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1 **Contrasting sea-surface responses between the western Mediterranean Sea and eastern**
2 **subtropical latitudes of the North Atlantic during abrupt climatic events of MIS 3**

3
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16 **ABSTRACT**

17 Dinoflagellate cyst (dinocyst) analysis was conducted on two cores from the SW Iberian margin
18 and central Alboran Sea from which high quality records of Marine Isotope Stage 3 have been
19 previously derived. Our aim in this study is to compare the dinocyst signature between 50 and 25
20 ka BP with existing datasets of foraminiferal and geochemical proxies related to hydrological
21 parameters. Quantitative reconstructions of sea-surface temperatures (SSTs) and salinities (SSS)
22 based on dinocysts are performed for the first time in this area. The results are compared to SSTs
23 derived from planktonic foraminifera and alkenone measurements, and to SSS calculated from
24 planktonic $\delta^{18}\text{O}$ and foraminiferal SST. Significant oscillations related to Dansgaard-Oeschger
25 cycles are recorded in both cores. Dinocyst-derived hydrological parameters exhibit synchronous
26 fluctuations and similar values to those derived from the other methods, in particular when
27 considering quantitative reconstructions for February based on foraminifera and dinocysts. Our
28 study shows that the influence of subpolar waters was felt during each Greenland Stadial (GS) off
29 Portugal, and that the amplification of the Heinrich Stadial cooling in the Alboran Sea was
30 related to the penetration of subpolar waters through the Strait of Gibraltar. During Greenland
31 Interstadials (GI), we provide evidence for the occurrence of warm and nutrient-rich sea-surface
32 waters in the Alboran Sea, probably due to gyre-induced upwelling. Finally, the difference
33 between August and February dinocyst SST estimates suggests higher seasonal contrasts during
34 GS compared to GI at the two core sites. Additionally, precession appears to have an imprint on
35 dinocyst-derived long-term seasonality record. However, this observation needs to be confirmed
36 by longer records.

37 *KEYWORDS: Dinocysts; SST and SSS quantification; Dansgaard-Oeschger; Greenland and*
38 *Heinrich stadials; Alboran Sea; Iberian margin.*

39 **1. Introduction**

40

41 The climate of the last glacial was characterised by a distinctive oscillatory mode, the so-called
42 Dansgaard-Oeschger (D-O) cycles, originally described and defined in Greenland ice cores
43 (Johnsen et al., 1992; Dansgaard et al., 1993; Grootes et al., 1993) and paced by an empirical
44 1470-year factor, the nature of which being presently not known (Schulz et al., 1999; Moreno et
45 al., 2005; Voelker et al., 2006). Greenland Interstadials (GI) and Greenland Stadials (GS)
46 including Heinrich Stadials (HS, following Sánchez-Goñi and Harrison, 2010), have previously
47 been observed in marine climate records worldwide (e.g. Voelker et al., 2002; Hemming, 2004;
48 Clement and Peterson, 2008). The impacts of HS, which primarily correspond to large ice-sheet
49 collapses, are now fairly well characterised in the NE Atlantic and Mediterranean areas. Their
50 occurrence is associated with enhanced aridity in the adjacent borderlands (Combourieu-Nebout
51 et al., 2002; Sánchez-Goñi et al., 2002), increases in northward Saharan dust transport (Moreno et
52 al., 2002; Bout-Roumazielles et al., 2007) and sea-surface temperature coolings in the western
53 Mediterranean Sea (e.g. Cacho et al., 1999) and on the Portuguese margin (e.g. Cayre et al.,
54 1999; Pailler and Bard, 2002; de Abreu et al., 2003). However, the forcing mechanisms behind
55 the initiation of HS and the inferred reorganisation of Meridional Overturning Circulation within
56 the last glacial period are still a matter of debate (Kageyama et al., 2009). Furthermore, a full
57 understanding of this rapid transmission of millennial-scale climatic variability from boreal to
58 subtropical latitudes, involving both atmospheric and oceanic processes, remains elusive. It has
59 been demonstrated that convection in the western Mediterranean, and thus export of
60 Mediterranean Outflow Water (MOW), was strengthened during GS of the last 50 ka (Cacho et
61 al., 2000, 2006; Sierro et al., 2005; Voelker et al., 2006; Toucanne et al., 2007; Frigola et al.,
62 2008). Consequently, it has been suggested that the Mediterranean could act as a major trigger for

63 the Meridional Overturning Circulation to switch from stadial to interstadial mode through the
64 influence of MOW, providing saline water to the North Atlantic at times when the Meridional
65 Overturning Circulation was partially interrupted (Johnson, 1997; Bigg and Wadley, 2001;
66 Rogerson et al., 2006; Voelker et al., 2006). The production of MOW is mainly controlled by the
67 salinity budget of the Mediterranean Sea which depends on the climate characteristics over the
68 region and therefore on dominant modes of climatic variability, notably the North Atlantic
69 Oscillation (NAO). At present, the NAO pattern, oscillating at decadal and centennial scales, has
70 a strong influence on climate over a large part of Europe during winter, by affecting the storm
71 tracks and the associated relative moisture over the Mediterranean and northern Europe (e.g.
72 Hurrell, 1995; Serreze et al., 1997; Osborn et al., 1999). Over the eastern Atlantic subtropical
73 sector, it has been argued that this oscillation also has an imprint at centennial or millennial
74 timescales, and has been implicated in abrupt climatic events of the last glacial (Sánchez-Goñi et
75 al., 2002; Moreno et al., 2005; Bout-Roumazeilles et al. 2007; Daniiau et al., 2007; Naughton et
76 al., 2009). Consequently, a change between two periods, each of them characterised by sustained
77 frequencies of a particular atmospheric configuration over several centuries, may have caused
78 significant variations of hydrological exchanges through time between the Atlantic and the
79 Mediterranean Sea (Moreno et al., 2005).

80 Previous high resolution palynological analysis of dinoflagellate cysts (dinocysts) and pollen on
81 core MD95-2042, collected off Portugal, indicate abrupt changes in sea surface and atmospheric
82 conditions, respectively, in response to D-O cycles (Eynaud, 1999; Sánchez-Goñi et al., 2000).
83 Additionally, a climatic contrast has been observed with respect to precipitation between the SW
84 Iberian margin and the Alboran region, with dryer conditions in southeastern Iberia than in its
85 southwestern part (Sánchez-Goñi et al., 2002). However, these studies do not quantify either
86 changes in SST and SSS, or discuss the possible impact of the observed precipitation gradient on

87 the hydrological conditions of both sides of the Strait of Gibraltar between 25 and 50 ka BP. To
88 fill this gap, we have analyzed dinocyst assemblages from core MD95-2043, located in the
89 Alboran Sea, and applied transfer functions. Furthermore, we have tested, for subtropical
90 latitudes, the robustness of quantitative paleotemperature and paleosalinity reconstructions
91 inferred from dinocysts against other proxy reconstructions (i.e. dinocyst- *versus* foraminifera-
92 *versus* alkenone- derived SST, and dinocyst- derived SSS *versus* SSS estimates calculated on the
93 basis of the $\delta^{18}\text{O}$ of *Globigerina bulloides* associated with foraminiferal SST).

94 **2. Core locations and present-day environmental settings**

95

96 Calypso cores MD95-2043 (36°8.6'N; 2°37.3'W; 1841 m water depth; 36 m long) and MD95-
97 2042 (37°48'N; 10°10'W; 3146 m water depth; 39.56 m long) were retrieved from the central
98 Alboran Sea and the SW Iberian margin, respectively (Fig. 1), and were both collected by the
99 oceanographic R/V Marion Dufresne during the 1995 International Marine Global Change
100 Studies I (IMAGES I) cruise (Bassinot and Labeyrie, 1996).

101 The Mediterranean Sea is located between the path of the mid-latitude westerlies which dominate
102 northern and central Europe, and the Azores High. At present, this basin experiences a typical
103 seasonal cycle marked by cool-wet winters, and warm-dry summers. The resulting Mediterranean
104 water budget is marked by a strong freshwater deficit due to a net excess evaporation (Béthoux,
105 1979, 1984). During winter and spring, intense cold and dry continental air outbursts induce
106 strong evaporation and cooling in the northern Mediterranean Sea, and thus an increased density
107 of surface water masses. As a result, surface waters sink in several specific Mediterranean
108 regions, flow westward, and finally form the Mediterranean Outflow Water (MOW). The outflow
109 current is exported at depth towards the Atlantic Ocean, through the narrow Strait of Gibraltar
110 (main sill depth of 280 m) and is split into two stability levels: an upper core centred between 500
111 m and 800 m, and a more saline and dense lower core found between 1000 and 1400 m (Ambar
112 et al., 2002). The MOW can be traced in the North Atlantic as a salinity and temperature
113 maximum at about 1 km depth (e.g. Hill and Mitchelson-Jacob, 1993; Iorga and Lorzier, 1999;
114 O'Neill-Baringer and Price, 1999), and mixes progressively with North Atlantic Intermediate
115 Waters (NAIW; Mauritzen, 1996). In the opposite direction, a North-Atlantic low-salinity surface
116 current penetrates the Alboran Sea and mixes with Levantine Intermediate Waters (which
117 resurface in the westernmost part of the Alboran Sea), forming the so-called Modified Atlantic

118 Waters (MAW). The strong flow of MAW along the coast of Spain initiates the formation of two
119 anticyclonic gyres, the Western and Eastern Alboran Gyres (WAG and EAG; Fig. 1) whose
120 position and intensity fluctuate at a seasonal scale. The degree of development of the EAG, under
121 which core MD95-2043 is located (Fig. 1), controls the position and intensity of the permanent
122 Almeria-Oran Front (AOF; Fig. 1) (Tintoré et al., 1988; Rohling et al., 1995, 2009; Viúdez and
123 Tintoré, 1995). Upwelling cells occur along the AOF, this front marking the deflection of MAW
124 along the Algerian margin, forming the Algerian Current (AC; Fig. 1).

125 The main modern hydrological structures and currents of the western Iberian margin form part of
126 the North Atlantic Eastern Boundary Current (Fig. 1; Peliz et al., 2005; Relvas et al., 2007). They
127 are driven by the North Atlantic subtropical gyre intensity in relation to the seasonal
128 displacement of the Azores High (e.g. Fiúza, 1984; Fiúza et al., 1998). Core MD95-2042 is
129 located in the seasonal coastal upwelling band of the Portugal-Canary eastern boundary
130 upwelling system that is active mainly from April to October (Aristegui et al., 2005; Peliz et al.,
131 2005). The upwelling predominantly receives North Atlantic Central Waters (NACW) and, in
132 part, also MOW (Sánchez and Relvas, 2003). Occasionally, during winter months, a warm
133 northward-flowing surface current known as the Iberian Poleward Current or Portugal Coastal
134 Counter Current is formed by coastal convergence along the western Iberian margin. This
135 phenomenon occurs when weak northerlies are interspersed with strong south-southwesterly
136 winds (e.g. Fiúza et al., 1998; Peliz et al., 2005). This current can be traced as far as the Bay of
137 Biscay where it is known as the “Navidad Current”. Winter warming in the southern Bay of
138 Biscay during Navidad years has been correlated with low values of the NAO index and this
139 current was found to extend from Portugal to Norway in exceptional Navidad years (Garcia-Soto
140 et al., 2002).

141 Modern sea-surface conditions of the SW Iberian margin and Alboran Sea are characterised
142 respectively by mean annual salinities of 36.5 and 37.1 psu (world dataset atlas compiled by
143 Schmidt, 1999 and Bigg and Rohling, 2000, <http://data.giss.nasa.gov/o18data>: Craig and Gordon
144 dataset for the Iberian margin and C. Pierre dataset for the Alboran Sea), February SSTs are
145 around 15.2 and 14.7°C, respectively, and August SSTs are around 20 and 23.7°C (WOA 2001).

146 **3. Methodology**

147

148 **3.1. Chronostratigraphy**

149

150 Cores MD95-2043 (central Alboran Sea) and MD95-2042 (SW Iberian margin), characterised by
151 high sedimentation rates, are composed mainly of calcareous hemipelagic clays and have yielded
152 high resolution paleoceanographic records (e.g. Cacho et al., 1999 and Shackleton et al., 2000,
153 respectively).

154 The age model of core MD95-2042 is derived from 16 AMS ^{14}C dates and, for the period beyond
155 AMS ^{14}C range, by graphical tuning of the MD95-2042 planktonic $\delta^{18}\text{O}$ record with Greenland
156 ice core $\delta^{18}\text{O}$ records. The GISP 2 chronology (Bard et al., 2004) was used between 26 and 47 ka,
157 and the GRIPSS09sea chronology (Shackleton et al., 2004) was applied between 47 and 77 ka.
158 Further details of the age models are given in Daniau et al. (2007) and Sánchez-Goñi et al.
159 (2008).

160 The age model of core MD95-2043 is based on 21 AMS ^{14}C dates and graphical tuning of the
161 MD95-2043 alkenone SST record to GISP2 $\delta^{18}\text{O}$ (Cacho et al., 1999). Due to the recent
162 improvement of the chronology of the SW Iberian margin core, some discrepancies in the two
163 age models appear in the dates of the climatic events at around 40 ka BP. In these cases, we have
164 slightly modified the chronology of the Alboran Sea record for the time interval between 40 and
165 50 ka, according to Sánchez-Goñi et al. (2009), in order to align the latter sedimentary sequence
166 with that of the SW Iberian margin.

167

168 **3.2. Dinoflagellate analysis**

169

170 Dinoflagellates are flagellate protists that occur in both marine and freshwater environments, and
171 which thrive in the depth range of 18-100 m in oceanic domains (Dodge and Harland, 1991;
172 Raine et al., 2002) and 0-10 m at the coastline. Water turbulence is greater in the neritic zone and
173 represents a limiting factor for light penetration and thus the maximum depth of dinoflagellate
174 habitats. Dinoflagellates reproduce primarily through fission, but sexual reproduction also occurs
175 resulting in a resting cyst which is preserved in sediments (Dodge et Harland, 1991; Head, 1996).
176 The distribution of dinoflagellate cysts (dinocysts) reflects physico-chemical parameters of the
177 overlying water masses (temperature, salinity, sea-ice cover, seasonality and nutrient availability)
178 (e.g. Turon, 1984; Mudie, 1992; Matthiessen, 1995; Rochon et al., 1999; Devillers and de Vernal,
179 2000; Zonneveld et al., 2001; Marret and Zonneveld, 2003; de Vernal and Marret, 2007).

180 Dinocyst analysis is therefore an essential tool for reconstructing Quaternary paleoenvironments
181 (e.g. Turon, 1984; Turon and Londeix, 1988; Eynaud et al., 2000, 2004, 2009; de Vernal et al.,
182 1997, 2001, 2005; Mudie et al., 2002, 2004; Grøsfjeld et al., 2006; Penaud et al., 2008, 2009,
183 2010).

184

185 Dinocyst assemblages were characterised at the species level on the sediment fraction smaller
186 than 150 μm on 61 palynological slides for core MD95-2043 (this study) and on 71 slides for
187 core MD95-2042 (Eynaud, 1999; Sánchez-Goñi et al., 2000). The preparation technique followed
188 the protocol described by de Vernal et al. (1999) and Rochon et al. (1999), slightly modified at
189 the EPOC laboratory (Castera and Turon, [http://www.epoc.u-](http://www.epoc.u-bordeaux.fr/index.php?lang=fr&page=eq_paleo26)
190 [bordeaux.fr/index.php?lang=fr&page=eq_paleo26](http://www.epoc.u-bordeaux.fr/index.php?lang=fr&page=eq_paleo26)). Each subsample of 8 cm^3 was weighed, dried
191 overnight and then weighed again to obtain dry weight. Subsamples were then washed through a
192 150 μm sieve and the fraction smaller than 150 μm was used for palynological analysis. After
193 chemical and physical treatments (cold HCl, cold HF and sieving through single-use 10 μm nylon

194 mesh screens), the final residue was mounted between slide and coverslip with glycerine jelly
195 coloured with fuschin. Identifications and counts were performed using a Leica DM 6000
196 microscope at $400\times$ magnification, and counts aimed to reach 300 specimens wherever possible
197 for each sample (cf. Appendix A). Taxonomic identifications are consistent with those of
198 Fensome et al. (1998) and Fensome and Williams (2004). *Brigantedinium* cysts are grouped
199 together and include all spherical brown cysts, since it is rarely possible to identify them at the
200 species level due to their crumbled aspect which masks the archeopyle.

201 Dinocyst assemblages were described by the percentages of each species calculated on the basis
202 of the total dinocyst sum including the few unidentified taxa and excluding pre-Quaternary cysts.
203 Palynomorph concentrations were calculated using the marker grain method (de Vernal et al.,
204 1999). Aliquot volumes of *Lycopodium* spores were added to each sample before chemical
205 treatments in order to obtain palynomorph concentrations.

206

207 **3.3. Quantitative reconstructions of sea-surface parameters**

208

209 **3.3.1. Dinocyst SST and SSS reconstructions**

210

211 We used a transfer function based on the Modern Analogue Technique (MAT) to reconstruct sea-
212 surface hydrological parameters from dinocysts. The MAT principally uses the statistical distance
213 between fossil (paleoceanographic record) and current (modern database) assemblages. The
214 calculation of past hydrological parameters relies on a weighted average of the SST values of the
215 best modern analogues found (minimum and maximum number of analogues imposed in the
216 transfer functions are 5; cf. Appendix B for the list of analogues found for each sample). The

217 maximum weight is given for the closest analogue in terms of statistical distance. The reader is
218 referred to Guiot and de Vernal (2007) for a review of theory of transfer functions and to de
219 Vernal et al. (2001, 2005) for a step by step description of the application of transfer functions to
220 dinocysts, including discussion about the degree of accuracy of the method.

221
222 The dinocyst transfer function used in this work (cf. de Vernal et al., 2005; GEOTOP website:
223 http://www.unites.uqam.ca/geotop/monographie_n940/eng/index.shtml) is derived from a
224 modern database comprising 60 dinocyst species and 940 stations from the North Atlantic, Arctic
225 and North Pacific oceans and their adjacent seas, including the Mediterranean Sea (84 stations
226 including station “M1039”, cf. Fig. 1; Mangin, 2002), as well as epicontinental environments
227 such as the Estuary and Gulf of St. Lawrence, the Bering Sea and the Hudson Bay. The transfer
228 function (n=940) is run under the “3Pbase” software (Guiot and Goeury, 1996). This software
229 was originally developed for pollen-based quantitative climate reconstruction ([http://www.imep-](http://www.imep-cnrs.com/pages/3pbase.htm)
230 [cnrs.com/pages/3pbase.htm](http://www.imep-cnrs.com/pages/3pbase.htm)) and was subsequently applied to dinocyst assemblages (e.g. de
231 Vernal et al., 2001, 2005). An index “Dmin”, provided by the software “3PBase”, allows testing
232 the reliability of the reconstructions (cf. de Vernal et al., 2005). This index describes, for each
233 sample analyzed, the distance between the closest analogue found by the transfer function and the
234 fossil assemblage. A threshold value is calculated from the calibration of the database for the
235 identification of non-similar or very bad analogues. This threshold value provided by the
236 software “3PBase” is 71.72 and, below this value, the similarity between the modern data and the
237 fossil record is considered significant (cf. Appendix B for the Dmin values calculated for each
238 sample). The authors caution that the full reference set has been used for the calculations of
239 hydrological parameters without any regional selection of samples within the modern database. In
240 this study, we present February and August mean sea-surface temperatures (SST, with prediction

241 errors of $\pm 1.2^{\circ}\text{C}$ and $\pm 1.8^{\circ}\text{C}$ respectively), and February and August mean sea-surface salinities
242 (SSS, with prediction errors of ± 1.7 for both).

243

244 **3.3.2. Foraminiferal SST and SSS reconstructions**

245

246 In the same way, we used a transfer function based on the MAT to reconstruct foraminiferal SST
247 from the foraminiferal assemblages of cores MD95-2042 (Cayre et al., 1999) and MD95-2043
248 (Pérez-Folgado et al., 2003). Calculation of past hydrological parameters relies on a weighted
249 average of the SST values of the best 5 modern analogues found systematically. This transfer
250 function has been developed at EPOC laboratory (“Environnements et Paléoenvironnements
251 OCéaniques”, Bordeaux1 University, France). The MAT (e.g. Kucera, 2007) is run under the “R”
252 software using a script first developed for dinocyst transfer functions
253 (http://www.cerege.fr/IMG/pdf/ECCOR_StatRAvr08.pdf). The modern database relies on a
254 modern database of 1007 modern assemblages and is derived from the ones developed separately
255 for the North Atlantic and the Mediterranean seas during the MARGO project (Kucera et al.,
256 2005; Hayes et al., 2005). These databases were merged together to offer a larger set of analogues
257 for subtropical reconstructions over the last glacial period notably (Eynaud et al., 2009;
258 Matsuzaki et al., in press). Modern hydrological parameters were requested from the WOA 2008
259 database using the tool developed during the MARGO project ([http://www.geo.uni-](http://www.geo.uni-bremen.de/geomod/staff/csn/woasample.html)
260 [bremen.de/geomod/staff/csn/woasample.html](http://www.geo.uni-bremen.de/geomod/staff/csn/woasample.html)). This method allows the reconstruction of annual
261 and seasonal (winter, spring, summer and fall) SST. In this paper, we present winter and summer
262 mean SST with a prediction error of $\pm 1.2^{\circ}\text{C}$ and $\pm 1.3^{\circ}\text{C}$ respectively, and annual SST with a
263 prediction error of $\pm 1.1^{\circ}\text{C}$ (Eynaud et al, 2009; Matsuzaki et al., in press).

264

265 SSS estimates have been derived using the approach summarised in Malaizé and Caley (2009) for
266 calibration of the salinity-water isotope relationship. Craig and Gordon (1965) established the
267 first salinity-water isotope relationship as follows:

$$268 \quad \delta^{18}\text{O}_{\text{sw}} = 0.66 \text{ SSS} - 23.5$$

269 Since this pioneer work, many measurements have been performed and many different
270 relationships, with different slopes, have been derived for different oceans (Ostlund et al., 1987,
271 Schmidt, 1999; LeGrande and Schmidt, 2006). Discrepancies in slope and in intercept values are
272 due to local characteristics, on spatial and temporal scales. For the Atlantic Ocean, an important
273 dataset, established from several decades of oceanographic measurements, allows revision of the
274 Craig and Gordon (1965) calibration and the determination of a new relationship, with a mean
275 slope component of 0.558.

276 In order to estimate past changes in oceanic $\delta^{18}\text{O}_{\text{sw}}$, Epstein et al. (1953) established a
277 paleotemperature equation which links temperature with the isotopic composition of calcite
278 ($\delta^{18}\text{O}_{\text{c}}$) in calcareous shells (e.g. foraminifera) and the ambient waters ($\delta^{18}\text{O}_{\text{sw}}$). Shackleton and
279 Opdyke (1973) have adapted the equation of Epstein et al. (1953) as follows:

$$280 \quad T = 16.9 - 4.38 (\delta^{18}\text{O}_{\text{c}} - \delta^{18}\text{O}_{\text{sw}}) + 0.13 (\delta^{18}\text{O}_{\text{c}} - \delta^{18}\text{O}_{\text{sw}})^2$$

281 Stable isotope $\delta^{18}\text{O}_{\text{c}}$ measurements in cores MD95-2042 (Cayre et al., 1999; Shackleton et al.,
282 2000) and MD95-2043 (Cacho et al., 1999) were carried out on *G. bulloides* monospecific
283 samples. Following the pioneer study of Duplessy et al. (1991), we corrected the summer
284 temperature by 1°C for *G. bulloides* species. By solving the Shackleton and Opdyke (1973)
285 equation, we obtain $\delta^{18}\text{O}_{\text{sw}}$ variations, which integrate the signal of both local and global
286 variations. To remove the global influence of continental ice volume, we used estimations of past
287 global $\delta^{18}\text{O}_{\text{sw}}$ changes based on benthic isotopic records and coral terrace growth (Waelbroeck et

288 al., 2002), and used the modern $\delta^{18}\text{O}_{\text{sw}}$ values (1.2 ‰ for the Alboran Sea and 0.97 ‰ for the
289 Iberian margin) extracted from the world dataset atlas compiled by Schmidt (1999) and Bigg and
290 Rohling (2000) (C. Pierre dataset for the Alboran Sea, and Craig and Gordon dataset for the
291 Iberian margin, from <http://data.giss.nasa.gov/o18data>). The residual $\delta^{18}\text{O}_{\text{sw}}$ “ice-corrected”
292 signal is then converted into quantitative SSS values for both cores, using a mean slope of 0.558.
293 SSS uncertainties of around 1 psu derive from the prediction error of the foraminiferal SST
294 reconstructions (1.8°C uncertainty linked with August SST reconstructions corresponds to a 0.45
295 ‰ uncertainty on a δ scale), as well as on uncertainties in the global $\delta^{18}\text{O}_{\text{sw}}$ changes linked to
296 sea level changes (0.15 ‰).

297 **4. Dinocyst assemblages through time**

298

299 Concerning the species of the genus *Brigantedinium* spp., round brown cysts formed by
300 heterotrophic dinoflagellates, it is important to take into account the fact that high occurrences of
301 *Brigantedinium* spp. can be linked to better preservation under hypoxic or anoxic bottom
302 conditions (Combourieu-Nebout et al., 1998; Zonneveld et al., 2001; Kodrans-Nsiah et al., 2008).
303 We can not exclude that *Brigantedinium* spp. may have been subjected to dissolution since deep
304 ventilation in the western Mediterranean Sea (and thus oxidation processes) during GS was
305 stronger than during GI of the last glacial (cf. benthic $\delta^{13}\text{C}$ in Fig. 4; Cacho et al., 2006).
306 However, *Brigantedinium* spp. are present throughout the Alboran core without any clear trend
307 between few percentages during GS and high percentages during GI. We can thus expect to
308 reconstruct dinocyst assemblages without significant preservational problems.

309 When comparing both cores (Fig. 2), we note extremely high relative abundances of
310 *Brigantedinium* spp. (30-89%) on the SW Iberian margin (Fig. 2b). This may reflect the high
311 productivity induced by permanent or intensified upwelling cells along the Portuguese margin
312 during the last glacial (e.g. Abrantes, 2000; Lebreiro et al., 1997); this zone today is characterised
313 by a seasonal upwelling system (Peliz et al., 2005) and *Brigantedinium* spp. percentages only
314 reach a maximum of 2-5% in the modern database. Percentages of these cysts fluctuate more in
315 the Alboran Sea, with pronounced oscillations (average of 19% and peaks reaching 30 to 45%)
316 (Fig. 2a).

317

318 **4.1. Dinocysts occurring during Greenland Stadials**

319

320 Dinocysts which feature prominently during GS include *Bitectatodinium tepikiense*, *Spiniferites*
321 *lazus*, *Spiniferites elongatus*, *Impagidinium aculeatum*, and *Operculodinium centrocarpum* in the
322 Alboran Sea (Fig. 2a), and *B. tepikiense*, *S. elongatus*, cysts of *Pentapharsodinium dalei*,
323 *Lingulodinium machaerophorum* and *Nematosphaeropsis labyrinthus* on the SW Iberian margin
324 (Fig. 2b). It is also interesting to note that maximal dinocyst concentrations occur during GS off
325 Portugal (most notably during HS 4; Fig. 2b) contrarily to what we observe in the Alboran record
326 (Fig. 2a).

327
328 In the Alboran Sea, *B. tepikiense* increases during HS events, especially HS 3 and HS 4 (Fig. 2a),
329 whereas it occurs during each GS on the SW Iberian margin, where it consistently makes up
330 almost 10% of the total dinocyst assemblage (Fig. 2b). Today, *B. tepikiense* is mainly distributed
331 between 40°N and 60°N in temperate to sub-arctic environments of the North Atlantic, with the
332 highest abundances found south of the Gulf of St. Lawrence in coastal environments of Nova
333 Scotia and the Gulf of Maine (Wall et al., 1977; Mudie, 1992). This species is characteristic of
334 areas marked by strong seasonal contrasts, with freezing winter SSTs and up to 16°C summer
335 SST (Rochon et al., 1999; de Vernal et al., 2005), and enhanced surface water stratification
336 (Rochon et al., 1999; Marret and Zonneveld, 2003). Previous results from the North Atlantic
337 (Zaragosi et al., 2001; Penaud et al., 2009), the SW Iberian margin (Sánchez-Goñi et al., 2000;
338 Turon et al., 2003), and the western Mediterranean Sea (Turon and Londeix, 1988; Combourieu-
339 Nebout et al., 2002) have shown increased abundances of *B. tepikiense* during HS.

340
341 *S. elongatus* develops during each GS with low percentages reaching 2-5% in the Alboran Sea
342 (Fig. 2a) and 1-2% on the western Iberian margin (Fig. 2b). Maximum present-day occurrences
343 of this species are observed in the Baffin Bay and Barents Sea, and this taxon is generally related

344 to cool to temperate conditions (Rochon et al., 1999). The significant occurrence of *B. tepikiense*
345 and *S. elongatus* is attributed to strong seasonality characterised by winter sea-surface
346 temperatures probably less than 5°C (Marret et al., 2004).

347

348 Cysts of *P. dalei* occur sporadically in the Alboran Sea core (Fig. 2a) but mark clearly the cold
349 events in core MD95-2042, most notably HS 2 and HS 4 (Fig. 2b). Cysts of *P. dalei* are well
350 represented in modern sediments from polar to subpolar environments that experience summer
351 sea-surface temperatures higher than 4°C (Rochon et al., 1999; Matthiessen, 1995; Marret et al.,
352 2004; de Vernal et al., 2005). They are particularly prevalent as part of the spring bloom within
353 North Atlantic fjord systems (Dale, 1977; Harland et al., 2004a,b).

354

355 In the Alboran Sea, we show that *S. lazus* is associated with cold HS and especially HS 5 (Fig.
356 2a). Today, the distribution of *S. lazus* is restricted to coastal regions of western Europe, always
357 with low abundances (less than 2% of the dinocyst assemblages) (Reid, 1974). This species can
358 be regarded as a neritic temperate species of regions characterised by oligotrophic to eutrophic
359 surface water conditions (Reid, 1974; Harland, 1983).

360

361 Finally, the most striking feature visible in the dinocyst distribution during the D-O cycles is the
362 systematic occurrence of *Impagidinium aculeatum* during GS in the Alboran Sea, with low
363 percentages reaching a maximum of 5% (Fig. 2a). On the western Iberian margin, this species
364 peaks during GI, although a local maximum is also observed during HS 5 (Fig. 2b). At present,
365 this taxon is associated with warm water dinocyst assemblages; high relative abundances are
366 found in tropical/subtropical oligotrophic open oceanic sites (Turon, 1984). In SW Iberian and
367 western Mediterranean paleoclimate records, this species is characteristic of the Holocene with

368 percentages close to 20% (Combourieu-Nebout et al., 1998; Turon et al., 2003; Rouis-Zargouni
369 et al., 2010). However, percentages close to 5% were also previously observed during cold
370 stadials (HS 1 and HS 2) in a core from the Sicilian-Tunisian Strait (Rouis-Zargouni et al., 2010).

371

372 **4.2. Dinocysts occurring during Greenland Interstadials**

373

374 Dinocysts which feature prominently during GI in the Alboran Sea include *Spiniferites mirabilis*,
375 *Impagidinium patulum*, *L. machaerophorum*, *N. labyrinthus* as well as some heterotrophic
376 species (Protoperidinioids, *Selenopemphix nephroides* and *Selenopemphix quanta*, but excluding
377 *Brigantedinium* spp. whose ecology is rather complex) (Fig. 2a). On the SW Iberian margin they
378 include *Impagidinium* species (*I. aculeatum* and *I. patulum*), *S. mirabilis*, *O. centrocarpum* and *S.*
379 *nephroides* (Fig. 2b). Unlike the SW Iberian margin (Fig. 2b), maximal dinocyst concentrations
380 occur during GI in the Alboran Sea (especially during GI 8 and GI 12; Fig. 2a).

381

382 In the Alboran Sea, GI are characterised by the species *S. mirabilis* (Fig. 2a) with percentages on
383 average four times higher than on the SW Iberian margin (Fig. 2b). The highest relative
384 abundances of *S. mirabilis* are recorded during GI 7 and GI 8. Today, *S. mirabilis* is mainly
385 distributed between 35°N and 50°N in warm temperate to temperate environments of the North
386 Atlantic with highest occurrences found off the coast of Portugal and in the Bay of Biscay
387 (Rochon et al., 1999). This species extends as far south as 10°N and is generally absent from
388 areas with summer SST below 12°C and salinity below 28.5, and thrives optimally when winter
389 SSTs are between 10°C and 15°C and summer SSTs are above 15°C.

390

391 On the SW Iberian margin, GI are characterised by *I. patulum* (Fig. 2b) with abundances four

392 times higher than in the Alboran Sea (Fig. 2a). Most *Impagidinium* species occur today with
393 maximum frequencies in tropical to warm temperate waters between 20°N and 35°N and are
394 representative of full-oceanic conditions (Harland, 1983; Turon, 1984; Bouimetarhan et al.,
395 2009).

396
397 *Spiniferites delicatus* is common on the SW Iberian margin and is mainly linked with interstadial
398 conditions (Fig. 2b), while it is rare in the Alboran Sea (Fig. 2a). *S. delicatus* is a temperate
399 species adapted to warm and neritic conditions (Wall et al., 1977; Harland, 1983; Marret, 1994).

400
401 *S. quanta*, *S. nephroides* and some Protoperidinioids show a distribution pattern closely linked to
402 GI in the Alboran Sea (Fig. 2a). This association also holds for *S. nephroides* on the SW Iberian
403 margin, but is less obvious for *S. quanta* and Protoperidinioids as they also increase during GS
404 (Fig. 2b). These latter taxa, represented by low relative abundances in the modern database and a
405 scattered distribution in the North Atlantic (Rochon et al., 1999), are derived from heterotrophic
406 dinoflagellates and are mainly related to high food resources. Their heterotrophic strategy of
407 nutrition probably links them to the presence of higher concentrations of nutrients in surface
408 waters. *S. quanta* has previously been linked to the dynamics of upwelling cells off NW Africa
409 (Dodge and Harland, 1991; Penaud et al., 2010).

410

411 **4.3. Opposite dinocyst patterns between Iberian and Alboran environments over D-O cycles**

412

413 Other dinocyst species not described above, *L. machaerophorum*, *N. labyrinthus*, and *O.*
414 *centrocarpum*, are an important component of the dinocyst assemblages and show opposite
415 patterns in both cores

416
417 *N. labyrinthus* shows very high percentages in the Alboran Sea (average 24%), where it occurs
418 generally within GI (Fig. 2a). Peaks of this species systematically exceed 40% and can reach 60
419 to 70% of the total dinocyst assemblage. On the SW Iberian margin, this taxon is less prevalent
420 (average of 4.5%) with peaks generally reaching 10% during GS (except during HS) (Fig. 2b). At
421 present, *N. labyrinthus* is a typical open-ocean species found predominantly between 45°N and
422 65°N in the North Atlantic Ocean (Rochon et al., 1999; Matthiessen, 1995; Marret et al., 2004; de
423 Vernal et al., 2005). Maximum abundances of this species are recorded off southern Greenland
424 where cold waters from the East Greenland and Labrador currents mix with warm North Atlantic
425 waters of the Irminger Current (Rochon et al., 1999; Marret et al., 2004). This species, in
426 association with *B. tepikiense*, was previously related to polar water incursions during MIS 5 cold
427 substages off Portugal (Sánchez-Goñi et al., 1999; Eynaud et al., 2000), and during the Younger
428 Dryas off Portugal (Turon et al., 2003), in the Mediterranean (Turon and Londeix, 1988; Rouis-
429 Zargouni et al., 2010) and off NW Morocco (Marret and Turon, 1994; Penaud et al., 2010). *N.*
430 *labyrinthus* has also been positively correlated with nutrient-rich and cool waters (Turon and
431 Londeix, 1988; Devillers and de Vernal, 2000).

432
433 *L. machaerophorum* occurs during GI in the Alboran Sea, particularly GI 8, 9 and 12 (Fig. 2a),
434 while it occurs during GS on the SW Iberian margin (Fig. 2b). *L. machaerophorum* is today
435 considered to be a temperate to tropical coastal euryhaline species (Mertens et al., 2009). It
436 dominates associations from the northern African and southern European Atlantic shelves, and it
437 is also found in abundance in North African coastal upwelling regions (Targarona et al., 1999;
438 Sprangers et al., 2004) and near the Congo outlet (Marret, 1994). Extremely high concentrations
439 of these cysts have been found in areas with typical seasonally stratified water columns such as

440 fjords, bays, and estuaries (e.g. Reid, 1972; Bradford and Wall, 1984; Dale, 1985; Lewis, 1988;
441 Morzadec-Kerfourn, 1988; Dale et al., 1999). This taxon has often been related to warm and
442 stratified surface waters (Marret and Zonneveld, 2003), and has also been used as a proxy for
443 fluvial inputs towards the ocean (Zaragosi et al., 2001; Holzwarth et al., 2010).

444
445 *O. centrocarpum* occurs frequently in the Alboran Sea record with percentages reaching 20 to
446 40% during almost every GS (Fig. 2a). On the SW Iberian margin, this species shows relatively
447 low percentages, with peaks generally less than 10% during GI and reaching a maximum of 20%
448 during GI 3 (Fig. 2b). This species is considered to be a cosmopolitan, cool to temperate taxon
449 (Turon, 1984; Rochon et al., 1999; Marret and Zonneveld, 2003), and an abundance pattern
450 following the route of the North Atlantic Drift (NAD) has been identified from its present
451 geographical distribution in North Atlantic surface sediments (Turon, 1984; Rochon et al., 1999).
452 This observation was previously used to interpret the presence of an active NAD at times when
453 this species was abundant in the Quaternary North Atlantic sediments (Zaragosi et al., 2001;
454 Eynaud et al., 2004; Penaud et al., 2008, 2009).

455

456 **5. Temperature and salinity records: convergences and discrepancies**

457

458 **5.1. SST: dinocysts versus planktonic foraminifera**

459

460 Quantitative reconstructions derived from the two transfer functions (dinocyst and foraminifera)
461 are not directly comparable since the reconstructed parameters are not exactly the same. The
462 dinocyst transfer function 3PBase-940 only provides February and August SST reconstructions
463 while the foraminiferal transfer function R-1007 only provides annual and seasonal-mean (i.e.

464 winter, spring, summer and fall) SST reconstructions. However, February and August
465 foraminiferal SST can be estimated using the MAT transfer function derived from Pflaumann et
466 al. (1996) which relies on a modern database of 692 modern assemblages (only Atlantic stations),
467 improved during the MARGO project (Kucera et al., 2005). It is, however, more appropriate for
468 us to discuss the foraminiferal data from the R-1007 transfer function (Atlantic and
469 Mediterranean modern databases) for the Alboran Sea core which is located at the boundary of
470 the two basins. The comparison of winter *versus* February SST values on the 664 common
471 stations from the foraminiferal north Atlantic databases (n=1007 *versus* n=692) generates a mean
472 difference of 0.26°C (with a maximum difference of 1.95°C). For the summer *versus* August SST
473 values, the mean difference is 0.37°C (with a maximum difference of 1.1°C). These values are
474 within the range of the error bars of the foraminiferal reconstructions and therefore encourage us
475 to consider that monthly and seasonal foraminiferal SST values can be discussed interchangeably.
476 Concerning dinocysts, the n=940 database also includes Mediterranean stations. It is worth noting
477 that, among the 84 Mediterranean analogues (including 17 stations in the Alboran Sea), the only
478 Mediterranean analogue found by the transfer function is located in the Alboran Sea and named
479 “M1039” (Fig. 1; Appendix B). This is mainly the case during GI conditions (GIs 3, 5, 7, 8, 9,
480 11, 12 and 13; cf. Appendix B), suggesting that GI conditions in the Alboran Sea between 25 and
481 50 ka were roughly equivalent to conditions prevailing at the “M1039” site today (Fig. 1).

482
483 When comparing SST reconstructions from both micropaleontological proxies on the two cores,
484 one can note that February *versus* winter SST values obtained with dinocysts (3PBase-940) and
485 foraminifera (R-1007), respectively, are closer than August *versus* summer ones (Fig. 3).
486 However, the general good consistency of February *versus* winter SST is not valid for the SW
487 Iberian margin during HS 5 with about 8°C difference between both micropaleontological

488 reconstructions (dinocyst SST estimates around 13°C and foraminiferal ones close to 5°C; Fig.
489 3). This offset is unrealistically large given the prediction error for each method (around 1.2°C).
490 This minor SST drop expressed in dinocyst populations may be underestimated, as also revealed
491 by the annual alkenone-derived SST values warmer than the February dinocyst-based SST ones
492 (Fig. 3). One reason may come from the occurrence of the dinocyst species *Impagidinium*
493 *aculeatum* during HS 5 (Fig. 2). Indeed, on the western Iberian margin, this species peaks during
494 GI, although a local maximum is observed during HS 5 (percentages slightly higher than 5%).
495 However, this species is associated with warm water dinocyst assemblages and is found today in
496 tropical/subtropical oligotrophic open oceanic sites (Turon, 1984). Although the occurrence of
497 this taxon during a cold HS is not understood, it may switch the transfer function towards warmer
498 SST values.

499
500 August *versus* summer SST reconstructions for both cores show similar overall trends but differ
501 more in terms of amplitudes, with dinocyst SSTs being 5 to 10°C higher than foraminiferal SST
502 estimates (Fig. 3). Two hypotheses can be put forward with respect to this observation. Either the
503 August/summer (dinocyst or/and foraminiferal) quantifications are less reliable or the
504 discrepancy reflects an ecological bias. Dinocysts are indeed produced by dinoflagellates that
505 thrive in the photic zone while foraminifera can migrate deeply in the water column with living
506 depths ranging from 0 to 1000 m. Dinoflagellates, being found in shallower water, would thus
507 record warmer SST consistently with a higher stratification during the warmest month, i.e.
508 August. Various biases between different micropaleontological reconstructions can thus occur in
509 relation to the ecological strategy (depth of habitat and growth seasons) of the different
510 planktonic populations (de Vernal et al., 2005). The following comparison with alkenones will
511 help us to decipher whether the foraminiferal or dinocyst signal is too cold or warm, respectively.

512 Finally, concerning the Alboran Sea core, one can note that dinocysts and foraminiferal show
513 closer August *versus* summer SST values between 31 and 38 ka and larger offsets between 38
514 and 50 ka. This point will be further discussed in section 6.2.3. of this manuscript.

515

516 **5.2. SST: transfer functions *versus* alkenones**

517

518 Alkenone-derived SST from cores MD95-2043 (Cacho et al., 1999) and MD95-2042 (Pailler and
519 Bard, 2002) has previously been compared with ice-core records, demonstrating a close linking
520 between SSTs in the western Mediterranean and temperature developments over the wider North
521 Atlantic region, including Greenland. Here, it appears that our dinocyst-based SST
522 reconstructions for both cores match peak to peak with the SSTs derived from alkenones,
523 showing minimum values during GS (Fig. 3). Furthermore, the alkenone-based SST, which
524 records an annual signal of temperature, fluctuates systematically in between the temperature
525 range given by seasonal dinocyst SST reconstructions and is closer to February than August SST
526 (Fig. 3). A co-variation between February dinocyst SST and alkenone-derived SST
527 reconstructions was previously observed off NW Morocco during the last glacial over the last 30
528 ka (Penaud et al., 2010). This would confirm the hypothesis that switches in mean annual
529 temperatures were dominated by, and thus weighted towards, the winter season during the last
530 glacial cycle as suggested by Denton et al. (2005).

531

532 The foraminiferal transfer function $n=1007$ also provides annual SST reconstructions that can
533 directly be compared with the alkenone signal. One can note that annual SST values
534 reconstructed with both proxies are closer during GI than during GS (Fig. 3). Foraminifera
535 routinely provide much colder temperatures during GS, and especially during HS that are

536 characterised by significant SST anomalies between 4 and 7°C (Fig. 3). This shift of annual SST
537 values towards cold SSTs is mainly due to extremely cold summer SSTs reconstructed with
538 foraminifera during GS (Fig. 3). This contrast has parallels to the previous observation on the
539 considerable offsets between dinocyst and foraminiferal August *versus* summer SST. Since
540 alkenones are synthesised by coccolithophorids which are single-celled algae, protists and
541 phytoplankton such as dinoflagellates, we can assume that they both give a signature of sea-
542 surface hydrological changes occurring in the photic zone, while foraminifera may yield a bias
543 towards colder SSTs, especially during GS. However, it will be crucial to understand if this
544 finding reflects a real ecological strategy of foraminifera (different depths of habitat following
545 different seasons) or a problem associated with the transfer function.

546

547 **5.3. SSS: dinocysts *versus* planktonic foraminifera**

548

549 Quantifications of salinity are of critical interest as they are fairly rare in paleoclimate studies. A
550 major challenge in paleoceanography is to increase the availability of SSS reconstructions for
551 comparison with climate models (e.g. MARGO project members, 2009). In our records, similar
552 changes, both in amplitude and timing, are revealed through the comparison of February
553 dinocyst-based SSS *versus* winter foraminiferal derived ones (Fig. 3). Numerical results are
554 always very close or at least within the prediction error of 1-1.8 psu, except during HS 5 and GI
555 12 at the SW Iberian margin which are marked by differences of 2 to 3 psu between both
556 micropaleontological proxies (Fig. 3). Comparison of the February/winter SSS data is
557 particularly striking because the methods of quantification are different. One method requires
558 MAT calculation of SST based on planktonic foraminifera and $\delta^{18}\text{O}$ analysis (e.g. Malaizé and

559 Caley, 2009), while the other one is obtained on the basis of MAT transfer function applied to
560 dinocyst assemblages (e.g. de Vernal et al., 2005). However, correlations between fluctuations in
561 August dinocyst-based SSS and summer foraminiferal derived ones are less evident with
562 fluctuations similar in timing but divergent in terms of amplitudes, especially for the Alboran Sea
563 (Fig. 3).

564
565 For the SW Iberian margin, our study shows that major low salinity events are recorded during
566 HS (Fig. 3b). HS 4 displays the maximum drop in February/winter SSS with values reaching 31
567 psu, corresponding to the largest fluxes of freshwater to the ocean over our study period. Our data
568 for HS 3 (around 31.5-32.5 psu, 4 psu lower than the modern value) also show a major low
569 salinity event and are consistent with previous results acquired further north along the Celtic
570 margin over the last 35 ka BP where this interval was also characterised by a 4 psu depletion in
571 SSS (Eynaoud et al., submitted).

572 For the Alboran Sea, the difference in the resolution of analysis between dinocyst and
573 foraminiferal/isotope reconstructions makes it difficult to compare some sections (Fig. 3a). This
574 is mainly due to the fact that foraminiferal SSS is calculated from both foraminiferal SST and
575 planktonic $\delta^{18}\text{O}$ data and the resolution of the $\delta^{18}\text{O}$ record is lower than the foraminiferal SST
576 record. For the whole Alboran Sea record, February/winter SSS shows changes of similar order
577 of magnitude between each GS, with the strongest signal of freshening recorded during HS 4
578 reaching 33 psu (4 psu lower than the modern value).

579

580 **6. Paleohydrological changes affecting subtropical Atlantic latitudes through D-O cycles**

581

582 **6.1. The new contribution of dinocyst assemblages in the Alboran Sea**

583

584 **6.1.1. Warm intervals (GI) in the Alboran Sea**

585

586 During GI, we observe an expansion of warm temperate to tropical species with a trend of
587 decreasing abundance, and decreasing February and alkenone SST, from immediately prior to a
588 Heinrich Stadial to the next HS (Fig. 4). This pattern is similar to that seen in the Greenland ice
589 core isotope records showing progressively shorter GI and smaller increases in Greenland air
590 temperatures between HS 5 and HS 4 and between HS 4 and HS 3 (Fig. 4, NGRIP GICC05).
591 This demonstrates a similar trend between Alboran SST and atmospheric temperatures over
592 Greenland, involving a rapid transmission of Northern Atlantic climate changes into the western
593 Mediterranean region.

594 When comparing the amplitude of warm taxa development during individual D-O warming (Fig.
595 4), we note less frequent occurrences of thermophilous taxa during the earlier interstadials (GI
596 12-9) than during the subsequent interstadials (GI 8-5). This is consistent with pollen analysis
597 conducted on the same core (Fletcher and Sánchez-Goñi, 2008) demonstrating high values for
598 Mediterranean forest during GI 8, 7, 6 and 5 reflecting the maximum in subtropical summer
599 insolation associated with the precession minimum centred around 30-35 ka (Fig. 4). Such a
600 development is also seen in paleo-vegetation records at nearby Alboran site ODP 976
601 (Combourieu-Nebout et al., 2002) and on the SW Iberian margin (MD95-2042: Sánchez-Goñi et
602 al., 2000).

603 The interstadials GI 8 and GI 12, immediately succeeding HS 4 and HS 5, respectively, exhibit
604 particularly long and warm periods. They are both marked by high relative abundances of warm
605 water dinocysts, high SST estimates (especially clear with alkenones; Fig. 4), and by the highest

606 paleoproductivity conditions indicated by heterotrophic taxa in the early interstadials and high
607 dinocyst concentrations in the second half of GI 8 and GI 12 (Fig. 4). In terms of sea-surface
608 paleohydrology, GI 8 and GI 12 are thus very similar with a comparable bipartite structure (Fig.
609 4). Palynological investigations of GI 8 and GI 12 in the same Alboran Sea core showed that
610 these periods were characterised by the strongest expansions of mixed oak forest between 48 and
611 15 ka (Fletcher and Sánchez-Goñi, 2008). Furthermore, Fletcher and Sánchez-Goñi (2008)
612 identified an Atlantic oceanic character during GI 12, in contrast to a markedly Mediterranean
613 character during GI 8, linked to the precession minimum that likely enhanced the Mediterranean
614 climate and caused an enhanced seasonal contrast between dry summers and wet winters
615 (Sánchez-Goñi et al., 2008, 2009; Fletcher and Sánchez-Goñi, 2008). Our dinocyst data also
616 reveal a strong expansion of the species *L. machaerophorum* in the second half of GI 8,
617 representing between 20 and 40% of the total dinocyst assemblage (Fig. 4). This species has
618 sometimes been used to trace fluvial inputs (Zaragosi et al., 2001; Holzwarth et al., 2010) and
619 could reflect higher river run-off to the Alboran Sea, providing further evidence for increased
620 winter precipitations during this interval.

621 Throughout our record, heterotrophic dinocyst species (*S. nephroides*, *S. quanta*, and
622 Protoperidinioids cysts) occur during each GI in the Alboran Sea (Fig. 4). Their occurrences
623 suggest sea-surface conditions characterised by increased productivity (Rochon et al., 1999), as
624 also indicated by increased total dinocyst concentrations (Fig. 4). This general pattern of higher
625 productivity during GI relative to GS has previously been discussed based on geochemical
626 evidence (calcium carbonate, barium excess, and total organic carbon) from the same core
627 (Moreno et al., 2004). Today, the two semi-permanent anticyclonic gyres found in the Alboran
628 Sea represent energetic mesoscale features and the main forcing maintaining these gyres is the
629 Atlantic jet which enters through the Strait of Gibraltar (Bormans and Garret, 1989; Benzohra

630 and Millot, 1995; Garcia-Lafuente et al., 1998; Macias et al., 2008). The intensity of the jet is
631 typically modulated by atmospheric pressure variations over the western Mediterranean. Indeed,
632 when atmospheric pressures are lower than average, configuration close to a NAO negative
633 mode, westerlies prevail above the Mediterranean, the Atlantic jet flows northward near the
634 Spanish coast and the western Alboran gyre is well developed in the entire western Alboran Sea.
635 The opposite (NAO positive mode) occurs when easterly winds prevail, the inflow of Atlantic
636 waters is lower and the Atlantic jet is directed southward, reducing the western Alboran gyre
637 extent (Candela et al., 1989; Garcia-Lafuente et al., 2002; Macias et al., 2008). At an annual
638 scale, on the basis of satellite imagery analysis (e.g. Garcia-Gorriz and Carr, 1999; Baldacci et
639 al., 2001; Macias et al., 2007, 2008) maximum surface chlorophyll concentrations were usually
640 found in winter and minimum values were observed in summer (July and August). Biological
641 patterns are thus also tightly coupled to atmospheric pressure above the Mediterranean Basin with
642 westerlies being shifted southward during winter. At the millennial-scale resolution of our study
643 and in agreement with Moreno et al. (2004), our data suggest that GI conditions would exhibit a
644 prolonged southward shift of the westerly wind belt, inducing a more intense Atlantic surface jet
645 that favoured gyre-induced upwelling in the Alboran Sea.

646

647 **6.1.2. Cold intervals (GS) recorded in the Alboran Sea**

648

649 The association *B. tepikiense* - *S. elongatus* represents an indicator for the incursion of subpolar
650 water masses at the Alboran site that is highly evident during HS (Fig. 4). This latter pattern is
651 similar to that of the subpolar foraminifera *Neogloboquadrina pachyderma* s. recorded in the
652 same core (Cacho et al., 1999) that shows higher percentages during HS (Fig. 4). This
653 foraminifer displays the highest percentages during HS 4, then HS 3 and finally HS 5, while *B.*

654 *tepikiense* shows higher percentages during HS 3 compared to HS 4 (cf. Figs. 2 and 4). *B.*
655 *tepikiense* is absent from the Mediterranean Sea today, and highest abundances of this species
656 occur in areas characterised by high-amplitude (10°C) seasonal temperature shifts (Rochon et al.,
657 1999). Therefore, its occurrence during HS in Alboran Sea surface waters implies enhanced
658 seasonal temperature contrast (15°C; cf. Fig. 5) compared to the present-day one (*i.e.* 10°C;
659 23.5°C in summer and 14.5°C in winter; cf. Fig. 5), caused by a strong decrease of winter SST
660 (Fig. 4). Our February dinocyst SST reconstructions reinforce this hypothesis by showing low
661 SST with values around 5.5°C and 6.5°C during HS 4 and HS 3, respectively, *i.e.* 9°C less than at
662 present (Fig. 4). It is thus important to note that, in contrast to *N. pachyderma* s., *B. tepiense*
663 does not mark systematically colder intervals but intervals with larger seasonal contrasts. Our
664 results are also in agreement with other observations made at ODP Site 976 (Alboran Sea) that
665 reflect coeval increases of *B. tepiense* with *N. pachyderma* s. percentages during HS, over the
666 last 50 ka BP (Turon and Londeix, 1988; Combourieu-Nebout et al., 2002). Our data thus
667 confirm that sea-surface cooling in the Alboran Sea was mainly linked to the advection of cold
668 Atlantic water to the western Mediterranean (Cacho et al., 1999). Such cold-water advection
669 occurred synchronously with regional cooling related to atmospheric conditions over the western
670 Mediterranean (Combourieu-Nebout et al., 2002; Sánchez-Goñi et al., 2002). Indeed, on the
671 adjacent continent, the Mediterranean forest (*i.e.* temperate taxa) collapsed (Fig. 4) and estimated
672 winter atmospheric conditions indicate a 10°C lowering and a decrease of 400 mm in
673 precipitation (Sánchez-Goñi et al., 2002).

674 The HS configuration contrasts with the other GS. In the Alboran Sea, before or during each GS
675 (especially those not associated with HS conditions) we note pronounced peaks of *O.*
676 *centrocarpum* (Fig. 4). The occurrence of *O. centrocarpum* in the Alboran Sea, whose present-
677 day distribution directly mirrors the flow path of the NAD, may result from the influx of cool

678 North Atlantic waters entering into the Mediterranean; temperatures of these waters, while
679 presumably low, remained above those of the subpolar waters that entered the Mediterranean
680 during HS. GS events of the last glacial have been demonstrated to coincide with intensification
681 of the deep circulation in the Mediterranean (Cacho et al., 2000, 2006; Sierro et al., 2005; Frigola
682 et al., 2008). Our data suggest that the intensification of deep ventilation in the Alboran Sea,
683 reflected in benthic $\delta^{13}\text{C}$ data (Fig. 4; Cacho et al., 2006), was synchronous with the advection of
684 North Atlantic waters to the Mediterranean, as reflected by highest relative abundances of cool-
685 water dinocyst taxa (Fig. 4).

686 We demonstrate two different patterns during GS: those associated with HS are marked by
687 increased abundances of cold water species associated with subpolar waters, and those not
688 associated with HS are marked by an expansion of cool North Atlantic species. Interestingly, it
689 has been demonstrated that the densest Western Mediterranean Deep Water was formed during
690 GS not associated with HS (Cacho et al., 2006; Frigola et al., 2008). More precisely, it has been
691 suggested that a strong mode of overturning prevailed during GS not associated with HS, an
692 intermediate mode of overturning during HS and a weak mode of overturning during GI (Sierro
693 et al., 2005; Frigola et al., 2008). Strong overturning was expected during HS since strong and
694 cold northern continental winds prevailed over the Mediterranean resulting in dry-cold conditions
695 on land (Fig. 4; Combourieu-Nebout et al., 2002; Sánchez-Goñi et al., 2002). However, the HS
696 intermediate mode has been linked with a strong influence of subpolar waters that lowered sea-
697 surface salinity thus reducing deep water formation and favouring water column stratification
698 (Sierro et al., 2005). The observation of subpolar species (*N. pachyderma* s. and *B. tepikiense*)
699 during HS (especially HS 3 and HS 4), as well as the low winter and summer salinities
700 reconstructed from dinocysts, lowered by around 0.5 to 1.5 psu compared to the other GS in both

701 seasonal configurations (Fig. 3), within low precipitation phases, reinforce the idea of sustained
702 cold conditions with subpolar water masses advection towards the Mediterranean Sea and
703 decreased deep water formation.

704

705 **6.2. Alboran Sea *versus* SW Iberian margin between 50 and 25 ka**

706

707 **6.2.1. SST reconstructions and paleoenvironmental signatures**

708

709 In both cores, lowest SST estimates are observed during HS 4 which represents the most
710 pronounced cold event between 25 and 50 ka, with February/winter SST of about 5.5°C and 4°C
711 recorded in the Alboran Sea and the SW Iberian margin, respectively (Fig. 3). Previously, the
712 coldest deep water temperatures in the Alboran Sea (Cacho et al., 2006), maximum
713 concentrations of IRD off the Portuguese margin (Thouveny et al., 2000), and a significant
714 increase in the transport of Saharan dust (Moreno et al., 2002) were observed to occur during HS
715 4. This confirms the magnitude of HS 4 (cf. Eynaud et al., 2009) in terms of expansion of polar
716 waters towards the Iberian margin (i.e. February SST 11°C lower than modern ones) and even
717 towards the western Mediterranean (i.e. February SST 9°C lower than modern ones), in phase
718 with a nearly complete shutdown of the thermohaline circulation (Maslin et al., 1995; Elliot et al.,
719 2002; Roche et al., 2004).

720 HS 3 is marked by a change in surface hydrological conditions with February/winter SST of
721 about 6.5°C in the Alboran Sea and 5.5°C off Portugal (Fig. 3). Winter conditions seem thus to
722 have been less severe during HS 3 than HS 4, although they are nevertheless characterised in both
723 paleo-records by significant cooling.

724 Finally, HS 5 is the event which displays the weakest changes both in the Alboran Sea and off
725 Portugal, where dinocyst February SST are about 9°C and 13°C, respectively (Fig. 4). However,
726 foraminiferal winter SST estimates are close to 6°C and 5°C in the Alboran Sea and off Portugal,
727 respectively. Quantifications derived from transfer functions appear less reliable during this
728 interval. The alkenone annual signal (around 10.5-11°C in the Alboran Sea and 12-12.5°C off
729 Portugal) probably provide an intermediate signature between dinocyst-based (too warm) and
730 foraminiferal-based (too cold) February/winter SST estimates.

731
732 Micropaleontological evidence in the subpolar North Atlantic has shown that abrupt SST changes
733 associated with the D-O events in Greenland were matched by SST variations of at least 3-5°C
734 (Bond et al., 1992, 1993; Elliot et al., 2002). In the subtropical North Atlantic, SST excursions of
735 4-5°C across stadial-interstadial transitions of the last glacial were recorded from the Bermuda
736 Rise (alkenones: Sachs and Lehman, 1999; isotopes: Keigwin and Boyle, 1999) and off Portugal
737 (alkenones: Bard et al., 2000; Martrat et al., 2007). SST changes of up to 6°C are also
738 documented in the western Mediterranean (alkenones: Martrat et al., 2004), due to southward
739 shifts in the position of the Polar Front. Consistently with these previous results, we show here
740 that the Alboran Sea and the SW Iberian margin experienced fluctuations of dinocyst SST of
741 around 5-6°C between GI and GS, except for HS events and more specifically HS 4 with a SST
742 drop of around 11°C off Portugal and of 8°C in the Alboran Sea. This demonstrates the extreme
743 sensitivity of dinocysts to climate fluctuations in subtropical latitudes.

744

745 **6.2.2. Multiproxy evidence for paleohydrological changes through time**

746

747 During GI, the expansion of the Mediterranean forest (Fig. 4) was attributed to atmospheric

748 conditions close to present-day ones with warm summer and wet winter conditions over south-
749 western Europe (e.g. Sánchez-Goñi et al., 2002; Combourieu-Nebout et al., 2002; Bout-
750 Roumazeilles et al., 2007; Daniau et al., 2007). During those times, we show that warm
751 temperate species *S. mirabilis* and *I. patulum* expanded in the Alboran Sea and on the SW Iberian
752 margin, respectively (Fig. 4), and high SST is recorded at both sites during GI revealing sea-
753 surface conditions closer to present-day ones (Figs. 3 and 4).

754
755 During GS, the decline of the Mediterranean forest (Fig. 4) and the development of steppe and
756 semi-desert vegetation over the south-western European borderlands were attributed to
757 intensified winter dryness with cold continental conditions affecting the western Mediterranean
758 area (e.g. Sánchez-Goñi et al., 2002; Combourieu-Nebout et al., 2002; Moreno et al., 2002, 2005;
759 Roucoux et al., 2005; Bout-Roumazeilles et al., 2007; Daniau et al., 2007). At that time,
760 assemblages of cold-water dinocyst species, including *B. tepikiense* and *S. elongatus* developed,
761 reflecting cold-water advection along the SW Iberian margin and towards the Alboran Sea (Fig.
762 4). However, unlike the western Iberian margin where *B. tepikiense* occurs during each GS, *B.*
763 *tepikiense* expands principally in the Alboran Sea during HS, and especially during HS 3 and HS
764 4. In the western Mediterranean Sea, HS impacts were therefore greater than those associated
765 with the other GS. Broecker (2006) has compiled data from several areas and has identified sites
766 where the impact associated with North Atlantic Heinrich events is larger (eastern Brazil, central
767 Florida, Arabian Sea, Chinese stalagmites, and western Mediterranean area), in contrast to sites
768 where impacts are similar to those observed during the other stadials (Greenland ice and Cariaco
769 Basin). To explain this discrepancy, Broecker (2006) involves the greater magnitude of north-
770 hemispheric sea ice expansion associated with North Atlantic Heinrich events which exceeded
771 that associated with the other stadials.

772 *B. tepikiense* was previously used to trace subpolar water masses on the western Iberian margin
773 (Eynaud et al., 2000; Turon et al., 2003) and in the Alboran Sea (Turon and Londeix, 1988;
774 Combourieu-Nebout et al., 2002) during HS. In our data, the observation of subpolar dinocysts in
775 both cores, in phase and synchronous with peaks of *N. pachyderma* s. (Fig. 4), confirms the idea
776 of the shift of the Polar Front (PF) towards southern latitudes during North Atlantic Heinrich
777 events. Indeed, Eynaud et al. (2009) proposed a conceptual scheme for the position of the PF on
778 the western Iberian margin during Heinrich events and its influence on the local hydrology, and
779 noticed that the protrusion of subpolar waters extended until approximately 40°N. Core MD95-
780 2042 (37°48'N), at the southern limb of the Ruddiman belt (i.e. between 40 and 55°N;
781 Ruddiman, 1977), was affected by subpolar waters and iceberg discharges. However, it is
782 surprising to observe that the peaks of *B. tepikiense* are even larger during GS not associated with
783 HS on the SW Iberian margin. When considering dinocyst, foraminiferal, and especially
784 alkenone-based SST reconstructions, HS are characterised by the coldest temperatures (Figs. 3
785 and 4). We can therefore assume that temperatures were probably too cold during summer
786 months to permit the expansion of *B. tepikiense*. This hypothesis is reinforced by the large
787 excursions towards cold summer temperatures observed with dinocyst and foraminiferal SST
788 reconstructions (Fig. 3). The huge advection of subpolar waters down to the SW Iberian margin
789 during each GS, and especially HS, is furthermore reinforced by the dinocyst SSS reconstructions
790 that show a generally more pronounced influence of meltwater in this sector compared with the
791 westernmost part of the Mediterranean Sea (average salinity offset of around 1 psu between the
792 sectors; Fig. 3). This is especially true for HS 4, characterised by salinities of around 31 on the
793 SW Iberian margin and 33 in the Alboran Sea (Fig. 3). We thus show the pronounced impact of
794 subpolar waters affecting the SW Iberian margin during each GS and especially during HS, while
795 the Alboran Sea is only impacted by huge freshwater discharges during HS.

796

797 **6.2.3. A first attempt to interpret the dinocyst seasonality signal**

798

799 Sea surface temperatures and precipitation are environmental parameters directly linked to
800 seasonality in the study region, and the difference between August and February SST estimates
801 derived from dinocysts may contain important information regarding the seasonality (Fig. 5).
802 Overall, the two cores show excursions in seasonality reconstructions towards a higher seasonal
803 contrast during GS due to extremely cold February SST recorded during these cold intervals. At
804 the SW Iberian margin, seasonal contrasts during GI appear similar to the present-day range,
805 while extremely pronounced seasonal contrasts of roughly similar magnitude are recorded during
806 each GS. The pattern of seasonality is very close to the relative abundance curve of *B. tepikiense*
807 (Fig. 5). In the Alboran Sea, higher seasonal contrasts are mainly noted during HS and also
808 correspond to higher percentages of *B. tepikiense*. It is not surprising to find parallels between
809 this species and the seasonal signal since highest abundances of this species are found in areas
810 today characterised by high-amplitude (10°C) seasonal temperature shifts (Rochon et al., 1999).
811 In the Alboran Sea, it confirms the establishment of an enhanced seasonal temperature contrast
812 compared to the present-day caused by a strong decrease in winter SST during HS (Combourieu-
813 Nebout et al., 2002). At the SW Iberian margin, it reveals very clearly the marked shift between
814 seasonal temperatures during each GS.

815 In the Alboran Sea, another climatic trend is superimposed on the general observations made
816 above with, in broad terms, stronger seasonal contrasts during the early part of the record
817 between 50 and 38 ka, compared to the period 38-31 ka (Fig. 5). Since seasonality and precession
818 are closely linked, we show the precession curve (Berger and Loutre, 1991) alongside the
819 reconstruction of seasonality (Fig. 5). It appears that lower seasonal contrasts are observed during

820 the precession minimum and *vice versa* (Fig. 5). We would have expected higher seasonal
821 contrasts during the precession minimum since this orbital parameter enhanced the Mediterranean
822 climate with warmer summer and wetter winters (Meijer and Tüenter, 2007). When looking at
823 February SST reconstructions in the Alboran Sea, one can note that fluctuations during GS and
824 GI are of similar magnitude (Fig. 5). The contrast is mainly due to August SST reconstructions
825 that show colder values between HS 4 and HS 3 than between HS 5 and HS 4. Our data would
826 suggest a link between the Alboran Sea paleohydrology and the precession signal through a
827 climatic forcing acting on August temperatures. A first hypothesis could involve the general
828 strengthening of the gyre-induced upwelling in the Alboran Sea during the time interval 38-31 ka
829 favouring a cooling of sea-surface waters compared to the period 50-38 ka. However, the
830 functioning of the gyres is mainly controlled today by winter conditions. Furthermore, no large
831 differences are observed between 50-38 ka and 38-31 ka in terms of total dinocyst concentrations
832 and heterotrophic dinocyst species, these latter proxies indicating paleoproductivity conditions
833 probably connected to gyre-induced upwelling intensity. Greater stratification of the Alboran
834 water column may also be suggested for the period 50-38 ka compared to the period 38-31 ka on
835 the basis of the observation of dinocyst and foraminiferal August/summer SST (Fig. 3). Indeed,
836 SSTs in August and summer are similar between 38 and 31 ka and are more distinct between 50
837 and 38 ka (Fig. 3). This might suggest a greater stratification of water masses during summers
838 within the interval 50-38 ka and a stronger mixing of water masses during summers within the
839 interval 38-31 ka. Warmer dinocyst SSTs might relate to sea-surface conditions while
840 foraminiferal SSTs would incorporate SST signals at greater depths in the water column for the
841 period 50-38 ka. Further investigation will be necessary to verify and explore on longer time-
842 scales: a) the imprint of precession on the seasonality changes inferred from dinocysts, and b) the
843 functioning of the mesoscale energetic features in the Alboran Sea represented by two

844 anticyclonic gyres today.

845 **7. CONCLUSION**

846

847 We have characterised glacial climate variability between 25 and 50 ka BP by comparing surface
848 paleohydrology signals on both sides of the Strait of Gibraltar (Alboran Sea and SW Iberian
849 margin). Comparison of dinocyst assemblages enables the reconstruction of hydrological features
850 at both locations and we present, in this study, the first quantitative dinocyst reconstructions (SST
851 and SSS) for MIS 3 obtained at mid-latitudes. Our hydrological quantifications acquired for both
852 cores reproduce millennial-scale changes correlated to the D-O climatic variability, with a pattern
853 of marked decrease in SST accompanied by a strong freshening of sea-surface waters evident
854 during each GS on the SW Iberian margin and during HS in the Alboran Sea. Furthermore, we
855 show similar patterns and amplitudes in SST reconstructions based on dinocysts, foraminifera
856 and alkenones, and in SSS derived from dinocysts and foraminiferal SST coupled with planktonic
857 $\delta^{18}\text{O}$. Larger discrepancies between dinocyst and foraminiferal estimates occur for summer
858 reconstructions than for winter reconstructions, which are very similar in amplitude. We
859 furthermore show the occurrence of cold taxa (including *B. tepikiense* and *S. elongatus*) during
860 GS and the presence of thermophilous ones (*S. mirabilis* and *Impagidinium* spp.) during GI.
861 However, the amplitude of variation in these taxa is not equivalent at both sites. *B. tepikiense*
862 characterises each GS in the SW Iberian margin while it only develops during HS in the Alboran
863 Sea. This pattern reflects, in the Alboran Sea, the maximum climatic deterioration during HS and
864 the incursion of low-salinity subpolar waters to the western Mediterranean. During other GS (i.e.
865 excluding HS), conditions were less severe in the Alboran Sea as is reflected by the occurrence of
866 *O. centrocarpum* at very high relative abundances in the Alboran Sea and low values on the SW
867 Iberian margin. This species conceivably reflects the inflow of cool North Atlantic waters to the

868 western Mediterranean, synchronously with stronger deep convection occurring at that time in the
869 western Mediterranean. Finally, expansions of temperate to tropical taxa testify to the installation
870 of warm sea-surface waters during GI. In the Alboran Sea, a high primary productivity pattern is
871 deduced from heterotrophic species and high total dinocyst concentrations, suggesting gyre-
872 induced upwelling due to prevailing southward-shifted westerlies above the Mediterranean at that
873 time. The functioning of the gyre on multi-millennial timescales has also been discussed in light
874 of seasonality reconstructions based on dinocysts. However, the relationships between gyre
875 dynamics, seasonality and precession need to be explored in longer records covering several
876 precession cycles. The dinocyst data thus exhibit regional trends and suggest distinct oscillations
877 of sea-surface temperature and salinity, documenting the combined influence of atmospheric and
878 hydrologic processes impacting on the western Mediterranean Sea and eastern subtropical
879 latitudes of the North Atlantic during the abrupt climatic events of MIS 3.

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889

890 9. References

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1504 organic-walled dinoflagellate cysts in surficial sediments of the Benguela upwelling region and
1505 their relationship to upper ocean conditions. *Progress in Oceanography* 48, 25-72.

1506 **10. Figure captions**

1507

1508 **Figure 1:** Area of interest with major sea-surface features. The studied cores MD95-2043
1509 (Alboran Sea; 36°8.6'N; 2°37.3'W; 1841 m water depth) and MD95-2042 (Iberian margin;
1510 37°48'N; 10°10'W; 3146 m water depth) are located on the large map, depicting also the
1511 bathymetry of the study area and the major surface currents within the Alboran sea; WAG:
1512 Western Alboran Gyre; EAG: Eastern Alboran Gyre; AOF: Almeria-Oran Front; AC: Algerian
1513 Current. The small map on the left shows large scale North-Atlantic currents with: the North
1514 Atlantic Drift (NAD), the Portugal Current (PC) flowing southward from 45°N to 30°N, the
1515 Azores Current (AzC) derived from the southern branch of the Gulf Stream and flowing eastward
1516 to the Gulf of Cadiz at about 35°N, and the Canary Current (CC) fed by both the AzC and the PC.
1517 Together, these currents form the Eastern Boundary Current of the North Atlantic subtropical
1518 gyre. A Mediterranean analogue of the dinocyst modern database has been located on the map
1519 (M1039). Colors from red to dark blue on the map reflect growing bathymetry towards the
1520 deepest areas.

1521

1522 **Figure 2:** MD95-2043 (a) and MD95-2042 (b). Core depths are displayed in centimetres along
1523 the vertical axis. The relative abundances of selected dinocyst species are compared with
1524 planktonic $\delta^{18}\text{O}$ data and $U^{k'}_{37}$ -SST (Core MD95-2043: Cacho et al., 1999; Core MD95-2042:
1525 Cayre et al., 1999; Shackleton et al., 2000; Paillet and Bard, 2002), providing the stratigraphical
1526 framework for the two cores. Quaternary dinocyst concentrations are also illustrated. HS:
1527 Heinrich Stadial; GI: Greenland Interstadial.

1528

1529 **Figure 3:** MD95-2043 (a) and MD95-2042 (b). Quantitative dinocyst reconstructions (February
1530 and August SST and SSS, 3PBase-940) compared with SST provided by foraminifera (winter,
1531 summer and annual, R-1007) and alkenones ($U^{k'}_{37} - SST$), and SSS provided by calculations
1532 based on planktonic $\delta^{18}O$ (*G. bulloides*) and foraminiferal SST estimates (winter and summer, R-
1533 1007). Error bars are shown in the figure for the different reconstructions. HS: Heinrich Stadial;
1534 GI: Greenland Interstadial.

1535
1536 **Figure 4:** Comparison between cores MD95-2043 and MD95-2042 of a multi-proxy compilation
1537 including: selected dinocyst species, total dinocyst concentrations, February dinocyst (3PBase-
1538 940) and alkenone SST (Cacho et al., 1999; Pailler and Bard, 2002) reconstructions, percentages
1539 of *N. pachyderma* s. (Cacho et al., 1999; Pérez-Folgado et al., 2003), benthic $\delta^{13}C$ (Cacho et al.,
1540 2006), and percentages of the Mediterranean forest (Sánchez-Goñi et al., 2000, 2002; Fletcher
1541 and Sánchez-Goñi, 2008). These data are compared with $\delta^{18}O$ ice core records and the precession
1542 signal (Berger and Loutre, 1991). HS: Heinrich Stadial; GI: Greenland Interstadial.

1543
1544 **Figure 5:** MD95-2043 (a) and MD95-2042 (b). Estimation of a parameter of seasonality based
1545 on the difference between February and August dinocyst SST reconstructions (3PBase-940),
1546 compared with percentages of the dinocyst species *Bitectatodinium tepikiense*, the precession
1547 signal (Berger and Loutre, 1991) and planktonic $\delta^{18}O$ records (*G. bulloides*). Dinocyst and
1548 alkenone SST records are also represented, as well as modern SST values. HS: Heinrich Stadial;
1549 GI: Greenland Interstadial.

1550

1551 **11. Appendices**

1552

1553 **Appendix A:** MD95-2042 (SW Iberian Margin) and MD95-2043 (Alboran Sea): individual
1554 counts of dinocyst species, total dinocysts counted and total dinocyst concentrations per sample.

1555

1556 **Appendix B:** MD95-2042 (SW Iberian Margin) and MD95-2043 (Alboran Sea): lists of the five
1557 best analogues found with the transfer function 3PBase-940. The geographical coordinates of
1558 each named analogue can be found on the GEOTOP website

1559 (http://www.geotop.ca/index.php?option=com_content&task=view&id=762&Itemid=226). The

1560 five analogues found by 3PBase-940 have systematically been used for the calculations, the

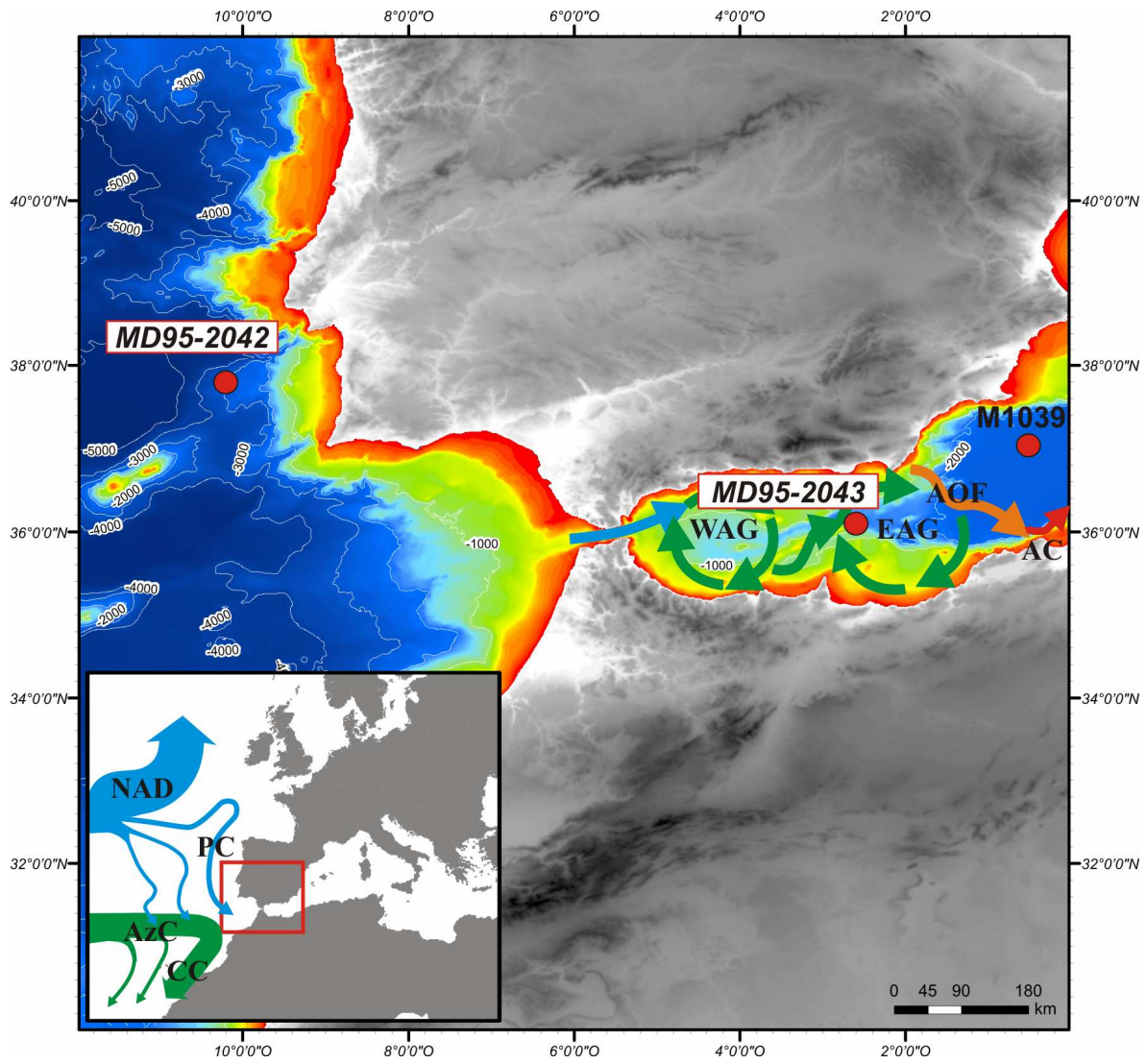
1561 threshold (Dmin value of 71.72) may then be considered to judge the good reliability of the

1562 reconstructed hydrological parameters. The M1039 analogue, highlighted in the tables in bold, is

1563 a Mediterranean analogue located in Figure 1.

1564

1565



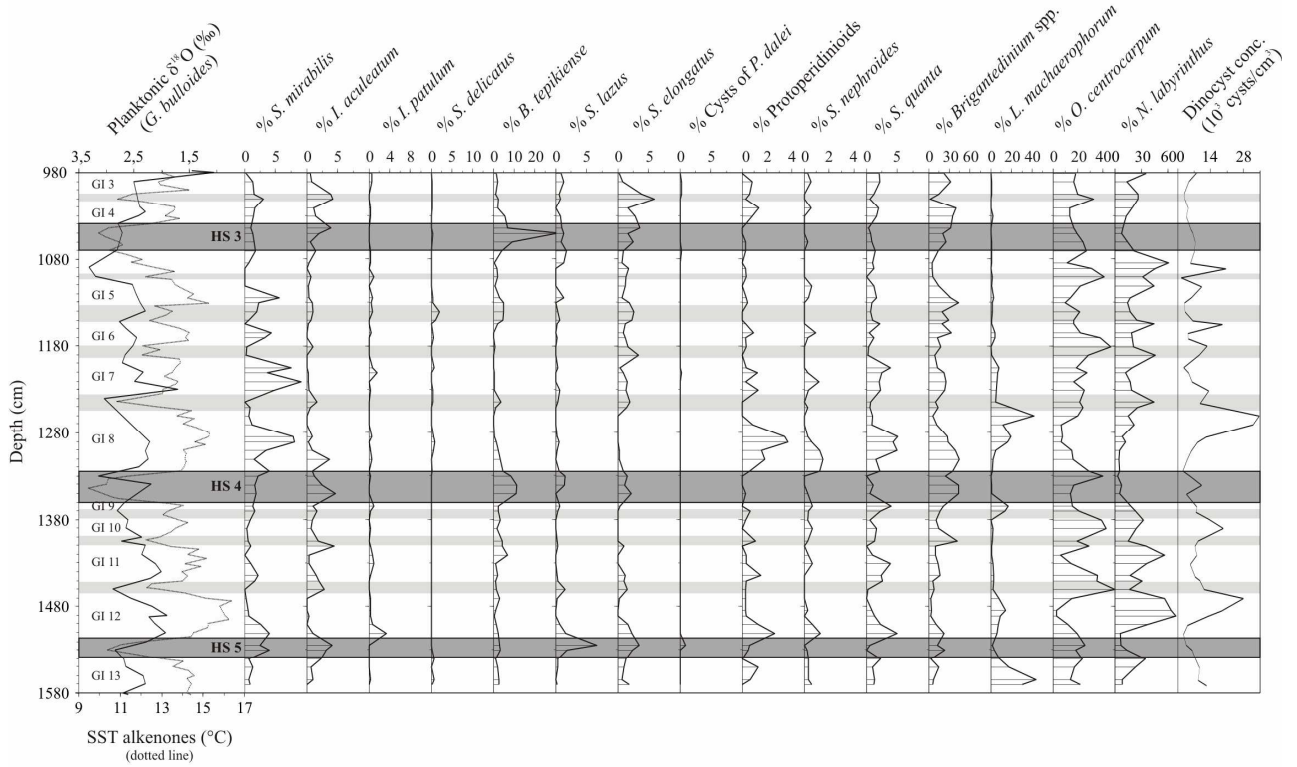
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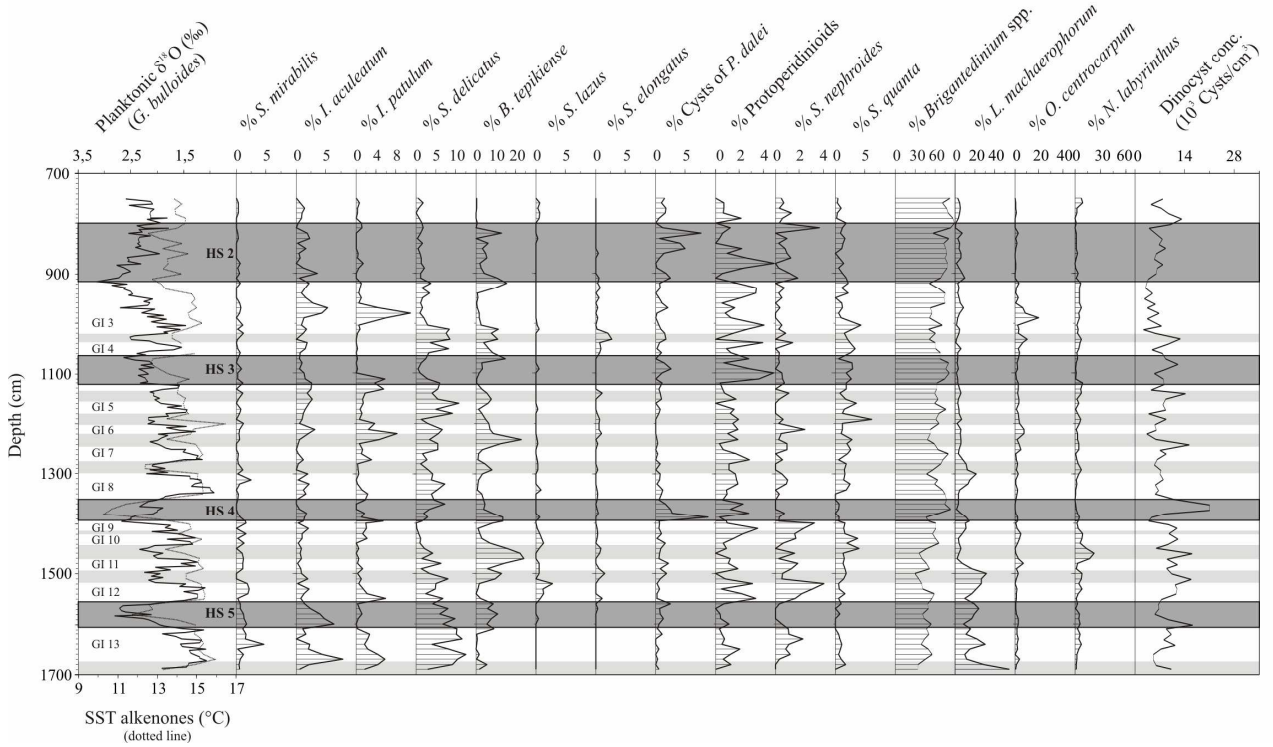
1568

1569 **Figure 1**

a) MD95-2043 - Alboran Sea



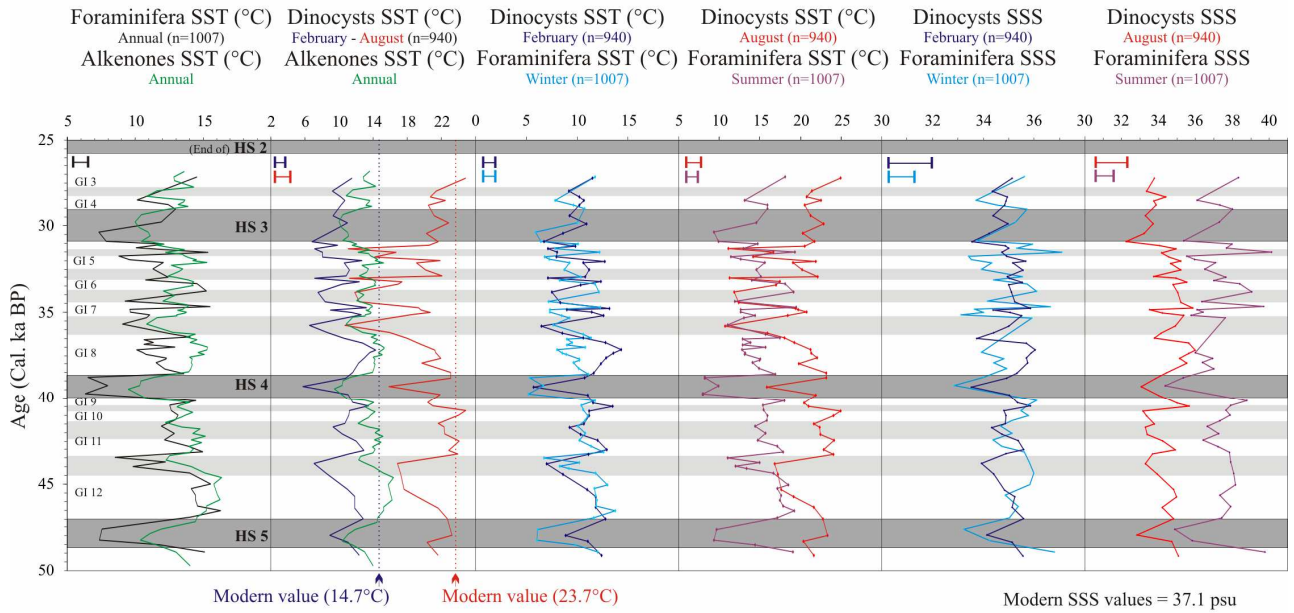
b) MD95-2042 - SW Iberian Margin



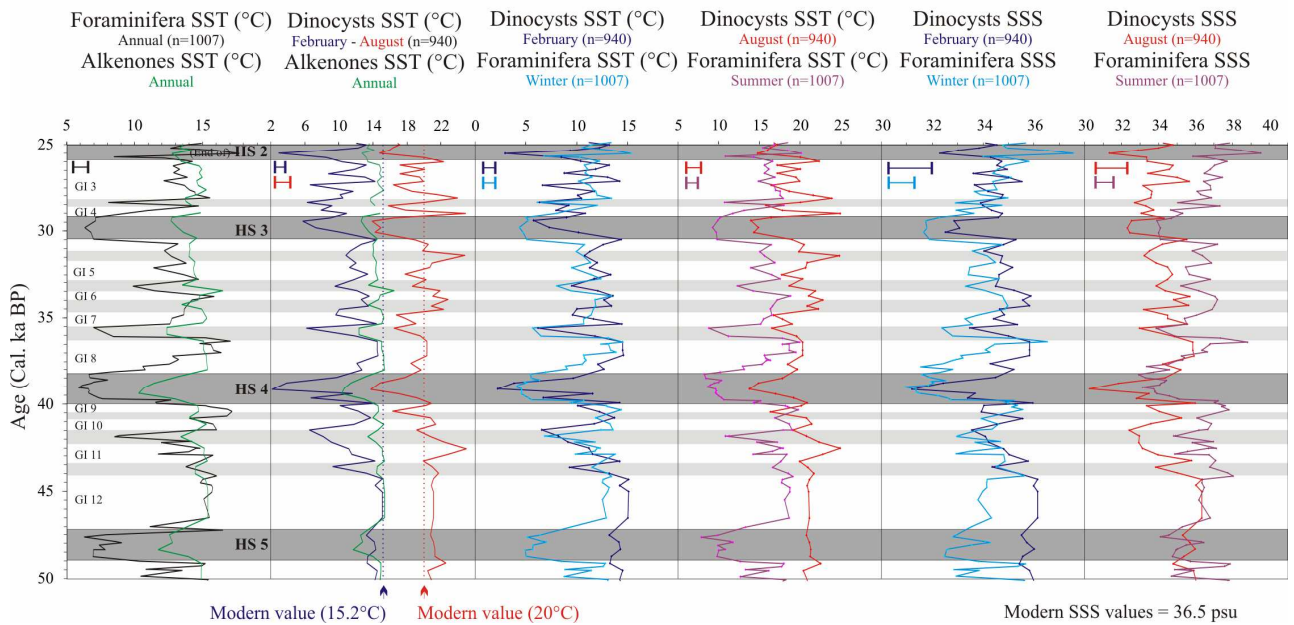
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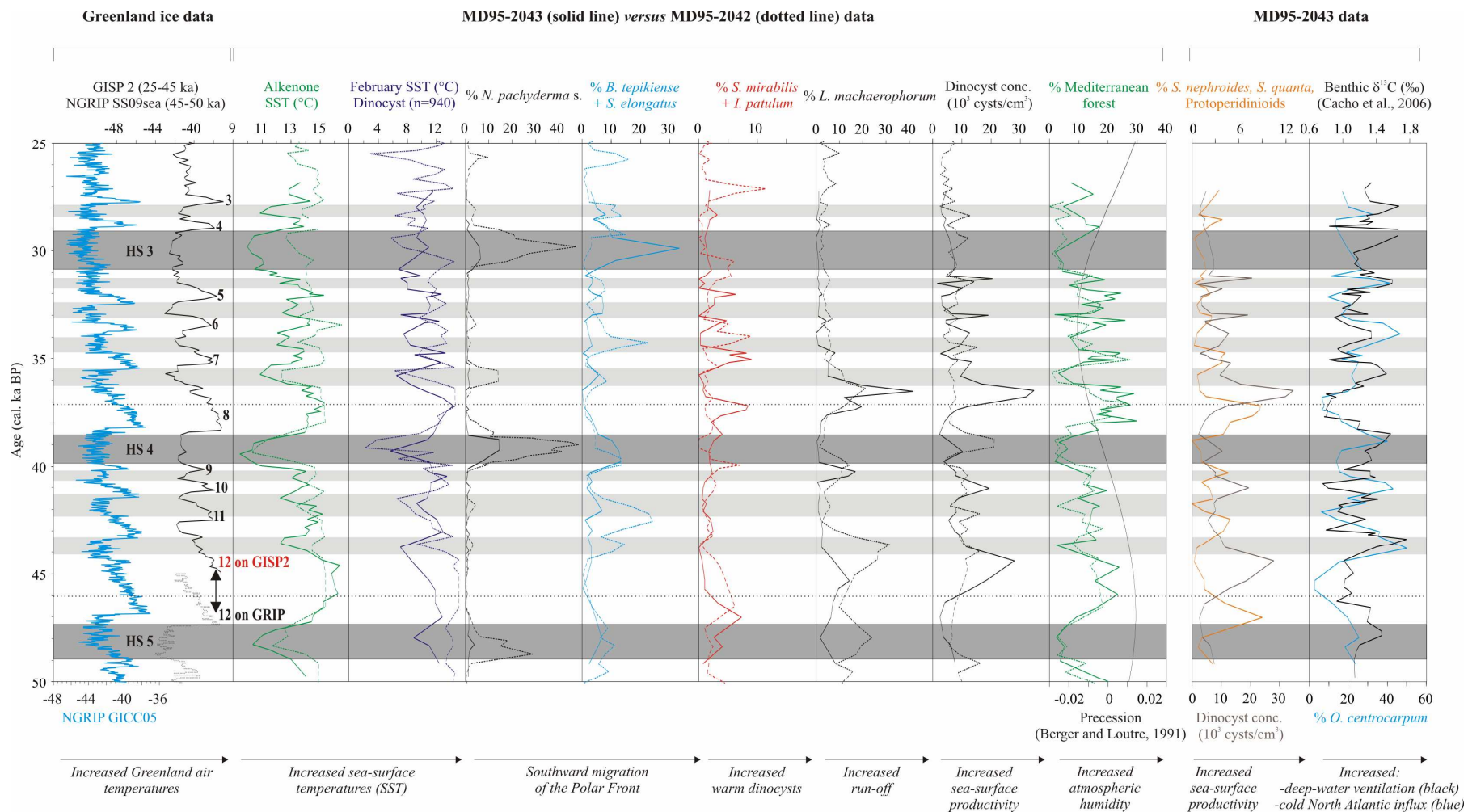
1571 **Figure 2**

a) MD95-2043 - Alboran Sea



b) MD95-2042 - Iberian Margin



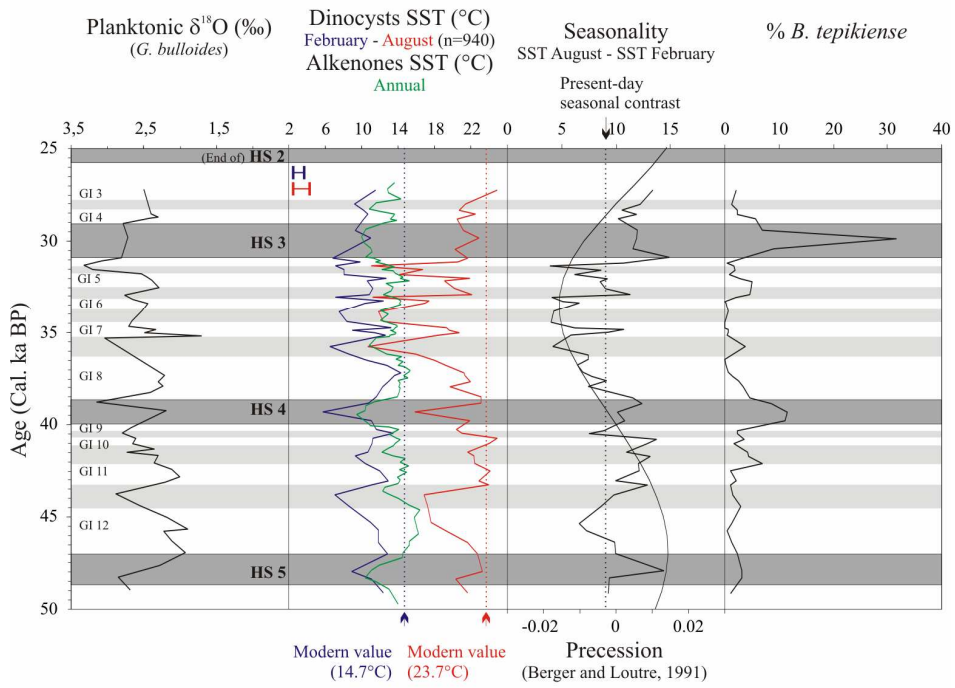


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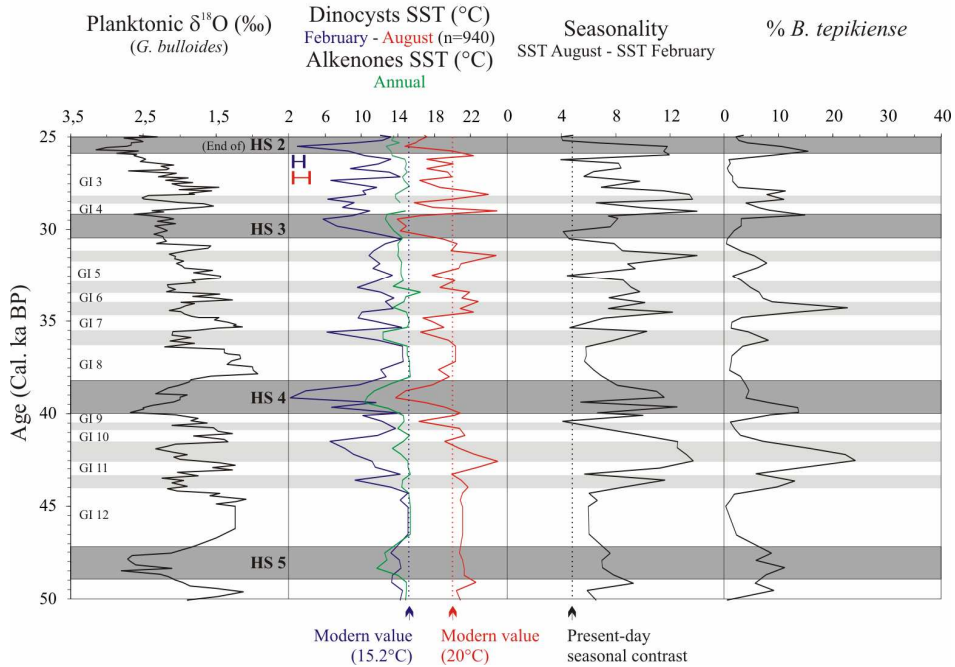
1577

1578 **Figure 4**

a) MD95-2043 - Alboran Sea



b) MD95-2042 - Iberian Margin



1579

1580 **Figure 5**

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1581