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# 1 Origin of $\delta^{13}\text{C}$ minimum events in thermocline and intermediate waters of the western 2 South Atlantic

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## 17 Abstract

18 Stable carbon isotopic ( $\delta^{13}\text{C}$ ) minimum events have been widely described in marine  
19 archives recording the properties of thermocline and intermediate waters during glacial termi-  
20 nations. However, the mechanisms associated with these events remain ambiguous. Here we  
21 present three high temporal resolution deep-dwelling planktonic foraminifera  $\delta^{13}\text{C}$  records  
22 from the main thermocline and one benthic  $\delta^{13}\text{C}$  record from the modern core of Antarctic  
23 Intermediate Water (AAIW). Our sediment cores are distributed along the western South At-  
24 lantic from the equator to the subtropics, with the longest record spanning the last ~300 kyr.  
25 The results show that  $\delta^{13}\text{C}$  minimum events were pervasive features of the last three glacial  
26 terminations and Marine Isotope Stage 4/3 transition in the western South Atlantic. Two dis-  
27 tinct mechanisms were responsible for the  $\delta^{13}\text{C}$  minima at the thermocline and intermediate  
28 depths of the Atlantic, respectively. We suggest that the  $\delta^{13}\text{C}$  minimum events at the thermo-  
29 cline were mostly driven by the thermodynamic ocean-atmosphere isotopic equilibration,  
30 which is supported by calculated  $\delta^{13}\text{C}$  of dissolved inorganic carbon in the subtropical western

31 South Atlantic as well as by previously published model simulations. On the other hand, inter-  
32 mediate depths  $\delta^{13}\text{C}$  minimum events in the tropics were likely caused by the slowdown of the  
33 Atlantic meridional overturning circulation and the associated accumulation of isotopically light  
34 carbon at mid and intermediate depths of the Atlantic Ocean.

## 35 **1. Introduction**

36 Upwelling in the Southern Ocean (SO) connects the deep ocean with the atmosphere,  
37 by drawing deep waters to the SO surface (Marshall and Speer, 2012). The deep ocean is  
38 hypothesized to have been the main sink of atmospheric  $\text{CO}_2$  during glacial periods, as evi-  
39 denced by the accumulation of  $^{13}\text{C}$ -depleted remineralized carbon in poorly ventilated deep  
40 waters below ~2500 m water depth (Hodell et al., 2003; Howe et al., 2016a; Curry and Oppo,  
41 2005). Across glacial terminations, the breakup of deep-ocean stratification (e.g., Du et al.,  
42 2018; Basak et al., 2018) and the intensification of upwelling in the SO have been linked to  
43 the advection of nutrient-rich  $^{13}\text{C}$ -depleted waters from the deep ocean to the SO surface (An-  
44 derson et al., 2009; Toggweiler et al., 2006). This was followed by atmospheric  $\text{CO}_2$  increase  
45 and stable carbon isotopic ( $\delta^{13}\text{C}$ ) minimum events at thermocline and intermediate depths of  
46 the oceans (between 100 and 1200 m water depth) (e.g., Martínez-Botí et al., 2015; Ziegler et  
47 al., 2013; Schmitt et al., 2012; Spero and Lea, 2002). A widely held hypothesis behind these  
48  $\delta^{13}\text{C}$  minimum events is the northward transport of  $^{13}\text{C}$ -depleted carbon from the SO surface  
49 towards the equatorial regions by Subantarctic Mode Water (SAMW) and Antarctic Intermedi-  
50 ate Water (AAIW) (Spero and Lea, 2002). This mechanism is named the “oceanic tunnel” (Liu  
51 and Yang, 2003). In the Atlantic Ocean, the “oceanic tunnel” hypothesis is supported by the  
52 occurrence of  $\delta^{13}\text{C}$  minimum events in thermocline and intermediate depth records from the  
53 tropical region (e.g., Poggemann et al., 2017; Mulitza et al., 1998). However,  $\delta^{13}\text{C}$  minimum  
54 events have also been observed in regions far from the influence of SAMW and AAIW as, for  
55 example, in the North Atlantic (e.g., Lynch-Stieglitz et al., 2019; Rickaby and Elderfield, 2005).  
56 Additionally, intermediate depth records in subtropical latitudes of the Southern Hemisphere

57 lack evidence of  $\delta^{13}\text{C}$  minimum events. For instance, depth transects along the southern Bra-  
58 zilian margin ( $\sim 27^\circ\text{S}$ ) do not support the presence of a  $^{13}\text{C}$ -depleted northward flowing AAIW  
59 during Termination I (Lund et al., 2015; Tessin and Lund, 2013; Oppo and Horowitz, 2000).  
60 Accordingly, other mechanisms have been raised to explain the occurrence of the  $\delta^{13}\text{C}$  mini-  
61 mum events.

62 The outgassing of  $^{13}\text{C}$ -depleted  $\text{CO}_2$  to the atmosphere during glacial terminations re-  
63 sulted in an atmospheric  $^{13}\text{C}$  depletion similar to that observed in the upper ocean (Eggleston  
64 et al., 2016; Schmitt et al., 2012). Consequently, it has been suggested that temperature-  
65 mediated ocean-atmosphere isotopic equilibration could explain the widespread occurrence  
66 of  $\delta^{13}\text{C}$  minimum events in the ocean (Shao et al., 2021; Lynch-Stieglitz et al., 2019; Ninne-  
67 mann and Charles, 1997). According to this mechanism, the ocean-atmosphere isotopic equi-  
68 libration would globally imprint the  $\delta^{13}\text{C}$  minimum signal in the upper ocean and possibly at  
69 intermediate and mid-depth waters at the formation region of these water masses.

70 Alternatively,  $\delta^{13}\text{C}$  minima at intermediate depths of the tropical Atlantic may also have  
71 been caused by changes in the dynamics of the Atlantic meridional overturning circulation  
72 (AMOC) during glacial terminations, as indicated by model simulations (Menviel et al., 2018;  
73 Schmittner and Lund, 2015). Some studies have suggested that the reduced ventilation rate  
74 of North Atlantic Deep Water (NADW), which forms the lower limb of the AMOC, caused the  
75 observed  $^{13}\text{C}$  depletion in the mid-depth of the Atlantic Ocean during the last glacial termina-  
76 tion (Campos et al., 2020; Lacerra et al., 2017; Schmittner and Lund, 2015; Lund et al., 2015).  
77 Accordingly, this northern sourced  $^{13}\text{C}$ -depletion may have expanded upwards into intermedi-  
78 ate depths ( $\sim 1000$  m water depth) in the tropical Atlantic (Voigt et al., 2017; Freeman et al.,  
79 2015). Ultimately, this could imply a northern origin for the  $\delta^{13}\text{C}$  minimum signal at intermediate  
80 depths of tropical regions.

81 Here we investigate the occurrence and origin of the  $\delta^{13}\text{C}$  minimum events at the ther-  
82 mocline and intermediate depths of the western South Atlantic during the last three glacial  
83 terminations. We present three new high-resolution planktonic foraminiferal  $\delta^{13}\text{C}$  records from  
84 South Atlantic Central Water (SACW) and one record from AAIW extending from the equator

85 to the southern subtropical latitudes along the Brazilian margin (Fig. 1). Our results point to  
86 two distinct and independent mechanisms governing the  $\delta^{13}\text{C}$  minimum events at the thermo-  
87 cline and intermediate depths. At the thermocline,  $^{13}\text{C}$ -depletions were mostly driven by the  
88 ocean-atmosphere isotopic equilibration, while AMOC slowdown was mostly responsible for  
89 the  $^{13}\text{C}$ -depletion at intermediate depths.

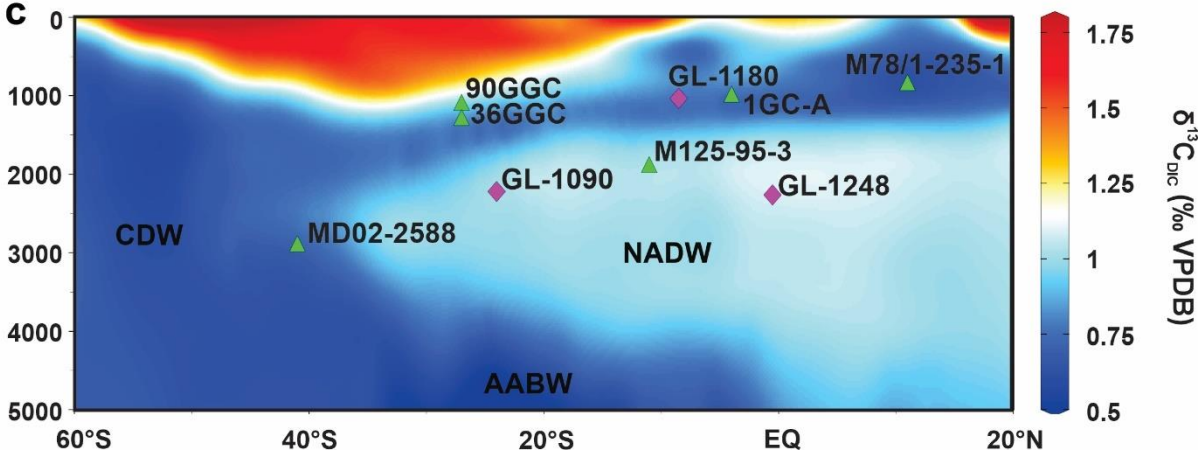
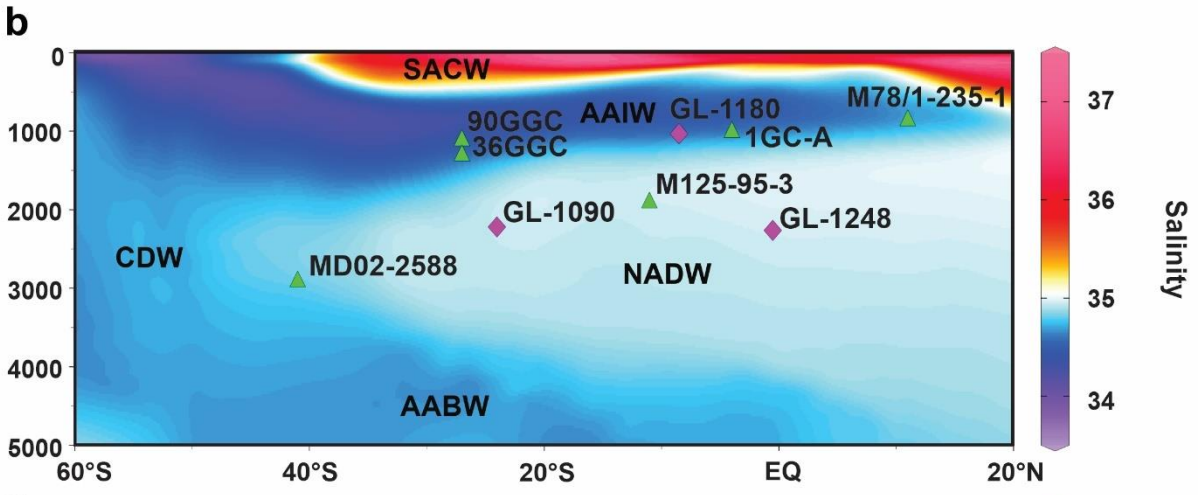
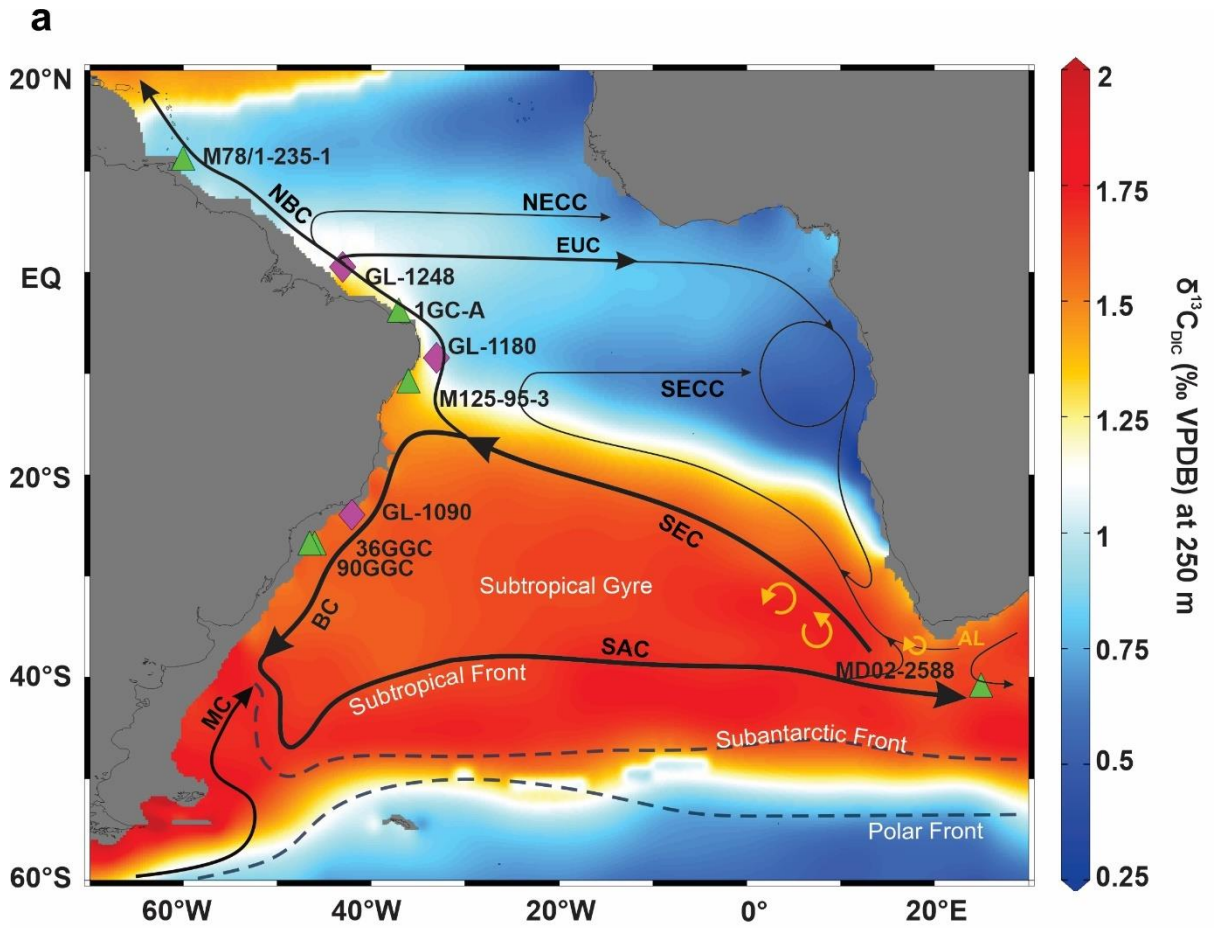
## 90 **2. Regional settings**

91 The South Atlantic upper-ocean circulation is dominated by the South Atlantic Sub-  
92 tropical Gyre (SASG) (Fig. 1a). The northernmost and the southernmost limits of the SASG  
93 are marked by the South Equatorial Current (SEC) and the South Atlantic Current (SAC),  
94 respectively. The westward-flowing SEC gives rise to the southward-flowing Brazil Current  
95 (BC) and the northward-flowing North Brazil Undercurrent and North Brazil Current. At about  
96  $38^\circ\text{S}$ , the BC converges with the northward-flowing Malvinas Current, giving rise to the SAC  
97 that flows eastwards. Ultimately, the SAC feeds the SEC, which returns to the Brazilian coast,  
98 closing the gyre (Peterson and Stramma, 1991).

99 The water masses filling thermocline and intermediate depths of the western South  
100 Atlantic are SACW and AAIW (Fig. 1b), together named Southern Ocean intermediate waters  
101 (SOIWs). The SACW occupies the South Atlantic thermocline between approximately 100 to  
102 500 m water depth. This water mass shows wide temperature ( $5^\circ\text{C}$  to  $20^\circ\text{C}$ ) and salinity (34.3  
103 to 36) ranges (Stramma and England, 1999). A portion of SACW is formed in the Brazil-Mal-  
104 vinas Confluence zone (BMC) region at the subtropical convergence (Sprintall and Tomczak,  
105 1993). During austral winter, the thick and homogeneous SAMW subducts in the subtropical  
106 convergence, producing SACW, which moves eastward within the SAC and recirculates in the  
107 SASG (Stramma and England, 1999). In the southeast Atlantic, another portion of SACW is  
108 formed by the injection of Indian Ocean waters into the South Atlantic in the Agulhas Current  
109 retroflection region (Poole and Tomczak, 1999; Sprintall and Tomczak, 1993; Gordon, 1986).  
110 Below the SACW, AAIW is marked by minimum salinity between approximately 500 to 1500  
111 m water depth (Fig. 1b). The temperature and salinity of the AAIW range between  $2^\circ\text{C}$  to  $6^\circ\text{C}$

112 and 33.8 to 34.8, respectively (Emery and Meincke, 1986). AAIW is mainly formed by over-  
113 turning convection in the extreme east of the South Pacific Ocean, enters the southwestern  
114 Atlantic through the Drake Passage, and is carried northward in the Malvinas Current (England  
115 et al. 1993). Northeast of the Drake Passage and at the BMC, mixing and eddy formation inject  
116 AAIW into the South Atlantic subsurface (Stramma and England 1999; Talley, 1996). Addi-  
117 tionally, substantial amounts of AAIW enter the Atlantic from the Indian Ocean through Agul-  
118 has Current leakage around the southern tip of Africa (Stramma and England, 1999).

119         Below AAIW, the NADW fills the mid-depth of the Atlantic Ocean between ~1500 and  
120 4000 m water depth. In the modern ocean, the core of NADW is centered at 2400 m water  
121 depth (Fig. 1b). However, during the Last Glacial Maximum (LGM), NADW is estimated to  
122 have shallowed to ~1800 m water depth (Gebbie, 2014). This water mass is formed in high  
123 latitudes of the North Atlantic through buoyancy loss and convection of dense upper ocean  
124 waters in the Labrador and Nordic Seas. NADW is featured by a salinity maximum (above  
125 34.9) at mid-depth of the Atlantic Ocean (Fig. 1b) and a temperature ranging between 2 and  
126 4°C (Tomczak and Godfrey, 1994).



128 **Figure 1.** Location of the sediment cores analyzed in this study (pink diamonds) and other  
129 sediment cores discussed within the text (green triangles). a) Atlantic Ocean pre-industrial  
130 stable carbon isotopic composition of the dissolved inorganic carbon ( $\delta^{13}\text{C}_{\text{DIC}}$ ) at 250 m water  
131 depth (Eide et al. 2017). Black arrows schematically represent the large-scale geostrophic  
132 currents between 100- and 500 m water depth. The dashed line depicts the Subantarctic and  
133 Polar Fronts (Orsi et al., 1995). Orange arrows depict the Agulhas leakage rings (adapted  
134 from Stramma and England, 1999). BC: Brazil Current; EUC: Equatorial Under Current; MC:  
135 Malvinas Current; NBC: North Brazil Current; NECC: North Equatorial Countercurrent; SAC:  
136 South Atlantic Current; SEC: South Equatorial Current; SECC: South Equatorial Countercur-  
137 rent. b) Western Atlantic Ocean meridional section of salinity (Zweng et al., 2018); c) Western  
138 Atlantic Ocean meridional section of the pre-industrial  $\delta^{13}\text{C}_{\text{DIC}}$  (Eide et al., 2017). In both sec-  
139 tions, AABW: Antarctic Bottom Water; AAIW: Antarctic Intermediate Water; CDW: Circumpolar  
140 Deep Water; NADW: North Atlantic Deep Water; SACW: South Atlantic Central Water. In all  
141 panels, purple diamonds indicate sediment cores GL-1248, GL-1180, and GL-1090 (this  
142 study), while green triangles indicate the sediment cores M125-95-3 (Campos et al., 2020),  
143 GS07-150-17/1GC-A (1GC-A in the figure) (Freeman et al., 2015), M78/1-235-1 (Poggemann  
144 et al., 2017), KNR159-5-90GGC (Lund et al., 2015), KNR-159-5-36GGC (Tessin and Lund,  
145 2013), and MD02-2588 (Ziegler et al., 2013). The figure was partially generated using the  
146 software Ocean Data View (Schlitzer, 2017).

147

148 The global deep ocean circulation can be separated in the upper and lower cells  
149 (Toggweiler et al., 2006; Talley, 2013). Over the SO, the westerlies force northward Ekman  
150 transport, which is compensated by the upwelling of Circumpolar Deep Water (CDW) in the  
151 Antarctic Zone (Marshall and Speer, 2012; Toggweiler and Samuels, 1995). In the upper cell,  
152 which is more limited to the Atlantic Ocean, SOIWs composed by the lightest portion of CDW  
153 will feed the formation of the NADW in the North Atlantic. In the lower cell, the densest portion  
154 of CDW moves towards the Antarctic continental shelf to form Antarctic Bottom Water  
155 (AABW), which flows northward, occupying the ocean abyss below approximately 4000 m



156 water depth (Pellichero et al., 2018; Abernathy et al., 2016; Talley, 2013). In the modern  
157 ocean, there is an interplay between these two cells (Talley, 2013).

158 The spatial distribution of South Atlantic stable carbon isotopic composition of dis-  
159 solved inorganic carbon ( $\delta^{13}\text{C}_{\text{DIC}}$ ) at 250 m water depth (Fig 1a) reveals the oligotrophic and  
160 high  $\delta^{13}\text{C}_{\text{DIC}}$  SASG (warm tones in Fig. 1a), while in the equatorial and eastern tropical Atlantic  
161 (Guinea and Angola domes), reduced ventilation and high productivity and remineralization of  
162 organic carbon result in  $^{13}\text{C}_{\text{DIC}}$ -depleted subsurface waters (cold tones in Fig. 1a). The merid-  
163 ional  $\delta^{13}\text{C}_{\text{DIC}}$  section from the western tropical and South Atlantic Ocean (Fig 1c) exhibits the  
164 pattern of  $^{13}\text{C}$ -depletion at thermocline and intermediate depths in the tropics. Oppo et al.  
165 (2018) show that remineralization is responsible for up to 1‰ of  $\delta^{13}\text{C}_{\text{DIC}}$  depletion at thermo-  
166 cline and intermediate depth in the western tropical Atlantic. Approximately, 60% of this re-  
167 mineralization occurs near the formation site of AAIW, while the rest occurs as AAIW advects  
168 northward. The vertical  $\delta^{13}\text{C}_{\text{DIC}}$  profile at about 20°S is marked by a decrease from the oligo-  
169 trophic SASG (~1.5 ‰) downward through the SACW and AAIW until reaching the minimum  
170 of ~0.7 ‰ at 1300 m water depth, reflecting the influence of Upper Circumpolar Deep Water.  
171 Below that,  $\delta^{13}\text{C}_{\text{DIC}}$  increases again due to the  $^{13}\text{C}_{\text{DIC}}$ -enriched NADW (~1 ‰) that flows south-  
172 ward between 1500 and 4000 m water depth. Finally, in the South Atlantic abyss, the  $^{13}\text{C}_{\text{DIC}}$ -  
173 depleted AABW (~0.5 ‰) flows northward, underlying NADW.

### 174 3. Methods

#### 175 3.1. Sediment cores

176 In this study, we present new data from sediment cores GL-1248 (0°55'20" S,  
177 43°24'10" W, 2264 m water depth, 19.29 m long) from the Brazilian equatorial margin, GL-  
178 1180 (8°27'18" S, 33°32'53" W, 1037 m water depth, 17.32 m long) from the Brazilian tropical  
179 margin and GL-1090 (24°55'12" S, 42°30'36" W, 2225 m water depth, 19.14 m long) from the  
180 Brazilian subtropical margin (Fig 1). All sediment cores were provided by the Brazilian oil com-  
181 pany Petrobras. A visual analysis of the core sections does not indicate any sedimentation  
182 disturbance, except for a hiatus between 2.18 and 1.70 m in core GL-1248 (Venancio et al.,

183 2018). The three cores were sampled at 2 cm resolution with sample volume of approximately  
184 10 cm<sup>3</sup>. Samples were wet-sieved to retain the fraction larger than 63 μm. The retained mate-  
185 rial was dried at 50°C for 24 hours and stored in acrylic containers. Foraminifera shells were  
186 handpicked using a binocular microscope.

### 187 **3.2. Age model**

188 We used the published age model of core GL-1248 (Venancio et al., 2018), based on  
189 twelve calibrated radiocarbon ages and the visual alignment of its Ti/Ca ratio to the Greenland  
190 stable oxygen isotope ( $\delta^{18}\text{O}$ ) record (NGRIP members, 2004), using the extended Greenland  
191 Ice Core Chronology (GICC05modelext) (Wolff et al., 2010). This alignment assumes that  
192 Ti/Ca peaks resulting from increased precipitation in the northeast Brazilian region correspond  
193 to North Atlantic cold events (e.g., Zhang et al., 2017). GL-1248 was sampled relatively close  
194 to the Parnaíba River mouth and was significantly influenced by terrigenous input during peri-  
195 ods of low sea levels and high precipitation periods in northeastern Brazil (Venancio et al.,  
196 2018). The age model was built using the software Clam 2.2 (Blaauw, 2010), with GL-1248  
197 spanning the last 124 ka, and was described in detail by Venancio et al. (2018).

198 The age model of core GL-1180 was published by Nascimento et al. (2021) and is  
199 based on six calibrated radiocarbon ages (Supporting Information, Table S1) and the align-  
200 ment of the benthic  $\delta^{18}\text{O}$  record of *Cibicides* spp. with the global  $\delta^{18}\text{O}$  stack LR04 (Lisiecki and  
201 Raymo, 2005) (Table S2 and Fig. S1). The benthic  $\delta^{18}\text{O}$  tie point uncertainties were calculated  
202 following the recommendations by Govin et al. (2015). The calculation used the uncertainties  
203 of the reference stack and of the alignment procedure, as well as the resolutions of the refer-  
204 ence stack and of the target record. Radiocarbon ages were calibrated using the IntCal13  
205 calibration curve (Reimer et al., 2013), with a reservoir effect of  $400 \pm 200$  without any addi-  
206 tional local reservoir effect. The age model (Fig. S1) was built using the software Bacon v 2.3  
207 (Blaauw and Christen, 2011); it shows that sediment core GL-1180 spans the last 300 kyr.  
208 More details about the age model can be found in Nascimento et al. (2021).

209 For core GL-1090, we use the age model previously published by Santos et al. (2017)  
210 and improved by Ballalai et al. (2019) and Santos et al. (2020). It is based on ten calibrated  
211 radiocarbon ages and the alignment of its benthic  $\delta^{18}\text{O}$  record of (*Cibicides wuellerstorfi*) with  
212 the benthic  $\delta^{18}\text{O}$  records from the reference curves LR04 (Lisiecki and Raymo, 2005) and  
213 MD95-2042 (Govin et al., 2014). The age model was built using the software Bacon v 2.3  
214 (Blaauw and Christen, 2011), with GL-1090 covering the last 182 kyr.

### 215 3.3. Foraminifera stable carbon isotopes ( $\delta^{13}\text{C}$ ) analysis

216 Our  $\delta^{13}\text{C}_{\text{DIC}}$  reconstruction of SACW is based on the  $\delta^{13}\text{C}$  data from thermocline-  
217 dwelling foraminifera *Neogloboquadrina dutertrei* (GL-1248), *Globorotalia truncatulinoides*  
218 (dextral) (GL-1180), and *Globorotalia inflata* (GL-1090). The apparent calcification depth  
219 (ACD) of *N. dutertrei* was estimated to be 150 m in the region of core GL-1248 (Venancio et  
220 al. 2018), in line with previous estimates in the tropical Atlantic (Cl  roux et al., 2013; Farmer  
221 et al., 2007; Steph et al., 2009). *G. truncatulinoides* presents a wide vertical migration range  
222 during its life cycle but primarily calcifies at the main thermocline (Steph et al., 2009; Regen-  
223 berg et al., 2009; Cl  roux et al., 2007; LeGrande et al., 2004), which is consistent with an ACD  
224 of ~250 m in the region of core GL-1180 (Nascimento et al., 2021). In the western South At-  
225 lantic, the *G. inflata* ACD stably ranges from 350 to 400 m water depth (Groeneveld and  
226 Chiessi, 2011).

227 Between 5 and 10 shells of *N. dutertrei* (350-415  $\mu\text{m}$ ), *G. truncatulinoides* (dextral;  
228 300-425  $\mu\text{m}$ ), and *G. inflata* (250-300  $\mu\text{m}$ ) were handpicked from samples of cores GL-1248,  
229 GL-1180, and GL-1090, respectively. The sampling resolution varied between 2 and 4 cm  
230 downcore.

231 Our reconstruction of seawater  $\delta^{13}\text{C}$  for intermediate depth is based on  $\delta^{13}\text{C}$  from ben-  
232 thic foraminifera *Cibicides* spp. from core GL-1180 (1037 m water-depth). To do so, between  
233 5 and 10 shells of *Cibicides* spp. (300-350  $\mu\text{m}$ ), composed by *Cibicides pachyderma*, *Cibicides*  
234 *refulgens*, and *Cibicides kullenbegi* (Supporting Information, Text S1), were handpicked at 4  
235 cm resolution.

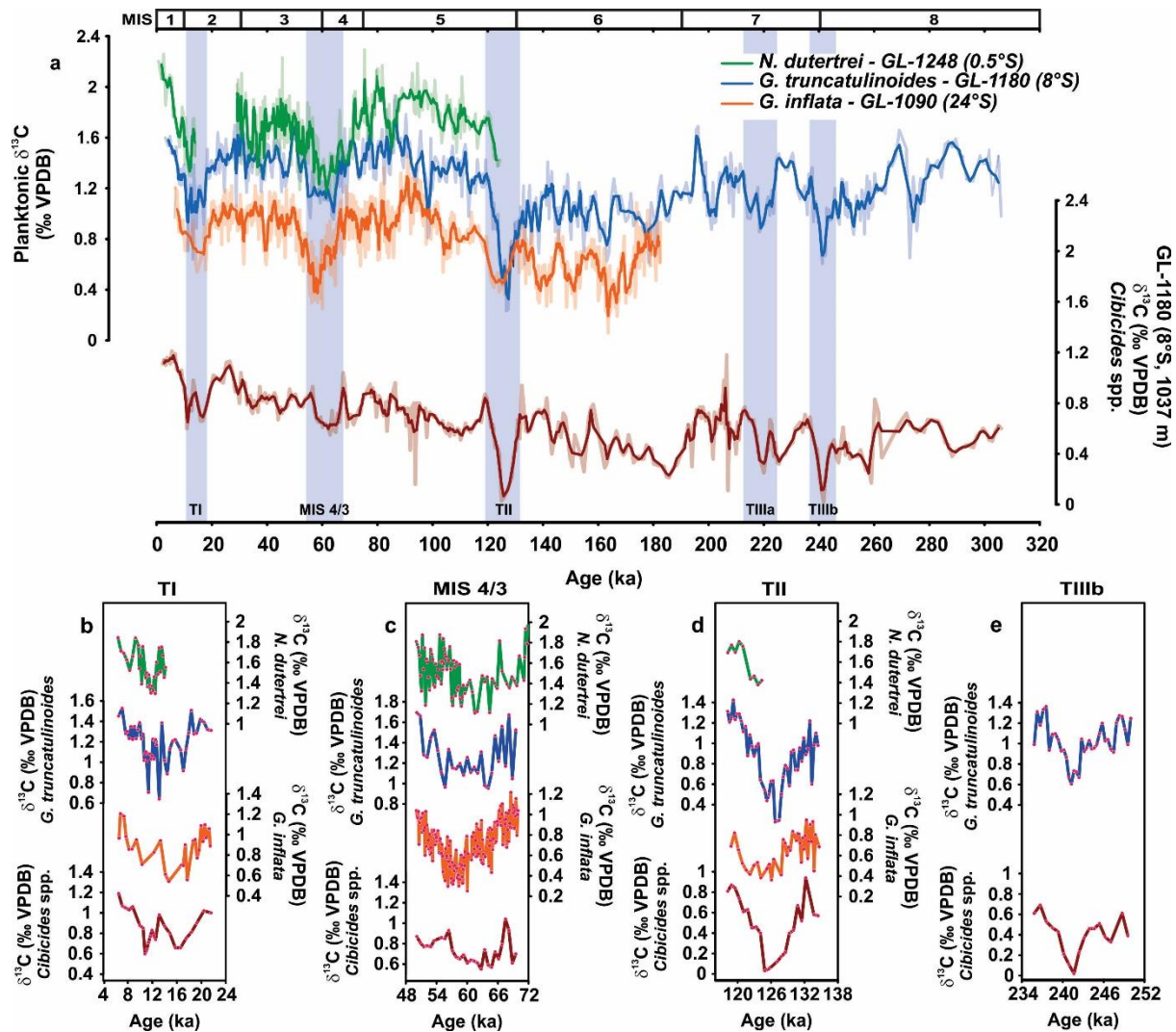
236 Analyses of planktonic foraminifera from the cores GL-1248 and GL-1090, and benthic  
237 foraminifera from GL-1180 were carried out in the Center for Marine Environmental Sciences  
238 (MARUM), University of Bremen, Germany, using a Finnigan MAT 251 isotope ratio mass  
239 spectrometer (IRMS) equipped with an automated carbonate preparation device (Kiel I). Data  
240 were calibrated against the in-house standard (Solnhofen limestone), and the standard devi-  
241 ation of the in-house standard over the measurement period was 0.03 ‰. Analyses of plank-  
242 tonic foraminifera from core GL-1180 were performed in the Paleooceanography and Paleocli-  
243 matology Laboratory, University of São Paulo, Brazil, using a Thermo™ Scientific MAT 253  
244 IRMS coupled to a Thermo™ Kiel IV automated carbonate preparation device (Kiel IV) (Crivel-  
245 lari et al., 2021). Data were calibrated against the NBS 19 standard. The standard deviation  
246 of NBS 19 over the measurement period was 0.02 ‰. All data are presented in parts per  
247 thousand (‰) relative to the Vienna Pee Dee belemnite (VPDB).

#### 248 4. Results

249 The  $\delta^{13}\text{C}$  of *N. dutertrei* (GL-1248) ranges between 1.11 and 2.29 ‰ (mean  $1.70 \pm$   
250  $0.17$  ‰), while the  $\delta^{13}\text{C}$  range is from 0.23 to 1.93 ‰ (mean  $1.21 \pm 0.21$  ‰) for *G. truncatuli-*  
251 *noides* (GL-1180), and between 0.06 to 1.47 ‰ (mean  $0.80 \pm 0.19$  ‰) for *G. inflata* (GL-1090)  
252 (Fig. 2). Planktonic  $\delta^{13}\text{C}$  records from all studied cores present a similar pattern with remark-  
253 able negative excursions during glacial terminations that only recover after the onset of the  
254 following interglacial period (Fig. 2). The records show little difference between glacial and  
255 interglacial  $\delta^{13}\text{C}$  values. A prominent negative excursion is also observed during the transition  
256 between Marine Isotope Stage (MIS) 4 and MIS 3. In general, the benthic  $\delta^{13}\text{C}$  record from  
257 core GL-1180 is similar to the thermocline  $\delta^{13}\text{C}$  records (Fig 2a). The average benthic  $\delta^{13}\text{C}$  is  
258  $0.64 \pm 0.17$  ‰, and the values range between 0.02 and 1.49 ‰, similar to *G. inflata*  $\delta^{13}\text{C}$  from  
259 core GL-1090.

260 Termination IIIb (Fig 2d) is only recorded in sediment core GL-1180, which covers the  
261 last 300 kyr. *G. truncatulinoides* and *Cibicides* spp.  $\delta^{13}\text{C}$  records show a drop of 0.4 and 0.5  
262 ‰, respectively, relative to MIS 8.  $\delta^{13}\text{C}$  minima are also observed during Termination IIIa at

263 the transition between MIS 7d and MIS 7c. A more pronounced  $\delta^{13}\text{C}$  minimum occurred during  
264 Termination II (Fig 2c), when *G. truncatulinoides* and *G. inflata*  $\delta^{13}\text{C}$  values drop by  $\sim 0.8$  and  
265  $\sim 0.4$  ‰, respectively. Similarly, during Termination II, the benthic  $\delta^{13}\text{C}$  record dropped by  $\sim 0.6$   
266 ‰ relative to MIS 6 (Fig 2c). The *N. dutertrei*  $\delta^{13}\text{C}$  record spans the last 124 ka only, and its  
267 oldest samples indicate that  $\delta^{13}\text{C}$  is recovering from a minimum of at least 1.40 ‰ to reach  
268 interglacial values of  $\sim 1.8$  ‰ at 121 ka. All records show negative  $\delta^{13}\text{C}$  excursions during  
269 MIS4/3 transition. These excursions are equivalent to those observed during glacial termina-  
270 tions. The largest drops are exhibited in *N. dutertrei* (GL-1248) and *G. inflata* (GL-1090), where  
271  $\delta^{13}\text{C}$  values decrease by  $\sim 0.5$  ‰. *G. truncatulinoides* and *Cibicides spp.* (GL-1180) show a  
272  $\delta^{13}\text{C}$  decrease of  $\sim 0.3$  ‰. During Termination I, *G. inflata* and *G. truncatulinoides*  $\delta^{13}\text{C}$  drops  
273 approximately 0.4 ‰ relative to MIS 2 (Fig. 2b). *G. truncatulinoides*  $\delta^{13}\text{C}$  presents several  
274 short-lived  $\delta^{13}\text{C}$  drops within a broad negative excursion that started at  $\sim 18$  ka. For *G. inflata*  
275 and *G. truncatulinoides* records, fully interglacial values are reached at  $\sim 7$  ka. GL-1248 pre-  
276 sents a hiatus between 29 and 14 ka, but we note that  $\delta^{13}\text{C}$  is low at the end of Termination I  
277 and reaches interglacial values in the mid-Holocene. During Termination I, the  $\delta^{13}\text{C}$  record of  
278 *Cibicides spp.* shows two negative excursions of up to  $\sim 0.4$  ‰ relative to the Last Glacial Max-  
279 imum (LGM) (Fig 2b).



280

281 **Figure 2.** Carbon isotopic composition ( $\delta^{13}\text{C}$ ) of planktonic and benthic foraminifera analyzed  
 282 in this study. a)  $\delta^{13}\text{C}$  of *Neogloboquadrina dutertrei* (GL-1248, green line), *Globorotalia trun-*  
 283 *catulinoides* (GL-1180, blue line), *Globorotalia inflata* (GL-1090, orange line), *Cibicides* spp.  
 284 (GL-1180, dark-red line). Blue bars indicate  $\delta^{13}\text{C}$  minimum events during glacial terminations  
 285 and the Marine Isotope Stage (MIS) 4/3 transition. Dark colored lines depict the three points  
 286 running average, while light colored lines depict the original data. Zooms of deep-dwelling and  
 287 benthic  $\delta^{13}\text{C}$  records for b) zoom into Termination I (TI); c) zoom into MIS4/3 transition; d)  
 288 zoom into Termination II (TII); e) zoom into Termination IIIb (TIIIb) colors as in a). MIS are  
 289 indicated at the top of panel a.

## 290 5. Discussion

### 291 5.1. $\delta^{13}\text{C}$ values of planktonic foraminifera

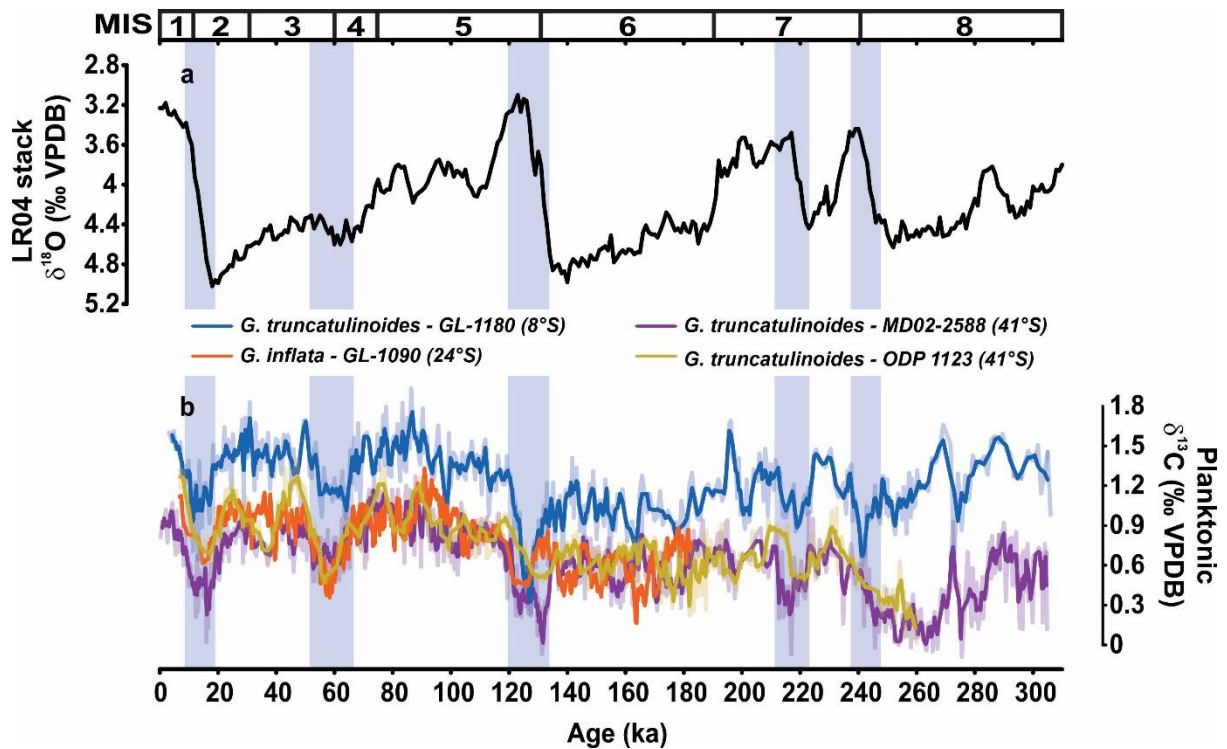
292 We present  $\delta^{13}\text{C}$  values of three deep-dwelling planktonic foraminifera species in order  
293 to reconstruct  $\delta^{13}\text{C}_{\text{DIC}}$  of thermocline waters in the western South Atlantic. Although  $\delta^{13}\text{C}$  of  
294 the calcite reflects  $\delta^{13}\text{C}_{\text{DIC}}$ , it is not in isotopic equilibrium with seawater (Ravelo and Hillaire-  
295 Marcel, 2007). Offsets between  $\delta^{13}\text{C}_{\text{DIC}}$  and  $\delta^{13}\text{C}$  of calcite are related to abiotic kinetic frac-  
296 tionation, which causes an offset of  $1.0 \pm 0.2$  ‰ (Romanek et al., 1992), changes in foraminif-  
297 era metabolic rate (Bemis et al., 2000), seawater carbonate ion concentration (Spero et al.,  
298 1997; Wilke et al., 2006), and photosymbiosis (Spero and Deniro, 1987). The  $\delta^{13}\text{C}$  from *G.*  
299 *truncatulinoides* and *G. inflata* is known to be dependent on shell size (Birch et al., 2013;  
300 Friedrich et al., 2012; Elderfield et al., 2002). This is likely an effect of high metabolic rates of  
301 individuals in early ontogenetic stages but analyzing shells larger than 250  $\mu\text{m}$  largely avoids  
302 this effect (Birch et al., 2013; Friedrich et al., 2012). Even so, the  $\delta^{13}\text{C}$  average of 11 core-tops  
303 ( $1.0$  ‰  $\pm$   $0.2$ ) from the western South Atlantic (Chiessi et al., 2007) shows that *G. inflata* is  
304  $\sim 0.5$  ‰ lighter than the pre-industrial  $\delta^{13}\text{C}_{\text{DIC}}$  provided by Eide et al. (2017) in the same region  
305 between 350 to 400 m water depth. The same result was found in sediment trap samples from  
306 the Cape Basin (Wilke et al., 2006), suggesting that this offset is a widespread feature of this  
307 species in the South Atlantic. In the opposite direction, plankton tow samples from the equa-  
308 torial Atlantic reveal that *N. dutertrei*  $\delta^{13}\text{C}$  is about 0.5 ‰ higher than  $\delta^{13}\text{C}_{\text{DIC}}$  (Mulitza et al.,  
309 1999). Therefore,  $\delta^{13}\text{C}$  differences or similarities between foraminifera species may not only  
310 result from habitat depth but also from offsets between  $\delta^{13}\text{C}_{\text{DIC}}$  and  $\delta^{13}\text{C}$  of calcite. Yet, these  
311 effects mostly affect the absolute  $\delta^{13}\text{C}$  values and are not expected to influence temporal  
312 changes in the downcore records, which is the central aspect discussed here.

313 The  $\delta^{13}\text{C}$  of *G. truncatulinoides* is considered a robust proxy for nutrient content in the  
314 thermocline (Mulitza et al., 1998). *G. truncatulinoides*  $\delta^{13}\text{C}$  record from GL-1180 presents a  
315 positive offset of  $\sim 0.5$  ‰ relative to the same species in the Agulhas Plateau (MD02-2588,  
316  $41^\circ\text{S}$ ) (Ziegler et al., 2013) and in the western South Pacific (ODP1123,  $41^\circ\text{S}$ ) (Hu et al., 2020)  
317 (Fig. 3b). Since sites MD02-2588 and ODP1123 are near central waters formation latitudes,  
318 we would presume their *G. truncatulinoides*  $\delta^{13}\text{C}$  values to be higher than in the aged central  
319 waters in the tropics. The positive offset of *G. truncatulinoides*  $\delta^{13}\text{C}$  in the tropics relative to

320 the subtropics may be related to ecological distinctions between organisms living in these  
321 regions, morphotype (dextral or sinistral), and the shell size. In the stratified tropical Atlantic,  
322 the ACD range of *G. truncatulinoides* is estimated to extend from the base of the seasonal  
323 thermocline to ~400 m water depth (Mulitza et al., 1997). In the subtropics, this species ex-  
324 tends its life cycle from the weakly stratified ocean surface to ~800 m water depth (Mulitza et  
325 al., 1997; Ujiie et al., 2010; Hu et al., 2020). In fact, this large difference in the depth range  
326 may favor the *G. truncatulinoides* secondary calcite crust, which forms deeper in the water  
327 column (Mulitza et al., 1997), to record lower  $\delta^{13}\text{C}$  values in the subtropics relative to the trop-  
328 ical ocean. Additionally, in contrast to Hu et al. (2020) and Ziegler et al. (2013) who analyzed  
329 the sinistral morphotype of *G. truncatulinoides*, we used the *G. truncatulinoides* dextral. In-  
330 deed, it was suggested that the dextral morphotype shows a shallower ACD and prefers a  
331 more stratified thermocline than the sinistral (Feldmeijer et al., 2014; Ujiie et al., 2010). Be-  
332 sides, the size fraction analyzed in our study is larger than in the aforementioned studies,  
333 reducing the effect of metabolic rate on the  $\delta^{13}\text{C}$  of calcite (Birch et al., 2013; Friedrich et al.,  
334 2012). Mixing with high  $\delta^{13}\text{C}$  of SASG waters (Fig. 1a, c) may also contribute to increasing  
335 the thermocline  $\delta^{13}\text{C}$  in the tropics relative to the subtropics (Fig. 3b).

336         Despite the difference in location and species measured within GL-1090 (i.e. *G. inflata*  
337 from the SASG) to those of Ziegler et al. (2013) and Hu et al. (2020) (i.e. *G. truncatulinoides*  
338 from Agulhas Plateau and western South Pacific, respectively), the three records present sim-  
339 ilar  $\delta^{13}\text{C}$  values (Fig. 3b). We ascribe this similarity to the negative ~0.5 ‰ offset of *G. inflata*  
340  $\delta^{13}\text{C}$  relative to  $\delta^{13}\text{C}_{\text{DIC}}$  (Wilke et al., 2006) and to the deep ACD of *G. inflata* in the western  
341 South Atlantic (Groeneveld and Chiessi, 2011). The higher averaged  $\delta^{13}\text{C}$  values observed in  
342 the *N. dutertrei* record (GL-1248) (Fig. 2a) may reflect the shallow habitat within the thermo-  
343 cline and the additional positive offset (0.5 ‰) relative to  $\delta^{13}\text{C}_{\text{DIC}}$  observed in the equatorial  
344 Atlantic (Mulitza, 1999).





345

346 **Figure 3.** Stable carbon isotopic ( $\delta^{13}\text{C}$ ) records presented in this study compared with pub-  
 347 lished thermocline (see locations in Fig. 1) records. a) LR04 global oxygen stable isotopic  
 348 ( $\delta^{18}\text{O}$ ) stack (Lisiecki and Raymo, 2005). b)  $\delta^{13}\text{C}$  of *Globorotalia truncatulinoides* from the  
 349 western tropical South Atlantic (GL-1180, blue line), *Globorotalia inflata* from the western sub-  
 350 tropical South Atlantic (GL-1090, orange line), *G. truncatulinoides* from the Agulhas Plateau  
 351 (MD02-2588, purple line) (Ziegler et al., 2013), and *G. truncatulinoides* from the western South  
 352 Pacific (ODP1123, dark yellow) (Hu et al., 2020). Dark colored lines depict the three points  
 353 running average, while light colored lines depict the original data. Blue bars indicate  $\delta^{13}\text{C}$  min-  
 354 imum events during glacial terminations and the Marine Isotope Stage (MIS) 4/3 transition.  
 355 MIS are indicated at the top of the panel a.

## 356 5.2. Mechanisms behind $\delta^{13}\text{C}$ minimum events at thermocline and intermediate depth

### 357 5.2.1. The oceanic tunnel hypothesis

358 The deglacial  $\delta^{13}\text{C}$  minimum events shown by our records can be potentially explained  
 359 by the transport of  $^{13}\text{C}$ -depleted carbon from the deep ocean to the SO surface and its equa-  
 360 torward advection by SOIW (Spero and Lea, 2002), the so-called “oceanic tunnel” hypothesis  
 361 (e.g., Pena et al., 2013), in reference to Liu and Yang (2003). The deep ocean is hypothesized

362 as the main sink of atmospheric CO<sub>2</sub> during glacial periods. During glacial terminations, the  
363 SO warmed (WAIS members, 2013; Barker et al. 2009), the sea-ice cover shrank (WAIS mem-  
364 bers, 2013; Gersonde and Zielinski, 2000), and the Southern Hemisphere westerlies are  
365 thought to have shifted southward and strengthened (Menviel et al., 2018; Toggweiler et al.,  
366 2006). The reduction in brine rejection due to reduced sea-ice cover, together with the stronger  
367 westerlies, resulted in the breakup of deep ocean stratification and enhancement of its venti-  
368 lation (Du et al., 2018; Basak et al., 2018; Skinner et al., 2010), transporting nutrient-rich and  
369 <sup>13</sup>C-depleted waters from the deep ocean to the SO surface (Anderson et al., 2009; Ziegler et  
370 al., 2013). Finally, the low δ<sup>13</sup>C signal was transmitted northward by the SOIWs (Spero and  
371 Lea, 2002). Spero and Lea (2002) hypothesized a correspondence between the δ<sup>13</sup>C<sub>DIC</sub> sig-  
372 nals of central waters and AAIW for the “oceanic tunnel” hypothesis to be valid. Indeed, our  
373 planktonic records share a similar structure with the benthic record from GL-1180 (Fig 2a).  
374 Since modern SACW and AAIW have a similar source and formation mechanism (Pellichero  
375 et al., 2018; Abernathey et al., 2016; Tomczak and Godfrey, 1994), our results, at least theo-  
376 retically, corroborate the subsurface link between the SO and low latitudes, which resulted in  
377 the δ<sup>13</sup>C minima observed in the western South Atlantic over the last three glacial terminations.

378 Ninnemann and Charles (1997) speculated that regions with little or no influence of  
379 SOIWs should have a different δ<sup>13</sup>C signal from that seen in the Southern Hemisphere. How-  
380 ever, δ<sup>13</sup>C minima have also been found in thermocline records of the subtropical North Atlan-  
381 tic, far from the influence of SOIWs (Lynch-Stieglitz et al., 2019). Besides, at intermediate  
382 depth, the occurrence of these events was limited to the tropical (e.g., Poggemann et al., 2017;  
383 Freeman et al., 2015; Arz et al., 1999) and North Atlantic (e.g., Oppo et al., 2015; Rickaby and  
384 Elderfield, 2005). In contrast, intermediate depth δ<sup>13</sup>C records from subtropical latitudes of the  
385 southern Brazilian margin (~27°S) revealed no evidence of <sup>13</sup>C-depleted northward-flowing  
386 AAIW during the last glacial termination (Lund et al., 2015; Tessin and Lund, 2013; Oppo and  
387 Horowitz, 2000). Analogous results were reported at intermediate depth of the eastern sub-  
388 tropical Pacific Ocean (Pahnke and Zahn, 2005; Bostock et al., 2004). Additionally, intermedi-  
389 ate depth Δ<sup>14</sup>C records (a proxy of water mass ventilation) from the SO do not point to a poorly

390 ventilated AAIW during the last termination (Burke and Robinson, 2012). Similarly,  $\Delta^{14}\text{C}$  rec-  
391 ords from the southern Brazilian margin (Sortor and Lund, 2011) and equatorial Atlantic (Clé-  
392 roux et al., 2011; Chen et al., 2020) give support to persistent well-ventilated thermocline and  
393 intermediate waters during Termination I. Together, stable and radiogenic carbon isotopes,  
394 suggest that the SOIWs were not a pathway for aged and  $^{13}\text{C}$ -depleted carbon accumulated  
395 in the deep ocean during glacial periods. Therefore, the “oceanic tunnel” mechanism is not  
396 expected to have played a major role in causing the  $\delta^{13}\text{C}$  minimum events in our records and  
397 in other records from the tropical and North Atlantic.

### 398 5.2.2. *The thermodynamic ocean-atmosphere isotopic equilibration hypothesis*

399  $\delta^{13}\text{C}$  minimum events in the atmosphere (Fig. 4a) (Eggleston et al., 2016; Schmitt et  
400 al., 2012) have also been ascribed to the enhanced upwelling around the Antarctica and the  
401 outgassing of  $^{13}\text{C}$ -depleted  $\text{CO}_2$  from the SO ocean surface (Schmitt et al., 2012). Previous  
402 studies have suggested that the atmosphere can work as a bridge, globalizing its low atmos-  
403 pheric  $\delta^{13}\text{C}$  signal in the upper ocean through the temperature-mediated ocean-atmosphere  
404 isotopic equilibration, also named thermodynamic equilibration (Shao et al., 2021; Lynch-  
405 Stieglitz et al., 2019). Accordingly, the deglacial  $\delta^{13}\text{C}$  minima observed in thermocline and in-  
406 termediate depths in the western South Atlantic (Fig. 2a) may have been caused by the ther-  
407 modynamic equilibration in the formation regions of the water masses that fill these depths  
408 (Lynch-Stieglitz et al. 2019; Lynch-Stieglitz et al., 1995; Lynch-Stieglitz and Fairbanks, 1994;  
409 Broecker and Maier-Reimer, 1992). This hypothesis explains, for example, the presence of  
410 low  $\delta^{13}\text{C}$  signal in oceanic regions where the nutrient concentrations are thought to have been  
411 always low and in regions that are not affected by the northward flowing SOIWs, as the North  
412 Atlantic (Lynch-Stieglitz et al., 2019; Ninnemann and Charles, 1997).

413 Empirical experiments indicate that the  $\delta^{13}\text{C}$  at the ocean surface tend to be 0.1 ‰  
414 higher relative to the atmospheric value per degree of cooling (Zhang et al., 1995). To estimate  
415 the influence of the thermodynamic equilibration in the western South Atlantic, we used the  
416 sea surface temperature (SST) record from site GL-1090 (Fig. 4b) (Santos et al., 2017) and

417 the isotopic composition of atmospheric CO<sub>2</sub> over the last 150 kyr (Fig. 4a) (Eggleston et al.,  
418 2016) to estimate the predicted  $\delta^{13}\text{C}_{\text{DIC}}$  at the ocean surface. We applied the thermodynamic  
419 isotopic-equilibrium equation published by Zhang et al. (1995), where  $\epsilon_{\text{DIC-g}} = (10.51 \pm 0.05)$   
420  $- (0.105 \pm 0.002) \times T$ . This equation was derived from direct measurement of the isotope  
421 fractionation between DIC of seawater and atmospheric CO<sub>2</sub> for a temperature and carbonate  
422 ion fraction range of 5 to 25°C and 0.05 to 0.2, respectively (Zhang et al., 1995). By using the  
423 SST record from GL-1090 (24°S), we assume that this region in the South Atlantic Subtropical  
424 Gyre is representative of where the ocean exchanges CO<sub>2</sub> with the atmosphere and that these  
425 surface waters will be eventually pumped to the South Atlantic thermocline. It is also noteworthy  
426 that this estimate only assumes the effect of thermodynamic equilibration on the  $\delta^{13}\text{C}_{\text{DIC}}$ .  
427 The predicted  $\delta^{13}\text{C}_{\text{DIC}}$  (Fig. 4d) strongly resembles the *G. inflata*  $\delta^{13}\text{C}$  record (Fig. 4c), showing  
428 similar millennial and long-term variations and with similar magnitude of  $\delta^{13}\text{C}$  drops to those  
429 presented by our planktonic foraminifera record. Therefore, we argue that the ocean-atmos-  
430 pheric thermodynamic equilibration can substantially modulate the thermocline  $\delta^{13}\text{C}$  in the  
431 western South Atlantic in orbital time-scales. Indeed, this result agrees with recent numerical  
432 modeling, which demonstrates that thermodynamic equilibration dominates the upper ocean  
433  $\delta^{13}\text{C}_{\text{DIC}}$  minimum anomaly during the last glacial termination (Shao et al., 2021). Previous mod-  
434 eling outputs also support the role of thermodynamic equilibration on the upper ocean  $\delta^{13}\text{C}_{\text{DIC}}$   
435 (Menviel et al., 2018; Schmittner and Lund, 2015).

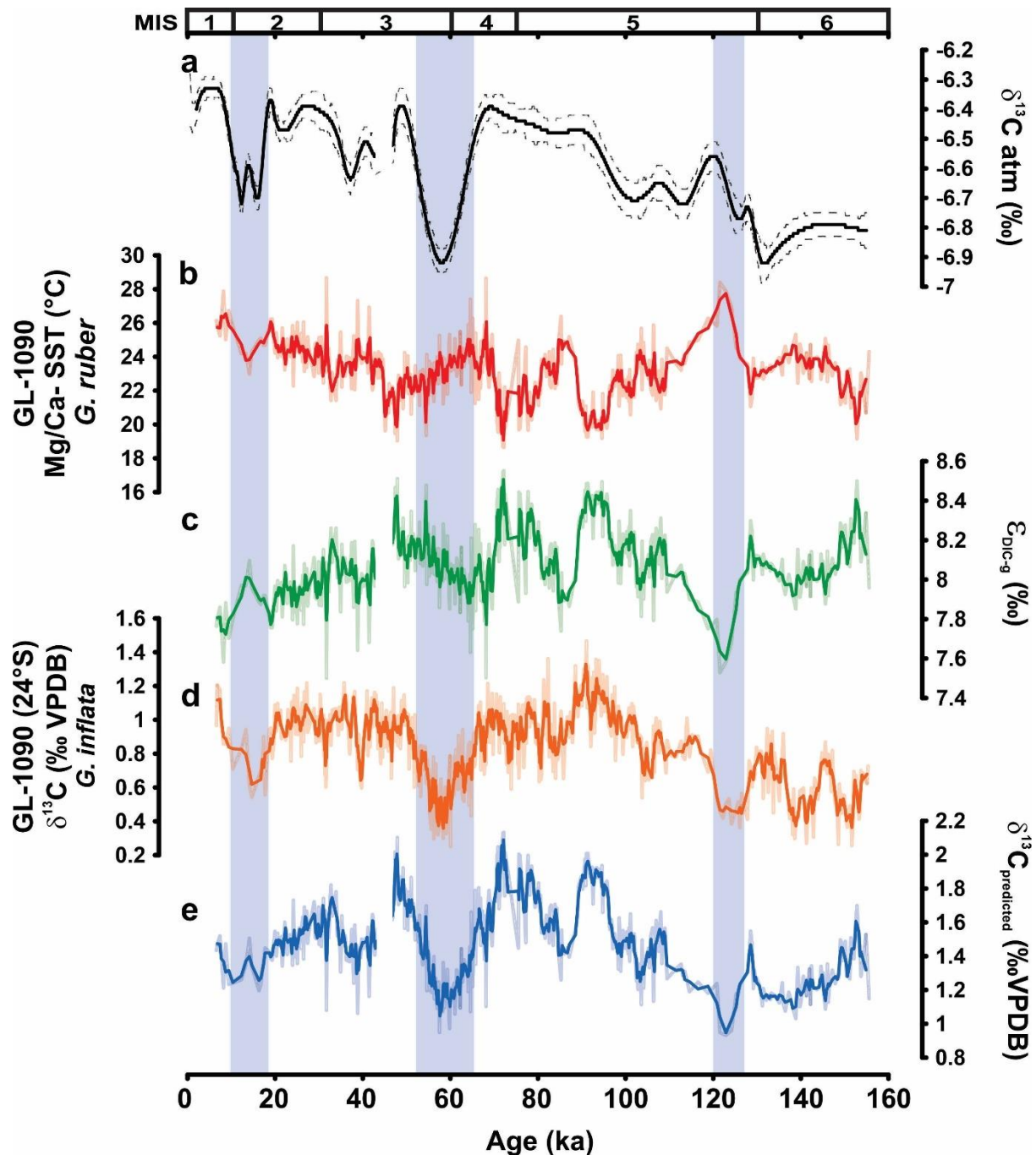
436 The average of our predicted  $\delta^{13}\text{C}_{\text{DIC}}$  record (1.48 ‰) is ~0.7 ‰ higher relative to the  
437 average of our *G. inflata*  $\delta^{13}\text{C}$  record (0.8 ‰). This result is expected since  $\delta^{13}\text{C}_{\text{DIC}}$  was pre-  
438 dicted for the ocean surface. Additionally, as mentioned in section 5.1., *G. inflata*  $\delta^{13}\text{C}$  shows  
439 a negative offset relative to  $\delta^{13}\text{C}_{\text{DIC}}$  (e.g., Wilke et al., 2006). Besides, the thermocline  $\delta^{13}\text{C}_{\text{DIC}}$   
440 in the western tropical and South Atlantic is also affected by the  $\delta^{13}\text{C}$  signal of the remineral-  
441 ized and preformed components of DIC (Oppo et al., 2018).

442 Both the predicted and the *G. inflata*  $\delta^{13}\text{C}$  records show a remarkable drop ( $\sim 0.5$  ‰)  
443 during the MIS 4/3 transition (Fig. 4c, d). This drop is similar to that found in  $\delta^{13}\text{C}$  of atmos-  
444 pheric  $\text{CO}_2$  ( $\sim 0.5$  ‰) (Eggleston et al., 2016). The MIS 4/3 transition can be considered an  
445 unfinished glacial termination following the maximum extension of Southern Hemisphere glac-  
446 iers at  $\sim 65$  ka (MIS 4) (Schaefer et al., 2015). Indeed, several lines of evidence point to the  
447 Southern Hemisphere under full glacial conditions during MIS 4 (Kohfeld and Chase, 2017;  
448 Schaefer et al., 2015; Barker and Diz, 2014). During MIS 4/3 transition, Antarctica showed a  
449 substantial warming, equivalent to the beginning of a full glacial termination (Wolff et al., 2009).  
450 In addition, the atmospheric  $\text{CO}_2$  concentration increased by  $\sim 30$  ppm (Bereiter et al. 2012).  
451 We argue that the climatic conditions around MIS 4/3 transition were akin to the beginning of  
452 a full termination. Consequently, the enhancement of the SO upwelling during this transition  
453 resulted in the transfer of  $^{13}\text{C}$ -depleted carbon from the deep ocean to the atmosphere, which  
454 ultimately imprinted its signal in the upper ocean.

455 A reduction of the biological pump efficiency may also have contributed to the negative  
456  $\delta^{13}\text{C}$  anomaly in the thermocline during glacial terminations, due to the accumulation of iso-  
457 topically light carbon in the upper ocean (Lund et al., 2019; Hertzberg et al., 2016; Schmittner  
458 and Lund, 2015). However,  $\delta^{13}\text{C}$  depletions are observed in regions where changes in the  
459 biological pump are not expected to have occurred, e.g., site GL-1180 ( $8^\circ\text{S}$ ), which is in the  
460 northern portion of the oligotrophic South Atlantic Subtropical Gyre. We also found thermocline  
461  $\delta^{13}\text{C}$  depletions in regions where the primary productivity is thought to have increased during  
462 the last termination, as indicated by Pereira et al. (2018) for the region of sediment core GL-  
463 1090 ( $24^\circ\text{S}$ ). However, a weaker biological pump at the Subantarctic Zone (SAZ), due to re-  
464 duced dust-borne iron supply during Termination I (Martínez-García et al., 2014; Jaccard et  
465 al., 2013), may have caused the advection of  $^{12}\text{C}$ -enriched waters to the western South Atlan-  
466 tic. These changes in the biological pump at the SAZ can be particularly relevant to the ther-  
467 mocline  $\delta^{13}\text{C}_{\text{DIC}}$  depletion in the domain of SAMW, but it would have minimal influence on the  
468 thermocline  $\delta^{13}\text{C}$  minima seen in the Northern Hemisphere (e.g., Lynch-Stieglitz et al., 2019).

469           Although the thermodynamic equilibration may have provided a substantial contribu-  
470 tion to the thermocline  $\delta^{13}\text{C}_{\text{DIC}}$  minimum during glacial terminations, the absence of a  $^{13}\text{C}$ -  
471 depleted signal at intermediate depth in the South Brazilian margin (27°S) (Tessin and Lund,  
472 2013; Lund et al., 2015) implies that the isotopic equilibration through southern-sourced wa-  
473 ters did not imprint the low  $\delta^{13}\text{C}_{\text{DIC}}$  signal at intermediate depth of the South Atlantic. Alterna-  
474 tively, the signal of the thermodynamic equilibration in AAIW may have been overcome by  
475 other processes. For example, a weaker biological pump at the SAZ may have reduced the  
476 export of organic matter and its remineralization at intermediate depth, thus causing the in-  
477 crease in the  $\delta^{13}\text{C}_{\text{DIC}}$  observed in the AAIW domain (Hertzberg et al., 2016). Such a hypothe-  
478 sized reduction of the biological pump at the SAZ agrees with the role of the regeneration of  
479 organic matter in modulating the  $\delta^{13}\text{C}_{\text{DIC}}$  signal at intermediate depth of the tropical and South  
480 Atlantic, as shown by Oppo et al. (2018).

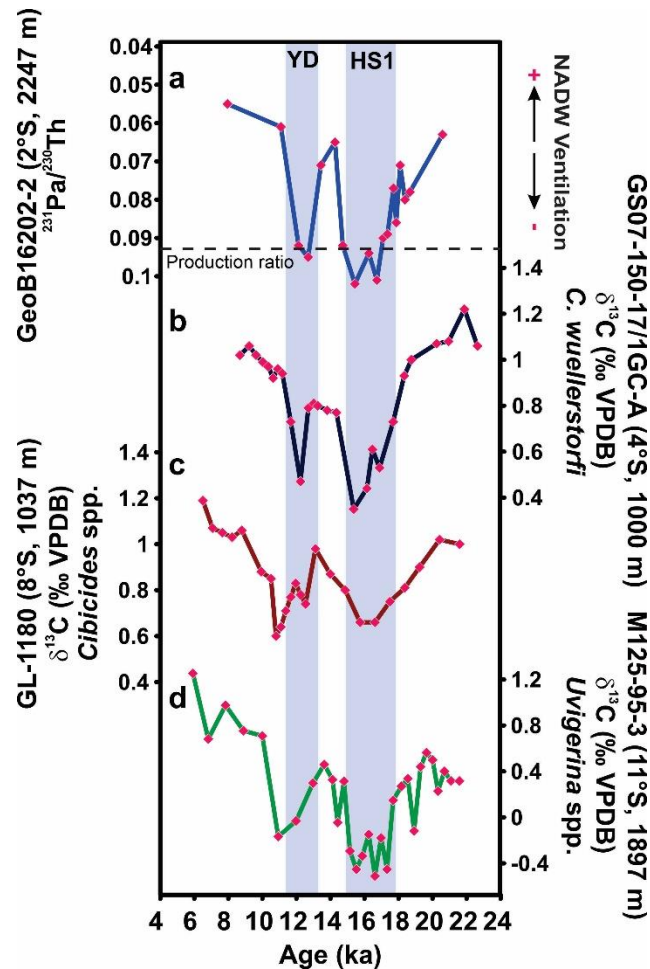
481           Therefore, in agreement with previous studies (e.g., Shao et al., 2021; Lynch-Stieglitz  
482 et al., 2019) we suggest that the atmosphere acted as a bridge during the early deglaciation,  
483 spreading the  $\delta^{13}\text{C}$  minimum signal globally in the upper ocean. Although thermodynamic equi-  
484 libration may have modulated the  $\delta^{13}\text{C}_{\text{DIC}}$  minimum events, changes in preformed and remin-  
485 eralized components of DIC can also affect  $\delta^{13}\text{C}_{\text{DIC}}$ , particularly in the domain of SAMW (Oppo  
486 et al., 2018). Finally, in the South Atlantic, the signal of thermodynamic equilibration was  
487 mostly seen in the thermocline so that an alternative mechanism and/or pathway is warranted  
488 to explain the  $\delta^{13}\text{C}$  minimum events observed in the benthic  $\delta^{13}\text{C}$  record from GL-1180 (1037  
489 m, 8° S).



490

491 **Figure 4.** Effect of temperature dependent ocean-atmosphere isotopic equilibration on the  
 492 stable carbon isotope ratio of the dissolved inorganic carbon ( $\delta^{13}\text{C}_{\text{DIC}}$ ) of ocean surface in site  
 493 GL-1090. a) Monte Carlo cubic spline average of atmospheric  $\delta^{13}\text{C}$  over the last 150 kyr (Eg-  
 494 gleston et al., 2016); b) Mg/Ca based sea surface temperature from sediment core GL-1090  
 495 (Santos et al. 2017), SST was recalculated using Mg/Ca-temperature equation from Gray and  
 496 Evans (2019). c) The  $^{13}\text{C}$  fractionation between ocean DIC and atmosphere CO<sub>2</sub> calculated  
 497 using the equation from Zhang et al. (1995). d)  $\delta^{13}\text{C}$  of *Globorotalia inflata* from site GL-1090

498 (this study). e) The predicted  $\delta^{13}\text{C}_{\text{DIC}}$  of ocean surface assuming the thermodynamic ocean-  
 499 atmosphere isotopic equilibration. Bars indicates times synchronic  $\delta^{13}\text{C}$  drops in the predicted  
 500  $\delta^{13}\text{C}_{\text{DIC}}$  and *G. inflata*  $\delta^{13}\text{C}$ . The thick lines in panels b, c, d, and e depict 3-points running  
 501 averages.



502  
 503 **Figure 5.** Comparison between the stable carbon isotopic composition ( $\delta^{13}\text{C}$ ) of benthic foram-  
 504 inifera from intermediate and mid-depth of the South Atlantic and radiochemical data from the  
 505 North Atlantic. a)  $^{231}\text{Pa}/^{230}\text{Th}$  record from sediment core GeoB16206-1 (Mulitza et al. 2017).  
 506 Note inverted y axis. b) Benthic  $\delta^{13}\text{C}$  from sediment core GS07-150-17/1GC-A (Freemann et  
 507 al., 2015). c) Benthic  $\delta^{13}\text{C}$  of sediment core GL-1180 (this study). d) Benthic  $\delta^{13}\text{C}$  of sediment  
 508 core M125-95-3 (Campos et al., 2020). Blue bars indicate Heinrich Stadial 1 (HS1) and the  
 509 Younger Dryas (YD).



510 5.2.3.  $\delta^{13}\text{C}$  minimum events at intermediate depth of the tropical Atlantic: the AMOC slowdown  
511 hypothesis

512 As discussed in sections 5.2.1 and 5.2.2, neither the “oceanic tunnel” nor thermody-  
513 namic equilibration through southern-sourced waters can explain the  $\delta^{13}\text{C}$  minimum events  
514 observed in our GL-1180 benthic  $\delta^{13}\text{C}$  record. Here we point to a North Atlantic origin of  $\delta^{13}\text{C}$   
515 minimum events at the intermediate depth of the tropical Atlantic. Over the last glacial termi-  
516 nation,  $^{13}\text{C}$ -depleted waters have been observed in the mid-depth of the North, tropical, and  
517 South Atlantic (Campos et al., 2020; Lacerra et al., 2017; Voigt et al., 2017; Oppo et al., 2015;  
518 Lund et al., 2015; Tessin and Lund, 2013). This depletion is likely linked to the reduced NADW  
519 ventilation rate during Heinrich Stadial 1 and the Younger Dryas, resulting in the accumulation  
520 of respired carbon at mid-depth of the Atlantic Ocean (Lacerra et al., 2017; Schmittner and  
521 Lund, 2015). Indeed, reduced NADW ventilation rate is supported by  $^{231}\text{Pa}/^{230}\text{Th}$  values (Fig.  
522 5a) from mid-depth of the Atlantic Ocean (e.g., Mulitza et al. 2017; Voigt et al. 2017). Accord-  
523 ingly, this mid-depth  $^{13}\text{C}$ -depleted signal could have affected intermediate depth (~1000 m)  
524  $\delta^{13}\text{C}_{\text{DIC}}$  signal at the Brazilian equatorial margin during Termination I, as observed in benthic  
525  $\delta^{13}\text{C}$  record (Fig. 5b) from site GS07-150-17/1GC-A (Fig. 1) (Freeman et al., 2015).

526 Numerical modeling supports a decrease of  $\delta^{13}\text{C}_{\text{DIC}}$  from the surface to mid-depth of  
527 the Atlantic Ocean during periods of AMOC slowdown, although this decrease is more expres-  
528 sive at intermediate and mid-depths of the tropical and North Atlantic (Menviel et al., 2018;  
529 Schmittner and Lund, 2015). These simulations agree with  $\delta^{13}\text{C}$  records of benthic foraminif-  
530 era (e.g., Campos et al., 2020; Voigt et al., 2017; Oppo et al., 2015). Therefore, we suggest  
531 that the  $\delta^{13}\text{C}$  depletion observed at intermediate depth of the western tropical South Atlantic  
532 (e.g., site GL-1180) was likely caused by the reduction of the North Atlantic overturning and  
533 accumulation of isotopically light carbon at intermediate and mid-depths. Indeed, over the last  
534 deglaciation, our benthic  $\delta^{13}\text{C}$  record (Fig. 5c) presents the same “W-like” shape observed in  
535 benthic  $\delta^{13}\text{C}$  record of core M125-95-3 (Fig. 5d), collected from the mid-depth of the tropical  
536 South Atlantic (Fig. 1) (Campos et al., 2020). The early  $\delta^{13}\text{C}$  decrease in the GL-1180 benthic  
537 record relative to other benthic records shown in Fig. 5 may result from the limited chronologic

538 control (6 radiocarbon ages) and the lower sampling resolution of GL-1180 over the early  
539 deglaciation. No substantial chronological difference is found by calibrating the radiocarbon  
540 ages of GL-1180 with the Marine20 calibration curve (Heaton et al. 2020) (Supporting Infor-  
541 mation, Text S2). Given the location and resemblance of GL-1180 benthic record relative to  
542 sediment cores M125-95-3 (Campos et al., 2020) and GS07-150-17/1GC-A (Freeman et al.,  
543 2015) (Fig. 1 and Fig. 5), it is reasonable to assume that over the Termination I, the bottom of  
544 the water column at these sites were influenced by a  $\delta^{13}\text{C}$  signal of similar origin. Together,  
545 these results support that the  $^{13}\text{C}$ -depletion at intermediate depth of the tropical Atlantic is  
546 mostly caused by the aged North Atlantic sourced waters. This interpretation can be applied  
547 to  $\delta^{13}\text{C}$  minimum events that occurred during previous glacial terminations, as seen in our  
548 benthic record (Fig. 2a). Importantly, this hypothesis is dissociated from the “ocean tunnel”  
549 and the ocean-atmosphere isotopic equilibration.

550         However, it has also been hypothesized that the thermodynamic equilibration could be  
551 equally responsible for the  $\delta^{13}\text{C}$  minimum observed at mid-depth of the Atlantic Ocean over  
552 the last termination by affecting the formation region of the NADW (Lynch-Stieglitz et al.,  
553 2019). However, a recent study showed that mid-depth benthic  $\delta^{13}\text{C}$  minimum leads the plank-  
554 tonic counterpart and atmospheric  $\delta^{13}\text{C}$  by ~800 years (Lund et al., 2019). If the ocean-atmos-  
555 phere isotopic equilibration was responsible for  $\delta^{13}\text{C}$  minimum at mid-depth of the Atlantic  
556 Ocean, one would expect the atmosphere to lead the ocean. Therefore, the lag of the atmos-  
557 phere relative to the ocean implies that the ocean-atmosphere isotopic equilibration through  
558 northern-sourced waters is unlikely to have been the main driver of the  $\delta^{13}\text{C}$  minimum event  
559 at intermediate and mid-depths of the Atlantic Ocean during Termination I.

560         In contrast to our reasoning of a northern-sourced  $\delta^{13}\text{C}$  minimum at intermediate depth,  
561 Poggemann et al. (2017) suggested an injection of nutrient-rich  $^{13}\text{C}$ -depleted AAIW at inter-  
562 mediate depth of the tropical North Atlantic during the last deglaciation (site M78/1-235-1; Fig.  
563 1). This conclusion was based on benthic  $\delta^{13}\text{C}$  and seawater-dissolved cadmium ( $\text{Cd}_w$ ), a  
564 proxy for phosphate concentration in seawater. However, as stressed above, the low  $\delta^{13}\text{C}$

565 signal contradicts depth transects from the southern Brazilian margin (Lund et al., 2015; Tes-  
566 sin and Lund, 2013; Oppo and Horowitz, 2000). While part of this  $Cd_w$  increase shown by  
567 Poggemann et al. (2017) may result from the remineralization of organic matter and nutrient  
568 accumulation due to the AMOC slowdown, another part must be decoupled from the  $\delta^{13}C_{DIC}$   
569 since records from the upper North Atlantic show that increases in  $Cd_w$  are not simultaneous  
570 with  $\delta^{13}C$  decreases (Lynch-Stieglitz et al., 2019). Indeed, a recent study showed that although  
571 the  $\delta^{13}C_{DIC}$  signal of upwelled deep waters is strongly affected by ocean-atmosphere isotopic  
572 equilibration in the SO surface, nutrients are still transported northward through the “oceanic  
573 tunnel” decoupled from  $\delta^{13}C_{DIC}$  signal (Shao et al., 2021). Therefore, this result can help to  
574 conciliate  $Cd_w$  and benthic  $\delta^{13}C$  records from intermediate depth of the tropical Atlantic.

## 575 **6. Conclusions**

576 We present three high temporal resolution thermocline and intermediate depth (1037  
577 m) foraminiferal  $\delta^{13}C$  records from the western South Atlantic spanning the last 300 kyr. The  
578 planktonic and benthic records show remarkable  $\delta^{13}C$  minimum events during the last three  
579 glacial terminations, as well as during the MIS 4/3 transition. Benthic  $\delta^{13}C$  depth transects  
580 from the south Brazilian margin, as well as intermediate depth  $\delta^{13}C$  records from the Pacific  
581 Ocean, contradict the long-held hypothesis of a northward advection of  $^{13}C$ -depleted carbon  
582 from the SO surface by SAMW and AAIW. Instead, we suggest that the  $\delta^{13}C$  minima at the  
583 thermocline and intermediate depths were governed by two distinct and independent mecha-  
584 nisms. The good correspondence between predicted  $\delta^{13}C_{DIC}$  and planktonic  $\delta^{13}C$  from the  
585 southern Brazilian margin indicates that the thermodynamic ocean-atmosphere isotopic equi-  
586 libration strongly modulated the  $\delta^{13}C_{DIC}$  signal in the western South Atlantic thermocline. This  
587 mechanism can explain the global occurrence of  $\delta^{13}C$  minimum events in the upper ocean  
588 and agrees with recent numerical modeling. Indeed, the thermodynamic equilibration is a lead-  
589 ing driver of the  $^{13}C$ -depleted signal in the western South Atlantic thermocline during glacial  
590 terminations and MIS 4/3 transition. However, in the domain of SAMW, changes in the biolog-  
591 ical pump in the SAZ must also be considered. We propose that the origin of the  $\delta^{13}C$  minimum

592 events at intermediate depths of the tropical Atlantic is related to the slow rate of NADW for-  
593 mation during glacial terminations and the accumulation of remineralized carbon at mid and  
594 intermediate depths of the tropical and North Atlantic. Importantly, the latter mechanism does  
595 not evoke the deep ocean and upwelling in the SO as a source of <sup>13</sup>C-depleted carbon to  
596 intermediate depths of the Atlantic Ocean.

#### 597 **Data availability**

598 All data presented in the manuscript are available at [LINK]

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