Increased abyssal ocean density stratification across the Middle Pleistocene Transition

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Abstract. We report ~~basinal~~basin-scale and ~~global~~globally distributed compilations of deep-water temperature and $\delta^{18}O_{seawater}$ for the past 1.5 million years using ~~tandem~~paired oxygen isotopic ($\delta^{18}O$) and Mg/Ca measurements of benthic foraminifera. Across the Middle Pleistocene Transition (MPT), interbasinal gradients suggest North Atlantic deep-water became colder ~~and~~while Pacific deep-water ~~saltier~~became more saline during glacial periods after ~~~900~~930 thousand years ago~~. Salinity in source areas~~ (ka). The $\delta^{18}O$ of seawater increased in both the deep Atlantic and Pacific across the MPT indicating increased continental ice volume. But the increase was greater in the Pacific than the Atlantic, which we suggest represents increased salinity of Southern Component Water (SCW). We propose the following sequence of changes during the MPT as a working hypothesis: (1) freshwater input to the marginal seas around Antarctica was reduced beginning at 930 ka by decreased ~~meltwater discharge from ice sheets~~melting of the Antarctic Ice Sheet and/or marine ice shelves and increased sea ice ~~extent, which led to increased~~ formation in the Southern Ocean; (2) the salinity and density of SCW increased, resulting in enhanced abyssal stratification ~~of~~and rendering the ~~abyssal ocean. The~~ deep ocean ~~became~~ a more effective carbon trap; (3) together with increased export production of organic matter and ~~lowered~~reduced deep-to-surface water exchange in the Southern Ocean, carbon storage in the deep ocean increased, lowering glacial atmospheric ~~carbon dioxide, leading to expansion of~~ $p$$CO_2$; (4) lower $p$$CO_2$ permitted the growth of larger continental ice sheets, which reached a critical size and ~~longer~~lengthened the glacial cycles. ~~Results support a physical~~Our hypothesis supports an important role for abyssal ocean stratification in ~~explaining the MPT. Collectively, our deep ocean stacks lend support to hypotheses proposing that the MPT resulted from a progressive drawdown in glacial atmospheric~~ ~~$p$$CO_2$, a conclusion that awaits verification~~the MPT, and requires further testing with additional benthic Mg/Ca-$\delta^{18}O$ records, numerical model simulations, and forthcoming atmospheric $p$$CO_2$ and mean ocean temperature results from the Beyond EPICA–Oldest Ice core from Antarctica.

1 Introduction

The Middle Pleistocene Transition (MPT), which occurred between ~1.25 and 0.64 million years ago (Ma), marks a major reorganisation of the global climate system and remains a central topic of palaeoclimate research since its first recognition over 50 years ago (Shackleton and Opdyke, 1976). Benthic foraminiferal oxygen isotope ($\delta^{18}O$) records are marked by a shift from dominant 41 thousand year (kyr) to quasi-100 kyr glacial-interglacial cycles across the MPT. There was no commensurate change in orbital forcing (Clark et al., 2006; Pisias and Moore, 1981), suggesting processes internal to the climate system were responsible. The exact cause(s) of the MPT remain(s) an open question (see reviews by (Berends et al., 2021a; Clark et al., 2006; Herbert, 2023; McClymont et al., 2013)). Uncertainties include whether the transition was gradual or abrupt (Sosdian and Rosenthal, 2009; Elderfield *et al.*, 2012; Ford *et al.*, 2016; Legrain, Parrenin and Capron, 2023) and whether it was driven primarily by Northern or Southern Hemisphere processes (An et al., 2024; Basak et al., 2026; Hines et al., 2024; Pérez-Montero et al., 2026; Starr et al., 2021; Williams et al., 2024; Wirths et al., 2025; Yehudai et al., 2021). Most hypotheses invoke a long-term global cooling, commonly attributed to a reduction in atmospheric carbon dioxide ($p$$CO_2$) (Berends et al., 2021a; Clark et al., 2006; Willeit et al., 2019). Yet, despite extensive work, the relative roles of ice-sheet dynamics and carbon-cycle feedbacks, and their timing remain poorly constrained.

Proposed explanations for the MPT can be broadly grouped into three categories: (i) internal ice-sheet dynamics, most notably the regolith hypothesis invoking progressive changes in basal conditions leading to the exposure of high-friction crystalline bedrock (Clark et al., 2006; Clark and Pollard, 1998); (ii) changes in ice-sheet geometry and extent, particularly involving North American and/or Antarctic ice sheets (An et al., 2024; Bintanja and Van De Wal, 2008; Elderfield et al., 2012; Raymo et al., 2006; Wirths et al., 2025); and (iii) feedbacks associated with long-term changes in the global carbon cycle, including deep-sea carbon storage and atmospheric $p$$CO_2$ variability (Berger et al., 1999; Chalk et al., 2017; Farmer et al., 2019; Lear et al., 2016; Qin et al., 2022; Shackleton, 2000; Thomas et al., 2022; Willeit et al., 2019). Increasing attention has been paid to the potential role of marine carbon cycle processes (Herbert, 2023), which are closely linked to changes in deep ocean circulation during the MPT (Hall et al., 2001; Hasenfratz et al., 2019; Kim et al., 2021; Li et al., 2026; Martínez-Garcia et al., 2010; Pena and Goldstein, 2014; Qin et al., 2022; Raymo et al., 1997). A "thermohaline circulation crisis" (Pena and Goldstein, 2014) (THC) was proposed for the deep Atlantic across the MPT, but the magnitude and extent of such an event has been questioned recently (Basak et al., 2026; Hines et al., 2024; Williams et al., 2024). Furthermore, recent findings from Allan Hills blue ice area of Antarctica show no significant decline in mean atmospheric $p$$CO_2$ over the MPT (Marks-Peterson et al., 2026), which does not necessarily preclude changes in the glacial values of atmospheric $p$$CO_2$ because the mean values are likely weighted towards interglacial periods. Forthcoming atmospheric $p$$CO_2$ measurements from the Beyond-EPICA ice core are highly anticipated, and will likely resolve the timing and magnitude of atmospheric $p$$CO_2$ changes across the MPT.

Whilst numerous studies implicate enhanced carbon storage in the deep sea and lowered glacial atmospheric $p$$CO_2$ across the MPT (Billups et al., 2018; Chalk et al., 2017; Farmer et al., 2019; Gildor and Tziperman, 2001; Hodell et al., 2003; Jaccard et al., 2013; Lear et al., 2016; Sigman et al., 2010; Starr et al., 2021; Thomas et al., 2022), no direct comparisons of the temperature and salinity controls on the density of deep water masses have been made between the North Atlantic and Pacific Ocean basins across the transition. This limits assessment of large-scale changes in abyssal density structure, and their potential role in carbon storage and atmospheric $p$$CO_2$ change (Adkins, 2013; Broecker and Peng, 1998, 1983). Here, we combine new and previously published records of Mg/Ca-derived deep-water temperature and oxygen isotopic composition of seawater

($\delta^{18}O_{seawater}$), a proxy for both ice volume and salinity, to evaluate how abyssal ocean density structure changed across the MPT.

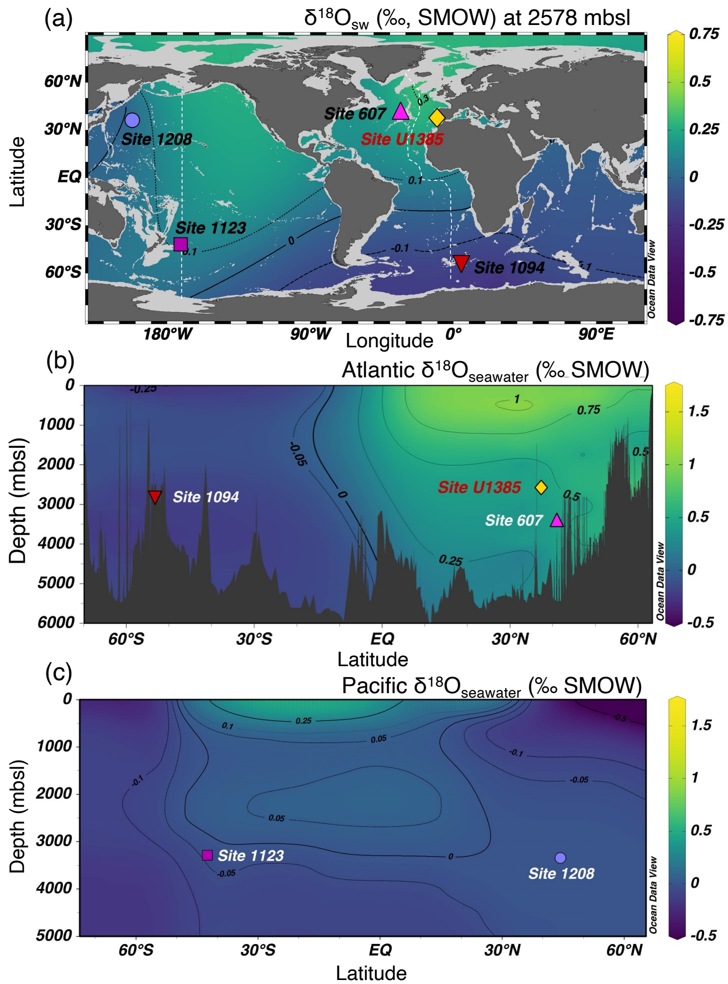
Benthic $\delta^{18}O$ records are invaluable tools for reconstructing glacial-interglacial variability of the Quaternary (Shackleton, 1967; Shackleton and Opdyke, 1973). However, interpretation of benthic $\delta^{18}O$ records is complicated because the signal is dependent on both temperature and the $\delta^{18}O_{seawater}$, the latter varying with global ice volume and regional hydrography including salinity changes at sites of deep-water formation (Waelbroeck et al., 2002). Deconvolution of benthic $\delta^{18}O$ into its temperature and $\delta^{18}O_{seawater}$ components has been accomplished by paired measurement of Mg/Ca and $\delta^{18}O$ of calcite (Elderfield et al., 2010, 2012; Ford et al., 2016; Ford and Raymo, 2020; Hasenfratz et al., 2017, 2019; Sosdian and Rosenthal, 2009), with estimates of propagated uncertainty (Thirumalai et al., 2016). Other indirect approaches include inverse modelling that combines benthic $\delta^{18}O$ data with ice sheet models of varying complexity (Berends et al., 2021b; Bintanja and Van De Wal, 2008) and an iterative process-based approach that assesses changes in relationships between the oxygen isotopic composition of ice ($\delta^{18}O_{ice}$), sea level, temperature, and $\delta^{18}O_{seawater}$ (Rohling et al., 2021). More recently, changes in global mean sea surface temperatures (ΔGMSST) combined with proxy-based deep ocean temperature estimates, have been used to infer changes in mean ocean temperature (ΔMOT), enabling separation of temperature and ice volume components from a global benthic $\delta^{18}O$ stack (Clark et al., 2025). These different approaches can yield widely divergent interpretations of Quaternary temperature and ice volume evolution. For example, during Marine Isotope Stage (MIS) 22 (~900 ka), Rohling et al. (2021) infer a substantial increase in ice volume with limited deep-water cooling, consistent with earlier Mg/Ca-based reconstructions (Elderfield et al., 2012; Ford and Raymo, 2020). In contrast, the reconstruction of Clark et al. (2025) shows pronounced deep-water cooling with relatively little change in ice volume across the MPT.

Here we present a new 1.5-million-year-long record of North Atlantic deep-water temperature, using Mg/Ca measured in calcite of two infaunal benthic foraminifera (*Uvigerina peregrina* and *Globobulimina affinis*) from sediment cores recovered at International Ocean Discovery Program (IODP) Site U1385 (37°34.3′N, 10°7.6′W, 2578 metres below sea level [mbsl]) from the Iberian Margin (Hodell et al., 2015, 2023) (Figs. 1, S1 and S2; Table 1). The Mg/Ca-derived temperature is paired with benthic $\delta^{18}O$ to isolate the $\delta^{18}O_{seawater}$ signal (Sect. 2.3; Figs. 2 and S3). Our record provides an important complement to existing North Atlantic data from Deep Sea Drilling Project (DSDP) Site 607 (Sosdian and Rosenthal, 2009), where Mg/Ca-derived temperature estimates were based partly on *Cibicidoides wuellerstorfi*. These results were called into question because this epibenthic taxa can be affected by changes in carbonate ion concentrations of bottom waters (Yu and Broecker, Wallace, 2010) (Sect. 3.1; Supplement). This criticism was subsequently refuted (Sosdian and Rosenthal, 2010) prompting additional Mg/Ca measurements of infaunal benthic taxa (*Uvigerina* spp.) at Site 607 (Ford et al., 2016).~~Benthic oxygen isotope ($\delta^{18}O$) records are invaluable tools for identifying glacial-interglacial cycles of the Quaternary and providing information about changes to Earth's energy imbalance (Baggenstos et al., 2019; Bereiter et al., 2018; Grimmer et al., 2025; Shackleton et al., 2023). The interpretation of benthic $\delta^{18}O$ records is complicated, however, because the signal is dependent on both temperature and the oxygen isotopic composition of seawater ($\delta^{18}O_{seawater}$), which varies with global ice volume and local hydrographic effects (Waelbroeck et al., 2002). The deconvolution of the global deep-water benthic $\delta^{18}O$ record into its temperature and~~

$\delta^{18}O_{seawater}$ components has been accomplished by tandem measurement of Mg/Ca and $\delta^{18}O$ (Elderfield et al., 2010, 2012; Ford et al., 2016; Ford and Raymo, 2020; Hasenfratz et al., 2017, 2019; Sosdian and Rosenthal, 2009), with estimates of propagated uncertainty (Thirumalai et al., 2016). Other indirect approaches have also been used previously including: inverse models combining observed benthic $\delta^{18}O$ data with ice sheet models of varying complexity (Berends et al., 2021; Bintanja and Van De Wal, 2008); or, an iterative process-based approach assessing changes in relationships between the oxygen isotopic composition of ice ($\delta^{18}O_{ice}$), sea level, temperature, and $\delta^{18}O_{seawater}$ (Rohling et al., 2021). More recently changes in global mean sea surface temperatures (ΔGMSST) along with proxy-based deep ocean temperatures have been used to infer changes in mean ocean temperature (ΔMOT), which are then used to extract the deep ocean temperature and ice volume components from a benthic $\delta^{18}O$ stack (Clark et al., 2025). Some of these approaches have led to vastly divergent interpretations of the temperature and ice volume history of the Quaternary (Clark et al., 2025).

Here we present a new 1.5-million-year-long record of North Atlantic deep-water temperature, reconstructed using Mg/Ca measured in calcite of infaunal benthic foraminifera (*Uvigerina peregrina* and *Globobulimina affinis*) from sediment cores recovered at International Ocean Discovery Program (IODP) Site U1385 from the Iberian Margin, spanning the Middle Pleistocene Transition (MPT) (Figs. 1, S1 and S2; Table 1 see Sect. 2 Materials and Methods). The MPT occurred between ~1.25 to 0.64 million years ago (Ma) when glacial-interglacial cycles in the Earth's climate system changed from dominant periods of 41 kyr to quasi-100 kyr without a commensurate change in orbital forcing (Clark et al., 2006; Pisias and Moore, 1981; Shackleton and Opdyke, 1976). The Site U1385 record from the Iberian Margin is important because of previous concerns raised about the reliability of the Mg/Ca record measured partially on epibenthic foraminifera (Sosdian and Rosenthal, 2009) at Site 607 from the central North Atlantic (Yu and Broecker, Wallace, 2010)

We combine (stack) North Atlantic records from Sites U1385 and 607 (Ford et al., 2016; Sosdian and Rosenthal, 2009) and compare them with equivalent Pacific records (Sites 1123 and 1208 (Elderfield et al., 2012; Ford and Raymo, 2020)) to reconstruct the basinal histories of temperature and salinity change (Sect. 2.5–2.6). The Atlantic and Pacific records are then combined with Site 1094 from the South Atlantic sector of the Southern Ocean (Hasenfratz et al., 2019) (Figs. 1, S1 and S2; Table 1) to produce globally distributed stacks. Although the number of deconvolved benthic Mg/Ca–$\delta^{18}O$ records available for stacking is few over the past 1.5 million years (Myr), some distinct differences emerge in the patterns of inferred deep ocean temperature and salinity change between the North Atlantic and Pacific basins.



~~(Sect. 3.1; Supplement). Subsequent Mg/Ca measurements of infaunal *Uvigerina* at Site 607 supported previous findings (Ford et al., 2016). The Mg/Ca-derived temperature is paired with benthic $\delta^{18}O$ to isolate the $\delta^{18}O_{seawater}$ signal (Figs. 2 and 3).~~

We combined (stacked) the two North Atlantic records (Sites U1385 and 607 (Ford et al., 2016; Sosdian and Rosenthal, 2009)) and compared them with similar stacked records from the deep Pacific (Elderfield et al., 2012; Ford and Raymo, 2020) to reconstruct the basinal histories of temperature and salinity change (Sect. 2.5–2.7). The Atlantic and Pacific records are then combined with Site 1094 from the South Atlantic sector of the Southern Ocean (Hasenfratz et al., 2019) (Figs. 1, S1 and S2; Table 1) to produce global stacks.

Table 1: Site location details and modern-day deep-water properties.

Site and Citation	Latitude	Longitude	Depth (m)	$\delta^{18}O_{sw}$ (‰)	Temperature (°C)	Salinity (psu)
Iberian Margin: **U1385** (This study; Birner et al., 2016) **MD99-2334** (Skinner et al., 2003) **MD01-2444** (Skinner et al., 2007)	37°34.285' N 37°48' N 37°33' N	10°7.562' W 10°10' W 10°08' W	2578 3146 2637	~0.05 ~0.05–0.10 ~0.2	~3.3	~35
DSDP 607 (Sosdian and Rosenthal, 2009; Ford et al., 2016) **Piston Core Chain 82-24-23PC** (Sosdian and Rosenthal, 2009)	41° N 43° N	33° W 31° W	3427 3406	~0.3	~2.6	~35
ODP 1123 (Elderfield et al., 2012)	41° 47.15'S	171°29.94' W	3290	-0.05	~1.3	34.73
ODP 1208 (Ford and Raymo, 2020)	36.1° N	158.2° E	3346	~–0.23	~1.5	34.67
ODP 1094 (Hasenfratz et al., 2019)	53.18° S	05.13° E	2807	~–0.05	~0.5	34.47

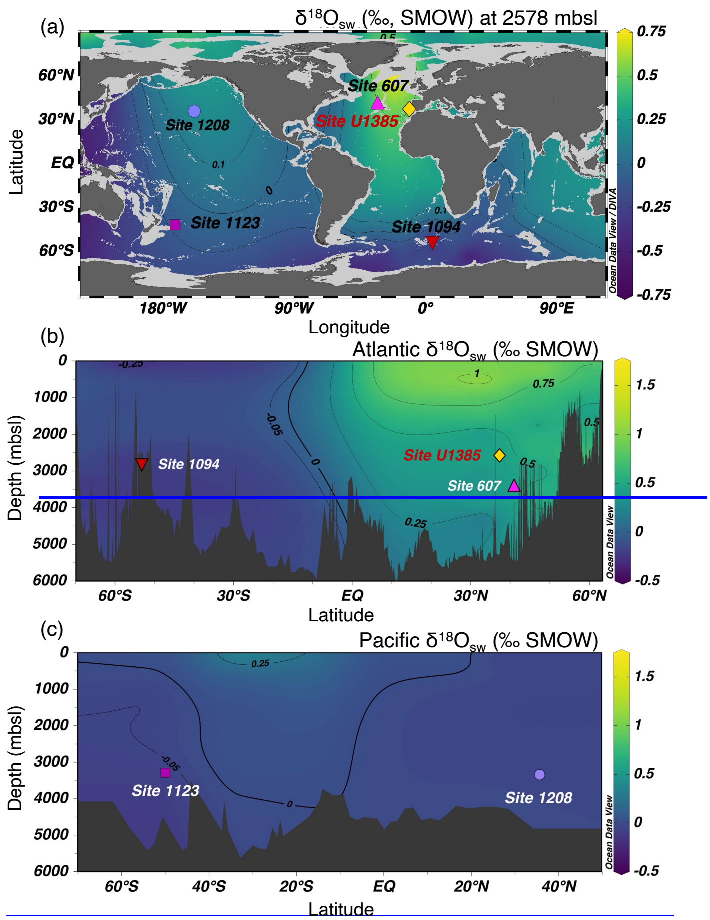
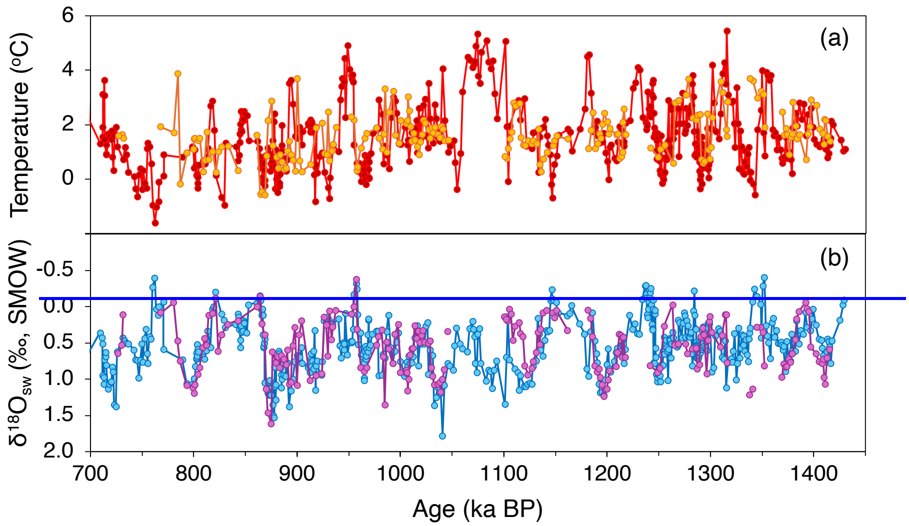


Figure 1: Meridional profiles of ~~the hydrographic isosurface~~ deep-water $\delta^{18}O_{seawater}$ showing the site locations. (a) $\delta^{18}O_{seawater}$ at 2578 metres below sea level (mbsl~~) for oceanic deep-water $\delta^{18}O_{seawater}$ relative to site locations. (a) $\delta^{18}O_{seawater}$ isosurface at the Northeast Atlantic depth (2578 mbsl)~~), corresponding to the depth of IODP Site U1385 (this study; yellow diamond) in the Northeast Atlantic, along with the geographic location of other sites referenced in the text: North Atlantic DSDP Site 607 (Ford et al., 2016; Sosdian and Rosenthal, 2009) (pink triangle) Southwestern Pacific ODP Site 1123 (Elderfield et al., 2012) (purple square); North Pacific ODP Site 1208 (Ford and Raymo, 2020) (lilac circle~~) and the Southern Ocean ODP Site 1094 (Hasenfratz et al., 2019)~~); and the Southern Ocean ODP Site 1094 (Hasenfratz et al., 2019) (red inverted triangle) located in the Atlantic sector of the Antarctic Zone. Meridional transects shown in panels (b) and (c) are indicated by white dashed lines. (b and c) Atlantic and Pacific meridional profiles of deep-

water $\delta^{18}O_{seawater}$ (‰, SMOW) showing site locations~~ and depths~~. Maps were created using GLODAPv2.2021 and GLODAPv2.2023 and GEOSECS in Ocean Data View (Schlitzer, 2019; http://odv.awi.de). Data for panel (c) are from (LeGrande and Schmidt, 2006).

Table 1: Site location details and modern-day deep-water properties.



~~**Figure 2: Comparison of deep-water temperature and $\delta^{18}O_{seawater}$ estimates using measurements of $\delta^{18}O$ and Mg/Ca of *Uvigerina peregrina* and *Globobulimina affinis* at Site U1385. (a) Mg/Ca-derived deep-water temperature for *Uvigerina peregrina* (red) and *Globobulimina affinis* (orange). (b) The corresponding estimated changes in $\delta^{18}O_{seawater}$ based on *Uvigerina peregrina* (blue) and *Globobulimina affinis* (purple).**~~

~~The rationale for stacking records of temperature and $\delta^{18}O_{seawater}$ is the same as for benthic $\delta^{18}O$~~

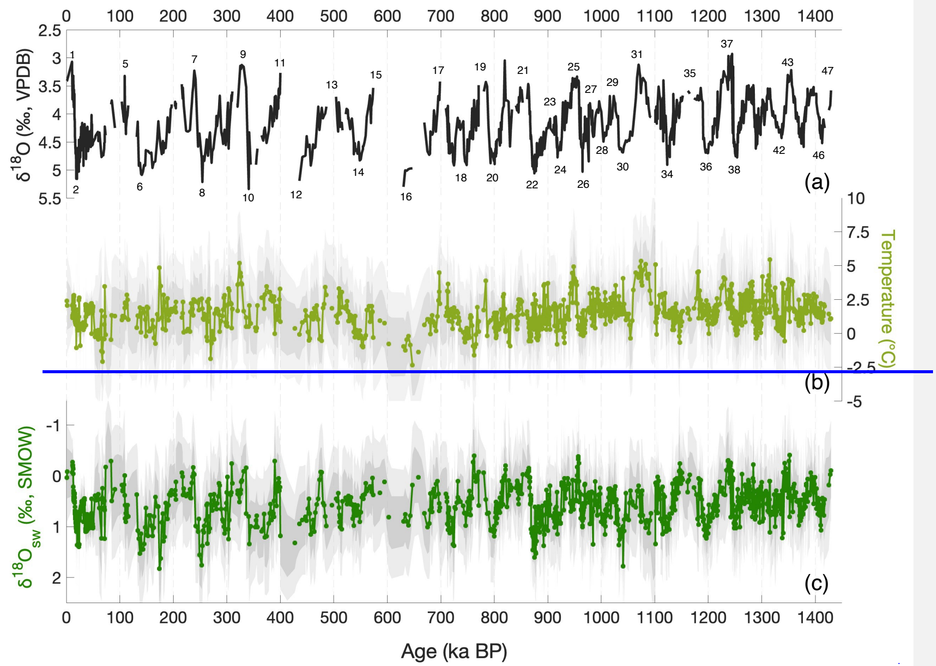
Site and Citation	**Latitude**	**Longitude**	**Depth (m)**	**$\delta^{18}O_{seawater}$ (‰)**	**Temperature (°C)**	**Salinity (PSS-78, unitless)***
Iberian Margin: U1385 (This study; Birner et al., 2016)	37°34.3′N	10°7.6′W	2578	~0.2	~3.3 (Hodell et al., 2014)	~35 (Hodell et al., 2014)
MD01-2444 (Skinner et al., 2007)	37°33′N	10°08′W	2637	~0.2		
MD99-2334 (Skinner et al., 2003)	37°48′N	10°10′W	3146	~0.2	~2.8	~34.9

				(Skinner et al., 2003)	(Skinner et al.,2003)	(Hodell et al., 2014)
DSDP 607 (Sosdian and Rosenthal, 2009; Ford et al., 2016) **Piston Core Chain 82-24-23PC** (Sosdian and Rosenthal, 2009)	41°00′N 43°00′N	32°58′W 31°00′W	3427 3406	~0.2	~2.6 (Sosdian & Rosenthal, 2009)	~35 (Sosdian & Rosenthal, 2009)
ODP 1123 (Elderfield et al., 2012)	41°47.2′S	171°29.9′W	3290	-0.05 (Elderfield et al., 2010)	~1.3 (Elderfield et al., 2010; Adkins et al., 2002)	34.73 (Adkins et al., 2002)
ODP 1208 (Ford and Raymo, 2020)	36°7.2′N	158°12.0′E	3346	~-0.03	~1.5	34.67
ODP 1094 (Hasenfratz et al., 2019)	53°10′S	5°07′E	2807	~-0.05	~0.5	34.47

*Note: All salinity values reported follow the Practical Salinity Scale (PSS-78) and are unitless.

Unless otherwise stated, temperature and salinity values are from GLODAPv2.2021 and GLODAPv2.2023 and GEOSECS accessed via Ocean Data View (Schlitzer, 2019; http://odv.awi.de), while $\delta^{18}O_{seawater}$ data for Sites 1123 and 1208 are from LeGrande and Schmidt (2006). Coordinates and water depths are taken from the original site publications cited in the first column.

~~(Imbrie et al., 1984; Lisiecki and Raymo, 2005); that is, to improve the signal/noise ratio and average out local signals to provide more representative regional and/or global mean signals of temperature and ice volume. Although the number of deconvolved benthic Mg/Ca–$\delta^{18}O$ records available for stacking is few over the past 1.5 million years (Myr), some distinct differences emerge in the patterns of deep-ocean temperature and salinity change between the North Atlantic and Pacific basins.~~



~~**Figure 3: Iberian Margin deep-water Mg/Ca-derived temperature and $\delta^{18}O_{seawater}$ estimated using PSU Solver. (a) Mixed benthic $\delta^{18}O$, mainly *Uvigerina peregrina* otherwise *Cibicidoides wuellerstorfi* and *Globobulimina affinis* offset to *Uvigerina* by +0.64 ‰ and -0.30 ‰ respectively (black line); Numbers denote Marine isotope stage (MIS) interglacials above (odd) and glacials below (even). (b) Mg/Ca-derived deep-water temperature (light green line and markers) with 1 and 2σ error bars (dark and light grey respectively). (c) Deconvolved $\delta^{18}O_{seawater}$ (dark green line and markers; note the inverted axis) error bars as in (b). All figures show combined records from Site U1385 (this study and (Birner et al., 2016), MD99-2334 (Skinner et al., 2003) and MD01-2444 (Skinner and Elderfield, 2007).**~~

All study sites are deeper than 2500 m and are therefore assumed to be representative of the ~~abyssal~~deep ocean (Figs. 1, S1 and S2; Table 1). Today, Sites U1385 (2578 ~~metres below sea level (~~mbsl~~)~~)) and 607 (3427 mbsl; (Ford et al., 2016; Sosdian and Rosenthal, 2009)) are bathed by North Atlantic Deep Water (NADW) and ~~have~~exhibit similar seawater physical properties~~.~~ (Table 1). Pacific Sites 1123 (3290 mbsl; (Elderfield et al., 2012)) and 1208 (3346 mbsl; (Ford and Raymo, 2020)) are located within Lower Circumpolar Deep Water (LCDW) and Pacific Deep Water (PDW) ~~(Elderfield et al., 2012; Ford and Raymo, 2020; Insua et al., 2014), respectively, and share very similar physical properties originating from the Southern Ocean~~

(Elderfield et al., 2012; Ford and Raymo, 2020; Insua et al., 2014), respectively, and share broadly similar physical properties originating from the Southern Ocean (Talley, 2013)~~. The temperature and salinity of the deep Pacific sites are ~1.2 ºC colder and ~0.3 practical salinity units (psu) lower than those sites in the North Atlantic, reflecting the formation of relatively warm and salty NADW (Figs. S1 and S2). Correspondingly, the $\delta^{18}O_{seawater}$ values of the Atlantic sites are on average ~0.5 ‰ greater than those of the deep Pacific sites today (Fig. 1). The Atlantic and Pacific records are merged with Site 1094 (2807 mbsl; (Hasenfratz et al., 2019)) from the South Atlantic sector of the Southern Ocean, which is currently bathed by LCDW~~. The deep western boundary current (DWBC) of the Southwest Pacific is the main pathway through which most of the cold deep water formed around Antarctica enters the Pacific Ocean (Chandler et al., 2024). The deep water gradually mixes as it flows northward modifying its properties, leading to the formation of Pacific Deep Water (PDW). Given the locations of Site 1123 within the DWBC and 1208 in the PDW, it is reasonable to assume they broadly reflect conditions in the deep Pacific.

Deep Pacific sites are ~1.2 °C colder and ~0.3 lower in salinity (PSS-78) than those in the North Atlantic, reflecting the formation of relatively warm and salty NADW (Figs. S1 and S2; Table 1). Correspondingly, the $\delta^{18}O_{seawater}$ values of the Atlantic sites are on average higher (~0.25 ‰) higher than those of the deep Pacific (Fig. 1; Table 1) (Frew et al., 2000). The Atlantic and Pacific records are merged with Site 1094 (2807 mbsl; (Hasenfratz et al., 2019)) from the South Atlantic sector of the Southern Ocean, which is currently bathed by LCDW.

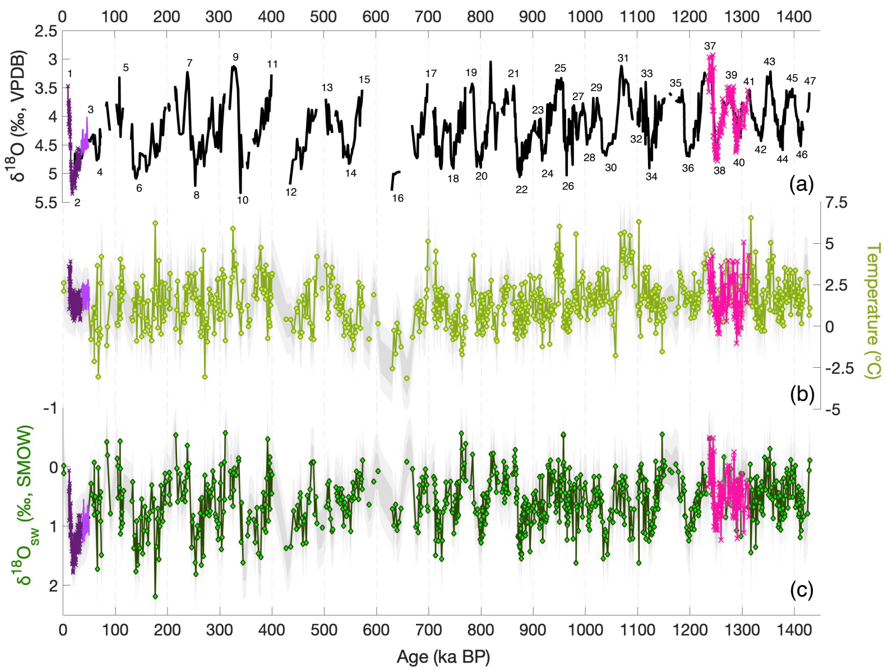


Figure 2: Iberian Margin deep-water Mg/Ca-derived temperature and $\delta^{18}O_{seawater}$ estimated using PSU Solver. (a) Composite benthic $\delta^{18}O$ record, primarily *Uvigerina peregrina*, with *Cibicidoides wuellerstorfi* and *Globobulimina affinis* values adjusted to *Uvigerina peregrina* by +0.64 ‰ and –0.30 ‰ respectively (black line). Numbers denote Marine Isotope Stage (MIS) interglacials (odd, above) and glacials (even, below). (b) Mg/Ca-derived deep-water temperature (light green line and markers) with 1σ and 2σ uncertainty envelopes (dark and light grey, respectively) incorporating both analytical and calibration uncertainties. (c) Deconvolved $\delta^{18}O_{seawater}$ (dark green line and markers; note the inverted axis) with uncertainties as in (b). All panels show Site U1385 (this study) together with records from Birner et al. (2016) (pink lines with crosses during MIS 41–37), MD01-2444 (Skinner and Elderfield, 2007) (light purple line) and MD99-2334 (Skinner et al., 2003) dark purple line with crosses).

2 Materials and Methods

2.1 Chronology

Here we report Mg/Ca measured on 870 samples of *Uvigerina peregrina*Site details and 202 samples of *Globobulimina affinis* picked from deep-sea sediments at IODP Site U1385 located on the Iberian Margin (37º34.285' N, 10º7.562' W, 2578 mbsl) (Hodell, 2014; Hodell et al., 2015) (Figs 1, S1 and S2). Four holes drilled at Site U1385 were spliced to

~~produce a continuous composite section extending to 166.5 metre composite depth (mcd), equivalent to ~1.45 Ma,~~chronology ~~during Marine Isotope Stage (MIS) 47 (Hodell et al., 2015, 2023). The chronology of Site U1385 records was established by aligning the benthic oxygen isotope ($\delta^{18}O$) record to the LR04 reference stack (Hodell et al., 2023; Lisiecki and Raymo, 2005). Chronostratigraphic ages were determined through linear interpolation between age-depth control points. All previously published records from the following sites are presented on their original timescales: Deep Sea Drilling Project (DSDP) Site 607 (Ford et al., 2016; Sosdian and Rosenthal, 2009); Ocean Drilling Program (ODP) Site 1123 (Elderfield et al., 2012); ODP Site 1208 (Ford and Raymo, 2020); ODP Site 1094 (Hasenfratz et al., 2019); MD99-2334 (Skinner et al., 2003), and MD01-2444 (Skinner and Elderfield, 2007). The high-resolution benthic $\delta^{18}O$ Prob-stack record displayed in Figs. 6–11 and S6–S9, represents the probabilistic stack that integrates a globally distributed compilation of benthic $\delta^{18}O$ records (Ahn et al., 2017). The timescale used for the Prob-stack is consistent with that of the LR04 reference (Lisiecki and Raymo, 2005).~~

Four holes drilled at Site U1385 were spliced to produce a continuous composite section extending to 166.5 metre composite depth (mcd), corresponding to a basal age of ~1.45 Ma during Marine Isotope Stage (MIS) 47 (Hodell et al., 2015, 2023). The Site U1385 age model was established by alignment of the benthic oxygen isotope ($\delta^{18}O$) record to the LR04 reference stack (Hodell et al., 2023; Lisiecki and Raymo, 2005), with chronostratigraphic ages assigned by linear interpolation between age-depth control points.

All previously published records used to compile the ocean stacks are presented on their original published age models. Most are based on alignment of benthic foraminiferal $\delta^{18}O$ to LR04, including DSDP Site 607 combined with Chain 82-24-23PC (Ford et al., 2016; Sosdian and Rosenthal, 2009), and Ocean Drilling Program (ODP) Site 1123 (Elderfield et al., 2012). ODP Site 1208 (Ford and Raymo, 2020) ~~Site U1385 benthic Mg/Ca, deep-water temperature and $\delta^{18}O_{seawater}$ records extend back to ~1.45 Ma, spanning MIS 1–47. Mg/Ca measurements from *Uvigerina peregrina* and *Globobulimina affinis* were conducted at ~20 cm intervals, except for two segments sampled at higher resolution: ~4 cm spacing between ~980–860 thousand years ago (ka) over MIS 27–21, and ~5 cm over MIS 40–37 (~1300–1240 ka) (Birner et al., 2016). The corresponding mean sampling intervals are ~1820 years, ~360 years, and ~450 years, respectively. Three noticeable breaks (>15 kyr) occur in the Mg/Ca-derived temperature and $\delta^{18}O_{seawater}$ records during key transitions. Two gaps, related to sampling limitations and low foraminiferal abundance, occur between MIS 6–5 (~132–116 ka), and MIS 16–15 (~630–603 ka). The third is a near 30-kyr hiatus between MIS 12–11 (~430-400 ka) at Termination V (Hodell *et al.*, 2015, 2023). To produce a near continuous Iberian Margin record spanning the past 1.5 Myr, including the MPT (Fig 3; Table.1), Site U1385 records were combined with Mg/Ca-derived temperature and $\delta^{18}O_{seawater}$ data from the Last Glacial Maximum (LGM), measured on *Globobulimina affinis* at nearby Sites MD01-2444 (Skinner and Elderfield, 2007) and MD99-2334 (Skinner et al., 2003). Values from MD99-2334 were adjusted by +0.6 ºC for temperature and +0.15 ‰ for $\delta^{18}O_{seawater}$ to account for its deeper water setting.~~ was aligned to the Prob-stack using the HMM-Stack MATLAB code (Ahn et al., 2017; Ford and Raymo, 2020; Lin et al., 2014: Butcher et al., 2017), which is also based on the LR04 age model. The ODP Site 1094 age model (Hasenfratz et al., 2019) was derived by graphical alignment of benthic $\delta^{18}O$ to LR04, with the original chronology of (Jaccard et al., 2013) retained where benthic foraminifera were sparse (e.g. MIS 4).

The Iberian Margin piston cores MD99-2334 and MD01-2444 were independently aligned to Greenland ice-core records through synchronisation of Dansgaard–Oeschger events recorded in planktonic $\delta^{18}O$ (Shackleton, 2000; Skinner et al., 2003;

Skinner and Elderfield, 2007). Owing to the close agreement in both structure and absolute benthic $\delta^{18}O$ values with Site U1385 over the interval of overlap, no further adjustments to these published age models were applied.

2.2 Stable isotopes

Measurements of benthic foraminiferal $\delta^{18}O$ at Site U1385, paired (same sample) with *Uvigerina peregrina* Mg/Ca analyses to reconstruct $\delta^{18}O_{seawater}$, were predominantly obtained from the same *Uvigerina* where sufficient calcite was available prior to cleaning for Mg/Ca analysis (Barker et al., 2003) (n = 560 representing 64 % of Mg/Ca samples). When *Uvigerina* calcite was unavailable, $\delta^{18}O$ values were substituted with previously measured *Cibicidoides wuellerstorfi* (n = 309) (Hodell et al., 2015, 2023) corrected to *Uvigerina* spp. by an offset of +0.64 ‰ (Hodell et al., 2023; Shackleton et al., 2000). To reconstruct $\delta^{18}O_{seawater}$ using *Globobulimina affinis* Mg/Ca analyses, paired $\delta^{18}O$ were obtained from corrected *Globobulimina affinis* (offset by -0.30 ‰; n = 86), *Cibicidoides wuellerstorfi* (n = 115) and *Uvigerina peregrina* (n = 1). Up to 25 and 53 specimens of *Uvigerina peregrina* and *Globobulimina affinis*, respectively, were selected from the 212–355 µm coarse fraction. Because shell calcite was also used for trace metal determination, tests were crushed between glass slides to facilitate cleaning. When the shell material exceeded ~240 µg, a portion (>80 µg) was separated prior to cleaning and reserved for paired isotope analysis. *Cibicidoidies wuellerstorfi* $\delta^{18}O$ and carbon isotopes ($\delta^{13}C$) have previously been conducted at 1 to 2 cm resolution using 1 to 5 specimens pickedOxygen and carbon isotopes were measured previously using *Cibicidoidies wuellerstorfi* from the >212 µm sediment coarse fraction (Hodell et al., 2015, 2023; Thomas et al., 2022).(Hodell et al., 2023, 2026). Additional analyses were made using *Uvigerina peregrina* and *Globobulimina affinis* selected from the 212–355 µm coarse fraction. All benthic oxygen isotope data were corrected to *Uvigerina* using an offset of 0.64‰ for *C. wuellerstorfi* and -0.3‰ for *G. affinis*. Tests were crushed between glass slides to facilitate cleaning. Isotopic measurements were made at the Godwin Laboratory for Palaeoclimate Research, Department of Earth Sciences, University of Cambridge following previously described methods (Hodell et al., 2015). Instrument precision was better than ±0.08 ‰ (1σ) for $\delta^{18}O$ and ±0.06 ‰ (1σ) for $\delta^{13}C$.

2.3 Foraminiferal Mg/Ca analyses

2.3 Foraminiferal Mg/Ca analyses

We determined Mg/Ca on 911 samples of *Uvigerina peregrina* and 264 samples of *Globobulimina affinis* which were oxidatively cleaned following the procedure of Barker, Greaves and Elderfield (2003). The Mg/Ca-derived temperature records from *Uvigerina peregrina* and *Globobulimina affinis* show good agreement well within error (Fig. 2a) which supports combining them. All Mg/Ca and other trace element ratios were measured using inductively coupled plasma-optical emission spectrophotometry (ICP-OES) at the Godwin Laboratory for Palaeoclimate Research (Cambridge, UK). The majority of analyses were performed on a Varian VISTA instrument (Birner et al., 2016) with the remaining ~180 samples measured using an Agilent 5100 ICP-OES. For both instruments, the intensity ratio calibration method of de Villiers, Greaves and Elderfield (2002) was applied. Interlaboratory comparison studies have validated the use of laboratory standards employed as reference materials for foraminiferal Mg/Ca analyses (Greaves et al., 2008). Repeated measurements of standards showed instrument precision for Mg/Ca determinations to be better than 0.5 % relative standard deviation (r.s.d.) (Greaves, 2008; Greaves et al., 2008; de Villiers et al., 2002). Based on 47 sets of replicate Mg/Ca determinations, including 4 sets of triplicates, the pooled standard deviation was calculated at 0.092 mmol mol^{-1} yielding an overall mean precision of 8.2 % r.s.d. (See Data Repository Table S4).

Benthic Mg/Ca, deep-water temperature and $\delta^{18}O_{seawater}$ records from Site U1385 extend back to ~1.45 Ma, spanning MIS 1–47. Mg/Ca measurements on *Uvigerina peregrina* and *Globobulimina affinis* were conducted at ~20 cm intervals, corresponding to a mean temporal resolution of ~2000 years (yr), except for two intervals sampled at higher resolution: ~4 cm spacing between ~980–860 thousand years ago (ka) over MIS 27–21, and ~5 cm spacing between ~1300–1240 ka (MIS 40–37) (Birner et al., 2016) yielding a temporal resolution of ~500 yr.

Three noticeable breaks (>15 kyr) occur in the Mg/Ca-derived temperature and $\delta^{18}O_{seawater}$ records because of low foraminiferal abundance between MIS 6–5 (~132–116 ka), and MIS 16–15 (~630–603 ka). In addition, a near 30-kyr hiatus removed the interval at Site 339-U1385 between MIS 12–11 (~430-400 ka) at Termination V (Hodell *et al.*, 2015, 2023), which has now been recovered in companion Site 397-U1385 (Hodell et al., 2026). To produce a near-continuous Iberian Margin record spanning the past 1.5 Myr, Site U1385 records were supplemented through the last glacial cycle with Mg/Ca-derived temperature and $\delta^{18}O_{seawater}$ data measured on *Globobulimina affinis* at nearby piston cores MD01-2444 covering 50–35 ka during the early glacial interval (Skinner and Elderfield, 2007), and MD99-2334 spanning 35–10 ka during the late glaciation (Skinner et al., 2003) (Fig. 2; Table.1). Following Skinner et al. (2007) a constant temperature offset of +0.6 °C was applied to Mg/Ca-derived temperatures from the deeper Core MD99-2334 to account for cooler modern deep-water temperature at that site location ((Hodell, 2014; Skinner et al., 2003, 2007; Skinner and Elderfield, 2007); Table 1).

After data quality screening (Sect. 2.4), we report Mg/Ca results for 864 samples of *Uvigerina peregrina* and 202 samples of *Globobulimina affinis*. All samples were oxidatively cleaned following the procedure of Barker, Greaves and Elderfield (2003). Trace elements were measured at the Godwin Laboratory for Palaeoclimate Research (Cambridge, UK) using inductively coupled plasma-optical emission spectroscopy (ICP-OES). Most analyses were performed on a Varian VISTA instrument (Birner et al., 2016) with the remaining ~180 samples measured using an Agilent 5100 ICP-OES. Both instruments used the intensity-ratio calibration method of de Villiers, Greaves and Elderfield (2002). Interlaboratory comparison studies have validated the use of laboratory standards employed as reference materials for foraminiferal Mg/Ca analyses (Greaves et al., 2008). Repeated measurements of standards showed instrument precision for Mg/Ca determinations to be better than 0.5 % relative standard deviation (r.s.d.) (Greaves, 2008; Greaves et al., 2008; de Villiers et al., 2002). Based on long-term standard reproducibility and 47 sets of replicate analyses (including four triplicates) measured on multiple aliquots of the same sample, the pooled standard deviation is ±0.09 mmol mol^{-1} (1σ), corresponding to a mean precision of 8.2 % r.s.d. (See Data Repository Table S4).

Mg/Ca–derived deep-water temperatures from *Uvigerina peregrina* and *Globobulimina affinis* show generally good agreement in both variability and mean values within uncertainty (1.4 ± 1.5 °C and 1.6 ± 1.0 °C, respectively) (Fig. S3a). Both species record similar mean $\delta^{18}O_{seawater}$ values (~ 0.6 ± 0.4 ‰; Fig. S3b). This agreement is consistent with the application of species-specific Mg/Ca–temperature calibrations applied in this study (Elderfield et al., 2006, 2010, 2012a; Weldeab et al., 2016) and justifies combining the two datasets. The individual *Uvigerina peregrina* and *Globobulimina affinis* time series were used as inputs for the PSU Solver calibration procedures described in Sect. 2.5 and combined into single records of deep-water temperature and $\delta^{18}O_{seawater}$ (Fig. 2).

2.4 Data quality evaluation

Mg/Ca ratios derived from foraminiferal calcite tests can be used to estimate ~~palaeo-ocean temperatures~~palaeotemperatures provided that stringent cleaning procedures are followed ~~(Barker et al., 2003). To evaluate the efficiency of these cleaning procedures and to assess the effects of diagenesis~~(Barker et al., 2003; Larsson and Jung, 2025). To evaluate the efficiency of these cleaning procedures and to assess the effects of authigenic precipitation, we analysed element/Ca ratios of co-measured trace elements—primarily Fe and Mn, but also Al, Ba, K, Na, Si, Ti and Zn (Barker et al., 2003; Elderfield et al., 2012; Greaves, 2008; Hasenfratz et al., 2017). Contamination by silicates and clay minerals can lead to significant uncertainty in measured Mg/Ca values. Therefore, the removal of contaminant silicates is ~~the most~~a critical step in the cleaning protocol, and its success can be evaluated through downcore covariation of Mg/Ca and Fe/Ca (or Al/Ca) records (Barker et al., 2003). Mg/Ca records from the two foraminiferal species were assessed separately, as *Globobulimina affinis* is known to exhibit elevated Mg/Ca ratios relative to other benthic foraminiferal species ~~(Skinner et al., 2003; Weldeab et al., 2016)~~(Skinner et al., 2003; Weldeab et al., 2016). At Site U1385~~ we excluded~~, five *Uvigerina peregrina* and three *Globobulimina affinis* samples were excluded from the Mg/Ca dataset following identification of anomalously high co-occurring Fe/Ca and Mg/Ca values (Barker et al., 2003). Sample exclusion was based on co-occurring Fe/Ca–Mg/Ca anomalies rather than a fixed Fe/Mg threshold (Fig. ~~S3~~S4). The ~~remaining~~ Mg/Ca dataset presented here includes *Uvigerina peregrina* samples ~~show~~with Fe/Ca and Fe/Mg values of up to 0.45 mmol mol^{-1} and 0.33 mol mol^{-1}, respectively (~~or~~and *Globobulimina affinis* values up to 0.79 mmol mol^{-1} and 0.28 mol mol^{-1}~~ for~~), ~~*Globobulimina affinis*),~~ which ~~are considerably higher than~~ exceed the typical ~~respective values of <~~Fe/Ca (<0.1 mmol mol^{-1}) and ~~<~~Fe/Mg (<0.03 mol mol^{-1}) values reported by Barker, Greaves and Elderfield (2003) and Elderfield et al. (2012).

Consistent with observations of *Uvigerina* spp. at Site 1123 (Elderfield et al., 2012), Mn/Ca values for *Uvigerina peregrina* at Site U1385 show a slight downcore increase, with lowest glacial values rising from ~0.012 mmol mol^{-1} at 60 ka to ~0.045 mmol mol^{-1} at ~1415 ka (Fig. ~~S4~~S5). Although the maximum Mn/Ca value ~~of~~for *Uvigerina peregrina* at Site U1385 (0.~~28 mmol mol^{-1}~~28 mmol mol^{-1}) is lower than the 0.~~36 mmol mol^{-1}~~36 mmol mol^{-1} reported ~~at~~for Site 1123 (Elderfield et al., 2012)~~ analyses suggest the potential presence of authigenic Mn-Fe-oxide coatings on the foraminiferal calcite. This interpretation is further supported by elevated Mn/Ca ratios measured in *Globobulimina affinis* specimens. However, the~~

~~absence of any positive correlation between Mn/Ca and Mg/Ca suggests that this contamination does not significantly affect Mg/Ca variability. Rather, the elevated Mn/Ca values are more likely attributable to ferromanganese oxide overgrowths (Skinner et al., 2003)~~. Mn/Ca values exceeding ~0.1 mmol mol^{-1} may suggest the presence of authigenic Mn–Fe-oxide coatings on some foraminiferal tests (Elderfield et al., 2012; Hasenfratz et al., 2017, 2019). The incorporation of Mn into calcite tests has been shown to be species-specific (van Dijk et al., 2025). In *Uvigerina peregrina*, Mn/Ca values range from 0.01 to 0.28 mmol mol^{-1}, while in *Globobulimina affinis*, values are generally higher, ranging from 0.04 to 1.88 mmol mol^{-1}, with two anomalous specimens exceeding this upper value. Despite these elevated Mn/Ca values, the absence of a positive correlation between Mn/Ca and Mg/Ca indicates that the presence of Mn–Fe oxides does not significantly influence Mg/Ca variability. We therefore interpret the higher Mn/Ca values as reflecting variable contributions from authigenic Mn–Fe-oxide coatings (Skinner et al., 2003), or from species-specific Mn incorporation during biomineralization (van Dijk et al., 2025).

Diagenetic Mn-Fe-oxide coatings commonly form on foraminiferal shells in deep-sea sediment cores used in palaeoceanographic reconstructions. In oxygenated porewater dissolved manganese is precipitated onto carbonate tests as Mn^{4+} oxides, later becoming remobilized as Mn^{2+} and forming Mn-rich oxide coatings when organic matter undergoes anaerobic decomposition deeper in the sediment (Barker et al., 2003; Boiteau et al., 2012). These Mn-rich coatings can substantially impact bulk Mg/Ca ratios and introduce bias into deep-water temperature reconstructions (Hasenfratz et al., 2017). Site specific studies reveal relatively consistent Mg/Mn ratios in foraminiferal coatings, ranging globally between 0.17–0.32 mol mol^{-1} [see Hasenfratz et al. (2017) Table 1 and references therein]. Regionally the upper end of this range—between 0.26 to 0.32 mol mol^{-1}—is most applicable to the Atlantic Ocean, although individual sites may deviate notably from basin-wide averages (Hasenfratz et al., 2017). These Mg/Mn ratios can be used to correct Mg/Ca ratios associated with the calcite lattice using the following equation (Hasenfratz et al., 2017):

$$\mathrm{Mg/Ca_{corrected}} = \mathrm{Mg/Ca_{measured}} - \left(\mathrm{Mn/Ca_{measured}} \times \mathrm{Mg/Mn_{coating}}\right) \quad (1)$$

To maintain consistency, we applied a ~~0.32 mmol mol^{-1}~~ $\mathrm{Mg/Mn_{coating}}$ correction following Equation 1, with the $\mathrm{Mg/Mn_{coating}}$ term fixed at 0.32 mol mol^{-1}, and applied uniformly to all Mg/Ca data (Hasenfratz et al., 2017; de Lange et al., 1992) (Fig. ~~S5~~S6). This value represents the maximum adjustment in the direction of lower temperatures and yields the least discrepancy between Site U1385 temperatures and those from the Pacific and South Atlantic. The correction has minimal influence on glacial Mg/Ca values.

~~Following~~In total, Mg/Ca was initially determined on 911 *Uvigerina peregrina* and 264 *Globobulimina affinis* samples. After quality control, we ~~report~~retain Mg/Ca results for ~~870~~864 samples of *Uvigerina peregrina* (including replicates) and 202 samples of *Globobulimina affinis*~~, after excluding a further 36~~ measurements, having excluded 42 and 59 samples, respectively, based on co-measured trace element ~~ratios indicative~~indicators of potential ~~Mg/Ca~~ contamination. ~~The~~Samples were rejected according to the following criteria~~ were used for exclusion~~:

- Clay contamination: Low [Ca] (<10 ppm) coupled with high Na/Ca (>5.5 mmol mol^{-1}), and K/Ca (>0.4 mmol mol^{-1}), almost always associated with elevated Mg/Ca.
- Additional indicators of clay influence: Al/Ca > 0.4 mmol mol^{-1}, Si/Ca > 0.4 mmol mol^{-1}, Ti/Ca > 0.005 mmol mol^{-1}, or very low Mn/Fe (< 0.1 mol mol^{-1}) due to elevated Fe/Ca.
- Analytical blank effects: Low [Ca] in combinationcombined with anomalously high Mg/Ca values, suggesting contamination during measurement.

2.5 PSU Solver: Calibration equations for Mg/Ca-derived deep-water temperature and $\delta^{18}O_{seawater}$

To calculate deep-water temperature and $\delta^{18}O_{seawater}$ estimates for Site U1385 over the past 1.5 Ma we used the Paleo-Seawater Uncertainty Solver (PSU Solver) MATLAB code (Thirumalai et al., 2016). This approach applies bootstrap Monte Carlo simulations (n = 1000) to propagate and constrain uncertainty across multiple proxies. Required input data include foraminiferal Mg/Ca and benthic $\delta^{18}O$ measurements with their respective 2 sigma (σ) analytical uncertainties, and species-specific Mg/Ca–temperature calibrations (Tables 2 and S1). Age model uncertainty was set at ±2 kyr, which represents both the optimal resolution for combining the records of differing resolution and the mean temporal resolution of all five records.

Deep-water temperature and $\delta^{18}O_{seawater}$ estimates were calculated for Site U1385 over the past 1.5 Ma using an updated Python version of Paleo-Seawater Uncertainty Solver (PSU Solver) originally developed in MATLAB (Thirumalai et al., 2016). Uncertainties associated with Mg/Ca–temperature calibration slopes and intercepts were propagated explicitly within a Monte Carlo framework by sampling calibration slopes and intercepts within their reported uncertainties during each iteration (n = 1000). Input data comprise foraminiferal Mg/Ca and benthic $\delta^{18}O$ measurements. Standard laboratory uncertainties of 0.08 ‰ for $\delta^{18}O$ and 0.05 mmol mol^{-1} for Mg/Ca were applied as inputs to the PSU Solver (Thirumalai et al., 2016). Species-specific Mg/Ca–temperature calibrations were employed within the PSU Solver framework (Table 2).

For *Uvigerina peregrina,* we used the linear Mg/Ca–temperature calibration of Elderfield et al. (2012):

$$\mathrm{Mg/Ca} = 1 + (0.1 \times \mathrm{T\ °C})$$

$$[\mathrm{mmol\ mol^{-1}}] = 1.0 + (0.1 \pm 0.013 \times \mathrm{T\ [°C]}) \quad (2)$$

Core-top Mg/Ca-derived temperatures (~2.5 °C; Figs. 2 and S3) are consistent with modern deep-water temperatures (~2–3 °C) at the Iberian Margin (Hodell et al., 2014; Skinner et al., 2003), supporting the applicability of the Elderfield et al. (2012) Mg/Ca–temperature calibration at this site.

For *Globobulimina affinis*, we applied the calibration of Weldeab, Arce and Kasten (2016):

$$Mg/Ca\ [mmol\ mol^{-1}] = 2.22 \pm 0.19 + (0.36 \times \pm 0.02 \times T\ °C)\ [°C]) \qquad (3)$$

To reconstruct $\delta^{18}O_{seawater}$ we used the oxygen-isotope palaeothermometry equation (Elderfield et al., 2010, 2012):

$$T = [°C] = 16.9 - 4.0 \times (\delta^{18}OO_{calcite}\ [‰] - \delta^{18}O_{seawater} + [‰] + 0.27) \qquad (4)$$

Analytical uncertainty for Mg/Ca at Site U1385 was calculated at ±0.185 mmol mol^{-1} (2σ), derived from the combined long-term standard error and pooled replicate error as reported in Elderfield et al. (2012) and Birner et al. (2016) (Tables 2 and S1). The analytical uncertainty for benthic $\delta^{18}O$ is ±0.16 ‰ (2σ).

Age model uncertainty was set at ±2 kyr, which represents both the optimal resolution for combining the records of differing resolution and the mean resolution of the five records.

Table 2: PSU Solver input data for 2σ error in analytical uncertaintySpecies-specific Mg/Ca–temperature calibrations.

Site	**Analytical Uncertainty (2σ)Species**	**Species specific Mg/Ca (mmol mol^{-1}) to temperature (°C) PSU eq 1**	**Lab; Citation**	
SiteIODP U1385	**Mg/Ca (mmol mol^{-1}) (long-term std + replicate error)***Uvigerina peregrina*	**δ^{18}O (‰)**$Mg/Ca = 1 + (0.1 \pm 0.013 \times T)$ Elderfield et al., 2012	Cambridge: This study (Birner et al., 2016)**Species specific Mg/Ca to temperature PSU eq 1**	**Lab; Citation**
IODP U1385	0.185 (Uvig)	0.16	$Mg/Ca = 1 + (0.1 \times T)$; Elderfield et al., 2012	Cambridge: This study (Birner et al., 2016)
IODP U1385	*Globobulimina affinis*	$Mg/Ca = 2.22 \pm 0.19 + (0.36 \pm 0.02 \times T)$ Weldeab et al., 2016	Cambridge: This study	
MD99-2334IODP U1385	0.185 (G. aff)*Globobulimina affinis* 0.16	$Mg/Ca = 2.22 \pm 0.19 + (0.36 \times T)$; $(0.36 \pm 0.02 \times T)$ Weldeab et al., 2016	Cambridge: This study(Skinner et al., 2003)	

Deleted Cells

Deleted Cells

MD01-2444~~**MD99-2334**~~	~~0.14 (G. aff)~~*Globobulimina affinis*	~~0.16~~	Mg/Ca = 2.22 ± 0.19 + ~~(0.36 × (T + 0.6))~~; (0.36 ± 0.02 × T) Weldeab et al., 2016 ~~(modified for water depth)~~	Cambridge: (Skinner et al., ~~2003)~~ -2007)
~~**MD01-2444**~~**ODP 1123**	~~0.14 (G. aff)~~*Uvigerina* spp.	~~0.16~~	Mg/Ca ~~= 2.22~~ = 1 + ~~(0.36 × T)~~; (0.1 ± 0.013 × T) ~~Weldeab~~Elderfield et al., ~~2016~~2012	Cambridge: (~~Skinner~~Elderfield et al., ~~2007~~2012)
~~**ODP 1123**~~	~~0.111 (Uvig)~~	~~0.16~~	~~Mg/Ca = 1 + (0.1 × T); Elderfield et al., 2012~~	~~Cambridge: (Elderfield et al., 2012)~~
DSDP 607 + Piston Core Chain 82-24-23PC	*Uvigerina* spp.		Mg/Ca = 0.9 + (0.1 ± 0.013 × T) Elderfield et al., 2012 (modified for reductive cleaning)	LDEO & Rutgers: (Ford et al., 2016)
~~**DSDP 607 + Piston Core Chain 82-24-23PC**~~	~~0.122 (Uvig):~~*Cibicidoides wuellerstorfi*	~~0.12~~	Mg/Ca ~~=~~ = 1.043 ± 0.9~~ + (0.1 × T);~~03 × e^(0.118 ± 0.1 × T) ~~Elderfield~~Barrientos et al., ~~2012 (modified for reductive cleaning)~~2018	LDEO & Rutgers: ~~(Ford et al., 2016)~~(Sosdian and Rosenthal, 2009)
	~~0.202 (Cibs & Orid)~~*Oridorsalis umbonatus*	~~0.12~~	Mg/Ca = 1.~~16 + (0.15 × T)~~;317 ± 0.03 × e^(0.102 ± 0.01 × T) ~~Sosdian and Rosenthal, 2009~~Barrientos et al., 2018	LDEO & Rutgers: (Sosdian and Rosenthal, 2009)
ODP 1208	~~0.115 (Uvig)~~*Uvigerina* spp.	~~0.12~~	Mg/Ca = 0.9 + ~~(0.1 × T)~~; (0.1 ± 0.013 × T) Elderfield et al., 2012 (modified for reductive cleaning)	LDEO & Rutgers: (Ford and Raymo, 2020)
ODP 1094	~~0.084 (Mpomp)~~*Melonis pompilioides*	~~0.14~~	Mg/Ca = 0.742 + ~~(0.119 × T)~~; (0.119 ± 0.003 × T) Hasenfratz et al., 2017	ETHZ & Cambridge: (Hasenfratz et al., 2019)

*Abbreviations: LDEO, Lamont-Doherty Earth Observatory (Columbia University, New York, USA); ETHZ, Geological Institute, ETH Zurich (Zurich, Switzerland).

2.6 Stacking temperature and $\delta^{18}O_{seawater}$ records

The rationale for stacking records of temperature and $\delta^{18}O_{seawater}$ is the same as that given for oxygen isotope records (Imbrie et al., 1984; Lisiecki and Raymo, 2005); that is, to improve the signal/noise ratio and Stacked deep-water temperature and $\delta^{18}O_{seawater}$ records were produced for the North Atlantic using PSU Solver estimates for Iberian Margin Sites U1385 (this study; Birner *et al.*, 2016), MD01-2444 (Skinner and Elderfield, 2007), and MD99-2334 (Skinner et al., 2003) along with Site 607 (Ford et al., 2016; Sosdian and Rosenthal, 2009). For the deep Pacific stack we used the PSU Solver output for Sites 1123 (Elderfield et al., 2012) and 1208 (Ford and Raymo, 2020) (Figs. S6 and S7). The global stacked records of mean deep ocean temperature (MDOT) and mean deep ocean $\delta^{18}O_{seawater}$ (Figs. 4a and 5a respectively) include all the above along with ODP Site 1094 located in the South Atlantic section of the Southern Ocean (Hasenfratz et al., 2019) (Figs 1, S1, S2 and S8). The PSU Solver-derived temperature and $\delta^{18}O_{seawater}$ estimates and their associated standard deviations were interpolated at a regular 3 kyr interval using MATLAB's `interp` function. These interpolated values were then bootstrapped (n = 1000) by randomly subsampling with a normal probability distribution using the calculated standard deviations to produce error estimates (Figs. 4 and 5). Stacks were manually constructed by first identifying gaps in each site's original data. Mean values and associated uncertainties were then calculated across each aligned age interval. Where age gaps exceeded ~6 kyr the interpolated and bootstrapped values were excluded from the stack alignment process.

2.7 Weighted stacks

The weighted global benthic $\delta^{18}O$, MDOT and $\delta^{18}O_{seawater}$ stacks were iteratively constructed using ocean basin volume-weighting percentages from Lisiecki and Stern (2016) (see their Table S2). At each age point, the percentage of total ocean volume represented by available values from each basin was assessed. Initially each basin was weighted according to its proportional contribution to total global ocean volume, with the North and South Atlantic and Pacific Oceans together comprising 43.6% (Table S2). However, because just over half of the global data points (n = 268) included contributions from all three basins, volume weighting was dynamically adjusted for the remaining data to accurately reflect the proportional volume each sample contributed to the total volume represented at that time step.

All stacks were smoothed using the LOWESS detrending/curve-fitting function in Acycle v2.7.1 (MATLAB) with a window of 16 kyr.

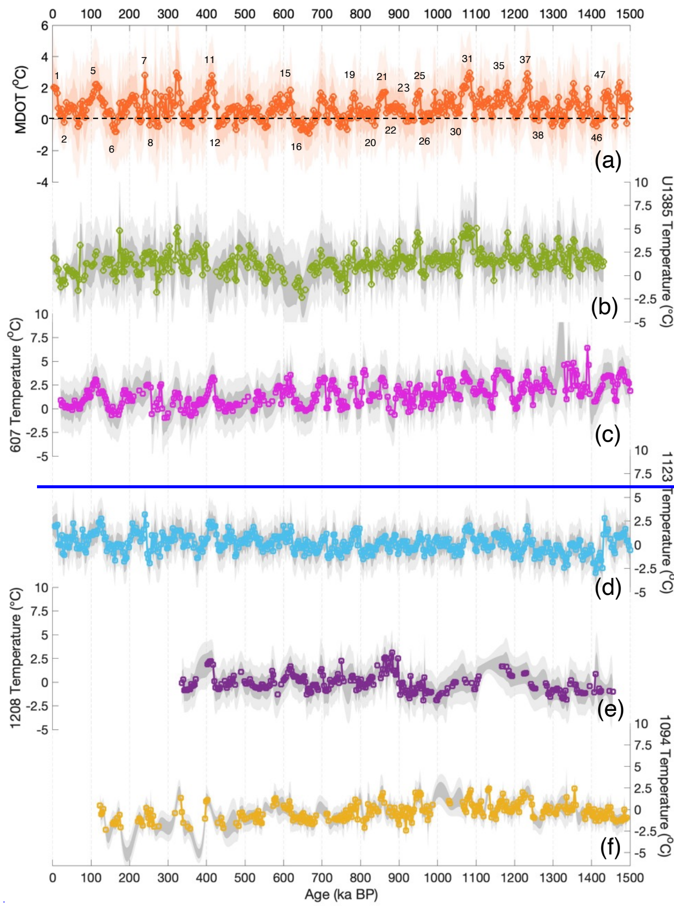


Figure 4: Interpolated and bootstrapped Mg/Ca-derived temperature records for all sites. (a) Mean deep ocean temperature stack (MDOT) (this study; orange line and markers) with MIS numbers denoting glacial (below) and interglacial (above) stages. Mg/Ca-derived temperature records for: (b) IODP Site U1385 (this study and including other Iberian Margin records (Birner et al., 2016; Skinner et al., 2003; Skinner and Elderfield, 2007)) (light green); (c) DSDP Site 607 (Ford et al., 2016; Sosdian and Rosenthal, 2009) (light pink); (d) ODP Site 1123 (Elderfield et al., 2012) (light blue); (e) ODP Site 1208 (Ford and Raymo, 2020) (dark purple); and (f) ODP Site 1094 (Hasenfratz et al., 2019) (yellow). Mg/Ca-derived temperature records were interpolated on a 3 kyr interval and bootstrapped. Shading represents 1 and 2σ error margins (dark and light grey respectively).

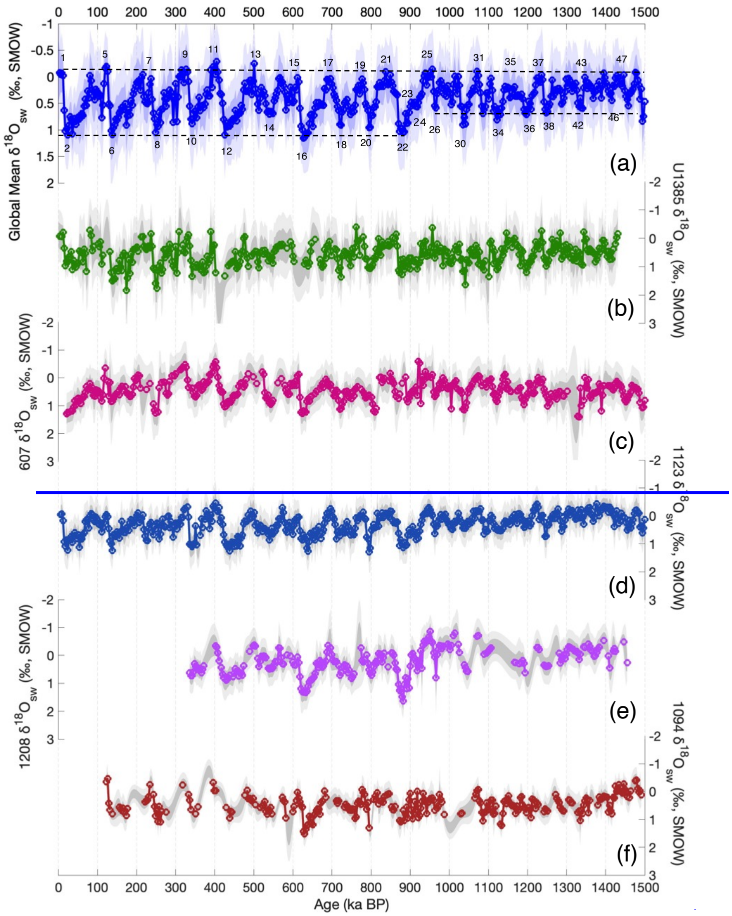


Figure 5: All sites interpolated and bootstrapped $\delta^{18}O_{seawater}$ records. (a) Global mean $\delta^{18}O_{seawater}$ stack (this study; blue line and markers) with MIS numbers denoting glacial stages (below) and interglacial stages (above). Interpolated and bootstrapped $\delta^{18}O_{seawater}$ records for: (b) IODP Site U1385 (this study and including (Birner et al., 2016; Skinner et al., 2003; Skinner and Elderfield, 2007)) (dark green); (c) DSDP Site 607 (Ford et al., 2016; Sosdian and Rosenthal, 2009) (dark pink); (d) ODP Site 1123 (Elderfield et al., 2012) (dark blue); (e)reduce the influence of local variability. Although stacked records provide more representative regional and/or global mean signals of temperature and ice volume they also tend to dampen the amplitude of the signal owing to misalignments and smoothing. Stacked deep-water temperature and $\delta^{18}O_{seawater}$ records were constructed for the North Atlantic using Sites U1385 (this study; Birner et al., 2016), MD01-2444 (Skinner and Elderfield, 2007), MD99-

2334 (Skinner et al., 2003) and Site 607 ODP Site 1208 (Ford and Raymo, 2020) **(purple); and (f) ODP Site 1094 (Hasenfratz et al., 2019) (dark red). All δ^{18}O$_{seawater}$ records were interpolated on a 3 kyr interval and bootstrapped. Shading represents 1 and 2σ error margins (dark and light grey respectively).**

3 Results

3.1 Comparison of Sites U1385 and 607 deep-water records

North Atlantic Site U1385 records confirm there are regional differences between the North Atlantic and South and North Pacific. Previously, the different temperature and δ^{18}O$_{seawater}$ signals at North Atlantic Site 607 and Southwest Pacific Site 1123 were attributed to the uncertainty in the reliability of the multispecies Mg/Ca-derived temperature record of Site 607 (Elderfield et al., 2012; Sosdian and Rosenthal, 2009, 2010; Yu and Broecker, Wallace, 2010); the high-resolution *Uvigerina* spp. record of Site 1123 was therefore considered more robust and a more global signal. However, the Site U1385 and North Pacific Site 1208 records (Ford and Raymo, 2020), both constructed using *Uvigerina* spp., confirm there are regional differences and no one site is a completely global representation of temperature and δ^{18}O$_{seawater}$ (ice volume) change over the MPT (Figs. 4, 5, S6 and S7).

North Atlantic Sites U1385 and 607 temperature and δ^{18}O$_{seawater}$ records show strong covariations in amplitude and frequency over the past 1.5 Ma (Figs. 6b and 7b respectively). There are however a few notable differences between the North Atlantic Mg/Ca-derived temperature records. Greater cooling is expressed at Site 607 which cools by ~0.9 ºC between 1500–900 ka and ~2.2 ºC over the past 1.5 Ma compared to respective cooling at Site U1385 of ~0.4 ºC and 1.0 ºC. At times divergence between the two North Atlantic temperature records (e.g. between 1180–1150 ka and ~400–350 ka) corresponds to temperatures derived from *Cibicidoides wuellerstorfi* and *Oridorsalis umbonatus* at Site 607. However, these differences cannot be solely attributed to a carbonate ion effect on the epifaunal species, as between ~1400–1300 ka, temperature values derived from *Uvigerina* spp. at Site 607 are anomalously high compared to data from other species at both sites (Ford et al., 2016; Sosdian and Rosenthal, 2009). In general, the close correspondenceThe deep Pacific stack was derived using Sites 1123 (Elderfield et al., 2012) and 1208 (Ford and Raymo, 2020) (Figs. S7 and S8). Globally distributed mean deep ocean temperature (MDOT) and mean deep ocean δ^{18}O$_{seawater}$ stacks (Figs. 3a and 4a respectively) incorporate these records together with ODP Site 1094 located in the South Atlantic sector of the Southern Ocean (Hasenfratz et al., 2019) (Fig. S9).

Prior to stacking, temperature and δ^{18}O$_{seawater}$ estimates were interpolated to a regular 3 kyr time step using MATLAB's `interp` function. Uncertainty envelopes were estimated using bootstrap resampling (n = 1000), assuming normally distributed errors defined by the reported standard deviations (1σ) (Figs. 3 and 4). Gaps in the original age models were identified, and interpolated values were excluded where age gaps exceeded ~6 kyr. At each interpolated time step, stacked estimates were calculated using either arithmetic or volumetrically weighted averaging of available basin-scale records. Uncertainties in non-weighted stacks were estimated by propagating the combined calibration and analytical errors in quadrature and incorporating inter-record variance, such that the uncertainty of the mean reflects both measurement uncertainty and the spread between paired estimates. For weighted stacks, basin weights were defined using fixed deep ocean volume fractions following the volumetric approach of Lisiecki and Stern (2016) (see their Table S2) and renormalised to unity at each time step to reflect the ocean volume represented by the available records. This ensures that stacked estimates reflect the relative deep ocean volume of the contributing basins without introducing bias when one or more basins are absent. Only half of the stacked data points (n = 268) include contributions from all three basins (North Atlantic, Southern Ocean and Pacific), together representing 43.6% of the global ocean volume (Table S1). The impact of data gaps in the Site 1094 record appears to exert little influence on the weighted MDOT and mean $\delta^{18}O_{seawater}$ reconstructions (Fig. S10).

At each time step t, volumetrically weighted stacked estimates were calculated as

$$x(t) = \sum_{i=1}^{N(t)} w_i(t)\, x_i(t), \tag{5}$$

where $x_i(t)$ is the basin-scale estimate (temperature or $\delta^{18}O_{seawater}$) and $N(t)$ is the number of available basins. Weights were defined as

$$w_i(t) = \frac{V_i}{\sum_{j=1}^{N(t)} V_j}, \tag{6}$$

where, V_i denotes the deep ocean volume fraction represented by basin i, and the denominator represents the total volume of all basins available at time t.

Total uncertainty was estimated by propagating intra-basin variance and incorporating an inter-basin variance term that captures the spread of basin-scale estimates about the weighted mean:

$$\sigma^2(t) = \sum_{i=1}^{N(t)} w_i^2(t)\, \sigma_i^2(t) + \sum_{i=1}^{N(t)} w_i(t)\left(x_i(t) - x(t)\right)^2, \tag{7}$$

where $\sigma_i(t)$ represents the reported 1σ uncertainty of basin i, and the second term represents inter-basin variability. Where only a single basin contributed to the stack, the uncertainty reduces to the reported basin 1σ value. This approach assumes that residual chronological misalignment between records is small relative to the interpolated 3 kyr resolution. Uncertainty envelopes incorporate analytical, calibration and age uncertainties propagated through a Monte Carlo framework, with basin-scale uncertainty additionally reflecting inter-site variability.

Change-point analysis was applied to the interpolated stacks (3 kyr resolution) using MATLAB's `findchangepts` function to identify statistically significant shifts in the mean state of the record. The analysis detects a primary breakpoint at ~930 ka in both mean averaged and basin-volume weighted $\delta^{18}O_{seawater}$ stacks, marking the most prominent transition in the time series.

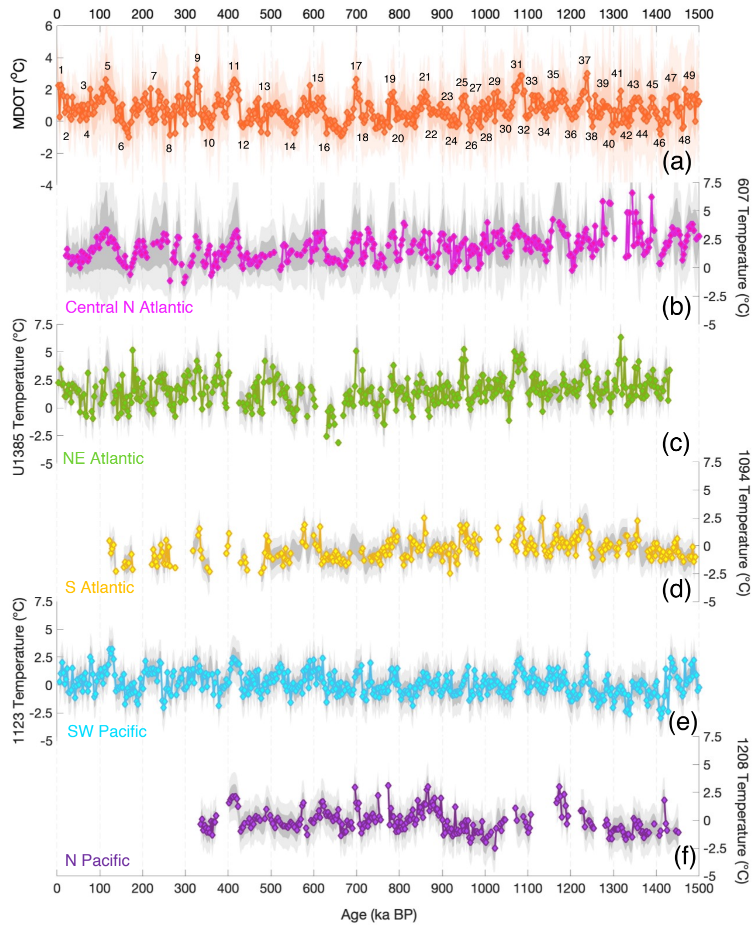


Figure 3: Interpolated and bootstrapped Mg/Ca-derived temperature records for all sites. (a) Mean deep ocean temperature stack (MDOT; this study, orange line and markers), with MIS numbers denoting glacial (below) and interglacial (above) stages. Mg/Ca-derived temperature records for: (b) DSDP Site 607 (Ford et al., 2016; Sosdian and Rosenthal, 2009) (light pink); (c) IODP Site U1385 (this study, including additional Iberian Margin records (Birner et al., 2016; Skinner et al., 2003; Skinner and Elderfield, 2007) (light green); (d) ODP Site 1094 (Hasenfratz et al., 2019) (yellow); (e) ODP Site 1123 (Elderfield et al., 2012) (light blue); and (f) ODP Site 1208 (Ford and Raymo, 2020) (dark purple). Mg/Ca-derived temperature records were interpolated on a 3 kyr interval

and subsequently bootstrapped. Shading represents 1 and 2σ error margins (dark and light grey respectively), propagated from both analytical and calibration sources.

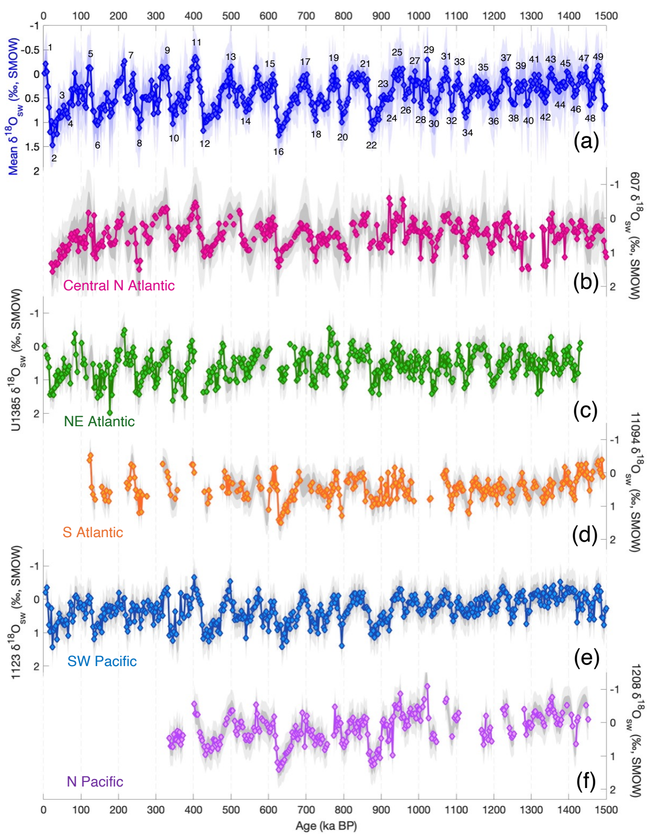


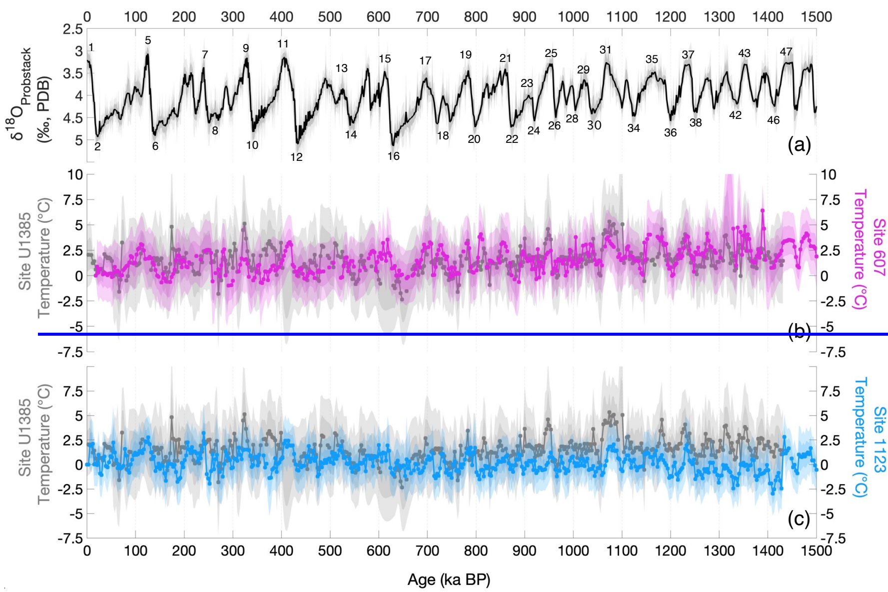
Figure 4: Interpolated and bootstrapped $\delta^{18}O_{seawater}$ records for all sites. (a) Mean $\delta^{18}O_{seawater}$ stack (this study; blue line and markers) with MIS numbers denoting glacial stages (below) and interglacial stages (above). Interpolated and bootstrapped $\delta^{18}O_{seawater}$ records for: (b) DSDP Site 607 (Ford et al., 2016; Sosdian and Rosenthal, 2009) (pink); (c) IODP Site U1385 (this study, including additional Iberian Margin records (Birner et al., 2016; Skinner et al., 2003; Skinner and Elderfield, 2007) (green); (d)

ODP Site 1094 (Hasenfratz et al., 2019) (orange); (e) ODP Site 1123 (Elderfield et al., 2012) (blue); and (f) ODP Site 1208 (Ford and Raymo, 2020) (purple). All $\delta^{18}O_{seawater}$ records were interpolated to a 3 kyr interval and subsequently bootstrapped. Shading denotes 1 and 2σ error margins (dark and light grey respectively), propagated from both analytical and calibration uncertainties.

3 Results

3.1 Comparison of Sites U1385 and 607 deep-water records

North Atlantic Sites U1385 and 607 temperature and $\delta^{18}O_{seawater}$ records generally show similar values and trends over the past 1.5 Myr (Figs. 5b and 6b), but there are some differences in the magnitude of the Mg/Ca-derived temperature changes. Greater cooling is expressed at Site 607 where temperatures show decreasing trends of ~1.0 °C between 1500–900 ka and ~1.7 °C over the past 1.5 Myr compared to Site U1385 which shows less cooling. At times, differences between the two temperature records (e.g. between 1180–1150 ka and ~400–350 ka) correspond to Mg/Ca measurements of *Cibicidoides wuellerstorfi* at Site 607, which is more prone to carbonate ion changes and dissolution than infaunal benthic taxa. However, the cooler temperatures at Site 607 cannot be attributed solely to a carbonate ion effect because *Uvigerina* spp. also give high temperatures between ~1400–1300 ka (Ford et al., 2016; Sosdian and Rosenthal, 2009) and may instead reflect hydrographic differences between the eastern and western North Atlantic basins (Chalk et al., 2019). Despite the subtle differences in the records, the close similarity in both temperature and $\delta^{18}O_{seawater}$ data from the two sites supports their integration into a North Atlantic stack. The magnitude of cooling is also supported by a decrease in alkenone SST in the high-latitude of the North Atlantic where deep water is formed (Lawrence et al., 2009).


δ18O Probstack (‰, PDB)
Site U1385 Temperature (°C)
Site 607 Temperature (°C)
Site 1123 Temperature (°C)
(a)
(b)
(c)
Age (ka BP)

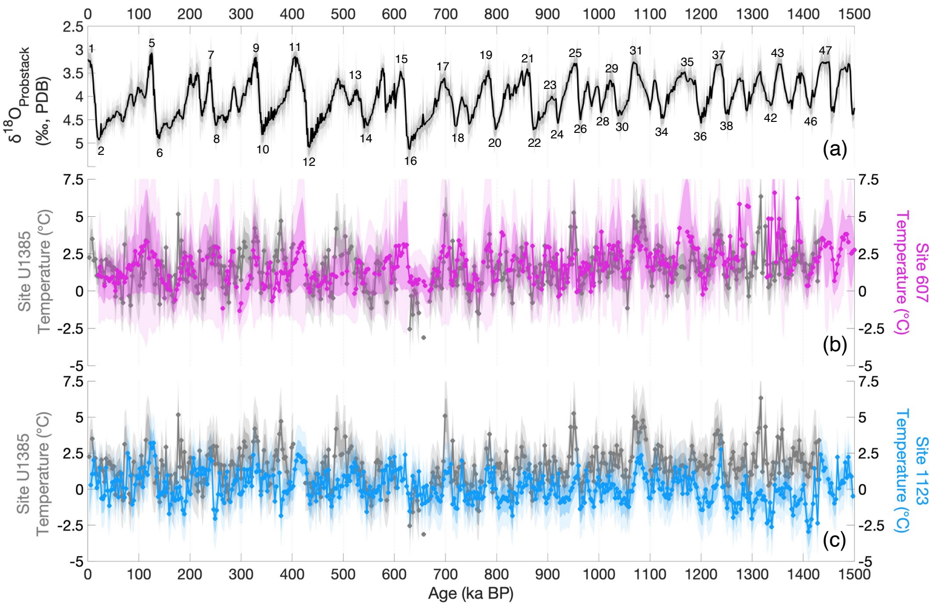


Figure ~~6~~5: Comparison of Mg/Ca-derived deep-water temperature from Site U1385 versus Sites 607 and 1123. (a) Prob-stack benthic δ18O (Ahn et al., 2017) (black line~~).~~), with MIS numbers ~~denote glacial stages~~ denoting glacials (below) and ~~interglacial stages~~ interglacials (above~~.~~). Mg/Ca-derived deep-water ~~temperature comparisons between~~temperatures from Site U1385 (grey) ~~against:~~are compared with (b) Site 607 (Ford et al., 2016; Sosdian and Rosenthal, 2009) (pink), and (c) Site 1123 (Elderfield et al., 2012) (blue). ~~Whereas the North Atlantic~~ Site 607 ~~reports~~exhibits a gradual ~~decrease in deep-water temperature (of ~1–3 °C)~~ cooling through time (Ford et al., 2016; Sosdian and Rosenthal, 2009), ~~the Southwestern Pacific~~whereas Site 1123 documents relatively stable, near-freezing~~,~~ temperatures during glacial maxima ~~temperatures~~ (Elderfield et al., 2012). ~~Same colour dark~~Dark and light shading ~~represents~~represent 1 and 2σ ~~error margins for each parameter~~uncertainties, respectively~~.~~, estimated by propagating analytical and calibration uncertainties in quadrature, and accounting for inter-record variance. Records are interpolated ~~on a~~to 3 kyr ~~interval~~resolution and bootstrapped.

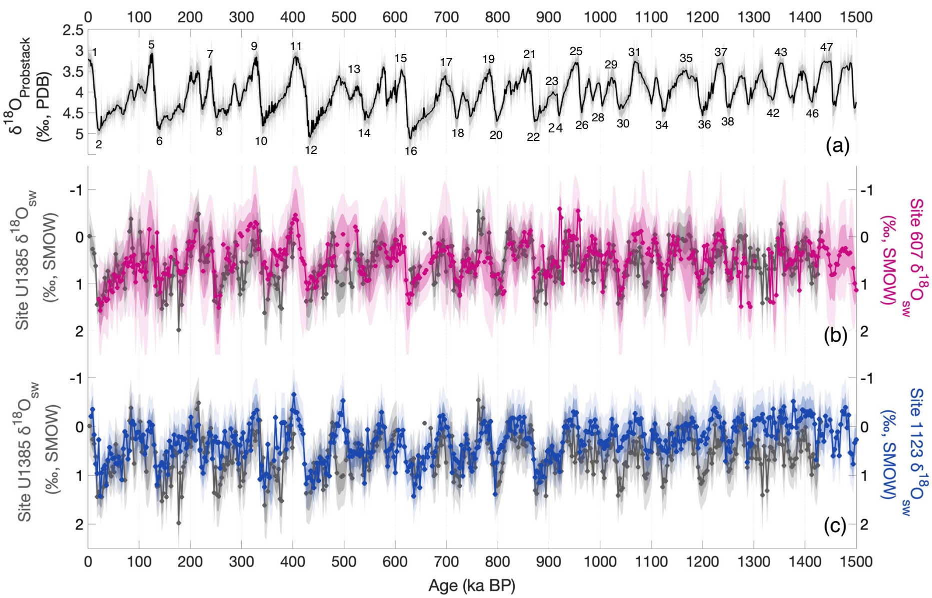
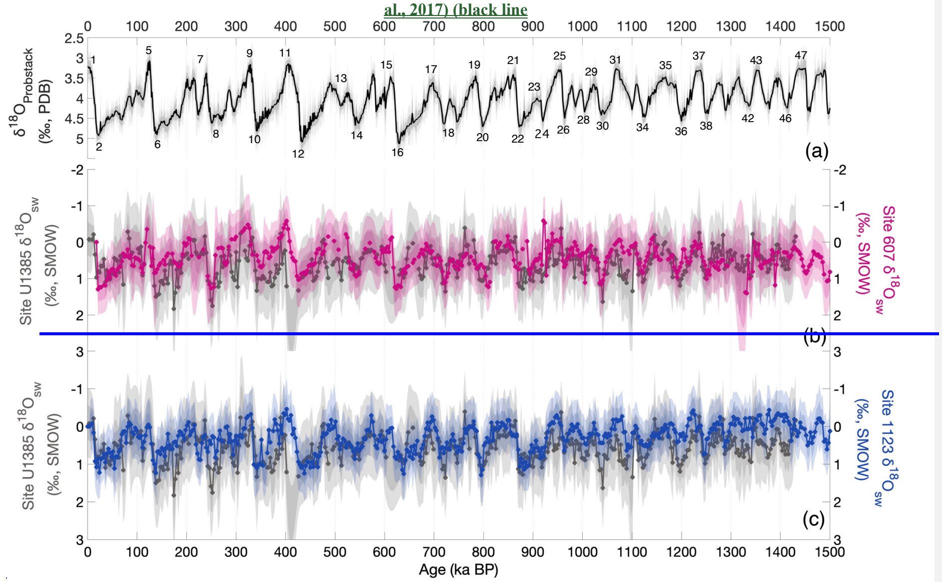

δ18O Probstack (‰, PDB)
Site U1385 δ18O sw (‰, SMOW)
Site 607 δ18O sw (‰, SMOW)
Site 1123 δ18O sw (‰, SMOW)
(a)
(b)
(c)
Age (ka BP)

Figure 6: Comparison of benthic $\delta^{18}O_{seawater}$ records from Site U1385 with Sites 607 and 1123. (a) Prob-stack benthic $\delta^{18}O$ (Ahn et al., 2017) (black line)



~~**Figure 7: Comparison of benthic $\delta^{18}O_{seawater}$ records between Site U1385 versus Sites 607 and 1123. (a)**~~**), with MIS numbers denoting glacial (even) and interglacial (odd) stages. Deep-water $\delta^{18}O_{seawater}$ from Site U1385 (dark grey) is compared with** ~~**Prob-stack benthic $\delta^{18}O$ (Ahn et al., 2017) (black line). MIS numbers denote glacial (even) and interglacial (odd) stages. Deep-water $\delta^{18}O_{seawater}$ comparisons between Site U1385 (dark grey) versus:**~~ **(b) Site 607 (Ford et al., 2016; Sosdian and Rosenthal, 2009) (dark pink), and (c) Site 1123 (Elderfield et al., 2012) (dark blue).** ~~**Interpolated on**~~**Records are interpolated at a 3 kyr** ~~**intervals**~~**interval and bootstrapped** ~~**with same colour dark**~~**. Dark and light shading** ~~**representing 1**~~**indicates 1σ and 2σ** ~~**error margins**~~**uncertainty envelopes, respectively** ~~**for each parameter.**~~**. No clear long-term trend in $\delta^{18}O_{seawater}$ (ice volume/salinity) is observed at Site 607 (Ford et al., 2016; Sosdian and Rosenthal, 2009)** ~~**but**~~**, whereas Site 1123** ~~**reports**~~**records an abrupt increase in Antarctic ice volume (salinity) ~900 ka (Elderfield et al., 2012).**

3.2 ~~Stacked records of~~Deep Atlantic-Pacific gradients in deep-water temperature and $\delta^{18}O_{seawater}$

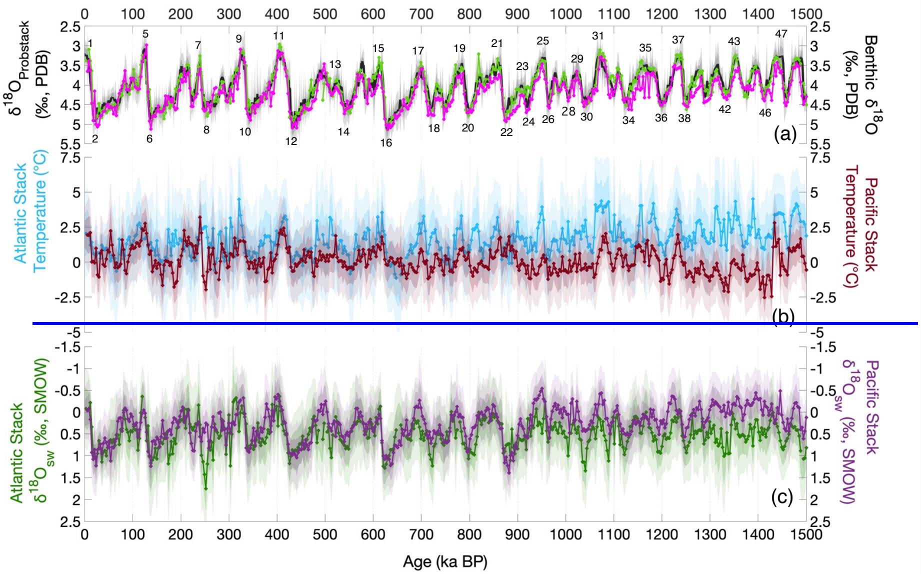
Whereas the benthic $\delta^{18}O_{calcite}$ records are similar between the Atlantic and Pacific (Fig. 7a), the deconvolved temperature and $\delta^{18}O_{seawater}$ records are not. Comparisons between Atlantic and Pacific deep-water temperature and $\delta^{18}O_{seawater}$ stacks show greater inter-basin differences before ~930 ka than ~~after ~900 ka.~~afterwards. From 1500 to ~~900~~930 ka (MIS 49 to ~~22~~24), North Atlantic deep-water was on average ~2.5 ~~°C~~± 1.5 °C warmer than the deep Pacific during both glacial and interglacial stages~~.~~ ~~Deep Pacific temperatures approach the freezing point of seawater (~-2 °C) during glacials before and after ~900~~ (Fig. 7b).

After ~930 ka, ~~whereas~~ this temperature difference decreases to ~1.1 ± 1.4 °C primarily reflecting cooling of glacial ~~temperatures in the deep~~ North Atlantic ~~cooled by more than 1 °C (Fig. 8b).~~

~~Deep~~ deep-water while deep Pacific temperatures show only minor changes. Prior to 930 ka, deep ocean glacial $\delta^{18}O_{seawater}$ values in the Pacific were distinctly less than those of the Atlantic. Across the MPT, deep ocean glacial $\delta^{18}O_{seawater}$ increased in both ~~oceans across the MPT reflecting increased glacial ice volume; however,~~ocean basins but the deep Pacific values increased ~~substantially~~ more than those of the deep Atlantic ~~at ~900 ka~~ (Fig. ~~8c). Thereafter~~7c). After 930 ka, glacial deep-water temperature and $\delta^{18}O_{seawater}$ values from the Atlantic and Pacific were more ~~like one another. We interpret the greater increase in deep Pacific $\delta^{18}O_{seawater}$ as indicating an increase in salinity, in addition to the increase from expansion of continental ice sheets.~~similar.

Deep-water temperatures at Site 1094 in the Atlantic sector of the Southern Ocean (Hasenfratz et al., 2019) are similar to those of the deep Pacific (Fig. S9), but slightly warmer consistent with the modern temperature difference between the two sites. No substantial change in glacial temperatures occurred at Site 1094 across the MPT (Fig. S9). The $\delta^{18}O$ values of seawater at Site 1094 are more similar to those of the deep North Atlantic record. Prior to the MPT, a gradient in $\delta^{18}O_{seawater}$ existed between Site 1094 and the deep Pacific, whereas the values converge after 930 ka.

All five records were stacked to produce global signals of deep ocean temperature (>2500 m) and $\delta^{18}O_{seawater}$ change. The MDOT stack (Fig. ~~S9c~~S11c) shows a modest gradual decrease in glacial deep-water temperatures ~~at~~from ~1050 to 925 ka primarily ~~due~~related to cooling in the deep North Atlantic. ~~In contrast~~At ~930 ka, the global mean $\delta^{18}O_{seawater}$ stack (Fig. ~~S9b) reflects an increase in continental ice volume and salinity ~900 ka. These changes are~~S11b) increases abruptly. This change is more pronounced in ocean volume-weighted records because of ~~a distinct warming~~ the proportionally greater influence of Pacific deep ocean temperature~~.~~, which shows a warming from MIS 24 through 22 (Figs. ~~9 and S10; Sect. 2.7~~8 and S11). Mean glacial $\delta^{18}O_{seawater}$ values in the ocean volume-weighted stack increase from 0.38 ± 0.08 ‰ prior to ~930 ka to 0.80 ± 0.11 ‰ thereafter, representing a mean difference of ~0.42 ‰ that is statistically significant ($p < 0.001$).


δ18O Probstack (‰, PDB)
Benthic δ18O (‰, PDB)
(a)
Atlantic Stack Temperature (°C)
Pacific Stack Temperature (°C)
(b)
Atlantic Stack δ18Osw (‰, SMOW)
Pacific Stack δ18Osw (‰, SMOW)
(c)
Age (ka BP)

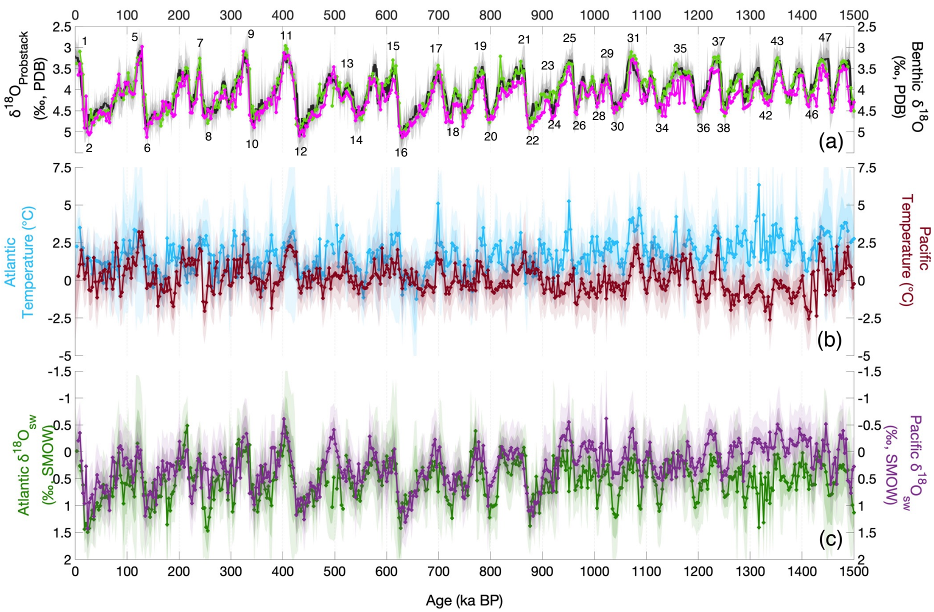


Figure ~~8~~7: Comparison of North Atlantic and Pacific Ocean deep-water temperature and benthic $\delta^{18}O_{seawater}$ stacks. (a) Prob-stack benthic $\delta^{18}O$ (Ahn et al., 2017) (black line~~)~~), overlain with contributing benthic $\delta^{18}O$ records ~~contributing to~~for the Atlantic (this study and (Ford et al., 2016; Sosdian and Rosenthal, 2009)) (bright green) and Pacific (Elderfield et al., 2012; Ford and Raymo, 2020) (bright pink) stacks in (b and c~~.~~). MIS numbers denote glacial (even) and interglacial (odd) stages. (b) ~~Deep~~Stacked deep-water ~~temperature stacks~~temperatures for the Pacific (red) and North Atlantic (blue) with ~~1 and 2σ error envelopes shown in same colour~~ dark and light shading indicating 1 and 2σ uncertainties respectively. (c) ~~Deep~~Stacked deep-water $\delta^{18}O_{seawater}$ ~~stack~~ for the Pacific (purple) and North Atlantic (green) with ~~error envelopes~~uncertainties as in b. Stacks and ~~their errors~~associated uncertainties represent ~~the means~~mean of ~~original,~~ manually aligned, interpolated (~~to~~ 3 kyr ~~interval~~resolution) and bootstrapped records, with uncertainties reflecting both analytical error and the spread between paired estimates (see Sect. ~~2 Materials and Methods).~~ 6).

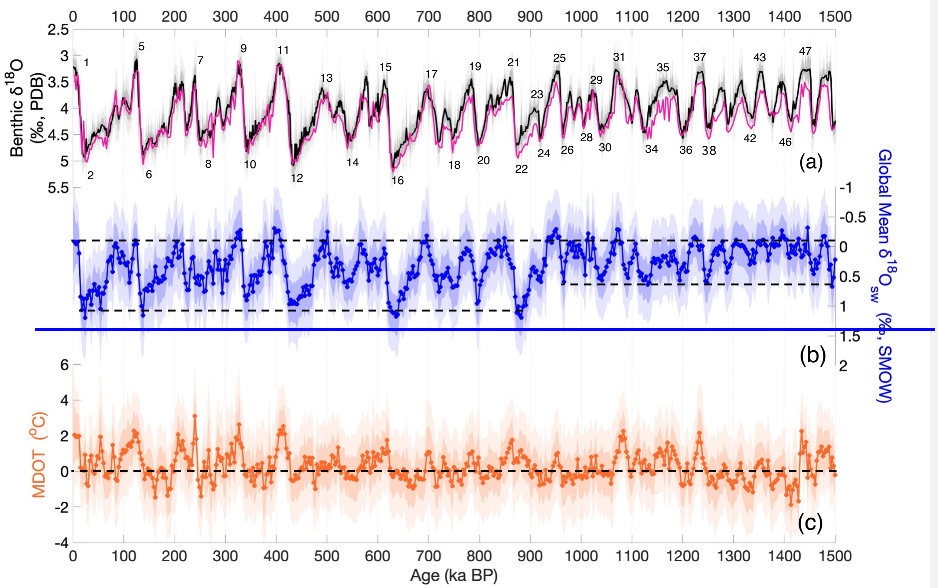


Figure 9: Global MDOT and mean $\delta^{18}O_{seawater}$ stacks weighted by ocean basin volume. (a) Prob-stack benthic $\delta^{18}O$ (Ahn et al., 2017) (black line

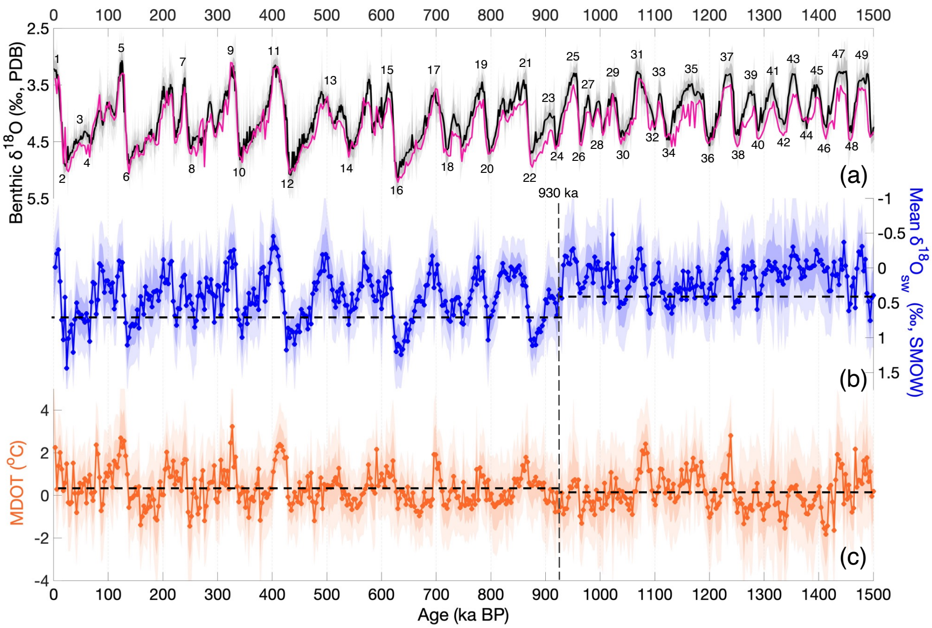


Figure 8: MDOT and mean $\delta^{18}O_{seawater}$ stacks weighted by ocean-basin volume. (a) Basin-volume–weighted benthic δ^{18}O stack compiled from Atlantic, Pacific and Southern Ocean records used in this study (pink) overlaid on Prob-stack benthic δ^{18}O (Ahn et al., 2017) (black). MIS numbers denote glacial (even) and interglacial (odd) stages.) overlain with the volume-weighted stack of global benthic δ^{18}O used in this study (pink); MIS numbers denoting glacial (even) and interglacial (odd) stages. (b) Global ocean volume-weighted stack of mean deep-water $\delta^{18}O_{seawater}$ (blue), showing an increase in ice volume during glacial stages from ~900 ka but no significant change in interglacial values (black dashed lines). (c) MDOT stack (orange) showing glacial deep-ocean temperatures have remained relatively constant (black dashed line) since their lowest point during MIS 46 (~1425 ka), and with a subtle cooling through the MPT (~1050–650 ka). Compared to the non-weighted records (Fig. S9), these records show a sharper increase in $\delta^{18}O_{seawater}$ (ice volume and salinity) at ~900 ka and colder deep-water temperatures derived from the Pacific Ocean in the early Pleistocene. The cooling between ~1050 to 925 ka, primarily due to the deep North Atlantic, is also more apparent in the weighted temperature record.

(b) Globally distributed basin-volume–weighted mean deep-water $\delta^{18}O_{seawater}$ stack (blue). Horizontal dashed lines indicate the glacial mean values for intervals before and after 930 ka. (c) Ocean volume–weighted MDOT stack (orange); the horizontal dashed lines indicate the total mean values for the intervals before and after 930 ka. The vertical dashed line marks 930 ka. Ocean-basin volume weighted stacks were constructed following the approach of Lisiecki and Stern (2016) (Sect. 2.6). Uncertainty envelopes incorporate analytical, calibration and age uncertainties propagated through a Monte Carlo framework, with basin-scale uncertainty additionally reflecting inter-record variability.

4 Discussion

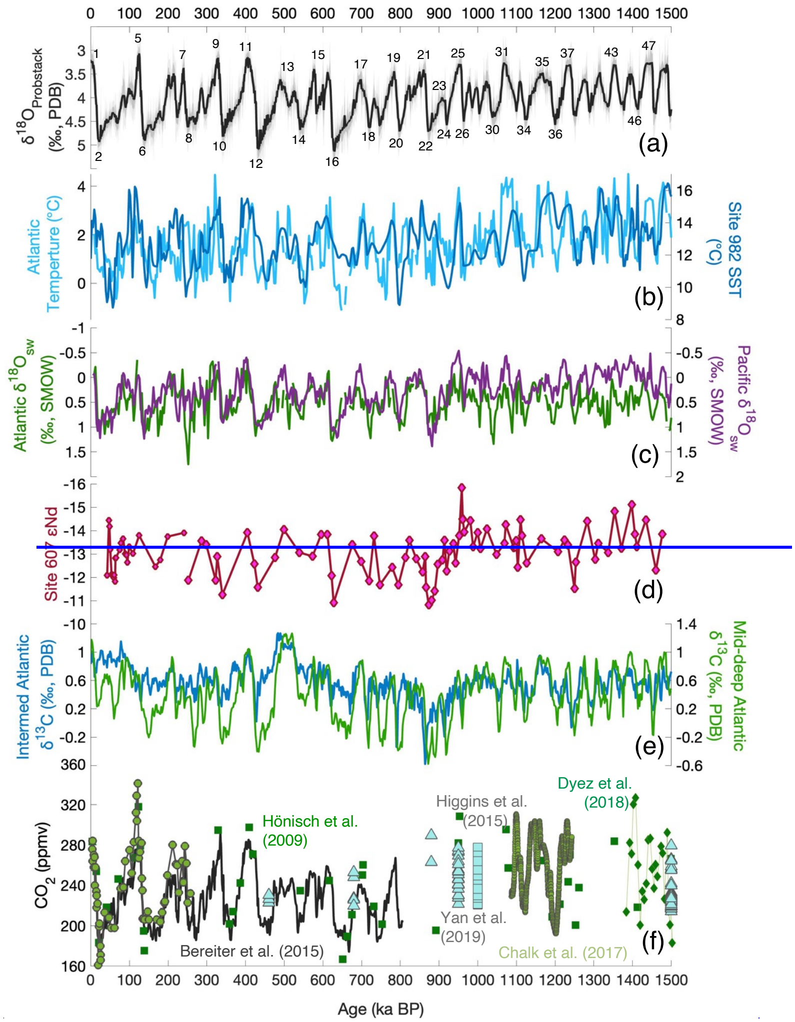
4.1 Basinal differences in abyssal temperature and salinity

Although the deep Atlantic and Pacific stacks contain only two records each, we suggest they are broadly representative of deep-water mass properties in the two basins. Comparison of Atlantic and Pacific temperature and $\delta^{18}O_{seawater}$ records ~~suggest an important change~~indicate differences in the ~~density structure~~magnitude of temperature and ~~stratification of~~ $\delta^{18}O_{seawater}$ changes across the ~~abyssal ocean at ~900 ka.~~MPT (Fig. 7b, c). The deep North Atlantic cooled during glacials between ~1050 to 925 ka ~~coinciding with~~whereas the deep Pacific remained cold, both before and after the MPT. The cooling in the deep North Atlantic is supported by a decrease (~2 °C) in high-~~-~~latitude North Atlantic sea surface temperature (SST) ~~as recorded~~ at Site 982 on the Rockall Plateau (58° N, 16° W~~)~~), close to the source area of deep-water formation (Lawrence et al., 2009) (Fig. ~~10b). Because deep water derived from the Southern Ocean is colder than that sourced from the North Atlantic, the cooling may also partly result from an expansion of southern sourced water into the deep North Atlantic. Cooling also coincides with an abrupt increase in neodymium isotopes (εNd) at Site 607 at ~900 to 870 ka (MIS 22) previously interpreted as indicating an increased influence of Southern Component Water (SCW) in the deep Atlantic (Kim et al., 2021; Pena and Goldstein, 2014) (Fig. 10d). Several recent studies have called this interpretation into question suggesting that the deep North Atlantic was bathed in Northern Component Water (NCW) during the last glaciation (Pöppelmeier et al., 2020) and in older glacial periods of the Quaternary (Hines et al., 2024; Williams et al., 2024)~~9b). Because deep water derived from the Southern Ocean is colder than that sourced from the North Atlantic, part of the cooling may also result from an expansion of southern sourced water into the deep North Atlantic during glacial stages. Cooling coincides with an abrupt increase in neodymium isotopes (εNd) at Site 607 that has been previously interpreted as indicating an increased contribution of Southern Component Water (SCW) in the deep Atlantic (Kim et al., 2021a; Pena and Goldstein, 2014) (Fig. 9d), with similar deep circulation changes inferred for the Pacific (Li et al., 2026) An increase in the $\delta^{13}C$ gradient between the intermediate and deep North Atlantic also supports a greater proportion of SCW in the North Atlantic across the MPT (Fig. 9e) (Lang et al., 2016; Lisiecki, 2014). Neodymium isotope changes in the Atlantic have alternatively been interpreted to represent changes in the end-member Nd value of NCW alongside more modest circulation changes (Pöppelmeier et al., 2020, 2022; Williams et al., 2024; Zhao et al., 2019).

Cooling of Northern Component Water (NCW) would increase its density and create a stratification inversion with deeper water masses in the absence of a density increase of SCW (Knorr et al., 2021). ~~The greater increase in deep Pacific $\delta^{18}O_{seawater}$ values relative to the deep Atlantic at ~900 ka suggests an increase in the salinity of the source areas for SCW formation (Figs. 8c and~~The $\delta^{18}O_{seawater}$ in both the deep North Atlantic and Pacific increases across the MPT reflecting expansion of continental ice sheets, but the increase is greater in the Pacific than the Atlantic (Fig. 7c). We attribute the greater increase in Pacific $\delta^{18}O_{seawater}$ to an increase in the salinity and density of SCW, which may have shoaled the boundary between NCW and SCW in the deep Atlantic if the density contrast between the two water masses increased (Figs. 9c and e) (Song et al., 2025). Glacial-interglacial change in the global ocean density stratification in much of the deep ocean (>2000 m) is controlled by surface freshwater forcing in the Southern Ocean (Sun et al., 2016). Increased salinity can be achieved by decreasing meltwater input to the marginal seas around Antarctica (e.g. Weddell and Ross Seas) either from the Antarctic ice sheet and/or grounded ice shelves. Sea ice expansion around Antarctica also raises salinity through brine rejection and expands the depth range of SCW (Ferrari et al., 2014), but sea ice formation does not significantly alter the $\delta^{18}O_{seawater}$ of seawater 10e). The $\delta^{18}O_{seawater}$ increase in the deep Pacific is attributed partly to ice volume and partly to increased salinity across the MPT. This requires a process to increase the salinity of surface waters in the marginal seas around Antarctica (e.g. Weddell and Ross Seas) where SCW forms. Decreasing freshwater input to these source areas can be achieved by either decreasing melting of the Antarctic ice sheet (Raymo et al., 2006) and/or reduced melting of grounded ice shelves (Adkins, 2013; Miller et al., 2012). Increased sea ice formation would also raise salinity, but this process has a small and opposite effect on the $\delta^{18}O_{seawater}$ (Kim and Timmermann, 2024; Toyota et al., 2013).

The exact mechanism by which the salinity in SCW source areas was increased is not certain. Raymo et al. (2006) suggested the East Antarctic ice sheet was terrestrial-based with melting margins before the MPT and transformed to a marine-based ice sheet afterward, which would have decreased freshwater input to the marginal seas around Antarctica. Adkins (2013) proposed that during glacial periods cooling of NADW decreased melting rates of floating Antarctic ice shelves, thereby increasing the salinity of SCW. In support of this, we observe cooling of the deep North Atlantic at the same time as inferred salinification of the deep Pacific.

Increased salinity would enhance the stratification of the deep ocean and render it a more effective carbon trap. Strengthening of SCW formation would also promote glacial deep ocean carbon storage through a standing volume effect (Skinner, 2009). Multiple proxies support decreased oxygen concentrations and increased nutrient concentrations in the deep Atlantic across the MPT (e.g. (Fig. 10e). (Farmer et al., 2019; Lear et al., 2016; Lisiecki, 2014; Thomas et al., 2022), which may have lowered the concentration of glacial atmospheric pCO_2 (Chalk et al., 2017; Higgins et al., 2015; Hönisch et al., 2009; Yan et al., 2019) (Fig. 10f).



Raymo et al. (2006) suggested the East Antarctic ice sheet was terrestrial-based with melting margins before the MPT and transformed to a marine-based ice sheet afterward, which would have decreased meltwater to marginal seas. An unconformity in the Ross Sea Embayment that spans the MPT has been interpreted to represent widespread expansion of a marine-based ice sheet (McKay et al., 2012) into the Ross Sea where AABW is formed today. Adkins (2013) proposed that cooling of NADW during glacial stages of the late Pleistocene decreased melting rates of floating Antarctic ice shelves. Adkins (2013) suggested decreased meltwater input led to the formation of a cold and salty version of Antarctic Bottom Water, and this water mass filled the deep ocean basins. We observe cooling of the high-latitude and deep North Atlantic across the MPT at the same time as the inferred salinity increase of the deep Pacific, which may support the mechanism outlined by Adkins (2013).

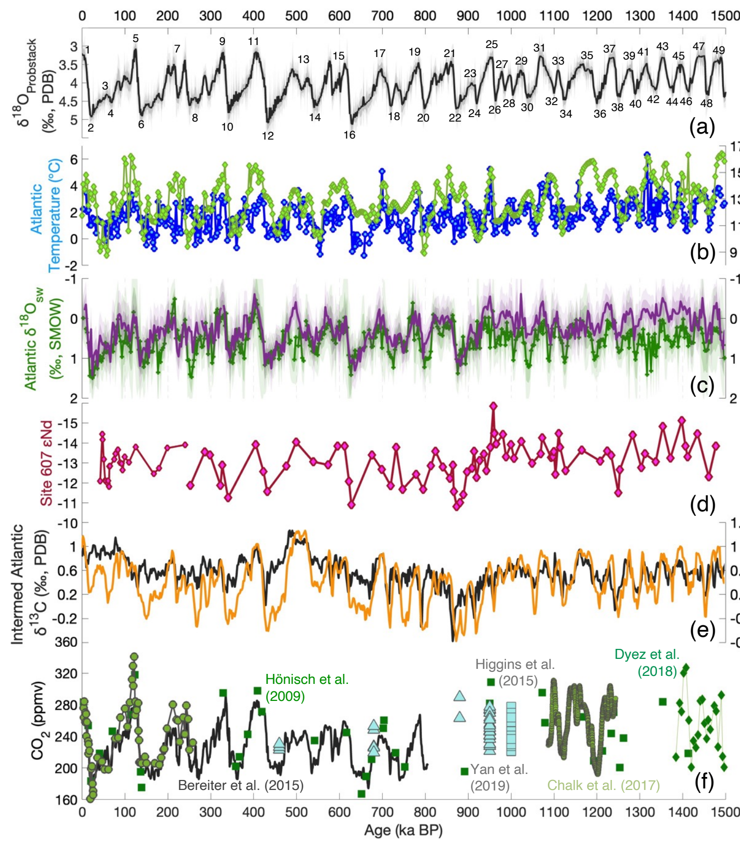


Figure ~~10~~9: Comparison of deep-water temperature, salinity, ocean circulation and atmospheric pCO_2 records over the last 1.5 Ma. (a) Benthic δ^{18}O Prob-stack (Ahn et al., 2017) (black line), MIS numbers denote glacial (below) and interglacial (above) stages. (b) Alkenone U$^{k'}_{37}$-derived, sea surface temperature (SST) record from ODP Site 982 (Lawrence et al., 2009) (~~dark blue~~green line with markers) overlaying the Atlantic stack for deep-water temperature (this study; ~~light~~ blue line with markers). (c) Atlantic (green) and Pacific (purple) δ^{18}O$_{seawater}$ stacks (this study). (d) DSDP Site 607/V30-97 εNd proxy for North Atlantic Ocean circulation changes ~~(Kim et al., 2021) (dark pink line with pink diamonds).~~(Kim et al., 2021b) (dark pink line with pink diamonds). (e) Carbon isotope gradient between regional δ^{13}C stacks (Lisiecki, 2014) for the intermediate depth North Atlantic (~~blue~~black) and middle deep Atlantic (~~green).~~orange). Low δ^{13}C reflects reduced ventilation in the deep Atlantic. (f) Atmospheric pCO_2 records (presented as in (Thomas et al., 2022)): Antarctic 800-kyr ice core data (Bereiter et al., 2015) (black line); Allan Hills Blue Ice (blue triangles (Yan et al., 2019) and blue squares (Higgins et al., 2015)); and δ^{11}B-based reconstructed pCO_2 [(Dyez et al., 2018) green line with diamonds; (Hönisch et al., 2009) solid green squares; (Chalk et al., 2017) early Pleistocene light green line/small circles, and late Pleistocene

light green circles]. Chronology for Atlantic and Pacific stacks is available in Sect. 2.1, all other records are presented on their original timescales. ~~Note: low $\delta^{13}C$ reflects ventilation in the deep Atlantic.~~

~~**4.2 Global versus Local Hydrographic Influences**~~

The freshwater fluxes to marginal seas can also be decreased by exporting freshwater to lower latitude via icebergs discharge and sea ice (Ferrari et al., 2014; Jansen, 2017). Starr et al. (2021) reported an increase in IRD in the Subantarctic Zone across the MPT and suggested this reflected increased transport of meteoric freshwater away from Antarctica. This process could have contributed to increased salinity in source areas of SCW formation, especially in the Weddell Sea sector. At the same time, increased transport of icebergs and sea ice from Antarctica increased the freshwater flux to the subantarctic, strengthened the halocline and slowed deep-to-surface exchange that controls carbon dioxide release to the atmosphere (Hasenfratz et al., 2019).

4.2 Implications for carbon storage

Although we cannot pinpoint which of the many processes discussed above was responsible for increased salinification and densification of SCW— some or all may have contributed— increased stratification of the glacial deep ocean may have played a role in enhancing ocean carbon storage and lowering atmospheric pCO_2 concentrations (Adkins 2013). Increased deep ocean stratification provides a physical mechanism by which sinking organic carbon can be remineralised and accumulate in the deep sea. In the process, oxygen is consumed, nutrients are released and total dissolved inorganic carbon (DIC) increases, although carbonate ion decreases. Several studies have reported an increase in deep-sea nutrients and lowered carbonate ion and oxygen concentrations in the deep Atlantic (e.g. (Fig. 9e) (Farmer et al., 2019; Lear et al., 2016; Lisiecki, 2014; Thomas et al., 2022) and Pacific (Diz et al., 2020; Peng et al., 2026; Qin et al., 2022) across the MPT.

Increased stratification of the deep ocean by itself is insufficient to change deep-ocean carbon storage. The biological pump exports organic matter from the surface ocean that is remineralised at depth. In the Southern Ocean, biological export production is thought to have increased owing to iron fertilization from dust over the MPT (Martínez-Garcia et al., 2011). The return of deep water towards the surface in the upwelling region of the Antarctic Circumpolar Current (ACC) ventilates the deep ocean through gas exchange in the Southern Ocean. This surface/deep exchange of water was reduced ("isolated") during glacial periods because of a strengthened halocline and reduced upwelling (Sigman et al., 2021). At Site 1094 across the MPT, Hasenfratz et al. (2019) suggested a reduction in deep water supply to the surface and a concomitant freshening of surface waters that strengthened the halocline, thereby reducing carbon dioxide release to the atmosphere. The strength of the Westerlies and ACC are also important for the upwelling of deep-water masses from the ocean interior in the Antarctic Zone, although changes may be asynchronous in different sectors of the Southern Ocean (Lamy et al., 2024; Sun et al., 2016).

Most hypotheses and models to explain the MPT invoke a glacial atmospheric pCO_2 decline to promote the growth of larger ice sheets; indeed, it's difficult to explain the MPT without this radiative cooling especially in the tropics (Herbert, 2023). Some studies have indicated a modest drop (~20 ppm) in concentrations of glacial atmospheric pCO_2 across the (Chalk et al., 2017; Higgins et al., 2015; Hönisch et al., 2009; Yan et al., 2019) (Fig. 9f), whereas others have found no mean change (Marks-Peterson et al., 2026). The uncertain magnitude of atmospheric pCO_2 decline across the MPT underscores the critical importance of new data from the BE-OI core, which has recovered continuous ice to 1.2 Ma and possibly to 1.5 Ma (Barbante etal., 2025).

4.3 Mean Deep Ocean Temperature

A critical question for the MPT is by how much and how fast did global deep ocean temperature and ice volume change? Global stacking of mean deep ocean temperature and $\delta^{18}O_{seawater}$ records (derived from paired benthic Mg/Ca-$\delta^{18}O$) can address this question and be compared to other more indirect approaches of estimating temperature and ice volume changes from benthic $\delta^{18}O$ records. Although only 5 long records of tandem benthic $\delta^{18}O$–Mg/Ca measurements exist spanning the last 1.5 Myr, we stacked the records first with equal weighting to reconstruct global records and then weighted by the volumes of different ocean basins (Lisiecki and Stern, 2016) (Sect. 2.~~7~~6) to better estimate global averages. Whereas the unweighted glacial MDOT record shows a slight decrease after ~900 ka, it is biased by the ~~strong~~ cooling of the deep Atlantic at Site 607 (Fig. ~~S9e~~S11c). Considering the much smaller volume of the deep North Atlantic compared to the deep Pacific, the volume-weighted MDOT stack is more representative and shows minimal change across the MPT (Fig. ~~9c). Both $\delta^{18}O_{seawater}$ stacks support previous findings for a step-like increase, reflecting increased continental ice volume during most glacial periods after ~900 ka (Elderfield et al., 2012; Ford and Raymo, 2020) (Figs. 9b and S9b~~8c).

We compare our stacked MDOT record with similar estimates derived by other methods. Rohling et al. (2021) (R21) calculated deep-sea temperatures (DST) using benthic $\delta^{18}O$ and estimates of $\delta^{18}O_{seawater}$ from sea level reconstructions. Our changes in MDOT agree well with R21 older than 1200 ka, after which the amplitude of R21 is greater than our MDOT (Fig. 10a). Neither record shows substantial cooling of deep ocean temperatures across the MPT. Chandler and Langebroek (2024) similarly used benthic $\delta^{18}O$, Mg/Ca and sea level reconstructions to estimate changes in mean Circumpolar Deep Water temperature (MCDWT) for the past 800 ka, including Sites 1123 and 1094 but using slightly different calibrations than this study (Fig. 10b). For the last 800 kyrs, our MDOT agrees well with MCDWT, indicating the slightly different approaches yield similar results. Lastly, we compare our MDOT estimates with changes in mean ocean temperature (MOT) reconstructed using noble gas (Xe/Kr) from the EPICA ice core (Grimmer et al., 2025) and shallow ice cores recovered from the Allan Hills blue ice area, Antarctica (Shackleton et al., 2026) (Fig. 10c). ~~We compare the MDOT and mean $\delta^{18}O_{seawater}$ stacks with similar estimates derived by other methods. Global deep-water stacked records agree with process-based estimates of temperature, sea level and $\delta^{18}O_{seawater}$ over the past 1.5 Myr (Rohling et al., 2021) (Figs. S11a and c), and for the last 800 ka with reconstructed Circumpolar Deep Water (CDW)~~The latter likely records a weighted averaging of glacial and interglacial conditions. There is good agreement between the two, especially considering that our MDOT does not include the entire ocean recorded by noble gas MOT. Both records show little change in MDOT or MOT across the MPT (Shackleton et al., 2026).

Both our MDOT and the noble gas MOT (Shackleton et al., 2026) results differ from reconstructions of Clark et al. (2024, 2025), which are based on using global mean sea surface temperatures (ΔGMSST) to estimate ΔMOT (Wolff, 2026). The ΔGMSST estimates are converted to ΔMOT using a non-constant ratio of ΔMOT/ΔGMSST over the MPT. For the period from 1.5 to ~0.9 Ma, these estimates give warmer deep-water temperatures relative to our study, followed by a marked cooling in deep ocean temperature (>2000 m) at ~900 ka (Fig. 10d). The Clark et al. compilation is strongly biased towards the deep North Atlantic (predominantly Site 607) with comparatively sparse Pacific deep ocean temperature (>2000 m) data over this interval (Clark et al., 2025; Lear et al., 2003). Site U1385 does not show cooling between 1.5 to ~0.9 Ma but the records from Sites 607 show a 1 °C long term cooling trend, which is unlikely to represent a global mean change in deep-ocean temperature (Figs. 2b, 3b, c, 5b, S3a, and S7b, c; Sect. 3.1). Instead, our findings support previous interpretations for near-freezing glacial bottom water temperatures in the Pacific and the South Atlantic Southern Ocean throughout the period from 1.5 to ~0.9 Ma, with no substantial change across the MPT (Elderfield et al., 2012a; Ford and Raymo, 2020; Hasenfratz et al., 2019; Siddall et al., 2010) (Figs. 8c, S9c and S12c). These results appear to be consistent with recent ice core noble gas-based reconstructions of MOT (Shackleton et al., 2026) (Fig. 10c), but higher resolution MOT data from BE-OI are needed to capture the full amplitude of glacial-interglacial variations (Barbante et al., 2025).

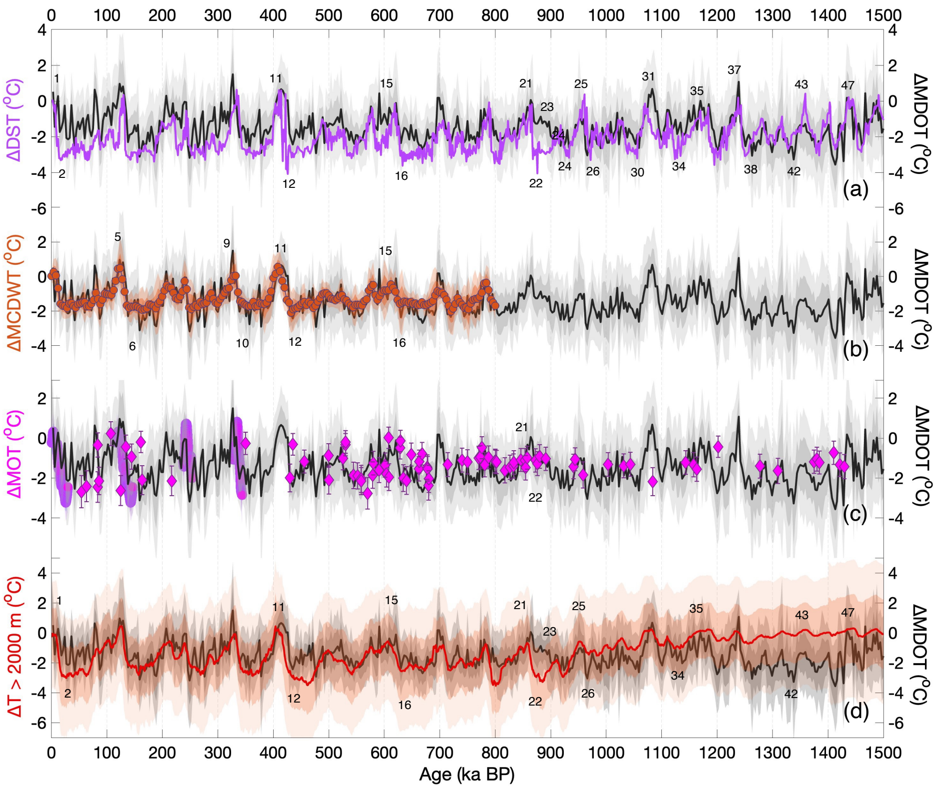


Figure 10: Comparison of MDOT with similar mean ocean and deep-water temperature reconstructions. All panels show the change in volume-weighted MDOT (ΔMDOT) (this study; black line) relative to an average present day value (1.73 ºC) versus: (a) the change in deep-sea temperature (ΔDST) estimated relative to the present (Rohling et al., 2021) (purple line); (b) the change in mean CDW temperature (ΔMCDWT) relative to present (Chandler and Langebroek, 2024) (dark orange circles); (c) the change in MOT (ΔMOT) derived from noble gas measurements in ice cores (Grimmer et al., 2025) (purple and pink circles) and, in Allan Hills blue ice (pink diamonds) (Shackleton et al., 2026), error bars on individual MOT estimates represent absolute uncertainties; and, (d) ΔMOT inferred from ΔGMSST and proxy-based deep-water temperature (Clark et al., 2025) (red), with both records referenced to the Early Holocene (0 °C at 10 ka) and offset by -1.73 0 °C following (Clark et al., 2025). Dark and light shading of the same colour represent 1σ and 2σ uncertainty envelopes, respectively. Uncertainties associated with ΔMOT are taken directly from the published datasets of Clark et al. (2025). MIS numbers mark many of the glacial (even) and interglacial (odd) stages.

4.4 Mean $\delta^{18}O_{seawater}$ and Ice Volume

Both weighted and unweighted mean ocean $\delta^{18}O_{seawater}$ stacks were compiled and show a step-like increase across the MPT at 930 ka (Figs. 8b and S11b), supporting previous interpretations of increased continental ice volume during most glacial periods after ~900 ka (Elderfield et al., 2012; Ford and Raymo, 2020).Fig. S11b) along with estimates derived from globally compiled records (Shakun et al., 2015) (Fig. S12). The global volume-weighted mean $\delta^{18}O_{seawater}$ stack also agrees well with reconstructions of relative sea level (RSL) Our volume-weighted mean $\delta^{18}O_{seawater}$ stack is consistent with process-based estimates of $\delta^{18}O_{seawater}$ over the past 1.5 Myr (Rohling et al., 2021) (Fig. 11a) and with reconstructions of relative sea level (RSL) change for the past ~500 ka (Grant et al., 2012, 2014) (Fig. 11b). Furthermore, changes in volume-weighted MDOT estimates over the last 350 ka relative to the Holocene are supported by available mean ocean temperature (MOT) changes derived from ice-core noble gas measurementsHowever, our value for MIS 22 is considerably greater than R21. This is the result of a warming trend of MDOT during MIS 22 that yields greater $\delta^{18}O_{seawater}$ values when the signal is deconvolved from the benthic $\delta^{18}O$ of calcite. Deep North Pacific Site 1208, in particular, shows anomalously high temperatures during MIS 22 that contribute to this discrepancy. (Grimmer et al., 2025) (Fig. 11c).

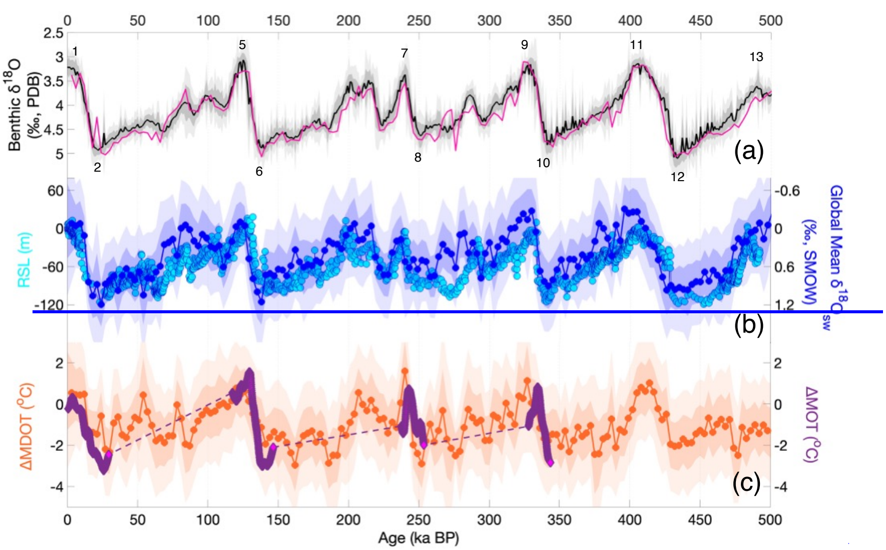


Figure 11: Comparison of volume-weighted MDOT and global mean $\delta^{18}O_{seawater}$ stacks with ice-core noble gas measurements of MOT and Red Sea RSL. (a) Prob-stack benthic $\delta^{18}O$ (Ahn et al., 2017) (black line) overlain with the ocean volume-weighted benthic $\delta^{18}O$ stack (this study; pink) with MIS numbers denoting glacial (below) and interglacial (above) stages. (b) The volume-weighted global mean $\delta^{18}O_{seawater}$ stack (this study; dark blue line with markers) versus Red Sea RSL reconstructions (Grant et al., 2012, 2014)

(light blue circles) scaled so that δ^{18}O$_{seawater}$ of 1.0 ‰ is equivalent to 10 m RSL. (c) The change in volume-weighted MDOT (ΔMDOT) relative to the Holocene (1.5 ºC (Goudsmit-Harzevoort et al., 2023; Rohling et al., 2021, 2022)) (this study; orange line with circles) overlain by ΔMOT derived from noble gas measurements in ice-cores (Grimmer et al., 2025) (purple line with pink diamonds).

Results differ substantially from estimates of changes in deep ocean temperature (>2000m) and mean δ^{18}O$_{seawater}$ constrained using ΔMOT inferred from proxy-based deep ocean temperatures and changes in global mean sea surface temperatures (ΔGMSST) (Clark et al., 2024, 2025). For the period from 1.5 to ~0.9 Ma prior to the MPT, warmer deep water temperatures and higher δ^{18}O$_{seawater}$ (greater ice volume) are calculated compared to our estimates. A large drop in deep ocean temperature (>2000 m) and no significant increase in ice volume are inferred at ~900 ka (Fig. 12). In contrast, the global MDOT stack shows near-freezing deep water temperatures during glacials both before and after the MPT with little evidence of cooling, except in the North Atlantic. The Clark et al. compilation is strongly biased towards the deep North Atlantic (predominantly Sites 607 and U1313) with only ~12 data points from the Pacific used to constrain deep ocean temperature (>2000 m) (Clark et al., 2025; Lear et al., 2003). Site U1385 does not show strong cooling between 1.5 to ~0.9 Ma (Figs. 2A, 3B, 4, 6, and S6; Sect. 3.1) but the record from Sites 607 and U1313 is unlikely to represent global mean changes and is complicated by multi-species measurementsOur results diverge from (Ford et al., 2016). Instead, our findings support previous interpretations for near-freezing glacial bottom water temperatures in the Pacific throughout the period from 1.5 to ~0.9 Ma with no significant change across the MPT (Elderfield et al., 2012; Ford and Raymo, 2020).

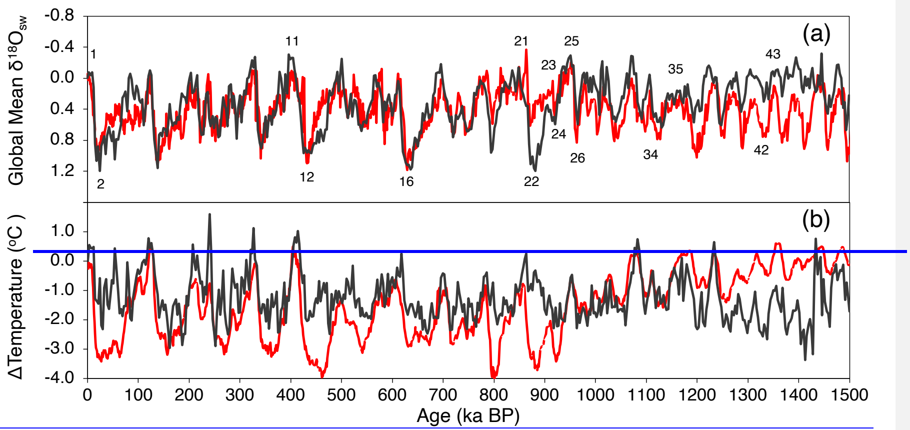


Figure 12: Comparison of MDOT and global δ^{18}O$_{seawater}$ against similar process-based stacks. (a) Volume-weighted global δ^{18}O$_{seawater}$ stack (this study; black) against δ^{18}O$_{seawater}$ derived from the deconvolution of the benthic δ^{18}O$_{ProbStack}$ using inferred changes in MOT (Clark et al., 2025) (red). Numbers denote MIS with glacial below and interglacial above. (b) Volume-weighted ΔMDOT (this

study; black) presented as the change from the preindustrial value (1.5 ºC) (Goudsmit-Harzevoort et al., 2023), versus ΔMOT inferred from ΔGMSST and proxy-based deep-water temperature (Clark et al., 2025) (red).

The interpretation of the MDOT history and global versus local/regional changes in global volume-weighted $\delta^{18}O_{seawater}$ has important implications for ice volume changes across the MPT (Fig. 12 and S13). Clark et al. (2025) infer little to no change in ice volume across the MPT which is contrary to most previous work. They discounted records of deep Pacific Ocean temperature and $\delta^{18}O_{seawater}$(Clark et al., 2024, 2025) in that our $\delta^{18}O_{seawater}$ values are lower prior to the MPT, particularly in the interval older than 1200 kyrs, suggesting less glacial ice volume before the MPT. They infer surprisingly large ice sheets before the MPT and relatively little change at ~900 ka, which is contrary to our results and most previous studies (Bintanja and Van De Wal, 2008; Clark et al., 2006; Elderfield et al., 2012a; Ford and Raymo, 2020; Rohling et al., 2021). Their method of reconstruction discounts the temperature and $\delta^{18}O_{seawater}$ records from Pacific Sites 1123 and 1208 (Elderfield et al., 2012; Ford and Raymo, 2020) claiming they are unrepresentative of global values and reflect regional hydrographic changes. While we agree there is a hydrographic (salinity) component to the increase in the glacial deep Pacific $\delta^{18}O_{seawater}$ signal that exceeds the Atlantic signal, there is also a global increase in $\delta^{18}O_{seawater}$ reflecting greater ice volume in most glacial periods after 0.9 Ma.across the MPT.

We acknowledge limitations to our study which underscores how comparable studies can improve upon this work in the future. The small number of tandem Mg/Ca–benthic $\delta^{18}O$ records available to stack leads to uncertainty in how well they represent global or basin-wide averages. The sites are all >2500 mbsl and a greater depth range would provide a better comparison with noble gas-based MOT estimates across the MPT.

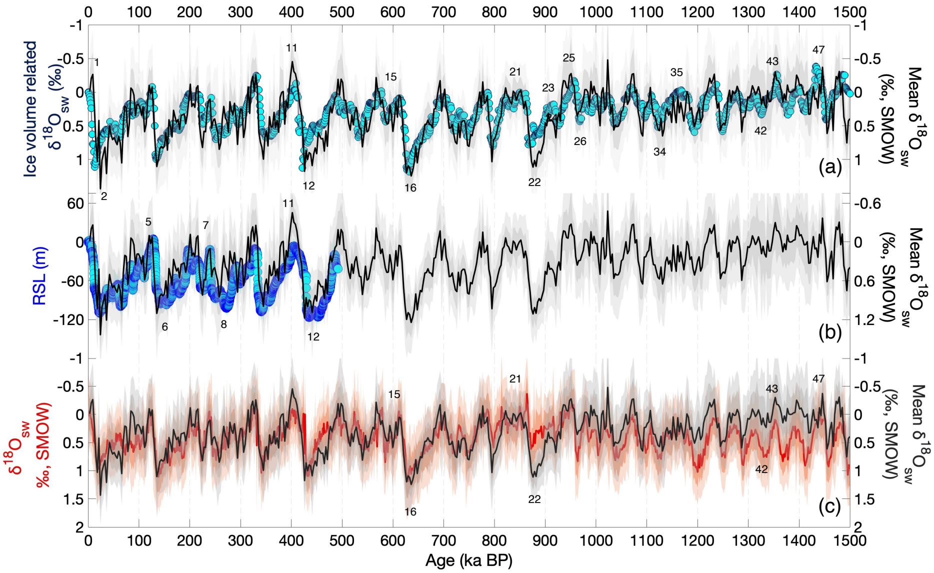


Figure 11: Comparison of mean $\delta^{18}O_{seawater}$ with similar process-based and relative sea level reconstructions. Volume-weighted mean $\delta^{18}O_{seawater}$ stack shown in each panel (this study; black) is compared against: (a) process-based estimated ice-volume related deep-sea $\delta^{18}O_{seawater}$ (Rohling et al., 2021) (light blue circles); (b) Red Sea relative sea level (RSL) reconstructions (Grant et al., 2012, 2014) (dark and light blue circles) scaled so that $\delta^{18}O_{seawater}$ of 1.0 ‰ is equivalent to 10 m RSL; and, (c) $\delta^{18}O_{seawater}$ derived from the deconvolution of the benthic $\delta^{18}O$ Prob-stack using inferred changes in MOT (Clark et al., 2025) (red). MIS numbers denote glacial (below) and interglacial (above) stages. Dark and light shading of the same colour represent 1σ and 2σ uncertainty envelopes, respectively. Uncertainties associated with $\delta^{18}O_{seawater}$ are taken directly from the published datasets of Clark et al. (2025). The $\delta^{18}O_{seawater}$ reconstruction of Clark et al. (2025) includes removal of a long-term increase of 0.083 ‰ Myr^{-1} in benthic $\delta^{18}O$, which is not applied to our records because such an arbitrary adjustment for diagenesis remains unproven. This adjustment is small relative to uncertainties in the $\delta^{18}O_{seawater}$ compilations and would increase the divergence between the two mean $\delta^{18}O_{seawater}$ curves.

4.5 Limitations and Future Opportunities

The small number of long, continuous benthic Mg/Ca–$\delta^{18}O$ records available across the MPT leads to questions about how well they represent basin-wide or global averages. Low abundance of preferred benthic foraminiferal taxa in some of the records results in time gaps where fewer than 5 sites are available for stacking. The water depths of the studied sites are all below 2500 mbsl and a greater depth range would provide a better comparison with noble gas-based MOT estimates across the MPT (Shackleton et al., 2026). The temporal resolution is relatively low (3 kyrs) and may not reflect the full amplitude of changes and certainly do not capture millennial-scale variability. Complementary B/Ca on the same samples would provide a proxy for carbonate ion changes, which could be used to assess if Mg/Ca palaeotemperatures of different benthic species of foraminifera have been affected by dissolution. The propagation of error when using paired benthic Mg/Ca–$\delta^{18}O$ to calculate the $\delta^{18}O_{seawater}$ is fairly large (±0.24 ‰ for *U. peregrina* and ±0.16 ‰ for *G. affinis* (1 σ), with the dominant contribution arising from temperature uncertainty). Using benthic foraminifera taxa with a greater sensitivity of Mg/Ca to temperature could potentially reduce the errors (Yang et al., 2025).

The recent recognition that the time rate of change of benthic $\delta^{18}O$ is a proxy for Earth's Energy Imbalance (EEI) offers the opportunity to investigate the processes that modify the global energy budget on long timescales (Shackleton et al., 2023). With more records, the derivative of the deconvolved stacks would permit separation of the sensible and latent heat components and provide valuable insight how the Earth's energy balance changed during the MPT.

All of the points discussed above underscore the importance of acquiring additional paired benthic Mg/Ca–$\delta^{18}O$ records to improve the records presented herein. Furthermore, the proposed causes for the changes in deep-ocean stratification and their impact on deep-ocean carbon storage should be tested with numerical models.

5 Conclusions

The causes of the Middle Pleistocene TransitionMPT have been sought since the frequency and amplitude change of glacial-interglacial cycles wereit was first recognized in oxygen isotope records over 50 years ago (Pisias and Moore, 1981; Shackleton, Nicholas John and Opdyke, 1973). The MPT involved a fundamental change in the interaction of deep-ocean circulation, pCO_2, and the cryosphere that permitted expansion of continental ice sheets (Chalk et al., 2017; Elderfield et al., 2012; Ford and Raymo, 2020; Higgins et al., 2015; Hönisch et al., 2009; Pena and Goldstein, 2014; Yan et al., 2019). Here we demonstrate the MPT involved changes in the physical properties of the abyssal ocean, including a cooling of the deep North Atlantic and salinification of the deep Pacific. This resulted in increased density stratification of the deep ocean in glacial periods after ~900 ka. The abyssal ocean became a more effective carbon trap that was fuelled partly by increased export production and surface stratification in the Southern Ocean (An et al., 2024; Billups et al., 2018; Ferrari et al., 2014; Hasenfratz et al., 2019; Jaccard et al., 2013; Martínez-Garcia et al., 2011; Qin et al., 2022; Rodríguez-Sanz et al., 2012; Starr et al., 2021; Weber et al., 2022).

Increased ocean carbon storage would have lowered glacial atmospheric pCO_2 (Chalk et al., 2017; Higgins et al., 2015; Hönisch et al., 2009; Yan et al., 2019), thereby permitting the ice sheets to grow larger and making them less susceptible to complete deglaciation with each obliquity cycle (Huybers, 2006; Huybers and Wunsch, 2005). Increased salinity of the deep Pacific implies decreased freshwater input into the source areas of SCW formation around Antarctica and increased sea ice extent. It is unclear whether this was brought about by a transition from terrestrial melting margins to marine-based ice sheets on Antarctica (Raymo et al., 2006) and/or decreased melting of floating ice shelves by a cooling of NADW (Adkins, 2013).

~~The stacked mean $\delta^{18}O_{seawater}$ record supports previous findings of an increase in glacial ice volume at ~900 ka and an intensification of the glacial cycles~~. The MPT involved a fundamental change in the interaction of deep-ocean circulation, atmospheric pCO_2, and the cryosphere that permitted continental ice sheets to grow large enough that they were able to survive modest rises in summer insolation that would have resulted in deglaciation prior to the MPT (Ganopolski, 2024; Raymo et al., 1997; Willeit et al., 2019). Here we suggest that the MPT involved changes in the physical properties of the abyssal ocean, including a cooling of the deep North Atlantic and salinification of the deep water in the Pacific. This resulted in increased density stratification of the deep ocean in glacial periods after ~900 ka.

Global ocean density stratification in much of the deep ocean (>2000 m) is controlled by surface freshwater forcing in the Southern Ocean (Sun et al., 2016). We suggest that across the MPT, meltwater input to the marginal seas around Antarctica decreased and sea ice increased during glacial periods, resulting in formation of cold (near freezing), salty deep water. Increased deep ocean stratification provides a physical mechanism by which sinking organic carbon from the surface can accumulate in the deep sea, especially if accompanied by increased export production and surface stratification in the Southern Ocean. Many complementary studies support increased carbon storage in the deep Atlantic (e.g. (Fig. 9e) (Farmer et al., 2019; Lear et al., 2016; Lisiecki, 2014; Thomas et al., 2022) and Pacific (Diz et al., 2020; Peng et al., 2026; Qin et al., 2022a) across the MPT, and some atmospheric pCO_2 proxy records show a moderate decline in glacial atmospheric concentrations (Chalk et al., 2017; Higgins et al., 2015; Hönisch et al., 2009; Yan et al., 2019). Forthcoming analyses of the newly recovered continuous Antarctic ice core spanning the MPT (Wolff et al., 2022) will provide critical information on changes in greenhouse gases and MOT, which are needed to test many of the inferences made herein.

~~(Elderfield et al., 2012; Ford and Raymo, 2020). The anticipated recovery of a continuous Antarctic ice core across the MPT (Wolff et al., 2022) will provide critical information, especially greenhouse gases and MOT, needed to test many of the inferences made herein on the basis of the stacked MDOT and $\delta^{18}O_{seawater}$ records.~~

Data availability. All data, not previously reported, has been deposited at the World Data Center PANGAEA repository (www.pangaea.de.) and will be publicly available when the paper is accepted and a DOI issued
Previously reported benthic $\delta^{18}O$ for IODP Site U1385 (Hodell et al., 2023) are archived at: (https://doi.org/10.1594/PANGAEA.951401~~)~~)).

Supplement link.

~~**Author contributions.** NCT and DAH designed the project. NCT conducted all analyses in collaboration with: DAH for stable isotopic analyses, MG for trace element analyses, and HLF for PSU Solver analyses. NCT and DAH wrote the original draft, and all co-authors contributed to the final manuscript.~~

Author contributions. Conceptualization: DAH conceived the study, led its design, and collected the samples; NCT co-designed the study: NCT and MG processed and prepared the new dataset; NCT organised the complete dataset and associated metadata for analysis and archiving. Formal analysis: NCT and HLF applied statistical and computational methods to analyse the data and develop the stacked reconstructions. Funding acquisition: DAH. Investigation: NCT and MG, collected new trace element data from Site U1385; NCT and DAH generated the stable isotope data; all authors contributed to data analysis. Methodology: DAH and NCT. Project administration: NCT and DAH coordinated the research activities and workflow.

Resources: DAH and MG provided laboratory facilities, materials and analytical support. Software: HLF and NCT implemented and executed the PSU Solver code. Visualization: NCT and DAH. Writing – original draft: NCT and DAH. Writing – review and editing: NCT, MG, HLF, and DAH.

Competing interests. One of the (co-)authors is a member of the editorial board of Climate of the Past. The authors declare that they have no other competing interests.

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