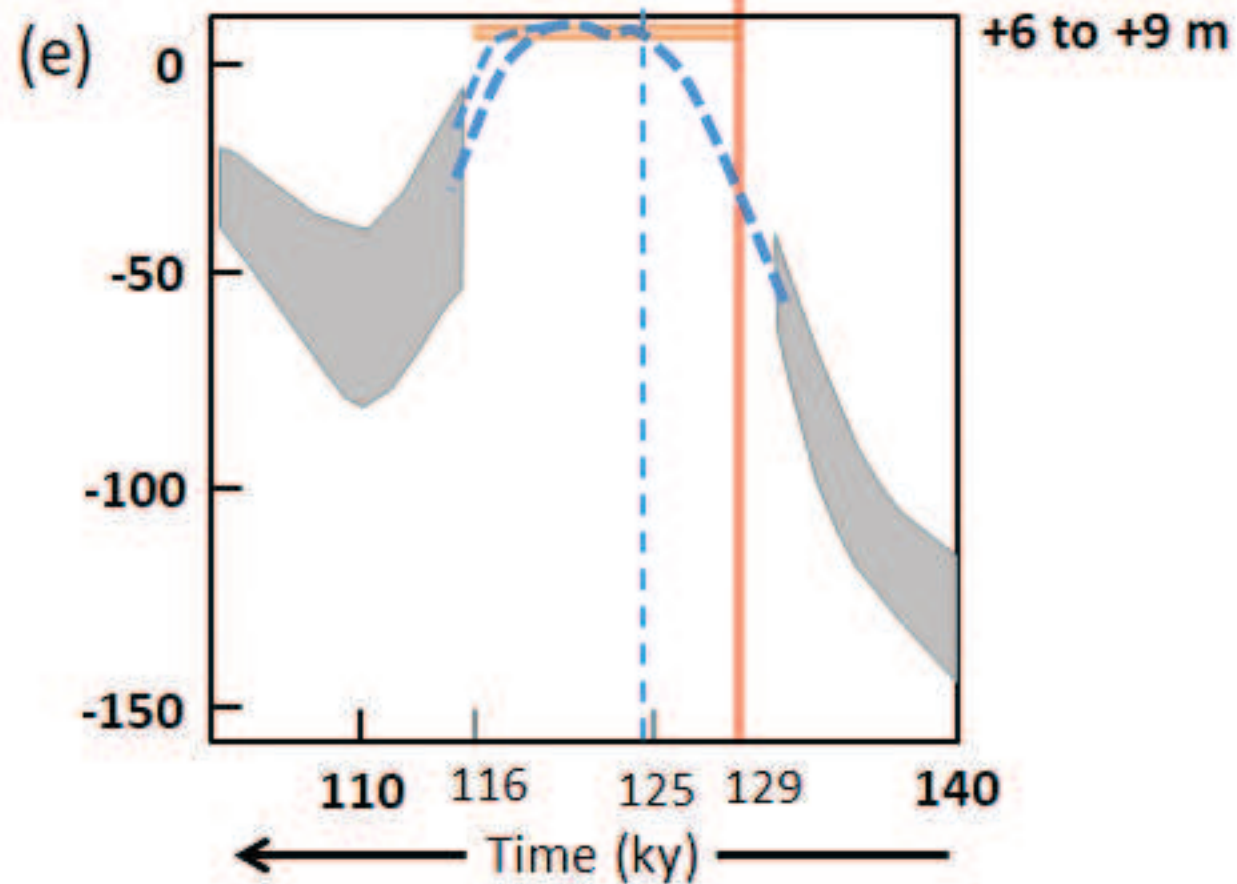
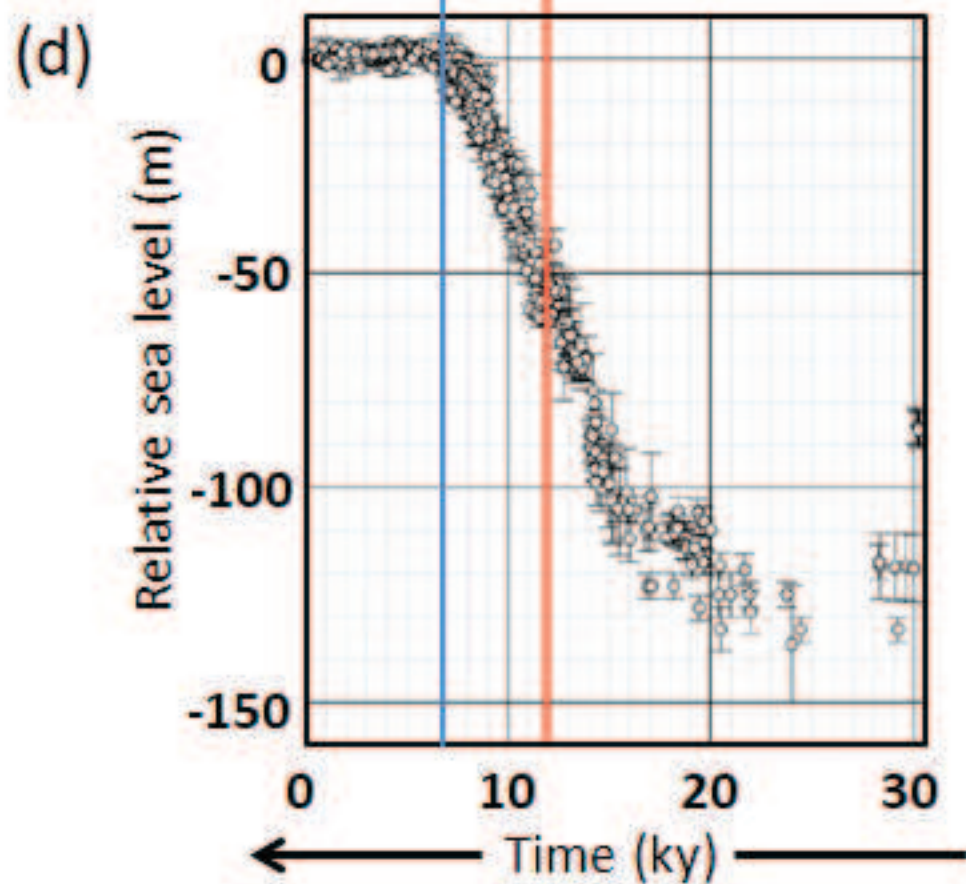
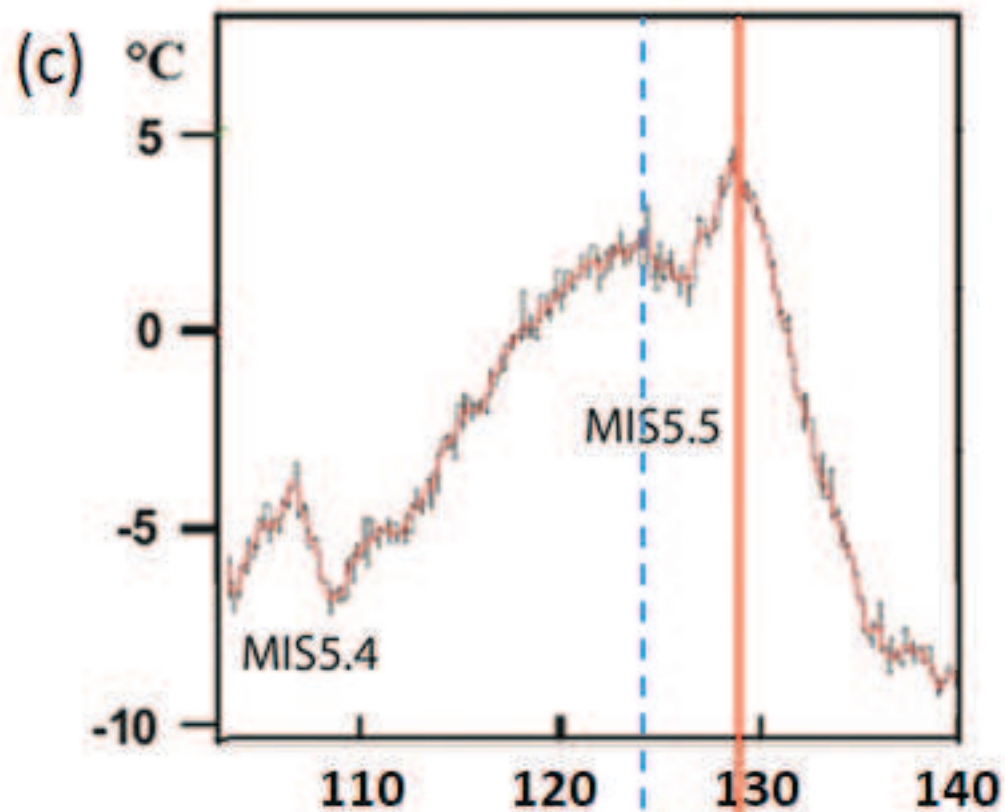
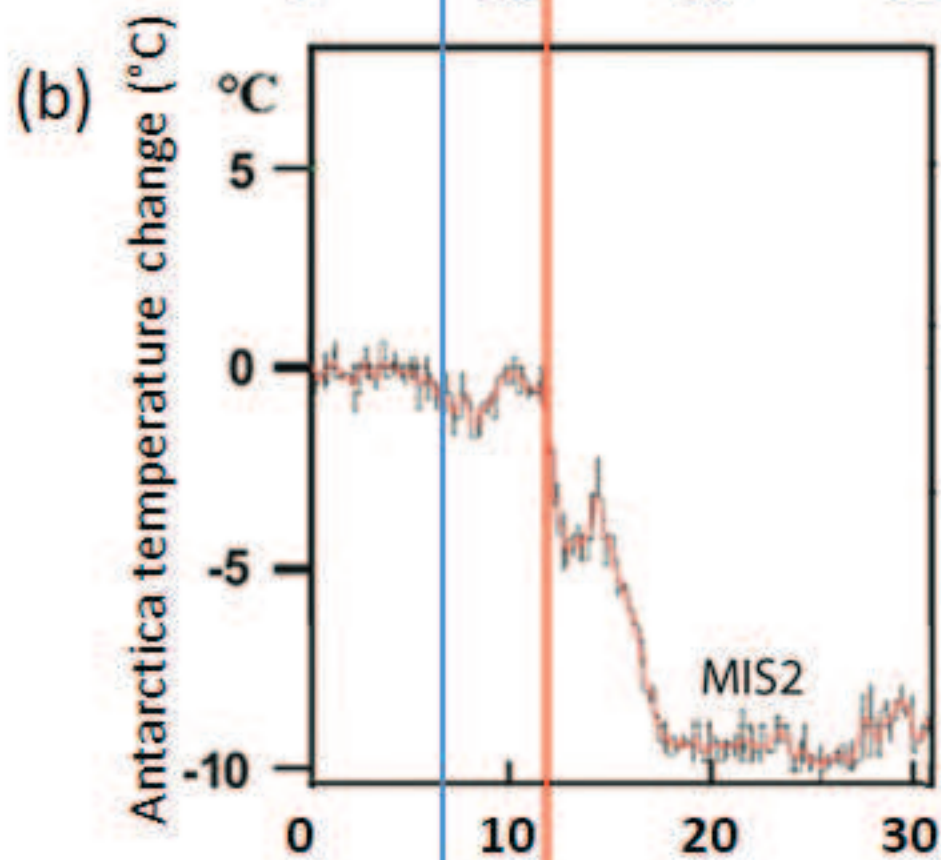
Modern geographical distribution :

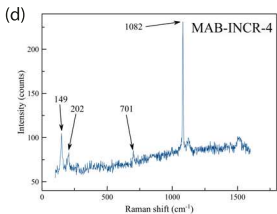
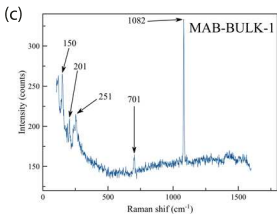
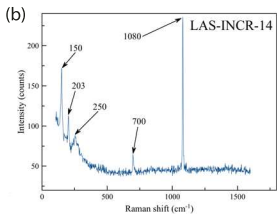
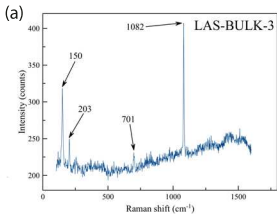


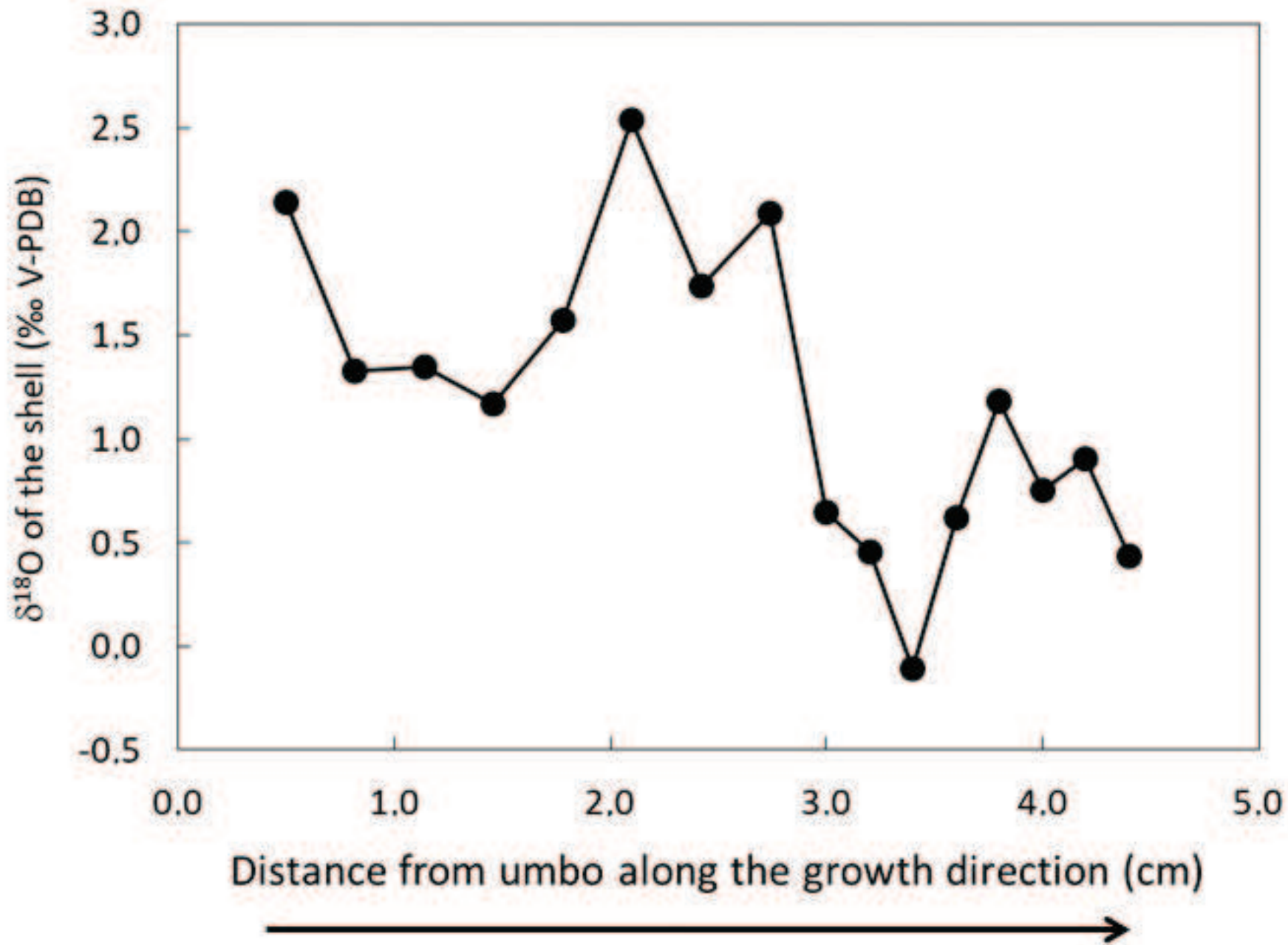
(a) Canary Island sediments



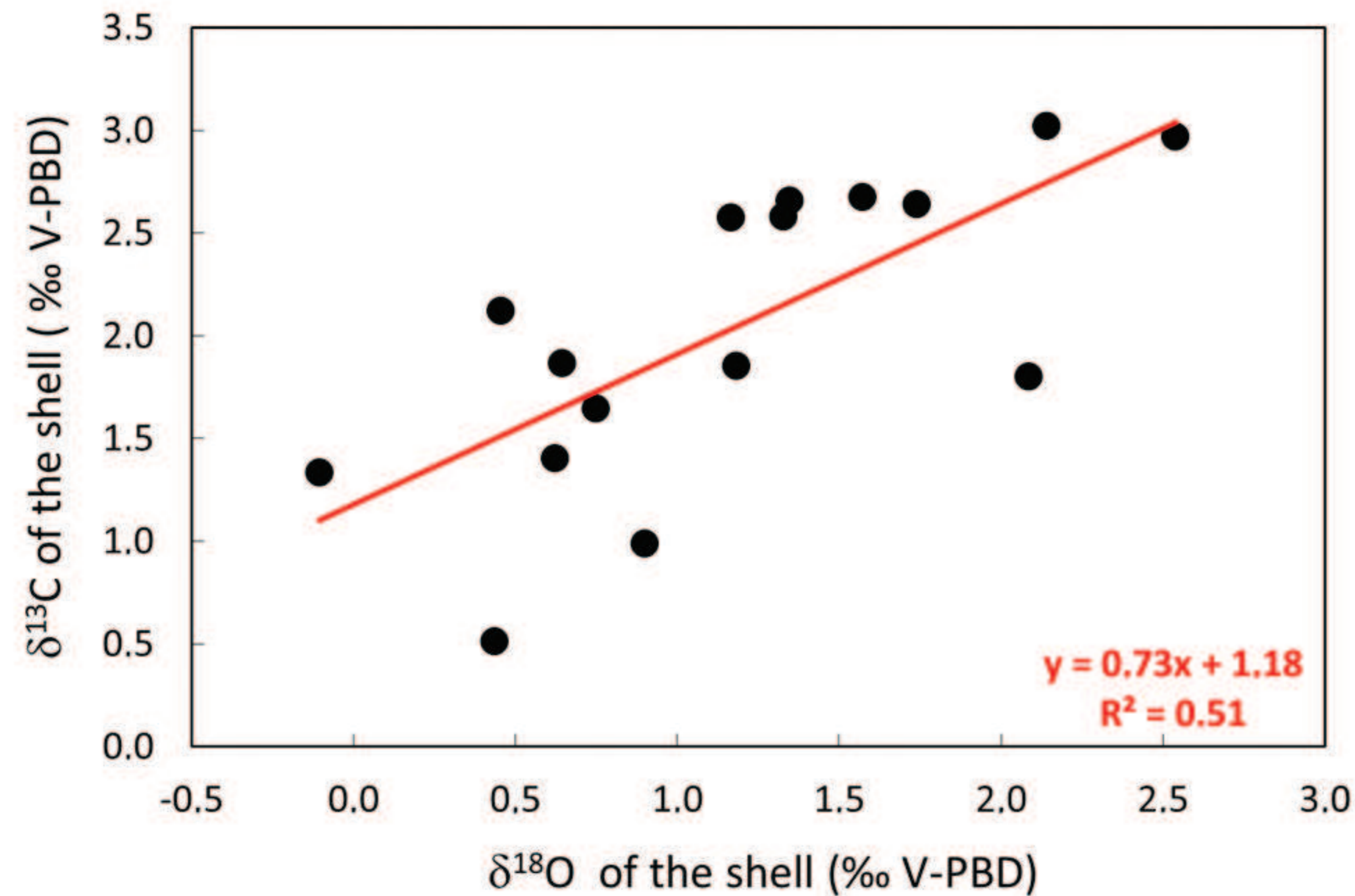
Marine deposits
Paleosols



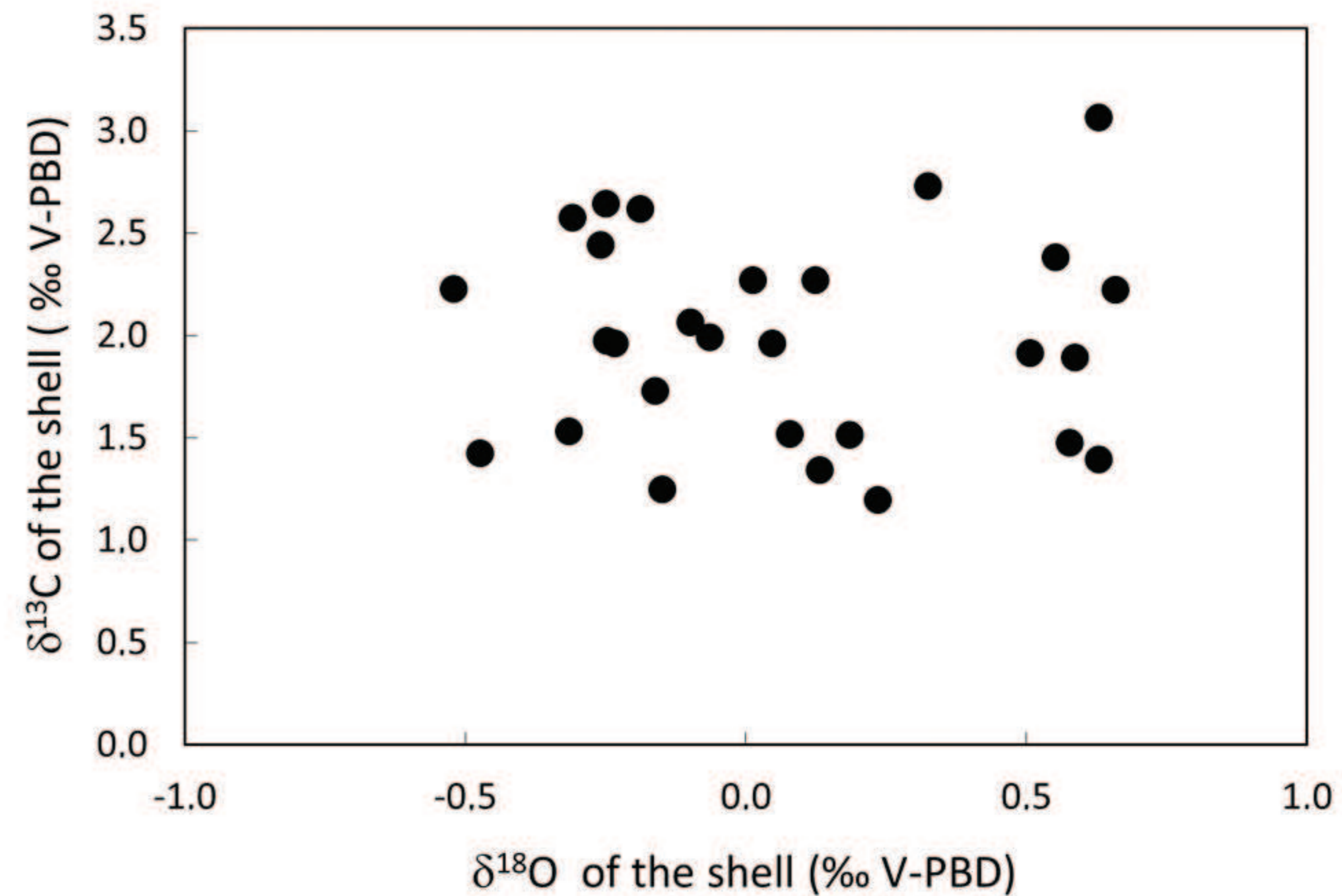


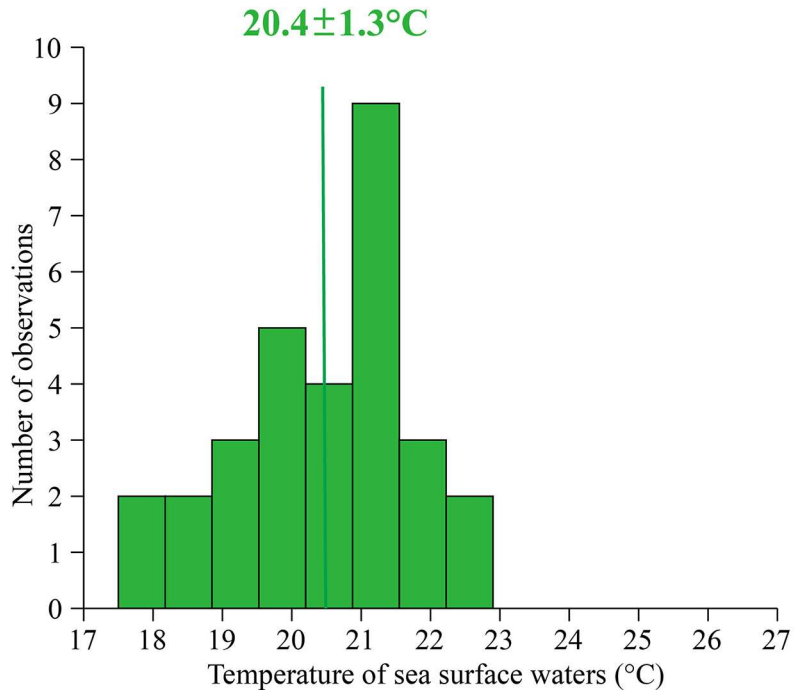
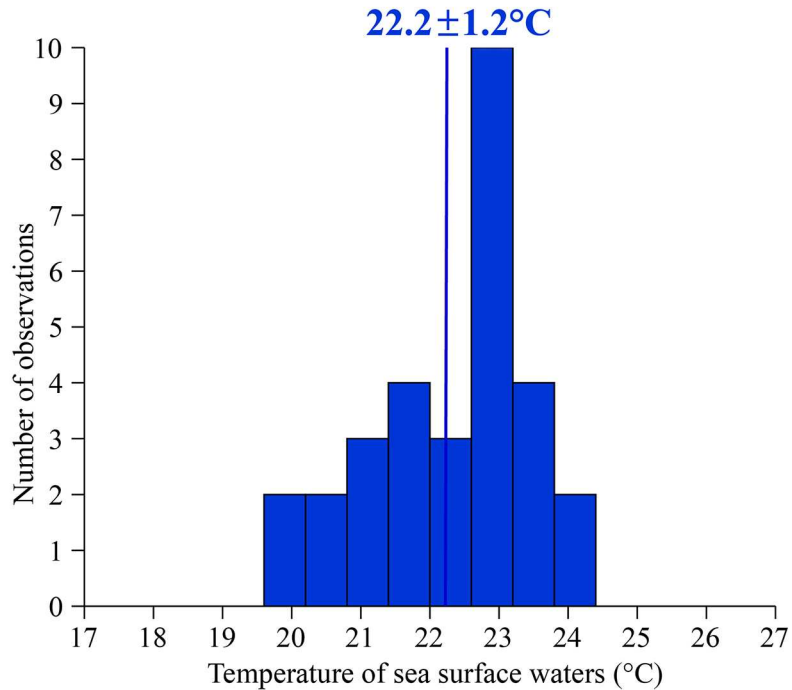


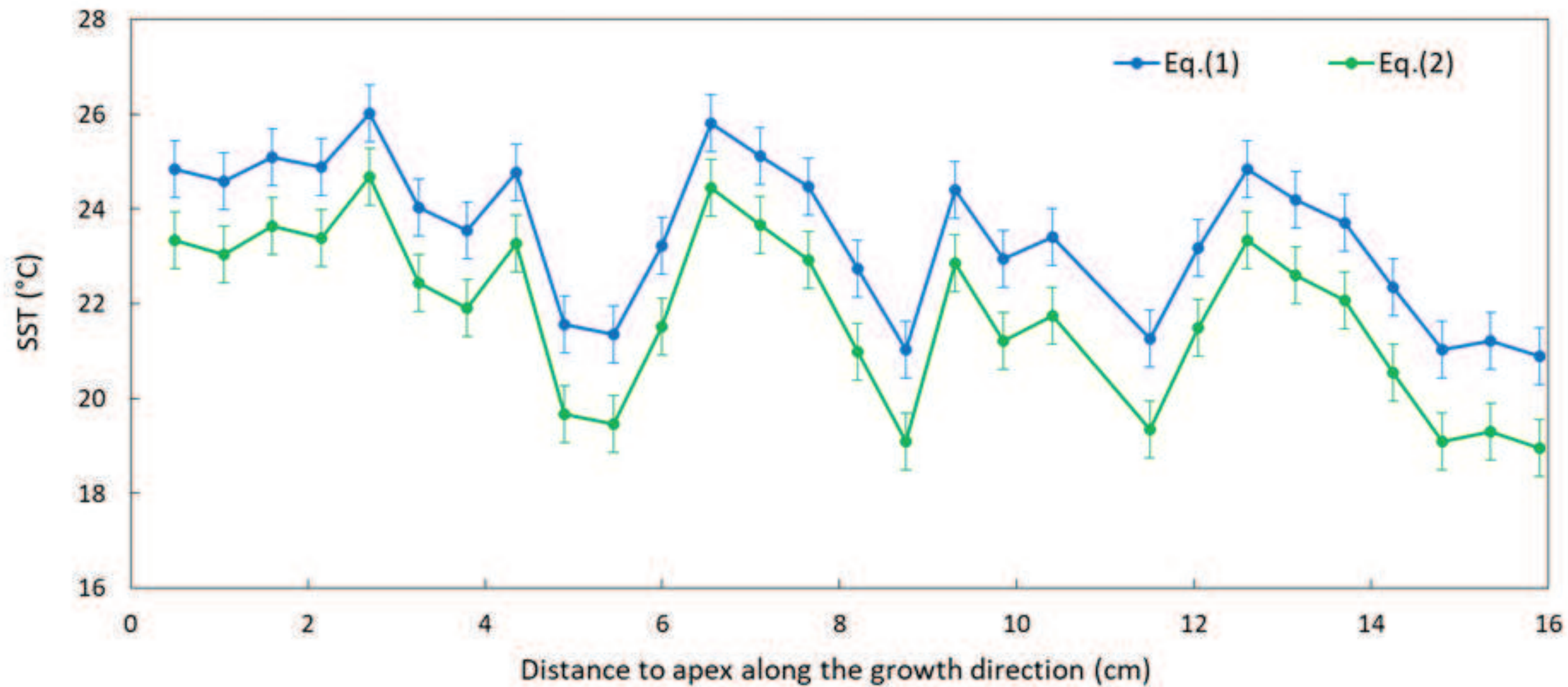
(a) Incremental series of *Cardium edule*

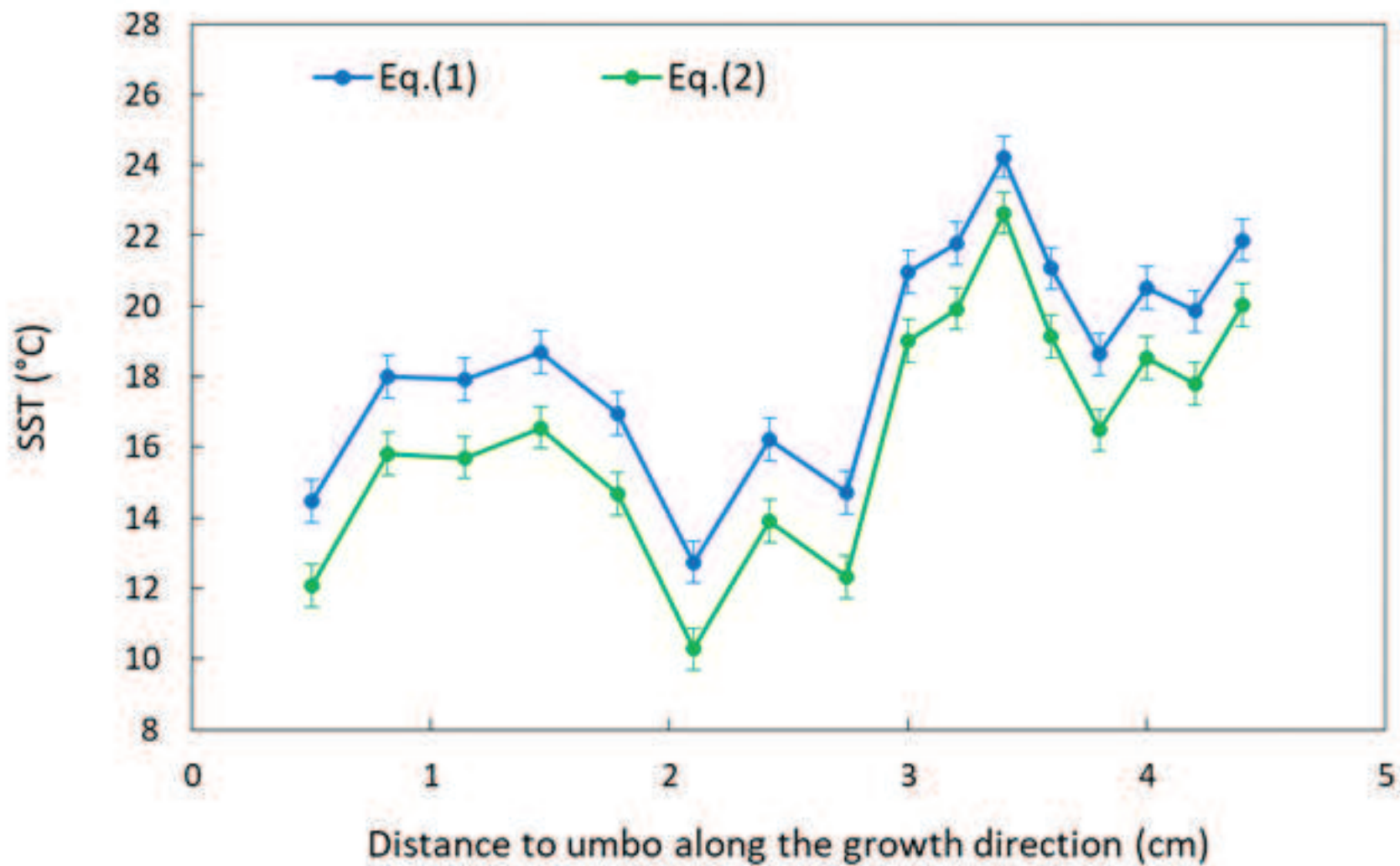


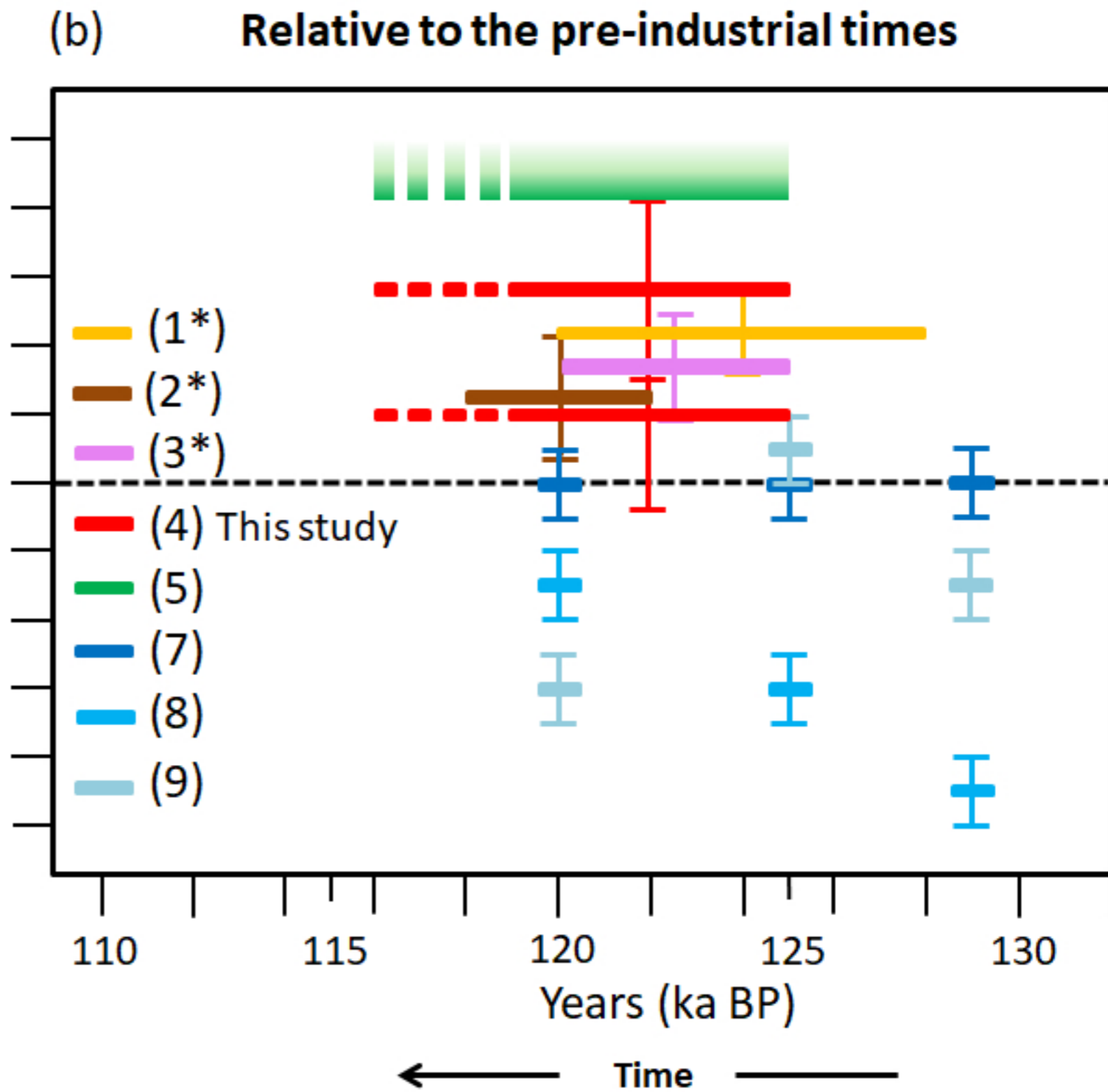
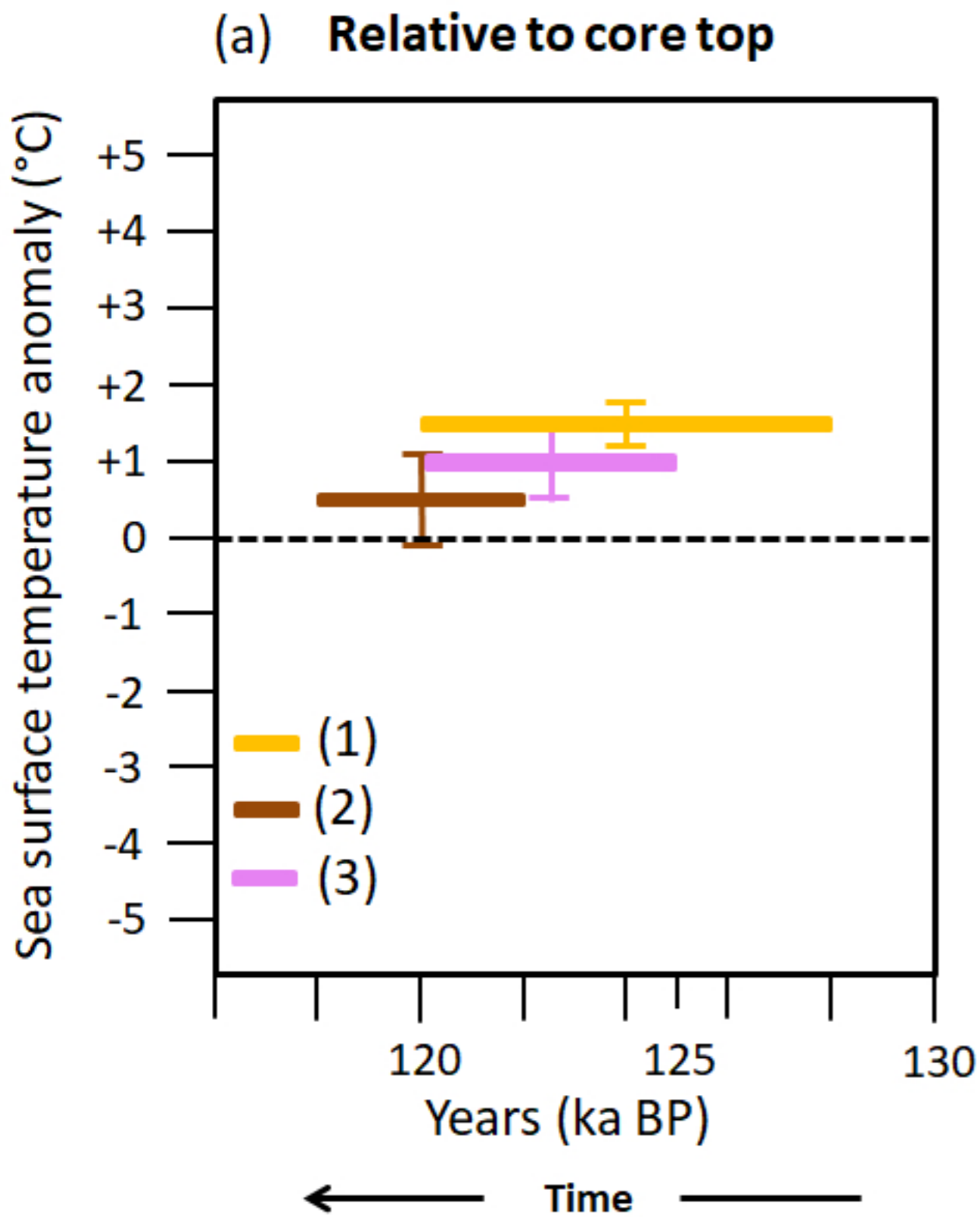
(b) Incremental series of *Stramonita haemastoma*











Period	Mean SST (°C)	Method	Authors
2007-2017	20.44°C	Satellite data and <i>in situ</i> SST data	Meco et al. (2018)
1993-2014	20.5±0.2°C (*)	Satellite data and <i>in situ</i> SST data	Gómez-Letona et al. (2017)
1976-2005	18.5±0.2°C (*)	GCMs simulations	Sousa et al. (2017)
1982-2012	20.3±0.2°C (*)	Satellite data	deCastro et al. (2014)
≤ 2001	≤ 20°C	Atlas 2001	Muhs et al. (2014) from Conkright et al. (2002)
1982-2012	20.3±0.2°C (*)	Satellite data	Santos et al. (2012)

Authors	Area	Proxy or Studies	SL anomaly	SL highstand(s)	Time-period
Kopp et al. (2013)	47 sites worldwide	Probabilistic analyses on $\delta^{18}\text{O}$ of foraminifera	> + 6.6 m	2 highstands separated of +4m	125-116 ky
Spratt and Lisiecki (2016)	sites worldwide	Stack estimates from $\delta^{18}\text{O}$ of foraminifera	+6 to + 9m ± 12 m		126-119 ky BP
Blanchon et al. (2009)	Yucatan, Mexico	Corals	+ 3 m : + 6m :		126-121 ky 120-117 ky
Thompson et al. (2011)	Bahamas	Corals	+4 m : + 6 m :		≈ 123 ky ≈ 119 ky
O'Leary et al. (2013)	Western Australia	Corals	+ 3-4 m : + 9.5 m :		127-119 ky < 118 ky
Dutton et al. (2015b)	Seychelles	Corals	rise : +7.6 m :		129-125 ky < 125 ky
Masson-Delmotte et al. (2013)	–	SL synthesis work	+ 6 m	2 highstands separated of +4m	129-116 ky
Dutton et al. (2015a)	–	SL synthesis work	+ 6 to + 9 m :	Peak :	129-116 ky < 125 ky (122-119 ky)
Capron et al. (2017a,b)	–	LIG studies	+ 6 to + 9 m :	2-phases, late peak :	122-119 ky

Reference	Material	SST Proxy
Sicre et al. (2000)	Sediment core	U_{37}^K
Matsuzaki et al. (2011)	Sediment core	Planktonic foraminiferal assemblages
Castañeda et al. (2009)	Sediment core	U_{37}^K
This study	Marine deposits	$\delta^{18}O$ from mollusk shells
Montesinos et al. (2014)	Marine deposits	Extralimital species
Muhs et al. (2014)	Marine deposits	Extralimital species
Hoffman et al. (2017)	Sediment cores	Planktonic foraminiferal assemblages