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1 **Holocene climate variability of the western Mediterranean: surface water dynamics inferred**
2 **from calcareous plankton assemblages**

3

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18

19 **Keywords**

20 Coccolithophores, Foraminifera, Alboran Sea, Holocene, Paleoclimate, Paleoproductivity,
21 Millennial-centennial scale climate variability

22

23 **Abstract**

24 A high resolution study (centennial-scale) has been performed on the calcareous plankton assemblage
25 of the Holocene portion of the Ocean Drilling Program Site 976 (Alboran Sea) with the aim to identify
26 main changes in surface water dynamic. The dataset also provided a Seasonal foraminiferal Sea
27 Surface Water Temperatures (SSTs), estimated using the modern analog technique SIMMAX 28, and
28 it was compared with available geochemical and pollen data at the site.

29 Three main climate shifts were identified: I) The increase in abundance of *Syracosphaera* spp. and
30 *Turborotalita quinqueloba* marks the early Holocene humid phase, during maximum summer
31 insolation and enhanced river runoff. It is concomitant with the expansion of *Quercus*, supporting
32 high humidity on land. It ends at 8.2 ka, registering a sudden temperature and humidity reduction; II)
33 The rise in the abundances of *Florisphaera profunda* and *Globorotalia inflata*, at ca. 8 ka, indicates
34 the development of the modern geostrophic front, gyre circulation and of a deep nutricline following
35 the sea-level rise; III) The increase of small *Gephyrocapsa* and *Globigerina bulloides* at 5.3 ka,

36 suggests enhanced nutrient availability in surface waters, related to more persistent wind-induced
37 upwelling conditions. Relatively higher winter SST in the last 3.5 kyr favored the increase of
38 *Trilobatus sacculifer*, likely connected to more stable surface water conditions. Over the main trends,
39 a short term cyclicality is registered in coccolithophore productivity during the last 8 kyr. Short periods
40 of increased productivity are in phase with Atlantic waters inflow, and more arid intervals on land.
41 This cyclicality has been related with periods of positive North Atlantic Oscillation (NAO) circulations.
42 Spectral analysis on coccolithophore productivity confirms the occurrence of millennial-scale
43 cyclicality suggesting an external (i.e. solar) and an internal (i.e. atmospheric/oceanic) forcing.

44

45 **Introduction**

46 An increasing number of climate records reveals that the Holocene has experienced a pervasive
47 millennial- and centennial-scale climate variability (e.g. Jalut et al., 2009; Magny et al., 2013;
48 Mayewski et al., 2004; Walker et al., 2012; Wanner et al., 2015), well-documented in both the North
49 Atlantic (e.g. Bond et al., 2001; Repschläger et al., 2017; Thornalley et al., 2009) and western
50 Mediterranean (e.g. Ausín et al., 2015a; Cacho et al., 2001; Català et al., 2018; Frigola et al., 2007;
51 Jalali et al., 2017, 2016; Nieto-Moreno et al., 2015; Rodrigo-Gámiz et al., 2011). The western
52 Mediterranean Sea is in fact extremely sensitive to the changes experienced in the North Atlantic and
53 is an ideal location for high-frequency climatic investigations, because water mass properties changes
54 and oceanographic and atmospheric circulation oscillations are usually amplified (Cacho et al., 1999,
55 2001; Català et al., 2018; Frigola et al., 2007, 2008; Jalali et al., 2016; Moreno et al., 2002; Nieto-
56 Moreno et al., 2015; Sierro et al., 2005; Toucanne et al., 2012). The millennial-scale climate
57 variability in the western Mediterranean is reflected in different water column configurations and
58 oceanographic features (e.g. fronts and eddies) that left a clear signature in the calcareous plankton
59 assemblages (Ausín et al., 2015a; Pérez-Folgado et al., 2003, 2004; Sbaffi et al., 2001). During the
60 Holocene, the Alboran Sea experienced relevant oceanographic perturbations, the most important of
61 which was the instauration of the modern geostrophic front and establishment of gyre anticyclonic
62 circulation dynamics, following sea level rising after the last deglaciation (Ausín et al., 2015b; Català
63 et al., 2018; Colmenero-Hidalgo et al., 2004; Heburn and La Violette, 1990; Rohling et al., 1995;
64 Weaver and Pujol, 1988). This important change also marked the end of the Organic Rich Layer 1
65 (ORL1) deposition in the western Mediterranean (Bárcena et al., 2001; Cacho et al., 2002; Jimenez-
66 Espejo et al., 2007, 2008; Rogerson et al., 2008). Targeting the high frequency oscillations
67 experienced during the Holocene, a growing attention has been focused on the impact of the North
68 Atlantic Oscillation (NAO) atmospheric pattern in the western Mediterranean Sea, in terms of
69 westerlies strength and deep water production, precipitation and river runoff, and coccolithophore

70 productivity (Ausín et al., 2015a; Fletcher et al., 2012; Frigola et al., 2007; Jalali et al., 2016; Moreno
71 et al., 2005; Smith et al., 2016; Trigo et al., 2004; Zielhofer et al., 2017). However, only a few studies
72 so far have documented the relationship between coccolithophore productivity and atmospheric
73 variability, like the present NAO index (e.g. Ausín et al., 2015a).

74 In this framework we carried out, over the last 12 ka, at the Ocean Drilling Program (ODP) Site 976,
75 an integrated study between coccolithophores and planktonic foraminifera by a centennial-scale
76 resolution, not available so far in the Alboran Sea. The aim was to reconstruct paleoenvironmental
77 fluctuations in the Alboran Sea and to discuss the mechanisms controlling fossil assemblage and
78 productivity variations at different time scales. A planktonic foraminifera-based Sea Surface
79 Temperature (SST) reconstruction is also provided, to have further insights on seasonal and annual
80 temperature variations. In addition, spectral and wavelet analyses of the coccolithophore
81 accumulation rates are performed to identify the different periodicities of coccolithophore
82 productivity fluctuations. The study also benefits from the comparison with additional inorganic and
83 organic geochemical proxies (Jiménez-Amat and Zahn, 2015; Martrat et al., 2014) and pollen data
84 (Combourieu-Nebout et al., 2009) available at the same site, improving the paleoclimate
85 reconstruction through a direct multi-proxy approach.

86

87 **Area of Study**

88 *Present hydrographical conditions*

89 The ODP Site 976 was recovered off the Spanish coast in the Alboran Sea, the westernmost basin of
90 the Mediterranean Sea, bordering the Atlantic Ocean (Fig. 1). Surrounding lands include the high
91 physiography of the Betic cordillera and Moroccan Rif mountains, that might provide a certain
92 riverine input, although subjected to high seasonality and extreme climatic events (Jimenez-Espejo
93 et al., 2008; Liqueste et al., 2005; Lobo et al., 2006). Surface Atlantic Water (AW) pours inside the
94 Alboran basin through the Strait of Gibraltar, as a constant stream of surface low-salinity waters
95 called the Atlantic Jet (AJ). The latter contributes to the creation of two quasi-permanent meso-scale
96 anticyclonic gyres: the Western Anticyclonic Gyre (WAG) and the Eastern Anticyclonic Gyre (EAG)
97 (Fig. 1) (Heburn and La Violette, 1990; Sarhan et al., 2000). In the area, two mechanisms are known
98 to be relevant for the upwelling dynamic: the southward drifting of the AJ, that would allow the water
99 from below to rise, and the wind stress (Sarhan et al., 2000). The influence of the vertical mixing of
100 AJ and deeper Mediterranean waters, concurrently with the complex bottom topography, forms areas
101 of geostrophic front and quasi-permanent upwelling: the Alboran front and the Almeria-Oran front
102 (Fig.1) (Perkins et al., 1990; Viúdez et al., 1996).

103 The combination of gyres and upwelling fronts results in increased nutrient availability and high
104 productivity waters, among the richest within the rest of the Mediterranean (D’Ortenzio and
105 D’Alcalà, 2009; Garcia-Gorriz and Carr, 1999). Counteracting the AW inflow, the denser and more
106 saline Mediterranean waters exit the basin through the Mediterranean Outflow Water (MOW), that
107 includes the western Mediterranean Deep Water (WMDW) from the Gulf of Lion and the Levantine
108 Intermediate Water (LIW) from the far East of the Mediterranean basin (Fig. 1) (Millot, 2008; Perkins
109 et al., 1990). Deep water formation is controlled by surface heat loss due to winds blowing from the
110 north and north-west (Font et al., 2007; Mertens and Schott, 1998; Rixen et al., 2005; Smith et al.,
111 2008).

112

113 *Present Climate Conditions*

114 The Alboran Sea climate conditions are under the influence of the Azores high pressure cell and its
115 seasonal latitudinal shift, resulting in mild wet winters and dry hot summers (Lionello, 2012; Moreno
116 et al., 2012; Rohling et al., 2015; Sumner et al., 2001). At decadal and inter-annual time scales,
117 atmospheric variability is regulated by the North Atlantic Oscillation (NAO) index, which is
118 characterized by positive (NAO+) and negative (NAO-) regimes (Hurrell, 1995; Olsen et al., 2012;
119 Smith et al., 2016; Trigo et al., 2004). During a NAO+ regime, stronger pressure difference between
120 the Azores High and Icelandic Low atmospheric cells brings storm trajectories to the north,
121 determining stormier and wetter weather in northwest Europe and dryer winters in southern Europe
122 and North Africa (Olsen et al., 2012; Smith et al., 2016; Zielhofer et al., 2017). At NAO- regime,
123 weaker difference between the two pressure cells leads storm tracks to the south, enhancing
124 precipitations over southern Europe and North Africa (Smith et al., 2016; Wanner et al., 2015). In the
125 Iberian Peninsula, winter precipitation mode has been related to air masses raised by atmospheric
126 instabilities and moisture supply from the tropical-subtropical North Atlantic corridor (Gimeno et al.,
127 2010).

128

129 **Materials and Methods**

130 *Core material and available data sets*

131 The ODP Site 976 (Comas et al., 1996) is located about 60 km south of the Iberian Peninsula and
132 about 110 km East of the Strait of Gibraltar (36°12.3' N, 4°18.7' W) (Fig. 1). The cores were recovered
133 on the lower part of a very gentle slope, dipping southward of the Spanish margin in the Alboran
134 Basin, at a depth of 1108 m. The investigated sediments are from Hole C – Core 1H – sections 1-3
135 (Comas et al., 1996), between 0.07 to 4.03 m below the sea floor. Main lithology is composed of
136 nannofossil rich clay, with slight to moderate bioturbation and common shell fragments (Comas et

137 al., 1996). One Organic Rich Layer (ORL) occurs in the studied interval and is identified on the basis
138 of the maximum accumulation of di- and tri- alkenones of 37 carbons (Martrat et al., 2004, 2014).
139 The age model used in the present study is from Martrat et al. (2014), based on available ¹⁴C calibrated
140 AMS radiocarbon dates (Combourieu-Nebout et al., 2002, 2009). Sampling resolution of calcareous
141 plankton dataset varies between 1 sample every 2 to 6 cm, according to the variable sedimentation
142 rate (20 to 60 cm), thus providing a temporal resolution of one sample every ca. 100 years.

143

144 *Calcareous plankton*

145 The coccolith assemblages were analyzed in 129 samples. Sample preparation for the coccolith
146 analysis follows the random settling technique by Flores and Sierro (1997). The quantitative analyses
147 were performed using a polarized light microscopy at 1000X magnification. The relative abundance
148 of taxa was estimated counting at least 500 specimens per sample, in variable fields of view.
149 Reworked calcareous nannofossils were estimated separately during this counting. The absolute
150 abundance of taxa is expressed as Nannofossil Accumulation Rate (NAR). The total NAR, used to
151 determine coccolithophore paleoproductivity (Baumann et al., 2004; Steinmetz, 1994), was estimated
152 following Flores and Sierro (1997):

153

$$\text{NAR} = \text{N} * \text{w} * \text{S}$$

154 where N is the number of coccoliths per gram of sediment (Ng^{-1}), w is the wet bulk density ($\text{g}\times\text{cm}^{-3}$)
155 (shipboard bulk density data, Comas et al., 1996), and S is the sedimentation rate ($\text{cm}\times\text{ky}^{-1}$). Wet
156 bulk density is frequently used as a substitute to dry bulk density, in the absence of the latter, to
157 estimate coccolithophore production (Grelaud et al., 2009; Marino et al., 2014; Stolz and Baumann,
158 2010). For taxonomic identification we referred to Young et al. (2003) and Jordan et al. (2004).
159 According to Flores et al. (2000): gephyrocapsids with high angle bridge ($>50^\circ$) and $>3\ \mu\text{m}$ in size
160 are indicated as *Gephyrocapsa oceanica*; gephyrocapsids $>3\ \mu\text{m}$ in size with a low angle bridge ($<$
161 25°) are indicated as *Gephyrocapsa muelleriae*; small *Gephyrocapsa* includes gephyrocapsids $< 3\ \mu\text{m}$
162 in size. Specimens of *Emiliania huxleyi* were differentiated into two main groups following size
163 criteria (Colmenero-Hidalgo et al., 2002): large *E. huxleyi* $> 4\ \mu\text{m}$ and small *E. huxleyi* $< 4\ \mu\text{m}$. Warm
164 water taxa are grouped according to their ecological preference for tropical-subtropical waters
165 (Baumann et al., 2004; Boeckel and Baumann, 2004; Winter and Siesser, 1994). The group includes:
166 *Calciosolenia* spp., *Discosphaera tubifera*, *Rhabdosphaera stylifera*, *Rhabdosphaera clavigera*,
167 *Umbilicosphaera foliosa*, *Umbilicosphaera sibogae*, *Umbellosphaera* spp., *Oolithotus* spp.
168 Planktonic foraminifera assemblages were analyzed in 122 samples washed through 63 and 150 μm
169 sieves. The residues ($>150\ \mu\text{m}$) were split, until a representative aliquot containing about 300
170 specimens has been obtained. All specimens were counted in the aliquots and species abundances

171 were quantified as percentages on the total number of planktonic foraminifers (relative abundance)
172 and as absolute abundances expressed as planktonic foraminifera Accumulation Rates (pfAR). The
173 pfAR has been calculated following Giraudeau et al. (2001):

$$174 \quad \text{pfAR} = \text{AA} * \mathbf{w} * \text{S}$$

175 where **AA** is the number of specimens per gram of dry sediment (nr/g), **w** is the wet bulk density
176 ($\text{g} \times \text{cm}^{-3}$) (shipboard bulk density data, Comas et al., 1996), and **S** is the sedimentation rate ($\text{cm} \times \text{ky}^{-1}$).
177 Sixteen species or species groups were distinguished. *Globigerinoides ruber* includes morphotypes
178 of *G. ruber* white, and *Globigerinoides elongatus* (sensu Aurahs et al., 2011); *Trilobatus sacculifer*
179 includes *Trilobatus trilobus*, *Trilobatus sacculifer* and *Trilobatus quadrilobatus* (sensu André et al.,
180 2012; Hemleben et al., 1989; Spezzaferri et al., 2015). *Globoturborotalita rubescens* includes
181 *Globoturborotalita tenella* because of their similar ecological preference (Capotondi et al., 1999).
182 The taxonomy of *Neogloboquadrina* spp. follows criteria by Darling et al., (2006):
183 *Neogloboquadrina incompta* includes specimens previously referred to *N. pachyderma* (dextral) and
184 intergrades between *N. pachyderma* (dextral) and *N. dutertrei*. *Neogloboquadrina pachyderma* only
185 includes the left coiling specimens.

186 According to their ecological preference (Hemleben et al., 1985; Kucera et al., 2005; Pujol and
187 Vergnaud-Grazzini, 1995) and to previous Mediterranean Sea paleoclimatic reconstructions (De Rijk
188 et al., 1999; Rohling et al., 1997), *G. ruber*, *T. sacculifer*, *Hastigerina pelagica*, *G. rubescens*,
189 *Orbulina universa*, *Beella digitata* and *Globigerinella siphonifera* have been grouped as warm water
190 taxa.

191 192 *Sea Surface Temperature estimation*

193 Planktic foraminifera assemblages were used to reconstruct annual, summer (July to September) and
194 winter (January to March) SST with the modern analog technique non distance-weighted SIMMAX
195 28 and 10 analogs (Pflaumann et al., 1996). Considering that the study site is influenced by Atlantic
196 and Mediterranean ocean circulation, following Schirrmacher et al. (2019), we use the combined
197 North Atlantic core-top database (Kucera et al., 2005; Salgueiro et al., 2010, 2014) and the
198 Mediterranean database (Hayes et al., 2005), and the root mean square error of both annual and
199 seasonal SST reconstructions is about 1.3°C (Schirrmacher et al., 2019).

200 201 *Power spectral and wavelet analysis*

202 Spectral and wavelet analyses were performed on the total NAR, displaying relevant high frequency
203 oscillations throughout the record. The analysis of the non-stationary (frequency changes along time)
204 and non-linear signals, was performed by applying the Empirical Mode Decomposition algorithm

205 (EMD) of Huang et al. (1998) in order to decompose multi-component signals into a series of
206 amplitude and frequency modulation (AM-FM) waves, each with slowly varying amplitude and
207 phase. A major advantage of EMD is that the basis functions are derived from the signal itself, hence
208 the analysis is adaptive, in contrast to the traditional methods where the basis functions are fixed as
209 sine and cosine for Fourier transform like methods and the mother wavelet functions for wavelet
210 analysis.

211 The signal and the Intrinsic Mode Functions (IMF) components are analysed without interpolation,
212 keeping the original evenly sampling intervals, with:

- 213 1. “REDFIT”, that computes the spectrum of a possibly unevenly sampled time-series, by using
214 the Lomb-Scargle Fourier transform. The spectrum is bias-corrected using spectra computed
215 from simulated AR(1) series and the theoretical AR(1) spectrum (Lomb, 1976; Scargle, 1982;
216 Schulz and Mudelsee, 2002).
- 217 2. Foster's (1996) weighted wavelet Z-transform (WWZ). To analyze non-stationary and
218 irregularly sampled signals, we need an extension of the classic wavelet formalism. Foster
219 (1996), who defines the WWZ, developed such extension as a suitable weighted projection
220 method re-orthogonalizing the three basic functions (real and imaginary part of the Morlet
221 wavelet and a constant) by rotating the matrix of their scalar products. Furthermore, he
222 introduces statistical F-tests to distinguish between periodic components and a noisy
223 background signal.

224

225 **Results**

226 *Calcareous nannofossils*

227 Calcareous nannofossils are generally abundant and well preserved and dissolution phenomena seem
228 not to be significant. Abundances of the most relevant taxa are presented in Fig. 2 and no major
229 discrepancies are observed between relative and absolute trends. The total NAR ranges between 0.5
230 $\times 10^{11}$ coccoliths $\times \text{cm}^{-2} \times \text{kyr}^{-1}$ and 2×10^{11} coccoliths $\times \text{cm}^{-2} \times \text{kyr}^{-1}$, with an average of 0.8×10^{11}
231 coccoliths $\times \text{cm}^{-2} \times \text{kyr}^{-1}$ (Fig. 2). A marked abundance peak occurs at about 8.2 ka, and an oscillating
232 pattern is recorded in the last 8 kyr (Fig. 2). Considering the relative abundances of the taxa (%), *E.*
233 *huxleyi* $< 4 \mu\text{m}$ represents the main taxon, having percentages between 40 and 60% and the highest
234 values between 10 and 8 ka (Fig. 2). Among geophyrocapsids, *G. muelleriae* results to be the most
235 abundant in the lower part of the record, with values reaching 30% of the assemblage, followed by a
236 descending trend (Fig. 2). Small *Gephyrocapsa* show an increase in abundance from 8%, between 10
237 and 7 ka, to 15% from 5.3 ka upward (Fig. 2). *Gephyrocapsa oceanica*, mainly represented by
238 morphotypes larger than $5 \mu\text{m}$, shows abundance fluctuations between 3 and 10% throughout the

239 Holocene (Fig. 2). *Florisphaera profunda*, although not a dominant component of the assemblages
240 (with very few isolated peaks greater than 15%), shows a distinct pattern, with very low percentage
241 values up to ~8 ka, that clearly increase upwards in fluctuating abundances (Fig. 2). Among less
242 abundant taxa, *Syracosphaera* spp. (*S. histrica* and *S. pulchra*), having a mean value of 1%, show a
243 distinct increase between 11 and 8 ka, reaching values of about 5% (Fig. 2). The warm water coccolith
244 taxa have very low abundances throughout the succession, with a mean value of 2%; the group shows
245 a gradual increase between 12 and 8 ka, and fluctuating pattern afterwards (Fig. 2). *Helicosphaera*
246 *carteri* and *E. huxleyi* > 4 µm show a similar pattern with abundance values ranging between 10 and
247 15% in the lowest part of the succession (during the Younger Dryas), followed by a clear decreasing
248 trend, with values around 1% (Fig. 2). Other taxa, not showing particular trends or significant
249 fluctuations, are represented by *Coccolithus pelagicus* ssp. *pelagicus*, *Gephyrocapsa caribbeanica*
250 and *Coronosphaera* spp., with percentages not higher than 5%. Subordinate taxa do not exceed the
251 3% of the assemblage and include *Coccolithus pelagicus* ssp. *braarudii*, *Coccolithus pelagicus* ssp.
252 *azorinus*, *Braarudosphaera bigelowii*, *Calcidiscus leptoporus* ssp. small (3–5 µm), *C. leptoporus* ssp.
253 *leptoporus* (5–8 µm), *C. leptoporus* ssp. *quadriperforatus* (8–10 µm), *Ceratolithus* spp.,
254 *Helicosphaera pavementum*, *Helicosphaera hyalina*, *Pontosphaera* spp., *Gladiolithus flabellatus*,
255 *Scyphosphaera* spp. and *Umbilicosphaera hulburtiana*. Reworked taxa occur in the samples with
256 variable abundances, never exceeding about 4% (Fig. 2).

257

258 *Planktonic foraminifera assemblages*

259 Planktonic foraminifera are well preserved and diversified. Relative and absolute abundances of the
260 most abundant/significant planktonic foraminifera taxa/ groups show comparable trends throughout
261 the entire succession. *Neogloboquadrina incompta* and *Turborotalita quinqueloba* are abundant in
262 the lower part of the record (between 12.5 ka and about 8 ka) and undergone a strong decreasing
263 upward (Fig. 3). Although with lower relative and absolute abundances, *G. ruber* and *G. bulloides*
264 are also abundant in this interval (Fig. 3). At about 8 ka, a prominent replacement of *G. inflata* at the
265 expense of *N. incompta* and *T. quinqueloba* occurs. Starting from 8 ka upwards, *G. inflata* together
266 with *G. bulloides* and *G. ruber*, became the most abundant taxon in the record (Fig. 3). *Globigerinita*
267 *glutinata*, with relative abundances not higher than 10%, doesn't show any relevant fluctuation in the
268 distribution pattern (Fig. 3). *Trilobatus sacculifer* became more abundant from about 8 ka upward,
269 showing a more prominent increase, as relative and absolute abundances, during the last 3.5 kyr (Fig.
270 3). A similar distribution pattern is also shown by *Truncorotalita truncatulinoidea* (Fig. 3). Other taxa
271 showing a very scattered distribution in the studied interval, with relative abundances < 3 % and any

272 significant fluctuations, are not shown in Fig. 3. They are represented by *Globorotalia scitula*,
273 *Globigerina falconensis*, *G. rubescens*, *N. pachyderma*, *G. siphonifera* and *O. universa*.

274

275 *Sea Surface Temperature*

276 Annual, summer and winter SST patterns show sharp fluctuations between 10.2 ka and 8.7 ka
277 characterized by strong drops of winter and summer temperature values greater than 10°C (Fig. 3).
278 During this interval an important increase of *T. quinqueloba* is observed, together with high
279 occurrence of *N. incompta* and *N. dutertrei* and increasing trend of warm water foraminifera taxa
280 (Fig. 5). In this interval, the similarity index slightly decreases (Fig. 3), indicating that this species
281 combination is not usual in the modern oceanographic condition for the North Atlantic and the
282 Mediterranean region. In this interval the average annual SST is about 13.9°C, while average winter
283 and summer temperatures are 11.7°C and 16.7°C respectively (Fig. 3). For the last 8 kyr, the average
284 annual SST is about 18.5°C, while winter SST in the Alboran Sea varies around ca. 15 °C, in
285 agreement with modern conditions (15.4 °C; Locarnini et al., 2013) (Fig. 3). The average summer
286 SST is 22.6°C, exceeding modern ones (21.4°C; Locarnini et al., 2013) (Fig. 3). Low temperatures
287 values are recorded between about 8.6 and 7.7 ka both in summer (ca. 20°C) and in winter (ca.13°C)
288 (Fig. 3). The highest temperatures are recorded between 7.7 ka and 5.8 ka with temperatures up to 23
289 °C during summer and up to 16 °C during winter (Fig. 3). During the last 5 kyr, summer SST weakly
290 decreases, with slightly oscillating values between 22°C and 23°C (Fig. 3). In the same interval,
291 winter SSTs are almost stable with average values of about 15°C (Fig. 3), although during the last 3
292 kyr, the winter temperatures are characterized by a slight increase.

293 These results suggest that, with the exception of the interval between 10.2 and 8.7 ka, our SST record
294 shows values comparable with those derived from the alkenone-SST at the same site (Martrat et al.,
295 2014), with the foram-based SST from other nearby records (Pérez-Folgado et al., 2003; Schirrmacher
296 et al., 2019), and with the present-day SST in the region (Locarnini et al., 2013). On the basis of these
297 considerations, only the last 8.6 kyr record has been considered for the climate interpretation.

298

299 *Power spectral and wavelet analysis*

300 The power spectrum of total NAR shows prominent peaks (over the 95% Confidence Level – C.L.)
301 of periodicity ranging between 1100 yr and 1700 yr (IMF3) (Fig. 4a, c). The wavelet analysis reveals
302 that periodicities are not evenly distributed through time and specifically the 1102 yr cycle occurs
303 from 12 ka to about 4 ka (Fig. 4c), while the 1693 yr periodicity emerges since about 5 ka upwards
304 (Fig. 4c). Significant peaks (over the 95% of C.L.) are observed at the periods of ~ 4300 yr and ~

305 8000 yr all along the record (IMF 4, 5) (Fig. 4 d,e). Scattered distribution of cycles between ~ 400
306 and ~ 700 yr are also observed (IMF 2) (Fig. 4b).

307

308 **Discussion**

309 *Main hydrographic and climate variations*

310 Surface water modifications occurring in the last 11 ka can be described by three main long-term
311 (between 3-5000 c.a. years-long) steps: Phase I, II and III (Figs. 5-6).

312

313 *Phase I – the early Holocene humid period*

314 This phase straddles the early Holocene, between 11.5 ka and 8 ka and is subsequent to the Younger
315 Dryas Stadial. The climate evolution of the latter stadial has been discussed in detail in Bazzicalupo
316 et al. (2018) based on the same proxies and therefore not discussed in the present study. Phase I is
317 marked by a gradual surface water temperature increase, well described by progressively growing
318 abundances of both coccolithophore and foraminifera warm-water taxa, associated with increasing
319 summer insolation (Fig. 5). During this phase, the distinct increase in both *Syracosphaera* spp. and
320 *T. quinqueloba* (Fig. 5) provides evidences of enhanced riverine/detrital input in surface waters.
321 *Syracosphaera* spp. has been, in fact, related to enhanced supply of fresher and turbid upper layer
322 (Ausín, et al., 2015b; Bazzicalupo et al., 2018; Colmenero-Hidalgo et al., 2004; Weaver and Pujol,
323 1988), while the cold taxon *T. quinqueloba* flourishes in high fertile and low density surface waters
324 (Aksu et al., 2002; Hemleben et al., 1985; Pujol and Vergnaud-Grazzini, 1995; Triantaphyllou et al.,
325 2010). Enhanced abundances of this taxon have been also related to areas influenced by continental
326 runoff (Bartels-Jónsdóttir et al., 2015; Girone et al., 2013; Jonkers et al., 2010; Margaritelli et al.,
327 2016; Rohling et al., 1997; Vallefucio et al., 2012) and, in the Eastern Mediterranean, the increase in
328 abundance of *T. quinqueloba*, during the deposition of sapropel layer S1, has been linked to a high
329 tolerance for low salinity and highly stratified water conditions coupled with the presence of high
330 nutrients and terrestrial organic material (Capotondi et al., 2004; Kontakiotis, 2016; Principato et al.,
331 2006; Rohling et al., 1997; Zachariasse et al., 1997). The high abundance of small *Gephyrocapsa*
332 during phase I (Fig. 6), also sustains nutrient availability in surface water (Gartner et al., 1987;
333 Hernández-Almeida et al., 2011; Okada and Wells, 1997; Takahashi and Okada, 2000). A
334 concomitant expansion of *Quercus* during phase I (Fig. 5) highlights enhanced humidity on land
335 (Combourieu-Nebout et al., 2009) likely in relation with extreme seasonality during precession
336 minima/insolation maxima (Fig. 5) and increased autumn/winter westerlies-carried rains over the
337 western Mediterranean, which supports enhanced supply of fresher water into the basin. This scenario
338 seems to reflect a regional climate condition since it is consistent with the establishment of the Early

339 Holocene humid phase, occurring between 11.5 and 7 ka (Jalut et al., 2009; Magny et al., 2002, 2013;
340 Peyron et al., 2017; Zanchetta et al., 2007; Zielhofer et al., 2017) and with additional evidences of
341 autumn/winter precipitation increase over the northern Mediterranean borderlands during northern
342 Hemisphere insolation maxima (Kutzbach et al., 2013; Meijer and Tuenter, 2007; Toucanne et al.,
343 2015; Tzedakis, 2007). Phase I is within the interval of sapropel layer S1 deposition in the Eastern
344 Mediterranean (10.8-6.1 ka, De Lange et al., 2008), developed during maximum summer insolation,
345 that contributed, through the enhanced monsoon precipitation, to increased runoff in the Eastern
346 Mediterranean (Howell and Thunell, 1992; Rohling et al., 2002, 2004, 2015; Rossignol-Strick, 1985;
347 Rossignol-Strick et al., 1982). Our data evidence that freshwater runoff during sapropel events was
348 not restricted to the Eastern Mediterranean but was rather widespread over the entire Mediterranean
349 Sea due to increased rainfall (Bard et al., 2002; Kallel et al., 2000; Kallel and Labeyrie, 1997;
350 Toucanne et al., 2015; Zanchetta et al., 2007), thus strengthening the connection between North
351 African summer monsoon and the increased western Mediterranean autumn/winter precipitation
352 during sapropel deposition (Toucanne et al., 2015). On the other hand, phase I straddles the final
353 phase of the deposition of ORL 1 as indicated by the decreasing, albeit still high, values of C₃₇ (Fig.
354 5), and reduced deep water ventilation in the western Mediterranean (Frigola et al., 2007). Given the
355 time offset between the beginning of the ORL1 formation (14.5 ka, Martrat et al., 2014), and the
356 recorded enhanced riverine input and humidity on land at c.a. 11.5 ka, it appears unlikely that excess
357 precipitation was the driving force of the ORL1 formation in the western Mediterranean (Rogerson
358 et al., 2008). As stated in Bazzicalupo et al. (2018), shoaling of the nutricline and increased export
359 production at the sea floor are relevant mechanism in the ORL1 deposition at the study core.

360

361 *The 8.2 ka event*

362 The transition between phase I and the following phase II is characterized by higher absolute
363 abundances of *N. incompta* between ~8.6 ka and ~8.1 ka (Fig. 6) and by a sharp warm water taxa
364 decrease (Fig. 5), suggesting water cooling. In more detail, a sharp and brief cooling event of about
365 3°C is recorded in annual, winter and summer SST (Fig. 6). An interruption of the surface water
366 warming trend is also indicated by a decrease of warm water coccolith taxa in the early stage (Fig.
367 5). A concomitant temperate forest regression (Fig. 6) marks a short-term precipitation decrease
368 episode.

369 This cooling episode is here related to the well-known cold and dry 8.2 ka event that punctuates the
370 early Holocene evolution and it is broadly recognized in Greenland ice core records (Alley and
371 Ágústsdóttir, 2005; Bond et al., 1997, 2001; Dansgaard et al., 1993; Lowe et al., 2008; Rasmussen et
372 al., 2006; Rohling and Pälike, 2005) and in the Mediterranean (e.g. De Rijk et al., 1999; Lirer et al.,

373 2013; Rohling et al., 2002; Sprovieri et al., 2003). In the ODP Site 976, the modification of the water
374 column structure indicated by calcareous plankton can be related to a southward displacement of the
375 ITCZ (Intertropical Convergence Zone) and an intensified impact of harsher, higher-latitude climate
376 conditions in the Mediterranean region (Rohling et al., 2002, 2004). Today, *N. incompta* does not
377 dwell in the Alboran Sea due to the occurrence of deep pycnocline and nutricline (located at a depth
378 of about 150-200m) and winter temperatures reaching 15°C (Pujol and Vergnaud Grazzini, 1995;
379 Rohling et al., 1995). On the other hand, this taxon is abundant in the Gulf of Lion where strong
380 winter mixing facilitates the advection of nutrients into the euphotic zone and, mainly, winter
381 temperatures reach 12°C (Pujol and Vergnaud-Grazzini, 1995; Rohling et al., 1995).

382 The marked increase of coccolithophore production within the upper part of the 8.2 ka event, as
383 indicated by the peak in total NAR (Fig. 2), is likely the result of an important hydrographic
384 modification occurring at this time, related to the enhanced Atlantic water inflow. This feature marks
385 the onset of the following phase II as discussed below and is very well comparable with a similar
386 peak in the coccolithophore absolute abundance at 8.2 ka recorded in the Alboran Sea by Colmenero-
387 Hidalgo et al. (2004) and related to the onset of gyre circulation into the basin.

388

389 *Phase II: the middle Holocene establishment of the modern oceanographic circulation*

390 Phase II represents the second major step in the hydrographic evolution of the basin. It develops
391 between 8.2 ka and about 5.3 ka, thus it nearly represents the middle Holocene portion of the record
392 (Bárcena et al., 2004; Giraudeau, 1993). It is marked by a distinct abundance increase of *F. profunda*
393 and a subsequent increase of *G. inflata* (Fig. 6) which replaces *N. incompta*. *Florisphaera profunda*
394 is a deep photic zone dweller and thrives with a deep nutricline and water column stratification
395 (Baumann et al., 2005; Incarbona et al., 2013; Sprovieri et al., 2012), while *G. inflata* is a deep living
396 taxon and benefits from water column stability, a deep pycnocline and reduced upwelling conditions.
397 The shift between *G. inflata* and *N. incompta* is in agreement with Rohling et al. (1995) that linked
398 this event to the establishment of the modern front-dominated conditions in the Alboran Sea, when
399 the amount of Atlantic water inflow was close to the present volume. At Site 976, the enhanced
400 Atlantic inflow, following the deglaciation and the sea level rise, would have deepened the nutricline
401 favoring *F. profunda*. In addition, it would have promoted both the development of the modern
402 geostrophic front, where *G. inflata* proliferates (Pujol and Vergnaud-Grazzini, 1995; Rohling et al.,
403 1995) and the establishment of WAG (Ausín et al., 2015b; Pérez-Folgado et al., 2003; Rohling et al.,
404 1995). This hydrographic evolution follows the culmination of the highest rate of global sea-level
405 rise (Lambeck et al., 2014). Concurrently to the development of a deep nutricline, high annual and
406 seasonal SSTs are recorded (Fig. 6) also marked by the increase of the tropical taxon *T. sacculifer*

407 (Fig. 6). Conversely, small *Gephyrocapsa* and neogloboquadrinids decrease (Fig. 6). We suggest that
408 anomalous sea-surface warmer conditions during this period promoted a prolonged water column
409 stratification, deepening of the thermocline (nutricline) and decrease of winter wind-induced mixing.
410 These conditions contributed to a higher increase of warm and oligotrophic taxa, that currently thrives
411 during mid-summer in the Mediterranean Sea (Bárcena et al., 2004; Pujol and Vergnaud-Grazzini,
412 1995), and decreasing of those taxa more related to nutrient-rich conditions such as small
413 *Gephyrocapsa* and neogloboquadrinids. The occurrence of *G. bulloides* during this phase (Fig. 6) is
414 consistent with its opportunistic behavior (Pujol and Vergnaud-Grazzini, 1995; Rohling et al., 1997;
415 Schiebel et al., 2001) and its favorite habitat, highly dependent on enhanced food availability, related
416 to strong seasonal contrast or river input. High abundances of temperate forest in the early stage of
417 phase II (Fig. 6) suggest still wet climate conditions on land. This phase, although coeval with the
418 younger portion of S1, is subsequent to the end of ORL 1 deposition in the Alboran Sea (Fig. 5). Deep
419 anoxia in the western basin is in fact independent of that of the eastern basin (Rogerson et al., 2008)
420 and ORL 1 termination is related to the 8.2 ka event and to the establishment of the modern front-
421 dominated conditions in the western Mediterranean (Cacho et al., 2002; Rogerson et al., 2008).
422 During phase II, differently from phase I occurring during ORL1 deposition and characterized by
423 shoaling of the nutricline and enhanced productivity in surface water (Bazzicalupo et al., 2018), the
424 calcareous plankton assemblages indicate stratified conditions in column water and deep nutricline,
425 which likely prevented productivity in surface water and export production at the sea floor. This
426 datum supports the hypothesis that productivity, although does not represent the triggering
427 mechanism, may provide a secondary control in the ORL formation (Rogerson et al., 2008).

428

429 *Phase III: the late Holocene reduced seasonality*

430 This phase is marked by the coeval increase of small *Gephyrocapsa* and *G. bulloides*, at c.a. 5.3 ka
431 (Fig. 6), suggesting increased nutrient availability in surface waters. These taxa are, in fact,
432 considered high surface water productivity proxies (Barcena et al., 2004; Colmenero-Hidalgo et al.,
433 2004; Gartner et al., 1987; Pujol and Vergnaud-Grazzini, 1995 Takahashi and Okada, 2000). The
434 enhanced abundances of the deep mixed dweller *T. truncatulinoides* (Fig. 6) support more intense
435 seasonal and prolonged mixing. Elevated abundances of *T. truncatulinodes* from sediment trap in the
436 Gulf of Lions have been related to increased winter mixing conditions (Rigual-Hernández et al.,
437 2012). On the other hand, the high abundance of *F. profunda* and *G. inflata* (Fig. 6) is still in relation
438 with the modern front-dominated conditions in the Alboran Sea and deep nutricline, originating at
439 the onset of phase II. Oscillations in the absolute abundances of *F. profunda* as well as of small
440 *Gephyrocapsa* (Fig. 6) are likely in relation with short-term fluctuations in total NAR, which are

441 discussed in more detail below. Foraminifera warm-water taxa, together with *G. ruber* group and to
442 a less degree warm water coccolith taxa, show a general decreasing trend (Fig. 5). The summer SST
443 record (Fig. 6) is in line with reducing summer insolation trend (Fig. 5) and with evidence from the
444 western Mediterranean terrestrial record of reduced seasonality (cooler summers and warmer winters)
445 during the transition to late Holocene (Ramos-Román et al., 2018). The aridification process,
446 accompanying the reduced seasonality at this time is highlighted, in the pollen record, by an increase
447 in *Artemisia* at around 4 ka at the studied core (Fig. 6), and by several coeval Mediterranean records
448 (Desprat et al., 2013; Fletcher et al., 2012; Fletcher and Sánchez Goñi, 2008; Jalali et al., 2016; Jalut
449 et al., 2000, 2009; Magny et al., 2013; Ramos-Román et al., 2018).

450 The last 3.5 kyr of phase III are marked by a relevant increase of *T. sacculifer* (Fig. 6). The distribution
451 pattern of this taxon is punctuated by three main short-term pulses (Ts1-Ts3 in Fig. 6), not previously
452 recorded in the western Mediterranean. *Trilobatus sacculifer* mainly occurs in warm and oligotrophic
453 tropical and sub-tropical waters with low seasonality (Bé and Hutson, 1977; Fraile et al., 2008;
454 Hemleben et al., 1989; Vincent and Berger, 1981). Today this taxon reaches its maximum abundance
455 in the Eastern Mediterranean basin and in the Red Sea, where low nutrient and warm surface waters
456 prevail throughout the year, due to the relatively stable deep pycnocline (Kallel and Labeyrie, 1997;
457 Kucera et al., 2005; Pujol and Vergnaud-Grazzini, 1995; Siccha et al., 2009). In the Red Sea, its
458 increasing trend, during the Holocene, has been also related to more arid conditions during reduced
459 monsoon climate system and prevailing eastern Mediterranean climate system (Edelman-Furstenberg
460 et al., 2009). In our record, the last 3.5 kyr are characterized by a reduction of seasonal thermal
461 gradient ($\Delta\text{SST}_{\text{sum-win}}$, Fig. 6) in the seawater, likely related to weak increase of winter SST,
462 concomitant with ameliorate climate condition on land, as suggested by coeval relative increases of
463 temperate forests in the pollen assemblages (Fig. 6). A positive correlation between *T. sacculifer* and
464 weaker winter conditions and stratification has been also found in the Arabian Sea (Munz et al.,
465 2015). We suggest that, at Site 976, relatively higher winter SSTs (with values exceeding 15°C) with
466 respect to the earlier interval, developed more stable year-round surface water conditions in the basin
467 favoring the increase of *T. sacculifer* in the last 3.5 kyr. Such conditions could probably represent the
468 response to changes in hydrological conditions in the adjacent Iberian basin, related to the reduction
469 of meltwater discharge in the North Atlantic (Bond et al., 2001). In the Gulf of Cádiz, according to
470 Schirrmacher et al. (2019), larger seasonal SST contrasts, during the Holocene, are related to periods
471 of enhanced iceberg discharge; the northward heat transport was blocked due to freshwater forcing
472 in the North Atlantic resulting in colder winter temperatures and higher summer temperatures due to
473 a seasonal northward migration of Intertropical Convergence Zone (ITCZ). This mechanism is similar
474 to the one proposed by Repschläger et al. (2017) for the early Holocene, when reinforcements of

475 northward heat transport and migration of Subtropical Gyre is found during periods of weak north
476 Atlantic meltwater discharge. Similarly, we retain that the decrease in the drift ice index during the
477 last 3 kyr (Bond et al., 2001), could have promoted higher northward advection of warmer water
478 masses that could have also reached the Alboran Sea through the Strait of Gibraltar, favoring the
479 instauration of a lower seasonal thermal gradient. The three distinct peaks of *T. sacculifer*, centered
480 at about 2.9 ka, 1.8 ka and 0.7 ka, trace the occurrence of short warm pulses at the core location. They
481 appear chronologically correlated with the short-term warm and dry events identified in northwestern
482 Africa lakes and in the Adriatic Sea (Piva et al., 2008; Zielhofer et al., 2017). The phase Ts1 is also
483 chronologically correlated with the warm phase recognized by Margaritelli et al. (2016) during the
484 Middle Bronze Age–Iron Age in the central Mediterranean.

485

486 *Millennial scale variations*

487 Since the WAG establishment in the Alboran Sea at ca. 8 ka, total NAR values show a series of
488 millennial-scale fluctuations over the middle and late Holocene at Site 976 (Fig. 7). This pattern
489 indicates that high-frequency variations in the coccolithophore productivity are superimposed to the
490 main climate phases. Connecting coccolithophore productivity to environmental proxies is a complex
491 task since multiple relationships might affect the link between the various signals. A certain
492 chronological uncertainty is also added, when comparing different sites with different age models. In
493 order to unravel the forcing mechanism responsible for coccolithophore productivity variations at the
494 studied core, we compared a few coccolithophore proxies with the pattern of $\delta^{18}\text{O}_{\text{seawater}}$ available at
495 the ODP Site 976 (Jiménez-Amat and Zahn, 2015), as a proxy of local surface water salinity variation.
496 We have also performed a comparison with the detrended $\delta^{18}\text{O}_{\text{speleothem}}$ curve (Smith et al., 2016),
497 which represents a high-resolution archive of quasi-cyclical events of relatively wet-to-dry climatic
498 conditions over Iberia, with a ~1500 year frequency (Smith et al., 2016). This trend is significantly
499 correlated with the NAO index (Olsen et al., 2012) (Fig. 7). A relationship between coccolithophore
500 productivity and NAO modes has been recently suggested in the Alboran Sea by Ausín et al. (2015a).
501 These authors indicate weakened (intensified) upwelling, related to weaker (stronger) westerlies,
502 responsible for reduced (reinforced) WMDW in the Gulf of Lions. In this scenario, the NAO
503 circulation mode is the forcing mechanism of coccolithophore variability. Intensified upwelling
504 would have been promoted by stronger westerlies blowing over the Gulf of Lions, during a NAO –
505 mode. These conditions would have promoted major WMDW formation and simultaneous
506 enhancement of the AJ influx, both fluctuating in-phase (Ausín et al., 2015a; García Lafuente et al.,
507 2002, 2007) . The AJ would have migrated southward, allowing the cool subsurface waters to fill the
508 area left behind the jet (Sarhan, 2000) and thus promoting upwelling. In the present work, a first

509 comparison between datasets indicates that low salinity phases (lighter $\delta^{18}\text{O}_{\text{seawater}}$) are concurrent
510 with high values of *G. oceanica* (Fig. 7). The occurrence of *G. oceanica* within the western
511 Mediterranean basin has been often related to Atlantic surface water inflows (Álvarez et al., 2010;
512 Bárcena et al., 2004; Bazzicalupo et al., 2018; Knappertsbusch, 1993; Oviedo et al., 2017). The
513 positive correlation of the coccolithophore taxon with salinity minima at Site 976, further supports
514 the relation between *G. oceanica* and Atlantic surface water inflows, since salinity minimum in the
515 Alboran Sea essentially traces phases of enhanced Atlantic water into the basin (Font et al., 1998;
516 Sarhan et al., 2000; Viúdez et al., 1996). Consequently, we use the lighter values of $\delta^{18}\text{O}_{\text{seawater}}$ and
517 the increased abundance of *G. oceanica* as proxies of Atlantic inflow in the Alboran Sea (Fig. 7),
518 which both provide a regime of cyclical Atlantic water inflow intensity in the basin. Coccolithophore
519 productivity variations, expressed as total NAR, display distinct pulses, well-correlated with the
520 pattern of the changing Atlantic inflow intensity and with the concomitant occurrence of alternating
521 dry/wet phases in the Iberia $\delta^{18}\text{O}_{\text{speleothem}}$ (Fig. 7). Therefore, the various proxies point out to a
522 coupling between enhanced coccolithophore productivity (high total NAR values), intensified
523 Atlantic waters inflow (lighter $\delta^{18}\text{O}_{\text{seawater}}$ and increased abundance of *G. oceanica*), and arid
524 conditions over the Iberia Peninsula (peaks in $\delta^{18}\text{O}_{\text{speleothem}}$), correlated with NAO+ phases (Fig. 7).
525 Our data support the model proposed by Ausín et al. (2015a), and specifically the relation between
526 coccolithophore productivity, Atlantic inflow and WMDW strength, although the dataset at site 976
527 indicate an opposite relationship between coccolithophore productivity and NAO mode. According
528 to the present results, enhanced Atlantic water inflow occurred during a persistent NAO+ index (Fig.
529 7); the latter would have strengthened the north-westerlies over the northwestern Mediterranean
530 basin, promoting a reinforcement of deepwater overturning and in turn increased the AJ (Fig. 8). The
531 suggested relation between NAO mode and WMDW strength is in agreement with results from
532 today's survey in the western Mediterranean (Rixen et al., 2005) and with the proposed relationship
533 between strengthening of the WMDW and NAO variability in the past. In fact, during the Holocene
534 and the Dansgaard-Oeschger events NAO + phases would have strengthen the northwesterlies over
535 the northwestern Mediterranean, enhancing the WMDW formation (Frigola et al., 2007; Moreno et
536 al., 2002, 2004, 2005; Nieto-Moreno et al., 2011). The scenario is also consistent with the observed
537 decadal-variability between NAO intensity and upwelling strength highlighted in the western
538 Mediterranean (Vargas-Yáñez et al., 2008). A possible reasonable explanation for the differing
539 interpretations between the present work and the Ausín et al. (2015a) study, is that the latter authors
540 based their paleoceanographic reconstruction on oscillations of the *F. profunda* NAR abundances in
541 the Alboran Sea. In Ausín et al. (2015a), *F. profunda* NAR peaks have been linked to the
542 intensification of the upwelling conditions in the area. Recent data establish a precise relationship

543 between *F. profunda* and primary productivity levels in today's low-latitude oceans (Hernández-
544 Almeida et al., 2019) and suggest that the link between *F. profunda* abundance and net primary
545 productivity in the Mediterranean Sea is not straight forward thus discouraging the use of this taxon
546 as a productivity indicator into the basin (Hernández-Almeida et al., 2019).

547

548 *External and internal forcing mechanisms of coccolithophore productivity cycles*

549 The time series analysis performed on the total NAR record confirms the occurrence of millennial-
550 scale periodicities in coccolithophore productivity during the Holocene and highlights two main
551 periodicities through the record: the first one of ~ 1100 yr (from 12.5 to c.a. 5 ka) and the second one
552 of ~ 1700 yr (from 5 to 0.19 ka) (Fig. 4c). The results of spectral and wavelength analyses indicate
553 that coccolithophore productivity changes in the Alboran Sea were modulated both by external (solar)
554 and internal (oceanic-atmospheric) forcing. In fact, the ~ 1100 yr-cycles appear close to the ~ 1000
555 yr cycle identified during the early Holocene in solar proxies in North Atlantic records and in IRD
556 record (Debret et al., 2007, 2009). They are also detected in the western Mediterranean pollen record,
557 which displays a periodic component of ~ 900 yr (Fletcher et al., 2012) and of ~ 1100 yr (Ramos-
558 Román et al., 2018) during the early and middle Holocene. On the other hand, cycles of ~ 1700 yr
559 are very close to the 1600-year cycle dominating during the last 5000 yr in several paleoclimate
560 records (Debret et al., 2007; 2009 and references therein) and related to internal
561 (oceanic/atmospheric) forcing. A similar shift in periodicity to a dominant ~ 1750 oscillation in the
562 last 6 ka (Fletcher et al., 2012) and ~ 1600 yr-cycle (Ramos-Román et al., 2018) in the last 4.7 ka has
563 been detected in the western Mediterranean in the pollen record and is related to the influence of
564 NAO-like circulation in the mid-late Holocene. The similar pattern in cyclicity observed in the
565 present study in NAR pattern in the Alboran Sea strengthens the relation between coccolithophore
566 productivity/hydrographic changes and atmospheric variability modulated by NAO fluctuations and
567 sustains the occurrence of a periodicity change through the Holocene from a dominant external (solar)
568 to a dominant internal (oceanic/atmospheric) forcing.

569

570 **Conclusions**

571 The calcareous plankton assemblage (coccolithophore and foraminifera) of the ODP Site 976 from
572 the Alboran Sea has been studied at a centennial-scale resolution, to investigate the climate variability
573 and the forcing mechanisms affecting the western Mediterranean basin during the last 12.5 ka.
574 Coccolithophore and planktonic foraminifera dataset is integrated with pollen and geochemical data
575 available at the site. During a first step, between 11.5 and 8.2 ka, calcareous plankton assemblage
576 clearly traces increasing temperature and freshwater arrival, related to riverine input in the basin,

577 during a period of insolation maxima. The timing of this phase in the Alboran Sea is quasi
578 concomitant with sapropel S1 deposition in the eastern Mediterranean, suggesting a connection
579 between the monsoonal mechanism for sapropel formation and high rainfall conditions in
580 Northwestern Europe. Following an abrupt surface water temperature decrease correlated with
581 humidity reduction and centered at 8.2 ka, the second phase (8-4.6 ka) is marked by a profound
582 change in the planktonic assemblages, reflecting a more stratified water column, the deepening of the
583 nutricline following a sea level rise and the instauration of the modern gyre circulation. The third
584 final phase (4.6-0.19 ka) is characterized by reduced seasonality (cooler summers and warmer
585 winters), enhanced surface water mixing and increased aridification on land related with a decrease
586 in summer insolation. Short-term cyclicality occurs in coccolithophore productivity, with a clear pattern
587 mainly occurring since the establishment of the modern circulation. Millennial-cycles of increased
588 coccolithophore productivity are associated with enhanced inflows of Atlantic water from the
589 Gibraltar strait modulated by NAO+ mode. The proposed scenario strengthens the role of
590 hydrographic changes and atmospheric variability modulated by NAO fluctuations on
591 coccolithophore productivity in the Alboran Sea. The results of the spectral analysis add information
592 on the value of coccolithophores in recording environmental changes and highlight that
593 coccolithophore productivity is modulated by both external (solar) and internal (oceanic-
594 atmospheric) forcing. A shift in periodicity from a dominant ~ 1100 yr oscillations to ~ 1600 yr
595 periodicity occurs at about 4 ka and appears in agreement with enhanced influence of NAO-like
596 circulation during the late Holocene.

597

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1141

1142 **Figure captions**

1143

1144 *Fig. 1:* Location of ODP Site 976 in the Alboran Sea (western Mediterranean), bathymetry of the
1145 area and modern-day oceanographic circulation. **AW** (Atlantic Water); **MOW** (Mediterranean
1146 Outflow Water); **WMDW** (western Mediterranean Deep Water); **LIW** (Levantine Intermediate
1147 Water); **WAG** (western Alboran Gyre); **EAG** (eastern Alboran Gyre). In violet shade: Alboran and
1148 Almeria-Oran upwelling fronts

1149

1150 *Fig. 2:* Downcore variations of calcareous nannofossil assemblages at Site 976 plotted as relative
1151 abundance (% , black line) and nannofossil accumulation rate - NAR (coccolith/cm² kyr, filled area).
1152 Sedimentation rate over time used for NAR calculation, from Martrat et al. (2014), is also shown.
1153 YD: Younger Dryas.

1154

1155 *Fig. 3:* Downcore variations of planktonic foraminifera assemblages at Site 976 plotted as relative
1156 abundance (% , black line) and planktonic foraminifera accumulation rate – pfAR (forams/cm²kyr,
1157 filled area), together with foraminifera-based summer, winter and annual SST and similarity index.
1158 Sedimentation rate over time, used for pfAR calculation, from Martrat et al. (2014). YD: Younger
1159 Dryas.

1160

1161 *Fig. 4:(a)* Signal of the Total NAR decomposed with CEEMD in five IMFs plus a residue (trend);
1162 (b), (c), (d), (e) spectral analysis made with “REDFIT” and Foster’s WWZ, of the IMFs extracted
1163 from Total NAR. The green and black line represent the 95% and 80% Confident Level
1164 respectively. Significantly periodicity (red dot) and relative values expressed in years were

1165 reported.

1166

1167 *Fig. 5: Abundances variations of calcareous plankton assemblage and additional proxies from Site*
1168 *976: accumulation rate of selected coccolithophores and planktonic foraminifera; relative*
1169 *abundance patterns of selected pollen taxa at Site 976 from Combourieu-Nebout et al. (2009); black*
1170 *line, 3 point average. Di- and tri-unsaturated alkenones of 37 carbons (C₃₇) from Martrat et al.*
1171 *(2014) and summer insolation curve (Laskar et al., 2004) are also shown. Younger Dryas (YD, grey*
1172 *bar); 8.2 ka event (light blue bar); dashed black lines are used to trace boundaries among phases I-*
1173 *III.*

1174

1175 *Fig. 6: Abundances variations of calcareous plankton assemblage and additional proxies from Site*
1176 *976: accumulation rate of selected coccolithophores and planktonic foraminifera; black line, 3 point*
1177 *average; foram based seasonal SST variations at Site 976; relative abundance patterns of selected*
1178 *pollen taxa at Site 976 from Combourieu-Nebout et al. (2009). 8.2 ka event (dotted bar),*
1179 *dashed black lines are used to trace boundaries among phases I-III.*

1180

1181 *Fig. 7: Abundances variation of coccolithophore assemblage and climate proxies from Site 976: *G.**
1182 **oceanica* absolute abundances (black line, 3 point average); $\delta^{18}\text{O}_{\text{seawater}}$ at Site 976 (green line, 3*
1183 *point average) (Jimenez-Amat and Zahn, 2015); $\delta^{18}\text{O}$ of combined and de-trended speleothems*
1184 *from Iberian Peninsula (Smith et al., 2016); coccolithophore productivity (total Nannofossil*
1185 *Accumulation Rate) at Site 976 (black line, 3 point average). Inferred NAO circulation pattern from*
1186 *redox variability from Lake SS1220, Greenland (Olsen et al., 2012) is also shown. Light blue bars*
1187 *represent periods of increased total NAR concomitant with enhanced Atlantic inflow and positive*
1188 *NAO index phases.*

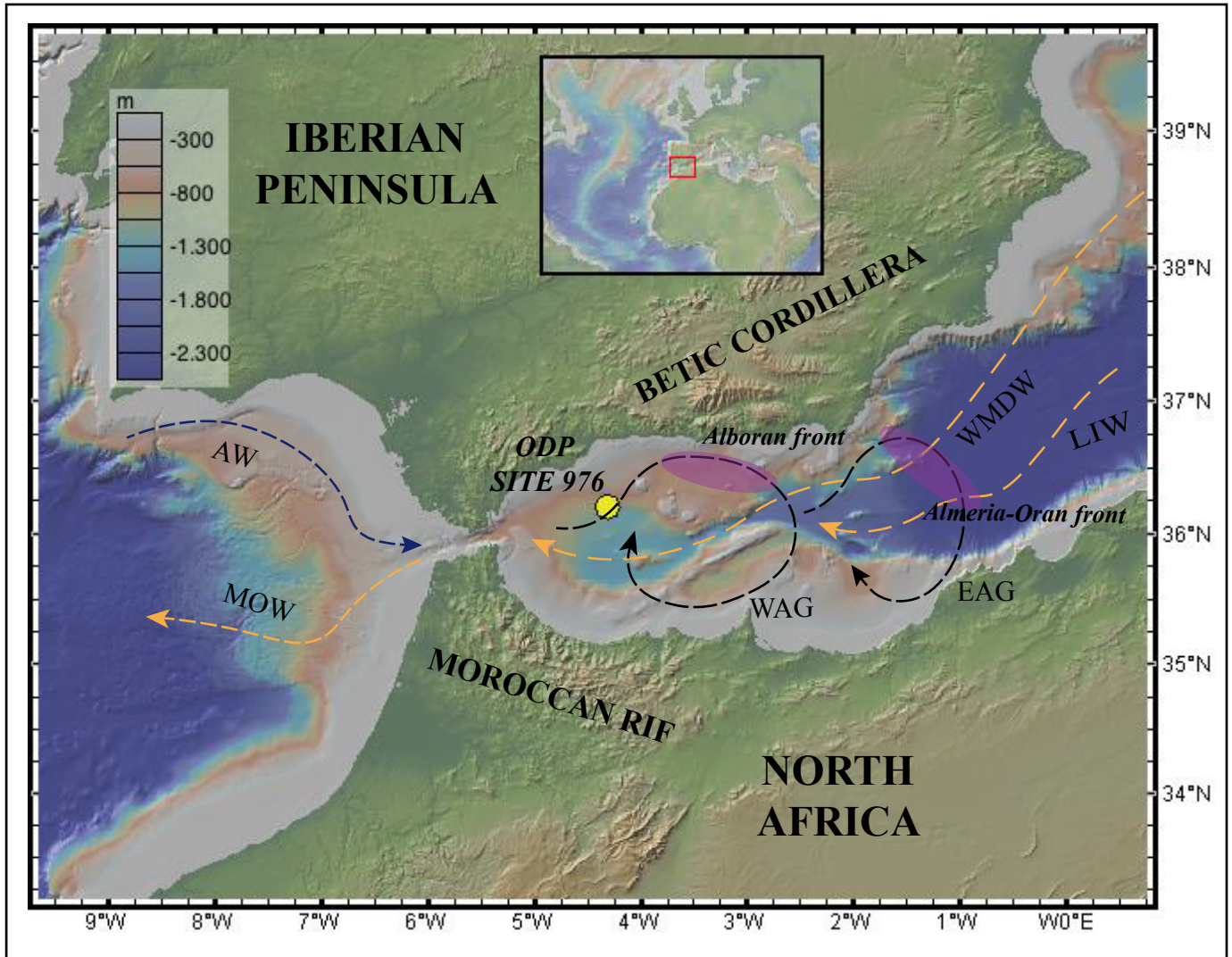
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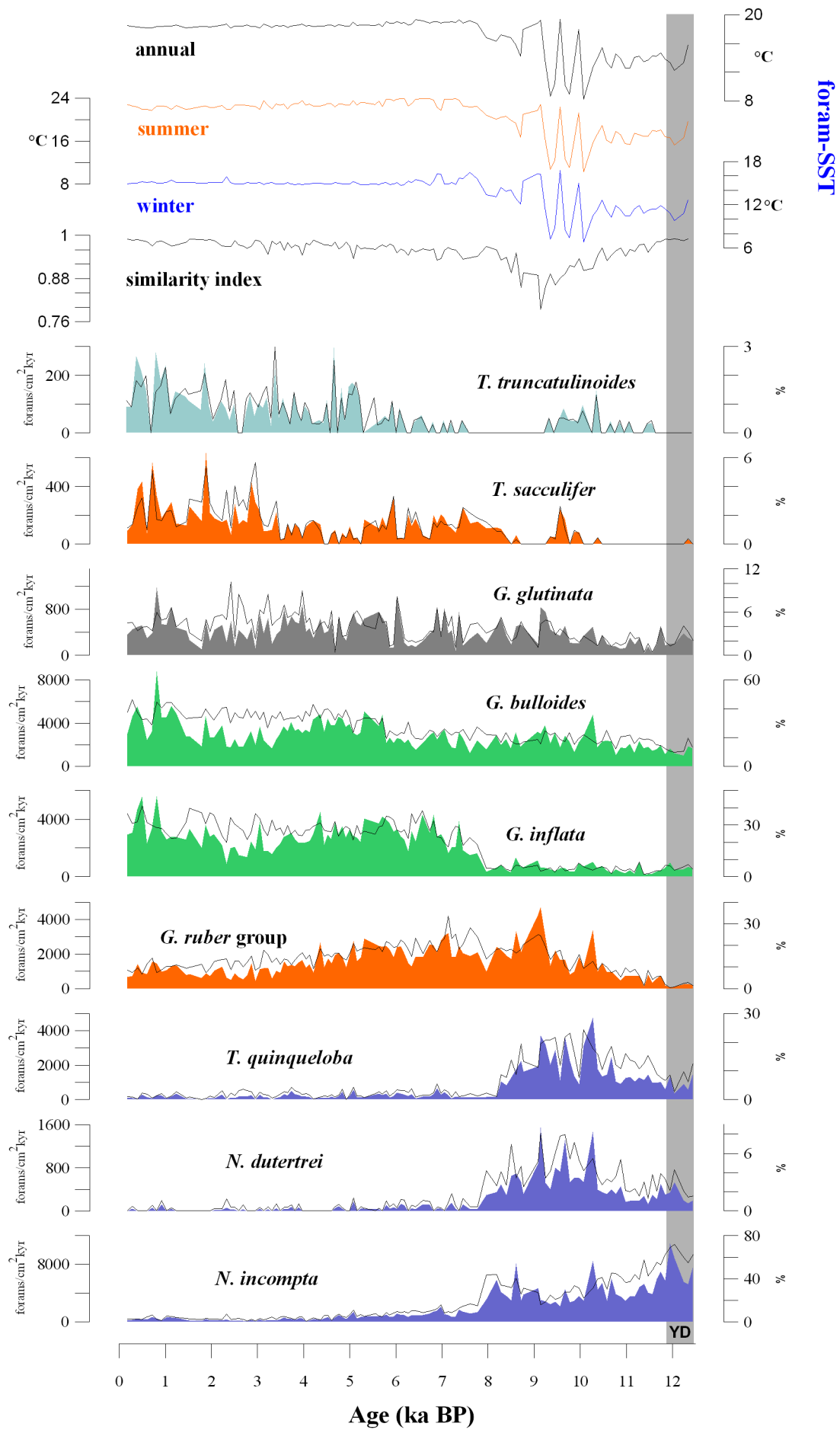
1190 *Fig. 8: Proposed different NAO circulations pattern scenarios as explained in the text: a) **NAO+***
1191 *enhanced northwesterly winds, deep water formation and Atlantic inflow inducing upwelling and*
1192 *coccolithophore productivity; b) **NAO-** reduced northwesterly winds, deep water formation and*
1193 *Atlantic inflow, inducing stratification and reduced coccolithophore productivity. **LIW** (Levantine*
1194 *Intermediate Water). **AJ** (Atlantic Jet); **WMDW** (western Mediterranean Deep Water). **MOW***
1195 *(Mediterranean Outflow Water). Diagram not to scale.*

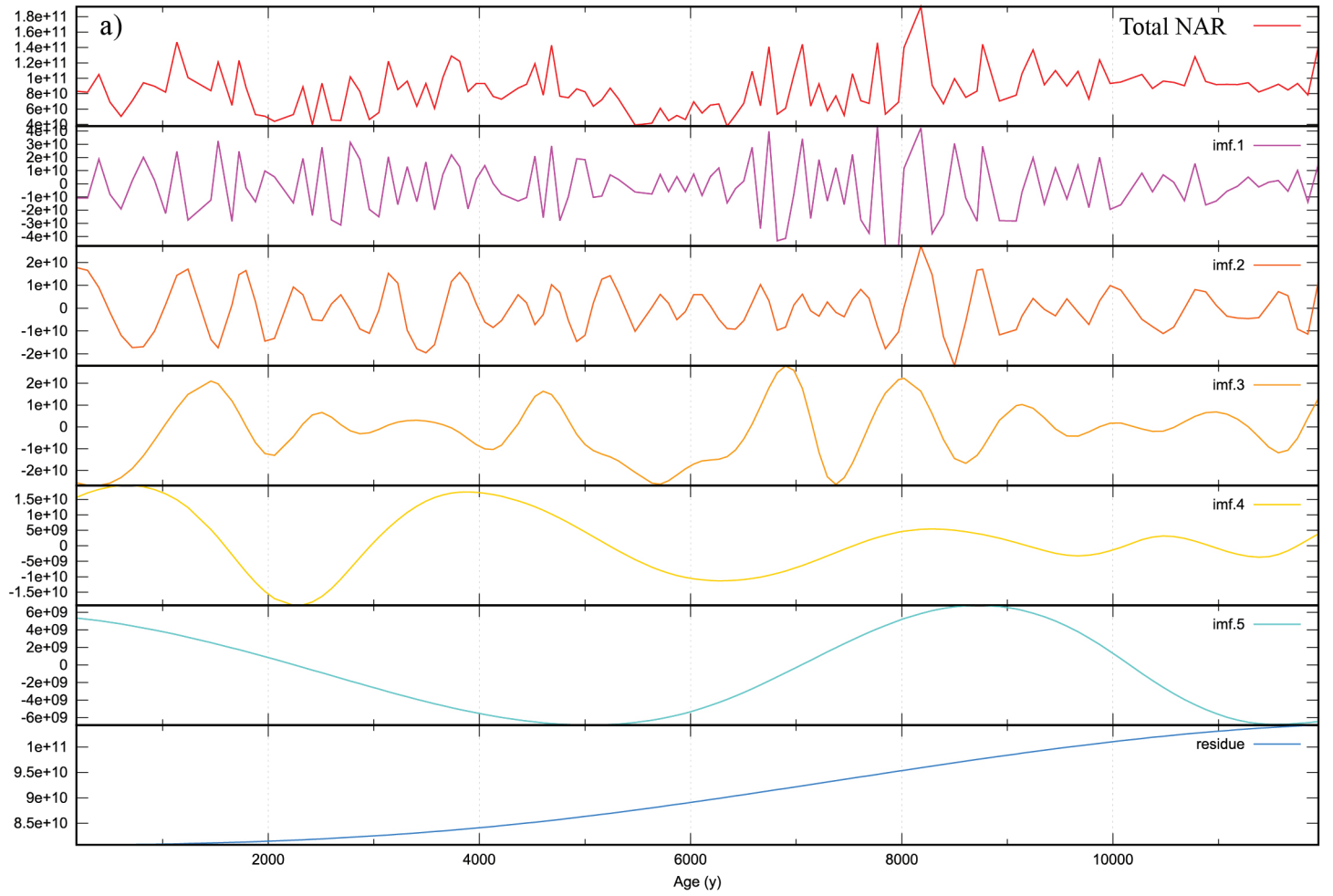
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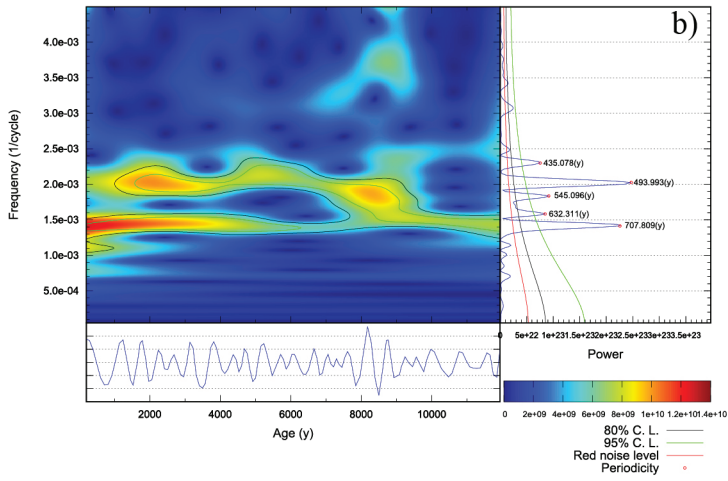
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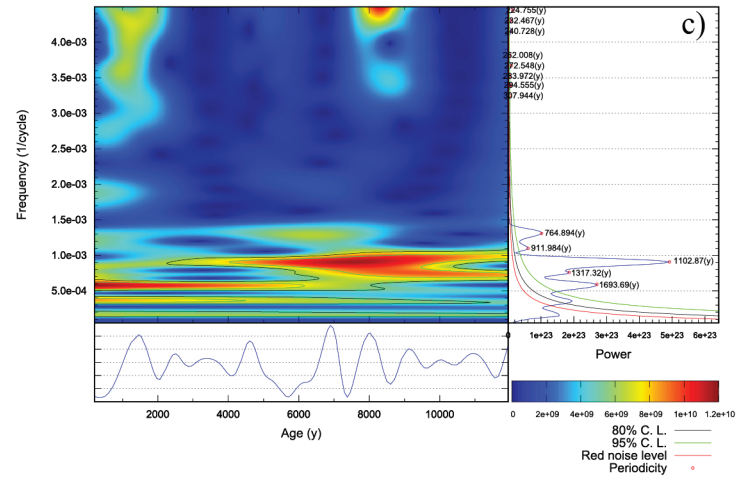




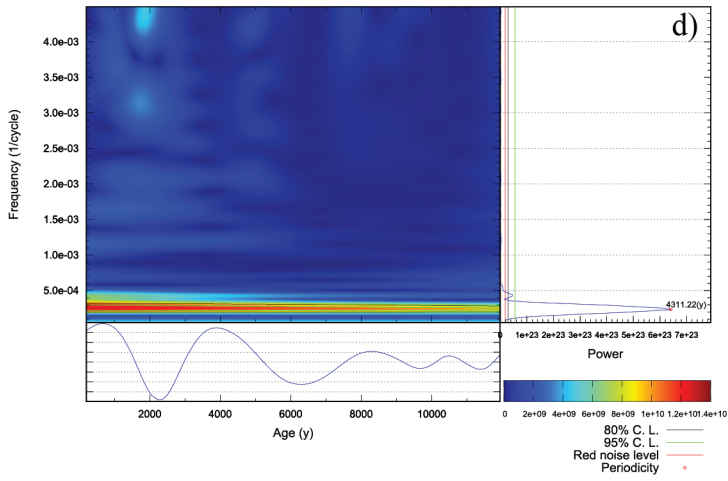
Total NAR IMF2



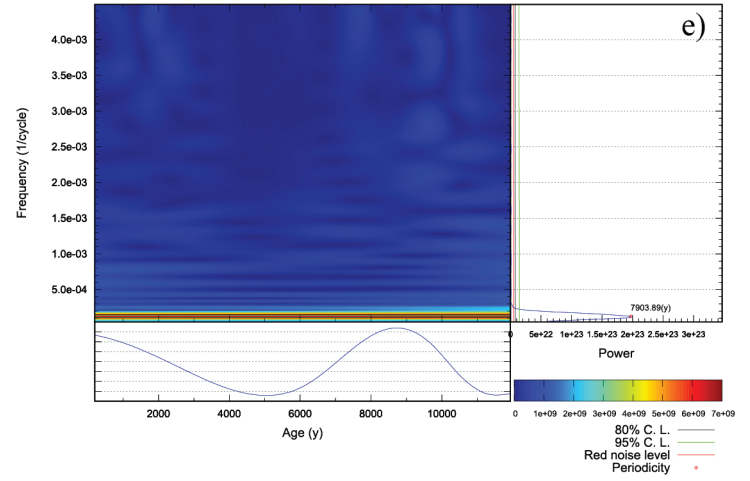
Total NAR IMF3

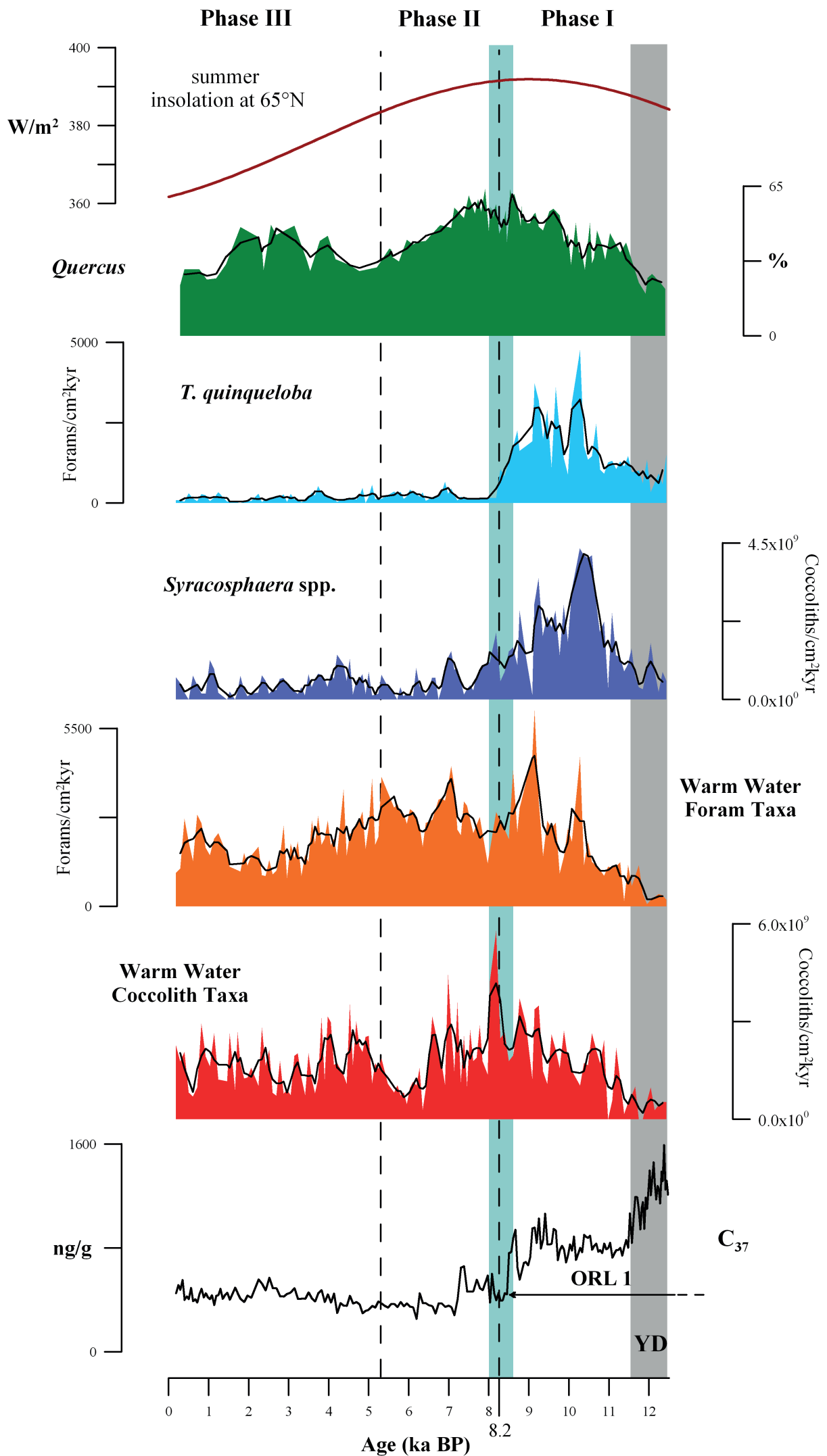


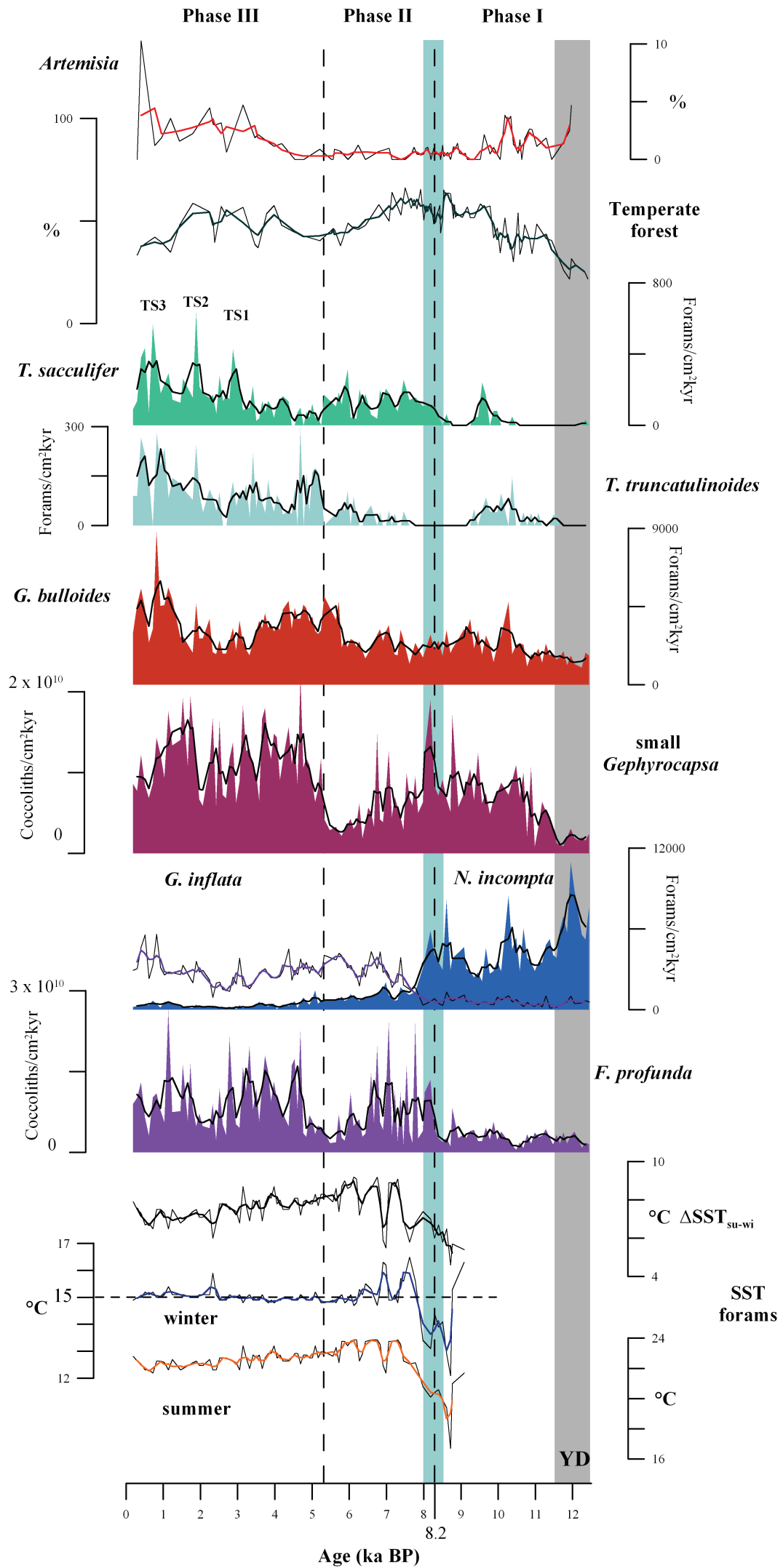
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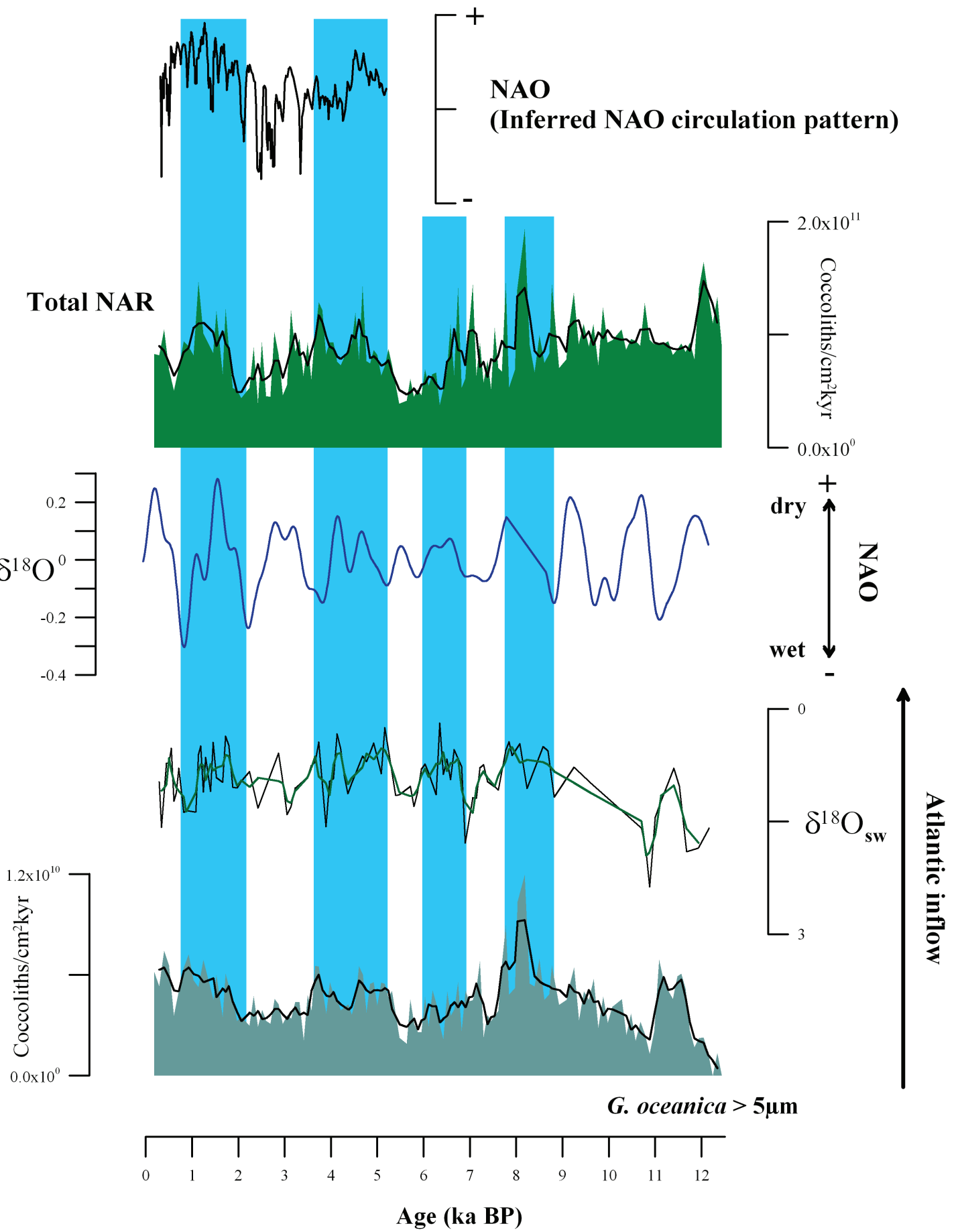
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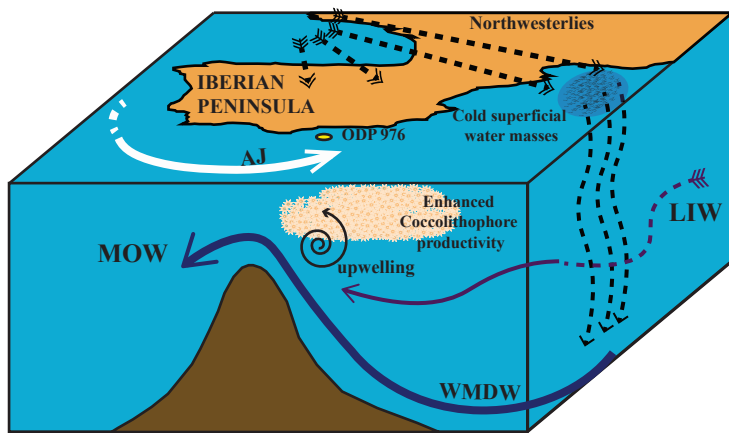




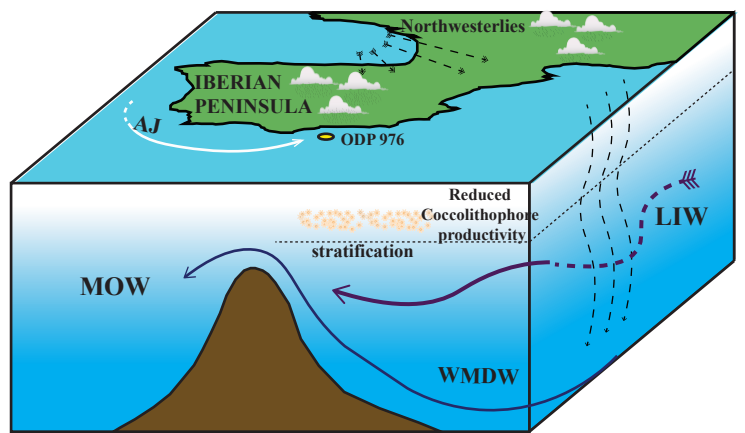


combined and de-trended speleothems
Iberian peninsula





a)



b)