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## Enhanced Carbonate Counter Pump and upwelling strengths in the Indian sector of the Southern Ocean during MIS 11

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### Abstract :

While numerous studies have highlighted the central role of Southern Ocean (SO) dynamics in modulating rapid increases in atmospheric CO<sub>2</sub> concentrations during deglaciations, fewer studies have yet focused on the impact of the Biological Carbon Pump - and more specifically the Carbonate Counter Pump (CCP) - in contributing to increase the CO<sub>2</sub> concentration in oceanic surface waters and thus, in the atmosphere. Here, we present micropaleontological (coccolith, planktonic foraminifera) and geochemical (CaCO<sub>3</sub>, CaXRF, δ<sup>13</sup>CN, pachyderma) constraints from sediment core MD04-2718 retrieved in the Polar Front Zone of the Indian Ocean covering the time interval spanning Marine Isotope Stage (MIS) 12 to MIS 10 (440,000–360,000 years). We compare our results with published records from the SO to reconstruct past changes in CCP and upwelling dynamics and understand their leverage on the ocean-atmosphere partitioning of CO<sub>2</sub>. We demonstrate that the sharp increase in atmospheric pCO<sub>2</sub> during Termination V was likely associated with enhanced deep-water ventilation in the SO, that promoted the release of previously sequestered CO<sub>2</sub> to the ocean surface as the westerly wind belt and the frontal system migrated southwards. Enhanced CCP is observed later, during MIS 11, and is likely the consequence of higher sea surface temperature and higher nutrient availability due to the reinvigoration of SO upwelling leading to increased coccolith (and to a lesser degree, planktonic foraminifera) production and export. The low eccentricity signal recorded during MIS 11 might have additionally strengthened the CCP, exerting a specific control on *Gephyrocapsa* morphotypes. In addition to the strong global biological productivity and higher carbon storage on land, these synergistic mechanisms may have permitted to shape the distinctive 30 ka-long pCO<sub>2</sub> plateau characteristic of MIS 11.

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## Highlights

► Multiproxy approach to reconstruct Carbonate Counter Pump strength and upwelling dynamics. ► Increase in pCO<sub>2</sub> during Termination V coincides with enhanced upwelling. ► Strong Carbonate Counter Pump during MIS 11 in the Indian Southern Ocean. ► Enhanced upwelling and Carbonate Counter Pump helped shape the pCO<sub>2</sub> plateau.

**Keywords** : Southern Ocean, Marine Isotope Stage 11, Coccolith, Planktonic foraminifera, Geochemistry, Carbonate Counter Pump, Upwelling, Atmospheric CO<sub>2</sub> concentration

## 42 1. Introduction

43

44 The Earth's climate system of the past 800 000 years (800 ka) has been characterised by  
45 glacial-interglacial cycles that are tightly coupled to pCO<sub>2</sub> changes (Lüthi et al., 2008; Bereiter et al.,  
46 2015). Causes underlying these changes, especially in conjunction with deglaciations (glacial  
47 Terminations) that registered pCO<sub>2</sub> rises of about 50-100 ppm, are still under debate. It is recognised  
48 that complex, synergistic interactions involving changes in sea-ice cover, surface water stratification  
49 and marine productivity in the SO together with the dynamics of the westerly winds may have played  
50 a major role in modulating the efficiency of the physical and biological carbon pumps (François et al.,  
51 1997; Stephens and Keeling, 2000; Sigman and Boyle, 2000; Hodell et al., 2003; Toggweiler et al.,  
52 2006; Jaccard et al., 2013; Ferrari et al., 2014; Abelmann et al., 2015; Galbraith & Jaccard, 2015;  
53 Menviel et al., 2018; Gottschalk et al., 2019; Stein et al., 2020; Sigman et al., 2021). The  
54 reinvigoration of SO upwelling has been held responsible for a substantial part of the abrupt pCO<sub>2</sub>  
55 rise observed during Terminations (Hodell et al., 2003; Anderson et al., 2009; Skinner et al., 2010;  
56 Siani et al., 2013, Jaccard et al., 2016; Rae et al., 2018; Gottschalk et al., 2020; Sigman et al., 2021)  
57 while an increase in the efficiency of the Biological Carbon Pump (BCP) might have accounted for up  
58 to 40-50 ppmv CO<sub>2</sub> drawdown during glacial maxima (Hain et al., 2010; Martinez-Garcia et al., 2009;  
59 Jaccard et al., 2013). However, due to a lack of high-resolution primary productivity reconstructions  
60 and, more generally, to the scarcity of multiproxy approaches (Duncan et al., 2016; Duchamp-  
61 Alphonse et al., 2018), it remains difficult to apprehend the spatio-temporal patterns underpinning  
62 changes in the BCP and compare them to the SO upwelling dynamics to understand their impact on  
63 pCO<sub>2</sub>. Furthermore, most of the studies aiming to reconstruct changes in the strength and efficiency  
64 of the BCP focus on the export of phytoplankton-derived organic carbon to the deep ocean (i.e., the  
65 Soft Tissue Pump (STP)) that directly lowers atmospheric CO<sub>2</sub> (Martin et al., 1990; Martinez-Garcia et  
66 al., 2009; Martinez-Garcia et al., 2014; Thöle et al., 2019). They typically overlook the export of  
67 phytoplankton- and zooplankton-derived calcium carbonate (CaCO<sub>3</sub>), which raises surface water CO<sub>2</sub>

68 (i.e., the CCP) (Baumann et al., 2004; Barker et al., 2006; Broecker and Clark, 2009; Salter et al., 2014;  
69 Rembauville et al., 2016), and contributes to the CaCO<sub>3</sub>/organic carbon ratio in the biogenic particle  
70 flux to the sea-floor (namely the rain ratio), especially in the SO (Duchamp-Alphonse et al., 2018;  
71 Krumhardt et al., 2020). This omission may relate to the fact that calcareous phytoplankton is in part  
72 represented by coccoliths, which detailed analysis requires sophisticated tools such as SYRACO  
73 (SYstème de Reconnaissance Automatique des Coccolithes), a software using automated optical  
74 microscope that allows the robust determination of coccolith mass and thus, coccolithophore  
75 calcification rates, based on the optical properties (birefringence) of calcite under cross-polarized  
76 light (Beaufort et al., 2005; 2014). It is also probably linked to the lack of well-preserved carbonate  
77 fraction in sediment cores of the SO, making the reconstruction of biogenic carbonate export in the  
78 past challenging (Barker et al., 2006; Jaccard et al., 2013; Thöle et al., 2019).

79         Amongst the interglacials of the past 800 ka, MIS 11 is the longest (~30 ka) as well as one of  
80 the warmest (Tzedakis et al., 2022 and references therein). Following the cold glacial MIS 12, MIS 11  
81 begins at the end of Termination V (TV), the first Termination recording high amplitude increases in  
82 both pCO<sub>2</sub> (~ 90 ppmv) (Bereiter et al., 2015; Nehrbass-Ahles et al., 2020) and Antarctic air  
83 temperatures (> 12°C) (Jouzel et al., 2007). MIS 11 occurs during an eccentricity minimum (Laskar et  
84 al., 2004) and coincides with strong global biosphere productivity (Brandon et al., 2020), a major  
85 phase in coral reef expansion (Husson et al., 2018), and a climax in coccolithophore (*Gephyrocapsa*  
86 sp.) production (Barker et al., 2006; Beaufort et al., 2022). Recent studies document the increase in  
87 coccolith abundances over this time interval in the Pacific and Atlantic sectors of the SO (Flores et al.,  
88 2003; Saavedra-Pellitero et al., 2017a; 2017b). However, no studies exist in the Indian sector of the  
89 SO yet, and none focus on the CCP and its impact on atmospheric pCO<sub>2</sub> based on combined coccolith  
90 and planktonic foraminifera carbonate export data.

91         Here, we provide coccolith and planktonic foraminifera abundance and mass from core MD04-  
92 2718 retrieved in the Indian sector of the SO. Using these data, we demonstrate the biogenic origin  
93 of the sedimentary CaCO<sub>3</sub> at this site and calibrate the carbonate signal to reconstruct high  
94 resolution changes in CCP strength across the time interval covering MIS 12 – MIS 10 (440-360 ka),  
95 including TV and MIS 11. Combined with reconstructed summer SST and  $\delta^{13}\text{C}_{N. pachyderma}$  data obtained  
96 from the same core, ~~as well as~~ previously published micropaleontological and geochemical data from  
97 the Pacific and Atlantic sectors of the SO as well as geochemical data from EPICA Dome C (EDC), our  
98 study further highlights the synergistic contribution of changing global productivity and terrestrial  
99 carbon storage, SO ventilation and CCP strength on atmospheric CO<sub>2</sub> during this key period of the  
100 past 800 ka.

101

## 102 2. Regional oceanic setting

103

104 The SO is delimited by the Antarctic continent to the South and by the Subtropical front (STF)  
105 to the North (Fig. 1a). The eastward-flowing Antarctic Circumpolar Current (ACC) connects the  
106 Atlantic, Pacific and Indian sectors of the SO. This current, driven by Coriolis and the Southern  
107 Hemisphere westerlies, is characterised by the largest mass transport of all ocean currents, averaging  
108 150 Sv ( $10^6\text{m}^3\text{s}^{-1}$ ) in the Indian Ocean (Park et al., 2009). Its Northern and Southern Boundaries (NB  
109 and SB) are often delimited around 40°S and 60°S respectively, but their exact location depends on  
110 bottom topography (Sokolov and Rintoul, 2009, Park et al., 2019) and varies latitudinally (Freeman et  
111 al., 2016). Three hydrological fronts generally characterised by stronger flows, are defined in the  
112 ACC: the Subantarctic Front (SAF), the Polar Front (PF) and the Southern Antarctic Circumpolar Front  
113 (SACCF) (Fig. 1). They help defining the Subantarctic Zone (SAZ) between the STF and the SAF, the  
114 Polar Front Zone (PFZ) between the SAF and the PF, and the Antarctic Zone (AAZ) delimited by the PF  
115 to the North, and the SB to the South (Orsi et al., 1995, Park et al., 2019). Within the Indian sector of  
116 the SO, the Kerguelen Plateau that extends from ~ 46°S to 63°S and between ~ 62°E and 85°E, is the  
117 second largest submarine plateau in the world. Due to its great meridional extent and topography, it  
118 constitutes a major bathymetric constrain to the ACC flow and thus exerts a significant impact on  
119 both ACC transport and front patterns (Park et al., 2009; Sokolov and Rintoul, 2009). In this context,  
120 the PF is located alongside the southern coast of the Kerguelen archipelago (Park et al., 2014). The  
121 circulation in the Southern Ocean is also regulated by the upwelling of CO<sub>2</sub>- and nutrient-rich  
122 Circumpolar Deep Waters (CDW) that resurface at the Antarctic divergence in the AAZ (Fig. 1c). These  
123 waters represent the source for the Subantarctic Mode Water (SAMW) and the Antarctic  
124 Intermediate Water (AAIW) that flow northward. Today, the change in the strength of upwelling  
125 affects the partitioning of CO<sub>2</sub> between the ocean subsurface and the atmosphere on decadal  
126 timescales (Lovenduski et al., 2008; Landschützer et al., 2015; Gruber et al., 2019).

127 In the SO, primary productivity is primarily steered by bottom-up control involving both  
128 physical and biogeochemical processes, which are spatially and temporally heterogeneous. Today,  
129 the presence/absence of sea ice together with the progressive temperature decrease polewards  
130 subdivide the SO into distinct provinces characterised by specific functional assemblages (Boyd,  
131 2002; Constable et al., 2014). In ecosystems that remain ice-free throughout the year, i.e., within the  
132 Permanently Open Ocean Zone (POOZ), primary productivity is commonly co-limited by light and  
133 micronutrient availability (Boyd et al., 2002; Moore et al., 2013). While the residual surface ocean  
134 macronutrient concentrations in the SO are among the highest globally (Pollard et al., 2006) largely  
135 due to the large-scale upwelling of CDW at the Antarctic divergence, primary productivity is relatively  
136 low, owing to the scarcity of bioavailable iron (Fe) imposes on phytoplankton growth (Martin et al.,

137 1990; Pollard et al., 2009). In all cases, phytoplankton that flourishes south of the PF is dominated by  
138 diatoms, with blooms generally occurring during austral summer (Rigual-Hernandez et al., 2015).  
139 Phytoplankton that thrives north of the PF is more heterogeneous and represented by nano- and  
140 picoplankton of different origins. One of the most striking patterns associated with phytoplankton in  
141 this area, is the significant coccolithophore blooms that occur during spring and summer between  
142 ~40 and ~60°S, and that forms the so-called Great Calcite Belt that enriches surface waters in CaCO<sub>3</sub>  
143 (Balch et al., 2011; 2016; Rembauville et al., 2016). Within the Indian sector of the SO, near the  
144 Kerguelen plateau, this Great Calcite Belt is observed in the PFZ (Smith et al., 2017) and the northern  
145 part of the AAZ (between Kerguelen and Heard and McDonald Islands) and may contribute ~ 60 to 85  
146 % of the CCP (Blain et al., 2007; Rembauville et al., 2016; Smith et al., 2017; Rigual-Hernandez et al.,  
147 2020). Relatively high primary productivity may also be observed locally, near the PF and the Crozet  
148 and Kerguelen plateaus, where oceanic current and/or bottom topography enriches surface waters  
149 with iron (Blain et al., 2007; Sokolov and Rintoul, 2007; Pollard et al., 2009; Quéguiner et al., 2013;  
150 Tagliabue et al., 2014).

151

## 152 3. Material and methods

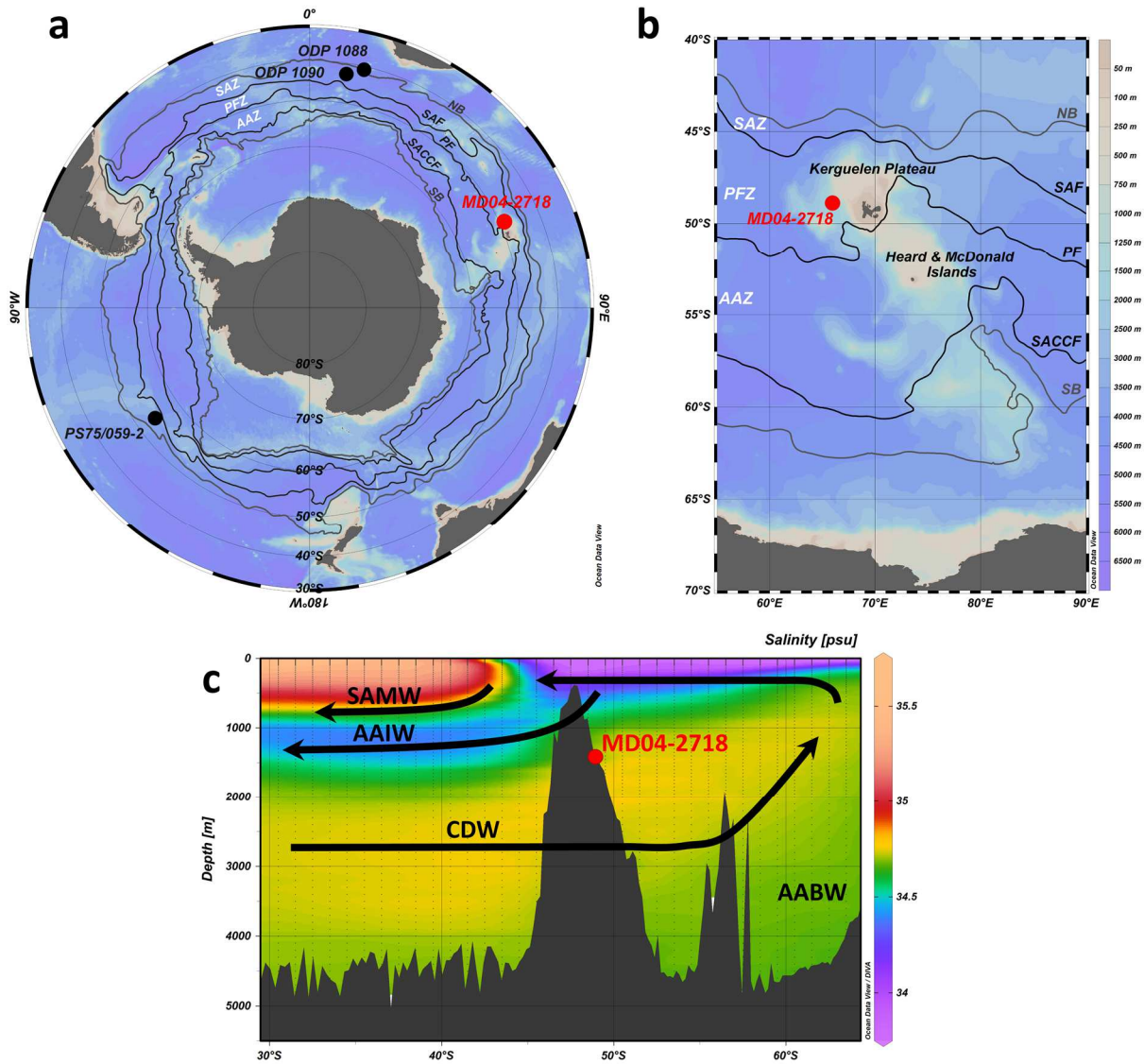
153

### 154 3.1 Material core

155

156 The 50-meter-long marine sediment core MD04-2718 (48°53,31S; 65°57,42E, 1428 m water  
157 depth) was retrieved from the NW flank of the Kerguelen Plateau in 2004, during the MD136/VIGO  
158 expedition aboard the R.V. Marion Dufresne (Fig. 1, Michel and Waelbroeck, 2017) within the PFZ.  
159 Surface water properties are largely controlled by CO<sub>2</sub>- and nutrient- rich waters upwelling in the  
160 Antarctic Zone, that are advected northward by the Ekman transport. Sediment recovered at this site  
161 is primarily siliciclastic with alternations of nannofossil (calcareous) and diatom (opal) oozes. MIS 11  
162 is clearly identifiable and represented by calcareous-rich ooze, between 11.5 and 13.1 m core depth.

163



164  
 165 Figure 1 Core location and study area. **a, b** Location of the studied core (MD04-2718, red dot) as well as cores from previous  
 166 studies (ODP sites 1088 and 1090 (Hodell et al., 2003) and PS75/059-2 (Saavedra-Pellitero et al., 2017a; 2017b) (black dots).  
 167 Oceanic fronts are from Park et al. (2019). NB: Northern Boundary of the ACC, SAZ: Subantarctic Zone, SAF: Subantarctic  
 168 Front, PFZ: Polar Front Zone, PF: Polar Front, AAZ: Antarctic Zone, SACCF: Southern Antarctic Circumpolar Front, SB:  
 169 Southern Boundary of the ACC. **c** Salinity (psu) section at 66°E using WOA18 (Zweng et al., 2018) on ODV software  
 170 (Schlitzer, 2018). SAMW: Subantarctic Mode Water, AAIW: Antarctic Intermediate Water, CDW: Circumpolar Deep Water,  
 171 AABW: Antarctic Bottom Water.

### 172 3.2 Geochemical analyses

173

174 Planktonic foraminifera  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  compositions (expressed in ‰) were measured at the  
 175 LSCE (Laboratoire des Sciences du Climat et de l'Environnement), based on 67 *Neogloboquadrina*  
 176 *pachyderma* senestral (200-250  $\mu\text{m}$  fraction, Supplementary Fig. 1) samples, subsampled every 3-4  
 177 cm between 11.03 and 13.3 m depth (except for 13.07, 13.10 and 13.13 m, which are barren of *N.*  
 178 *pachyderma* s). For each sample, 30 specimens were hand-picked, cleaned using methanol,



179 ultrasonicated, weighted, and analysed using an IRMS GV ISOPRIME mass spectrometer coupled with  
180 an extraction line. The measurements are reported versus the Vienna Pee Dee Belemnite standard  
181 (VPDB) with respect to NBS-19 standard, with a mean external reproducibility ( $1\sigma$ ) for carbonate  
182 standards of  $\pm 0.04$  ‰ and  $0.06$  ‰ for  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$ , respectively. The reproducibility of isotopic  
183 measurements on *N. pachyderma* s. are  $0.06$  and  $0.1$  ‰, for  $\delta^{13}\text{C}$  and  $\delta^{18}\text{O}$ , respectively.

184 High-resolution, relative sedimentary calcium (Ca) concentrations were measured every 1 cm using  
185 an Aavatech profiling X-ray Fluorescence (XRF) core scanner at ETH Zurich, Switzerland with a 10s  
186 integration at 10 kV and 500  $\mu\text{A}$ . In parallel, sedimentary  $\text{CaCO}_3$  concentrations were determined on  
187 26 samples at GEOPS Laboratory with a  $\pm 2$  % precision, using the vacuum-gasometric technique  
188 described in Duchamp-Alphonse et al., (2018) where the amount of  $\text{CO}_2$  degassed during the reaction  
189 of  $100 \text{ mg} \pm 2 \text{ mg}$  of dried sediments with 6N HCl is converted into  $\text{CaCO}_3$  percentages. Comparison  
190 between Ca counts and discrete  $\text{CaCO}_3$  contents show a highly significant statistical correlation  
191 ( $r^2=0.97$ ), and  $\text{Ca}_{\text{XRF}}$  signal is therefore directly converted into  $\text{CaCO}_3$  content (Jaccard et al., 2009;  
192 2010; Duchamp-Alphonse et al., 2018; Supplementary Fig. 2).

193

### 194 3.3 Micropaleontological analyses

195

#### 196 3.3.1 Coccoliths

197

198 Coccolith abundance and mass were determined at the GEOPS (GEOscience Paris Saclay)  
199 Laboratory. Smear slides were prepared for 68 samples, based on the protocol detailed in Duchamp-  
200 Alphonse et al., (2018), after Beaufort et al., (2014). For each sample, the abundance and mass of  
201 individual coccoliths (with an average of 6,172 coccoliths per sample (176 to 13,233)) were  
202 automatically determined with the help of the SYRACO software using a Leica DM6000B microscope.  
203 SYRACO detects and classifies most of the coccoliths (*Gephyrocapsa caribbeanica*, small  
204 *Gephyrocapsa*, *G. oceanica*, *G. muelleriae*, *Calcidiscus leptoporus* and *Coccolitus pelagicus*) present on  
205 the slides. Coccolith mass was also directly obtained using SYRACO based on a quasi-linear  
206 relationship between their brightness (birefringence in grey scale) and their thickness under cross-  
207 polarized light. Because this method applies to coccoliths thinner than  $1.55 \mu\text{m}$ , we interpret only  
208 thickness and mass data for *Gephyrocapsa* specimens, as *C. leptoporus* and *C. pelagicus* display third-  
209 order interferences. The presence of relatively large detrital particles may lead to an underestimation  
210 of small specimens including *Gephyrocapsa* sp. and an overestimation of larger ones including *C.*  
211 *pelagicus*. To correct for these potential biases, the abundance of each morphotype was verified  
212 manually on 18 samples covering the full spectrum of coccolith abundance observed from MIS 12 to

213 MIS 10. Visual counts were performed on the same view fields analysed by SYRACO and were  
214 compared to the results obtained by the software. *Gephyrocapsa* sp. abundances from SYRACO  
215 counts were calibrated using the correlation coefficients between the two datasets, while *C.*  
216 *pelagicus* abundances have been obtained visually for the entire sample set because of the confusion  
217 made by SYRACO between detrital particles and this species. *C. leptoporus* abundances did not differ  
218 significantly between SYRACO and visual counts were not corrected.

219

220 The number of coccoliths per gram of sediment (absolute coccolith abundance – ACA) was  
221 determined for each morphotype/species using the equation of Grelaud et al., (2009) as follows:

$$222 \quad ACA = \frac{N_c * S_f}{N_o * S_o * W_s}$$

223 where  $N_c$  represents the total coccolith number per sample,  $S_f$ , the flat beaker area (3117 mm<sup>2</sup>),  $N_o$ ,  
224 the number of view fields (165),  $S_o$ , the view field area (0.01 mm<sup>2</sup>) and  $W_s$ , the dry sediment weight.

225

226 The coccolith calcite mass per gram of sediment (per g. sed) was determined using the same  
227 equation, where  $N_c$  is replaced by  $M_c$ , that represents the coccolith calcite mass per view fields.  
228 *Gephyrocapsa* calcite mass ( $CM_{Gephy}$ ) was obtained based on the estimation of specimen abundance  
229 and mean mass as follows:

$$230 \quad CM_{Gephy} = ACA * mean\ mass_{Gephy}$$

231

232

### 233 3.3.2 Planktonic foraminifera

234

235 The abundance and planktonic foraminifera species composition were determined at the  
236 LSCE with a binocular microscope, using the > 150 µm fraction of 57 samples ranging between 11.30  
237 m and 13.3 m depth. For each sample, an average of 450 planktonic foraminifera were counted. The  
238 absolute planktonic foraminifera abundance (AFA) per gram of sediment was determined as follows:

$$239 \quad AFA = \frac{N_f * 2^{split}}{W_{dry}}$$

240 where  $N_f$  represents the number of planktonic foraminifera counted per sample (> 300), split is the  
241 number of splits performed on the sediment before counting ( $0 \leq split \leq 6$ ), and  $W_{dry}$ , the dry  
242 sediment sample weight.

243

244 The planktonic foraminifera calcite mass was determined at the LSCE using mean weights of the  
245 major planktonic foraminifera species from the >150 µm fraction. The foraminifera calcite mass is

246 thus underestimated as it does not consider juveniles planktonic foraminifera or smaller species such  
 247 as *G. uvula*. Yet the weight variations of this fraction are considered representative of the weight  
 248 variations of the total planktic foraminifera fraction. The calcite mass produced by planktonic  
 249 foraminifera,  $CaCO_{3plank.foram}$ , is (in mg/g):

$$250 \quad CaCO_{3plank.foram} = \frac{(\sum_i (\sum_j m_{ji} * size_{ji}) * X_i) * AFA}{10^7}$$

251 Where  $m_{ji}$  represents the mean weight in size range  $j$  of specie  $i$  (in  $\mu g$ ),  $size_{ji}$  the percentage of  
 252 population in size range  $j$  for specie  $i$ , and  $X_i$  the percentage of the specie  $i$  in the sample. The error  
 253 on the mass will be:

$$254 \quad \frac{1}{10^7} * (\sum_i X_i * (\sum_j size_{ji} * err_{m_{ij}} + \sum_j err_{size_{ji}} * m_{ij}) + \sum_i err_{X_i} * (\sum_j size_{ji} * m_{ji}))$$

255  
 256 where  $err$  is the error on each of the variables.

257  
 258 The size distribution, considering eight intervals (150-200  $\mu m$ , 200-250  $\mu m$ , 250-315  $\mu m$ , 315-355  
 259  $\mu m$ , 355-400  $\mu m$ , 400-450  $\mu m$ , 450-500  $\mu m$  and >500  $\mu m$ ), and the mean weight for each size has  
 260 been determined at six depths for the eight species representing > 99 % of the assemblage. The  
 261 mean weight has been determined from 15 weighings of 1 to 4 individuals (to reach weight between  
 262 10 to 40  $\mu g$ ), to decrease the weight error to 19 %  $[(1/(2(n-1)))^{0.5}]$ . The eight species include  
 263 *Neogloboquadrina pachyderma* sinistral and dextral, *Globigerinita glutinata*, *Turborotalia*  
 264 *quinqueloba*, *Globigerina bulloides*, *Globorotalia inflata*, *Globorotalia crassaformis* and *Globorotalia*  
 265 *truncatulinoides*. We determine mean weights of :  $3.55 \pm 0.32 \mu g$  for *N. pachyderma*,  $9.42 \pm 2.15 \mu g$   
 266 for *G. bulloides*,  $3.48 \pm 0.42$  for *G. glutinata*,  $13.9 \pm 2.37 \mu g$  for *G. crassaformis* and  $22.4 \pm 5.39 \mu g$  for  
 267 *G. inflata*. We assumed a similar weight to *G. glutinata* for *T. quinqueloba*. For *G. truncatulinoides*,  
 268 we assumed a mean weight similar to *G. crassaformis*. For *N. pachyderma*, 90 to 97 % of the  
 269 population ranges in the minimum measured size range (150-200  $\mu m$ ), and between 71 to 100 % for  
 270 *G. glutinata* and *T. quinqueloba*. For the other species the size range distribution varies with depth  
 271 with the main size range varying from 150-200  $\mu m$  to 250-315  $\mu m$  for *G. bulloides*, *G. crassaformis*  
 272 and *G. truncatulinoides* and from 150-200  $\mu m$  to 315-355  $\mu m$  for *G. inflata*. The error on the calcite  
 273 mass of the planktonic foraminifera depends thus mainly on the error on the population fraction in  
 274 the different size range for the larger species. The error on the calcite mass of the planktonic  
 275 foraminifera also considers the variations in planktonic foraminifera weight and size with depth  
 276 observed in the core, that can be linked with the changes in atmospheric  $CO_2$  concentration (Barker  
 277 and Elderfield, 2002; Moy et al., 2009; Henehan et al., 2017). For the 57 samples, the calcite mass of  
 278 the planktonic foraminifera and its error have been calculated from the counted abundance of the

279 different species and their mean size and mean weight calculated from the six depths. The error on  
280 this calcite mass varies from 26 to 57 % and is linearly linked with the abundance of *G. bulloides* ( $r^2=$   
281 0.91). Although large, the error remains negligible compared to the observed variations that are  
282 almost a factor of a thousand.

283  
284 Summer SST estimates were determined on the same 57 samples as those used for planktonic  
285 foraminifera assemblages based on two independent approaches: the Modern Analog Technique and  
286 the percentage of *N. pachyderma* s. The former one relies on the use of a database of modern  
287 foraminifera assemblage (Haddam et al., 2016) and the PaleoAnalog software (Therón et al., 2004) to  
288 find the closest modern analogues for the fossil assemblage (Prell, 1985; Haddam et al., 2016). The  
289 latter one is based on the linear relation between SST and the percentage of *N. pachyderma* s. in the  
290 total planktonic foraminifera assemblage following the equation of Govin et al. (2009):

$$291 \quad T^{\circ}\text{C} (\% \text{ pachy s.}) = -0.087 (\% \text{ pachy s.}) + 11.339$$

292 No SST could be determined between 13.13 and 13.07 m depth as the number of planktonic  
293 foraminifera was insufficient (<300) for a robust temperature estimate. During glacial periods the  
294 number of foraminifera per gram of sediment decreases drastically and, in these three levels, the  
295 >150  $\mu\text{m}$  fraction contains almost exclusively ice rafted detritus and radiolarians.

296

## 297 4. Results

298

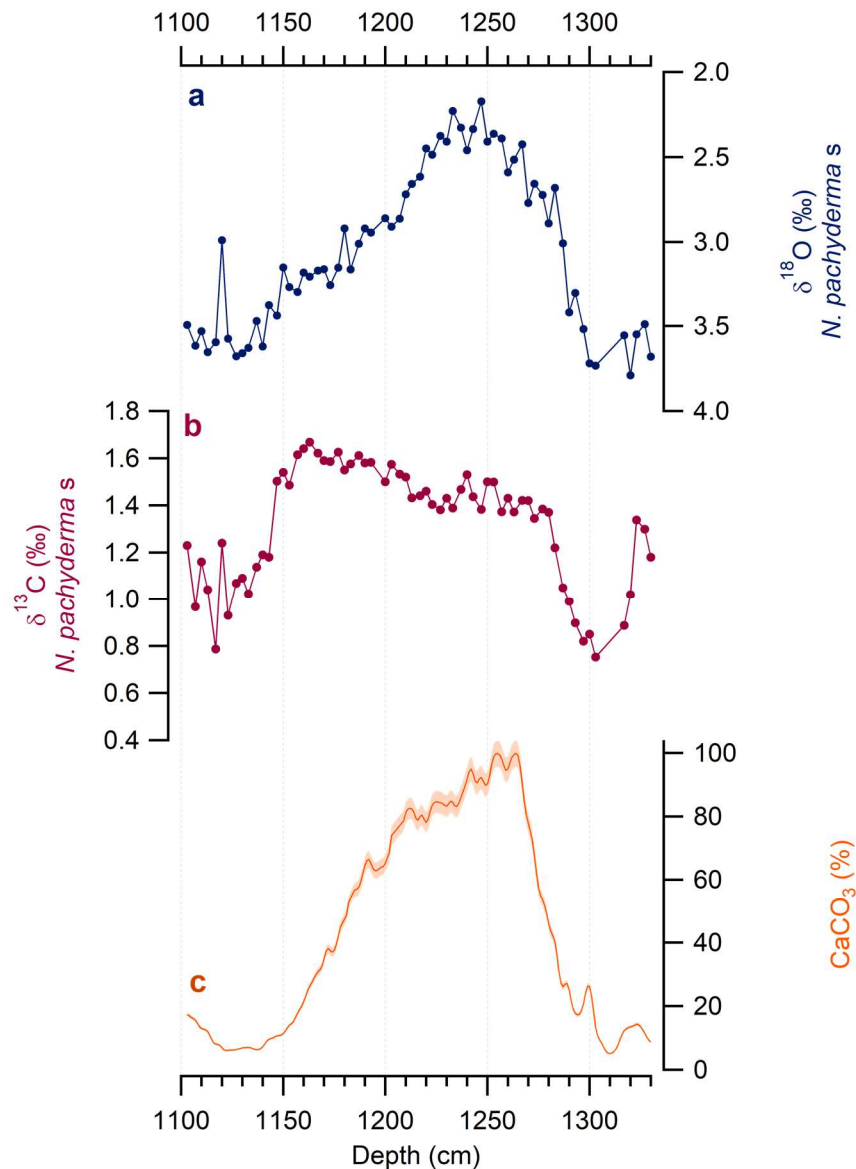
### 299 4.1 Geochemical signals

300

301 Planktonic foraminifera  $\delta^{18}\text{O}$  values range between 2.2 ‰ and 3.8 ‰ (Fig. 2a). After a rapid  
302 decrease from 3.7 to 2.2 ‰ between 13.00 and 12.33 m, values gradually increase to reach around  
303 3.6 ‰ at 11.33 m. The  $\delta^{13}\text{C}$  signal ranges from 0.75 ‰ to 1.67 ‰ (Fig. 2b).  $\delta^{13}\text{C}$  values are  
304 characterised by a slightly increasing plateau between 12.80 and 12.23 m and two sharp decreases to  
305 the minimum value of 0.8 ‰ from 13.23 to 13.03 m and from 11.57 to 11.33 m.

306 Sedimentary  $\text{CaCO}_3$  concentrations range between 5 % and 99 % (Fig. 2c). Values never  
307 exceed 20 % between 13.30 and 13.10 m and between 11.30 and 11.03 m, while they average 85 %  
308 between 12.67 and 11.90 m. The most striking trends are observed between 13.10 and 12.66 m and  
309 12.03 and 11.41 m, where  $\text{CaCO}_3$  concentrations undergo a drastic increase to maximum values of 99  
310 % at 12.65 m and a two-fold decreasing trend to reach minimum values of 7 % at 11.38.

311



312

313 Figure 2 Bulk carbonate signals and stable isotope records of planktonic foraminifera *N. pachyderma s.* from sediment core  
 314 MD04-2718. **a**  $\delta^{18}\text{O}_{N. pachy s}$  (‰), **b**  $\delta^{13}\text{C}_{N. pachy s}$  (‰), **c**  $\text{CaCO}_3$  (%) derived from  $\text{Ca}_{\text{XRF}}$  signal, the envelope representing the  
 315 error on the calculation of  $\text{CaCO}_3$ .

316

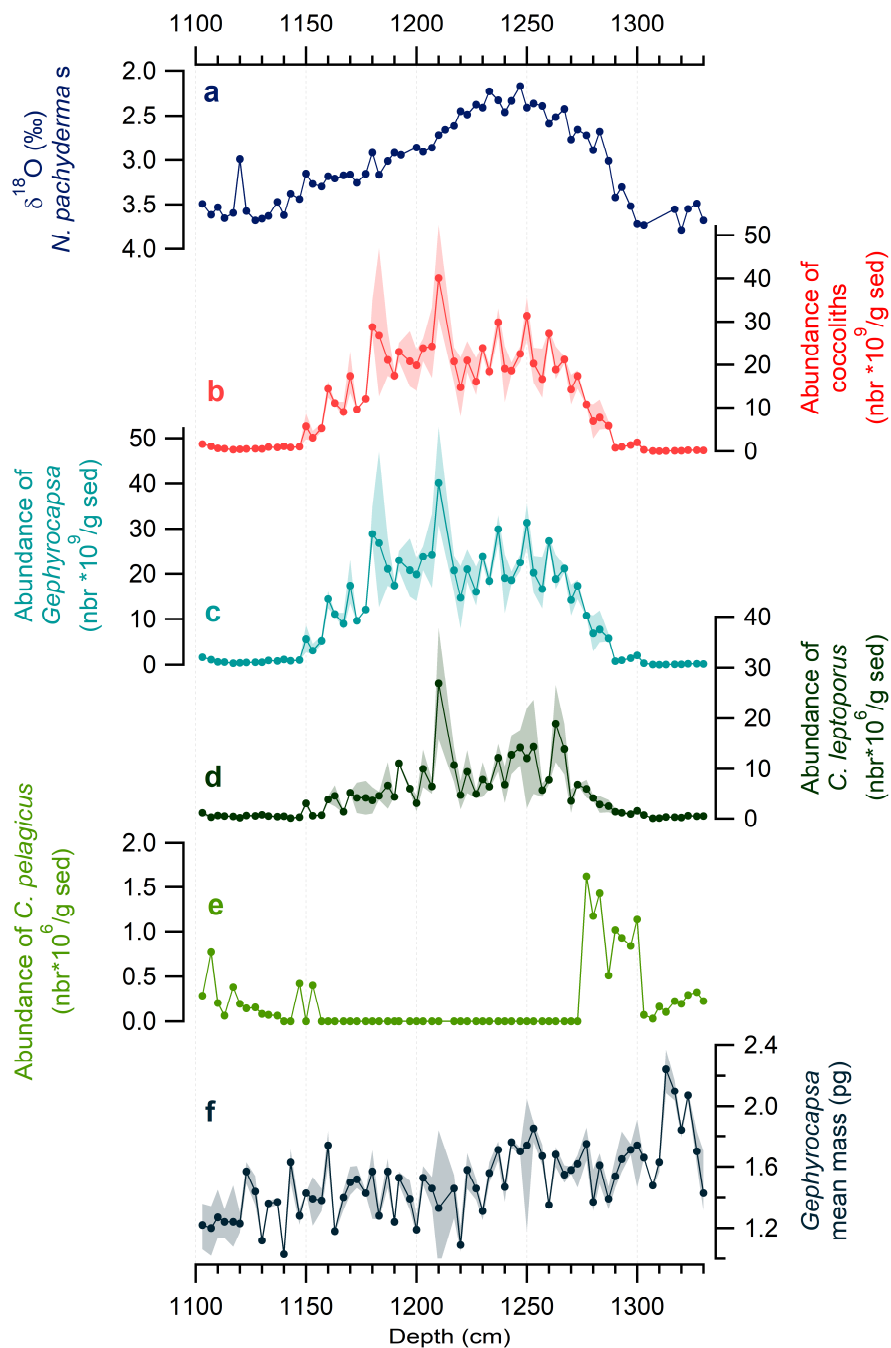
## 317 4.2 Micropaleontological signals

318

319 Coccolith assemblages are predominantly composed of *Gephyrocapsa* sp. i.e., specimens from  
 320 the Noelaerhabdaceae family whose proportions exceed 98.5 %, reaching maximum values (99.9 %)   
 321 between 12.87 and 11.33 m depths (Supplementary Fig. 3). *C. pelagicus* and *C. leptoporus* are  
 322 present as well, but in much lower proportions. They are mainly represented by the subspecies *C.*  
 323 *pelagicus* ssp. *braarudii* (10-14  $\mu\text{m}$ ) and *C. leptoporus* (5-8  $\mu\text{m}$ ) and range between 0 and 0.9 %, and  
 324 0 and 0.6 % of the assemblages, respectively. The number of total coccoliths per gram of sediments

325 (per g. sed.) ranges from  $0.017 \cdot 10^9$  (at 13.10 m) to  $40 \cdot 10^9$  (at 12.10 m) and is mainly represented by  
 326 the number of *Gephyrocapsa* sp. per g. sed. that fluctuates between  $0.016 \cdot 10^9$  and  $40 \cdot 10^9$  (Fig. 3b  
 327 and 3c).

328



329

330 *Figure 3 Coccolith and  $\delta^{18}O_{N. pachys}$  signals between 1103 and 1330 cm of the sediment core MD04-2718. a.  $\delta^{18}O_{N. pachys}$  (‰),*  
 331 *abundances of : b coccoliths (number of coccoliths  $\cdot 10^9$  per g sed.), c *Gephyrocapsa* (number  $\cdot 10^9$  per g sed.), d *C. leptoporus**  
 332 *(number  $\cdot 10^6$  per g sed.). e *C. pelagicus* (number  $\cdot 10^6$  per g sed.), f *Gephyrocapsa* mean mass (pg). Envelopes in b, c, d and f*  
 333 *show the error on SYRACO abundance for each specie and for *Gephyrocapsa* mean mass.*

334

335 After an increasing trend from 12.90 to 12.67 m, values show a plateau around  $22.5 \cdot 10^9$   
336 specimens per g. sed. between 12.67 and 11.80 m and decreases to reach values as low as  $0.35 \cdot 10^9$   
337 specimens per g. sed. at 11.17 m. *C. leptoporus* abundance fluctuates between  $0.06 \cdot 10^6$  and  
338  $21.6 \cdot 10^6$  specimens per g. sed. and its distribution generally follows that of *Gephyrocapsa* sp. (and  
339 thus of total coccoliths), with an increasing trend between 13.07 and 12.67 cm followed by a long-  
340 term decreasing trend up to 11.47 m, interrupted by a transient a peak at 12.10 m (Fig. 3d). *C.*  
341 *pelagicus* abundance varies between 0 and  $1.6 \cdot 10^6$  coccoliths per g. sed (Fig. 3e). It is totally absent  
342 from the sediment between 12.73 and 11.54 m and is thus only present between 13.30 and 12.77 m  
343 and between 11.53 and 11.03 m, where it reaches values as high as 1.6 and  $0.77 \cdot 10^6$  specimens per  
344 g. sed, respectively. *Gephyrocapsa* sp. mean mass varies between 1.03 pg and 2.24 pg (Fig. 3f). After  
345 reaching maximum values (from 2.07 to 2.24 pg) between 13.23 and 13.13 m, data are consistent  
346 with a long-term decrease to reach minimum values around 11.40 m depth.

347 The planktonic foraminifera assemblages are mainly composed of *N. pachyderma* sinistral and  
348 *G. bulloides* with proportions ranging between 0 % and 99.8 % and between 0 and 75.1 %,  
349 respectively. The other species, represented by *G. glutinata*, *G. inflata*, *G. crassaformis*, *N.*  
350 *pachyderma* dextral and *T. quinqueloba* never exceed 6.8 %. (Supplementary Fig. 4). The total  
351 number of planktonic foraminifera fluctuates between 0 and  $18.5 \cdot 10^3$  specimens per g. sed (Fig. 4b).  
352 No specimen is found between 13.13 and 13.07 m, and the total abundance of planktonic  
353 foraminifera never exceeds  $5.7 \cdot 10^3$  between 13.30 and 12.9 m and between 11.5 and 11.3 m.  
354 Between 12.9 and 11.5 depth the number of *N. pachyderma* s. varies between 0.3 and  $12.4 \cdot 10^3$   
355 specimens per g. sed (Fig. 4c), and significantly influences the total number of planktonic  
356 foraminifera which generally shows the same trends. Its abundance shows an overall plateau around  
357  $3 \cdot 10^3$  specimens per g. sed, and 2 maxima ( $12.4 \cdot 10^3$  and  $10.9 \cdot 10^3$  specimens per g. sed) at 12.70  
358 and 11.73 m, respectively. The number of *G. bulloides* fluctuates between 2.10 and  $4.7 \cdot 10^3$   
359 specimens per g. sed (Fig. 4d). After reaching maximum value at 12.70 m, it depicts a long-term  
360 decreasing trend to the minimum abundance of 2.1 specimens per g. sed. at 11.3 m. The number of  
361 *N. pachyderma* d. varies between 0.14 and 548.6 specimens per g. sed. and follows the general same  
362 trends as those of total planktonic foraminifera and *N. pachyderma* s. (Fig. 4e). *G. glutinata*  
363 abundance ranges between 0.2 and 551 specimens per g. sed (Fig. 4f). It depicts rather scattered  
364 values with two peaks (512 and 551 specimens per g. sed.) at 12.70 and 12.33 m. *G. crassaformis*  
365 abundance fluctuates between 0 and 259.5 specimens per g. sed, with values averaging 150  
366 specimens per g. sed between 12.70 and 12.13 m (Fig. 4g). *T. quinqueloba* abundance fluctuates  
367 between 0 and 219.4 specimens per g. sed, and no specific trend may be highlighted (Fig. 4h). Finally,

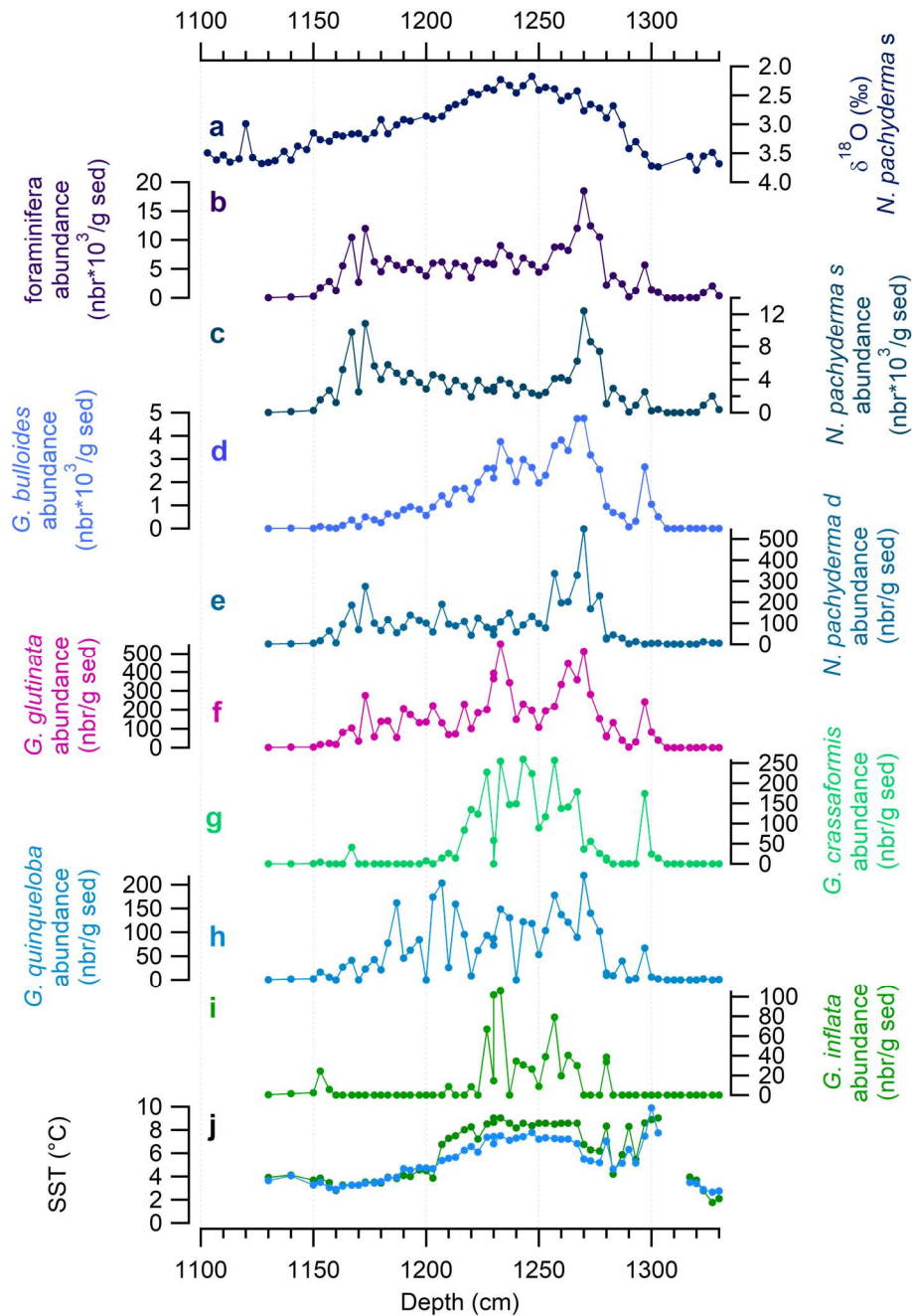
368 *G. inflata* abundance ranges between 0 and 106 specimens per g. sed (Fig. 4i). Similarly, to *G.*  
369 *crassaformis*, no or very few specimens are found between 12.23 and 11.60 m.

370

371 The reconstructed summer SSTs fluctuate between 1.7 and 9.9 °C (Fig. 4j). The maximum  
372 temperature is reached after a drastic warming amounting ~8 °C between 13.27 and 13.00 m. It is  
373 followed by temperatures fluctuating between 4.6 and 8.3°C from 12.93 to 12.70 m, and a plateau  
374 around 8°C between 12.67 and 12.27 m. Then, summer SSTs gradually decrease and reach down to 3  
375 °C between 12.03 and 11.30 m.

376





377

378 Figure 4 Planktonic foraminifera and  $\delta^{18}O_{N. pachy s}$  signals between 1103 and 1330 cm of the sediment core MD04-2718. **a**  
 379 Signal of  $\delta^{18}O_{N. pachy s}$  (‰), **b** Number of total planktonic foraminifera (per g sed.), Number (per g. of sed.) of: **c** *N.*  
 380 *pachyderma s.*, **d** *G. bulloides*, **e** *N. pachyderma d.*, **f** *G. glutinata*, **g** *G. crassaformis*, **h** *T. quinqueloba*, **i** *G. inflata*, **j**  
 381 Reconstructed summer SST ( $^{\circ}C$ ) using *N. pachyderma s.* percentages (blue curve) and based on Modern Analog Technique  
 382 method (green curve).

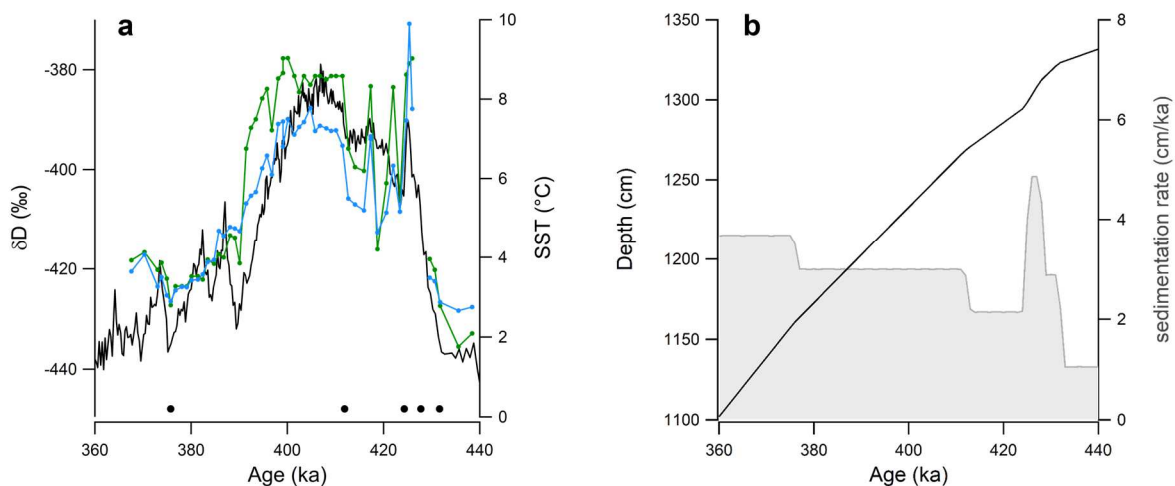
383

## 384 5. Age model

385

386 The age model of core MD04-2718 is based on the planktonic foraminifera  $\delta^{18}O$  record and  
 387 sedimentary  $CaCO_3$  signal obtained on the uppermost 30 m of the core (Supplementary Fig. 5). These  
 388 records helped identifying glacial-interglacial cycles and determining that the uppermost 20 m of the

389 core covered the last 800 ka. It also allowed us to define the depth interval 13.00 to 11.00 m as  
 390 representing the MIS 12 – MIS 10 time-interval and confirm our assumption that the nannofossil  
 391 ooze between 11.5 and 13.1m was deposited during MIS 11 (see section 3.1). Numerous studies  
 392 showed that the correlation between reconstructed SSTs in the Southern Ocean and air temperature  
 393 reconstructed from Antarctic ice cores was helpful to better constrain sediment core age model (e.g.,  
 394 Pichon et al., 1992; Sowers et al. 1993; Waelbroeck et al., 1995; Thöle et al., 2019; Ai et al., 2020).  
 395 Here, we used the record of  $\delta D$  of the EPICA Dome C ice core as a proxy for Antarctic surface  
 396 temperature (Jouzel et al., 2007). As such, for the interval MIS 12 – 10, the age model of core MD04-  
 397 2718 was refined based on the graphical alignment of our reconstructed summer SST with ~~record~~  
 398 ~~of~~ the  $\delta D$  record on the AICC2012 timescale (Bazin et al., 2013), assuming an in-phase relationship  
 399 (Fig. 5a). Using Analyseries software (Paillard et al., 1996), we set 5 tie points which helped us  
 400 identifying the 11.03- 13.3 m core section as representing the time interval between 438.4 and 360.2  
 401 ka i.e., between the end of MIS 12 and the beginning of MIS 10. Over this period, the sedimentation  
 402 rate averages 3.1 cm.ka<sup>-1</sup> with lowest values (1 cm.ka<sup>-1</sup>) being recorded during MIS 12, and highest  
 403 ones (4.9 cm.ka<sup>-1</sup>) being observed at the beginning of MIS 11 (Fig. 5b). Given these sedimentation  
 404 rates, the temporal resolution of our record ranges from ~ 0.6 to 3.8 ka for micropaleontological and  
 405 stable isotope data and of 0.2 to 2 ka for the CaCO<sub>3</sub> signal.  
 406



407  
 408 *Figure 5 a* Age model of MD04-2718 based on the correlation between MD04-2718 summer SST reconstructions and EDC  $\delta D$   
 409 *signal on AICC2012 timescale. Blue curve is the reconstructed summer SST using *N. pachyderma s.* percentages. Green curve*  
 410 *is the reconstructed summer SST based on Modern Analog Technique method. Black curve is the  $\delta D$  measured in EDC ice*  
 411 *core (Jouzel et al., 2007) on the AICC2012 timescale (Bazin et al., 2013). Black dots are the tie points between SST and  $\delta D$*   
 412 *Estimated sedimentation rate of MD04-2718 core. Black curve represents the depth versus the calibrated age (ka BP on*  
 413 *AICC2012) and the grey shaded area represent the variation of sedimentation rate with time.*

414

## 415 6. Discussion

416

### 417 6.1 CaCO<sub>3</sub> as a record of CCP strength

418

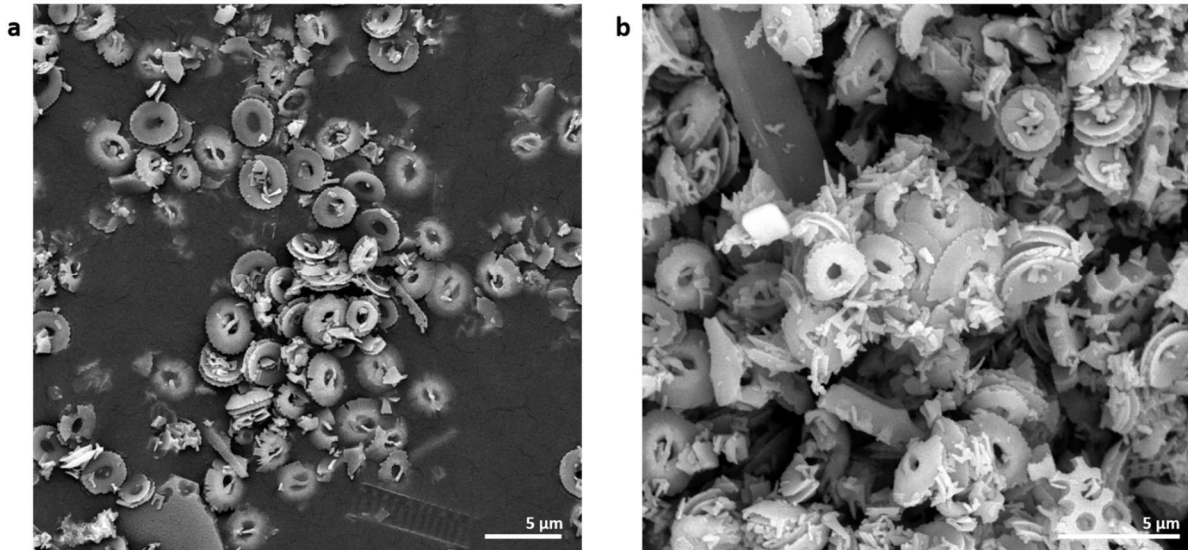
419 In pelagic environments of the SO, sedimentary carbonate concentration and/or coccolith  
420 abundances have been assumed to reflect biogenic carbonate export production - and by inference  
421 the strength of the CCP - based on the paradigm that in modern environments, and more particularly  
422 in the Great Calcite Belt, coccolithophores represent an important – if not dominant – part of  
423 exported mineralized organisms (Salter et al., 2014; Rembauville et al., 2016). However, the bulk  
424 sedimentary carbonate signal may include detrital or diagenetic phases and within the biogenic  
425 fraction, contributions of planktonic foraminifera and pteropods, the latter often considered  
426 negligible (Salter et al., 2014; Rembauville et al., 2016). Therefore, special attention must be devoted  
427 to assessing the significance of bulk sedimentary carbonate signals before making interpretations in  
428 terms of planktonic carbonate productivity and, by inference, of the CCP strength. It requires to  
429 carefully select the sedimentary cores that robustly reflect biogenic carbonate accumulation and to  
430 consider the potentially obfuscating impact of diagenesis.

431 Core MD07-2418 was retrieved at 1,428 m water depth i.e., well above the modern lysocline  
432 located at ~4,300 m (Kolla et al., 1976; Howard and Prell, 1994). The area only receives carbonate-  
433 free terrigenous material sourced from volcanic rocks of the Crozet and Kerguelen plateau (Nougier,  
434 1970). Thus, the sedimented carbonate material directly relates to local biogenic carbonate  
435 production, for which the diagenetic overprint and potential detrital contribution are minimal. These  
436 assumptions are corroborated by Scanning Electron Microscopy (SEM) observations highlighting that  
437 the sedimentary carbonate fraction is predominantly constituted by very well-preserved foraminifera  
438 and coccoliths, sometimes represented as intact coccospheres (Fig. 6). In addition, we measured the  
439 relative proportion of fragmented planktonic foraminifera shells versus intact foraminifers, a well-  
440 characterised preservation index (Vazquez-Riveiros et al., 2010; Flores et al., 2003) (Supplementary  
441 Fig. 6). This foraminifera fragmentation index varies between 0 and 15.2 % with values higher than 3  
442 % observed only between 426 and 417 ka. These results clearly indicate that planktonic foraminiferal  
443 preservation is generally good (Le and Shackleton, 1992).

444 Dissolution processes could result in differential sedimentary preservation in the coccolith  
445 assemblages as well, preferentially preserving resistant specimens. However, the assemblage is  
446 mainly constituted by well-preserved *Gephyrocapsa* sp., as seen on SEM observations. Furthermore,  
447 within this genus, small *Gephyrocapsa* sp. (<3 µm) follow the exact same trends as the larger ones,

448 which suggests that differential dissolution does not significantly affect coccolith assemblages in  
449 sediments from core MD04-2718 (Supplementary Fig. 6).

450



451

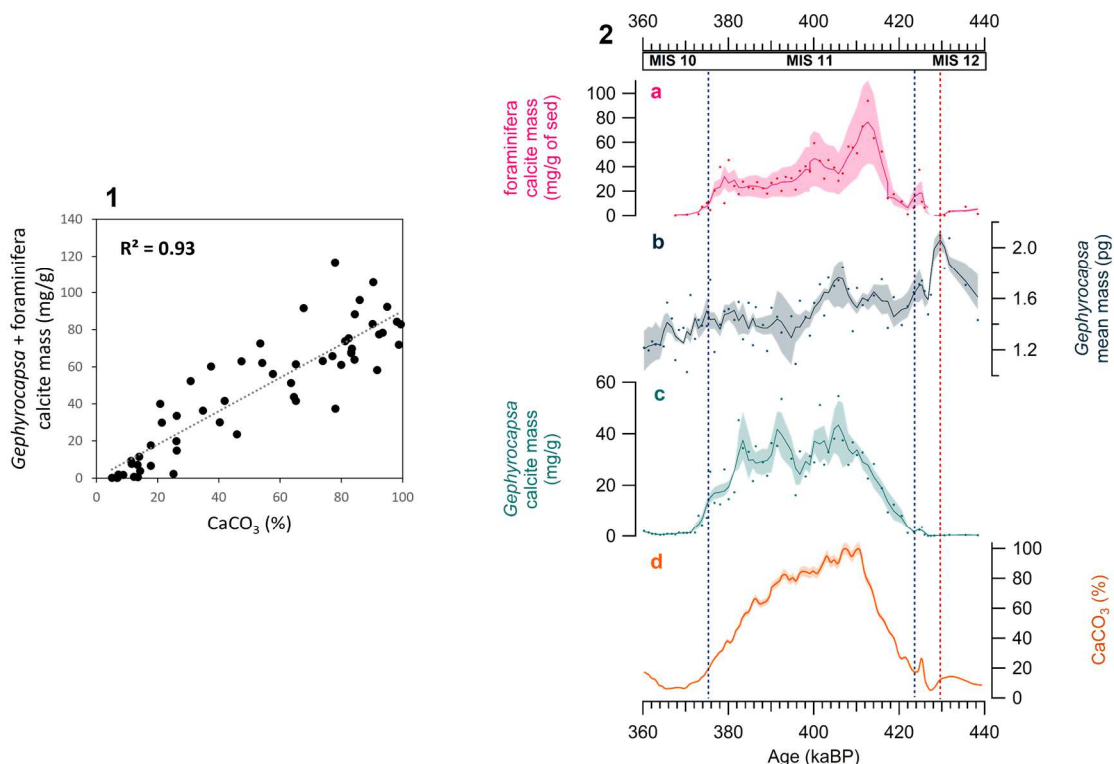
452 *Figure 6 Visual estimation of coccolith preservation of sediment core MD04-2718 based on Tescan VEGA II SEM observation*  
453 *(Museum National d'Histoire Naturelle, MNHN, Paris) showing good preservation a at 389.1 ka and b at 405.8 ka.*

454 At site MD04-2718, *Gephyrocapsa* calcite mass varies between 0.03 mg/g and 54.62 mg/g. The  
455 lowest values are recorded during MIS 12 (0.03 mg/g) and MIS 10 (0.43 mg/g) and the highest ones  
456 are documented during MIS 11. In detail, *Gephyrocapsa* calcite mass increases from 0.03 to 36.68  
457 mg/g at the beginning of MIS 11, averages 30 mg/g during MIS 11, and decreases gradually from 382  
458 ka to 372 ka to reach low values of 0.43 mg/g during glacial MIS 10. The *Gephyrocapsa* calcite mass  
459 record closely resembles the one of *Gephyrocapsa* sp. abundance, whereas no relationship may be  
460 observed with *Gephyrocapsa* mean mass (Fig. 7 and Supplementary Fig. 7). Moreover, while the  
461 variability in *Gephyrocapsa* sp. mean mass along the records are of a factor of two, the observed  
462 variations in abundance are more than a factor of two thousand. We therefore posit that changes in  
463 *Gephyrocapsa* calcite mass are mainly the consequence of changes in *Gephyrocapsa* abundance  
464 rather than in shell mean mass linked with shift in calcification or morphotypes.—The planktonic  
465 foraminifera (>150 µm) calcite mass varies between 0 and 93.9 mg/g. Briefly, it is characterised by  
466 null or low values during MIS 12 (< 7.2 mg/g) and MIS 10 (< 1.08 mg/g) and generally higher values  
467 during MIS 11 and the beginning of MIS 10 (> 20 mg/g). Maximum values are observed around 412.7  
468 ka i.e., at the beginning of MIS 11. According to the significant linear correlation that exists between  
469 the *Gephyrocapsa* sp. + planktonic foraminifera calcite mass and sedimentary CaCO<sub>3</sub> content ( $r^2 =$   
470 0.93), it appears that both plankton groups constitute the major component of the biogenic  
471 carbonate signal. *Gephyrocapsa* sp. seems to have a relatively larger contribution on the sedimentary

472 CaCO<sub>3</sub> content as shown by its greater correlation to CaCO<sub>3</sub> ( $r^2 = 0.89$ ) than planktonic foraminifera  
 473 calcite mass itself ( $r^2 = 0.83$ ) (Supplementary Fig. 2). Such results are consistent with the major  
 474 control of coccoliths and more particularly morphotypes of the Noelaerhabdaceae family exert on  
 475 PIC export today (Rembauville et al., 2016; Patil et al., 2020).

476 In the following, changes in the downcore CaCO<sub>3</sub> signal at site MD04-2718 will be interpreted  
 477 in terms of changes in planktonic carbonate export, and by inference changes in CCP strength.

478



479

480 *Figure 7 Origin of the carbonate fraction. 1 Correlation between CaCO<sub>3</sub> and Gephyrocapsa + planktonic foraminifera calcite*  
 481 *mass, 2 a planktonic foraminifera calcite mass (mg/g), b Gephyrocapsa mean mass (pg), c Gephyrocapsa calcite mass*  
 482 *(mg/g), d CaCO<sub>3</sub> (%). Smoothed curves use a three-point average using IGOR software. The shaded areas represent the error*  
 483 *on the curve. The dotted blue lines represent the beginning and end of MIS 11 from right to left respectively, and the dotted*  
 484 *red line represent the middle of the rising slope of CO<sub>2</sub> from glacial to interglacial.*

485

## 486 6.2 Increased CCP during MIS 11 as a response to the southward migration of 487 westerly winds and ACC front positions

488

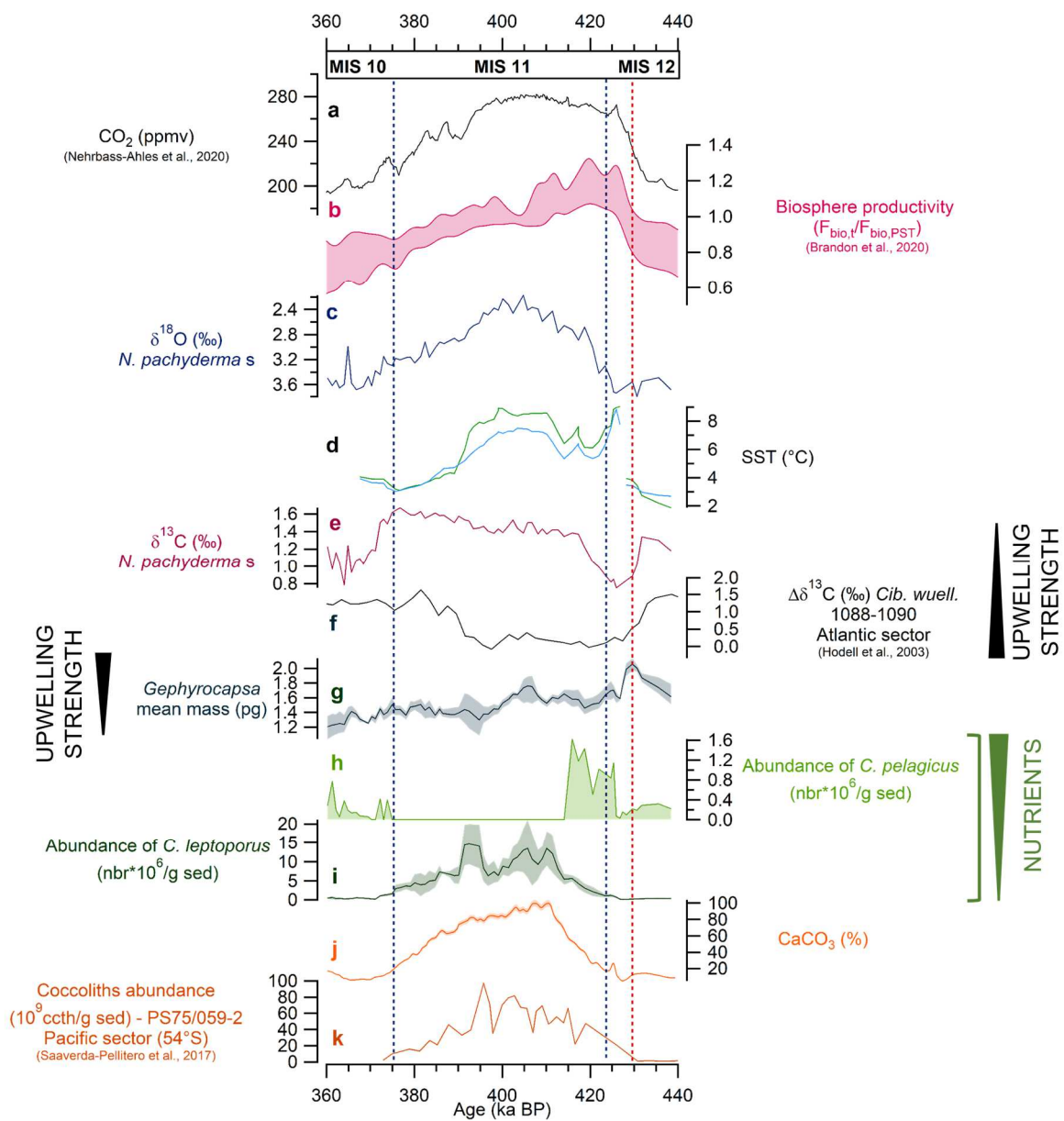
489 As represented by the variation in sedimentary CaCO<sub>3</sub> content, it appears that the CCP was  
 490 strengthened during MIS 11 compared to MIS 12 and MIS 10 (Fig. 8). By analogy to the modern SO  
 491 (Baumann et al., 1999; Zondervan et al., 2007) and the Pacific sector of the SAZ during MIS 12-MIS 10

492 (Schaefer et al., 2005; Ullermann, 2015; Saavedra-Pellitero et al., 2017a; 2017b), it is probable that  
493 such recorded pattern is related to substantial changes in SST and nutrient conditions. In addition to  
494 reconstructed SSTs, the abundances of *C. pelagicus* and *C. leptoporus* can be used to indirectly assess  
495 past changes in sea surface nutrient and temperature conditions. Indeed, while *C. pelagicus* seems to  
496 proliferate under colder conditions than *C. leptoporus* (Cobianchi et al., 2012), both these species  
497 appear to have affinities with relatively nutrient-rich waters (Andruleit, 1997; Balestra et al., 2004;  
498 Amore et al., 2012). This behaviour has been well documented on glacial/interglacial timescales in  
499 the SO (Flores et al., 1999; Findlay and Flores, 2000; Marino et al., 2009; Saavedra-Pellitero et al.,  
500 2017a; 2017b) to characterize frontal zone dynamics (Flores et al., 1999; Findlay and Giraudeau;  
501 2000; Flores et al., 2003; Baumann et al., 2004; Malinverno et al., 2015; Marino et al., 2008;  
502 Saavedra-Pellitero et al., 2017a; 2017b). At core site MD04-2718, reconstructed SSTs range between  
503  $\sim 2\text{-}4^{\circ}\text{C}$  during glacial MIS 12 and MIS 10, and  $\sim 8^{\circ}\text{C}$  during MIS 11 (Fig. 8). In parallel, relatively fertile  
504 conditions are inferred during TV and MIS 11 by the coccolith species *C. pelagicus* and *C. leptoporus*  
505 that peak between 426 and 415 ka and between 412 and 392 ka, respectively. The CCP was thus  
506 stronger during the relatively warm and nutrient-rich MIS 11 interglacial compared to the relatively  
507 cold and nutrient-depleted MIS 12 and MIS 10. The CCP trend shows however some temporal lag  
508 with the SST and nutrient records, highlighting other mechanisms in potentially regulating the  
509 variability of CCP strength. This is particularly obvious during TV (430-424 ka), when the CCP remains  
510 relatively weak despite drastic increases in SSTs and nutrient conditions. At that time, decreasing  
511  $\delta^{13}\text{C}_{\text{N.pachyderma}}$  values and increasing mean mass of *Gephyrocapsa* are recorded (Fig. 8). The rapid  $\delta^{13}\text{C}$   
512 decrease during the deglaciation is also documented in the South Pacific (Saavedra-Pellitero et al.,  
513 2017b) and South Atlantic (Hodell et al., 2003) (Fig. 8). This isotopic signal is consistent with the  
514 resumption of SO upwelling, which supplies abundant nutrients, previously sequestered in the ocean  
515 interior owing to greater water-column stratification, to the productive ocean surface (François et al.,  
516 1997; Spero et al., 2002; Adkins, 2013; Siani et al., 2013). Coeval increase in *Gephyrocapsa* mean  
517 mass has also been associated with increase in  $\text{CO}_2$ -rich upwelled waters in the SO (Duchamp-  
518 Alphonse et al., 2018). The reason behind these staggered and seemingly disconnected patterns  
519 could relate to the ecological competition between diatoms and coccolithophores in the SO. During  
520 glacial periods, favoured by higher iron (Fe) availability, diatom productivity outcompetes  
521 coccolithophore growth (Martin, 1990; Kohfeld et al., 2005; Martinez-Garcia et al., 2009; Thöle et al.,  
522 2019), and call for a weaker CCP. While the availability of Fe dwindled over deglacial periods,  
523 reinvigorated upwelling of nutrient-rich waters would have promoted diatom growth during  
524 Termination V. In such a scenario, we hypothesize that it is only during peak MIS 11 that silicic acid  
525 concentrations became too low, thereby limiting diatom growth (Sarmiento et al., 2004), and thus  
526 had the potential to favour coccolith production, enhancing the CCP. Overall, since the upwelled



527 waters that (re)surface in the AAZ act under the forcing of the prevailing westerlies (Anderson et al.,  
 528 2009; Lauderdale et al., 2013; Menviel et al., 2018), it is probable that winds migrated poleward  
 529 between MIS 12 and MIS 11, concomitantly forcing the poleward migration of the entire frontal  
 530 system (Flores and Sierro, 2007; Kemp et al., 2010; Saavedra-Pellitero et al., 2017a; 2017b).  
 531 According to the reconstructed SST ( $< 8^{\circ}\text{C}$  in mean) as well as the relatively low percentages of *C.*  
 532 *leptoporus*, while close to the SAF, it is likely that site MD04-2718 remained in the PFZ. That is only  
 533 around 390 ka, when SSTs decreased, that westerlies (along with the frontal system) might have  
 534 shifted back to the North, leading to a decrease in the CCP.

535



536

537

538 *Figure 8 Records of CCP and upwelling strengths in the SO and relation with variation of atmospheric CO<sub>2</sub> concentration. a*  
539 *CO<sub>2</sub> (ppm) (Nehrbass-Ahles et al., 2020), b Biosphere productivity (F<sub>bio,t</sub>/F<sub>bio,PST</sub>) (Brandon et al., 2020), c δ<sup>18</sup>O of N.*  
540 *pachyderma s. from MD04-2718, d summer SST reconstructed using N. pachyderma s percentages (blue curve) and using*  
541 *Modern Analog Technique method (green curve), e δ<sup>13</sup>C of N. pachyderma s. from MD04-2718, f Δδ<sup>13</sup>C of benthic*  
542 *Cibicidoides wuellerstorfi from ODP Site 1088- ODP Site 1090 (Hodell et al., 2003), g Gephyrocapsa mean mass (pg), h C.*  
543 *pelagicus abundance (\*10<sup>6</sup>/g sed), i C. leptoporus abundance (\*10<sup>6</sup>/g sed), j CaCO<sub>3</sub> (%) derived from Ca<sub>XRF</sub>, k PS75/059-2*  
544 *Coccoliths abundance (10<sup>9</sup>coccoliths/g of sediment, Saavedra-Pellitero et al., 2017b), showed on AICC2012 chronology.*

545

### 546 6.3 Increased CCP during TV and MIS 11 and its impact on atmospheric CO<sub>2</sub>

547

548 The onset of deep-water ventilation following MIS 12 is recorded at ~432 ka, while the CCP  
549 starts to increase a few thousand years later, around 426 ka (Fig. 8). This distinct timing suggests that  
550 the rapid increase in atmospheric pCO<sub>2</sub> during Termination V was triggered, in part, by the  
551 reinvigoration of the SO upwelling early in the deglacial progression. Concomitant with the  
552 reinvigoration of SO upwelling, the increase in global biosphere productivity (Brandon et al., 2020)  
553 that reflects in part, an increase in vegetation cover and terrestrial carbon storage (Tzedakis et al.,  
554 2006; de Vernal and Hillaire-Marcel, 2008; Prokopenko et al., 2009; Melles et al., 2012; Kousis et al.,  
555 2018) has probably acted as a sink of atmospheric CO<sub>2</sub>, thereby transiently buffering the impact of  
556 oceanic CO<sub>2</sub> degassing (Fig. 8). pCO<sub>2</sub> reached maximum concentration (up to ~ 282 ppm) during the  
557 second part of MIS 11, as the CCP functioned at maximum strength. Enhanced CCP in the SO  
558 contributed to reduce the efficiency of the BCP. Overall, enhanced *Gephyrocapsa* sp. export and  
559 burial both in the SO (Saavedra-Pellitero et al., 2017a; 2017b) and at low latitudes (Barker et al.,  
560 2006; Rickaby et al., 2007; Alvarez et al., 2010; Amore et al., 2012; Saavedra-Pellitero et al., 2019),  
561 would have led to a global increase in carbonate dissolution at the seafloor (Barker et al., 2006; Qin  
562 et al., 2018) and to a significant change in the global rain ratio, eventually impacting atmospheric  
563 pCO<sub>2</sub> (Sigman et al., 1998; Sigman and Boyle, 2000; Antia et al., 2001; Barker et al., 2006). This  
564 hypothesis warrants further investigations to fully quantify changes in the rain ratio and their impact  
565 on atmospheric pCO<sub>2</sub>. From 398 ka to 381 ka, the gradual, ~35 ppm decrease in atmospheric pCO<sub>2</sub>  
566 coincided well with the rapid increase in benthic foraminifera Δδ<sup>13</sup>C in the South Atlantic (Hodell et  
567 al., 2003), consistent with enhanced stratification of the SO and increased carbon storage in the  
568 ocean interior. The second part of atmospheric CO<sub>2</sub> drawdown during MIS 10 would be supported in  
569 part by the reduction in the strength of the CCP combined with a globally more efficient marine BCP.

570 At a global scale, the most striking pattern of MIS 11 relates to the low eccentricity context  
571 that is documented around 400 ka (Laskar et al., 2004). Low eccentricity, through the lengthening of



572 the growing season, might have triggered an acme of primary producers such as *Gephyrocapsa* sp., at  
573 both high and low latitudes (Rickaby et al., 2007). One cannot exclude either that our record reflects  
574 the long-term cyclic Noelaerhabdaceae evolutionary pattern recently highlighted for the tropical area  
575 over the Pleistocene (Beaufort et al., 2022), with *Gephyrocapsa* blooms during MIS 11 being the  
576 consequence of reduced seasonality and phytoplankton niches. In all cases, we propose that, under  
577 this particularly low eccentricity context, enhanced terrestrial productivity and carbon storage on  
578 land, together with enhanced biological carbonate productivity and CCP strength in the SO may have  
579 contributed to shape the 30 ka-long CO<sub>2</sub> plateau characteristic of MIS 11.

580

## 581 7. Conclusion

582

583 Our study documents temporal variations in the strength of the CCP based on Southern Ocean  
584 sediments and their impact on atmospheric CO<sub>2</sub> concentrations over the period ranging between MIS  
585 12 and MIS 10. First, using micropaleontological and geochemical data, we demonstrate that the  
586 sedimentary CaCO<sub>3</sub> signal robustly reflects the production and export of *Gephyrocapsa* sp. and (to a  
587 lesser degree) planktonic foraminifera and can therefore be used as a high-resolution proxy of the  
588 CCP. Second, using planktic foraminifera, *C. leptoporus* and *C. pelagicus* abundances combined with  
589  $\delta^{13}\text{C}$  of *N. pachyderma* s. from the same site, we reconstruct changes in sea surface temperature and  
590 nutrient conditions in conjunction with the regional upwelling dynamics, which are closely linked to  
591 the latitudinal migration of the westerly winds. Third, the combination of the results obtained in this  
592 study together with previous micropaleontological and geochemical data from literature allow us to  
593 highlight the sequence of events leading to the deglaciation and to understand the specific  
594 atmospheric pCO<sub>2</sub> pattern that characterises this key interval of the past. We show that the abrupt  
595 atmospheric CO<sub>2</sub> rise observed during TV is led by the reinvigoration of SO upwelling probably due to  
596 a poleward migration of the westerlies and ACC fronts, while the sustained CO<sub>2</sub> plateau observed  
597 during MIS 11 is the result of enhanced global primary productivity and increased CCP due to the  
598 exceptionally low eccentricity context.

599

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612

### 613 **Data availability**

614 Data will be made available in the Pangaea database prior to publication.

615

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