

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

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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
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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 planktonic δ^{18} O and foraminiferal SST. Significant oscillations related to Dansgaard-Oeschger 24 25 cycles are recorded in both cores. Dinocyst-derived hydrological parameters exhibit synchronous fluctuations and similar values to those derived from the other methods, in particular when 26 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-Roumazeilles 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

the Meridional Overturning Circulation to switch from stadial to interstadial mode through the 63 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 region and therefore on dominant modes of climatic variability, notably the North Atlantic 68 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; Daniau 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).

Previous high resolution palynological analysis of dinoflagellate cysts (dinocysts) and pollen on core MD95-2042, collected off Portugal, indicate abrupt changes in sea surface and atmospheric conditions, respectively, in response to D-O cycles (Eynaud, 1999; Sánchez-Goñi et al., 2000). Additionally, a climatic contrast has been observed with respect to precipitation between the SW lberian margin and the Alboran region, with dryer conditions in southeastern Iberia than in its southwestern part (Sánchez-Goñi et al., 2002). However, these studies do not quantify either 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 δ^{18} O of <i>Globigerina bulloides</i> associated with foraminiferal SST).

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

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; Jorga 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
- around 15.2 and 14.7°C, respectively, and August SSTs are around 20 and 23.7°C (WOA 2001).

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148 **3.1. Chronostratigraphy**

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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). The age model of core MD95-2042 is derived from 16 AMS ¹⁴C dates and, for the period beyond 154 AMS ¹⁴C range, by graphical tuning of the MD95-2042 planktonic δ^{18} O record with Greenland 155 156 ice core δ^{18} 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 ¹⁴C dates and graphical tuning of the

161 MD95-2043 alkenone SST record to GISP2 δ^{18} 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

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

167

166

168 **3.2. Dinoflagellate analysis**

with that of the SW Iberian margin.

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	

Dinocyst assemblages were characterised at the species level on the sediment fraction smaller
than 150 µm on 61 palynological slides for core MD95-2043 (this study) and on 71 slides for
core MD95-2042 (Eynaud, 1999; Sánchez-Goñi et al., 2000). The preparation technique followed
the protocol described by de Vernal et al. (1999) and Rochon et al. (1999), slightly modified at
the EPOC laboratory (Castera and Turon, http://www.epoc.u-

bordeaux.fr/index.php?lang=fr&page=eq_paleo26). Each subsample of 8 cm³ was weighed, dried
overnight and then weighed again to obtain dry weight. Subsamples were then washed through a
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 $10 \,\mu m$ 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

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

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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 including station "M1039", cf. Fig. 1; Mangin, 2002), as well as epicontinental environments 226 227 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-230 cnrs.com/pages/3pbase.htm) and was subsequently applied to dinocyst assemblages (e.g. de Vernal et al., 2001, 2005). An index "Dmin", provided by the software "3PBase", allows testing 231 232 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 $\pm 1.2^{\circ}$ C and $\pm 1.8^{\circ}$ 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 from the foraminiferal assemblages of cores MD95-2042 (Cayre et al., 1999) and MD95-2043 247 248 (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-260 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 mean SST with a prediction error of $\pm 1.2^{\circ}$ C and $\pm 1.3^{\circ}$ C respectively, and annual SST with a 262 263 prediction error of $\pm 1.1^{\circ}$ C (Eynaud et al, 2009; Matsuzaki et al., in press).

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

268

$$\delta^{18}$$
Osw = 0.66 SSS - 23.5

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

In order to estimate past changes in oceanic δ^{18} Osw, Epstein et al. (1953) established a paleotemperature equation which links temperature with the isotopic composition of calcite (δ^{18} Oc) in calcareous shells (e.g. foraminifera) and the ambient waters (δ^{18} Osw). Shackleton and Opdyke (1973) have adapted the equation of Epstein et al. (1953) as follows:

280

$$T = 16.9 - 4.38 (\delta^{18}Oc - \delta^{18}Osw) + 0.13 (\delta^{18}Oc - \delta^{18}Osw)^2$$

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

al., 2002), and used the modern δ^{18} Osw values (1.2 ‰ for the Alboran Sea and 0.97 ‰ for the 288 289 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 Iberian margin, from http://data.giss.nasa.gov/o18data). The residual δ^{18} Osw "ice-corrected" 291 292 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 ‰ uncertainty on a δ scale), as well as on uncertainties in the global δ^{18} Osw changes linked to 295 296 sea level changes (0.15 %).

4. Dinocyst assemblages through time

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 δ^{13} 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

- 17
- 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 Scotia and the Gulf of Maine (Wall et al., 1977; Mudie, 1992). This species is characteristic of 333 334 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 (Rochon et al., 1999; Marret and Zonneveld, 2003). Previous results from the North Atlantic 336 (Zaragosi et al., 2001; Penaud et al., 2009), the SW Iberian margin (Sánchez-Goñi et al., 2000; 337 338 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

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

347

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

354

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

360

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

times higher than in the A

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

396

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

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 opposite415 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

L. machaerophorum occurs during GI in the Alboran Sea, particularly GI 8, 9 and 12 (Fig. 2a),
while it occurs during GS on the SW Iberian margin (Fig. 2b). *L. machaerophorum* is today
considered to be a temperate to tropical coastal euryhaline species (Mertens et al., 2009). It
dominates associations from the northern African and southern European Atlantic shelves, and it
is also found in abundance in North African coastal upwelling regions (Targarona et al., 1999;
Sprangers et al., 2004) and near the Congo outlet (Marret, 1994). Extremely high concentrations
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	

When comparing SST reconstructions from both micropaleontological proxies on the two cores,
one can note that February *versus* winter SST values obtained with dinocysts (3PBase-940) and
foraminifera (R-1007), respectively, are closer than August *versus* summer ones (Fig. 3).
However, the general good consistency of February *versus* winter SST is not valid for the SW
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 for a 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.

Finally, concerning the Alboran Sea core, one can note that dinocysts and foraminiferal show
closer August *versus* summer SST values between 31 and 38 ka and larger offsets between 38
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

The foraminiferal transfer function n=1007 also provides annual SST reconstructions that can directly be compared with the alkenone signal. One can note that annual SST values reconstructed with both proxies are closer during GI than during GS (Fig. 3). Foraminifera 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: dinocyts 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
- 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 for aminifera and δ^{18} O analysis (e.g. Malaizé and

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

564

For the SW Iberian margin, our study shows that major low salinity events are recorded during 565 566 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 (Eynaud et al., submitted). For the Alboran Sea, the difference in the resolution of analysis between dinocyst and 572 573 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 planktonic δ^{18} O data and the resolution of the δ^{18} O record is lower than the foraminiferal SST 575 576 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

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

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

2	1
J	T

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>i.e.</i> 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>i.e.</i> 9°C less than at
662	present (Fig. 4). It is thus important to note that, in contrast to N. pachyderma s., B. tepikiense
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. tepikiense 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 δ^{13} 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

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

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, for a miniferal, 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

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 Tuenter, 2007). When looking at 823 February SST reconstructions in the Alboran Sea, one can note that fluctuations during GS and GI are of similar magnitude (Fig. 5). The contrast is mainly due to August SST reconstructions 824 825 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 and 38 ka (Fig. 3). This might suggest a greater stratification of water masses during summers 837 838 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 for a signal straight for the straight of 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

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 δ^{18} O. Larger discrepancies between dinocyst and foraminiferal estimates occur for summer 857 858 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 characterises each GS in the SW Iberian margin while it only develops during HS in the Alboran 862 863 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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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 $37^{\circ}48$ 'N; $10^{\circ}10$ '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 to the Gulf of Cadiz at about 35°N, and the Canary Current (CC) fed by both the AzC and the PC. 1516 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 Figure 2: MD95-2043 (a) and MD95-2042 (b). Core depths are displayed in centimetres along 1522 the vertical axis. The relative abundances of selected dinocyst species are compared with 1523 planktonic δ^{18} O data and U^{k'}₃₇-SST (Core MD95-2043: Cacho et al., 1999; Core MD95-2042: 1524

1525 Cayre et al., 1999; Shackleton et al., 2000; Pailler 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.

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 δ^{18} O (<i>G. bulloides</i>) 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 <i>N. pachyderma</i> s. (Cacho et al., 1999; Pérez-Folgado et al., 2003), benthic δ^{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 δ^{18} O records (<i>G. bulloides</i>). 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

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	










Figure 2 1571

1570



SPP

a) MD95-2043 - Alboran Sea

500 Coloo

a) MD95-2043 - Alboran Sea







1575 Figure 3

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Figure 4

a) MD95-2043 - Alboran Sea



b) MD95-2042 - Iberian Margin



