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Climate evolution in the Adriatic Sea across the last deglaciation: a multiproxy approach combining biomarkers and calcareous plankton

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Abstract:	A multiproxy study combining calcareous plankton assemblages (coccolithophore and planktonic foraminifera), and terrestrial (n-alkanes) and marine (alkenones) biomarkers was carried out in a sediment core (ND14Q-AR2) from the South Adriatic Sea. The focus of the study is to investigate millennial-to-centennial scale climate variability in the Eastern Mediterranean during the last deglaciation, between 20 and 11 ka BP. The high-resolution reconstruction allows for the characterization of the impact of the Heinrich Stadial 1 (HS1, here identified between 17.1 and 14.9 ka BP) at a centennial/multi-decadal time scale resolution. Based on terrestrial proxies, the HS1 interval has a two-fold partition: HS1a (17.1-15.9 ka BP), characterized by decreasing temperatures and relative high humidity, and HS1b (15.9-14.9 ka BP), characterized by the coldest temperatures and drier conditions on land. Terrestrial proxies suggest changes in moisture availability on land and ice melting delivery from the Alps during HS1. This period is followed by the Bølling-Allerød record (B-A, 14.9-12.3 ka BP) indicating ameliorated climate conditions and distinct local hydrological signals, related to global melt event similar to the Melt Water Pulse-1A. Finally, the comparison of our results with other Northern Hemisphere climatic records shows a good correspondence between the temperature variations in the South Adriatic Sea and the Greenland ice core oxygen isotope record, highlighting the climatic response of the South Adriatic to global climate variations. Comparison with other Mediterranean paleoclimatic records suggests a two-steps reorganization of the ocean/atmospheric circulation during the HS1 in the mid- and low-latitudes during the HS1 interval, but also a connection with the tropical Northern Hemisphere climate.
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Highlights:

- Planktonic assemblages and marine/terrestrial biomarker comparison.
- Late deglaciation, Heinrich Stadial 1 and Bølling-Allerød in the South Adriatic Sea.
- Two-step division of Heinrich Stadial 1, following changes in humidity on land.
- Melting water during the deglaciation inferred by planktonic assemblage proxies.
- Heinrich Stadial 1 partition well correlated with other paleoclimatic records.

1 **Climate evolution in the Adriatic Sea across the last deglaciation: a multiproxy approach combining** 2 **biomarkers and calcareous plankton**

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17

18 **Keywords**

19 Deglaciation, Heinrich Stadial 1, Adriatic Sea, terrestrial and marine biomarkers, calcareous plankton, climate
20 forcing.

21

22 **Abstract**

23 A multiproxy study combining calcareous plankton assemblages (coccolithophore and planktonic
24 foraminifera), and terrestrial (n-alkanes) and marine (alkenones) biomarkers was carried out in a sediment core
25 (ND14Q-AR2) from the South Adriatic Sea. The focus of the study is to investigate millennial-to-centennial
26 scale climate variability in the Eastern Mediterranean during the last deglaciation, between 20 and 11 ka BP.
27 The high-resolution reconstruction allows for the characterization of the impact of the Heinrich Stadial 1 (HS1,
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32 and ice melting delivery from the Alps during HS1. This period is followed by the Bølling-Allerød record (B-
33 A, 14.9-12.3 ka BP) indicating ameliorated climate conditions and distinct local hydrological signals, related
34 to global melt event similar to the Melt Water Pulse-1A. Finally, the comparison of our results with other
35 Northern Hemisphere climatic records shows a good correspondence between the temperature variations in
36 the South Adriatic Sea and the Greenland ice core oxygen isotope record, highlighting the climatic response
37 of the South Adriatic to global climate variations. Comparison with other Mediterranean paleoclimatic records
38 suggests a two-steps reorganization of the ocean/atmospheric circulation during the HS1 in the mid- and low-
39 latitudes during the HS1 interval, but also a connection with the tropical Northern Hemisphere climate.

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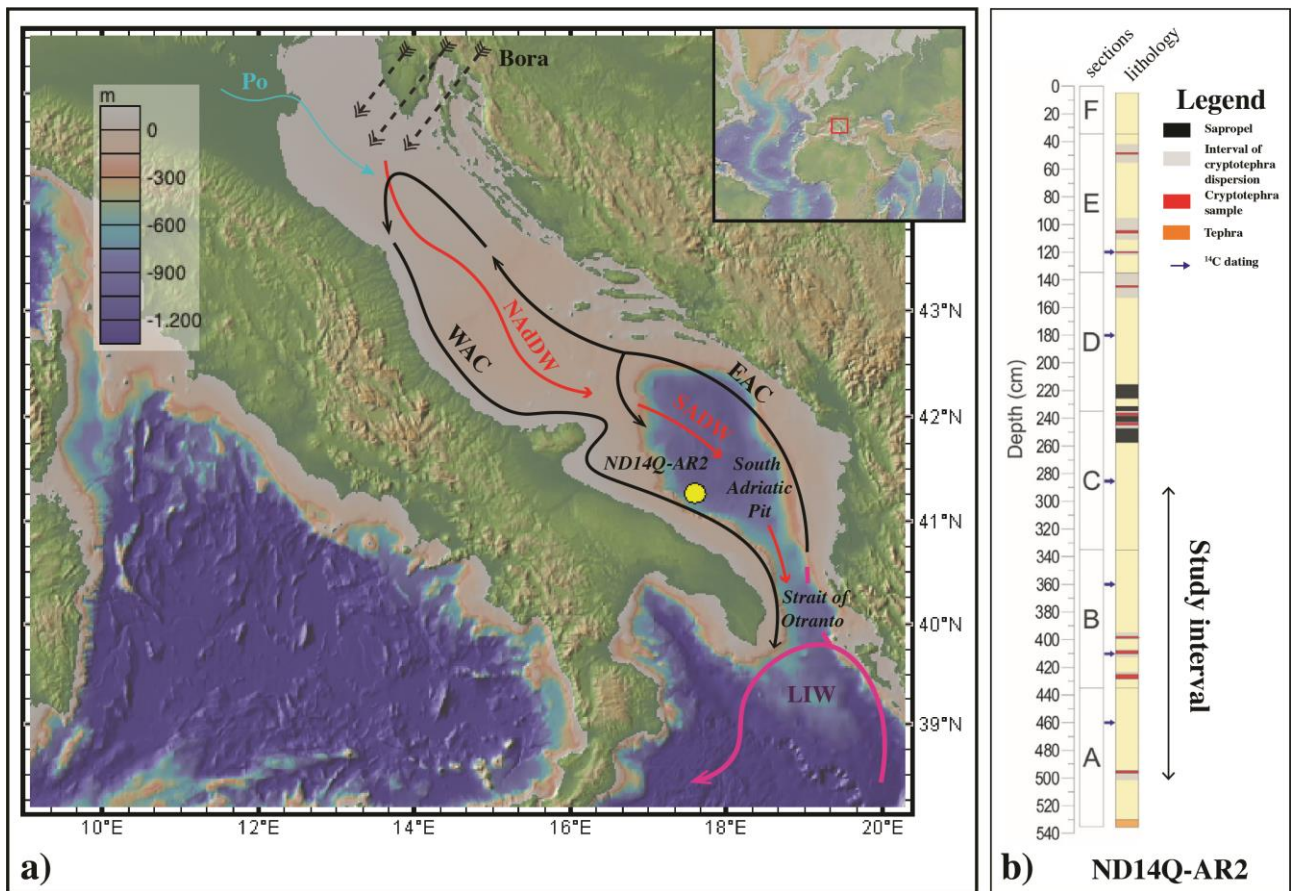
42 1. Introduction

43

44 The last deglaciation was punctuated by a series of large amplitude oscillations that left a distinct imprint in
45 the North Atlantic and Greenland records (Bond et al., 1992; Broecker et al., 1992; Dansgaard et al., 1993;
46 Bard et al., 2000; Rasmussen et al., 2006; Stanford et al., 2011; Clark et al., 2012; Schwab et al., 2012;
47 Toucanne et al., 2015; Repschläger et al., 2015; Hodell et al., 2017). Across this transition, the most severe
48 climatic period known as the Heinrich Stadial 1 (HS1) is well documented in the western Mediterranean (e.g.
49 Martrat et al., 2004, 2014; Sierro et al., 2005; Combourieu Nebout et al., 2009; Rogerson et al., 2010; Ausín
50 et al., 2015; Naughton et al., 2016; Bazzicalupo et al., 2018). Cold and fresher water inflow from iceberg
51 melting through Strait of Gibraltar most likely caused significant modification of the water column (Cacho et
52 al., 1999; Pérez-Folgado et al., 2003; Colmenero-Hidalgo et al., 2004; Martrat et al., 2004, 2014; Sierro et al.,
53 2005; Jiménez-Amat and Zahn, 2015; Toucanne et al., 2015) and notable changes on land in southern Europe
54 (Allen et al., 1999; Sanchez Goñi et al., 2002; Combourieu Nebout et al., 2009; Naughton et al., 2016). Several
55 records of HS1 (ca. 17.5–14.5 ka BP) in the western Mediterranean pointed out the complexity of this interval
56 involving changes in the atmospheric circulation and inflow of cold Atlantic waters (Moreno et al., 2005;
57 Fletcher and Sánchez Goñi, 2008; Frigola et al., 2008; Melki et al., 2009; Rodrigo-Gámiz et al., 2011;
58 Martínez-Ruiz et al., 2015; Bazzicalupo et al., 2018). Other studies in the central and eastern Mediterranean
59 revealed that the impact of HS1 was not confined to the western Mediterranean and Atlantic region and that
60 the eastern basin experienced marked cold and arid climate conditions during this period (Kwiecien et al.,
61 2009; Kotthoff et al., 2011; Desprat et al., 2013; Sicre et al., 2013; Castañeda et al., 2016). In fact, in the
62 Adriatic Sea, HS1 was characterized by increased aridity and extremely cold conditions (Asioli et al., 2001)
63 and some other records highlight distinct change in surface water temperatures, such as a two-step HS1 cooling
64 (Sicre et al., 2013).

65 The present study provides the first centennial/multi-decadal time scale reconstruction of paleoenvironmental
66 conditions across the last deglaciation, between 20 and 11 ka BP, in the South Adriatic Sea (SAS), combining
67 terrestrial and marine biomarkers with calcareous plankton assemblages (coccolithophores and planktonic
68 foraminifera). The sediment core ND14Q-AR2 used for this study was retrieved from the deepest part of the
69 South Adriatic basin. This area provides an ideal location for high resolution paleoclimatic reconstructions
70 because of thick deglaciation and Holocene sedimentary sequences and recurrent tephra layers for
71 geochronological constraints (Siani et al., 2004, 2013; Lowe et al., 2007; Jalali et al., 2018; Cascella et al.,
72 2019; Totaro et al., 2022). Terrestrial and marine biomarkers were used to infer environmental changes such
73 as Sea Surface Temperatures (SSTs), in tandem with terrestrial input associated with humidity changes on
74 adjacent lands (Jalali et al., 2018; Ternois et al., 2000). Recent studies have proven the value of cross-analysis
75 of marine and terrestrial biomarkers to interpret multidecadal climatic variations contained in paleoclimate
76 signals (Jalali et al., 2016, 2017, 2018; Schirrmacher et al., 2019). In fact, paired high-resolution marine and
77 terrestrial biomarker records in the SAS over the last three millennia have shown the impact of North Atlantic
78 Oscillation (NAO) and East Atlantic (EA) on SSTs centennial-scale variability (Jalali et al., 2018). On the

79 other hand, detailed calcareous plankton description from SAS are either scarce or mostly focused on
 80 planktonic foraminifera across the latest part of the deglaciation (Younger Dryas) to Holocene interval (Asioli
 81 et al., 1999, 2001; Capotondi et al., 1999; Sangiorgi et al., 2003; Favaretto et al., 2008; Piva et al., 2008;
 82 Narciso et al., 2012; Siani et al., 2013). In the same interval very few data exist on the coccolithophore
 83 assemblages (Giunta et al., 2003; Sangiorgi et al., 2003; Narciso et al., 2012; Cascella et al., 2019) and do not
 84 focus on the detail changes that occurred during the deglaciation. The study of core ND14Q-AR2 provides the
 85 first paleoenvironmental reconstruction of the deglaciation in the SAS that comprehends coccolith, planktonic
 86 foraminifera and biomarker analysis at high-temporal resolution to explore multidecadal changes at a regional
 87 scale during the deglaciation.



88 **Fig. 1:** a) Modern oceanographic circulation in the Adriatic Sea, bathymetry of the area and location of the
 89 core ND14Q-AR2 (yellow dot). Black lines, surface circulation: **EAC** (East Adriatic Current), **WAC** (West
 90 Adriatic Current). Purple lines, middle water circulation: **LIW** (Levantine Intermediate Water). Red lines,
 91 deep water circulation: **NAdDW** (North Adriatic Deep Water), **SADW** (South Adriatic Deep Water). b)
 92 Lithological log of core ND14Q-AR2 from Totaro et al. (2022). The black line represents the portion of the
 93 core studied here.

94
 95
 96 **2. Study area**

97 *2.1. Oceanographic and climate setting*

98 The Adriatic Sea is a semi-closed elongated basin located between the Balkans and the east coast of the Italian
 99 Peninsula, connected to the Mediterranean Sea through the Strait of Otranto (Fig.1a). The Adriatic bathymetry

100 is characterized by a shallow northern basin, with an average depth of ca. 35 m, which increases in the southern
101 sector to an average depth of ca. 800 m (Ridente and Trincardi, 2002, 2006; Trincardi et al., 2007) and to about
102 1200 m in the deepest portion of the basin, the South Adriatic Pit (Artegiani et al., 1997). Surface water
103 circulation in the Adriatic Sea is characterized by the cyclonic circulation conditions of the South Adriatic
104 Gyre (SAG, Gačić et al., 1997), involving the East Adriatic Current (EAC) flowing northwards from the
105 eastern margin of the Strait of Otranto, along the Balkan coast, and descending as West Adriatic Current
106 (WAC), alongside the east Italian peninsula (Fig. 1a). The WAC, through freshwater discharges mostly from
107 the Po River and subordinately by small rivers along the Italian coast, delivers low-salinity water southwards,
108 together with suspended and dissolved substances enriching the basin with continental-derived nutrients
109 (Lipizer et al., 2014). Intermediate waters, flowing between 200 and 600 m, are composed by the Levantine
110 Intermediate Water (LIW) that flows in the Strait of Otranto from the Levantine Basin (Artegiani et al., 1997)
111 (Fig. 1a). The deepest water mass is represented by a dense water mass flowing southward from the northern
112 sector (Northern Adriatic Dense Water, NAdDW) sustained by Bora wind, blowing from the northwest during
113 winter (Gačić et al., 2002; Turchetto et al., 2007). The NAdDW combines with the Southern Adriatic Dense
114 Water (SADW) in the SAS (Fig. 1a), leading to one of the main source of the Mediterranean deep waters
115 (Lascaratos, 1993; Pinardi and Masetti, 2000; Manca et al., 2002).

116 With regard to the atmospheric circulation, the Adriatic Sea is situated between the subtropical high-pressure
117 zone and the mid-latitude westerlies belt (Orlić et al., 1992). Westerlies dominate most of the year, although
118 during summer, the subtropical high-pressure zone step over and the cyclonic and anticyclonic disturbances
119 of the westerlies belt almost disappear from the Adriatic (Orlić et al., 1992). These atmospheric influences,
120 together with winter Bora winds, result in a typically Mediterranean climate characterized by stormy and
121 relatively cold winter and hot and dry summer.

122

123 **3. Materials and Methods**

124 *3.1. Core ND14Q-AR2*

125 Core ND14Q-AR2, located in the southern Adriatic Sea (41°16' N, 17° 37' E), about 40 km from the Apulian
126 coast (Fig. 1a), was retrieved at 1013 m below sea level during the Next-Data 2014 expedition on board of the
127 Italian R/V Urania. The coring was possible thanks to the NextData project, a scientific collaboration among
128 several national research agencies, with the goal to acquire knowledge and know-how from climate archives
129 in the Italian territory (<http://www.nextdataprotect.it/>). The entire core is 540 cm long (Fig. 1b), while the
130 investigated section here extends from cm 285 to cm 505. Main lithology consists of gray/brown silty-clays,
131 with ten interbedded cryptotephra layers and one tephra (Fig. 1b). The adopted age model (Totaro et al., 2022)
132 relies on 10 radiocarbon (¹⁴C) AMS measurements and 6 tephra (cryptotephra and tephra) layers and was
133 constructed using the Bayesian model with the Bayesian statistic software Bacon with the statistical package
134 R (Blaauw and Christen, 2011).

135

136 *3.2. Coccolith assemblage analysis*

137 A total of 99 samples were prepared and analysed for quantification of coccolithophore assemblages at a
 138 sampling step of 2 cm over an interval ca. 220 cm long, with an average temporal resolution of ca. 80
 139 yrs/sample. Samples were prepared following Flores and Sierro (1997) in order to estimate both the relative
 140 and absolute coccolith abundances. Quantitative analyses were performed using a polarized light microscope
 141 at 1000× magnification. At least 500 specimens were counted per sample. Reworked calcareous nanofossils
 142 were counted separately in the investigated fields of view and their relative % abundances determined against
 143 the 500 non-reworked coccoliths. The absolute abundances of taxa (Total N) expressed as the number of
 144 coccoliths per gram of dry sediment (Ng⁻¹) were calculated with the following equation:

$$145 \quad \mathbf{N} = \mathbf{n} \times \mathbf{R}^2 \times \mathbf{V} \times \mathbf{r}^{-2} \times \mathbf{g}^{-1} \times \mathbf{v}^{-1}$$

146 where **N** is the absolute number of nanofossils per gram of dry sediment, **n** is the counted number of
 147 nanofossils in a random scanned area, **R** is the radius of the Petri dish, **V** is the volume of buffered water
 148 added to the weighted dry sediment, **r** is the radius of the visual field used in the counting, **g** is the gram of dry
 149 sediment, **v** is the volume of the pipetted sample (water and sediment) in the Petri dish.

150 Coccolith taxonomy mostly uses Young et al. (2003) and Jordan et al. (2004). In addition, the taxon *Emiliana*
 151 *huxleyi* was separated into two main morphotypes following size criteria: small *Emiliana* (SE) < 4µm and
 152 large *Emiliana* (LE) > 4µm (Colmenero-Hidalgo et al., 2002). Taxonomic subdivisions among
 153 geophyrocapsids follow Flores *et al.* (2000). The algebraic sum of *Calciosolenia* spp., *Discosphaera tubifera*,
 154 *Rhabdosphaera clavigera*, *Umbilicosphaera foliosa*, *Umbilicosphaera sibogae*, *Umbellosphaera* spp. and
 155 *Oolithotus* spp., i.e. taxa having preference for oligotrophic, tropical and sub-tropical waters (Winter and
 156 Siesser, 1994; Baumann et al., 2004; Boeckel and Baumann, 2004), is plotted as warm-water coccolith taxa
 157 (WWCT) and used as warm surface water proxy.

158

159 3.3. Biomarker analysis

160 Ninety-three samples were analysed for biomarkers at a sampling step of about 2 cm, with the exception of
 161 the bottom 10 cm where sampling step was increased to 4 cm. This resolution provides a temporal resolution
 162 of 1 sample every ca. 80 years, and 1 sample every ca. 160 years for the bottom 10 cm. A few grams of
 163 sediment were freeze-dried to extract lipids following Ternois et al. (2000). Both alkenones and n-alkanes
 164 were isolated from the total lipid extract by silica gel chromatography. A Varian CX 3400 gaschromatograph
 165 was used to quantify n-alkanes and alkenones after addition of 5 α-cholestane prior injection. Long-chain
 166 alkenones in open sea waters are known to be mainly produced by the coccolith taxon *Emiliana huxleyi*, the
 167 most abundant coccolithophore in the modern ocean (Paasche, 2002). The unsaturation index of C₃₇ alkenones
 168 (U₃₇^{k'}) and calibration of Conte et al. (2006) were used to derive SSTs using the following equation:

$$169 \quad T[^\circ\text{C}] = -0.957 + 54.293 [U_{37}^{k'}] - 52.894 [U_{37}^{k'}]^2 + 28.321 [U_{37}^{k'}]^3$$

170 High-molecular weight n-alkanes with an odd number of carbon atoms (Σ[C27]+[C29]+[C31]+[C33],
 171 hereafter named TERR-alkanes) produced by higher plants were used to estimate terrestrial inputs from the
 172 vegetation. These compounds are constituents of epicuticular wax of higher plants used by the vegetation to
 173 reduce water loss by evapotranspiration (Gagosian and Peltzer, 1986). The Average Chain Length (ACL) of

174 n-alkanes was calculated between C₂₇ to C₃₃ to assess past moisture conditions on land, based on the
175 assumption that longer chains are synthesized by plants in response to limited water availability (Gagosian and
176 Peltzer, 1986; Eglinton and Eglinton, 2008; Jalali et al., 2018)

177

178 3.4. Planktonic foraminiferal analysis

179 Planktonic foraminifera were identified in 101 samples at sampling step of 1 cm focused on the interval
180 between 19.6 and 13.7 ka BP the average temporal resolution is of ca. 50 yrs/sample. The samples were dried
181 in an oven at 50°C and washed with deionized water using a 63 µm mesh sieve. Quantitative analyses on
182 planktonic foraminifera assemblage were carried out on 125 µm mesh sieved fractions. About 200-300
183 specimens of planktonic foraminifera were counted and identified in each sample to have a significant statistic.
184 The taxonomic concepts and ecological inferences are based on Hemleben et al. (1989) and Pujol and
185 Vergnaud-Grazzini (1995). Fifteen species of planktonic foraminifera were identified. Some planktonic
186 species were grouped as follows: *Orbulina* spp. includes both *O. universa* and *O. suturalis*; *Globigerinoides*
187 *quadrilobatus* includes *G. trilobus* and *G. sacculifer*; *Globigerinoides ruber* includes *G. gomitulus*;
188 *Globigerina bulloides* includes *G. falconensis*; *Globigerinatella siphonifera* includes *G. calida*. Analyses
189 discriminated between left and right coiling of *Globorotalia truncatulinoides*, *G. inflata*, *G. scitula* and
190 *Neogloboquadrina pachyderma*. The data obtained were expressed in relative abundances (%).

191 A planktonic foraminiferal paleoclimate curve was constructed (Cita et al., 1977; Sanvoisin et al., 1993;
192 Capotondi et al., 2016) from the algebraic sum of warm-water species percentages (expressed as positive
193 values) and cold-water species percentages (expressed as negative values), featuring ecological preferences
194 and modern habitat characteristics as reported in Hemleben et al. (1989), Rohling et al. (1993), and Pujol and
195 Vergnaud-Grazzini (1995). Warm-water species include *G. ruber* (white and pink varieties), *G. quadrilobatus*,
196 *G. siphonifera* and *O. universa*. Cold-water species encompass *G. bulloides*, *Globigerinita glutinata*,
197 *Turborotalita quinqueloba*, *G. inflata*, *G. scitula*, *G. truncatulinoides* left coiled and *N. pachyderma* right
198 coiled. Here, we only show the abundance profile of *G. truncatulinoides* left coiled. Negative and positive
199 values of the curve correspond to the cold and warm surface waters, respectively.

200

201 3.5. X-ray Fluorescence (XRF)

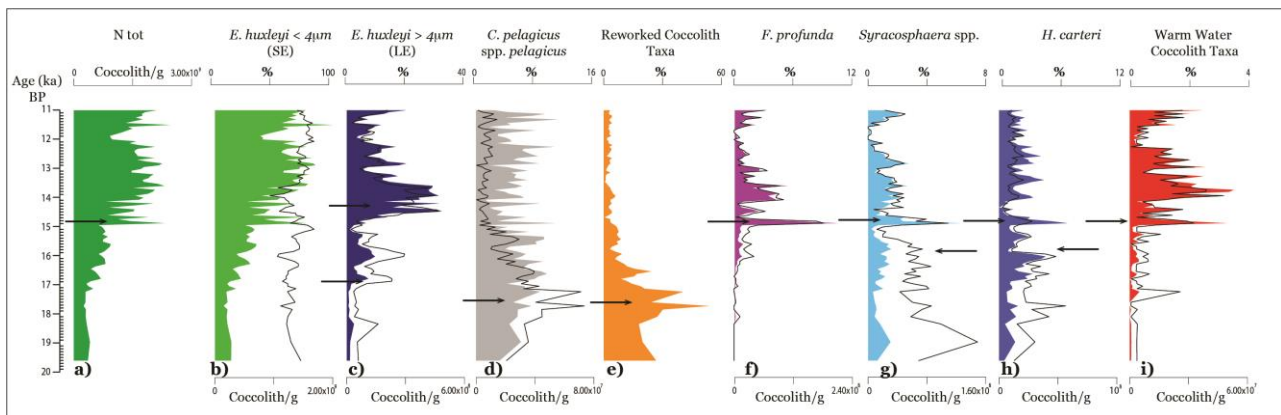
202 Bulk elemental geochemical composition of core ND14Q-AR2 was measured at the CORELAB laboratory of
203 the University of Barcelona with an Avaatech XRF core scanner. The sediment core was measured at 1 cm
204 resolution with excitation conditions of 10 kV, 0.5 mA and 10 s for major elements. For the scope of this paper
205 only the log-ratio of K and Al is shown. Log-ratio of element intensities has demonstrated to provide the most
206 easily interpretable signal of relative changes in chemical composition (Weltje and Tjallingii, 2008).

207

208 4. Results

209 4.1. Coccolithophore assemblage

210 Total N values range between 2×10^8 coccoliths/g and 2.7×10^9 coccoliths/g, with an average of 1×10^9
 211 coccoliths/g (Fig. 2a). In the lower part of the record the coccolith abundance is extremely low, while it
 212 gradually increases at ca. 17 ka BP and has highest values from 15 ka BP till the top of the record, with a brief
 213 decrease at 12 ka BP (Fig. 2a). SE represents the most abundant taxon, with percentages lying between 60 and
 214 80 % (Fig. 2b). The relative abundance of LE is generally not higher than 10% although pronounced increases
 215 occur at 16.8 ka BP, 15.9 ka BP, and between 14.4 and 13.6 ka BP (Fig. 2c). *Coccolithus pelagicus* ssp.
 216 *pelagicus*, generally representing <10% of the assemblage, shows relative abundances of about 15% during
 217 the 18-16 ka BP interval (Fig. 2d). Reworked coccolith taxa show a similar relative abundance pattern, with
 218 high values in the older part of the interval and declining at 18-16 ka BP (Fig. 2e). Although not a dominant
 219 component of the assemblages (never > 15 %), *Florisphaera profunda* exhibits very low percentages
 220 (generally < 3%) along the whole record, while rising with fluctuating abundances, between 14.8 and 13 ka
 221 BP (Fig. 2f). Among the less abundant taxa, *Syracosphaera* spp. (*Syracosphaera histrica* and *Syracosphaera*
 222 *pulchra*) and *Helicosphaera carteri* show a similar pattern of distinct decrease from higher relative abundances
 223 in the oldest part of the record (ca. 20-15.7 ka BP) to a decrease onward, with an isolated peaking values at ca.
 224 14.8 ka BP (Fig. 2g, h). The WWCT indicate very low abundances throughout the core, generally < 4 %, with
 225 increasing values between 14.8 and 12.3 ka BP, and from 11.5 ka BP up to the top of the studied interval (Fig.
 226 2i). Other taxa of minor abundances, not exceeding 3% of the assemblage, are not shown. They include
 227 *Coccolithus pelagicus* ssp. *braarudii*, *Coccolithus pelagicus* ssp. *azorinus*, *Braarudosphaera bigelowii*,
 228 *Calcidiscus leptoporus* ssp. *small* (3–5 μm), *C. leptoporus* ssp. *leptoporus* (5–8 μm), *C. leptoporus* ssp.
 229 *quadriperforatus* (8–10 μm), *Ceratolithus* spp., *Helicosphaera hyalina*, *Gephyrocapsa* spp., *Gladiolithus*
 230 *flabellatus*, *Umbilicosphaera hulburtiana*, *Pontosphaera* spp., *Scyphosphaera* spp.



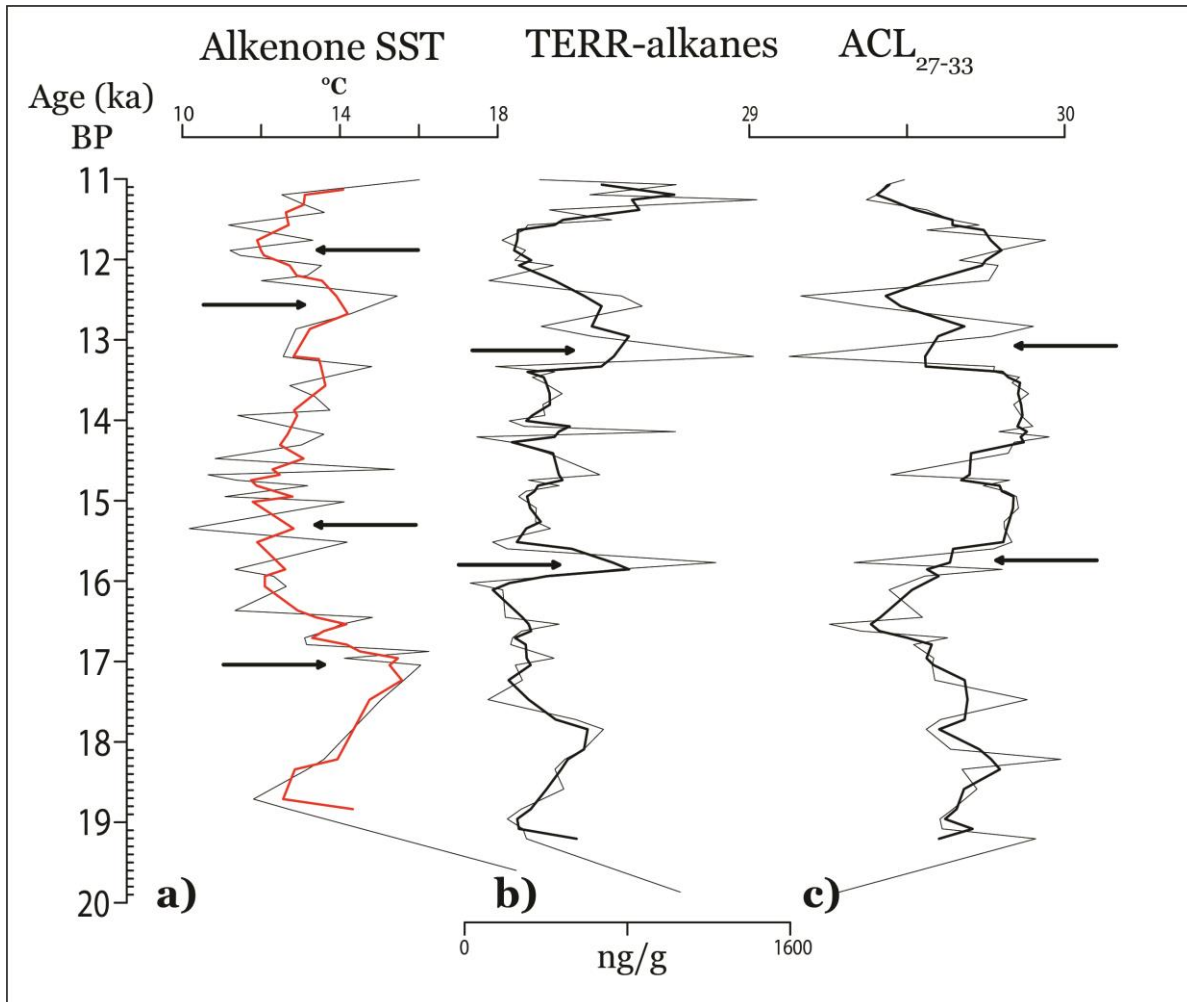
231
 232 **Fig. 2:** Coccolithophore assemblage variations from core ND14Q-AR2 (thick lines relative abundances, %;
 233 filled area absolute abundance, Coccolith/g). Reworked coccolith taxa are plotted as relative abundance (%).
 234 Total N: total coccolithophore abundance variations. Black arrows indicate significant increases/decreases in
 235 abundances as signalled in the text.

236

237 4.2. Biomarker data

238 The SST reconstruction reveals multidecadal variability over the 11-20 ka BP interval featuring rapid surface
 239 hydrological changes (Fig. 3a). Values between 20 and 17 ka BP, where temporal resolution is lower, range

240 between 19 and 12°C. Values decrease from 16 to 10 °C between 17 to 15.5 ka BP, to reach a minimum value
 241 of 10°C at 16 - 15 ka BP (Fig. 3a). A second cold interval (ca. 11°C) is also recorded between 12 and 11 ka
 242 BP (Fig. 3a). Between these cold periods (15 to 12.5 ka BP) SST gradually rises to nearly 16°C (Fig. 3a).
 243 TERR-alkane concentrations along the core ND14Q-AR2 do not show a clear trend but rapid shifts
 244 highlighting short-term terrestrial inputs to the site (Fig. 3b). ACL values show relatively high values during
 245 the entire interval (nearly 30). Lower ALC appeared to coincide with high TERR-alkanes (Fig. 3c).



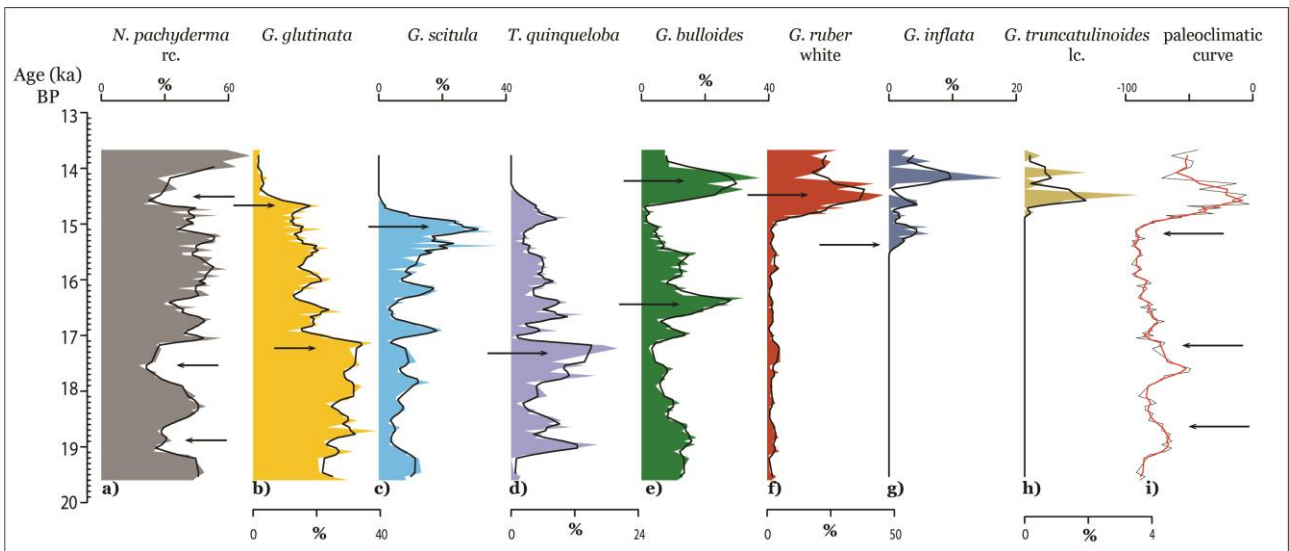
246
 247 **Fig. 3:** Variations through time of marine (a) and terrestrial (b, c) biomarkers from core ND14Q-AR2. Thick
 248 lines 3-points average. Black arrows indicate significant increases/decreases in abundances as signalled in the
 249 text.

250

251 4.3. Planktonic foraminiferal assemblage

252 Planktonic foraminifera are well preserved and diversified. Assemblages are mainly represented by *N.*
 253 *pachyderma* right coiled (rc) and characterized by three intervals of lower values between 19.01-18.6, 17.7-
 254 17.1 and 14.7-14.07 ka BP (Fig. 4a). *Globigerinita glutinata*, *G. scitula* and *T. quinqueloba* were identified till
 255 ca. 14.5 ka BP (Fig. 4b, c, d). *G. glutinata* abundance profile is characterized by two stepwise decreases at
 256 17.1 and 14.5 ka BP (Fig. 4b); *G. scitula* abundances fluctuate between 5 and 20% and reach maximum values
 257 between 15.42 and 14.8 ka BP before an important decrease (Fig. 4c). *Turborotalita quinqueloba* shows a

258 prominent peak at 17.2 ka BP and a progressive decrease with time opposed to the *G. scitula* abundances (Fig.
 259 4d). *Globigerina bulloides* is present throughout the studied section and shows two major peaks (ca. 30%) at
 260 14.1 ka and 16.2 ka BP (Fig. 4e). *Globigerinoides ruber* white shows a characteristic increase in abundance
 261 from ca. 3% to ca. 45% at 14.9 ka BP (Fig. 4f). *Globorotalia inflata* is only present in the uppermost part of
 262 the study section between 13.6 and 15.3 ka BP with abundance lying between 2 and 18% (Fig. 4g).
 263 *Neogloboquadrina dutertrei*, *Orbulina* spp., *G. siphonifera*, *G. rubescens*, *G. quadrilobatus*, *G. ruber* pink and
 264 *G. truncatulinoides* left coiled occasionally reach values between 3 to 6 %. The planktonic foraminiferal
 265 paleoclimatic curve shows a main shift from cold to warm conditions at ca. 15 ka while the older part shows
 266 relatively constant values except for two warmer intervals at 18.7-19.1 and 17.2-17.8 ka BP (Fig. 4i).



267 **Fig. 4:** Foraminifera assemblage relative abundance from core ND14Q-AR2 (%; thick line 3-points average),
 268 and foram-based paleoclimate curve (thick line 3 points average). Black arrows indicate significant
 269 increases/decreases in abundances as signalled in the text.
 270

271
 272 **5. Discussion**

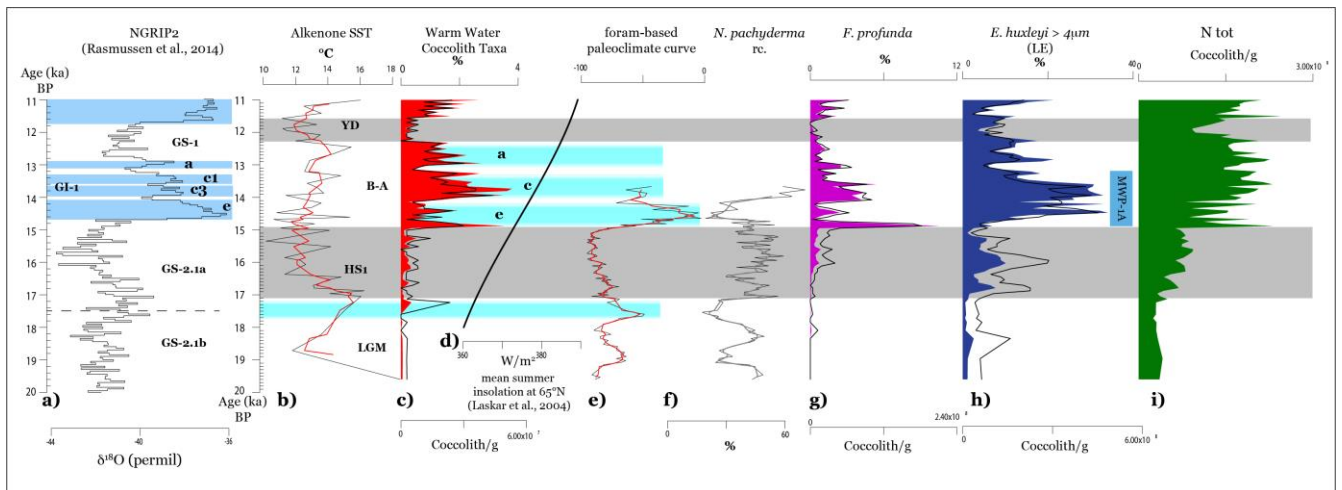
273 *5.1. Surface water conditions in the SAS*

274 *5.1.1. Last Glacial Maximum (LGM)*

275 Comparison between alkenone-derived SSTs and WWCT relative abundances through the record reveals
 276 several millennial-scale climate phases (Fig. 5b, c). Within the uncertainty of age models, well-known climate
 277 period from the Last Glacial Maximum to the Early Holocene period that are reflected in the Stadial and
 278 Interstadials of the Greenland ice core oxygen isotope record (Rasmussen et al., 2014) (Fig. 5a), match with
 279 SST changes as expected from ocean and atmospheric linkages with the North Atlantic region. They include
 280 the latest part of the Last Glacial Maximum (LGM, 20-17.1 ka), the Heinrich Stadial 1 (HS1, 17.1-14.9 ka),
 281 the Bølling-Allerød period (BA, 14.9-12.3 ka) and the Younger Dryas (YD 12.3-11 ka).

282 SSTs during the LGM indicate a marked cooling (Fig. 5b). The low temporal resolution of the SST signal
 283 during this time interval, due to insufficient amount of alkenones, is consistent with the low absolute
 284 abundances of *E. huxleyi* (Fig. 2b, c) pointing out non-favourable growth conditions. The WWCT signal and

285 planktonic forams paleoclimate curve provide a more detailed picture of the surface water conditions in the
 286 SAS during the LGM and notably the extremely low and negative values respectively, witnessing cold surface
 287 water conditions (Fig. 5c, e). A warm interval starts at ca. 17.6 ka BP according to the SSTs, WWCT and the
 288 planktonic forams paleoclimate curve, and reaches 16°C between 17.5-17.1 ka BP (Fig. 5b c, e). This episode
 289 seems to be coeval with lighter $\delta^{18}\text{O}$ in Greenland ice cores indicating warmer temperature in the earlier GS-
 290 2.1a (Fig. 5a). During the LGM, the high but decreasing abundances of reworked coccoliths (Fig. 6i) suggests
 291 enhanced erosion in the borderland area and continental detrital inputs to the core location (Flores et al., 1997;
 292 Bonomo et al., 2014, 2016; Maiorano et al., 2016). At that time of lower sea level, the northern Adriatic basin
 293 was almost completely subaerially exposed and the Po river mouth likely located closer to the ND14Q-AR2
 294 site (Correggiari et al., 1996; Vai and Cantelli, 2004). Synchronous low total coccolith abundances indicate
 295 that cold and turbid surface waters were not favourable to coccolith growth (Fig. 5i). Coeval increased *G.*
 296 *glutinata* and *T. quinqueloba* (Fig. 6h) seems to support the occurrence of land-derived nutrient in surface
 297 waters (Cita et al., 1977; Corselli et al., 2002; Geraga et al., 2008; Jonkers et al., 2010; Margaritelli et al.,
 298 2018).



299
 300 **Fig. 5:** Core ND14Q-AR2 records: a) Greenland $\delta^{18}\text{O}$ (Rasmussen et al., 2014); b) alkenone base SST (red
 301 line represents a 3-points running average); c) warm coccolith taxa (WWCT) distribution pattern (black line,
 302 relative abundances, %; colored area, absolute abundances, Coccolith/g); d) summer insolation curve (Laskar
 303 et al., 2004); e) foram-based paleoclimatic curve (thick line, 3 points average); f) *N. pachyderma* rc relative
 304 abundances (%; dark lines, 3-points average); g) *F. profunda* distribution pattern (black line, relative
 305 abundances, %; colored area, absolute abundances, Coccolith/g); h) *E. huxleyi* > 4 μm distribution pattern
 306 (black line, relative abundances, %; colored area, absolute abundances, Coccolith/g); i) total coccolithophore
 307 abundance variations. **LGM** (Last Glacial Maximum); **HS1** (Heinrich Stadial 1, grey bar); **B-A** (Bølling-
 308 Allerød); **YD** (Younger Dryas, grey bar); **GS** (Greenland Stadial and sub events 2.1a, b); **GI** (Greenland
 309 Interstadial and sub events: **a, c, e**); **MWP-1A** (Melt Water Pulse).

310

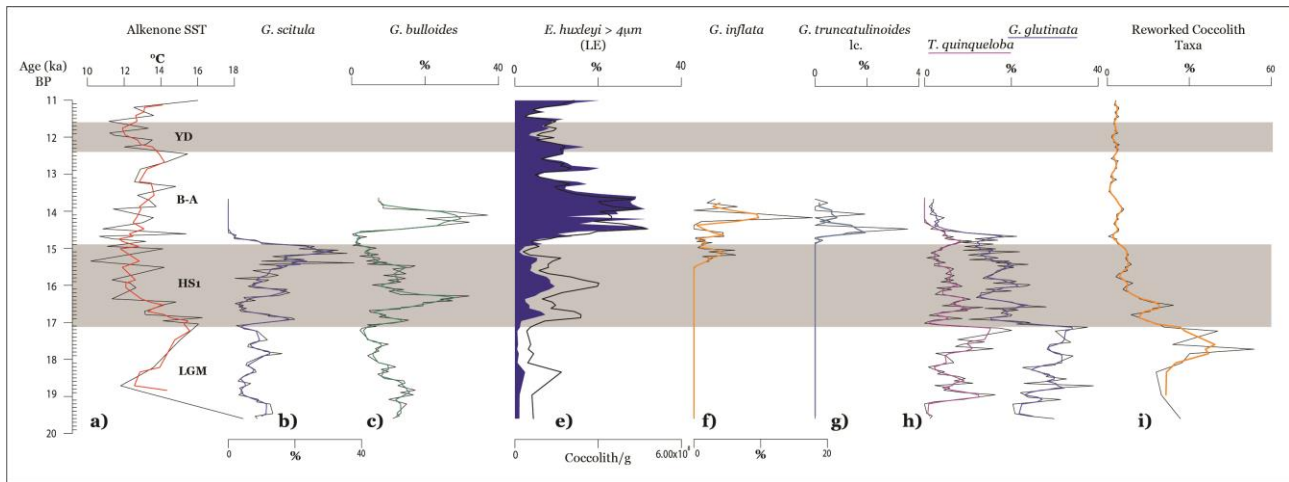
311

312 5.1.2. Heinrich Stadial 1 (HS1)

313 The SST decline at 17.1 ka (Fig. 5b) reflects the climate response of the Adriatic Sea to the onset of the HS1
314 event, in agreement with earlier studies in the Mediterranean (Cacho et al., 1999; Sierro et al., 2005; Kotthoff
315 et al., 2011). SST cooling during Heinrich events in the Mediterranean has been linked to cold water inflow
316 from the Atlantic Ocean (Cacho et al., 1999, 2001) combined with intensified cold winds in winter (Rohling
317 et al., 1998; Cacho et al., 2000). The HS1 is identified in our core between 17.1 and 14.9 ka by a $\sim 6^{\circ}\text{C}$ SST
318 drop to the lowest values of the record (10°C) at 16-15 ka (Fig. 5b). The WWCT and foraminiferal paleoclimate
319 curves do not show remarkable changes at the onset of HS1 (Fig. 5b) while conversely, HS1 cooling is clearly
320 manifested by an abrupt increase of % *N. pachyderma* rc (Fig. 5f), a progressive increase of polar-subpolar
321 taxon *G. scitula* (Ortiz et al., 1996; Schiebel et al., 2002) (Fig. 6b). The occurrence of *N. pachyderma* rc is
322 indicative of cold and eutrophic conditions (Hemleben, et al., 1989; Pujol and Vergnaud-Grazzini, 1995;
323 Rigual-Hernández et al., 2012). Increase abundances of opportunistic species *G. bulloides*, during this interval
324 (Fig. 6c), further support enhanced nutrient availability in the upper surface ocean, most likely promoted by
325 seasonal winter mixing and/or continental runoff (Cita et al., 1977; Schiebel et al., 2001; Corselli et al., 2002;
326 Geraga et al., 2008; Jonkers et al., 2010; Margaritelli et al., 2018, 2020). In the Adriatic Sea, several studies
327 reported the occurrence of *N. pachyderma* rc during the cold events of the last deglaciation (Asioli et al., 2001;
328 Trincardi et al., 2007; Favaretto et al., 2008; Piva et al., 2008; Siani et al., 2010; Narciso et al., 2012). This
329 taxon has also been found during HS1 in the Ionian Sea (Geraga et al., 2008), in the Sicily Channel (Sprovieri
330 et al., 2003) and in the North Aegean Sea (Kotthoff et al., 2011). Finally, cold-water species *G. inflata* (De
331 Castro Coppa et al., 1980; Pujol and Vergnaud-Grazzini, 1995) combined to highest values of *G. scitula* (Ortiz
332 et al., 1996; Schiebel et al., 2002) between 15.3 and 14.9 ka (Fig. 6f, b) points to the coldest phase of HS1
333 (Fig. 6a).

334 During HS1, coccolith assemblages are also characterized by increased LE abundances (Fig. 5h). This
335 morphotype is absent in present day assemblages with the exception of cold ocean waters in the Arctic and
336 Subarctic regions (Hagino et al., 2005; Flores et al., 2010). Its presence in the western Mediterranean basin
337 during the late Pleistocene and Holocene has been used to capture cold/fresher surface water masses during
338 HS events (Colmenero-Hidalgo et al., 2002, 2004; Flores et al., 2010; Ausín et al., 2015; Di Stefano et al.,
339 2015; Bazzicalupo et al., 2018). The LE morphotype has never been documented before in the central
340 Mediterranean nor in the Adriatic Sea, where its ecological behaviour and its potential value as
341 ecostratigraphical pattern is unknown. Its occurrence during HS1 in our core supports the signature of cold-
342 fresher surface water most probably linked to meltwater from local continental ice and their delivery by the Po
343 River. Highest relative abundances of this taxon are also recorded in younger sediment layers of the core and
344 will be discussed in the following sections.

345 The HS1, as identified at ND14Q-AR2, matches well with the cold portion of Greenland Ice counterpart GS-
346 2.1a (Fig. 5a). The SST decrease in the later phase of the HS1 is in good agreement with the extreme cooling
347 in the eastern Mediterranean and Red Sea (Arz et al., 2003; Castañeda et al., 2010, 2016).



348
 349 **Fig. 6:** Core ND14Q-AR2 records: a) alkenone base SST (red line represents a 3-points running average); b,
 350 c, f, g, h) foraminifera taxon distribution patterns (%; dark lines, 3-points average); e) *E. huxleyi* > 4 μm
 351 distribution pattern (black line, relative abundances, %; colored area, absolute abundances, Coccolith/g); i)
 352 reworked coccolith taxa distribution pattern (black line, relative abundances, %; orange line, 3-points average).
 353 **LGM** (Last Glacial Maximum); **HS1** (Heinrich Stadial 1, grey bar); **B-A** (Bølling-Allerød); **YD** (Younger
 354 Dryas, grey bar).

355

356 5.1.3. The Bølling-Allerød (B-A)

357 The B-A interstadial spans between 14.9 and 12.3 ka BP, as revealed by the general increase of WWCT and
 358 the gradual alkenone-SST rise from 11 to 15.5 °C (Fig. 5b), as well as by a positive shift of the foraminifera
 359 paleclimatic curve (Fig. 5e). During the B-A event, the WWCT exhibit three centennial-scale oscillations that
 360 are partially registered by planktonic foraminiferal curve as well (Fig. 5c, e). These three warmings are nearly
 361 contemporaneous to the succession of warm intervals known as GI-1e, c1-3, a (Fig. 5a) (Rasmussen et al.,
 362 2014). These high frequency fluctuations were also previously observed in other planktonic foraminiferal
 363 assemblage records from the SAS (Asioli et al., 2001; Siani et al., 2010).

364 During the B-A we observe an increase in the abundance of the deep dwelling taxon *F. profunda* (Fig. 5g),
 365 pointing to seasonal stratification and deep nutricline/thermocline conditions (Molfinio and McIntyre, 1990;
 366 Beaufort et al., 1997; Di Stefano and Incarbona, 2004; Balestra et al., 2008) likely related to gradual increase
 367 of insolation and related surface water warming (Fig. 5d) (Beaufort et al., 1997, 2001). Indeed, the increase of
 368 Total N (Fig. 5i) suggests ameliorated surface water conditions for the coccolithophore assemblage i.e.,
 369 warmer surface waters and seasonal nutrients availability, likely related to seasonal mixing. The strong
 370 increase of *G. bulloides* (Fig. 6c) together with the peaks in abundance of *G. truncatulinoides* and of *G. inflata*
 371 (Fig. 6f, g), suggest a deep vertical mixing during the winter season in the study sites at the onset of the B-A,
 372 with a strong advection of nutrients from the deeper layers to the surface (Schiebel et al., 2001; Margaritelli et
 373 al., 2020).

374 LE occurrences rising between 14.7 and 13.4 ka BP is also observed. As mentioned before the distribution of
 375 this taxon is not known in the central Mediterranean but well documented in the western Mediterranean
 376 throughout this interval and has been related to the inflow of colder and fresher Atlantic water in the Alboran

377 Sea, likely deriving from iceberg melting (Colmenero-Hidalgo et al., 2002, 2004; Flores et al., 2010; Ausín et
378 al., 2015; Bazzicalupo et al., 2018). Given its ecological preference, the LE increase in our record seems to
379 point out to cold freshwater outburst in the SAS through the Po River due to Alpine and Apennine ice sheets
380 melting resulted from ameliorated climate in the North Atlantic (as revealed by Greenland ice). Indeed
381 previous data provide evidence of ice melting in the Alps mountains surrounding the Adriatic basin and
382 subsequent increased run off of the Po River (Combourieu-Nebout et al., 1998; Asioli et al., 2001). This
383 hydrological episode recorded in the SAS would reflect the more global ice melt event known as the MWP-
384 1A (Bard et al., 1996; Fairbanks et al., 2005) (Fig. 5h), occurring around 14 ka BP (Liu and Milliman, 2004;
385 Stanford et al., 2006). Altogether, our results suggest that the LE may be a valuable indicator of meltwater
386 episodes into the SAS.

387 A closer look of the centennial scale variability shows that the major peak of *G. bulloides* and *G. inflata*
388 occurring between 14.5 and 14.0 ka BP coincides with low relative abundances of LE (Fig. 6e, f). This opposite
389 behaviour between taxa might feature decadal-scale shifts from freshwater driven stratified conditions
390 (increase of LE) to cold and strong winds allowing for strong water column mixing and higher surface water
391 nutrients (increase of *G. inflata* and *G. bulloides*).

392

393 5.1.4. Younger Dryas (YD)

394 Between 12.3 and 11.6 ka BP, alkenone SSTs show an abrupt cooling from 15.5 °C to values oscillating
395 between 11-13°C (Fig. 5b) evidencing the return to cold water conditions of the YD period in the
396 Mediterranean Sea (Cacho et al., 1999; Asioli et al., 2001; Di Stefano and Incarbona, 2004; Siani et al., 2010;
397 Kotthoff et al., 2011; Sicre et al., 2013; Martrat et al., 2014). The coeval low relative abundances of WWCT
398 are consistent with surface water cooling at the time of the GS-1 in Greenland ice (Fig. 5a).

399

400 5.2. Terrestrial inputs to the SAS

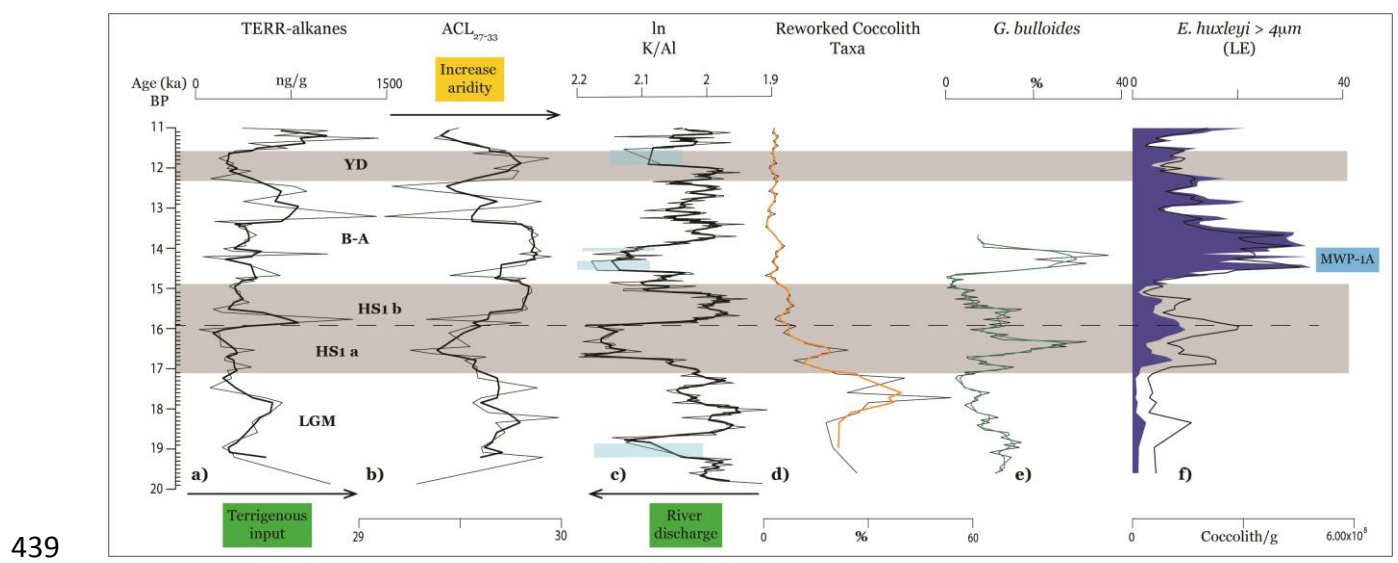
401 5.2.1. Terrestrial inputs during the HSI

402 Climate variability on land can be derived from ACL and TERR-alkanes content of the sediment. These
403 biomarker proxies produced by higher plants allow to diagnose moisture conditions (Gagosian and Peltzer,
404 1986; Sicre and Peltzer, 2004) and have been successfully used for paleoclimate reconstructions in the
405 Mediterranean (Jalali et al., 2016, 2017, 2018). K/Al ratio provides a complementary information on detrital
406 riverine discharge (Frigola et al., 2008; Nieto-Moreno et al., 2011; Rodrigo-Gámiz et al., 2011; Martinez-Ruiz
407 et al., 2015) Indeed, the northern Mediterranean regions are a major source of illite. River-derived increase of
408 illite supply leads to higher K/Al ratios in the sediments during humid intervals (Martinez-Ruiz et al., 2015).
409 The relatively high ACL values (Fig. 7b) suggest prevalent arid conditions on adjacent land and catchment
410 basins, in agreement with the dominance of desert and semidesert vegetation type during the deglaciation in
411 the central Mediterranean (Combourieu-Nebout et al., 1998; Magri and Sadori, 1999; Desprat et al., 2013).
412 This general trend is interrupted by short humid phases as indicated by rapid ACL decrease (Fig. 7b). These

413 brief periods are generally accompanied by peaking TERR-alkanes, that might suggest a link with increased
414 river run-off (Fig. 7a).

415 Of particular interest is a two-phase structure of HS1 portrayed by HS1a (17.1-15.9 ka BP) and HS1b (15.9-
416 14.9 ka BP) (Fig. 7) not previously identified in the SAS. Low ACL values during HS1a suggest relatively
417 humid conditions, while concurrent low TERR-alkanes are indicative of reduced higher plant material export
418 to the SAS (Fig. 7d). Concomitant high values of K/Al ratio in the sediment point out to enhanced riverine
419 discharge under relatively wet conditions (Fig. 7c). Abundance peak of reworked coccolith taxa during HS1a
420 seems to support enhanced detrital inputs in the SAS (Fig. 7d). Altogether these data suggest enhanced delivery
421 of detrital inputs with a lower contribution of higher plant with lower ACL, likely from an Alpine source. In
422 addition, the herbivorous opportunistic planktonic foraminifera *G. bulloides*, often associated with high
423 continental runoff (Pujol and Vergnaud-Grazzini, 1995; Rohling et al., 1997; Schiebel et al., 1997; Thiede et
424 al., 1997; Capotondi et al., 2004; Schmidt et al., 2004; Principato et al., 2006) is also abundant (Fig. 7e)
425 confirming enhanced continental input during HS1a. In contrast, during HS1b, ACL pattern shifts to higher
426 values suggesting reduction in moisture availability on land (Fig. 7b). The low K/Al ratio during HS1b well
427 supports reduced river input (Fig. 7c) as well as low TERR-alkanes (Fig. 7a).

428 Between HS1a and HS1b, peaking TERR-alkanes and low ACL (15.8 ka BP, Fig. 7a) suggests enhanced
429 transport of higher plant derived material to the SAS, grown under humid conditions. The concomitant increase
430 of LE abundances suggests a link with the supply of cold fresher waters (Fig. 7f). An almost identical event in
431 the calcareous nannofossil assemblage has been identified in the western Mediterranean (Bazzicalupo et al.,
432 2018) in the middle of the HS1, during a phase of increased melt water in the North Atlantic and/or in the
433 Mediterranean via the Rhone River caused by Eurasian/Alps Ice Sheet complex melting (Cacho et al., 1999;
434 Sierro et al., 2005; Frigola et al., 2008; Melki, 2011). The brief fresh water supply seen at our core location
435 between HS1a and HS1b can thus reasonably be assigned to Alpine and Apennine melting ice (Asioli et al.,
436 2001; Storms et al., 2008). The imprint of the short-lived episode in the middle of HS1 is also seen in the
437 western Mediterranean (Martrat et al., 2014; Hodell et al., 2017) and in the Gulf of Cadiz (Sierro et al., 2020)
438 showing lighter seawater $\delta^{18}\text{O}$ and warmer SST.



439

440 **Fig. 7:** Core ND14Q-AR2 records: a, b) terrestrial biomarker (TERR-alkanes and ACL₂₇₋₃₃; thick lines 3-points
441 average); c) K/Al log ratio (thick line 3-point average, azure areas represents tephra layers levels removed
442 from the figure to facilitate the observation); d) reworked coccolith taxa distribution pattern (black line, relative
443 abundances, %; orange line, 3-points average); e) *G. bulloides* distribution pattern (%; dark lines, 3-points
444 average); f) *E. huxleyi* > 4 μm distribution pattern (black line, relative abundances, %; colored area, absolute
445 abundances, Coccolith/g). **LGM** (Last Glacial Maximum); **HS1** (Heinrich Stadial 1, grey bar); **B-A** (Bølling-
446 Allerød); **YD** (Younger Dryas, grey bar); **MWP-1A** (Melt Water Pulse).

447

448 5.2.2. Terrestrial inputs during the B-A and YD.

449 During the earlier phase of B-A (14.9-13.4 ka BP), ACL values were generally lower (Fig. 7b) whereas TERR-
450 alkanes show fluctuating values, indicating drier conditions on land. This finding is in apparent disagreement
451 with the general indication of wetter climate conditions on land during the B-A, in contrast to HS1 and YD,
452 based on central Mediterranean pollen records (Combourieu-Nebout et al., 1998; Magri and Sadori, 1999;
453 Asioli et al., 2001; Desprat et al., 2013). However, it seems consistent with evidence of relatively warm climate
454 and limited water availability, as suggested by dominant semi-desert plant remains in the central Mediterranean
455 (Pollen Zone PZ2 between 13.7 and 13.1 ka BP, Desprat et al., 2013). Relatively high concentrations of the
456 steppe taxon *Artemisia* (up to 30%) during the early stage of the B-A are also reported in the SAS (labelled as
457 GI-1 e-d in Asioli et al., 2001).

458 The most recent part of B-A (13.4-12.3 ka BP) is characterized by two TERR-alkane increases concomitant
459 with low ACL-values suggesting increased moisture availability (Fig. 7b). This increased humid local
460 condition is well known throughout the Mediterranean as a major attribute of climate ameliorating in the basin
461 (Combourieu-Nebout et al., 1998; Magri and Sadori, 1999; Combourieu Nebout et al., 2009; Fletcher et al.,
462 2010; Desprat et al., 2013). The B-A interval ends with an increase of ACL indicating drier conditions again
463 supported by evidence of prevailing dry conditions in the Mediterranean during the YD (Combourieu-Nebout
464 et al., 1998; Fletcher and Sánchez Goñi, 2008; Combourieu Nebout et al., 2009; Fletcher et al., 2010; Desprat
465 et al., 2013).

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467 5.3. The Two-phase structure of the HS1

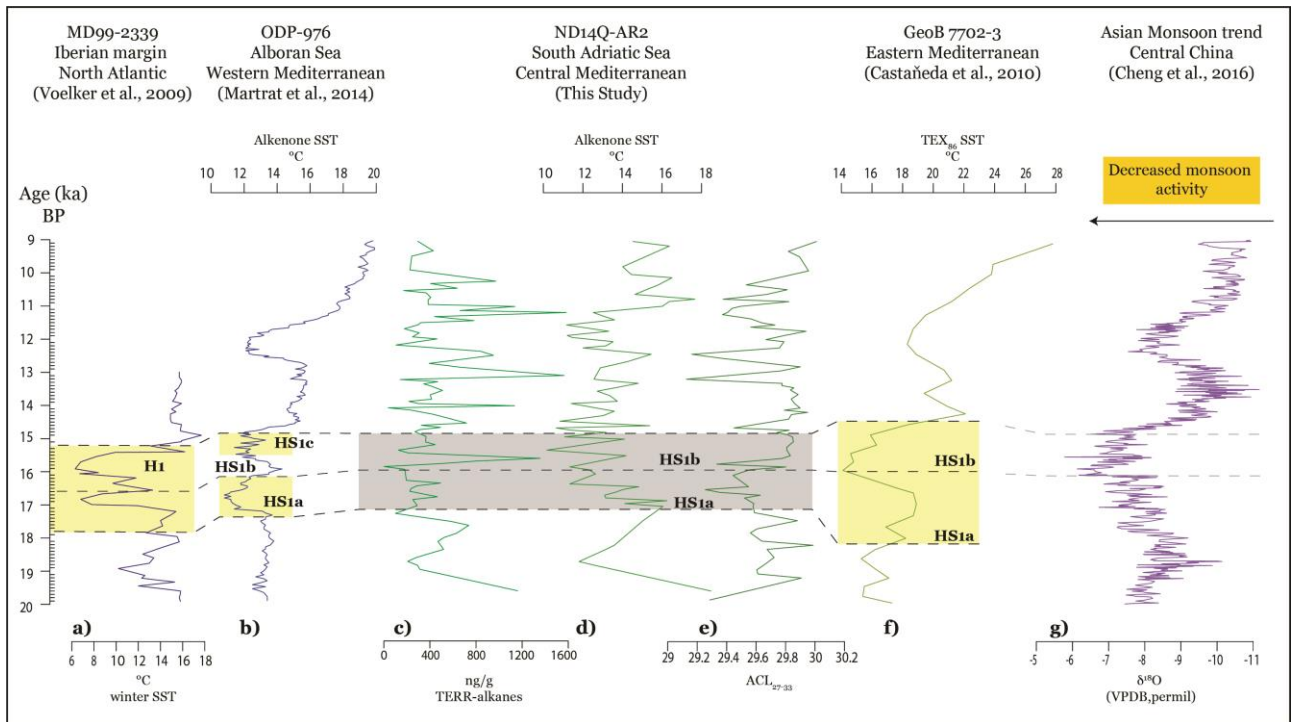
468 Our results in the Adriatic Sea reinforce the two-fold structure of HS1 found in many records around the world
469 (Huang et al., 2019 and references therein), yet the forcing and physical mechanisms at play are still debated.
470 Currently, in the central Mediterranean the detailed structure of HS1 is poorly documented and few records
471 reported on prolonged aridity in the region across the interval embracing HS1 (Combourieu-Nebout et al.,
472 1998; Asioli et al., 2001; Desprat et al., 2013), although some authors did not detail different phases within
473 the HS1.

474 The distinct two-phase structure in our core, however, are mainly fingerprinted by terrestrial proxies that
475 appear to reflect freshwater delivery in the SAS due to ice melting from the Alps mountains. A similar pattern,
476 within the limit of comparing different age models, was also observed in the Alboran Sea (Fig. 8b)

477 (Bazzicalupo et al., 2018) and in the north-western Iberia margin (Fig. 8a) (Naughton et al., 2009, 2016;
478 Voelker et al., 2009) during the HS1a, although well recorded in the marine proxies as well, similarly related
479 to freshwater arrival. A shift in the European moisture source distribution was also inferred during HS1 based
480 on evidences in the Black Sea (Kwiecien et al., 2009) and in the Aegean Sea (Kotthoff et al., 2011).

481 The SAS is under the influence of both mid-latitude westerlies wind belt and the subtropical high-pressure
482 zone (Orlić et al., 1992). Moister conditions during the HS1a are likely related to the strength and latitudinal
483 position of the westerly winds, with their southern position leading to relatively wetter conditions in southern
484 Europe and in the Mediterranean region.

485 At the onset HS1 cold and fresh water from ice sheet melting was released in the North Atlantic, affecting the
486 surface ocean circulation (Bond et al., 1992; Stanford et al., 2011). Subsequent collapse of the AMOC,
487 diminished transport of warm and moist from low to high latitudes (Rahmstorf, 1994). This alteration of the
488 AMOC resulted in North Atlantic/western Europe cooling (e.g. Seidov and Maslin, 1999; McManus et al.,
489 2004). The associated southward displacement of the ITCZ and moisture transport to mid-latitudes resembles
490 the negative NAO state (Trigo et al., 2004). Variability in precipitation and flood activity in the Alpine region
491 and the northern catchment basin of the SAS have been related to the NAO variability (Bartolini et al., 2009;
492 Wirth et al., 2013). In the eastern Mediterranean, Castañeda et al. (2016) (Fig. 8f) describes a two-phase HS1
493 nearly synchronous with core ND14Q-AR2: HS1a (c.a. 17.5-16 ka BP) characterized by relatively warmer
494 SSTs and HS1b (16-14.5 ka BP) corresponding to colder and driest conditions on land, which is in good
495 agreement with the HS1 interval in our core (Fig 8e, f). Huang et al. (2019) hypothesized a southward migration
496 of the ITCZ reaching its southernmost position during the HS1b in agreement with climate modelling
497 highlighting the southward shift of the ITCZ as the AMOC weakened (Kageyama et al., 2009). During HS1b,
498 colder conditions in the Adriatic Sea are consistent with those of the easternmost Mediterranean (Fig. 8f)
499 (Essallami et al., 2007; Castañeda et al., 2010, 2016) and with the weakest Asian monsoon (Fig. 8g) (Cheng
500 et al., 2016; Huang et al., 2019) pointing to a linkage with the tropical Northern Hemisphere climate. Colder
501 SSTs likely caused a further weakening of the AMOC (Huang et al., 2019) leading a southernmost position of
502 the ITCZ (Deplazes et al., 2013; Zhang et al., 2014) explaining drier conditions in the northern-low to middle
503 latitudes during the HS1b (Huang et al., 2019). The comparison between the Adriatic Sea, eastern
504 Mediterranean and Southern-central Asian domain (Fig. 8) provides evidence of a mid- and low-latitudes
505 reorganization of the atmospheric circulation during HS1 associated with the AMOC weakening.



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Fig. 8: Comparison of climate records across the Last Deglaciation in the Northern Hemisphere. a) Forams-based MAT winter SST curve from Core MD99-2339, Iberian Margin, North Atlantic (NA) (Voelker et al., 2009); b) Alkenone SST curve from ODP Core 976, Alboran Sea, Western Mediterranean (WM) (Martrat et al., 2014); c) Terrestrial TERR-alkanes curve, d) Alkenone SST curve and e) ACL₂₇₋₃₃ curve, from Core ND14Q-AR2, South Adriatic Sea (This study); f) TEX₈₆ SST curve from Core GeoB 7702-3, Eastern Mediterranean (Castañeda et al., 2010); g) Composite $\delta^{18}\text{O}$ Asian Monsoon record, Central China (Cheng et al., 2016). Yellow shaded areas indicate HS1 interval as indicated by the various cited authors: a) **H1** (Voelker et al., 2009); b) **HS1a, b, c** (as indicated in Bazzicalupo *et al.*, 2018); f) **HS1a, b** (Castaneda et al., 2016). Grey shaded area indicates the **HS1** division in the ND14Q-AR2 core (see text). Black Dashed line correlate HS1 divisions among climate records; grey dashed line is a tentative correlation between lowest monsoon activity registered in Asia (Cheng et al., 2016) and **HS1b** in our core (as indicated in the text).

6. Conclusions

In this work we present the first multiproxy record from the SAS that combines terrestrial and marine biomarkers with calcareous plankton assemblages in order to investigate centennial timescale climate variability during the last deglaciation. Alkenone SSTs, coccolithophore and planktonic foraminiferal assemblages along core ND14Q-AR2 evidence the influence of cold and fresher waters from melting continental alpine ice on the hydrology of the SAS that are consistent with Greenland ice core and SST reconstruction from other Mediterranean records. Terrestrial biomarkers suggest prevalent arid conditions on adjacent land and catchment basins interrupted by short humid phases accompanied by enhanced river run-off. Most importantly, our high-resolution reconstruction provides a detailed description of the HS1 (17-14.9 ka) that identifies a two-fold structure very similar to other records from the Mediterranean and other locations of the Northern Hemisphere. This feature is particularly expressed in the terrestrial proxy record highlighting the

530 role of atmospheric influence on the SAS, reflecting interactions with the cryosphere, e.g., ice melting from
531 the Alps and delivery by the Po River. Shifts of the ITCZ associated to changes of the AMOC strength seems
532 to be driving centennial scale changes during HS1, as further supported by the good correspondence of
533 Mediterranean records with those of Central Asia indicating decreased monsoon activity. A brief event of
534 Alpine and Apennine ice melting seems to have interrupted the humid/arid phases of HS1, highlighted by
535 increased LE abundances. Finally, our results demonstrate that LE might be a valuable indicator of cold and
536 low salinity waters from Alpine and Apennine glaciers melting and therefore a potential ecostratigraphical
537 marker in the SAS during the deglaciation.

538

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547

548 *Authors contributions*

549 Pietro Bazzicalupo conceived the idea of the study with the support of Patrizia Maiorano and Marie-
550 Alexandrine Sicre. Pietro Bazzicalupo carried out calcareous nannofossil analyses and interpretation with
551 supervision of Patrizia Maiorano, and biomarker investigations and interpretation under the supervision of
552 Marie-Alexandrine Sicre and Vincent Klein. Helena Checa carried out foraminiferal analysis and interpretation
553 with supervision of Giulia Margaritelli and Leopoldo David Pena. Jamie Frigola carried out XRF core scanner
554 analysis and interpretation. Pietro Bazzicalupo wrote the manuscript and developed the interpretative
555 framework of the work with the support of Fabrizio Lirer, Patrizia Maiorano, Giulia Margaritelli and Marie-
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557 critical revisions of the manuscript.

558 All the authors contributed to the discussions of the results.

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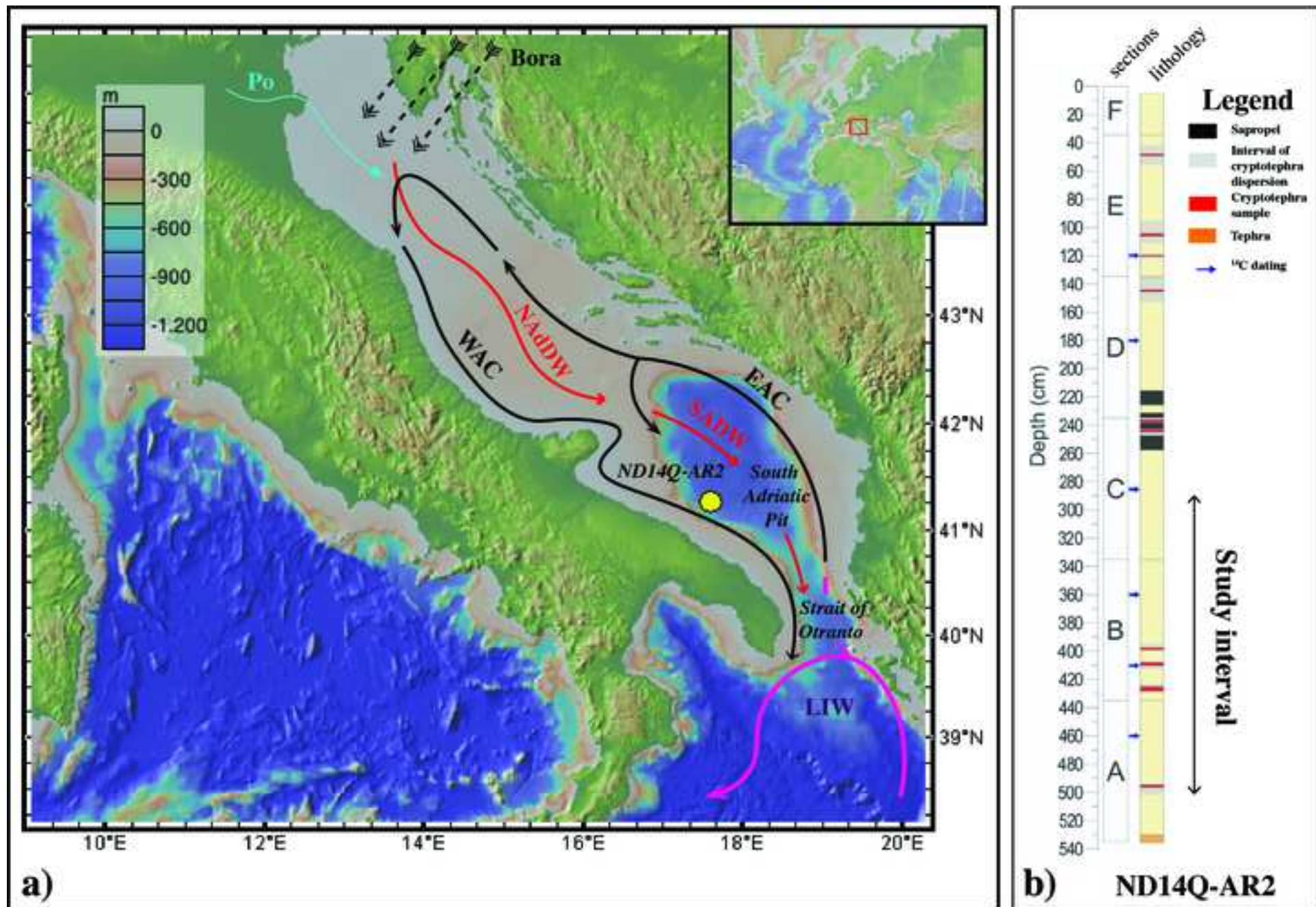
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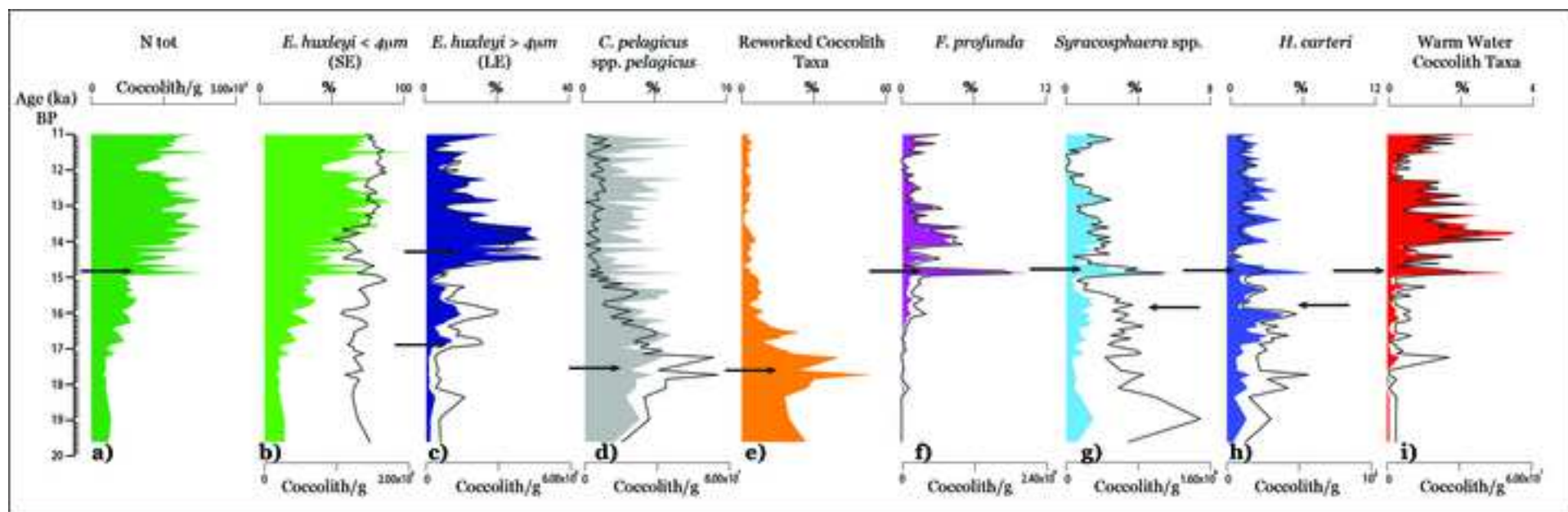
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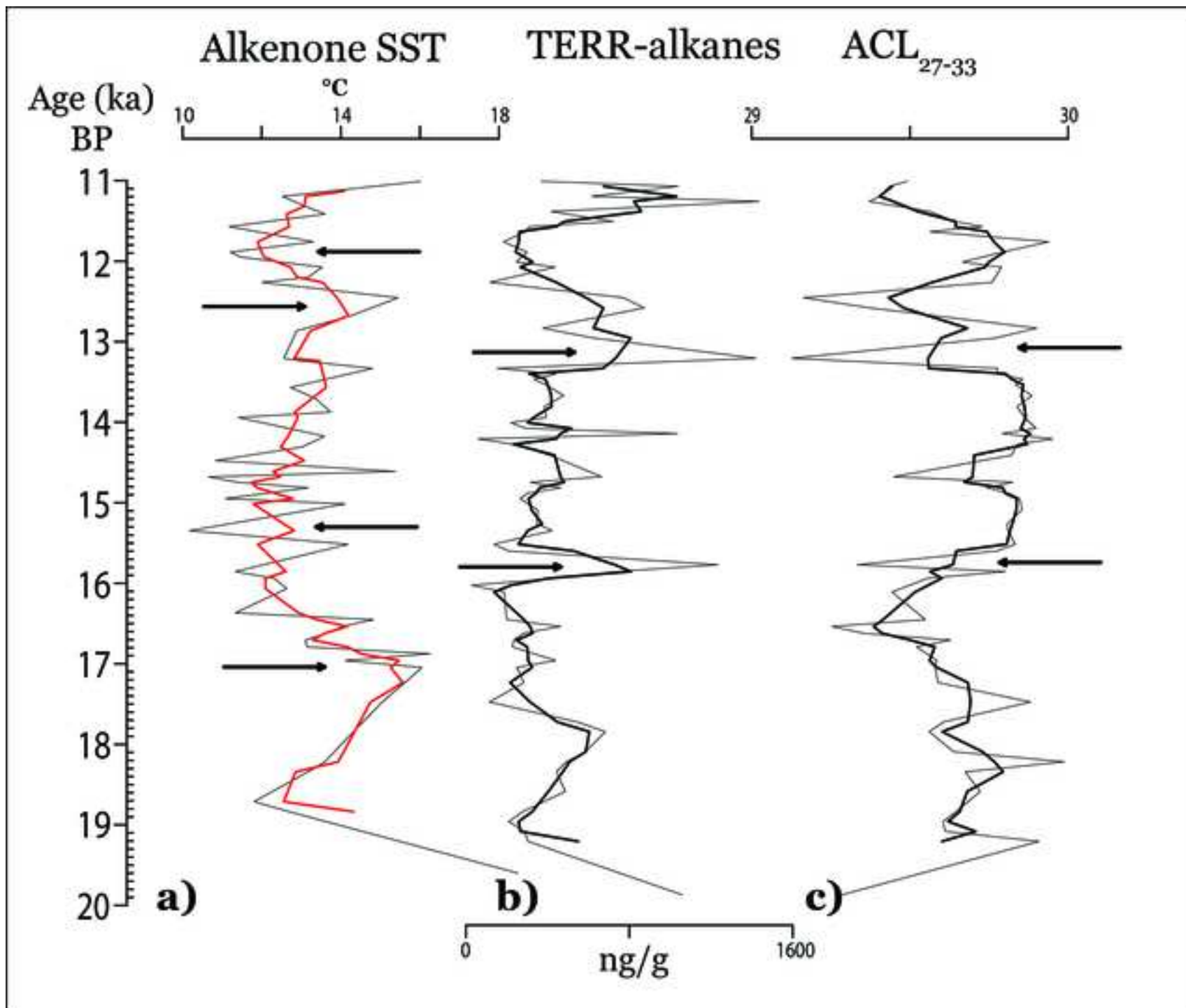
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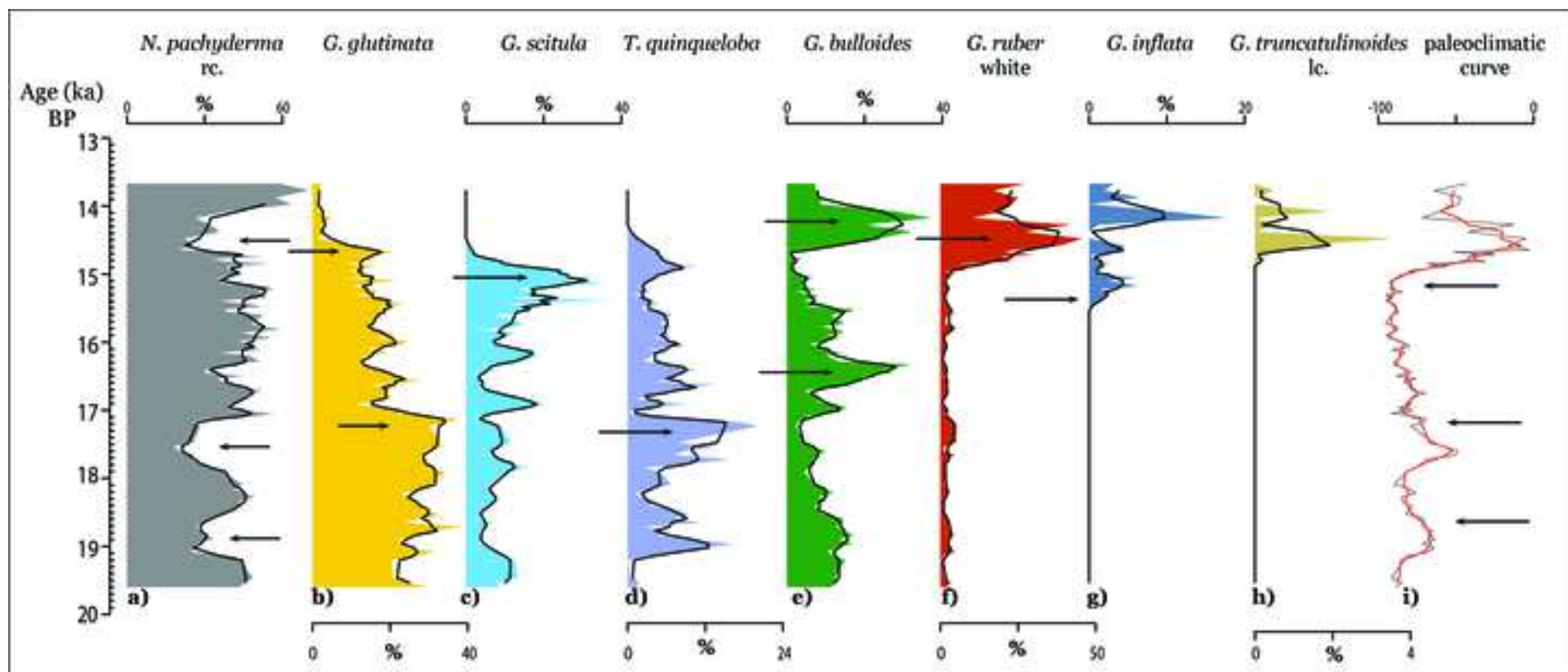
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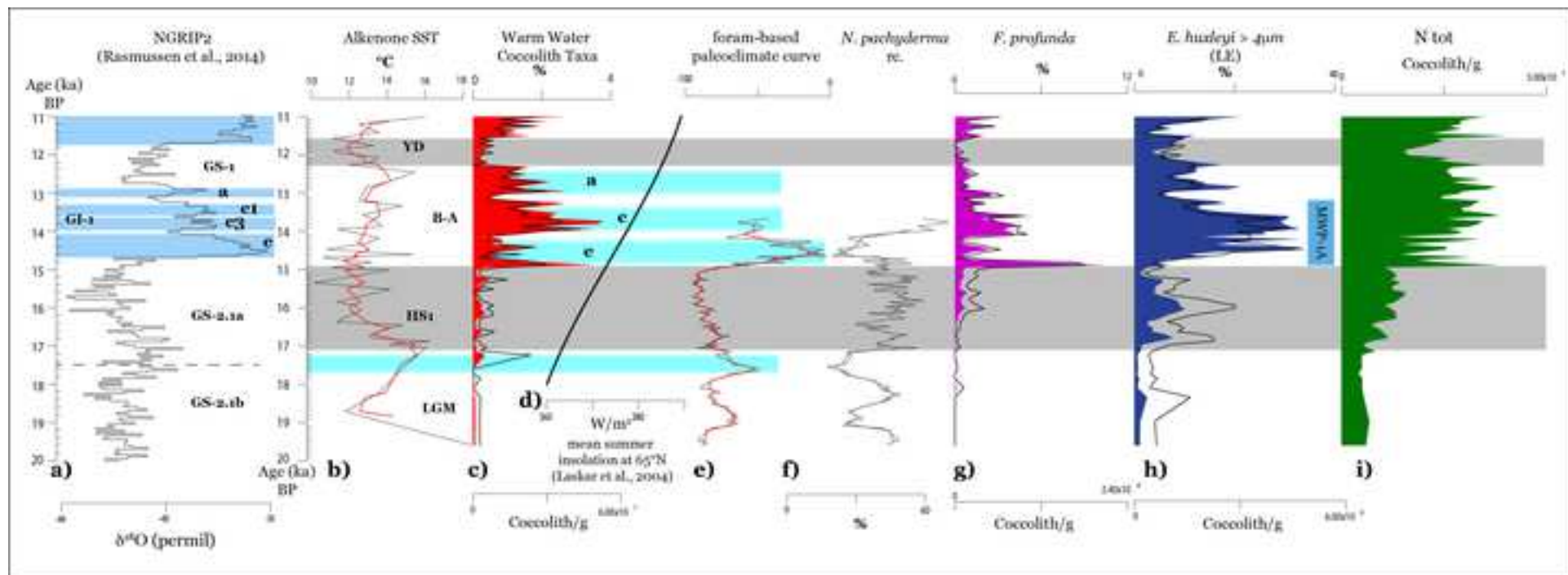
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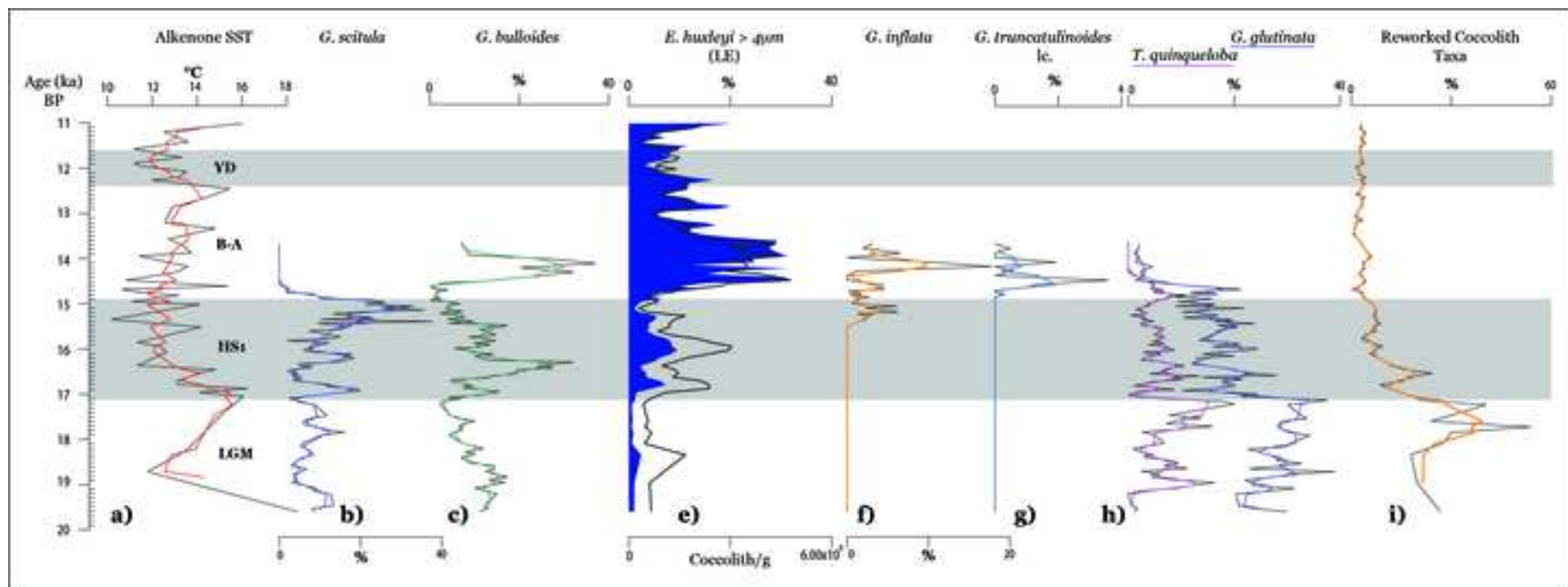


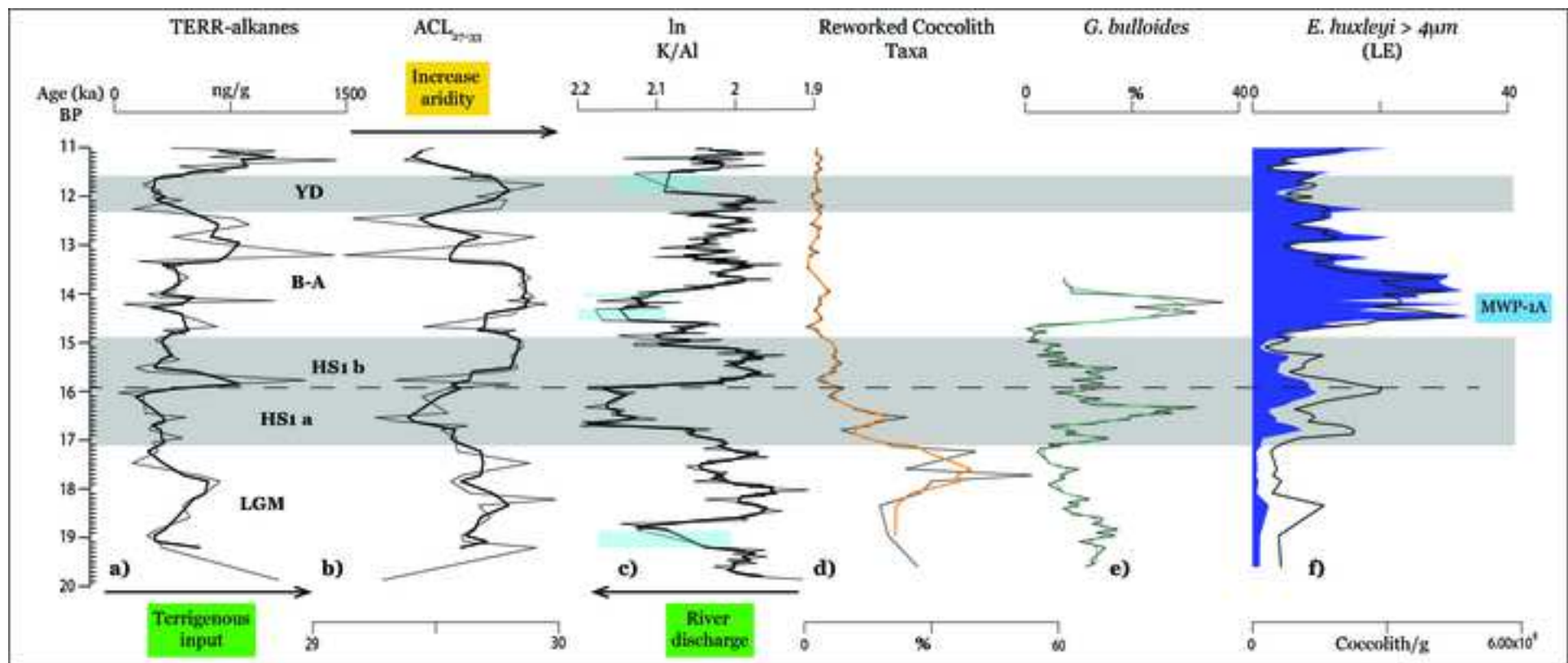


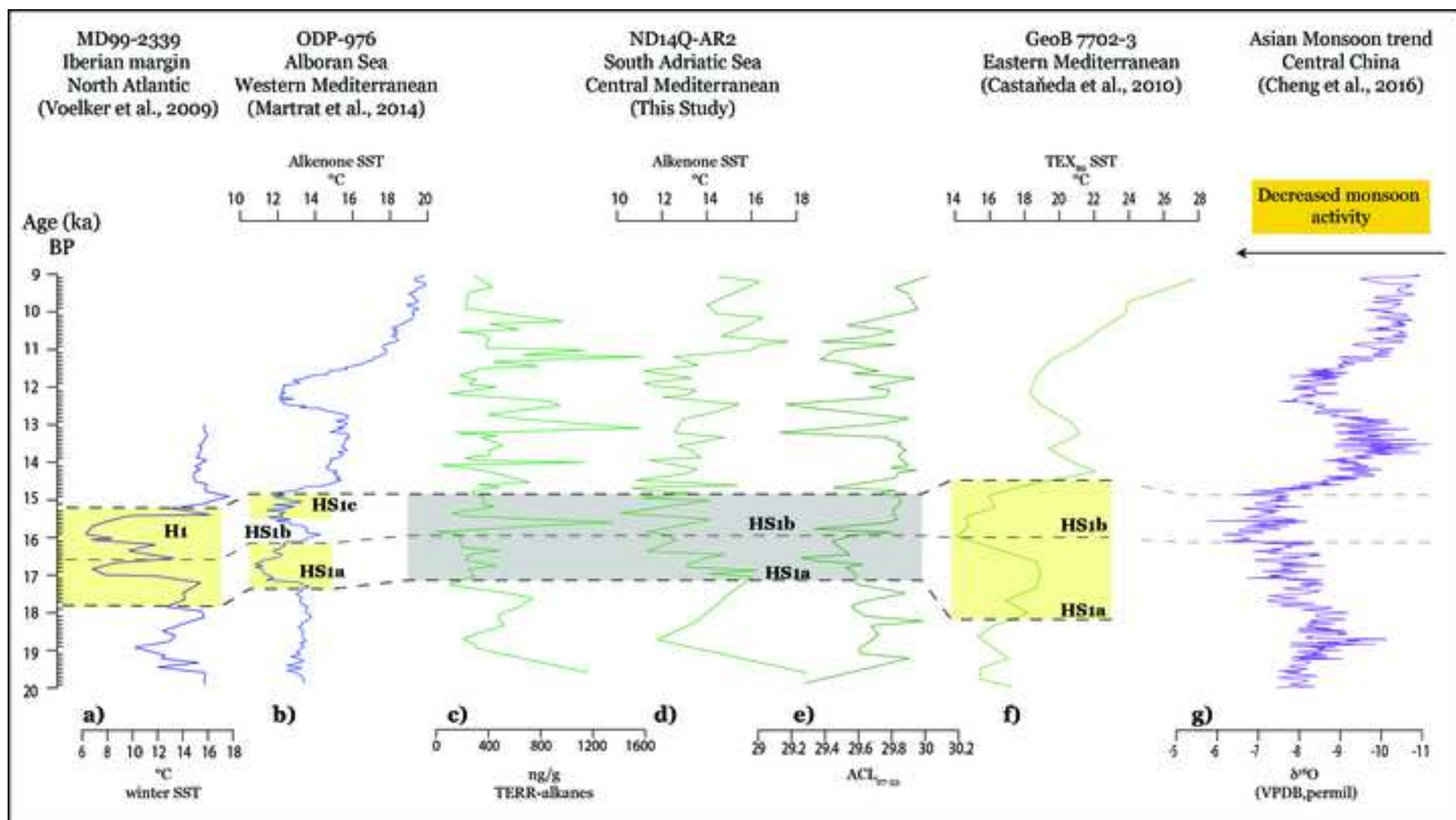












Declaration of interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: