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Subsurface warming in the tropical Atlantic for the last 3 deglaciations: insights from organic molecular proxies

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Abstract

During glacial terminations, the role of cross-equatorial currents in the northward spreading of warm and salty waters is essential to the reach of the following interglacial state. In the western tropical Atlantic, where the interhemispheric North Brazil Current develops, most of the existing temperature records focused on the last deglaciation. Here, using a marine sediment core covering the last 305 kyr before present (BP), we present two records of the surface (mixed layer) and the subsurface (upper thermocline) temperatures based on organic proxies $U^{K'}_{37}$ and TEX^{H}_{86} , respectively. During the last three glacial terminations, these records exhibit large temperature increases especially in the subsurface layer. We suggest the combined influences of the thermal bipolar seesaw and the Agulhas Leakage, acting together to warm the interior South Atlantic, to explain the strong warming of the tropical western Atlantic thermocline (+ 5°C) during these periods of reduced Atlantic meridional overturning circulation (AMOC). In addition, we propose an enhanced cross-equatorial gyre to elucidate the synchronous deglacial warming of the western and eastern sides of the tropical Atlantic. The resumption of the AMOC at the onset of the following interglacial leads to the release of heat stored in thermocline waters mainly in the whole South Atlantic, as highlighted by the sharp upper ocean cooling recorded at the end of the deglacial interval. The combination of this temperature record and previously published ones in the western tropical Atlantic confirms the strong sensitivity of the upper ocean to AMOC changes in this region.

Keywords

Quaternary, Deglaciation, Paleoceanography, South Atlantic, Upper Ocean Temperature, Marine biomarkers, $U^{K'}_{37}$, TEX^{H}_{86} , Cross-equatorial gyre

1. Introduction

The Atlantic meridional overturning circulation (AMOC) plays a key role in the Earth's climate system (Buckley and Marshall, 2016; Trenberth and Fasullo, 2017). The AMOC experienced strong changes at both millennial and orbital scales over the late Pleistocene (Kageyama et al., 2010; Menviel et al., 2020; Paillard, 1998). Changes in AMOC intensity affected the spatial distribution of heat and water budgets (e.g. Castañeda et al., 2009; Wang et al., 2004), the atmospheric concentration

of carbon dioxide (e.g. Deaney et al., 2017; Ma et al., 2021) and the intensity of monsoon regimes over the globe (e.g. Rodríguez-Zorro et al., 2020; Simon et al., 2015). Future projections suggest an AMOC slowdown in response to the anthropogenic climate change (Liu et al., 2020). Therefore, it appears crucial to fully understand AMOC behaviour and associated climatic changes to elucidate, among others, the future evolution of Earth's system.

40 In the course of the late Pleistocene, millennial-scale events of massive icebergs discharges in the North Atlantic (Heinrich, 1988) were responsible of substantial AMOC weakening (McManus et al., 2004). The melting of these large ice blocks led to considerable freshwater inputs to the North Atlantic surface waters reducing the North Atlantic Deep Water production (Lynch-Stieglitz, 2017). As a result, the AMOC weakening induces a decrease of northward heat advection and thus a warming (cooling) in the South (North) Atlantic. Attempts to explain this opposite climatic behaviour between the North and the South
45 Atlantic rely on the bipolar seesaw thermodynamic model (Broecker et al., 1990; Crowley, 1992; Stocker and Johnsen, 2003). These "Heinrich events" occurred during some of the cold phase of the Dansgaard-Oeschger cycles, i.e. Greenland stadials (Bond et al., 1992; Rasmussen et al., 2014; Stocker, 2000) and deglaciations (Denton et al., 2010; Knorr and Lohmann, 2007). Glacial terminations are also associated with a poleward shift of the southern subtropical front, leading to an increase in the Agulhas Leakage (Bard and Rickaby, 2009). Indeed, this sudden deglacial outflow of saline and warm waters originating from
50 the Indo-Pacific oceans and propagating over the South Atlantic through the Cape basin is proposed to play a key role in the triggering of interglacial periods (Caley et al., 2011; Durgadoo et al., 2013; Peeters et al., 2004). As a consequence, the South Atlantic turns into a large heat reservoir during deglacial periods and abrupt climatic changes, with major reorganizations of the Earth's climate system such as a southward migration of the intertropical convergence zone (ITCZ) over the tropical South Atlantic (McGee et al., 2014; Portilho-Ramos et al., 2017).

55 The western tropical Atlantic is the oceanic gateway between the North and the South Atlantic where the interhemispheric North Brazil Current (NBC) forms (Rodrigues et al., 2007). This region experienced strong changes in response to Heinrich events and associated decrease in AMOC intensity. The reduced cross-equatorial transport by the North Brazil Current (Vink et al., 2001) led to an accumulation of warm waters in the western tropical Atlantic, as generally exhibited by surface temperature records (Crivellari et al., 2019; Nace et al., 2014; Weldeab et al., 2006). However, some temperature proxies
60 disagree with such warming trend, probably due to non-thermal influences on these proxies (e.g. Jaeschke et al., 2007). This

general warming of the tropical and South Atlantic then caused a southward migration of the ITCZ (Portilho-Ramos et al., 2017), a weakening of southeasterly (SE) trades (Kim and Schneider, 2003) and a shoaling of the thermocline (Venancio et al., 2018). Moreover, in freshwater hosing experiments, the modelled tropical surface and thermocline layers exhibit large temperature increases and circulation changes (Chang et al., 2008; Dahl et al., 2005; Laurian and Drijfhout, 2011).

65 Paleotemperature proxies based on lipid biomarkers, namely long-chain C_{37} -alkenones and isoprenoid glycerol dialkyl glycerol tetraethers (iGDGTs), are among the most largely used to reconstruct past upper ocean temperatures. The distribution of long-chain C_{37} -alkenones, produced by *Prymnesiophyceae* algae, has been related to seawater temperature in both culture (Prahl and Wakeham, 1987), marine particles (Conte et al., 2006) and sediment studies (Müller et al., 1998; Tierney and Tingley, 2018) through the $U^{K'}_{37}$ index. *Prymnesiophyceae* can thrive in different depths into the euphotic zone. In the
70 oligotrophic (nutrient-poor) South Atlantic waters, they mostly develop in the lower euphotic zone (Cordeiro et al., 2013), where they find both light and nutrients (the deep chlorophyll maximum, Latasa et al., 2017) supplied by underlying waters (Benthien and Müller, 2000; Gérikas Ribeiro et al., 2016; Latasa et al., 2017). In the western tropical Atlantic, alkenones producers have been reported to thrive mainly above the thermocline, between 50 and 100 m depth (Dauner et al., 2019; Poulton et al., 2017) into the mixed layer. The C_{37} -alkenones $U^{K'}_{37}$ index can be used as a temperature proxy of this layer (later
75 referred as surface, Figure 1).

Isoprenoid glycerol dialkyl glycerol tetraethers (iGDGTs) are produced by marine *Thaumarchaeota*, and their distribution, reflected in the TEX_{86} index, is strongly correlated with seawater temperature (Kim et al., 2010; Schouten et al., 2002; Tierney and Tingley, 2014). These chemo-autotroph ammonia-oxidizers archaea may develop within the whole water column (Ceccopieri et al., 2018; Zhang and Liu, 2018) but are prominent below the euphotic zone (Hurley et al., 2018), close to the
80 NO_2^- maximum. Depending on local settings, this maximum is observed in the mixed layer (surface; Dauner et al., 2019) or below, in the upper thermocline (subsurface; Zakem et al., 2018). In oligotrophic and highly stratified waters like the western tropical Atlantic, *Thaumarchaeota* develop in the upper thermocline because of a shallow thermocline and competition with primary producers (Inglis and Tierney, 2020; Schouten et al., 2013; Smith et al., 2014). This implies that the TEX_{86} can be used as a proxy of the upper thermocline temperature (hereafter referred as subsurface, Figure 1). The complementarity of the
85 $U^{K'}_{37}$ and TEX_{86} temperature proxies was shown by previous studies carried out in the eastern tropical Atlantic (Lopes dos

Santos et al., 2010), in the subtropical South Atlantic (Dauner et al., 2019), in the South China Sea (Li et al., 2013) and in the gulf of California (McClymont et al., 2012).

The sediment cores investigated so far in the western tropical Atlantic cover at most the last 200 kyr BP, mainly focusing on the last glacial cycle (Behling et al., 2000; Dauner et al., 2019) and the last deglaciation (Chiessi et al., 2015; Meier et al., 90 2021; Santos et al., 2022). In this study, we reconstructed the variations of upper ocean temperature using a marine sediment core collected in the western tropical Atlantic and covering the last 305 kyr BP. The surface and subsurface temperature records presented here, which are based on the $U^{K'}_{37}$ and TEX^H_{86} organic proxies, respectively, cover, for the first time in this area, the last three glacial terminations, with a higher temporal resolution during interglacial and deglacial times. We aimed to examine the response of the upper ocean temperature (surface and subsurface) in conjunction with changes in ocean stratification and 95 atmospheric reorganizations, especially during deglacial AMOC slowdowns. We investigated the role of both the bipolar seesaw and Agulhas Leakage outflow to explain the magnitude of the strong upper ocean warmings reconstructed during glacial terminations in the western tropical Atlantic.

2. Material and Methods

2.1 Regional oceanographic settings

100 South Atlantic surface waters originate from both Indian, Pacific and Southern Oceans (Mignac et al., 2018). The subtropical gyre is the main circulation feature in the South Atlantic (Figure 1). This anticyclonic gyre flows, on average at the surface, between $\sim 10^\circ\text{S}$ and $\sim 20^\circ\text{S}$ (NW boundary) and 40°S (southern boundary) (Schettini et al., 2017; Stramma and England, 1999). In the first 800 m depth, the South Atlantic Central Waters (SACW) are the dominant water mass and they are fed by exchanges with the Indian, Pacific and Southern Oceans. Southern Ocean waters enter the subtropical gyre through 105 the Malvinas Current ($\sim 200 - 400$ m depth, SW corner) to supply the South Atlantic Current (SAC) flowing eastward (Figure 1). Around the Cape Basin, the South Atlantic Current meets the Indo-Pacific waters spreading through the Agulhas Leakage (AL) rings and eddies ($\sim 100 - 600$ m depth) (Souza et al., 2018). Agulhas waters are an important supply of heat and salt to the South Atlantic (Beal et al., 2011). These waters mix with the South Atlantic Current to form the South Equatorial Current (SEC) flowing across the South Atlantic in a northwestward direction (Figure 1). The southern branch of the South Equatorial

110 Current (sSEC) bifurcates at the northwestern boundary and forms the Brazil Current and the North Brazil Undercurrent (NBUC) flowing southward and northward respectively (Rodrigues et al., 2007). The NBUC is a subsurface current along the Northeast Brazil that mixes with the South Equatorial Current to form the North Brazil Current (NBC; Goes et al., 2005; Krelling et al., 2020; Schott et al., 1998). This North Brazil Current is the main surface interhemispheric pathway linking the South and North Atlantic basins.

115 The core studied here is located at the northernmost position of the bifurcation of the southern branch of the South Equatorial Current, around 10°S (Rodrigues et al., 2007). At this location, surface waters are predominantly tropical waters (high salinity and temperature) with minor contributions (3 – 5%) of fresh and relatively cold coastal waters of riverine origin (Schettini et al., 2017). Mean annual sea surface temperature (SST) and salinity (SSS) at the core site are 27.7°C and 36.3 psu respectively (Locarnini et al., 2013). The mixed layer is 70 – 90 m deep and overlies on a thermocline that extends up to 300
120 m (Figure 1). Seasonal variability of temperature and stratification of the upper layer is very low at this location. The temperature of the mixed layer oscillates between 26.3°C and 28.1°C and its width varies between 70 and 120 m, while the thermocline depth and temperature are overall constant over the year (80 – 120 m depth for the upper thermocline with an average temperature of 25°C).

2.2 Sediment core and age model

125 The marine sediment core GL-1180 (8°27'18"S, 33°32'53"W, 1 037 m water depth) was collected on the Northeastern Brazilian margin, in the western tropical Atlantic ocean by the Petrobras oil company as described by Nascimento et al. (2021b) and stored at low temperature (4°C). 135 depths were subsampled corresponding to a resolution of 1.5 kyr and 7.1 kyr for deglacial/interglacial times and glacial periods, respectively. After sampling, sediments were freeze-dried and then stored at room temperature prior to preparation for lipid biomarker analyses.

130 We use the published age model of GL-1180 core, which is detailed in Nascimento et al. (2021b). Briefly, 6 depths were selected for radiocarbon (¹⁴C) dating on *Globigerinoides ruber* and *Trilobatus sacculifer* shells by AMS (Accelerator Mass Spectrometry). The age model is also based on the visual alignment between the benthic foraminifera (*Cibicidoides sp.*) $\delta^{18}\text{O}$ and the global reference stack LR04 (Lisiecki and Raymo, 2005). It was then constructed using Bacon software v2.3 (Blaauw and Christen, 2011). Records of stable oxygen ($\delta^{18}\text{O}$) and carbon ($\delta^{13}\text{C}$) isotopic composition measured on the thermocline

135 (*Globorotalia truncatulinoides*) and benthic (*Cibicides spp.*) foraminifera species are available along the GL-1180 core (Nascimento et al., 2021a).

2.3 Sample preparation

For each sample of core GL-1180 taken for lipid biomarker analyses, between 5 and 10 g of dry sediment (corresponding to 1 cm of sediment), were extracted using a mixture of dichloromethane/methanol ($\text{CH}_2\text{Cl}_2/\text{CH}_3\text{OH}$), 9/1 (v/v) in an ultrasonic bath. The extraction consisted in five ultrasonication cycles (10 min each) followed by centrifugation (5 min \times 3 000 rpm) using 20 mL, 15 mL and 10 mL (for the three last cycles only) of solvent mixture. The supernatants were combined, concentrated to dryness using rotary evaporator and dissolved into 8 mL of $\text{CH}_2\text{Cl}_2/\text{CH}_3\text{OH}$, 9/1 (v/v).

Separation of the total lipid extract into three fractions of increasing polarity was performed over activated (150°C , ≥ 2 h) and extracted (CH_2Cl_2 followed by *n*-heptane) silica (0.8 g of SiO_2 per column). The first fraction (apolar fraction) was eluted with 5.5 mL of *n*-heptane, the second fraction containing alkenones with 8 mL of a mixture of heptane/ethyl acetate, 92.5/7.5 (v/v) and finally the third fraction containing GDGT with 8 mL of $\text{CH}_2\text{Cl}_2/\text{CH}_3\text{OH}$, 1/1 (v/v). The fractions were concentrated at 35°C to dryness under gentle N_2 flux and dissolved into 2 mL of solvent corresponding to the mixture used for the elution of the respective fraction.

2.4 Lipid biomarker analyses

Alkenone analyses were performed using a gas chromatograph coupled with a flame ionization detector (GC-FID) HP 6890 Series plus and equipped with an autosampler. Separation was carried out using an Agilent CP-Sil 5 CB column (50 m \times 0.32 mm, 0.25 μm inner diameter film thickness). Injections were performed with 5- α -cholestane (4 ppm final concentration) as injection standard in on-column mode. Helium was used as carrier gas at a constant flow of 2 mL/min. The GC oven program was: 50°C to 140°C at $30^\circ\text{C}/\text{min}$, then to 280°C at $20^\circ\text{C}/\text{min}$ and finally to 305°C at $0.5^\circ\text{C}/\text{min}$. Individual alkenones were identified based on their retention times. For 14 samples (0 – 45, 75 – 105, 135, 150, 560 and 660 cm), alkenones were below the detection limit.

GDGTs were analysed by a high-pressure liquid chromatography coupled with mass spectrometry with an atmospheric pressure chemical ionization source (HPLC–APCI–MS) using a Shimadzu LCMS 2020 in selected ion monitoring mode, with

a procedure slightly modified from Hopmans et al. (2016) and Huguet et al. (2019). Briefly, GDGT analysis was performed using two Acquity UHPLC BEH HILIC columns in tandem (150 mm × 2.1 mm, 1.7 μm; Waters, USA), thermally controlled at 40°C. Injection volume was 30 μL. The flow rate was set at 0.2 mL/min. GDGTs were eluted isocratically for 25 min with 82% A/18% B, followed by a linear gradient to 65% A/35% B in 25 min, then a linear gradient to 100% B in 30 min, and back to 82% A/18% B in 4 min, maintained for 50 min (A= hexane, B=hexane/isopropanol 9/1, v/v). Semi-quantification of iGDGTs and brGDGTs was performed by comparing the integrated signal of the respective compound with the signal of a C₄₆ synthesised internal standard (Huguet et al., 2006) assuming their response factors to be identical.

2.5 Lipid-based temperature proxies

2.5.1 Core tops used for the selection of temperature proxy calibrations

We used published lipid-based data on core tops (0–1 cm) close to the location of core GL-1180 (–8.5°S, –34°W) to determine the most accurate calibration (i.e. exhibiting the lowest residual error) to convert the U^{K'}₃₇ and TEX₈₆ values into temperatures. U^{K'}₃₇ and TEX₈₆ values from these core tops were obtained from dataset published in Benthien and Müller (2000) and Kim et al. (2008), respectively. We tested different U^{K'}₃₇ and TEX₈₆ calibrations and compared the calculated temperature estimates to modern *in situ* sea temperatures values from the World Ocean Atlas 13 (WOA 13; Locarnini et al., 2013). We used both mean annual and seasonal temperatures over the period 1955-2014 for the comparison. When needed, spatial interpolation with DIVA gridding was performed using the Ocean Data View software (Schlitzer, 2020) to better fit with the location of core top temperature data.

2.5.2 The U^{K'}₃₇ index as a mixed layer temperature proxy

Past sea surface (mixed layer) temperatures in core GL-1180 were reconstructed using the long-chain alkenones (C₃₇ di- or tri-unsaturated methyl-ketones) through calculation of the U^{K'}₃₇ index (Eq. 1; Prahl and Wakeham, 1987):

$$U_{37}^{K'} = \frac{[C_{37:2}]}{[C_{37:2}] + [C_{37:3}]} \quad (eq. 1)$$

where [C_{37:x}] refers to the fractional abundance of C₃₇ alkenone with x insaturations

The main environmental variable influencing the $U^{K'}_{37}$ index is seawater temperature (Prahl and Wakeham, 1987). Nevertheless, non-thermal factors (i.e. lateral transport, seasonality and nutrient availability) can impact the $U^{K'}_{37}$ signal and all these processes have to be constrained before application of the index as a temperature proxy. They were examined close to the GL-1180 location by Benthien and Müller (2000), who highlighted the main influence of seawater temperature on the $U^{K'}_{37}$ signal in the western tropical Atlantic.

Various global and regional calibrations were developed to convert the $U^{K'}_{37}$ values into sea surface temperature (SST) estimates (Conte et al., 2006; Tierney and Tingley, 2015 and references therein). Based on the core top data in our study area, we found that the global calibration (60°N-60°S, core top calibration, mean annual SST at 0 m depth, Müller et al., 1998; eq. 2) yields the most accurate reconstruction of modern SST (mean residual error of 0.3°C), even though all tested calibrations provided realistic values (residual errors < 1°C; see Supplementary Table S1 and Figure S2).

$$SST = \frac{(U^{K'}_{37} - 0.044)}{0.033} \quad (R^2 = 0.96, n = 370, Error: \pm 1^\circ C)(eq. 2)$$

In the oligotrophic waters of the western tropical Atlantic, alkenones producers thrive preferentially in the lower mixed layer (see 1. Introduction). As the main export of alkenones is reported within the mixed layer (Dauner et al., 2019), we assume that the $U^{K'}_{37}$ is a surface proxy reconstructing the temperature of this layer (surface, $U^{K'}_{37}$ -SST), as generally observed in the study area (Crivellari et al., 2019; Müller and Fischer, 2004; Sikes et al., 1991). The calibration standard error ($\pm 1\sigma$; Müller et al., 1998) and total analytical error were respectively of 1°C (0.03 $U^{K'}_{37}$ unit) and 0.31°C (0.01 $U^{K'}_{37}$ unit). The analytical error was evaluated on the replicated injection of 10 different samples.

2.5.3 The TEX_{86} index as an upper thermocline temperature proxy

Past temperature of the subsurface (upper thermocline) was reconstructed based on the TEX_{86} index, calculated as follows (Schouten et al., 2002):

$$TEX_{86} = \frac{[GDGT - 2] + [GDGT - 3] + [Cren']}{[GDGT - 1] + [GDGT - 2] + [GDGT - 3] + [Cren']} \quad (eq. 3)$$

where $[GDGT-X]$ refers to the fractional abundance (relative to total iGDGTs) of compound GDGT-X and Cren' is the regioisomer of crenarchaeol. See Figure 2 for respective molecular structures.

205 Other factors than temperature were reported to have an influence on the distribution of iGDGTs. Thus, even though the
TEX₈₆ index is based on the distribution of iGDGTs produced by marine *Thaumarchaeota*, these lipids are synthesised by all
archaea (Inglis and Tierney, 2020), notably terrestrial *Thaumarchaeota* (Hopmans et al., 2004), methanogenic (Inglis et al.,
2015) and methanotrophic archaea (Zhang et al., 2011). In order to constrain the effect of non-marine *Thaumarchaeota*-derived
iGDGTs on the TEX₈₆ signal, we calculated the diagnostic ratios previously developed (Weijers et al., 2014; Zhang et al.,
210 2016, 2011). The values obtained (Supplementary Figure S1, Table S2) confirmed that in our record, marine *Thaumarchaeota*
are the main iGDGT producers and that seawater temperature is the main factor influencing the iGDGT distribution.

Among all developed surface and subsurface calibrations based on the TEX₈₆ index, the calibration based on the logarithmic
transformation of the TEX₈₆ i.e. TEX^H₈₆ was specifically designed for “warm environments” i.e. SST > 15°C (Kim et al., 2010)
and the Bayesian calibration developed by (Tierney and Tingley, 2014) is the most up to date.

215 Using the same approach as for U^K₃₇ temperature calibration selection, we selected the calibration based on the TEX^H₈₆
index (eq. 4; Kim et al., 2010) as this calibration provides the most accurate (residual error of -1.5°C on average; see
Supplementary Table S3 and Figure S2) SST estimates for our study area.

$$TEX_{86}^H = \log(TEX_{86}) \quad ; \quad SST = 68.4 * TEX_{86}^H + 38.6 \quad (R^2 = 0.87, n = 255, \text{Error: } \pm 2.5^\circ\text{C}, \pm 1\sigma) \quad (\text{eq. 4})$$

220 Since in the south western tropical Atlantic the *Thaumarchaeota* preferentially develop below the mixed layer and within
the upper thermocline (see 1. Introduction) we use the TEX^H₈₆ as a subsurface i.e. upper thermocline temperature proxy
(hereafter mentioned as TEX^H₈₆-SubT). The total analytical error is of 0.01 TEX₈₆ unit (0.8°C for TEX^H₈₆ temperature) based
on duplicate injections of 3 different samples.

3. Results

U^K₃₇-derived sea surface temperatures varied between 26.1 and 28.9°C (Figure 3). They show a typical glacial-interglacial
225 pattern with high values (28.1°C on average) found during the 230 – 185 and 132 – 70 kyr BP intervals, corresponding to MIS
7 (penultimate interglacial) and MIS 5 (last interglacial), respectively. The sharp increase of SST before these intervals of high
SST correspond to the deglacial periods of Terminations II and III (TII and TIII), beginning at 133 and 250 kyr BP,

respectively. Even though the record is not continuous over the last 30 kyr, the SST increased starting from around 20 kyr BP and remained high ($\sim 28.5^{\circ}\text{C}$) during the Holocene (Figure 3).

230 The $\text{TEX}_{86}^{\text{H}}$ upper thermocline temperature values ranged between 23.5°C and 29.4°C (Figure 3). The upper thermocline temperature values reconstructed with $\text{TEX}_{86}^{\text{H}}$ are generally lower (1.4°C on average) and show a larger range of variations than the $\text{U}^{\text{K}'}_{37}\text{-SST}$ record. $\text{TEX}_{86}^{\text{H}}\text{-SubT}$ also shows a glacial-interglacial pattern, with high temperature values during MIS 7, MIS 5 and the Holocene. However, deglaciation intervals exhibit the highest temperature increases in the whole $\text{TEX}_{86}^{\text{H}}\text{-SubT}$ and $\text{U}^{\text{K}'}_{37}\text{-SST}$ records.

235 During Termination III, $\text{U}^{\text{K}'}_{37}\text{-SST}$ and $\text{TEX}_{86}^{\text{H}}\text{-SubT}$ records increased to reach a maximum around 245 kyr BP and started to decrease at the end of the deglacial interval (Figure 3). A similar temperature evolution is observed in both $\text{U}^{\text{K}'}_{37}\text{-SST}$ and $\text{TEX}_{86}^{\text{H}}\text{-SubT}$ records but the upper thermocline temperature only decreases before the onset of the following interglacial. Termination I shows a decrease in the $\text{TEX}_{86}^{\text{H}}\text{-SubT}$ record around 14 kyr BP corresponding to the Bølling-Allerød period, while the Younger Dryas and Heinrich Event 1 are associated with high $\text{TEX}_{86}^{\text{H}}\text{-SubT}$ values (Figure 3). Besides, $\text{U}^{\text{K}'}_{37}\text{-SST}$ and $\text{TEX}_{86}^{\text{H}}\text{-SubT}$ increased at the onset of Termination II and remained high over the whole interval. The temperature decrease of the $\text{TEX}_{86}^{\text{H}}\text{-SubT}$ record preceded the one of $\text{U}^{\text{K}'}_{37}\text{-SST}$ by approximately three thousand years.

245 Furthermore, the $\text{TEX}_{86}^{\text{H}}\text{-SubT}$ (but not the $\text{U}^{\text{K}'}_{37}\text{-SST}$) record exhibited strong increases during the transitions between the substages of the MIS 7 (MIS7a-7b and MIS 7c-7d). This temperature pattern was not observed during the MIS 5 and is specific to MIS 7 (Figure 3). Finally, at the onset of interglacial conditions, the $\text{TEX}_{86}^{\text{H}}\text{-SubT}$ records starts to decrease which contrasts to the relatively stable $\text{U}^{\text{K}'}_{37}\text{-SST}$ evolution (for TI and TII) during optimum interglacial periods. This feature is particularly visible during the Last Interglacial (Figure 3).

4. Discussion

4.1 Upper Ocean warmings during glacial terminations of the last 305 kyr on the western boundary of the South Atlantic

250 Based on the GL-1180 core, the western tropical Atlantic experienced a warming in its upper limb over the last three glacial terminations (Figures 3 and 4). The warming of the ocean upper limb is a well described feature for the last two terminations as reported by previous studies along the Brazilian margin (Figure 4). During Termination 1, the upper thermocline experienced

temperature increase (Heinrich Event 1, H1, Younger Dryas, YD) and decrease (Bølling-Allerød, B/A) related to changes in upper layer heat transport. The H1 and YD events are associated with strong AMOC reduction (Lynch-Stieglitz, 2017) leading to a decrease in the cross-equatorial heat transport (Laurian and Drijfhout, 2011) and warm water accumulation in the South Atlantic (Arz et al., 1999; Chiessi et al., 2008; Santos et al., 2017). The deglacial seawater warming is common to the whole western South Atlantic, as exhibited by a stack of 17 surface temperature records (Santos et al., 2022, Figure 4C) from the Brazilian margins. However, in our records (Figure 3), the amplitude of the warming is lower in surface (+1.5°C) compared to upper thermocline (+4°C) waters, indicating warm water accumulation at deeper levels, as observed in freshwater hosing experiments (e.g. Rühlemann et al., 2004).

Over Termination II, the GL-1180 temperature records present similarities with those of the GL-1090 core (Brazilian margin, ~25°S) (see Figure 1 for core locations) in respect to: (i) the occurrence of temperature increase, (ii) the larger amplitude of the warming at the subsurface (upper and main thermocline) compared to the surface and (iii) the surface and subsurface decrease at the onset of the last interglacial (Figure 4A, B). In the GL-1090 temperature records, the vertical temperature gradient is stronger for organic proxies (U^{K}_{37} and TEX^{H}_{86}) than for foraminifera-based (*G. ruber*, *G. inflata*) Mg/Ca records. At this location, these differences can be explained by distinct habitat depths between *G. ruber* and the Thaumarchaeota (mixed layer), the haptophytes (upper thermocline) and *G. inflata* (main thermocline). The habitat depth is actually controlled by nutrient and lights supplies, themselves constrained by the structure of the water column (Hurley et al., 2018; Latasa et al., 2017). Such differences between organic and inorganic proxies have already been highlighted and attributed to distinct habitat depths, seasonality and transport to the sediment (Dauner et al., 2019; Richey et al., 2011). In addition, temporal changes in the structure of the upper water column could also lead to diverging temperature estimates from organisms dwelling at distinct depths.

In contrast, the GL-1180 temperature record of the main thermocline (*G. truncatulinoides*) differs clearly from the surface and upper thermocline records (Figure 3). First, the temperature estimates based on the TEX^{H}_{86} index are higher than those based on the Mg/Ca ratio (*G. truncatulinoides*). This is in agreement with a shallower signal recorded by Thaumarchaeota (upper thermocline) compared to *G. truncatulinoides* (main thermocline). Secondly, there is no deglacial warming indicated by the main thermocline temperature record. In contrast, a precessional pace have been identified in the stratification proxy

$\Delta\delta^{18}\text{O}$ record (between *G. ruber* and *G. truncatulinoides*) whose variations have been attributed mainly to main thermocline temperature changes (Nascimento et al., 2021b). These differences between the temperature of the upper (TEX^H₈₆-SubT) and
280 main thermocline (Mg/Ca-ThT) are attributed to different mechanisms controlling subsurface variations within distinct layers of the thermocline. The South Atlantic Central Waters are cold and nutrient rich intermediate waters underlying the warm tropical waters as observed in local vertical nutrient profiles (Supplementary materials, Figure S4). The high nutrient concentrations in the lower half of the thermocline while tropical surface waters are nutrient-poor indicate the vertical incursions of SACW up to the main thermocline. Such SACW incursions into the thermocline have been reported along
285 Brazilian margins, in response to ocean-atmosphere dynamics, mainly associated with wind and coastal upwelling on the west margin (Gérikas Ribeiro et al., 2016; Lessa et al., 2016). We suggest here that in contrast to *G. truncatulinoides*, the TEX^H₈₆ records the temperature of the upper thermocline which is not under the influence of SACW.

The observed deglacial warming are common features attributed, as for the Termination I, to the accumulation of warm waters in the South Atlantic during times of weakened AMOC. A similar thermal evolution is observed here within GL-1180
290 temperature records during the Termination III, indicating a recurring feature in this region for at least the last 3 deglaciations. In addition, the GL-1180 upper thermocline temperature records exhibit significant increases during the transition from “cold” (MIS 7d and 7b) to “warm” (MIS 7c and 7a) sub-stages of the MIS 7 but with smaller amplitude of changes compared to the deglaciations (Figure 3). Regarding to the previous assumptions, these “sub-deglaciations” are smaller analogous to the deglaciations, driven by similar mechanisms.

295 Two principal climatic processes can explain the deglacial temperature changes observed within the western South Atlantic: (i) the thermodynamic model of the bipolar seesaw (Broecker et al., 1990) which was proposed to explain the opposite thermal behaviour between the (cooling) North and the (warming) South Atlantic during periods of reduced AMOC (see 1. Introduction; Pedro et al., 2018) and (ii) the increase in Agulhas Leakage outflow during deglaciations of the Late Pleistocene (Peeters et al., 2004), suggested to be responsible of large inputs of saline and warm waters to the South Atlantic (Scussolini
300 et al., 2015). The upper ocean warming recorded by the GL-1180 temperature record during the past 3 glacial terminations, at the extreme north-western boundary of the South Atlantic, could be explained by one (bipolar seesaw) or the other (Agulhas

Leakage) hypothesis, or both of them. We investigated subsequently which of these hypotheses is the most likely to explain the water temperature variations observed in the western South Atlantic.

4.2 Bipolar seesaw influence on millennial-scale changes in upper western tropical Atlantic temperature

305 As suggested above, the upper ocean warming observed in core GL-1180 during the last three glacial terminations could be due (at least partly) to the thermal bipolar seesaw concept, an interhemispheric coupling in response to millennial-scale abrupt climatic events related to AMOC changes (Broecker et al., 1990; Crowley, 1992; Stocker and Johnsen, 2003). To fully assess the impact of millennial-scale abrupt events on the upper western Atlantic temperature, we examined the variations in GL-1180 temperature record over millennial-scale climatic events (Figure 5). We focus here on the upper thermocline record
310 (TEX^H₈₆-SubT), which exhibits larger variations compared to the U^K₃₇-SST, and investigate the period between ~70 and 125 kyr BP, which has a higher temporal resolution than the last glacial period. Within our age model uncertainties (Nascimento et al., 2021b), a subsurface warming (+ 2.5°C on average) is observed at 74, 84 and 107 kyr BP corresponding to Greenland Stadials (GS), i.e. cold phases of DO cycles (Rasmussen et al., 2014) 20, 22 and 25, respectively (Figure 5E). During these periods of cold North Atlantic (Martrat et al., 2007) and weakened AMOC (McManus et al., 2004; Piotrowski et al., 2005),
315 the western tropical Atlantic experienced water mass ventilation (Waelbroeck et al., 2018), stratification (Venancio et al., 2018) and upper ocean temperature changes (Jaeschke et al., 2007; Portilho-Ramos et al., 2017; Rama-Corredor et al., 2015) (Figures 5B, C, D). Within the main thermocline, the temperature record (Mg/Ca-ThT) of GL-1090 core (Santos et al., 2020) exhibits a seawater warming during GS19, GS20 and GS21 (Figure 5F). In addition, the Mg/Ca-ThT record of GL-1180 core (Nascimento et al., 2021b) also shows a temperature increase during GS20 (Figure 5G). Both warmings were induced by the
320 reduced meridional heat transport and the accumulation of warm waters within the South Atlantic basin during such periods.

Based on the bipolar seesaw model (Stocker and Johnsen, 2003), the extension of sea-ice cover and the reduction of deep-water formation in the Nordic seas (Boers et al., 2018) led to a reduction of the interhemispheric northward heat advection and thus, to a warming of the South Atlantic basin (Landais et al., 2015), as observed in GL-1180 and GL-1090 temperature records. This thermal response is also well represented by freshwater hosing model experiments, i.e. forcing the reduction of
325 deep water formation in the Nordic seas by large inputs of freshwater within the North Atlantic (Chang et al., 2008). The AMOC collapse leads to the accumulation of warm waters (Dahl et al., 2005) in the South and tropical Atlantic because of the

reduction of the cross-equatorial North Brazil Current transport, especially at thermocline levels (Laurian and Drijfhout, 2011), thus inducing the upper thermocline warming ($\text{TEX}_{86}^{\text{H}}\text{-SubT}$) observed in the GL-1180 core (Figure 5E).

During glacial terminations, the occurrence of millennial-scale events, i.e. Heinrich events, led to a substantial warming of the South Atlantic as explained by the thermal bipolar seesaw model. Nevertheless, the upper thermocline temperature record of GL-1180 core exhibits an increase of $+2^{\circ}\text{C}$ during Greenland Stadials which accounts only for the half of the deglacial warming (Figures 3, Figure 5). Thus, we cannot exclude that the bipolar see saw is responsible for, at least, part of the observed temperature increase during glacial terminations (Barker and Diz, 2014). However, deglaciations of the late Pleistocene are also associated with an increase of the Agulhas Leakage outflow, a major contributor of warm and saline thermocline waters to the South Atlantic (Peeters et al., 2004). Assuming the role of the bipolar seesaw in the upper ocean warming in the western tropical Atlantic, we therefore investigate the potential of Agulhas Leakage waters to explain these temperature increases.

4.3 Impact on western tropical Atlantic upper ocean temperatures of the expansion of Agulhas Leakage waters during past terminations

In the GL-1180 core, the difference between the $\text{U}_{37}^{\text{K}}\text{-SST}$ and $\text{TEX}_{86}^{\text{H}}\text{-SubT}$ temperatures (vertical gradient, ΔST , Figure 5C) exhibited minimal values over the last three deglaciations. A similar decrease in the vertical thermal contrast between the mixed layer and the upper thermocline over glacial terminations of the late Pleistocene is also observed in U_{37}^{K} and $\text{TEX}_{86}^{\text{H}}$ temperature records from cores collected within the Agulhas Current pathway along southern Africa (Caley et al., 2011; Figures 5A, B) and in the Cape Basin (Cartagena- Sierra et al., 2021). A similar deglacial temperature pattern was previously reported in the Cape Basin records (CBR cores, Figure 5D) (Peeters et al., 2004). Deglaciations are associated with an increase in the Agulhas Leakage, which is the outflow of warm and saline waters from the Agulhas Current in the South Atlantic (Nirmal et al., 2023; Peeters et al., 2004) (Figure 5E). The intensity of the Agulhas Leakage strongly depends on the position of the southern hemisphere mid-latitude westerlies (Biaostoch et al., 2009). This wind belt related with the subtropical front exhibits latitudinal migrations over orbital (Bard and Rickaby, 2009) and longer timescales (Cartagena- Sierra et al., 2021). During full glacial states, these westerlies were suggested to blow at a latitude of $\sim 35^{\circ}\text{S}$ in the Cape Basin region (Bard and Rickaby, 2009), therefore blocking the Agulhas Leakage outflow in the South Atlantic. At the onset of the deglaciation, the poleward shift of the mid-latitude westerlies initiates the leakage of the Agulhas Current, i.e. the outflow of these warm and

salty waters, to the South Atlantic. These waters then spread in the South Atlantic and mixed with South Atlantic central waters, imprinting their physical properties, high salinity (Scussolini et al., 2015), high temperatures (Kasper et al., 2014) and low vertical gradient (ΔST , Figure 5A) over the South Atlantic.

355 The changes in Agulhas Leakage outflow have first been reconstructed using subtropical foraminiferal assemblages (Peeters et al., 2004) including *G. menardii*, which is transported by Agulhas waters outflow from the Indian Ocean to the South Atlantic (Caley et al., 2014). On the western side of the South Atlantic, an increase in *G. menardii* abundance was also observed during Termination I (Santos et al., 2014) and imputed to the spreading of Agulhas Leakage warm and salty waters. The increase outflow of Agulhas waters during glacial terminations has also been suggested to explain the increase in upper
360 ocean temperature and salinity reconstructed in the western South Atlantic during TII (Ballalai et al., 2019) and the eastern South Atlantic during TI and TII (Kasper et al., 2014; Scussolini et al., 2015). Agulhas warm and salty mode waters are spreading into the South Atlantic mainly within the thermocline (Guerra et al., 2022), and constitute the main source of heat and salt to the upper limb layers of this basin (Castellanos et al., 2017). The abovementioned records reported a stronger temperature and salinity increase at upper (Kasper et al., 2014) and permanent thermocline (Ballalai et al., 2019; Scussolini et
365 al., 2015) levels compared to the surface.

Similarly, the temperature increase recorded in core GL-1180 during deglaciations has a higher amplitude in the upper thermocline (+ 5°C) than in surface (+2°C) water. In addition, the upper thermocline temperature maximum is reached during the glacial termination, while, at the surface, the temperature remains relatively high during the later interglacial. Upstream, the same temperature pattern was reported in the Mozambique Channel (Caley et al., 2011), and in the Southeast Atlantic
370 (Kasper et al., 2014; Peeters et al., 2004).

Therefore, these results indicate that the signature of Agulhas Leakage waters was observed during past glacial terminations along their pathway from the Indian Ocean to the western boundary of the South Atlantic through the Cape Basin. In addition, Agulhas Leakage waters were reported to reach the western South Atlantic on modern times (Castellanos et al., 2017; Guerra et al., 2022; Lübbecke et al., 2015) and over past deglaciations (Ballalai et al., 2019; Santos et al., 2014). Altogether, these
375 observations suggest that the upper ocean warming in the western tropical Atlantic during the last three glacial terminations is due, at least partly, to the spreading of Agulhas Leakage waters in the South Atlantic.

4.4 Proposed mechanisms to explain the upper western tropical Atlantic warming during the last 3 glacial terminations

The warming of western tropical Atlantic subsurface waters occurred, at the onset of deglaciations (~19, ~135 and ~251 kyr BP respectively for TI, TII and TIII) with a large amplitude (+ 3 to + 5°C), especially at the upper thermocline, in a relatively short time interval (< 2 kyr) indicating the advection of considerable volumes of warm waters (Figure 7). A similar temperature pattern was also reported within the Mozambique Channel (Caley et al., 2011) and in the southeastern South Atlantic (Kasper et al., 2014), on the course of the cross-oceanic Agulhas Leakage (Figure 7). As discussed in section 4.3, the poleward shift of the southern westerlies at the onset of the deglaciation led to massive outflow of Indo-Pacific waters in the South Atlantic (Bard and Rickaby, 2009; Peeters et al., 2004). During the preceding glacial period, Indian Ocean waters were circulating in closed system because of the latitude of the southern westerlies (Beal et al., 2011), increasing gradually their heat (Sun et al., 2022) and salt contents. The millennial-scale and large increase observed in these temperature records is a consequence of the progressive advection and accumulation of these waters, especially at thermocline levels (Guerra et al., 2022) in the South Atlantic (Figure 8, left panel). In the South Atlantic, the subtropical gyre was suggested to intensify during deglaciations (Laurian and Drijfhout, 2011), concomitantly with a (i) a southward migration of the ITCZ, (ii) an intensification of the SE trades and (iii) a reduction of the cross-equatorial North Brazil Current in response to large iceberg discharges in the North Atlantic (Nace et al., 2014). These mechanisms allowed for the accumulation of warm and salty waters in the whole South Atlantic basin up to tropical latitudes (Figure 8), thus explaining the pattern observed in temperature records on eastern and western boundaries of the South Atlantic (Figures 4, 7). The restricted transport by Agulhas rings and eddies on short timescales (Guerra et al., 2022; Wang et al., 2015) would require, during deglaciations, centuries to millennia to spread the Agulhas heat over the whole South Atlantic at thermocline and intermediate levels and warm this basin as indicated by temperature records along Brazilian margins (Figure 4).

In addition, a similar temperature pattern, i.e. surface and subsurface deglacial warmings, was reported in the tropical North Atlantic, thus indicating the spreading of these warm waters across the equator (Lopes dos Santos et al., 2010; Reiig et al., 2019; Schmidt et al., 2012) (Figure 7). However, the main cross-equatorial upper ocean current in the Atlantic i.e. the North Brazil Current was suggested to strongly slowdown (Laurian and Drijfhout, 2011) during abrupt AMOC collapses of the last

deglaciation (Arz et al., 1999; Bahr et al., 2018; Weldeab et al., 2006). In parallel, because the North Atlantic is cold during AMOC reduction events of glacial terminations, the eastward and cross-equatorial advection of the above-mentioned warm waters is required to explain the mirrored temperature pattern observed on the western (GL-1180 core, this study) and eastern 405 (GeoB9528-3 core, Lopes dos Santos et al., 2010) sides of the tropical Atlantic (Figure 7).

We propose here a schematic overview of surface and thermocline circulations in the tropical Atlantic to shed light on the main currents and associated temperature patterns (Figure 8). During the deglaciation, the advection of warm waters from the Indo-Pacific basins warmed the South Atlantic in response to the increase of the Agulhas Leakage. In addition, the reduced cross-equatorial AMOC transport led to the accumulation of these warm waters within the South Atlantic. The South Atlantic 410 basin warmed up, while the North Atlantic cooled during events of meltwater input and AMOC reduction, thus establishing a strong interhemispheric SST dipole in the Atlantic. The combined action of these processes over the deglaciation leads to the development of this tropical SST dipole during several millennia and maintains favourable conditions for reorganizations in the wind fields, associated surface currents and subsurface heat storages (Joyce et al., 2004; Figure 8, right panel), as observed nowadays (Chang et al., 1997; Lee and Wang, 2008). In response to the increased cross-equatorial temperature gradient, a 415 stronger than normal cross-equatorial gyre develops (Laurian and Drijfhout, 2011). It is characterized by changes in upper ocean circulation in response to wind stresses (Joyce et al., 2004; Wu et al., 2008), with the formation of a westerly wind anomaly on the western part of the tropical Atlantic. The SST anomaly strengthens, alimanted by Bjerknes-type feedbacks observed on decadal timescales, and sustains these wind stresses (Lee and Wang, 2008). Together with the reduction of the North Brazil Current (Nace et al., 2014), these tropical westerlies would promote a prominent eastward current (Xie et al., 420 1999), probably an analogous of the North Brazil Current retroflexion, which is a precursor of the North Equatorial Countercurrent. This seasonal retroflexion (Lumpkin and Garzoli, 2005) was at least prolonged over the year or even stronger (Bahr et al., 2018; Vink et al., 2001) and persistent in response to the weakening of the North Brazil Current (Wilson et al., 2011) (Figure 8). An enhanced North Equatorial Countercurrent, propagating rings and eddies within the upper layers (Lumpkin and Garzoli, 2005; Schott et al., 1995), crossed zonally the tropical Atlantic to its eastern boundary and diverged 425 towards the African continent. Its southward branch re-circulated in the South Atlantic, while its northward one supplied warm waters to the eastern tropical North Atlantic (Meier et al., 2021). Finally, the strengthened northeast trades dragged these warm

waters westward and limited their northward spreading into the North Atlantic. Existing records in the eastern (GeoB9528-3, Lopes dos Santos et al., 2010) and western (SO164-03-4, M78/1-235-1, VM12-107 Reißig et al., 2019; Schmidt et al., 2012) tropical North Atlantic exhibiting a pronounced surface and subsurface warming over glacial terminations indicates the propagation of South Atlantic waters, although interhemispheric upper ocean transport is reduced (Laurian and Drijfhout, 2011) (Figure 8). These South Atlantic waters spread meridionally and zonally within the tropical Atlantic, through mode waters at thermocline and intermediate levels (Guerra et al., 2022; Laurian and Drijfhout, 2011), explaining the large accumulation of warm waters up to 20°N (Schmidt et al., 2012; Weldeab et al., 2006). Due to air-sea interactions, the atmospheric and oceanic reorganizations greatly impact the heat transport in the tropical thermocline (Laurian and Drijfhout, 2011; Xie et al., 1999). At the end of the deglacial interval, the resumption of the AMOC and northward upper ocean transport drained out warm waters, especially at thermocline levels, explaining the subsurface i.e. upper thermocline sharp cooling (compared to the surface) observed in the low-latitude temperature records (Lopes dos Santos et al., 2010; Reißig et al., 2019; Schmidt et al., 2012) including those of GL-1180.

5. Conclusions

In this study, the past upper ocean temperature (mixed layer i.e. surface, and upper thermocline i.e. subsurface) in the western tropical Atlantic were reconstructed over the last 305 kyr using organic temperature proxies, $U^{K'}_{37}$ and TEX^H_{86} respectively. Besides a glacial-interglacial variability, the temperature records exhibited increases over the last 3 deglaciations especially at subsurface levels. Similar deglacial warmings were previously reported on the Brazilian margins and in the southeastern South Atlantic. The accumulation of warm waters in the whole South Atlantic basin were attributed, at least partly, to the reduction of the meridional cross-equatorial heat transport linked with deglacial Heinrich events. Nevertheless, the large magnitude of the subsurface warming is suggested here to be also due to the spreading, into the South Atlantic, of the Agulhas Leakage warm and salty waters in response to the poleward shift of the southern westerlies. These mode waters are propagating especially at thermocline level and were accumulating into the South Atlantic with the intensified subtropical gyre during deglacial periods. In addition, we attributed the interhemispheric spreading of these waters on the eastern and western tropical North Atlantic to the development of a cross-equatorial gyre in parallel of the reduction of the transport by

the North Brazil Current. This gyre exhibits a strong eastward upper ocean current responding to an anomalous wind stress and allow the expansion of South Atlantic warm waters to the eastern boundary of the tropical North Atlantic during, at least, the last three glacial terminations of the Late Pleistocene.

Competing interests

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

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Captions of Figures and Tables

Figure 1: Regional oceanographic settings. Left panel: Upper South Atlantic Ocean currents and investigated (previously published) sediments cores plotted on the mean annual sea temperature at 150 m depth (World Ocean Atlas 2013) plotted at 1°x1° grid and spatially interpolated using DIVA gridding (Ocean Data View software, Schlitzer, 2005). AC = Agulhas Current, AL = Agulhas Leakage, SEC = South Equatorial Current, SECC = South Equatorial Counter Current, NECC = North Equatorial Counter Current, NEC = North Equatorial Current, NBUC = North Brazil Undercurrent, NBC = North Brazil Current, BC = Brazil Current, MC = Malvinas Current, SAC = South Atlantic Current. GL-1180 (this study), MD96-2048 (Caley et al., 2011), MD96-2080 and MD02-2594 (Kasper et al., 2014), CBR (Peeters et al., 2004), 64PE-174P13 (Scussolini et al., 2015), GL-1090 (Santos et al., 2017), GeoB3910-2 (Waelbroeck et al., 2018), GL-1248 (Venancio et al., 2018), VM12-107 (Schmidt et al., 2012), SO164-03-4 (Reiðig et al., 2019), GeoB9528-3 (Lopes dos Santos et al., 2010). Right panel: Annual and seasonal temperature profiles at 8.5°S, 33.5°W (World Ocean Atlas 13).

Figure 2: Molecular structure of long-chain C₃₇ alkenones and isoprenoid GDGTs for the calculation of temperature proxies

Figure 3: New temperature records (this study) compared to previously published records in the GL-1180 core: A) U^K₃₇-derived sea surface temperature (U^K₃₇-SST) using the global calibration at 0 m (Müller et al., 1998) record of GL-1180 core, this study; B) TEX^H₈₆-derived upper thermocline temperature (TEX^H₈₆-SubT) TEX₈₆ calculated with the equation from Schouten et al., (2002) and inferred upper thermocline temperature (Kim et al., 2010) record of GL-1180 core, this study; C) Mg/Ca-main thermocline temperature (Mg/Ca-ThT) based on *G. truncatulinoides*, GL-1180 core (Nascimento et al., 2021b), thick black line representing 3-points running-average. The y-scale is reduced twice compared to organic proxies; D) benthic foraminifera δ¹³C (*Cibicides spp.*), GL-1180 core (Nascimento et al., 2021b); E) benthic foraminifera δ¹⁸O (*Cibicides spp.*) in black, GL-1180 core (Nascimento et al., 2021a), thick black line representing 3-points running-average. Dark and light shaded areas represent interglacial periods and glacial terminations (TI, TII and TIII) respectively. Squared numbers refers to Marine Isotope Stages (MIS 1 to MIS 9) identified on the benthic foraminifera δ¹⁸O record of GL-1180. Total mean (calibration + analytical) errors are represented along the Y-axes for U^K₃₇-SST and TEX^H₈₆-SubT.

Figure 4: Western South Atlantic records of sea temperature -ST- (SST for surface, SubT for upper thermocline, ThT for main thermocline) over the last 3 glacial terminations, on their respective original chronologies. A) SST (TEX^H₈₆) and SubT (U^K₃₇) temperature records of GL-1090 core (Cruz et al., 2023). B) SST (Mg/Ca, *G. ruber*) and ThT (Mg/Ca *G. inflata*) temperature records of GL-1090 core

(Ballalai et al., 2019; Santos et al., 2017). C) SST stack (presented in temperature anomaly versus the late Holocene) for the Brazilian margin (17 records) (Santos et al., 2022). D) SST ($U^{K'_{37}}$) and SubT ($TEX^{H_{86}}$) temperature records of GL-1180 core, this study. E) Benthic foraminifera $\delta^{18}O$ records for the above-mentioned cores. Light shaded areas represent glacial terminations (identified on respective benthic foraminifera $\delta^{18}O$ records). Dark shaded areas represent the Heinrich Event 1 (H1) and the Younger Dryas (YD). The Bølling-Allerød (B/A) event is also indicated.

Figure 5: Investigating the Bipolar Seesaw impacts in the equatorial and subtropical western South Atlantic for the last 110 kyr. Above dash lines, selected paleoclimate records illustrating regional oceanic response to last glacial AMOC changes: A) Ice $\delta^{18}O$ of NGRIP Greenland Ice core project (NorthGRIP Comm. Members, 2004); B) C_{37} alkenone derived-sea surface temperature ($U^{K'_{37}}$ -SST) of GeoB3910-2 (Jaeschke et al., 2007); C) benthic foraminifera (*C. wuellerstorfi*) $\delta^{13}C$ of core GeoB3910-2 (Waelbroeck et al., 2018). Below dash lines, selected records showing thermocline changes on the Brazilian margins: D) upper ocean stratification $\Delta\delta^{18}O$ (between *G. ruber* and *N. dutertrei*) in core GL-1248 (Venancio et al., 2018); E) upper thermocline temperature ($TEX^{H_{86}}$ -SubT) of GL-1180 record (this study); F) thermocline dwelling *Globorotalia inflata* Mg/Ca-derived main thermocline temperature (Mg/Ca-ThT) from GL-1090 record (Santos et al., 2020); G) $\delta^{18}O$ sea-water corrected for ice-volume changes and Mg/Ca-derived main thermocline temperature (*G. truncatulinoides*, Mg/Ca-ThT) of core GL-1180 (Nascimento et al., 2021b).

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Figure 6: Records of the Agulhas Leakage. A) Difference between $U^{K'_{37}}$ and $TEX^{H_{86}}$ temperatures (ΔST) in the MD96-2048 record (Caley et al., 2011); B) SST records of the MD96-2048 (Caley et al., 2011); C) Difference between $U^{K'_{37}}$ -SST and $TEX^{H_{86}}$ -SubT temperature (ΔST) in the GL-1180 core (this study); D) $U^{K'_{37}}$ -SST record in CBR (*Cape Basin Records*) cores (Peeters et al., 2004); E) Agulhas Leakage fauna assemblages in the CBR cores (Peeters et al., 2004); F) benthic foraminifera $\delta^{18}O$ records of cores MD96-2048 (Caley et al., 2011), CBR (Peeters et al., 2004) and GL-1180 (this study).

Figure 7: Sea Temperature (ST) records exhibiting a strong subsurface warming over deglacial interval from the Indian Ocean to the western tropical North Atlantic. A) SST (Mg/Ca, *G. ruber*) and main thermocline seawater temperature (Mg/Ca-ThT, *G. truncatulinoides*) records of core SO164-03-4 (Reiðig et al., 2019). B) SST ($U^{K'_{37}}$) and SubT ($TEX^{H_{86}}$) temperature records of GeoB9528-3 core (Lopes dos Santos et al., 2010). C) SST ($U^{K'_{37}}$) and SubT ($TEX^{H_{86}}$) temperature records of MD96-2080 (Termination I) and MD02-2594 (TII) core (Kasper et al., 2014). D) SST ($U^{K'_{37}}$) and SubT ($TEX^{H_{86}}$) temperature records of GL-1180 core (this study). E) SST ($U^{K'_{37}}$) and SubT ($TEX^{H_{86}}$) temperature records of MD96-2048 core (Caley et al., 2011). F) Benthic foraminifera $\delta^{18}O$ records for the above-mentioned cores for which data are available. Dark shaded areas represent deglacial intervals (noted TI, TII and TIII) and blue

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shaded areas represent the Heinrich Event 1 (H1) and the Younger Dryas (YD). The Bølling-Allerød (B/A) event is also indicated.

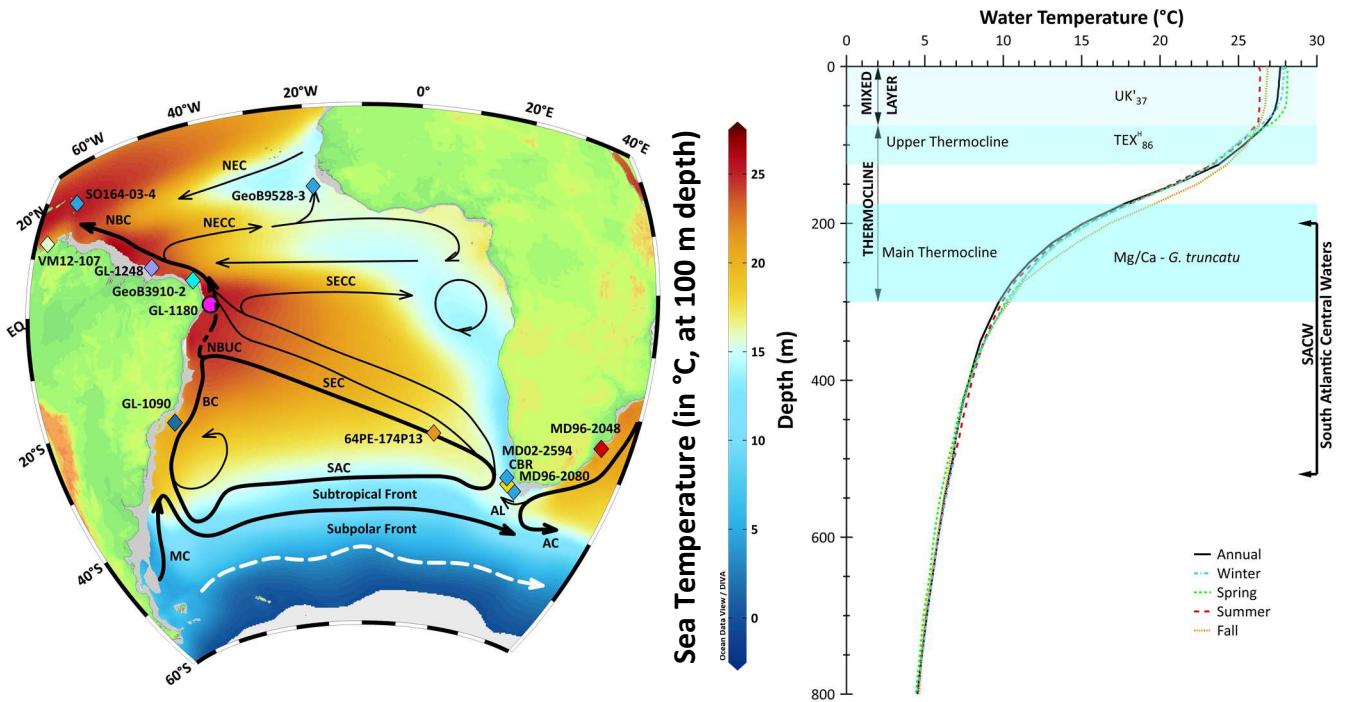
75 **Figure 8:** Schematic view of upper ocean temperature and surface currents during classical fullglacial interval (left) and the later deglacial period (right). Position of core exhibiting a surface and subsurface warmings over at least, one glacial termination are shown: VM12-107 (VM12 on the map, Schmidt et al., 2012), M78/1-235-1 (M78/1 on the map, Reißig et al., 2019), GeoB9528-3 (Lopes dos Santos et al., 2010), GL-1180 (*this study*), GL-1090 (Santos et al., 2017), 64Pe-174P13 (64PE on the map, Scussolini et al., 2015), MD02-2594 (2594 on the map, Kasper et al., 2014), MD96-2080 (2082 on the map, Kasper et al., 2014) and MD96-2048 (2048 on the map, Caley et al., 2011) cores used to explain the temperature pattern and circulations changes. Dotted lines correspond to winds (North-East trades and mid-latitudes westerlies). Solid arrows (thick when the current is strong, thin when it is weak) correspond to surface current: Agulhas Current (AC), Agulhas Leakage (AL), South-Equatorial Current (SEC), Brazil Current (BC), South Atlantic Current (SAC), North Brazil Current (NBC), North Equatorial Counter Current (NECC) and North Equatorial Current (NEC). Colour represent roughly upper ocean temperature between hot (red), temperate (orange) and cold (blue).

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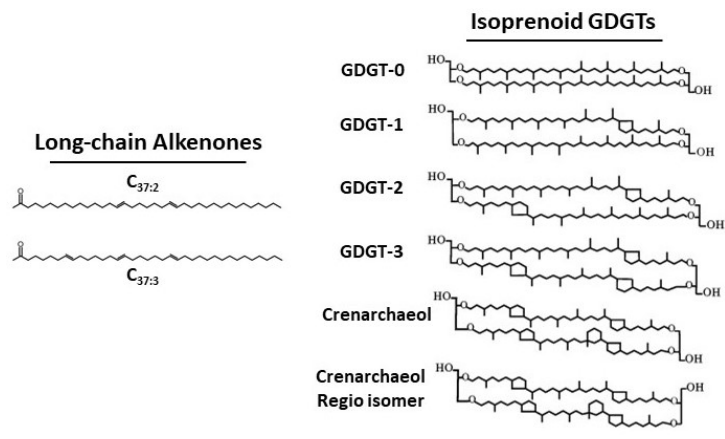
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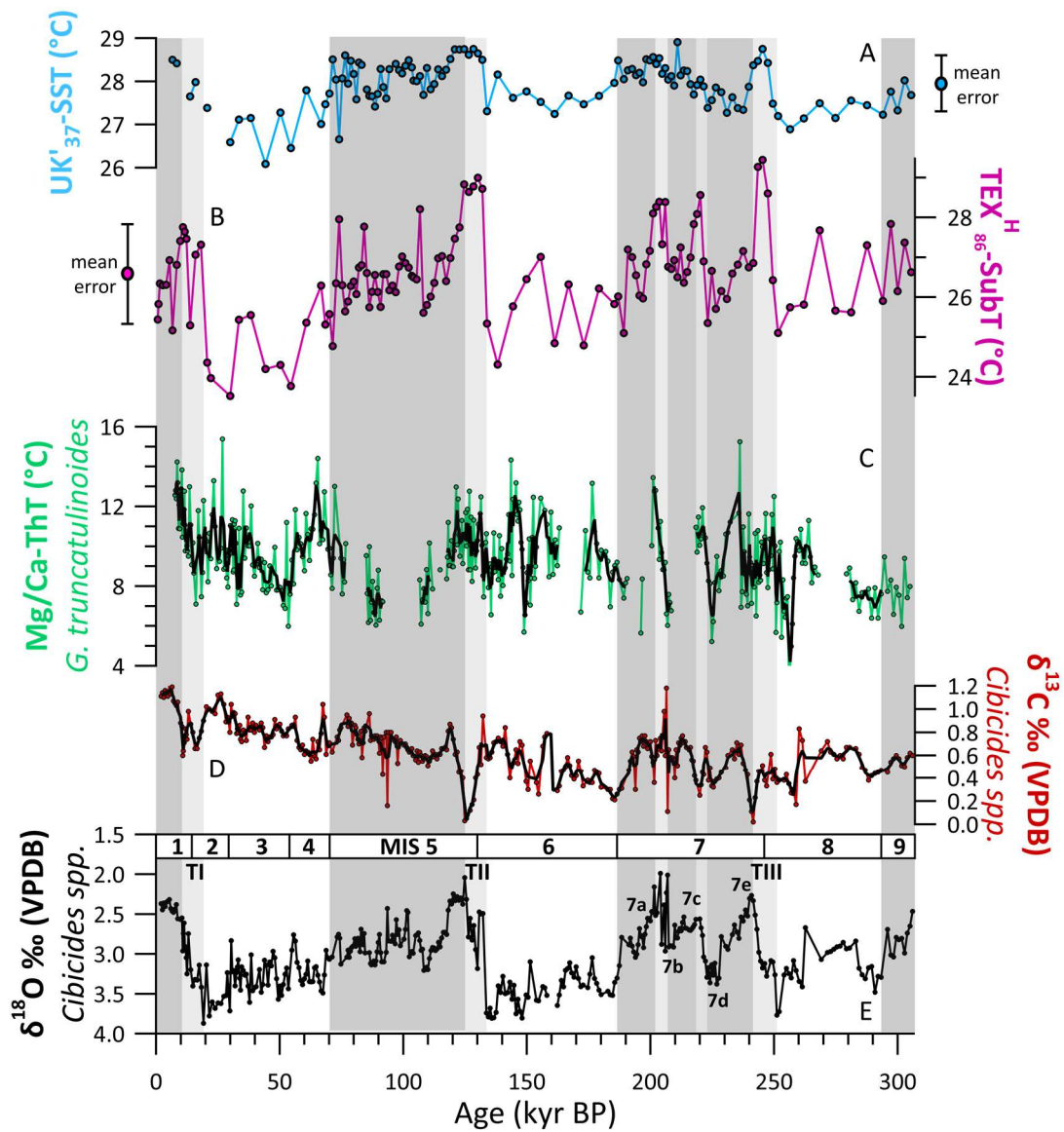


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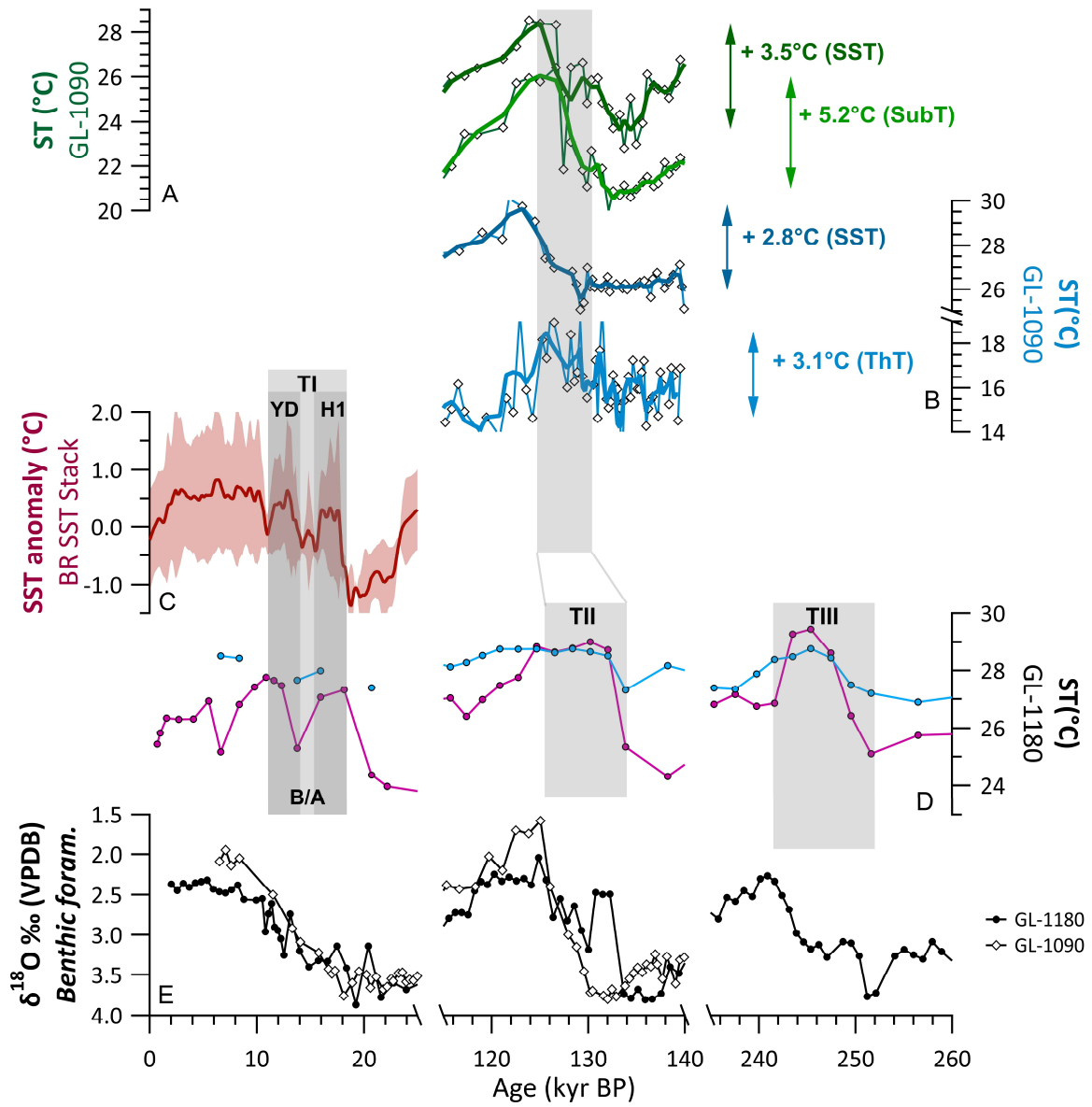


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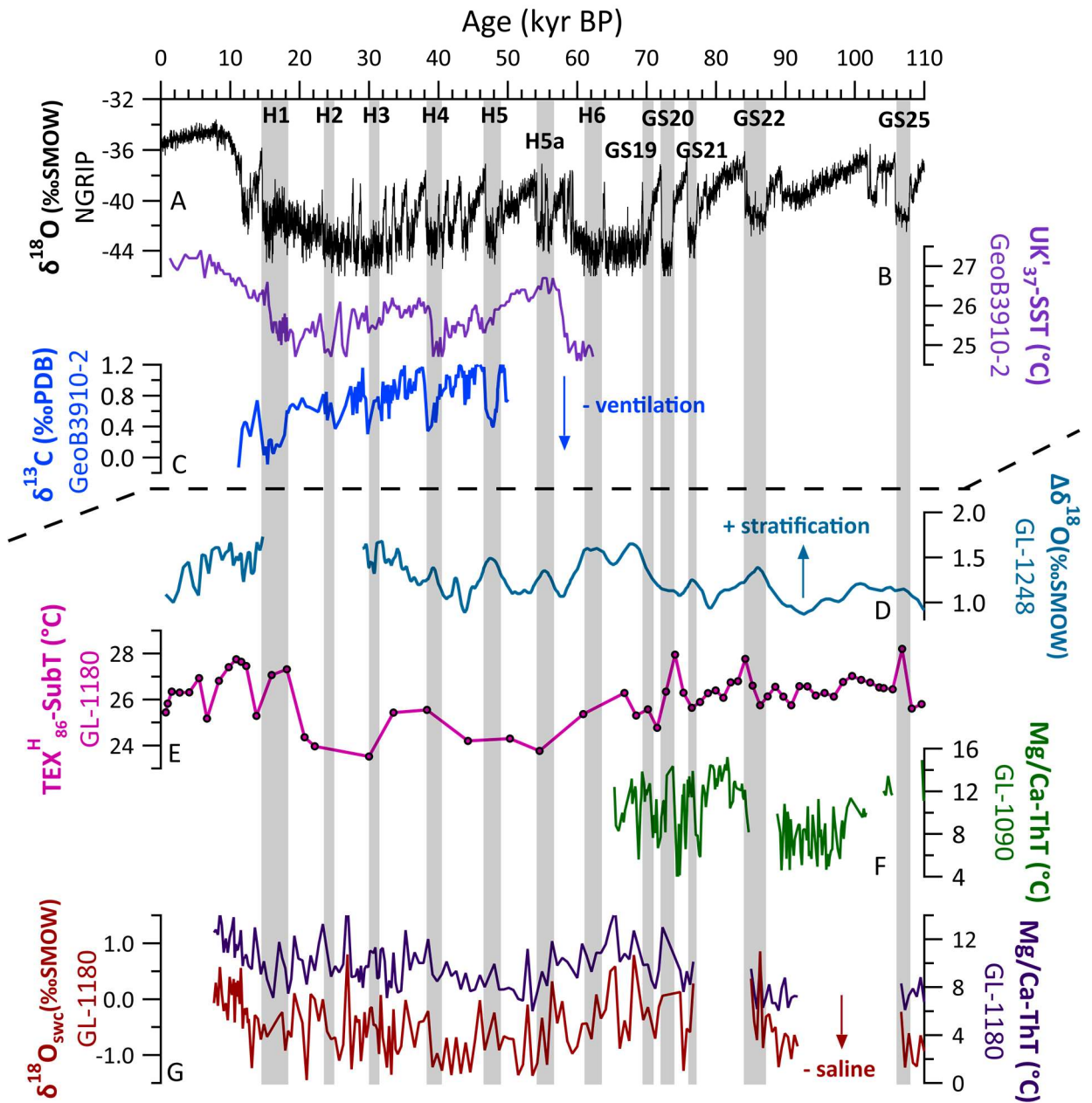
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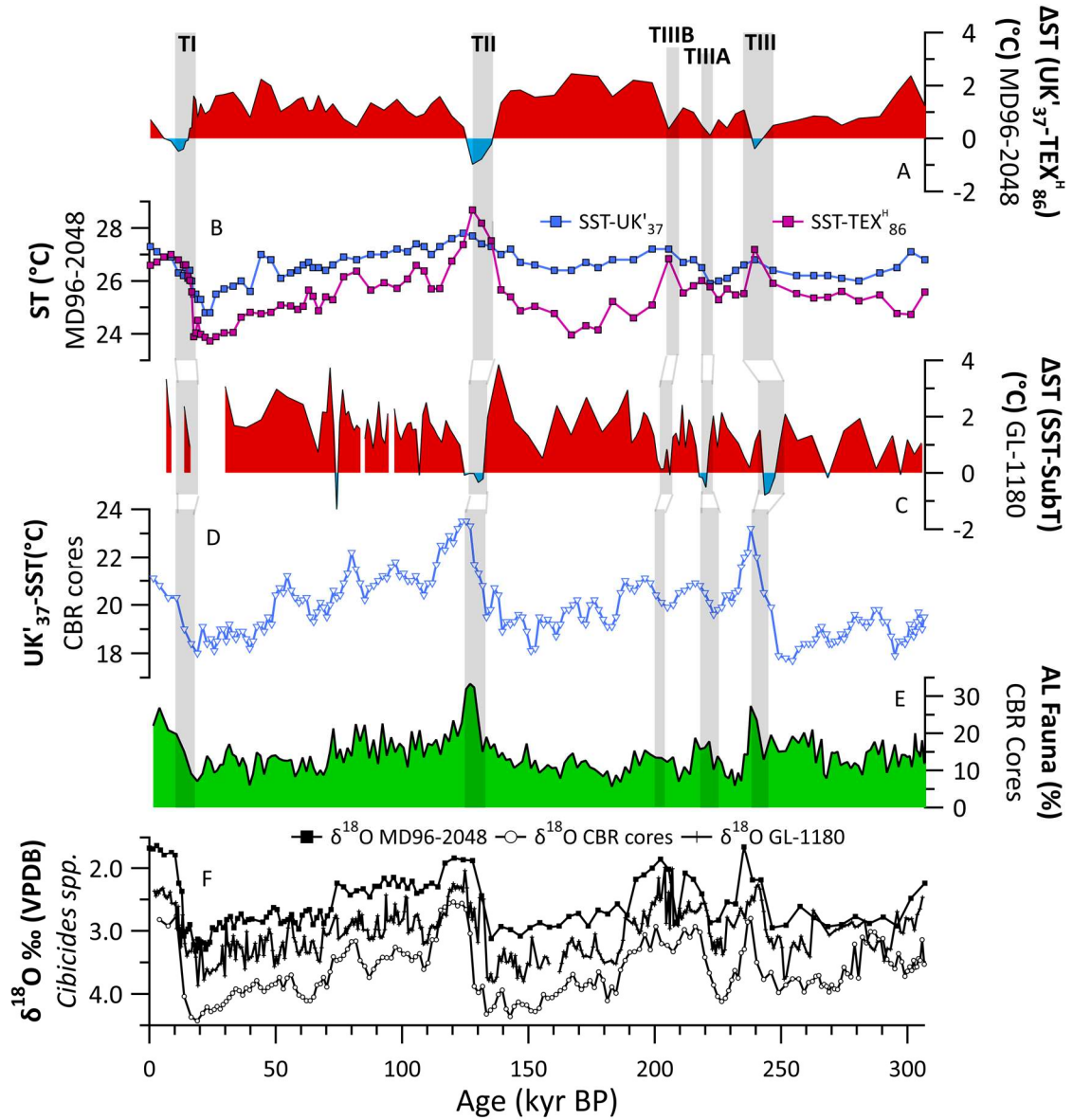


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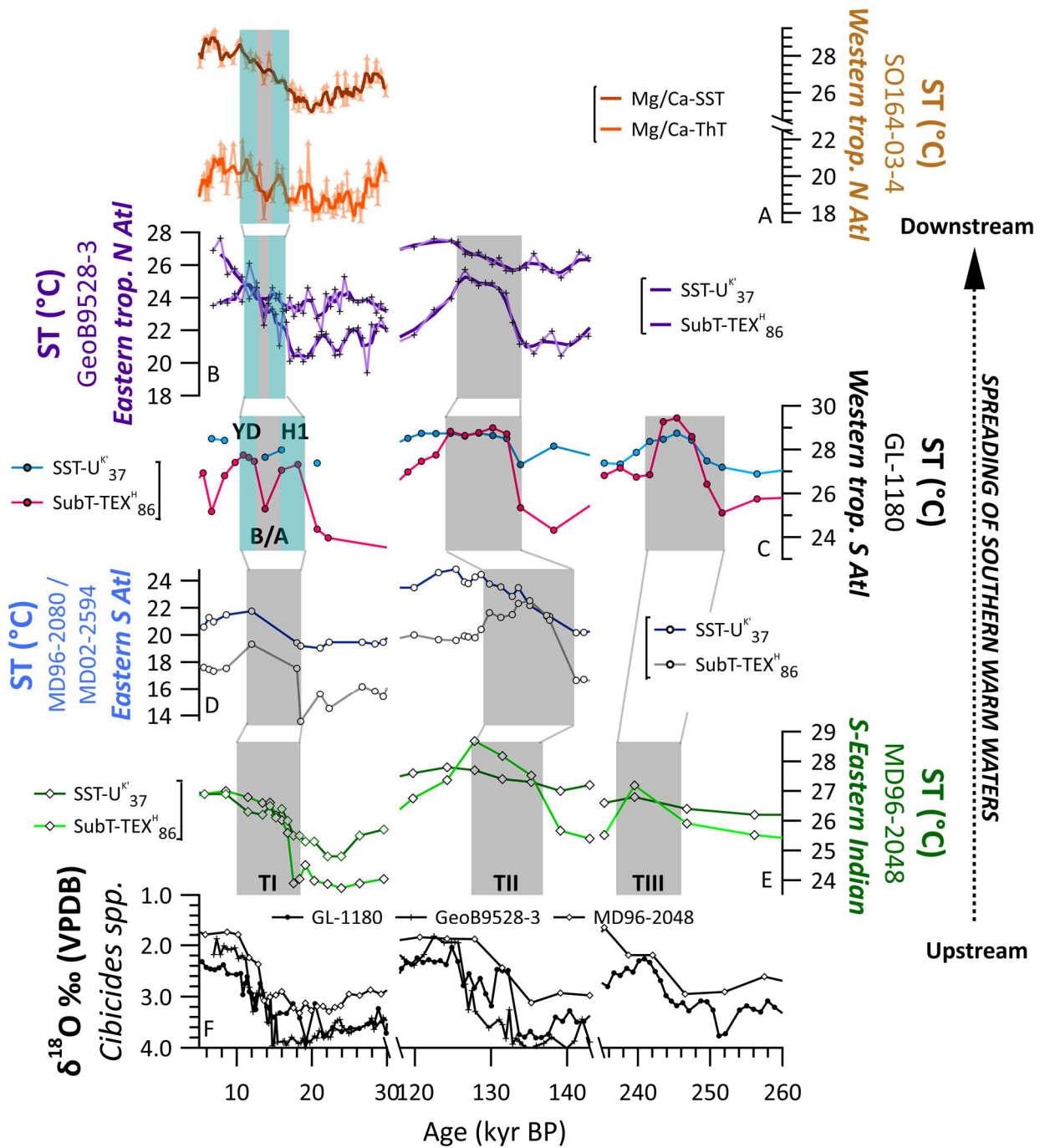


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ThT) from GL-1090 record (Santos et al., 2020); G) $\delta^{18}\text{O}$ sea-water corrected for ice-volume changes and Mg/Ca-derived main thermocline temperature (*G. truncatulinoides*, Mg/Ca-ThT) of core GL-1180 (Nascimento et al., 2021b).



140 Figure 14: Records of the Agulhas Leakage. A) Difference between UK'_{37} and $\text{TEX}^{\text{H}_{86}}$ temperatures (ΔST) in the MD96-2048 record (Caley et al., 2011); B) SST records of the MD96-2048 (Caley et al., 2011); C) Difference between UK'_{37} -SST and $\text{TEX}^{\text{H}_{86}}$ -SubT temperature (ΔST) in the GL-1180 core (this study); D) UK'_{37} -SST record in CBR (*Cape Basin Records*) cores (Peeters et al., 2004); E) Agulhas Leakage fauna assemblages in the CBR cores (Peeters et al., 2004); F) benthic foraminifera $\delta^{18}\text{O}$ records of cores MD96-2048 (Caley et al., 2011), CBR (Peeters et al., 2004) and GL-1180 (this study).



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Figure 15: Sea Temperature (ST) records exhibiting a strong subsurface warming over deglacial interval from the Indian Ocean to the western tropical North Atlantic. A) SST (Mg/Ca, *G. ruber*) and main thermocline seawater temperature (Mg/Ca-ThT, *G. truncatulinoides*) records of core SO164-03-4 (Reißig et al., 2019). B) SST ($U^{K'_{37}}$) and SubT ($TEX^{H_{86}}$) temperature records of GeoB9528-3 core (Lopes dos Santos et al., 2010). C) SST ($U^{K'_{37}}$) and SubT ($TEX^{H_{86}}$) temperature records of MD96-2080 (Termination I) and MD02-2594 (TII) core (Kasper et al., 2014). D) SST ($U^{K'_{37}}$) and SubT ($TEX^{H_{86}}$) temperature records of GL-

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1180 core (this study). E) SST ($U^{K_{37}}$) and SubT ($TEX^{H_{86}}$) temperature records of MD96-2048 core (Caley et al., 2011). F) Benthic foraminifera $\delta^{18}O$ records for the above-mentioned cores for which data are available. Dark shaded areas represent deglacial intervals (noted TI, TII and TIII) and blue shaded areas represent the Heinrich Event 1 (H1) and the Younger Dryas (YD). The Bølling-Allerød (B/A) event is also indicated.

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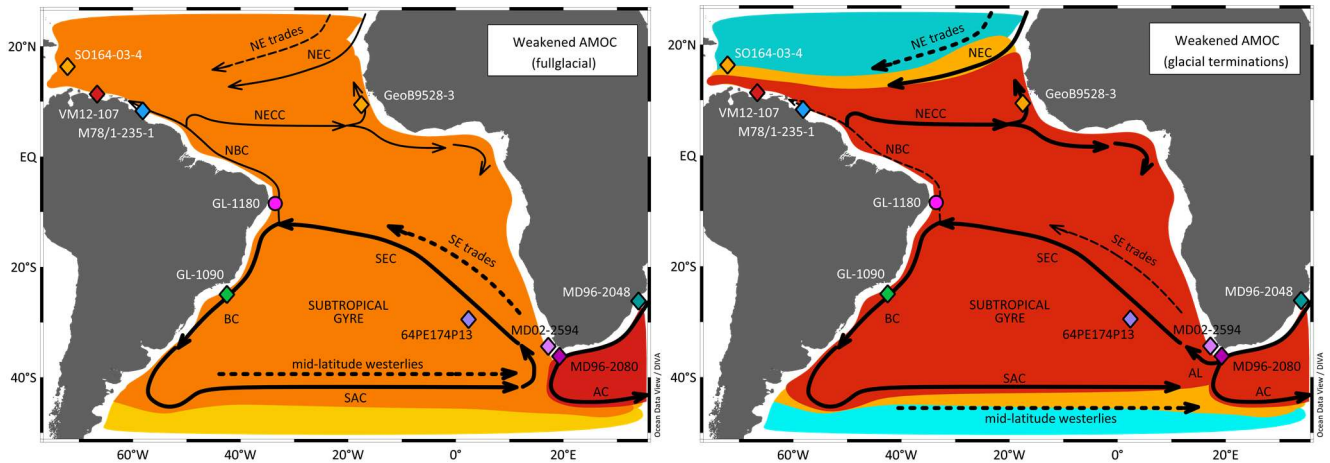


Figure 16: Schematic view of upper ocean temperature and surface currents during classical fullglacial interval (left) and the later deglacial period (right). Position of core exhibiting a surface and subsurface warmings over at least, one glacial termination are shown: VM12-107 (VM12 on the map, Schmidt et al., 2012), M78/1-235-1 (M78/1 on the map, Reißig et al., 2019), GeoB9528-3 (Lopes dos Santos et al., 2010), GL-1180 (*this study*), GL-1090 (Santos et al., 2017), 64Pe-174P13 (64PE on the map, Scussolini et al., 2015), MD02-2594 (2594 on the map, Kasper et al., 2014), MD96-2080 (2082 on the map, Kasper et al., 2014) and MD96-2048 (2048 on the map, Caley et al., 2011) cores used to explain the temperature pattern and circulations changes. Dotted lines correspond to winds (North-East trades and mid-latitudes westerlies). Solid arrows (thick when the current is strong, thin when it is weak) correspond to surface current: Agulhas Current (AC), Agulhas Leakage (AL), South-Equatorial Current (SEC), Brazil Current (BC), South Atlantic Current (SAC), North Brazil Current (NBC), North Equatorial Counter Current (NECC) and North Equatorial Current (NEC). Colour represent roughly upper ocean temperature between hot (red), temperate (orange) and cold (blue).

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