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The Atlantic Meridional Overturning Circulation as productivity regulator of the North Atlantic Subtropical Gyre

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36

37

38 **Abstract**

39 Spatially extensive and intense phytoplankton blooms observed off Iberia, in satellite pictures,
40 are driven by significant nutrient supply by upper ocean vertical mesoscale activity rather than
41 by horizontal advection by coastal upwelling. Productivity of oligotrophic regions is still poorly
42 depicted by discrete instrumental and model data sets. The paleoproductivity reconstructions
43 of these areas represent the mean productivity over long-periods bringing new insights into the
44 total biomass fluxes.

45 Here we present paleoproductivity records from the oceanic Tore Seamount region, covering
46 the period from 140 to 60 Ka. They show higher nutrient supplies during Termination II, Marine
47 Isotopic Stage (MIS)4, MIS6, and warming transitions of the MIS5 sub-stages. The highest
48 nutrient content (higher productivity) in phase with tracers of bottom-water ventilation
49 (benthic $\delta^{13}\text{C}$, $^{231}\text{Pa}/^{230}\text{Th}$) establishes a strong linkage with Southern Ocean-sourced Waters
50 variability. Low productivity and ventilation over warm sub-stages of MIS5 respond instead to
51 North Atlantic Deep Water.

52 Assuming that the Tore Seamount is representative of oligotrophic regions, the glacial-
53 interglacial relationship observed between paleoproductivity and Atlantic Meridional
54 Overturning Circulation strength opens new insights into the importance of estimating the total
55 biomass in these regions. The subtropical gyres might play a considerable role in the carbon
56 cycle over (sub-) glacial-interglacial time scales than thought before.

57

58 Keywords: Tore Seamount; Productivity; AMOC; Mesoscale eddies

59

60 **1. Introduction**

61 Most paleoproductivity reconstructions have mainly focused on upwelling areas due to their
62 unique characteristics as major productive ecosystems, contributing to about 25% of the world's
63 fisheries landings (Jennings et al., 2001). Nevertheless, these regions constitute only a small part
64 of the ocean (roughly 5%). Past estimates for the nutrient-depleted subtropical North Pacific
65 Ocean, typically considered as the ocean's desert, suggest that this extensive region significantly
66 contributes to the carbon sequestered from the surface ocean by the ocean biological carbon
67 pump (Emerson et al., 1997; Sarmiento and Gruber, 2002). If illustrative of other subtropical
68 regions, such as the North Atlantic Subtropical Gyre, these large nutrient-depleted areas may be
69 responsible for up to half of the biological pump of carbon to the deep ocean (Emerson et al.,
70 1997; Oschlies, 2002). Moreover, studies based on the mid-latitude North Atlantic (Gil et al.,

71 2009; Schwab et al., 2012) suggest the importance of offshore (oligotrophic) productivity
72 changes over glacial-interglacial cycles to be addressed. These changes likely account for central
73 mechanisms in the past global carbon cycle (Emerson et al., 1997; Oschlies and Garcon, 1998;
74 Williams and Follows, 1998). It is thus essential to understand the processes behind the organic-
75 carbon export in these oligotrophic areas in order to improve comprehensive earth system
76 models that allow predicting future climate changes more accurately.

77 The Iberian margin is a key region for paleoceanographic studies (e.g. Shackleton et al., 2000).
78 It is represented by the North Atlantic Eastern Boundary Upwelling System, which has been
79 widely targeted for paleoproductivity reconstructions over different periods based on nearshore
80 cores (e.g., Abrantes, 1991; Cayre et al., 1999; Incarbona et al., 2010; Salgueiro et al., 2010). In
81 contrast, past changes in productivity within the adjacent offshore region have received far little
82 attention and only been studied at low temporal resolution (Lebreiro et al., 1997; Marino et al.,
83 2014; Martins et al. 2017).

84 Presently, subtropical regions can no longer be considered barren deserts as they account as
85 important productive areas due to mesoscale eddy activity (Dufois et al., 2016; Pelegri et al.,
86 2005; Oschlies and Garcon, 1998). Here, we emphasize the importance of phytoplankton
87 biomass generated by means of sub-mesoscale eddy-driven vertical nutrient fluxes (Lévy, 2003;
88 Dufois et al., 2016), due to upper ocean mesoscale activity (Supplementary Fig. 1) over glacial-
89 interglacial timescales.

90 We report on a multi-proxy high-resolution study of a sedimentary sequence taken at an open
91 ocean site off Portugal [core MD01-2446; Tore Seamount]. The area is part of the eastern
92 boundary of the North Atlantic Subtropical Gyre. The core site is located about 300 km from the
93 coast within a transitional/oligotrophic setting and is ideally positioned to investigate the role
94 of oceanic primary production in surface waters linked to upper ocean mesoscale activity in
95 comparison to onshore productivity driven by variable upwelling [e.g. Abrantes, 1991; Pailler
96 and Bard, 2002]. The combination of several parameters (reflecting oceanic productivity,
97 bottom water flow speed, surface and deep-water masses properties and terrestrial input)
98 enables us to fully characterize the evolution of offshore conditions at the Iberian margin over
99 the period of 140-60 Ka, where climate was as warm or warmer than today (Kukla et al., 2002)
100 and the response of Atlantic Meridional Overturning Circulation (AMOC) enhanced around the
101 transitions from warm to cold sub-stages (Guihou et al., 2011).

102

103 **2. Core location and modern hydrographic setting**

104 Core MD01-2446 [39°03'N, 12°37'W] is a Calypso Piston Core, 26.60 m in length, taken within
105 the Pole-Ocean-Pole project in 2001 during the MD123 Geosciences cruise on board the RV

106 *Marion Dufresne*. It was retrieved from the eastern external slope rim of the Tore Seamount, a
107 crater-like feature, about 100 km in diameter, that rises ~2800 m above the adjacent basin (Fig.
108 1). The sediment core was recovered at 3547 m water depth. Its location and depth on the
109 external rim of the seamount guarantees negligible influence from lateral downslope sediment
110 transport from the Iberian continental shelf. The far distance to the coast, enables monitoring
111 the past local (though transitional / oligotrophic) oceanic productivity changes (Behrenfeld et
112 al., 2006).

113 The Iberian Margin is bathed at the surface by the southward branch of the North Atlantic Drift
114 off Iberia (named Portugal Current) and by the northward Portugal coastal counter-current
115 (Fiúza, 1984; Peliz et al., 2005) (Fig. 1). During spring/summer seasons, Iberian coastal surface
116 waters are sourced in the Eastern North Atlantic Central Water that upwells (forming cold and
117 relatively more nutrient-rich surface waters) due to the influence of the trade winds, typically of
118 the Eastern Boundary Upwelling Systems (Fiúza, 1984). The offshore extent of these upwelled
119 cold waters is about 50 km wide (Fiúza, 1983). Present-day offshore influence of upwelled water-
120 related filaments and plumes, formed in the vicinity of capes, does not usually exceed 150-200
121 km into the open ocean (Sousa and Bricaud, 1992). In contrast, during the winter season, the
122 poleward slope current, described as a narrow slope trapped flow structure, has a turbulent
123 character with an associated mesoscale eddy field (Oliveira et al., 2004; Peliz et al., 2005). The
124 long-lived anticyclones formed during the separation of a poleward slope current at prominent
125 capes are observed off northern Iberia, extending down to water depths > 1000 meters,
126 migrating off slope and into the deep ocean (Peliz et al., 2005) (Fig. 1).

127 Deep-water circulation in the vicinity of the Tore Seamount is influenced by the lower North
128 Eastern Atlantic Deep Water (NEADW), characterized by high oxygen and low nutrient contents,
129 and the Southern Source Bottom Water with high nutrients and lower oxygen contents (van
130 Aken, 2000). At present, the NEADW is a mixture of the Iceland Scotland Overflow Water (23%),
131 Labrador Sea Water (27%) and (47%) Antarctic Bottom Water (AABW) (van Aken, 2000).

132 Core MD01-2446 is in the range of the North Atlantic Deep Water (NADW) during interglacials
133 and the Southern sourced Ocean Water (SOW) during glacials, allowing reconstruction of past
134 deep oceanic circulation (Adkins et al. 2005).

135

136 **3. Methods**

137

138 **3.1 Stable carbon and oxygen Isotopes**

139 Surface and deep water records of carbon and oxygen isotopic composition were based, in our
140 study, on the planktonic foraminifera *G. bulloides* and benthic foraminifera *C. wuellerstorfi*,

141 respectively. A total number of 186 raw samples (2 cm resolution of sampling interval) were
142 freeze-dried, later washed with distilled water, sieved at 63 and 150 μm , and then dried in paper
143 filters for 48 hours in an oven at 40°C. Specimens of *G. bulloides* were picked from the 250-315
144 μm fraction and *C. wuellerstorfi* from the fraction greater than 250 μm . The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of *G.*
145 *bulloides* (3-4 specimen per analysis) and *C. wuellerstorfi* (1-2 specimens per analysis) were
146 measured at the Laboratoire des Sciences du Climat et de l'Environnement (LSCE) in Gif-sur-
147 Yvette, on a Finnegan DELTA plus XP and GV OPTIMA and IsoPrime mass spectrometers. Data is
148 expressed in ‰ versus Vienna Pee-Dee Belemnite (VPDB). VPDB is defined with respect to
149 NBS19 calcite standard (Coplen, 1988). The mean external reproducibility (1σ) of carbonate
150 standards is $\pm 0.05\text{‰}$ and $\pm 0.03\text{‰}$ for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ respectively; measured NBS18 $\delta^{18}\text{O}$ is –
151 $23.2 \pm 0.2\text{‰}$ VPDB. The pooled standard deviation replicate measurements for the studied period
152 is 0.31 ‰ and 0.26 ‰ for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ respectively for *G. bulloides*, and 0.17 ‰ and 0.24 ‰
153 for *C. wuellerstorfi*. Oxygen isotope values for *C. wuellerstorfi* were corrected to the ambient
154 seawater equilibrium by 0.64 ‰ (Shackleton and Opdyke, 1973).

155 Analyses of bulk $\delta^{18}\text{O}$ were performed in about 50 mg of crushed sediment, treated with 100%
156 phosphoric acid in a Kiel-carbonate device type and loaded into a Finnigan MAT 251 mass
157 spectrometer at MARUM (University of Bremen, Germany). The precision of the equipment is
158 $\pm 0.07\text{‰}$ for standard repeated measurements.

159

160 **3.2 Grain size analyses**

161 A volume of 8-25 cm^3 of sediment sample was freeze-dried. When necessary, sample material
162 was disaggregated in a 0.033 mol sodium hexametaphosphate (Calgon) solution, then washed,
163 and sieved over a 63 μm mesh size. In the $<63 \mu\text{m}$ fraction, organic matter was removed with
164 hydrogen peroxide, excess reagent released in a 60°C warm water bath, and the samples washed
165 with distilled water through diatom ceramic candles. Each sample was then homogenised by
166 stirring and immediately measured in a Micrometrics Sedigraph 5100 at the LNEG laboratory. A
167 set of samples was treated with $\text{C}_2\text{H}_4\text{O}_2$ and $\text{C}_2\text{H}_7\text{NO}_2$ to eliminate carbonates, stirred and re-run
168 in the Sedigraph.

169 Both bulk (189) and carbonate-free (117) samples were analysed. The error of replicated
170 samples is $\pm 0.12 \mu\text{m}$ for bulk and $\pm 0.35 \mu\text{m}$ for carbonate-free analysis. Statistics were
171 calculated for the grain size distribution interval of 0-63 μm . The data is presented as two
172 parameters: percentage of sortable silt (SS%), i.e. $\%(10-63 \text{ mm})/<63 \text{ mm}$ (McCave et al., 1995;
173 McCave and Hall, 2006), and mean carbonate-free $<63 \mu\text{m}$. We have used mean carbonate-free
174 $<63 \mu\text{m}$, rather than mean sortable silt ($\overline{\text{SS}}$, mean grain size of the carbonate-free non-cohesive

175 10-63 mm interval) because samples of MD01-2446 contain <12% of %SS. However,
176 Supplementary Fig.3 illustrates a good cross-correlation between %SS and carb-free <63% of
177 R2=0.68, therefore achieving the purpose of tracing strength of bottom currents.

178

179 **3.3 Magnetic analyses**

180 To reconstruct past changes in the average magnetic grain size, ARM (Anhysteretic Remanent
181 Magnetization), IRM (Isothermal Remanent Magnetization), κ (volume low-field magnetic
182 susceptibility) data and hysteresis parameters have been used.

183 The low field magnetic susceptibility has been measured on u-channels with a 45-mm diameter
184 MS2-C Bartington coil at LSCE. The data were generated every 2 cm with a resolution close to 4
185 cm and then normalized by volume to get the volume susceptibility (κ). ARM was also acquired
186 on u-channels using a 50 μ T DC field with a superimposed 0.1T AF peak field along the axis of the
187 u-channel. The translation speed during acquisition was 1 cm/s to secure the full acquisition of
188 ARM (Brachfeld et al., 2004). The κ_{ARM} parameter was obtained by dividing the value of ARM by
189 that of the DC field. Saturated IRM (SIRM) was then acquired, also along the vertical geographical
190 axis at 1 T using a 2G 1.6 m long pulsed solenoid. The measurements of ARM and IRM were
191 made using cryogenic 755-R 2G magnetometers equipped with high-resolution pick-up coils
192 (Weeks et al., 1993) and placed in the μ -metal shielded room at LSCE. As for κ , the
193 measurements were made every 2 cm with a spatial resolution of about 4 cm.

194 Both ratios $\kappa_{\text{ARM}} / \kappa$ and ARM/IRM are used as proxies for magnetic grain size changes as they
195 are smaller (greater) for coarse (fine) populations of magnetic particles (King et al., 1982). The
196 above ratios are only empirically related to the magnetic grain size. Therefore, we also
197 performed magnetic hysteresis measurements using an AGM 2900 from Princeton
198 Measurements Corporation every 5 cm. The hysteresis loops were made between +1 and -1 T.
199 After adjustment of the high field slope, the saturation magnetization (M_s), the remanent
200 saturation magnetization (M_{rs}), and the coercitive field (H_c) were calculated. The remanent
201 coercitive field (H_{cr}) was then determined by a remanent curve starting by saturation at 1T
202 followed by a stepwise decreasing field, opposed to the saturating one. The M_r/M_s and H_{cr}/H_c
203 ratio are both related to the domain-state of the magnetic grains, in turn related to the magnetic
204 grain size. These data, although at slightly lower resolution were used to compare with the
205 continuously acquired $\kappa_{\text{ARM}} / \kappa$ and ARM/IRM ratios.

206

207 **3.4 Bulk Geochemistry**

208 The biogenic opal content in each sample was measured at LNEG laboratory following the
209 reduction colorimetric technique by molybdate-blue spectrophotometry (Mortlock and
210 Froelich, 1989). All the samples, standards and blanks were duplicated, and therefore opal data
211 was plotted, for each depth, as average value between sample and replicate. The long-term
212 average of the standard deviation of all the pairs of samples/replicates is ± 0.03 wt%.

213 The siliceous fraction is prone to dissolution in the North Atlantic, and therefore the
214 determination of biogenic opal in pelagic regions, where typically siliceous productivity is low, is
215 known to be limited (Mortlock and Froelich, 1989). For this reason, we have used an additional
216 (and independent) proxy for productivity – the organic carbon (Corg), allowing for comparison
217 with opal data. Even though dependent on bottom water redox conditions, the Corg record
218 generally shows a positive correlation between surface water productivity and organic carbon
219 accumulation in the sediments (Stein, 1990; Nave et al., 2001; Lebreiro et al., 2006; Romero et
220 al., 2008).

221 The total carbon content was determined in three replicates of 2 mg of freeze-dried, ground and
222 homogenized bulk sediment sample per level, using CHNS-932 LECO elemental equipment at
223 LNEG laboratory. For each replicate, total carbon was analysed, then organic carbon combusted
224 at 400°C temperature (4 steps of 100°C/h plus 3h at 400°C) and analysed for inorganic carbon
225 content (CaCO₃). Organic carbon was calculated, for each level, as the difference between total
226 carbon and inorganic carbon (CaCO₃) in weight percentage (wt %).

227 Sedimentary fluxes of productivity data (opal, organic carbon) were determined by using ²³⁰Th
228 normalization as it provides a means to achieve more accurate vertical mass accumulation rates
229 (François et al., 2004). Indeed, vertical mass accumulation rates, when corrected to ²³⁰Th
230 normalized flux avoids misinterpretations of sediment mass accumulation rates due to sediment
231 focusing on the seafloor by bottom currents (François et al., 2004). Thus, all productivity data
232 was corrected to ²³⁰Th normalized flux (Guihou et al., 2010; Guihou et al., 2011), instead of using
233 the sediment Mass Accumulation Rates.

234

235 **3.5 X-ray analyses**

236 Bulk sediment samples were freeze dried overnight and hand crushed in an opal mortar. The
237 mineralogical characterization of sediment samples by X-ray diffraction (XRD) was carried out
238 using a Philips PW 1710 powder diffractometer with Bragg-Brentano geometry, equipped with
239 a large anode copper tube and a graphite single crystal monochromator. These assays were
240 performed at the X-ray laboratory installed in the Unit of Mineral Resources and Geophysics of
241 LNEG.

242 Six elements were selected for X-ray fluorescence (XRF) spectrometry tests due to their
243 relevance in marine sediment records: Ca, Sr, Al, Fe, Ti, K and Ba. Powdered samples (approx.
244 200 mg) were irradiated using an automated wavelength dispersive spectrometer Philips PW
245 1400 equipped with a rhodium tube to collect five successive counting (30 sec fixed counting-
246 time) at the diagnosis emission line of each one of these elements ($L\alpha$ for Ba and $K\alpha$ for the
247 others) and at the corresponding spectral background using suitable analysing crystals. After
248 subtracting the background from the peak counts, selected counting ratios were calculated: Ca,
249 Sr, Ba vs. Al, and K, Fe, Ti vs. Ca.

250 Because such quick XRF tests have essentially a semi-quantitative character, comparison with
251 11 discrete samples that were analyzed by ICP-AES at an external certified laboratory (Aveiro
252 University) allowed to validate the geochemical data obtained at the local X-ray laboratory. For
253 funds reasons, only a limited number of samples (11 samples) and elements (Ba as proxy for
254 productivity) could be analyzed and we chose Al and Ba which yielded average RSD (%) of 2.4 %
255 and 1.7 %, respectively.

256

257 **4. Chronology**

258 In order to achieve a precise chronological framework to define sub-stages (millennial-scale
259 precision) we compared cores located in a narrow region instead of using a global benthic stack
260 (Lisiecki and Raymo, 2005). Indeed, it has been shown that there might be significant differences
261 in large glacial-interglacial transitions of the benthic $\delta^{18}\text{O}$ records of different depth/water
262 masses/ocean basins (Waelbroeck et al., 2011).

263 Over the period 140-60 ka 4 tie-points have been defined between *G. bulloides* oxygen isotopic
264 records of each core (MD95-2040 and MD01-2446) and the *G. bulloides* oxygen isotopic record
265 of core MD95-2042 (Supplementary Fig. 2). MD95-2040 and MD01-2446 indicate a mean
266 sedimentation rate of 5cm/ka over stage 5.

267 At present, and likely during the time span of the record, the Tore Seamount location is within
268 the same water masses as core MD95-2042, which has a well-established and precise
269 chronology. Thus, we have used the revisited MD95-2042 age model (Govin et al., 2014) based
270 on the AICC2012 ice core chronology (Bazin et al., 2013; Veres et al., 2013), as the reference to
271 build the MD01-2446 chronology, by direct relationship between the planktonic $\delta^{18}\text{O}$ records
272 of both cores. Ages between the age control points were calculated by linear interpolation using
273 Analyseries software (Paillard et al., 1996). Errors for the current correlation range between 0.5
274 to 1.5 ka on the age scale. Assuming that rapid transitions are synchronous, the error is smaller
275 near rapid transitions (and the tie-points) and larger in between.

276

277 **5. Results**

278 Results of the analysed core, MD01-2446, are presented in comparison with previously
 279 published data of coastal sites (cores MD95-2040: 40°34.91'N, 9°51.67'W, 2465 m water depth
 280 and MD95-2042: 37°47.99'N, 10°9.99'W, 3146 m) and a nearby offshore core (D11957: 39°03'N,
 281 12°35'W, 3585 m water depth) (table 1).

282

Core	Reference	Data
MD01-2446	This study	Planktonic $\delta^{18}\text{O}$ (<i>G. bulloides</i>) Benthic $\delta^{13}\text{C}$ (<i>C. wuellerstorfi</i>) Benthic $\delta^{18}\text{O}$ (<i>C. wuellerstorfi</i>) Paleomagnetic data: ARM/IRM, $\kappa\text{ARM}/\kappa$ and Mr/Ms SS% and mean carbonate-free <63 μm . $\delta^{18}\text{O}$ Bulk sediment Opal content (weight %) Organic carbon (weight %) CaCO ₃ (weight %)
	Guihou et al., 2010	²³¹ Pa/ ²³⁰ Th
MD95-2042	Cayre et al., 1999	SST °C (summer) based on foraminifera MAT Transfer Function $\delta^{18}\text{O}$ (<i>G. bulloides</i>)
	Shackleton et al., 2000	Benthic $\delta^{18}\text{O}$ (<i>Globobulimina affinis</i> , <i>C. wuellerstorfi</i> , <i>Uvigerina</i> spp.) Benthic $\delta^{13}\text{C}$ (<i>Globobulimina affinis</i> , <i>C. wuellerstorfi</i> , <i>Uvigerina</i> spp)
	Salgueiro et al., 2010	Export Production (gC.m ⁻² .yr ⁻¹)
MD95-2040	Abreu et al., 2003	SST °C (summer) based on foraminifera Simmax.28 Transfer Function $\delta^{18}\text{O}$ (<i>G. bulloides</i>)
D11957	Lebreiro et al., 1997	Primary Productivity (SIMMAX.28 Transfer Function) Polar (SIMMAX.28 Transfer Function) Gyre Margin (SIMMAX.28 Transfer Function) Upwelling (FA20 Transfer Function)

283

284

285 **5.1 Coastal versus offshore surface water masses**

286 Comparison between nearshore and offshore areas of the Iberian margin (table 1) was based on
 287 changes in $\delta^{18}\text{O}$ of planktonic foraminifer *G. bulloides* (Fig. 2A). Over the 140-60 ka interval,
 288 oxygen isotopic values from core MD01-2446 vary approximately between 1 and 3 ‰. Typical
 289 glacial-interglacial isotopic changes are well recorded with heavier $\delta^{18}\text{O}$ values at MIS 6 and 4
 290 and the onset of MIS 5a, and lighter values standing equally for 5e, 5c, and relatively heavier for
 291 5a (Fig. 2A).

292 During warm phases (well-marked during 5e) isotopic values are heavier offshore (D11957) than
 293 nearshore (MD95-2040 and MD95-2042) (Fig. 2A). Oxygen isotopic differences between sites
 294 (north-south and east-west) are nearly negligible during glacial MIS 4, and sub-glacial stages 5b
 295 and 5d (Fig. 2A).

296 Sea Surface Temperatures (SST) based on foraminifera (table 1) show warmer temperatures
297 offshore (D11957) than at the northern nearshore site MD95-2040 and more similar to the
298 southern nearshore site MD95-2042 (Fig. 2B).

299

300 **5.2 Deep water masses**

301 The benthic oxygen isotopic record ($\delta^{18}\text{O}$ *C. wuellerstorfi*) from core MD01-2446 varies between
302 3 and 5 ‰ over the studied time interval showing the typical glacial-interglacial variability (Fig.
303 2C). Compared to the previously published MD95-2042 benthic $\delta^{18}\text{O}$ record (table 1), values are
304 very similar along MIS 5, but with marked lighter values during MIS 6 (Fig. 2C).

305 The comparison of benthic $\delta^{13}\text{C}$ (an indirect indicator of deep-water ventilation; Curry and Oppo,
306 2005) between MD01-2446 and nearshore core MD95-2042, points to heavier $\delta^{13}\text{C}$ values
307 (nutrient depleted) offshore, except during the short-time intervals: 106-101 ka and 83-77 ka
308 (Fig. 2D). This observation is reinforced by the similarity of the benthic $\delta^{13}\text{C}$ with the $^{231}\text{Pa}/^{230}\text{Th}$
309 record, a proxy for AMOC export variability (Guihou et al., 2010), both obtained from core
310 MD01-2446 (Fig. 3A-C, table 1).

311 The ARM/IRM, $\kappa\text{ARM}/\kappa$, and Mrs/Ms ratios, used to reconstruct past changes in the average
312 magnetic grain size, indicate relatively coarser magnetic grains coincident with periods of
313 depleted $\delta^{13}\text{C}$ at MD01-2446 (Fig. 3D). The SS% and mean carb-free $<63\ \mu\text{m}$, proxies for sorting
314 and speed of bottom water (Supplementary Fig.3), show an enhanced and well-sorted deep
315 current during MIS 6, MIS 4 and MIS 5b, but slower flow during MIS 5d and 5a and intermediate
316 MIS 5e (Fig. 3E).

317

318 **5.3 Continental Input**

319 Oxygen isotope ratios of bulk carbonate have been used as an efficient proxy for land
320 transported detrital carbonate to the seafloor during periods of low sea-level stands (Lebreiro
321 et al., 2009). Bulk oxygen isotopes of MD01-2446 show a long term trend toward heavier values
322 through MIS 5 consistent with the relative sea-level curve (Waelbroeck et al., 2002), although
323 the sub-stages of MIS 5 are not clearly defined (Figs. 4A and B).

324

325 **5.4 Paleoproductivity records**

326 Paleoproductivity variations were estimated using several independent proxies like biogenic
327 opal content, $\delta^{13}\text{C}$ difference (*G. bulloides-C. wuellerstorfi*) and organic carbon (Corg) (Fig. 5A-D).

328 The opal content (%) from core MD01-2446 follows a glacial-interglacial and sub-millennial
329 (transitions in sub-stages) variability showing highest siliceous primary productivity mainly

330 during cold periods, MIS4 and 6, and during transitions 5d to 5c and 5b to 5a (Fig. 5A). However,
331 in order to avoid biased interpretations of the opal % record, commonly associated with
332 sediment focusing, the ^{230}Th -normalised opal fluxes (hereafter referred as opal flux) was also
333 used. The opal flux undergoes a similar trend as the opal content (wt %), with larger differences
334 at MIS4 where despite the increased opal %, enhancement of opal flux is not observed (Fig. 5A-
335 B).

336 The difference between $\delta^{13}\text{C}$ recorded in planktonic and benthic foraminifera has been long
337 used as a useful indicator of deep ocean carbon storage, and its variations may be explained by
338 changes in the relative strength of export production (Broecker, 1982b; Shackleton et al., 1983;
339 Olivier et al., 2010). $\delta^{13}\text{C}$ difference _(planktonic-benthic) from MD01-2446, presents general similar
340 trend to opal % record (higher values during MIS4 and 6, and during 5d to 5c and 5b to 5a
341 transitions, Fig. 5C). Moreover, the organic carbon variability is also generally consistent with
342 the opal content, however, with higher variability, mainly from 5b to MIS4 period (Fig. 5A-B).

343 $\text{Log}(\text{Ca}/\text{Ti})$ which reflect relative changes of biogenic carbonate and detrital sediment (Hodell et
344 al., 2015) resembles CaCO_3 flux (Fig. 5D) presenting an inverse trend in comparison to
345 paleoproductivity proxies (Opal, Corg and $\delta^{13}\text{C}$ difference _(planktonic-benthic)).

346 The correlation between productivity and $\delta^{13}\text{C}$ was assessed by statistical analysis. The ^{230}Th -
347 normalized opal is not available for the last $\delta^{13}\text{C}$ negative anomaly (MIS6, Fig. 3, Fig. 5), since
348 bulk flux was not determined down the 621 cm core depth (age model ≥ 128 ka). For this reason,
349 the correlation between opal (wt%) and $\delta^{13}\text{C}$ was used (Supplementary Fig.4), after confirming
350 that the correlation between opal and opal flux are statistically significant (Supplementary
351 Fig.5).

352 The relationship between opal (wt%) and $\delta^{13}\text{C}$ (Supplementary Fig.4), estimated through the
353 Pearson correlation coefficient r for 152 samples, is moderate negative ($r = -0.52$) and
354 statistically significant (with a p-value clearly below a significance level of $\alpha = 0.001$). This
355 negative correlation confirms higher productivity coeval lower $\delta^{13}\text{C}$ values (i.e., coeval with
356 stronger influence of southern ocean waters).

357 The relationship between opal (wt%) and ^{230}Th -normalised opal flux (Supplementary Fig.5),
358 estimated through the Pearson correlation coefficient r for 144 samples, is moderate positive (r
359 = 0.55) and statistically significant (with a p-value clearly below a significance level of $\alpha = 0.001$).
360 Thus, mean to high values of ^{230}Th -normalised Opal flux are expected to coincide with the last
361 observed $\delta^{13}\text{C}$ negative anomaly (MIS6, Fig. 3, Fig. 5).

362 Note that opal and $\delta^{13}\text{C}$ were generally analysed in different support samples (odd and even cm
363 depths respectively), coinciding for nine depth-samples only; this number of samples is generally

364 not enough to measure the strength of a relationship, namely a trustful value for the Pearson
365 correlation coefficient. The remaining opal values used to assess the relationship between opal
366 and $\delta^{13}\text{C}$ (152-9 = 143) have been estimated, with each value corresponding to the opal average
367 of the previous and the following cm-samples. This is likely to cause an underestimation of the
368 actual Pearson correlation coefficient.

369

370 **6. Discussion**

371 **6.1 Open ocean versus nearshore hydrographic conditions off Iberia**

372 According to modern conditions in the North Atlantic, the classic temperature gradients of
373 latitudinal sea surface migrate North-South, on a seasonal basis, and tilt NE-SW during summer
374 and autumn (Supplementary Fig. 4). This is reflected by summer SSTs that indicate similar
375 temperatures in core MD95-2042 and MD01-2446 (Supplementary Fig. 6). By contrast, during
376 winters, there is an increase in temperature gradients along an N-S transect but in absence of
377 upwelling no E-W, i.e. nearshore-offshore, strong temperature gradient occurs (Supplementary
378 Fig. 6).

379 The $\delta^{18}\text{O}$ *G. bulloides* record (Fig. 2A) is more representative of the spring conditions as this is
380 the dominant species during spring in this area (Lombard et al., 2011). In modern spring, the
381 temperature and salinity differences between sites MD01-2446 and MD95-2040 have opposite
382 effects on the calcite $\delta^{18}\text{O}$ (Supplementary Fig. 6). Since $\delta^{18}\text{O}$ *G. bulloides* is similar at both sites
383 during most of the stage 5, it indicates that similar temperature and salinity gradients have
384 prevailed on the long-term. On a shorter time-scale, the nearshore record (MD95-2040)
385 indicates higher oxygen isotopic values than the offshore record at the end of 5c, 5d-5c, and 5b-
386 5a transitions. This could be due to the transient southward migration of the polar front system,
387 before it reaches a position south of the three sites during stage 5b and 4. This assumption is
388 supported by the highest summer surface temperature, at southern cores, during those periods.
389 Even though the Tore Seamount location is currently a subtropical region, the presence of polar
390 (foraminifera) fauna during glacial stages 4 and 6 and during the 85 ka cold MIS 5b sub-stage
391 (Fig. 6E) also confirms that this area is episodically part of the subpolar region (Lebreiro et al.,
392 1997; Cayre et al., 1999; Abreu et al., 2003) due to the southward migration of the polar frontal
393 system. During most of stage 5e the nearshore record of core MD95-2040 exhibits lighter oxygen
394 isotopic values than offshore MD01-2446 record, while summer SSTs are still lower nearshore
395 than offshore. Unless a large change in seasonal temperature evolution had occurred, this
396 particular trend may result from a salinity decrease at the nearshore site. This could be related
397 to the increased annual precipitation at NW Iberian and its possible influence on the shelf area
398 by river runoff (Sánchez Goñi et al., 2005).

399 The oxygen isotopic planktonic difference between the south nearshore MD95-2042 site and
400 offshore site during sub-stage 5e, as well 5c and 5a (Fig. 2B), is larger than expected from the
401 present spring temperature and salinity differences between the two sites (Supplementary Fig.
402 7). It implies therefore a larger winter and spring temperature gradient between the two sites,
403 potentially corresponding to a northward placement of the Azores front compared to its modern
404 spring position.

405 Concerning the deep-ocean, grain size trends are the result of sediment transport and
406 deposition processes. At depth, variations of the grain size in the sortable silt range have been
407 used to infer the dynamic of near-bottom paleo-current strength (McCave et al., 1995; Bianchi
408 et al., 1999; Hall and McCave, 2000; McCave and Hall, 2006; McCave et al., 2017).

409 We observe a trend of higher mean carbonate-free <63 μm grain size and %SS in MD01-2446
410 indicating stronger deep water currents during MIS 6 and 4 and sub-stadial 5b, together with
411 early interstadial 5e (Fig. 3E). Hall and McCave (2000) reported high sortable silt (\overline{SS}) at ~ 2465 m
412 in MD95-2040 on the Portuguese margin during warmer substages of MIS 5, as a result of faster
413 southern flow of the NEADW paleo-current. The apparently opposed dominant speed during
414 glacials in MD01-2446 can only be explained if the grain size parameters are recording a
415 northward bottom (~ 3500 m) flow at the Tore slope, which we link with AABW (Adkins et al.,
416 2005). This scenario is coherent with the observed variation of depleted $\delta^{13}\text{C}$ benthic traces of
417 the AABW (Fig. 3A, E).

418 On the other hand, the magnetic grain size parameters suggest efficient transport of detrital
419 magnetite by bottom-water flow (corresponding to coarser magnetic grains) during periods of
420 low benthic $\delta^{13}\text{C}$ (mainly MIS4 and MIS6, but also at warming transition 5d-c) (Figs. 3A-D). As it
421 has been widely reported for the North Atlantic, NADW production was weaker and slower
422 during these periods, (e.g. Michel et al., 1995; Shackleton et al., 2000; Kissel, 2005) despite
423 increased production above 2000 m depth of Glacial North Atlantic intermediate waters (Michel
424 et al., 1995; Shackleton et al., 2000; Guihou et al., 2011). As a result, the magnetic data appear
425 to selectively illustrate (sub)orbital variability associated with a southern bottom-water at 3500
426 mwd in the Eastern Atlantic Margin, seen at MD01-2446. The volcanic Canary Islands and/or
427 Cape Verde Islands might be the source of these magnetic grains, commonly Fe-Ti in oxides and
428 silicates.

429 Channell et al. (2013) report bacterial magnetite on two cores onshore at the SW Iberian margin
430 (MD01-2443 and MD01-2444) varying on stadial-interstadial timescales, similarly to our data.
431 The detrital/biogenic magnetite ratio increased with hematite concentration together with
432 coarser magnetites during stadials and glacial isotopic stages. However, the cores from Channell
433 et al. (2013) are located at the Iberian margin at 2600-2900 m water depth, where a lot of

434 detrital carbonate could have originated from the continental margin (Lebreiro et al., 2009)
435 whereas core MD01-2446, is a deeper core (3547 water depth) and out of Iberian margin
436 influence at is deeper.

437 Another hypothesis supporting the magnetic signal of this region was recently raised by
438 Yamazaki and Ikehara (2012), in which the resemblance of the pattern of the SUSAS stack in the
439 subtropical Atlantic with dust records from Antarctic ice cores (Schmieder et al., 2000), would
440 be explained by the redistribution to the subtropical Atlantic of biogenic magnetites produced
441 in circum-Antarctic Ocean sediments under iron fertilization by bottom water currents
442 (Yamazaki and Ikehara, 2012).

443 Whatever is the source of the magnetite particles, the two grain-size data sets converge if the
444 mineralogical components of coarse magnetic size are contained in the silt grain size fractions,
445 transported a long way, northward by bottom currents.

446 Taking into account the multi proxy approach presented in our study, our preferred
447 interpretation is that the coarse magnetic grain size at site MD01-2446 likely records the
448 northward SOW bottom current during colder periods (more SOW influence at these latitudes)
449 also coinciding with periods of lighter $\delta^{13}\text{C}$ (less contribution of NADW) and lower $^{231}\text{Pa}/^{230}\text{Th}$
450 values. In fact, our record reveals that warmer periods of weaker influence of SOW correspond
451 to southward intensification of NEADW.

452 In any case, further investigations are needed to test the potential of magnetic grain size as a
453 proxy for northwards SOW deep current in the Eastern North Atlantic as far from the coast as
454 the Tore Seamount.

455

456 **6.2 Source of nutrients sustaining paleoproductivity in oligotrophic areas**

457 A paleoproductivity signal consistent with glacial-interglacial climate variability is observed at
458 orbital timescales in the Tore Seamount region. The main question to be addressed is the origin
459 of the nutrients supporting this oceanic productivity.

460 Given the geographic location and water depth (about 3500 m) of the studied core, we do not
461 expect any contribution from sediments transported across the Tagus abyssal plain upslope to
462 the Tore Seamount by bottom currents (Fig. 1).

463 Two main alternative sources are therefore possible: 1) sinking material from coastal advection
464 by means of offshore extension of filaments associated with coastal upwelling, similarly to
465 observed in nearby southward regions (Aristegui et al., 2004), and 2) local offshore productivity
466 generated by mesoscale eddy activity at the North Atlantic subtropical gyre. The relative
467 changes in the contribution of both sources may also constitute an additional and possible
468 scenario.

469 Lighter $\delta^{18}\text{O}$ values of bulk sediment in the nearshore Iberian margin, core MD03-2698 (Lebreiro
470 et al., 2009), are associated with enhanced continental carbonate input during periods of low
471 sea-level (colder periods). These inputs should also imply a greater nutrient transport from
472 continental-sourced material (Broecker, 1982; Filippelli et al., 2007) by lateral advection of
473 surface waters to the ocean realm. Interestingly, the offshore bulk $\delta^{18}\text{O}$ copies the planktonic
474 foraminifera $\delta^{18}\text{O}$ with lighter values during warm periods (Fig. 4), opposed to coastal core
475 MD03-2698 (Lebreiro et al., 2009). This suggests that the bulk $\delta^{18}\text{O}$ is a proxy limited to
476 continental input at nearshore regions, and that the Tore Seamount record is out of major
477 continental influence.

478 Enhanced diatom production in regions such as the tropical Atlantic, has been attributed to be
479 driven by enhanced wind-driven upwelling or density-driven vertical mixing, or by elevated
480 thermocline concentrations of silicic acid supplied to the surface at a constant rate (e.g
481 Bradtmiller et al. 2016; Hendry et al., 2016).

482 The location of core MD01-2446 is beyond the 50 km extension of coastal upwelling influence
483 (Fiúza, 1983), where current seasonal filaments rarely extends 150-200 km offshore (Sousa and
484 Bricaud, 1992). Considering that during cold phases of MIS5 and during MIS4 and MIS6 the sea
485 level was lower than today (-40m during cold phases of MIS5, and about -80 and -120 during
486 MIS4 and MIS6, respectively), the seasonal filaments, during the low stand sea level, would
487 reach further offshore than present scenario. Considering the position of the isobath of -120 m,
488 the filaments could reach about 30 km further offshore but, still, that would put the site out of
489 the reach of the filaments. Although it has been suggested that filament activity is likely
490 intensified by stronger trade winds during cold glacial periods (Lebreiro et al., 1997), considering
491 a far reaching of 200 km offshore, with a low stand of -120m of sea level, their contribution to
492 the studied site would not be significant. It therefore does not support, by itself, the overall
493 offshore productivity.

494 Consequently, to explain the overall productivity at the offshore site we propose local sources
495 of nutrients generated by the upper ocean mesoscale eddy activity (Dufois, F. et al., 2016; Lévy,
496 2003) (Supplementary Fig. 1). This is consistent with the growing body of evidences accumulated
497 over the last decade pointing towards a significant primary productivity in oligotrophic areas
498 (McGillicuddy et al., 1998; Oschlies and Garcon, 1998; Williams and Follows, 1998; Oschlies,
499 2002; McGillicuddy et al., 2007; Incarbona et al., 2010; Williams, 2011).

500 Though mesoscale eddies have been shown to be sites of higher nutrient levels and/or biological
501 productivity in the mixed layer compared with the surrounding waters in the ocean. Our
502 multiproxy study, integrating time and space, is a strong evidence that oceanic paleoproductivity

503 variation (likely related to sub-mesoscale driven vertical nutrient fluxes) echoes that of glacial-
504 interglacial climate variability, in strong agreement particularly with the benthic $\delta^{13}\text{C}$ and
505 $^{231}\text{Pa}/^{230}\text{Th}$ records (Fig. 5A,C).

506 The higher contribution of nutrient-rich AAIW type southern sourced waters and the state of a
507 deeper upper mixed layer during cold periods would enrich the nutrient supply to the photic
508 zone favouring enhanced productivity in the Tore Seamount region.

509 As recently documented, biogenic opal export at low-latitudes is likely to rely on silicate from
510 the underlying thermocline, the concentration of which is related with the circulation of the
511 ocean interior (Meckler et al., 2013). A pronounced opal maxima during each glacial termination
512 over the past 550,000 years was reported off northwest Africa (Meckler et al., 2013), during the
513 times of a consistent strong deglacial reduction in the formation of silicate-poor glacial North
514 Atlantic intermediate water (GNAIW). Even though Meckler et al. (2013) focus on a coastal
515 upwelling system, where productivity is dependent of nutrients reaching the surface due to
516 upwelled waters related to trade winds intensification, the concentration of silicate at ocean
517 interior would be favourable to increased productivity also at oligotrophic regions if sub-
518 mesoscale driven vertical nutrient fluxes occur. This enhanced productivity, at glacial
519 terminations, was documented reaching latitudes at the Iberian margin, northwards of Tore
520 Seamount position, as evidenced by the diatom and barium records from core MD95-2039
521 (Thomson et al., 2000).

522 Enhanced northward advection of AAIW during periods of reduced North Atlantic overturning
523 circulation has also been recently documented (Rickaby and Elderfield, 2005; Pahnke et al.,
524 2008). Increased nutrient supply via intermediate water masses feeding into the thermocline
525 has been suggested to explain the observed opal changes in the E-Atlantic (Meckler et al., 2013),
526 which are consistent with Si isotope distribution patterns from several Atlantic sediment cores
527 (Hendry et al., 2016).

528 In parallel, our data reveals that at the eastern North Atlantic, periods of weakened North
529 Atlantic Deep Water formation (lighter benthic $\delta^{13}\text{C}$) are coeval with higher productivity,
530 suggesting a close relationship between oceanic productivity and resumption of southern ocean
531 waters (Fig. 5; Supplementary Fig.4).

532 This relationship was previously suggested through simulations with a coupled climate-
533 ecosystem model of intermediate complexity, revealing that during periods of abrupt climatic
534 changes (at millennial timescales), global ocean productivity is sensitive to changes in the
535 Atlantic Meridional Overturning Circulation (Schmittner, 2005).

536 In the context of our dataset, the coupling between $\delta^{13}\text{C}$ and productivity records could be
537 compromised by a circular issue effect, given that the $\delta^{13}\text{C}$ spreading of dissolved inorganic
538 carbon is primarily dependent on the interaction of biological uptake at the sea surface.
539 However, as confirmed for the Canary Islands region, the benthic foraminifera mean $\delta^{13}\text{C}$ values
540 of *C. wuellerstorfi* are not influenced by sustained high organic matter fluxes, validating this
541 species as a reliable recorder of bottom water $\delta^{13}\text{C}$ (Eberwein and Mackensen, 2006). It can thus
542 be assumed a factual affiliation between oceanic productivity changes and AMOC.
543 We therefore suggest that besides the upwelling intensity, the presence of southern sourced
544 silica-rich waters play a major role on the low-latitude biological productivity record as an
545 important source of nutrients. The deep NADW is slightly reduced during cold sub-stages of MIS
546 5 but the intermediate water formation (down to 2000 m depth) is greatly enhanced during cold
547 sub-stages (Guihou et al., 2011).

548

549 **6.3 The 85 ka and 105 ka productivity-events: oceanic nutrient supply**

550 The benthic $\delta^{13}\text{C}$ record show lower values during millennial warming transitions 5d to 5c, and
551 to a lesser extent 5b to 5a (Fig. 3A). These nutrient enriched periods are mostly coeval with
552 higher $^{231}\text{Pa}/^{230}\text{Th}$ values illustrating weakening of the AMOC export at MIS 5d to 5c and at MIS
553 5b to 5a transitions (Guihou et al., 2010; Guihou et al., 2011). Intensification of offshore
554 productivity is well-marked by different proxies, such as high opal content, $\delta^{13}\text{C}$ difference
555 (planktonic-benthic) and organic carbon content at the same periods (Figs. 5 A-D). It is also
556 supported by the productivity foraminifera-based SIMMAX.28 (Figs. 6B-C-D-E) pointing to rapid
557 and significant environmental changes.

558 Primary Productivity (Fig. 6D) reflected in Export Production at the open ocean site is different
559 from that at the nearshore sites (Fig. 6B). This disparity illustrates the development of high
560 nutrient levels in coastal zone waters compared to the offshore (oceanic) areas, where its direct
561 influence decreases. Similar records in both sites would imply an influence from terrestrial input
562 or erosion from sea level lowering toward an echo paleoproductivity but with much lower
563 magnitude at the open ocean environment (Thomson et al., 2000). The disparity between the
564 two sites, therefore, sustains productivity originating from the open ocean.

565 As shown in a more detailed view of the time sequences of events off Iberia over the last climate
566 cycle, cold SST episodes after relatively warm and largely ice-free periods occurred when the
567 predominance changed from northern deep waters (NADW) to southern (AABW) (Martrat et al.,
568 2007). Different origin of bottom waters are likely to be reflected on the surface record, not only
569 on SST but also on productivity as it is suggested by the coeval times of 85 ka and 105 ka

570 productivity-events and GS-22 and GS-24 at Martrat et al. (2007). The similar pattern of the
571 various proxies responding to hydrography and productivity variations is very consistent for the
572 analysed period. Therefore, the harmony of the records reflecting the ocean dynamics and
573 underlying biomass seems to point to a coupling between oceanic productivity and the SOW.

574

575 **7. Conclusions**

576 Interglacial conditions of MIS5 are compared to glacial stages MIS4 and MIS6 of higher
577 productivity, in areas under the potential influence of a coastal upwelling regime (Pailler and
578 Bard, 2002; Salgueiro et al., 2010). Site MD01-2446 shows higher nutrient supply during
579 Termination II, MIS4, MIS6, and warming transitions of the MIS5 sub-stages. The highest nutrient
580 content (higher productivity) is in phase with tracers of deep-water ventilation variations
581 (benthic $\delta^{13}\text{C}$ and $^{231}\text{Pa}/^{230}\text{Th}$), and therefore establishes evidence of the coupling between
582 productivity variations and the SOW.

583

584 The paleoproductivity record explained mainly by the variability of the sub-mesoscale driven
585 vertical nutrient fluxes is in agreement with benthic $\delta^{13}\text{C}$ and $^{231}\text{Pa}/^{230}\text{Th}$. Periods of low benthic
586 $\delta^{13}\text{C}$ and high $^{231}\text{Pa}/^{230}\text{Th}$ (less North Atlantic Deep Water formation, more GNAIW and
587 potentially AAIW and/or mode waters formation) during warming transitions of MIS5 sub-stages
588 which are coeval with higher productivity, suggest a link between oceanic productivity changes
589 and the Atlantic Meridional Overturning Circulation.

590

591 If such a link exists, as suggested by the data, it highlights the potential role of the subtropical
592 Atlantic Ocean in the ocean biological carbon pump, over (sub-) glacial-interglacial time scales.

593

594 Changes of coarse magnetic grain size in the sediment, combined with enhanced mean carb-
595 free $<63\ \mu\text{m}$ and SS%, suggests a distinct source of bottom flow. Coarse magnetic grains are
596 concentrated in silt grain size, and Ti, Fe and K are mainly adsorbed into clays, thus probably
597 recording the long way transport and deposition of sediments during colder periods by the
598 northward SOW current. Both parameters are signs of the dynamics of the bottom current flow
599 and its pathway.

600

601 From the Tore Seamount comparison to Iberian nearshore environments [cores MD95-2042 and
602 MD95-2040], a relationship between offshore and overturning circulation is proposed, raising

603 the need of increasing the number of studies on paleoproductivity variations in oligotrophic
604 areas in future research.

605

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618 improved the quality of the manuscript.

619

620 **Figure Captions**

621 Figure 1: Location and oceanography

622 Position of cores MD01-2446 [39.06°N, 12.62°W, 3547 m], D11957P [39.05°N, 12.58°W, 3585
623 m], MD95-2042 [37.8°N, 10.17°W, 3146 m] and MD95-2040 [40.58°N, 9.86°W, 2465 m].

624 A) Digital elevation model from (Farr et al., 2007), bathymetry from IOC, IHO and BODC, 2003.
625 Blue dashed arrows indicate the main deep-water mass off Iberian margin, adapted from (van
626 Aken, 2000): Northeast Atlantic Deep Water (NEADW) and the Lower Deep Water (LDW),
627 southern-sourced water mass derived from Antarctic Bottom Water (AABW);

628 B) Upper ocean winter-time circulation (adapted from (Peliz et al., 2005), where: WIWiF -
629 Western Iberia Winter Front; IPC – Iberian Poleward Current; IPC-A – Alternative path of the
630 Iberian Poleward Current; PC – Portugal Current; GCNR – Gulf of Cadiz Northern Recirculation;
631 RAME – Recurrent Anticyclonic meander/eddy, Swoddies – Slope Water Oceanic Eddies;
632 Meddies – Mediterranean eddies; WIBP – Western Iberia Buoyant Plume.

633

634 Figure 2: Surface and deep water masses: offshore (MD01-2446 in red and D11957 in cyan)
635 versus nearshore (MD95-2042 in blue and MD95-2040 in green) of the Iberian margin: A-
636 Planktonic $\delta^{18}\text{O}$ (‰) VPDB (*G. bulloides*) from MD01-2446 (this study), MD95-2042 (Cayre et.al.
637 1999) and MD95-2024 (Abreu et al., 2003); B- SST °C (summer) based on foraminifera Simmax

638 Transfer Function of cores D11957 (Lebreiro et al., 1997), MD95-2042 (Cayre et al., 1999) and
639 MD95-2040 (Salgueiro et al., 2010); C- Benthic $\delta^{18}\text{O}$ (‰) VPDB of cores MD01-2446 (red, this
640 study) and MD95-2042 (blue, Shackleton et al., 2000); D- Benthic $\delta^{13}\text{C}$ (‰) VPDB record of cores
641 MD01-2446 (red, this study) and MD95-2042 (blue, Shackleton et al., 2000). Bold lines for each
642 curve indicate 3-point average smoothing.

643

644 Figure 3: Core MD01-2446: A- benthic $\delta^{13}\text{C}$ record; B- benthic $\delta^{18}\text{O}$ record; C- decay corrected
645 excess sedimentary ($^{231}\text{Pa}/^{230}\text{Th}$) activity ratio (Guihou et al., 2010); bold line indicates 3-point
646 average smoothing; D- paleomagnetic data ARM/IRM, $\kappa\text{ARM}/\kappa$ and Mr/Ms (D); E- mean of the
647 carbonate-free $<63\ \mu\text{m}$ is black colour (at right-axis) and SS% (percentage of sortable silt i.e.
648 $\%(10-63\ \mu\text{m})/ <63\ \mu\text{m}$) is grey colour (at left-axis).

649

650 Figure 4: A- Relative Sea Level Variation [RSL (m)] from (Waelbroeck et al., 2002); B- $\delta^{18}\text{O}$ bulk
651 sediment from core MD01-2446.

652

653 Figure 5: Productivity proxies from core MD01-2446: A- Opal content (weight %). Right-side axes
654 are the same parameter ^{230}Th normalized flux ($\text{g}\cdot\text{cm}^{-2}\text{ka}^{-1}$) (bold line); B- $\delta^{13}\text{C}$ difference (*G. bulloides*
655 - *C. wuellerstorfi*); C- organic carbon (weight %); D- CaCO_3 (weight %). Right-side axes are the same
656 parameter ^{230}Th normalized flux ($\text{g}\cdot\text{cm}^{-2}\text{ka}^{-1}$) (bold line); E- ^{230}Th normalized bulk flux ($\text{g}\cdot\text{cm}^{-2}\text{ka}^{-1}$)
657 (Guihou et al., 2010); F- Benthic $\delta^{13}\text{C}$ record.

658

659 Figure 6: Open ocean eddy-pumping versus coastal upwelling productivity.

660 A- Opal content (weight %) and Opal ^{230}Th normalized flux ($\text{g}\cdot\text{cm}^{-2}\text{ka}^{-1}$) (thick line) from core
661 MD01-2446; B- Export Production ($\text{gC}\cdot\text{m}^{-2}\cdot\text{yr}^{-1}$) from core MD95-2042 (thick black line) (Salgueiro
662 et al., 2010) and from core D11957 (grey thin line) (Lebreiro et al., 1997); C- Core D11957
663 (Lebreiro et al., 1997): Polar, and Gyre Margin (SIMMAX.28 Transfer Function); D- core D11957:
664 Primary Productivity SIMMAX.28; E- Core D11957 (Lebreiro et al., 1997): Upwelling
665 (SIMMAX.28).

666 Sea Surface Temperatures (SST), Paleoproductivity (PP) and factors FA20 (Polar, Gyre margin,
667 and Upwelling) were estimated through the transfer function SIMMAX28, based on modern
668 analogues (Modern Analogue Technique using a similarity index). For more details see
669 Pflaumann et al. (1996, 2003) and Lebreiro et al. (1997).

670

671 Supplementary Figures:

672 Supplementary Figure 1:
673 A large phytoplankton bloom off Iberia forming a bluish-green mantle in surface water. Credits
674 from Visible Earth (<http://visibleearth.nasa.gov>), part of EOS Project Science Office located at
675 NASA Goddard Space Flight Center [Image courtesy Jacques Descloitres, MODIS Land Rapid
676 Response Team at NASA GSFC].

677
678 Supplementary Figure 2: Age-depth plot of MD01-2446. Symbols correspond to tie-points of
679 correlation of *G. bulloides* oxygen isotopic records of each core (MD95-2040 and MD01-2446)
680 and the *G. bulloides* oxygen isotopic record of core MD95-2042.

681
682 Supplementary Figure 3:
683 Cross correlation between SS% and “Mean carbonate-free <63 μm fraction” from core MD01-
684 2446.

685
686 Supplementary Figure 4:
687 Relationship between opal (wt%) and $\delta^{13}\text{C}$, estimated through the Pearson correlation
688 coefficient r and computed for 152 samples.

689
690 Supplementary Figure 5:
691 Relationship between opal (wt%) and ^{230}Th -normalised opal flux estimated through the Pearson
692 correlation coefficient r and computed for 144 samples.

693
694 Supplementary Figure 6:
695 Sea Surface Temperature distributions off Iberian margin for winter (A- January) and summer
696 (B- July) 2009. Source: POET Data Viewer – AVHRR Pathfinder Version 5 data
697 (<http://poet.jpl.nasa.gov>).

698
699 Supplementary Figure 7:
700 Sea Surface Temperature ($^{\circ}\text{C}$) and Salinity (psu) distributions off Iberian margin for spring season
701 (April-June). Figure generated with Ocean Data View (see <http://odv.awi.de/>).with the World
702 Ocean Atlas 2009 (WOA09).

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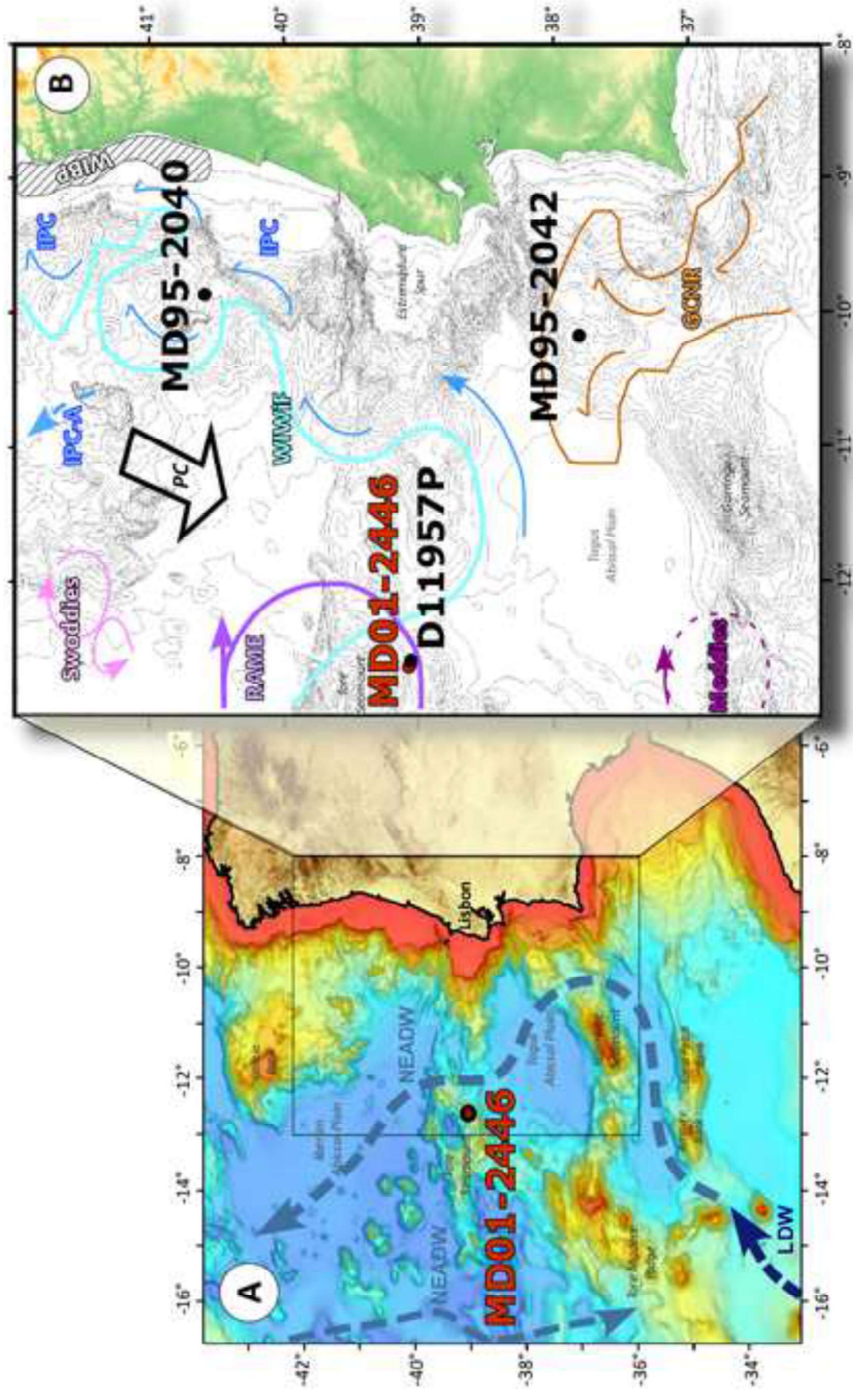
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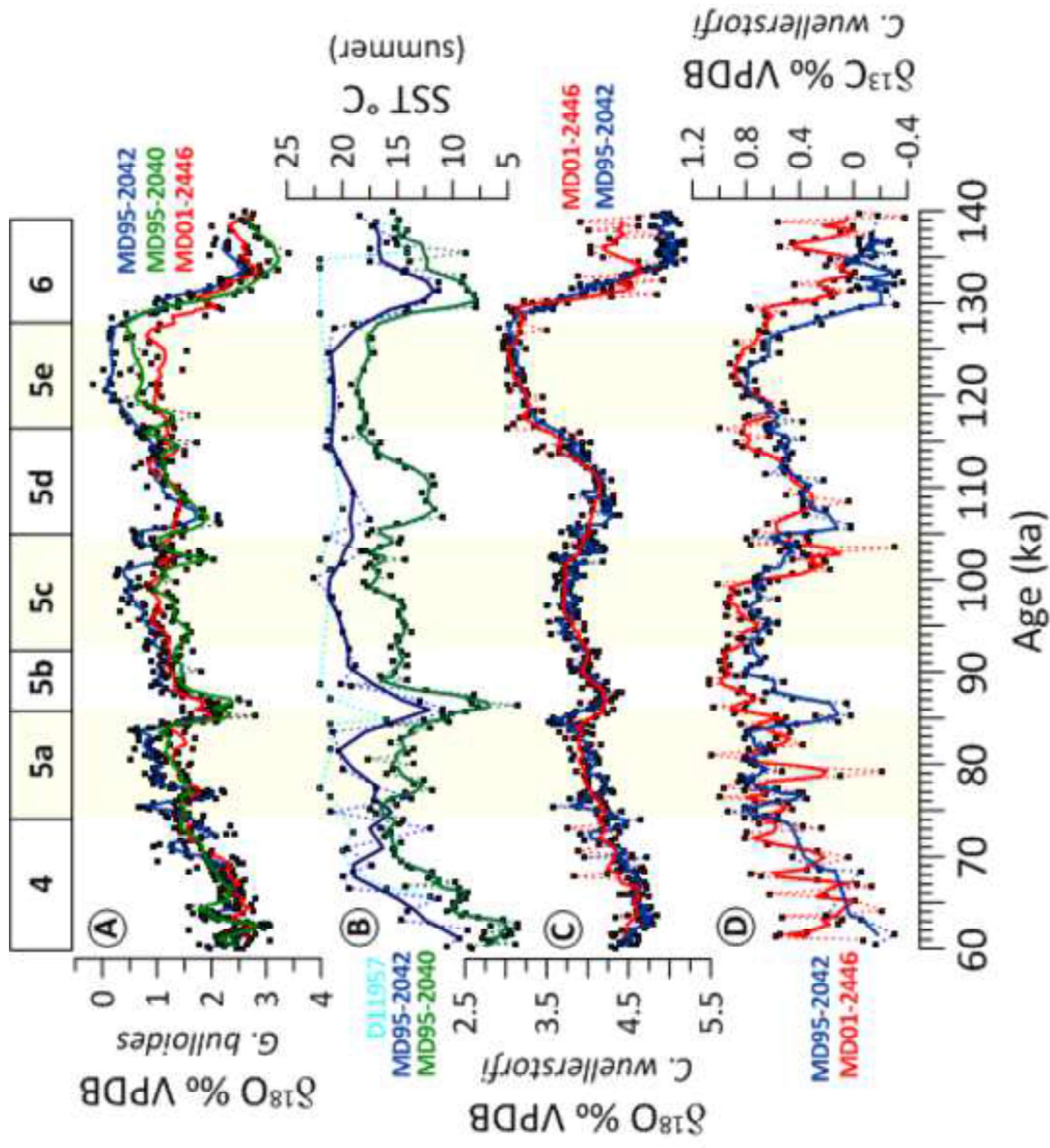
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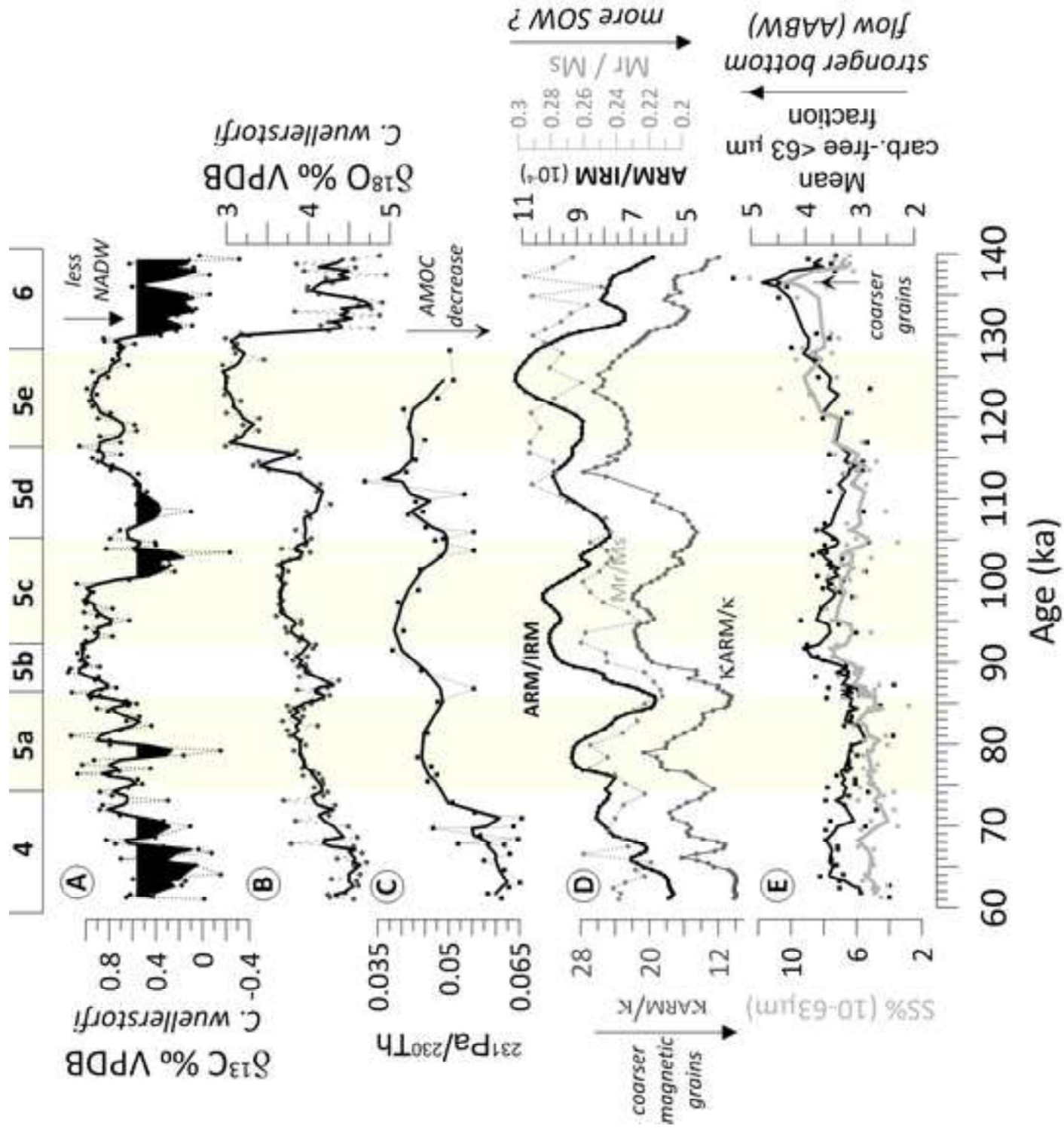
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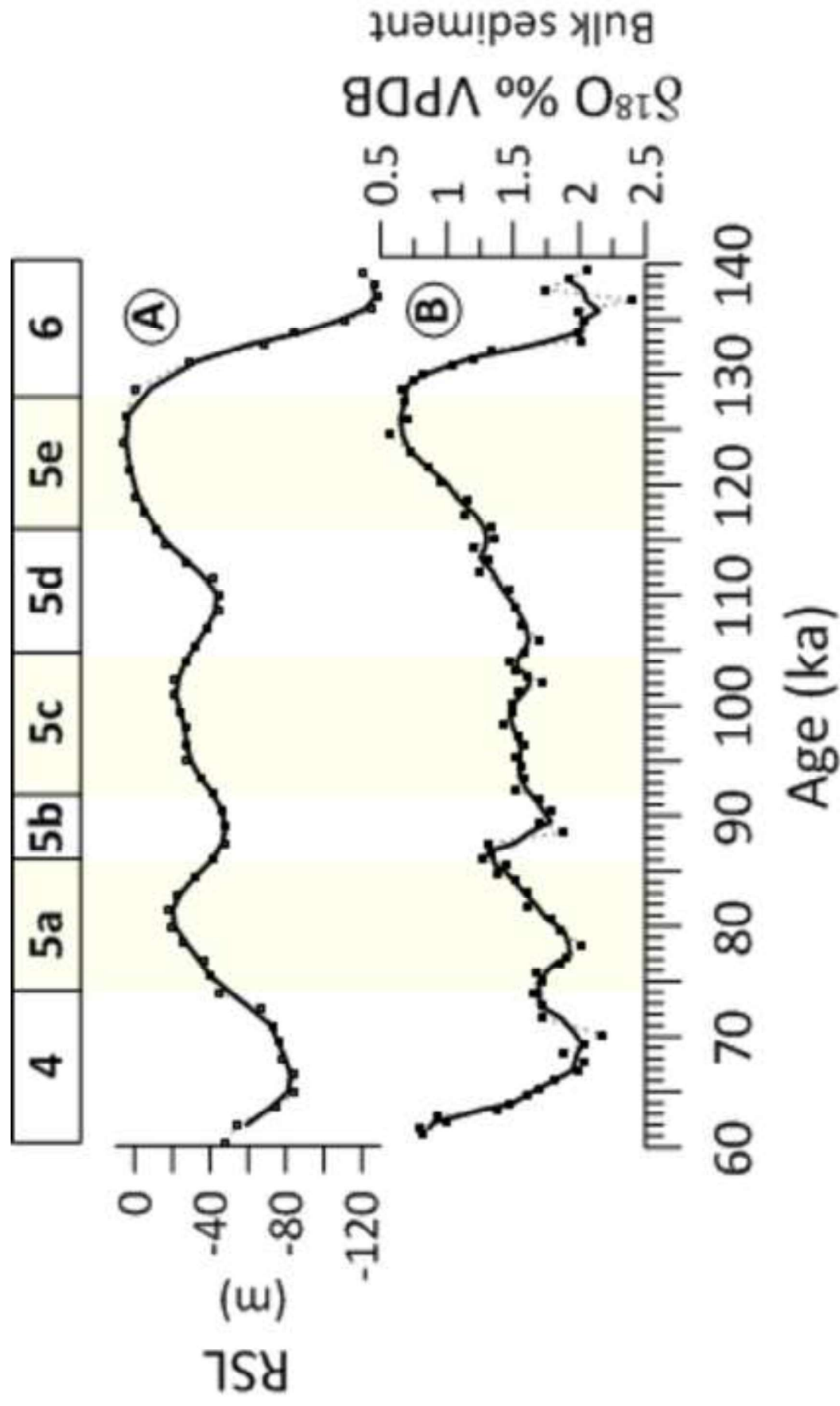
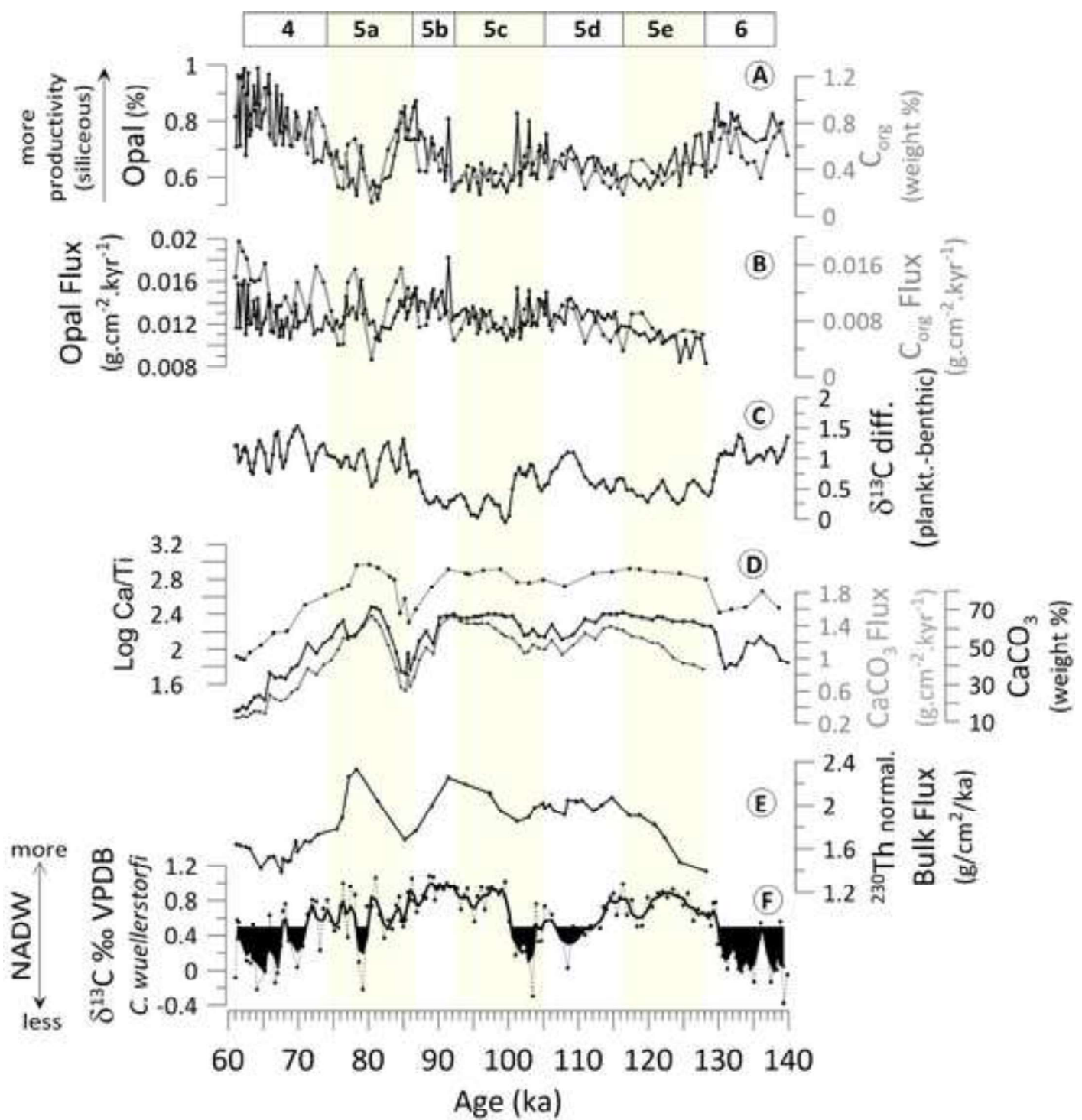
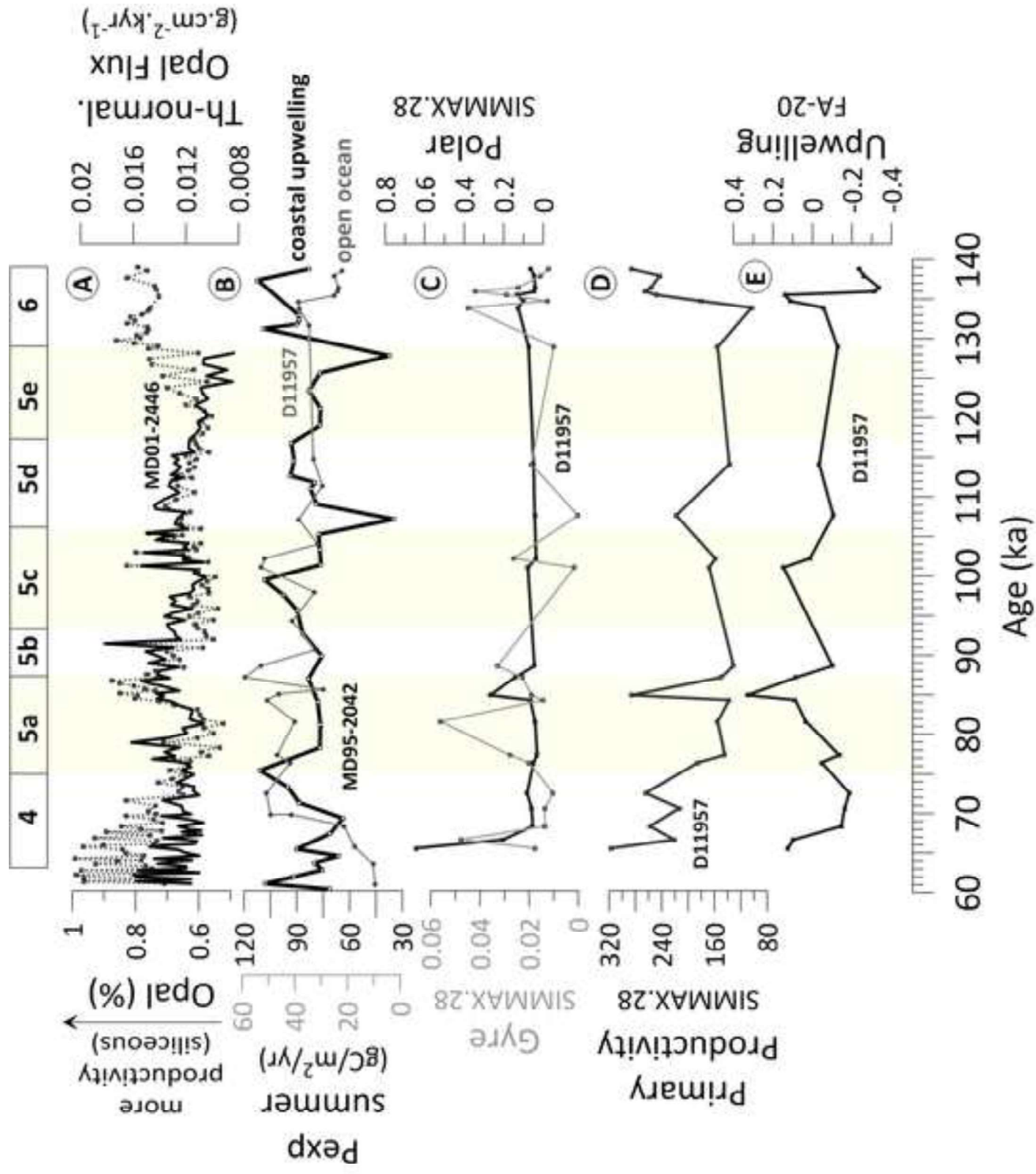
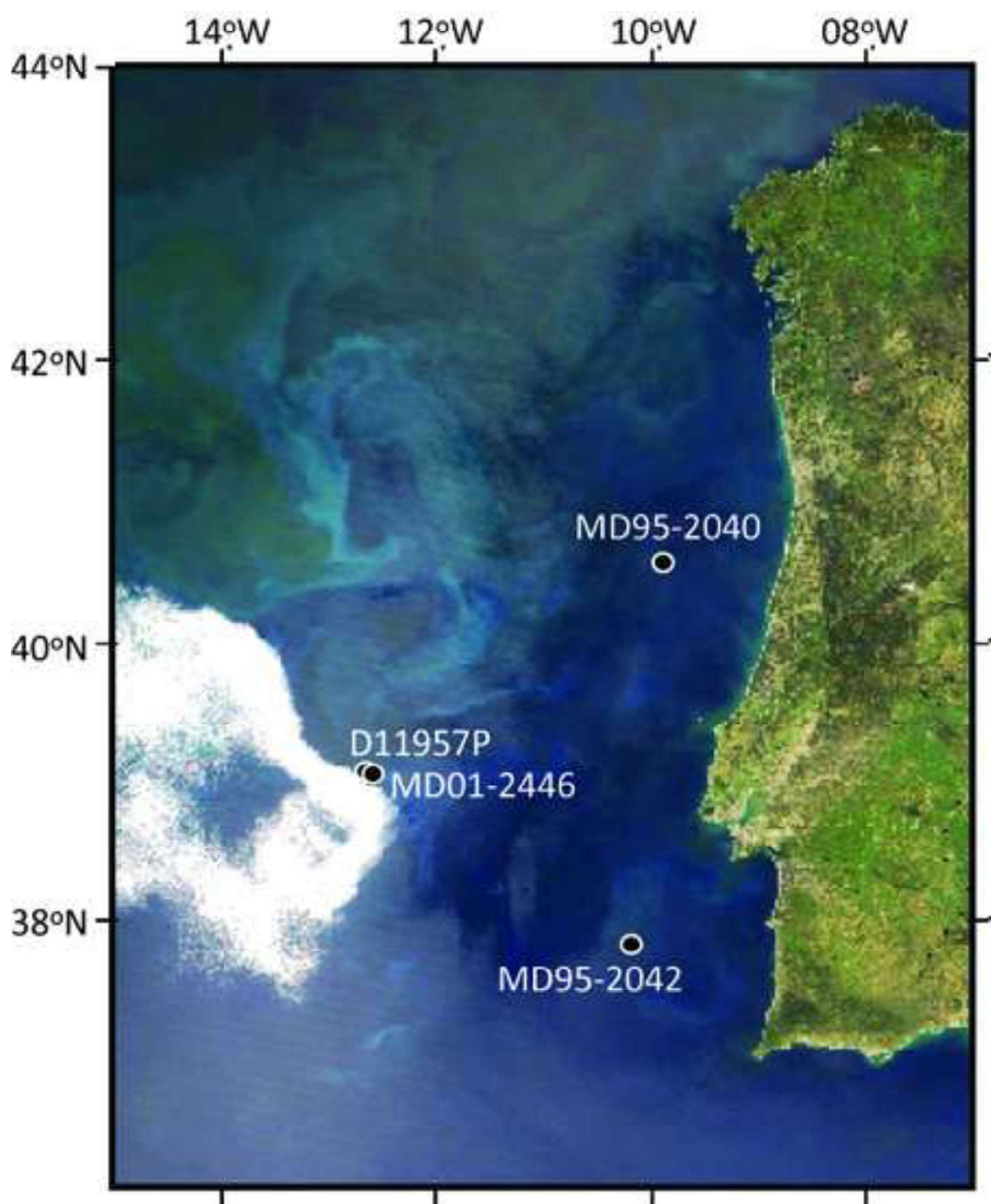


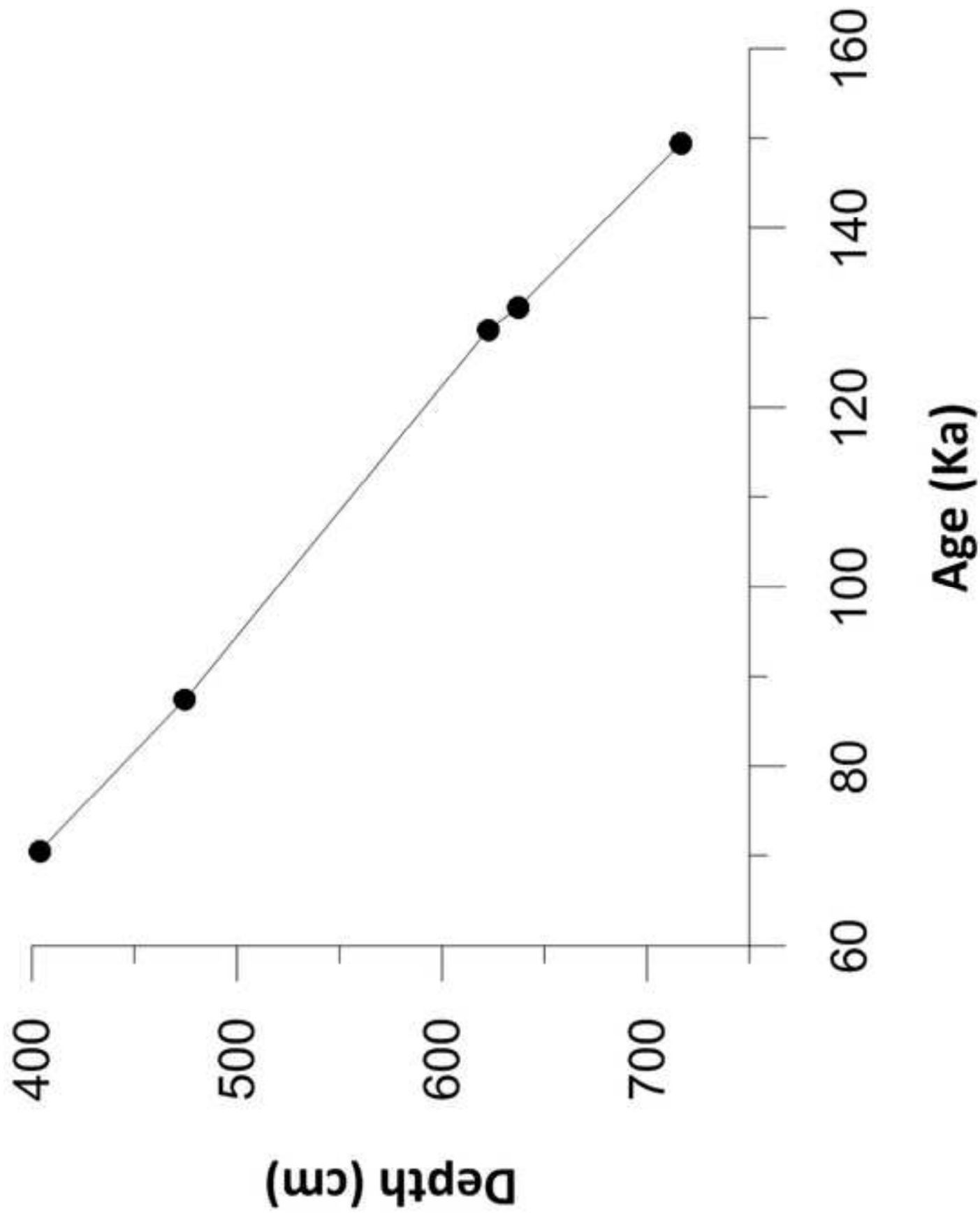
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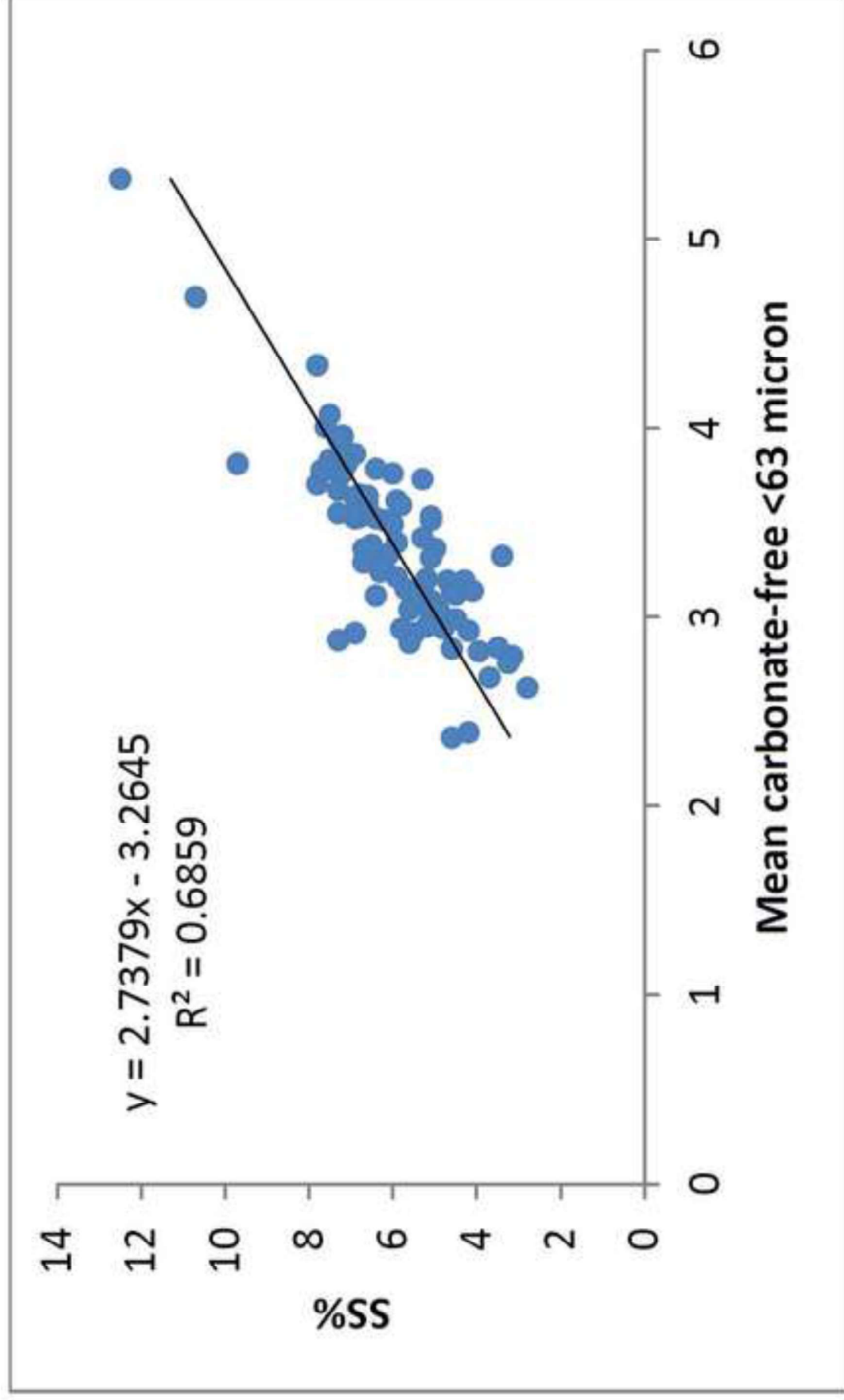
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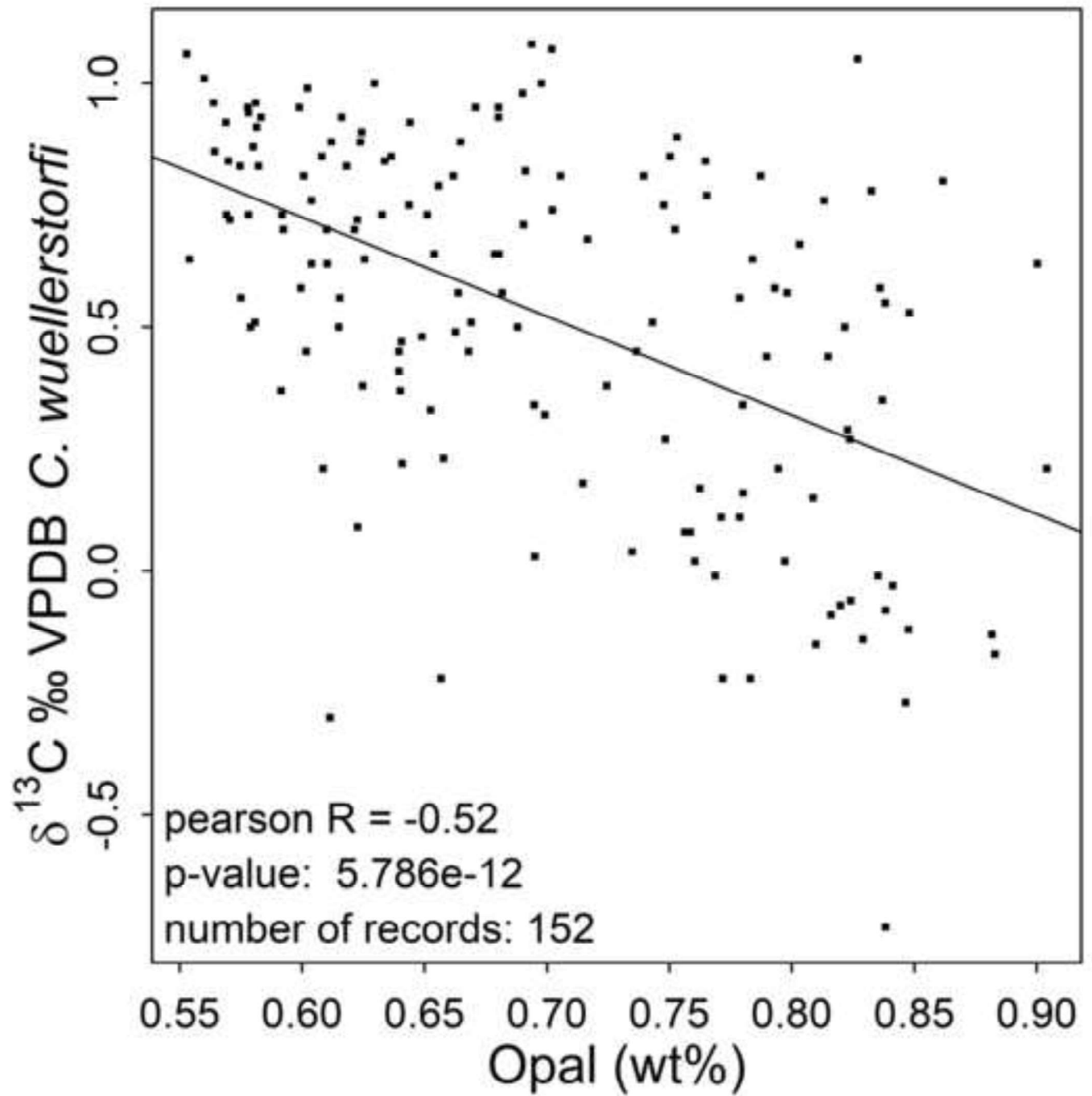


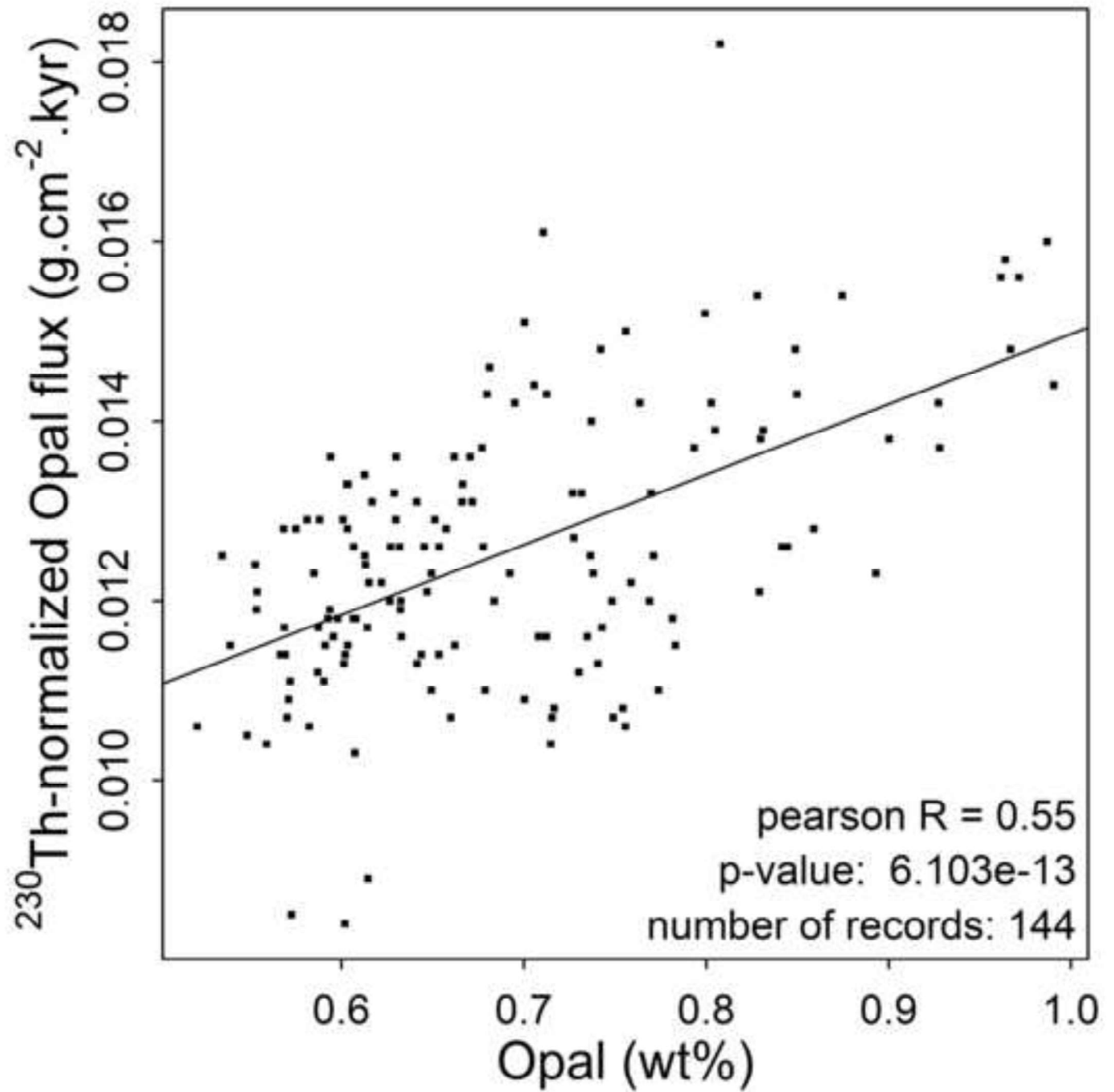


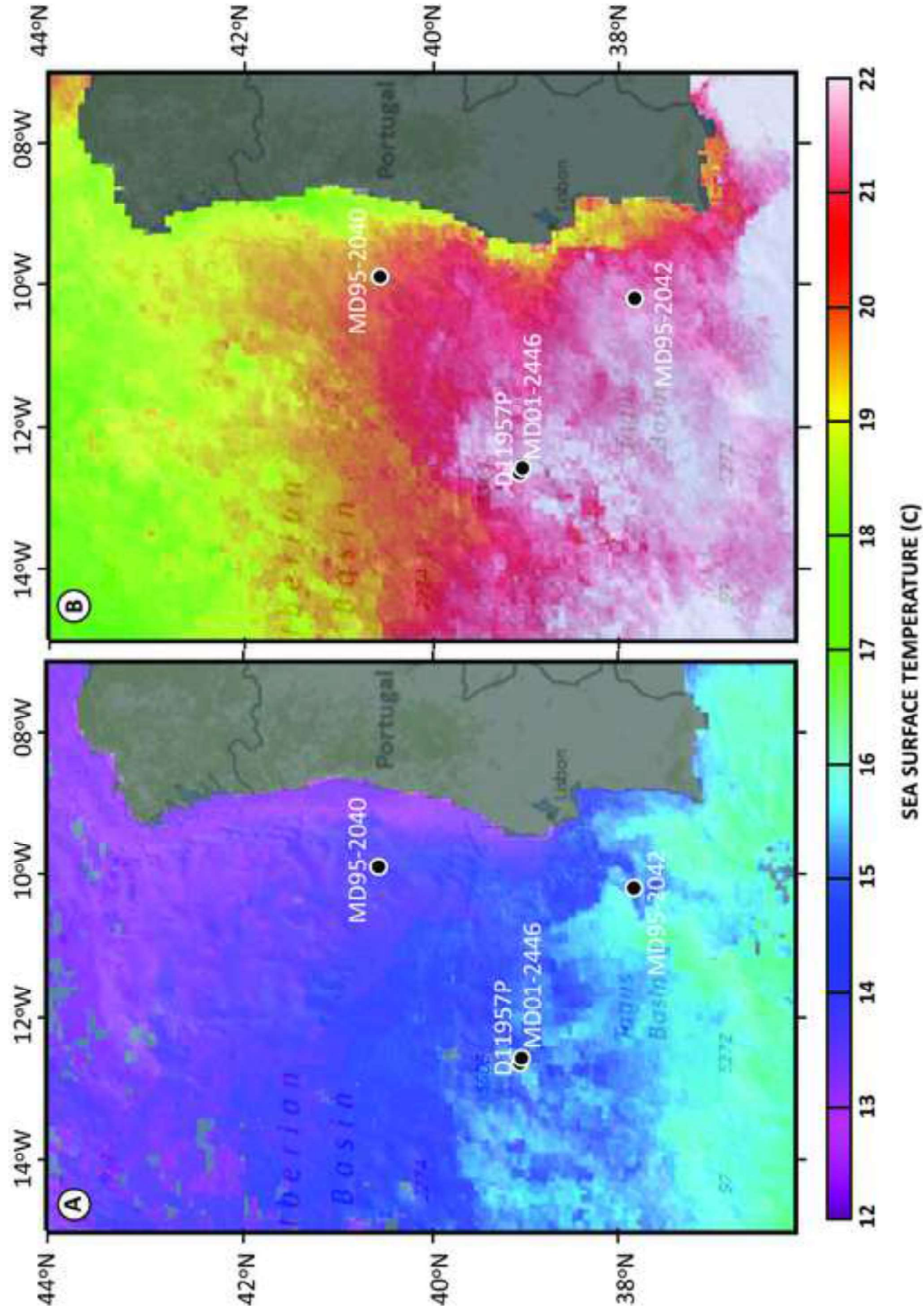












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