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

Brandon Margaux ^{1,2,*}, Duchamp-Alphonse Stéphanie ¹, Michel Elisabeth ², Landais Amaëlle ², Isguder Gulay ², Richard Patricia ², Pige Nicolas ³, Bassinot Franck ², Jaccard Samuel L. ⁴, Bartolini Annachiara ⁵

¹ Université Paris-Saclay, CNRS, GEOPS, 91405, Orsay, France

² Université Paris-Saclay, CNRS, CEA, UVSQ, Laboratoire des Sciences du climat et de l'environnement, 91191, Gif-sur-Yvette, France

³ Univ Lyon, UCBL, ENSL, UJM, CNRS, LGL-TPE, F-69622, Villeurbanne, France

⁴ Institute of Earth Sciences, University of Lausanne, 1015, Lausanne, Switzerland

⁵ Muséum National d'Histoire Naturelle, Département Origines & Évolution, CR2P MNHN, CNRS, Sorbonne Université, 8 rue Buffon CP38, 75005, Paris, France

* Corresponding author : Margaux Brandon, email address : margaux.brandon@locean.ipsl.fr

Abstract :

While numerous studies have highlighted the central role of Southern Ocean (SO) dynamics in modulating rapid increases in atmospheric CO₂ 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 CO₂ concentration in oceanic surface waters and thus, in the atmosphere. Here, we present micropaleontological (coccolith, planktonic foraminifera) and geochemical (CaCO₃, CaXRF, δ¹³CN, 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 CO₂. We demonstrate that the sharp increase in atmospheric pCO₂ during Termination V was likely associated with enhanced deep-water ventilation in the SO, that promoted the release of previously sequestered CO₂ 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 pCO₂ plateau characteristic of MIS 11.

Highlights

► Multiproxy approach to reconstruct Carbonate Counter Pump strength and upwelling dynamics. ► Increase in pCO₂ 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 pCO₂ plateau.

Keywords : Southern Ocean, Marine Isotope Stage 11, Coccolith, Planktonic foraminifera, Geochemistry, Carbonate Counter Pump, Upwelling, Atmospheric CO₂ 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 pCO₂ changes (Lüthi et al., 2008; Bereiter et al.,
46 2015). Causes underlying these changes, especially in conjunction with deglaciations (glacial
47 Terminations) that registered pCO₂ 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 pCO₂
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 CO₂ 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 pCO₂. 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 CO₂ (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 (CaCO₃), which raises surface water CO₂

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 CaCO₃/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 pCO₂ (~ 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 pCO₂ 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 CaCO₃ 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 $\delta^{13}\text{C}_{\text{N. pachyderma}}$ 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 CO₂ 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 CO₂- 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 CO₂ 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 CaCO₃
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

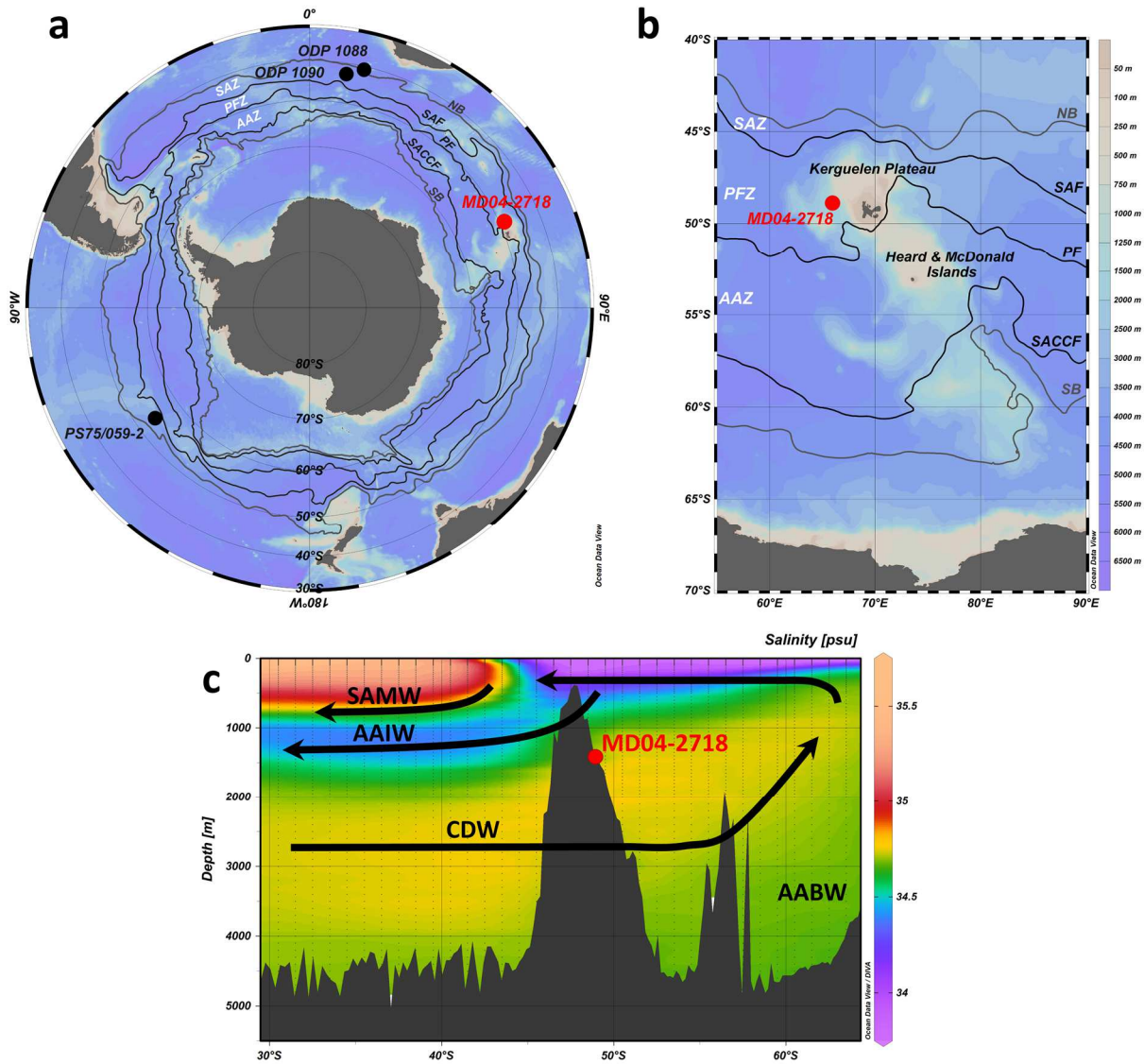
153

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 CO₂- 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 μm 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 (1σ) for carbonate
182 standards of ± 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 μA . In parallel, sedimentary CaCO_3 concentrations were determined on
187 26 samples at GEOPS Laboratory with a ± 2 % precision, using the vacuum-gasometric technique
188 described in Duchamp-Alphonse et al., (2018) where the amount of CO_2 degassed during the reaction
189 of $100 \text{ mg} \pm 2 \text{ mg}$ of dried sediments with 6N HCl is converted into CaCO_3 percentages. Comparison
190 between Ca counts and discrete CaCO_3 contents show a highly significant statistical correlation
191 ($r^2=0.97$), and Ca_{XRF} signal is therefore directly converted into 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 mm²), N_o ,
224 the number of view fields (165), S_o , the view field area (0.01 mm²) 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 μg), $size_{ji}$ 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 μm , 200-250 μm , 250-315 μm , 315-355
 259 μm , 355-400 μm , 400-450 μm , 450-500 μm and >500 μm), 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 μg), 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 ± 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 μm), 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 μm to 250-315 μm for *G. bulloides*, *G. crassaformis*
 272 and *G. truncatulinoides* and from 150-200 μm to 315-355 μm 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 μm fraction contains almost exclusively ice rafted detritus and radiolarians.

296

297 4. Results

298

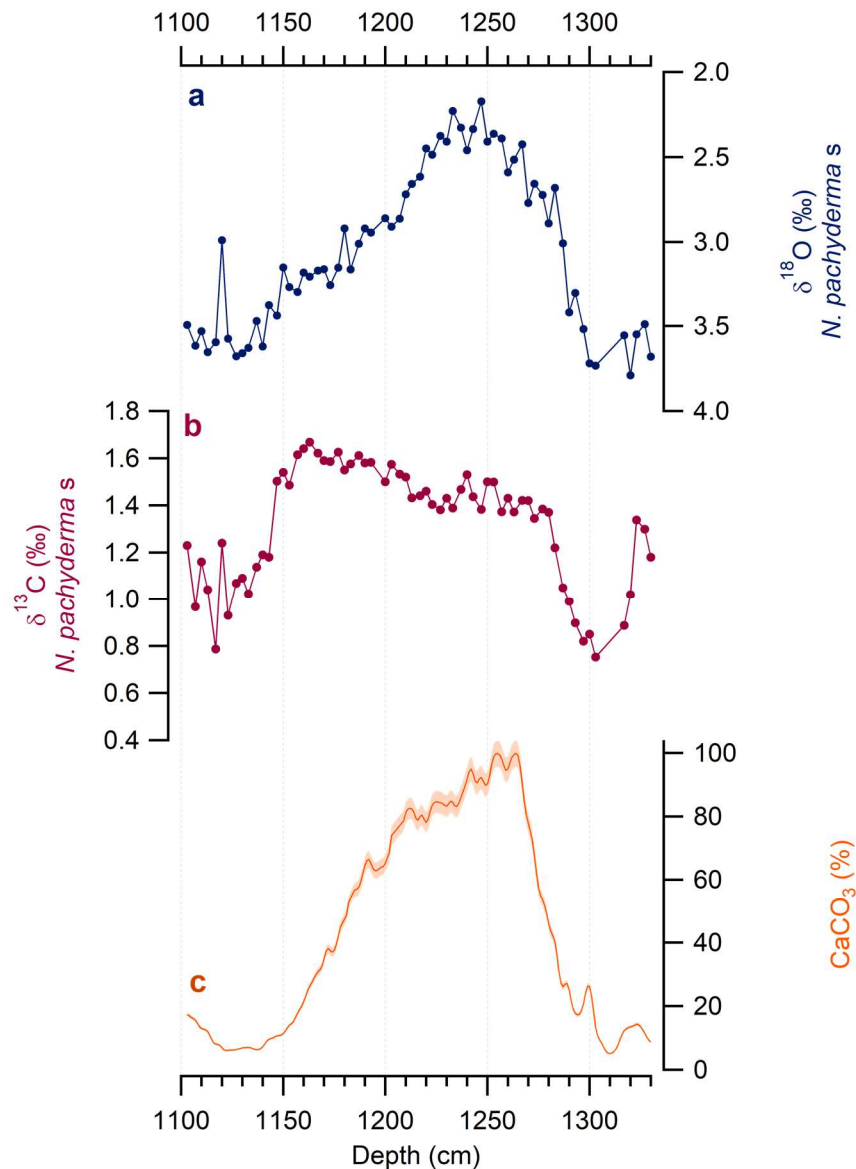
299 4.1 Geochemical signals

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 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 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** CaCO_3 (%) derived from Ca_{XRF} signal, the envelope representing the
 315 error on the calculation of CaCO_3 .

316

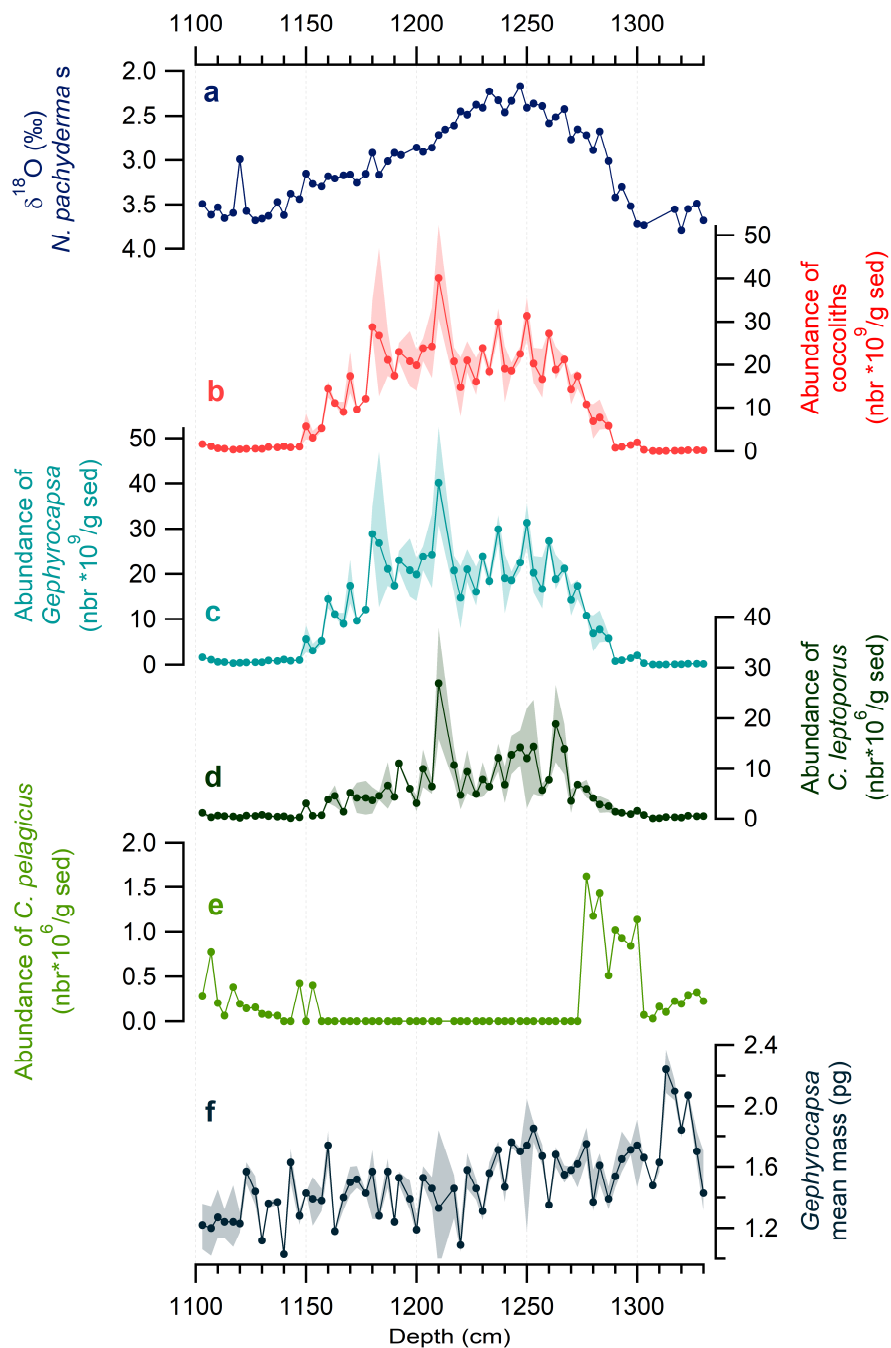
317 4.2 Micropaleontological signals

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 μm) and *C. leptoporus* (5-8 μm) 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.

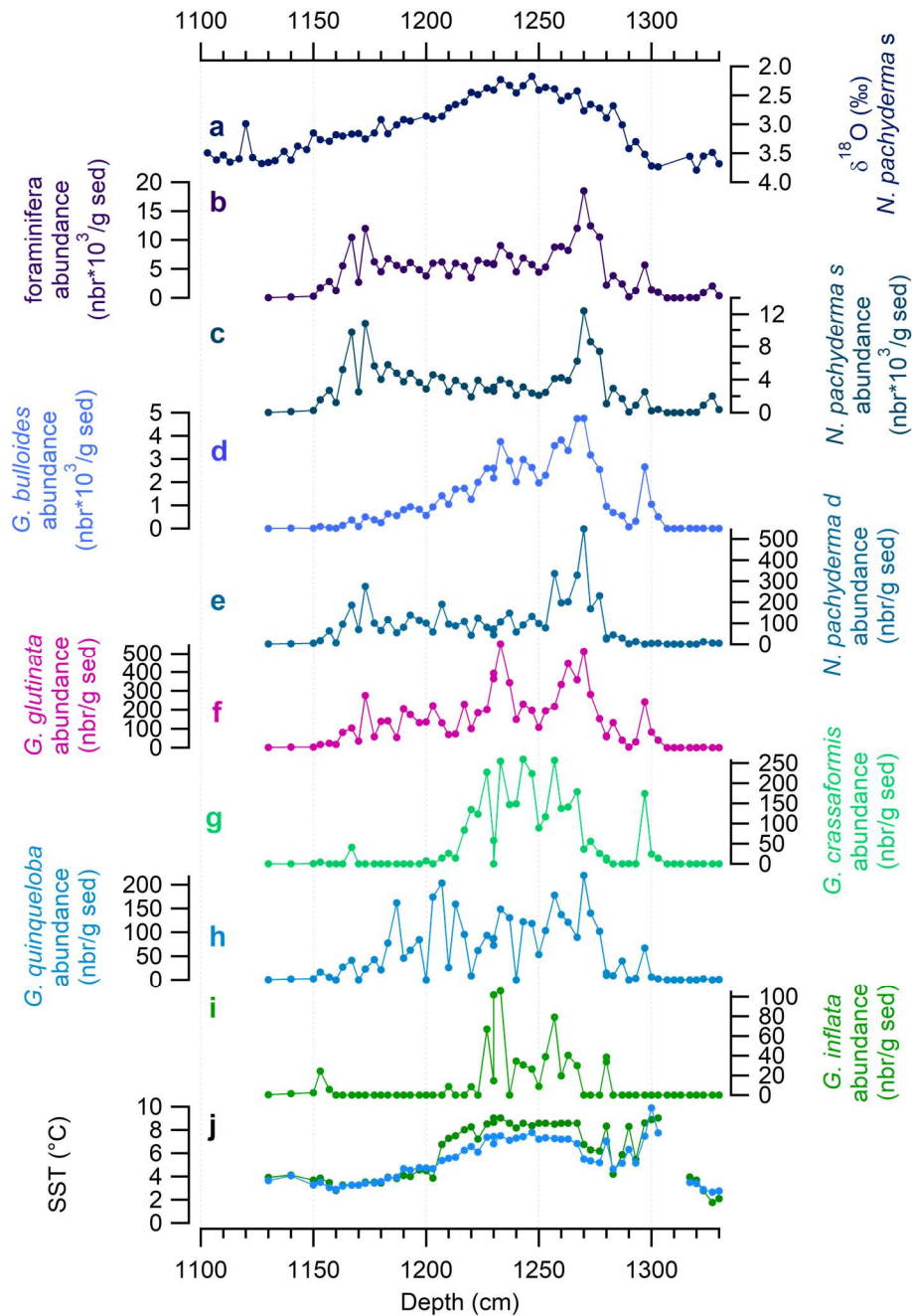
347 The planktonic foraminifera assemblages are mainly composed of *N. pachyderma* sinistral and
348 *G. bulloides* with proportions ranging between 0 % and 99.8 % and between 0 and 75.1 %,
349 respectively. The other species, represented by *G. glutinata*, *G. inflata*, *G. crassaformis*, *N.*
350 *pachyderma* 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 $\delta^{18}O_{N. pachys}$ signals between 1103 and 1330 cm of the sediment core MD04-2718. **a**
 379 Signal of $\delta^{18}O_{N. pachys}$ (‰), **b** Number of total planktonic foraminifera (per g sed.), **c** 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).

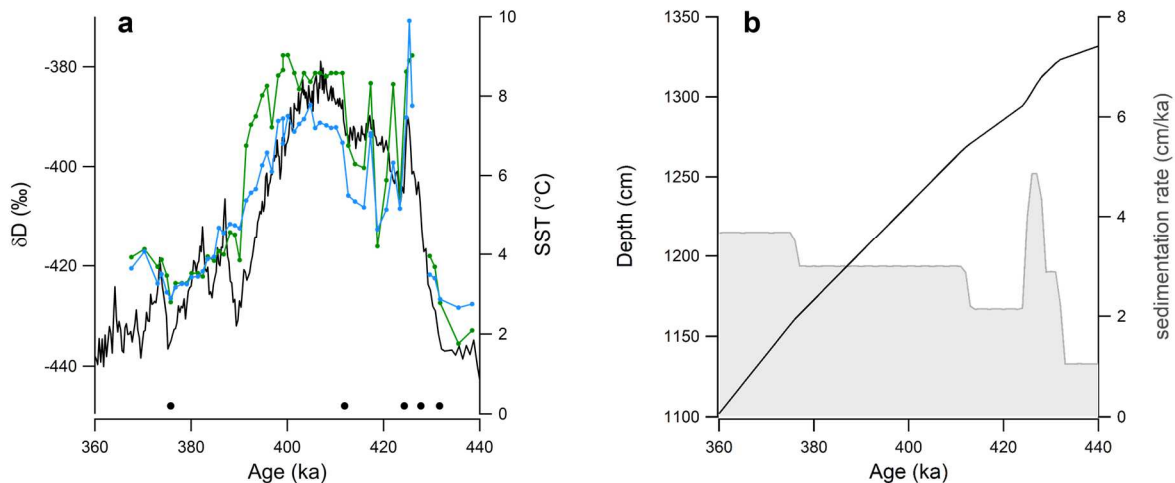
383

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 δD 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 δD 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⁻¹ with lowest values (1 cm.ka⁻¹) being recorded during MIS 12, and highest
 403 ones (4.9 cm.ka⁻¹) 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 CaCO₃ signal.
 406



407
 408 *Figure 5 a* Age model of MD04-2718 based on the correlation between MD04-2718 summer SST reconstructions and EDC δD
 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 δD 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 δD*
 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 CaCO₃ as a record of CCP strength

418

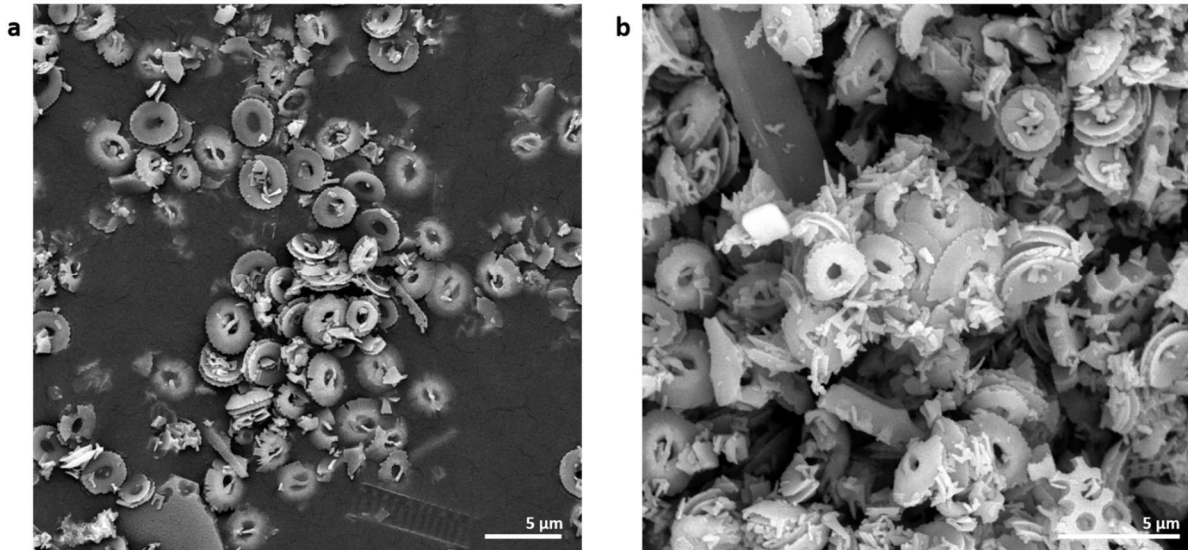
419 In pelagic environments of the SO, sedimentary carbonate concentration and/or coccolith
420 abundances have been assumed to reflect biogenic carbonate export production - and by inference
421 the strength of the CCP - based on the paradigm that in modern environments, and more particularly
422 in 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

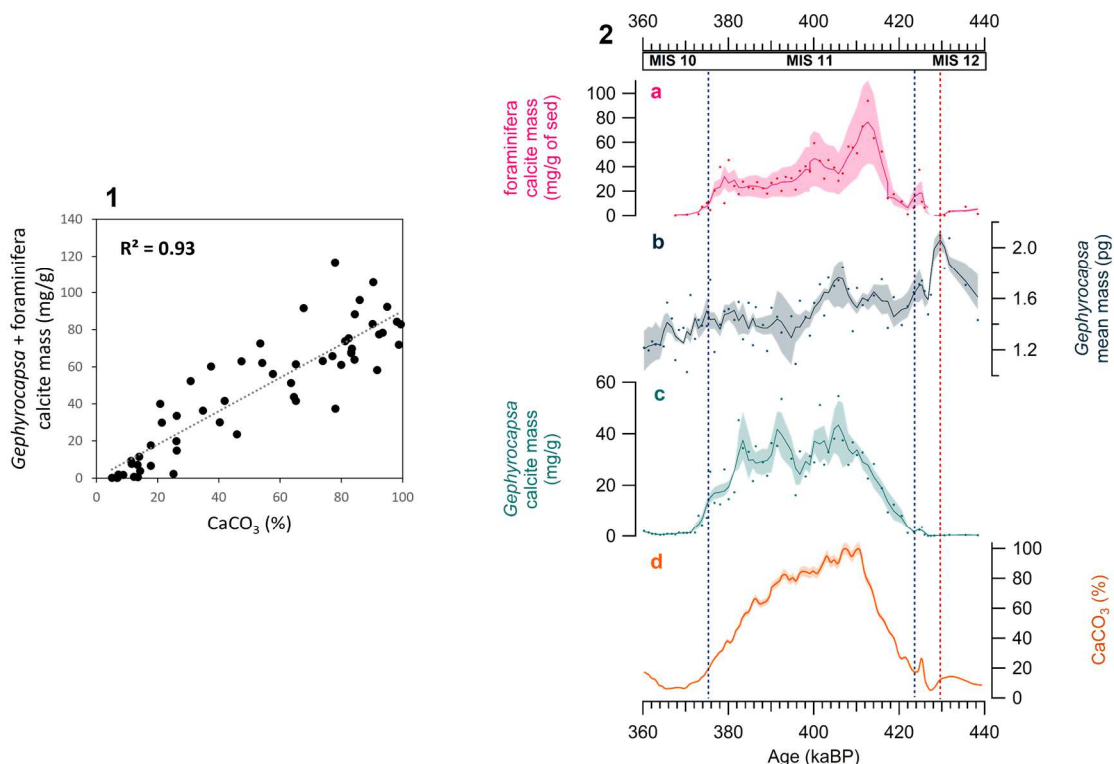
452 *Figure 6 Visual estimation of coccolith preservation of sediment core MD04-2718 based on Tescan VEGA II SEM observation*
453 *(Museum National d'Histoire Naturelle, MNHN, Paris) showing good preservation a at 389.1 ka and b at 405.8 ka.*

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 CaCO₃ 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 CaCO₃ content as shown by its greater correlation to CaCO₃ ($r^2 = 0.89$) 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 CaCO₃ 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 CaCO₃ 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 CaCO₃ (%). 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 CO₂ 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

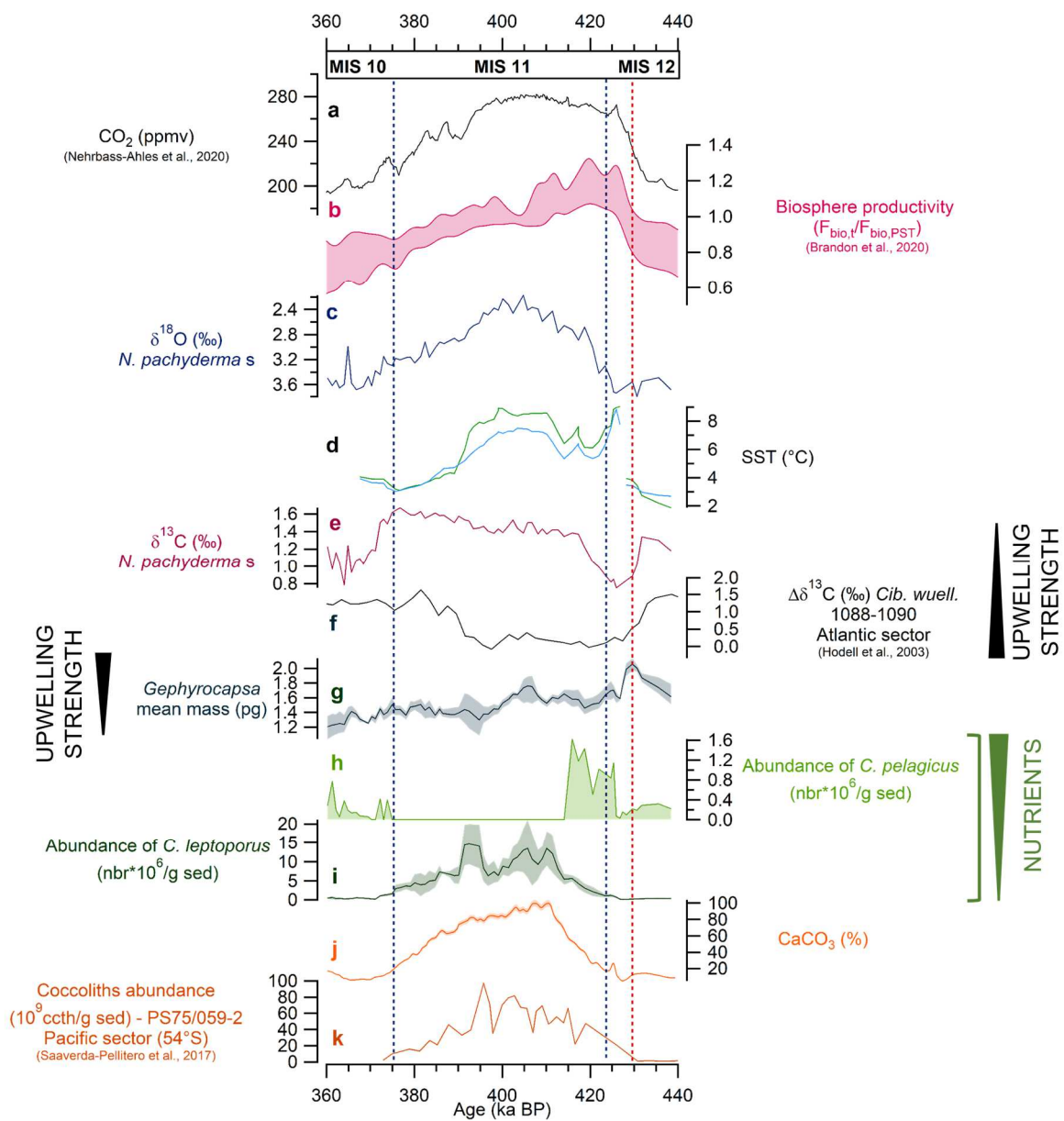
488

489 As represented by the variation in sedimentary CaCO₃ 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 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 CO₂ concentration. a*
539 *CO₂ (ppm) (Nehrbass-Ahles et al., 2020), b Biosphere productivity (F_{bio,t}/F_{bio,PST}) (Brandon et al., 2020), c δ¹⁸O 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 δ¹³C of N. pachyderma s. from MD04-2718, f Δδ¹³C 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 (*10⁶/g sed), i C. leptoporus abundance (*10⁶/g sed), j CaCO₃ (%) derived from Ca_{XRF}, k PS75/059-2*
544 *Coccoliths abundance (10⁹coccoliths/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 CO₂

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 pCO₂ 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 CO₂, thereby transiently buffering the impact of
556 oceanic CO₂ degassing (Fig. 8). pCO₂ 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 pCO₂ (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 pCO₂. From 398 ka to 381 ka, the gradual, ~35 ppm decrease in atmospheric pCO₂
566 coincided well with the rapid increase in benthic foraminifera Δδ¹³C 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 CO₂ 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 CO₂ 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 CO₂ concentrations over the period ranging between MIS
585 12 and MIS 10. First, using micropaleontological and geochemical data, we demonstrate that the
586 sedimentary CaCO₃ 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 pCO₂ pattern that characterises this key interval of the past. We show that the abrupt
595 atmospheric CO₂ 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 CO₂ 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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