



Regional low-latitude planktic $\delta^{18}\text{O}$ and Mg/Ca proxy response to changes in middle Miocene global ocean overturning circulation

Anta-Clarisse Sarr^{1,2}, Kaustubh Thirumalai^{1,3}, R. Paul Acosta⁴, Daeun Lee², Christopher J. Poulsen²

5 ¹ Aix Marseille Université, CNRS, IRD, INRAE, CEREGE, Aix-en-Provence, France

² Department of Earth Science, University of Oregon, Eugene, OR, USA

³ Department of Geosciences, University of Arizona, Tucson, AZ, USA

⁴ Atmospheric, Oceanic & Earth Sciences Department, George Mason University, Fairfax, VA, USA

10 *Correspondence to:* Anta-Clarisse Sarr (sarr@cerege.fr)

Abstract. Palaeoceanographic records suggest that a reorganization of low-latitude ocean-atmosphere dynamics occurred across the Middle Miocene Climate Transition (MMCT, 14.7-13.8 million years ago), but the mechanisms connecting global climate forcing to regional patterns are yet uncertain. The current paradigm for explaining this transition is tied to increasing latitudinal temperature gradients forced by global climate cooling due to decreasing atmospheric CO₂ concentrations. One overlooked possibility is the impact of changes in global ocean circulation patterns. Here, we use mid-Miocene simulations with an isotope-enabled model to investigate the fingerprint of global ocean circulation versus CO₂ levels on regional temperature, salinity, and the oxygen isotopic composition ($\delta^{18}\text{O}$) of seawater. We then explore how these parameters would be encoded in multi- and individual-test planktic foraminiferal $\delta^{18}\text{O}$ and Mg/Ca analyses using a proxy system model. Our results show that seawater $\delta^{18}\text{O}$, as opposed to temperature reconstructions, has the potential to delineate regional changes driven by perturbations to global ocean circulation. These predictions help evaluate where and how effectively tropical records inform us about the relative contributions of CO₂ forcing versus ocean circulation changes in driving the MMCT and other climate transitions across the Miocene.

1 Introduction

Climate system evolution since the Neogene (from 23 million years ago, Ma) provides a unique opportunity to better understand the complex interactions between (marine) carbon cycling, high-latitude ice extent, and ocean-atmospheric dynamics (e.g., Flower and Kennett, J. P., 1994; Sosdian et al., 2020; Holbourn et al., 2022). One interval that stands out is the Middle Miocene Climate Transition (MMCT, 14.7 to 13.8 Ma), which represents a major step in the late Cenozoic descent toward the present-day “icehouse” in the aftermath of the Miocene Climate Optimum (16.9 to 14.7 Ma) (Flower and Kennett, J. P., 1994; Westerhold et al., 2020). The MMCT is characterized by a stepwise increase in the oxygen isotope ratio ($\delta^{18}\text{O}$) of



30 benthic foraminifera, typically linked to bottom-water cooling (e.g., Leutert et al., 2021; Meckler et al., 2022; Hou et al., 2023) and/or the expansion of the Antarctic cryosphere (e.g., Levy et al., 2016, 2019; Sangiorgi et al., 2018). The MMCT also overlaps with the end of the global carbon cycle event known as the Monterey excursion (~16.7-13.5 Ma) (Flower and Kennett, 1993; Vincent and Berger, 1985). Most studies presume a link between this climate transition and $p\text{CO}_2$ drawdown, but the precise mechanisms at play and the respective contribution of various climatic forcing factors versus internal climate feedback are still debated (e.g., Flower and Kennett, J. P., 1994; Foster et al., 2012; Raitzsch et al., 2021; Sosdian et al., 2020; Herbert et al., 2022; Holbourn et al., 2014, 2022; Woodruff and Savin, 1989; Wright et al., 1992).

Alongside these global-scale changes, growing evidence suggests that the MMCT and post-MMCT periods (14.7 to 12.6 Ma) were associated with rearrangement(s) of low-latitude ocean-atmosphere dynamics that perturbed the upper tropical ocean physical and biogeochemical characteristics (e.g., Auderset et al., 2022; Auer et al., 2023; Fox et al., 2021; Hess et al., 2023, 2026; Holbourn et al., 2010, 2014; Matsui et al., 2017; Nathan and Leckie, 2009; Podder et al., 2024; Sosdian and Lear, 2020; Yang et al., 2020b; Zou et al., 2024). A striking feature is the apparent freshening of the tropical upper ocean, as evidenced by the decrease in the oxygen isotopic composition of seawater ($\delta^{18}\text{O}_{\text{sw}}$), reconstructed from paired measurements of $\delta^{18}\text{O}$ and the Mg/Ca ratio on planktic foraminiferal tests at multiple locations (Fox et al., 2021; Holbourn et al., 2010; Podder et al., 2024; Sosdian and Lear, 2020; Yang et al., 2020b; Zou et al., 2024). Mg/Ca provides an independent paleothermometer to deconvolve the $\delta^{18}\text{O}_{\text{sw}}$ from the temperature signal incorporated into foraminiferal calcite $\delta^{18}\text{O}$ ($\delta^{18}\text{O}_c$). Such changes in hydrography have mostly been attributed to modifications of regional surface-ocean circulation (e.g., Fox et al., 2021; Podder et al., 2024; Sosdian and Lear, 2020; Yang et al., 2020b; Zou et al., 2024) and/or to changes in the hydrological cycle related to latitudinal migrations of the Inter-Tropical Convergence Zone (ITCZ) (e.g., Fox et al., 2021; Holbourn et al., 2010) directly forced by $p\text{CO}_2$ -driven global cooling, Antarctic glaciation, and sharper latitudinal temperature gradients. Mechanisms linking $p\text{CO}_2$ and high-latitude cryosphere changes to regional climate patterns are yet uncertain, and a coherent explanation to reconcile multi-proxy regional records remains a significant gap in our understanding of Miocene climate change. In the absence of a coherent hypothesis, regional changes are often attributed to bathymetric changes in seaway regions (e.g., Crichton et al., 2021; Gourlan et al., 2008; Hess et al., 2026; Nathan and Leckie, 2009; Sosdian and Lear, 2020).

55 A potential global-scale driver of regional climatic change during the Miocene that has received little attention is a reorganization of the oceanic Global Overturning Circulation (GOC). Several studies have inferred intensified deep-water circulation across the MMCT (e.g., Diester-Haass et al., 2013; Holbourn et al., 2024; Holbourn, Kuhnt, Frank, et al., 2013), invoking increased poleward heat and moisture transport. Implicit in this interpretation, however, is the assumption that the large-scale pattern of the GOC remained fixed throughout the middle Miocene. Evidence for persistent stationarity is limited. The evolution of the GOC over million-year timescales remains debated, owing to challenges in interpreting regional benthic $\delta^{13}\text{C}$ records and other paleo-circulation tracers (e.g., Cortese et al., 2004; Cramer et al., 2009; Denk et al., 2013; Müller-Michaelis et al., 2013; Poore et al., 2006; Scher and Martin, 2008; Thomas and Via, 2007; Uenzelmann-Neben and Gruetzner,



2018; Woodruff and Savin, 1989; Wright et al., 1992). Paleoclimate model simulations also lack consensus, with diverse
 65 representations of GOC in different Earth System Models and middle Miocene simulation ensembles (Hutchinson et al., 2025;
 Liu et al., 2024; Tan et al., 2026; Naik et al., 2025), including support for multiple quasi-stable states of the Atlantic Meridional
 Overturning Circulation (AMOC), which is a characteristic part of the GOC (Lee et al., 2025).

Regional reconstructions, though, point to substantial changes in ocean circulation around the North Atlantic region, where
 70 Northern Atlantic Deep Water (NADW) and its precursor, Northern Component Water (NCW) form, wherein temperature
 reconstructions suggest a reorganization of the gyre circulation (e.g., Denk et al., 2013; Galochkina et al., 2023; Sangiorgi et
 al., 2021; Super et al., 2018). In the Indo-Pacific, restructuring of deep and intermediate water-masses has been attributed to
 changes in the Pacific meridional overturning circulation (e.g., Holbourn et al., 2013a, 2024; Kender et al., 2018; Ma et al.,
 2018; Zhang et al., 2025). In the Southern Ocean, the northward migration of frontal systems (Hou et al., 2023; Kuhnert et al.,
 75 2009; Leutert et al., 2020; Sangiorgi et al., 2018; Verducci et al., 2009) has been linked to changes in Antarctic Circumpolar
 Current (ACC) dynamics during the MMCT. Because each of these regions represents an integral component of the GOC
 (Talley, 2013), their concurrent reorganization raises the possibility that the MMCT involved a reconfiguration of the global
 meridional overturning circulation.

80 Changes in GOC are known to affect tropical ocean-atmosphere dynamics and upper ocean temperature and hydrography in
 ways that are distinct from those due to changes in atmospheric $p\text{CO}_2$ (e.g., Chiang et al., 2008; Liu et al., 2017; Timmermann
 et al., 2007). Here, we investigate the fingerprint of equilibrium-state change in GOC versus $p\text{CO}_2$ -driven cooling on tropical
 upper-ocean hydrography ($\delta^{18}\text{O}_{\text{sw}}$ and salinity) and temperature under middle-Miocene boundary conditions. To do so, we
 leverage a unique set of isotope-enabled CESM1.2 simulations that includes two quasi-equilibrium AMOC states at identical
 85 atmospheric $p\text{CO}_2$ levels (Lee et al., 2025) (Table 1), allowing us to isolate the climatic effects of GOC reorganization from
 those of CO_2 forcing. Taking advantage of the model's ability to simulate oxygen isotopes, we then use its output to forward
 model $\delta^{18}\text{O}_\text{c}$ and Mg/Ca proxy systems in planktic foraminifera. These predictions allow us to evaluate where and how
 effectively these proxy systems preserve the signature of global-scale circulation change and, in turn, to reevaluate the
 mechanisms responsible for tropical climate patterns across the MMCT.

90 2 Materials and Methods

2.1 Climate model simulations

We use three mid-Miocene simulations performed with the isotope-enabled Community Earth System Model 1.2 (iCESM1.2)
 (Brady et al., 2019) (Table 1). This setup includes the Community Atmosphere Model version 5.3 (CAM5.3) and the
 Community Land Model version 4.0 (CLM4) (1.9° by 2.5° finite volume grid and 30 uneven vertical levels in the atmosphere),



95 and the Parallel Ocean Program version 2 (POP2) and the Community Ice Code version 4.0 (CICE4) ($1^\circ \times 1^\circ$ and 60 uneven
 vertical levels in the ocean). Simulations were run with the mid-Miocene paleogeography, land-ice, and vegetation boundary
 conditions from Goldner et al. (2014) (see also Burls et al., 2021) and have been detailed in previous studies (Acosta et al.,
 2022; Gaskell et al., 2022; Lee et al., 2025). The main differences from the present-day configuration are the absence of the
 Greenland Ice Sheet, a smaller ice sheet than the modern Antarctic Ice Sheet, open Central American and Tethyan Seaways
 100 that allow inter-basin tropical ocean circulation, a narrower North Atlantic Gateway, and a wider Indonesian Gateway (Fig.
 1). Atmospheric CO_2 was set to 400 ppm for WARM cases with a strong (WARM-*on*) versus weak (WARM-*off*) GOC, and
 to 280 ppm for the COLD case (COLD-*on*). Other greenhouse gases and orbital parameters were set to their preindustrial
 values. Isotopic composition of the seawater ($\delta^{18}\text{O}_{\text{sw}}$) was initialized as a uniform value of 0.0‰. WARM-*on* and WARM-*off*
 simulations have different initial conditions, which result in distinct GOC states (see Lee et al., 2025 for details on the
 105 simulation setup) (Fig. S2): WARM-*on* and COLD-*on* simulations exhibits a vigorous AMOC (>25 Sverdrup, Sv; $1\text{ Sv} = 10^6$
 m^3/s ; Fig. S2a, d) and a weak and shallow PMOC (< 6 Sv). On the other hand, the WARM-*off* simulation exhibits a very weak
 AMOC (< 6 Sv, Fig. S2b, e), and no PMOC. AMOC is slightly stronger (+8 Sv) and PMOC weaker in the COLD-*on* relative
 to WARM-*on* (Fig. S2c, f). We note that we do not discuss here mechanisms that sustain different patterns of AMOC or PMOC
 (see Lee et al., 2025) but instead use the different simulated states as possible Miocene climate states to discuss how GOC
 110 states can overprint upper-ocean characteristics.

Simulations were run for 1,600 to 2,100 years (Lee et al., 2025), and we analysed the final 400 years of each simulation for
 the analysis, except where stated otherwise. We do not address here the mechanisms that sustain the different AMOC and
 PMOC states (see Lee et al., 2025). Instead, we treat these simulated circulation regimes as plausible Miocene climate states
 115 and use them to evaluate how changes in the GOC are expressed in upper-ocean temperature and hydrography. We consider
 three scenarios: (1) AMOC reinvigoration (ΔGOC : WARM-*on* vs WARM-*off*); (2) $p\text{CO}_2$ -driven cooling accompanied by
 strengthening of a pre-existing AMOC (ΔCO_2 : COLD-*on* vs WARM-*on*) and (3) AMOC reinvigoration combined with global
 cooling ($\Delta\text{GOC_CO}_2$: COLD-*on* vs WARM-*off*) (Table 1). To explore how these changes would be recorded by species
 occupying different upper-ocean habitats, we averaged model outputs both between 0- and 50 and 50- and 80 meters depths.

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Table 1: Summary table of simulations and scenarios. CESM simulations are from Lee et al. (2025).

CESM simulations	Boundary conditions	$p\text{CO}_2$ (ppm)	AMOC State	Scenarios build from the simulations		
				ΔGOC	ΔCO_2	$\Delta\text{GOC_CO}_2$
WARM- <i>on</i>	Goldner et al. (2014)	400	Strong ($> 25\text{Sv}$)	x	x	
WARM- <i>off</i>	Goldner et al. (2014)	400	Weak ($< 6\text{Sv}$)	x		x
COLD- <i>on</i>	Goldner et al. (2014)	280	Strong ($> 25\text{Sv}$)		x	x



2.2 Proxy System Forward Modeling

To evaluate how the simulated ocean states would be expressed in planktic foraminiferal geochemistry, we forward-modeled the $\delta^{18}\text{O}_c$ and Mg/Ca proxy systems from the simulated temperature, $\delta^{18}\text{O}_{sw}$, and salinity fields, sampled at the model grid point nearest to each core site. We use the Bemis et al. (1998) low-light equation for the $\delta^{18}\text{O}_c$:

$$T(^{\circ}\text{C}) = 16.5 - 4.8 * (\delta^{18}\text{O}_c - \delta^{18}\text{O}_{sw}) \quad (1)$$

Simulated $\delta^{18}\text{O}_{sw}$ (‰) was corrected for the 0.27 ‰ equilibration of PDB-evolved CO_2 with SMOW for the above paleotemperature equation (Hut, G, 1987).

We calculate Mg/Ca with the Gray and Evans (2019) multivariate calibration, using their *Trilobatus sacculifer* parameters:

$$\text{Mg/Ca} = \exp[a * (S - 35) + b * T + c] \quad (2)$$

where $a = 0.054 \text{ psu}^{-1}$ is the salinity sensitivity, $b = 0.062 \text{ }^{\circ}\text{C}^{-1}$ is the temperature sensitivity, and $c = -0.24$ is the calibration constant (the intercept d of Gray and Evans (2019)). We do not include a carbonate-ion term because *T. sacculifer* appears to show no resolvable pH sensitivity in the underlying culture data ($0.01 \pm 0.08 \text{ mmol mol}^{-1}$ per pH unit; Gray and Evans (2019)), in contrast to the -0.5 to -0.9 sensitivities of other planktic species. We note that Holbourn et al. (2010) used *G. obliquus* and *G. subquadratus* for ODP site 1146 paired Mg/Ca– $\delta^{18}\text{O}_{sw}$ measurements and found that for a limited interval, these values were the same as *T. sacculifer*. Accordingly, in this study, we ignore the pH effect (Gray and Evans, 2019) of the former species belonging to the *G. ruber* complex (Spezzaferri et al., 2015). Because the Gray and Evans (2019) calibration is defined for modern seawater Mg/Ca, we scaled the forward-modeled values to middle-Miocene Mg/Ca seawater composition following Martinot et al. (2022), and references therein:

$$\text{Mg/Ca} = \text{Mg/Ca}_{\text{calc}} \times (\text{Mg/Ca}_{sw}^{\text{Miocene}} / \text{Mg/Ca}_{sw}^{\text{modern}})^H \quad (3)$$

where $\text{Mg/Ca}_{\text{calc}}$ is the value from Eq. (2), $\text{Mg/Ca}_{sw}^{\text{modern}} = 5.2 \text{ mol mol}^{-1}$, $\text{Mg/Ca}_{sw}^{\text{Miocene}} \approx 2.46 \text{ mol mol}^{-1}$ at $\sim 14 \text{ Ma}$, and the equation was linearized by setting $H = 1$ (Tierney et al., 2019). This correction scales absolute Mg/Ca down uniformly in both simulated states and therefore does not affect the WARM-on–WARM-off (ΔGOC) pattern comparisons but affects reported absolute Mg/Ca values (Fig. S5). We stress that this is a theoretical exercise to explore proxy sensitivity and is not intended to compare published records with our simulations—instead, this is performed using inverted temperature and $\delta^{18}\text{O}_{sw}$ values (see below).



155 2.3 Revisiting published MMCT paired Mg/Ca– $\delta^{18}\text{O}_\text{c}$ records using PSU Solver

We reevaluate published paired $\delta^{18}\text{O}_\text{c}$ –Mg/Ca tropical records spanning ca. 15–13 Ma using a bootstrap Monte Carlo inversion built on the Paleo-Seawater Uncertainty (PSU) Solver framework (PSU Solver; Thirumalai et al. (2016)). We include the following paired records developed from cores at Ocean Drilling Program (ODP) site 1146 (South China Sea; Holbourn et al. (2010)), Integrated Ocean Drilling Program (IODP) site U1338 (eastern equatorial Pacific; Fox et al. (2021)), ODP site 761
 160 (western Australian margin; Sosdian et al. (2020), Sosdian and Lear (2020)), National Gas Hydrate Program Expedition 01 (NGHP-01) site 01A (eastern Arabian Sea; Yang et al. (2020b), Zou et al. (2024)) and ODP site 806 (central equatorial Pacific; Sosdian and Lear (2020)) (Fig. 1a). Other paired records exist but do not span the MMCT (e.g., Sunbird-1; Nairn et al. (2021)) and were excluded. We retain the published age model for each record, acknowledging potential inaccuracies in absolute ages (Westerhold et al., 2024). Because our intent is to compare the magnitude of trends and variability within each
 165 record against the simulations rather than aligning records temporally to one another, we do not consider this a limitation.

PSU Solver constrains temperature, salinity and $\delta^{18}\text{O}_\text{sw}$ from paired Mg/Ca– $\delta^{18}\text{O}_\text{c}$ measurements by solving, at each time step, a system of proxy equations simultaneously within a bootstrap Monte-Carlo framework that propagates analytical, calibration, sea-level and age uncertainties into probabilistic reconstructions (Thirumalai et al., 2016). At each iteration, we solve for
 170 temperature (T), salinity (S), and $\delta^{18}\text{O}_\text{sw}$ from three coupled relationships: the Bemis et al. (1998) $\delta^{18}\text{O}_\text{c}$ –T equation, the Gray and Evans (2019) temperature- and salinity-dependent Mg/Ca calibration for *T. sacculifer*, and a local, simulation-derived $\delta^{18}\text{O}_\text{sw}$ –salinity relationship; $\delta^{18}\text{O}_\text{sw}$ was also corrected for changes in global ice volume (Rohling et al., 2021). Calibration coefficients, analytical errors ($2\sigma = 0.16\text{‰}$ for $\delta^{18}\text{O}_\text{c}$ and 3‰ mmol mol^{−1} for Mg/Ca), and associated inputs below were obtained from normal distributions; we report the median and its 16th–84th-percentile range from $n = 3000$ iterations. We
 175 diagnose local $\delta^{18}\text{O}_\text{sw}$ –salinity relationships by ordinary least-squares regression of monthly iCESM1.2 $\delta^{18}\text{O}_\text{sw}$ against salinity at each site (Table S1). Regressions were calculated separately for WARM-on and WARM-off, and for the 0–50 and 50–80 m intervals ($n = 4,800$ monthly values). The primary inversion used WARM-on at 50–80 m, consistent with the assumed thermocline habitat of *T. sacculifer*, whereas the other combinations were retained as sensitivity experiments. These relationships were compared with the modern regional regressions of LeGrande and Schmidt (2006): $\delta^{18}\text{O}_\text{sw} = 0.27S - 8.88$
 180 for the tropical Pacific sites and $\delta^{18}\text{O}_\text{sw} = 0.16S - 5.31$ for the Indian Ocean sites. To isolate variability at timescales shorter than eccentricity and obliquity, each irregularly sampled reconstruction was interpolated and filtered using a zero-phase Fourier high-pass filter with cutoff periods of 100 and 41 kyr. Filtered values were interpolated back to the observed ages before calculating range and sample standard deviation. We required at least four median-spaced observations per cutoff period; otherwise, the statistic was designated as unresolved.

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Overall, our implementation follows the Thirumalai et al. (2016) framework but adapts it to the middle Miocene in four respects. First, the Mg/Ca procedure uses the salinity-dependent (Gray and Evans, 2019) calibration. Second, we corrected



each Mg/Ca measurement for the secular evolution of seawater Mg/Ca (Tierney et al., 2019). Third, we removed the ice-volume component of $\delta^{18}\text{O}_{\text{sw}}$ using the Neogene sea-level reconstruction of Rohling et al. (2021), and report the residual, ice-volume-corrected $\delta^{18}\text{O}_{\text{sw}}$ (Table 2). Fourth, in place of modern $\delta^{18}\text{O}_{\text{sw}}$ –SSS relationships (e.g., LeGrande and Schmidt, 2006), we prescribe site-specific $\delta^{18}\text{O}_{\text{sw}}$ –salinity regressions obtained directly from the iCESM1.2 simulations at each core location (Table S1, Fig. S3), so that the inversion is constrained by a self-consistent, physically simulated hydrographic relationship. We additionally assess the ice-volume term by comparing reconstructions computed with and without it, isolating the fraction of reconstructed $\delta^{18}\text{O}_{\text{sw}}$ attributable to changes in ice volume versus local hydrography. For each record, we then compute the overall range and standard deviation of the reconstructed, ice-volume-corrected $\delta^{18}\text{O}_{\text{sw}}$, and subsequently compare these with equivalent statistics derived from the ΔGOC , ΔCO_2 , and $\Delta\text{GOC_CO}_2$ simulation scenarios (Sect. 2.1, Table 1). Monthly forward-modeled $\delta^{18}\text{O}_\text{c}$ were concatenated to obtain an 800-year-long time timeseries with 9,600 time steps for each scenario. Simulated values at each site location were extracted from the nearest grid point using paleo-coordinates at 14 Ma, retrieved using pyGplates (Mather et al., 2024) and the plate rotation model from Seton et al. (2012) (Table S1). The outcomes of this exercise are reported in Tables 2 and S1.

2.4. Construction of synthetic distributions of individual foraminiferal analyses

The geochemistry of planktic foraminiferal tests integrates the temperature, salinity, and $\delta^{18}\text{O}_{\text{sw}}$ of the upper ocean in which they form, and we use forward-modeling of these simulated fields (Sect. 2.2) to evaluate how ΔGOC and ΔCO_2 are expressed in $\delta^{18}\text{O}_\text{c}$ and Mg/Ca proxy systems. Published middle Miocene records rest exclusively on multi-test analyses (hereafter referred to as “bulk” analysis), in which many tests are pooled and measured together; the resulting values approximate a long-term mean state, time-averaged over the sampled stratigraphic interval and further smoothed by bioturbation. Individual foraminiferal analyses (IFA), by contrast, resolve signals spanning the lifespan of foraminifera (Ganssen et al., 2011; Koutavas et al., 2006; Thirumalai et al., 2013). Because each planktic foraminiferal test calcifies over a few weeks to months (Bijma et al., 1990; Fehrenbacher et al., 2024), an IFA population samples the distribution of conditions experienced by numerous individuals rather than an averaged value, and the ensuing distributional statistics can be used to track seasonal or interannual variability, whichever dominates at a given site (Leduc et al., 2009; Thirumalai et al., 2024, 2025). We therefore examine both the mean-state fingerprint relevant to existing bulk records (Sect. 3.2.1) and the variability fingerprint that IFA would be able to access (Sect. 3.2.2). To evaluate how simulated upper-ocean variability could be expressed in IFA distributions, we constructed idealized synthetic $\delta^{18}\text{O}_\text{c}$ distributions from the monthly, forward-modeled proxy values at each grid cell and for both the 0–50 m and 50–80 m depth intervals. For each GOC state, the distribution comprised the monthly values from the final 400 years of the simulation, yielding 4,800 modeled observations per grid cell and depth interval. Each monthly value was weighted equally, and no explicit representation of species-specific seasonality, calcification-depth variability, sediment mixing, or differential preservation was applied (Bienzobas Montávez et al., 2024; Dolman and Laepple, 2018).



Whereas the $\sigma S/\sigma I$ metrics above use the full monthly populations without added noise, to explore how these distributions would be sampled by IFA measurements across GOC states, we applied a bootstrap resampling procedure, following the quantile-based treatment of IFA introduced by Ford et al., (2015). In each of 1,000 iterations, we “picked” 50 monthly values at random, without replacement—reflecting the potential geochemical analysis of 50 individual tests—and added Gaussian analytical uncertainty (1σ) of $\pm 0.08\text{ ‰}$ to $\delta^{18}\text{O}_c$ and $\pm 1\text{ ‰}$ to Mg/Ca (Thirumalai et al. 2013; Glaubke et al. 2021). The 50 “picked” values were then summarized into 25 quantile bins (i.e., reflecting two pseudo-measurements per quantile), and the bin means were taken as the sample quantiles (Glaubke et al. 2021). Quantile–quantile (Q–Q) lines were constructed for the WARM-on and WARM-off quantiles from independent picks, and we report the median across 1,000 iterations as the central Q–Q relationship and show the full ensemble of realizations as the uncertainty envelope (grey lines in Fig. 6 and Supplementary Fig. S5).

3 Results

3.1 Tropical upper ocean response to GOC change

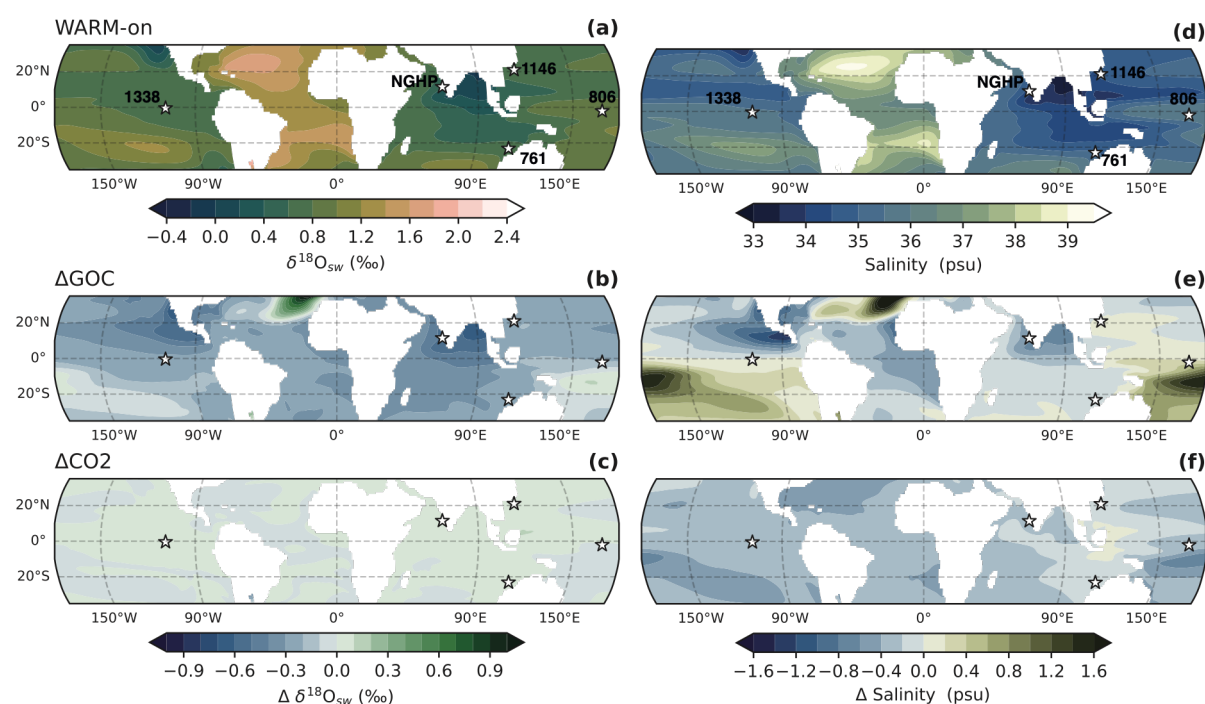


Figure 1. Mean change in simulated $\delta^{18}\text{O}_{sw}$ (‰) (a-c) and salinity (psu) (d-f), averaged between 0-50m. Absolute for WARM-on (right); Difference for ΔGOC (middle) and ΔCO_2 (right) scenarios. Stars indicate the location of drilling sites with published records (see Sect. 2.3) – ODP 806, ODP 1146, ODP 761, NGHP-01-01A and IODP 1336.



We first evaluate how a change in the AMOC equilibrium state (ΔGOC) affects mean upper-ocean characteristics ($\delta^{18}\text{O}_{\text{sw}}$, salinity, and temperature). The $\delta^{18}\text{O}_{\text{sw}}$ decreases across most of the tropics (0.3 to 0.6 ‰ on average), with a more muted response in the south-central Pacific under ΔGOC (Fig. 1a-b). The largest decreases (> 0.5 ‰) occur in the Bay of Bengal and the East Pacific Warm Pool (EPWP), regions characterized by relatively depleted $^{18}\text{O}_{\text{sw}}$ (Fig. 1a-b). Salinity changes broadly parallel the $\delta^{18}\text{O}_{\text{sw}}$ response. Much of the Northern Hemisphere tropics freshens by 0.5 to 1 psu, while the southern tropical Atlantic freshens by approximately 0.4 to 0.6 psu. These regions largely overlap with areas of decreasing $\delta^{18}\text{O}_{\text{sw}}$ (Fig. 1d-e). In contrast, salinity increases in the south-central tropical Pacific, where changes in $\delta^{18}\text{O}_{\text{sw}}$ remain small. This region is notable as a site where subtropical waters subduct and enter the subtropical cells (Qu et al., 2013; Qu and Gao, 2017). Interestingly, as the $\delta^{18}\text{O}_{\text{sw}}$ changes do not totally overlap with those of salinity, the SSS- $\delta^{18}\text{O}_{\text{sw}}$ relationship is not stationary at the local scale (Fig. 1b, e and Fig. S3). This decoupling has important implications for deconvolving and interpreting paired foraminiferal proxy records.

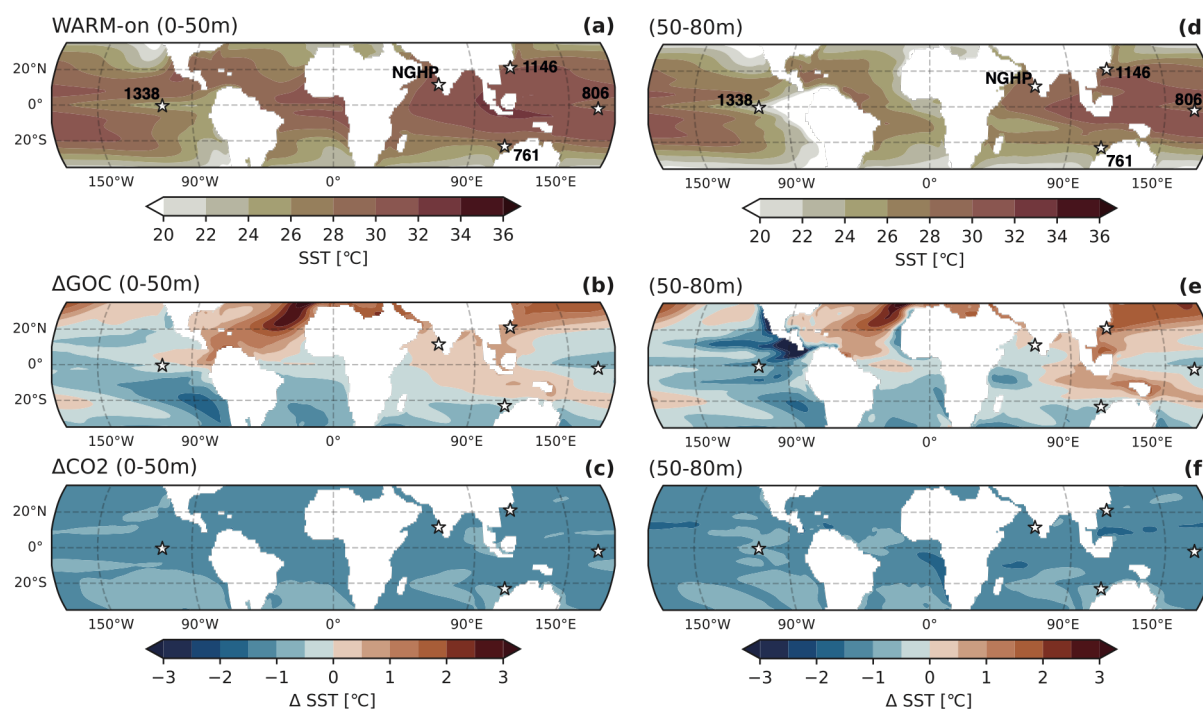


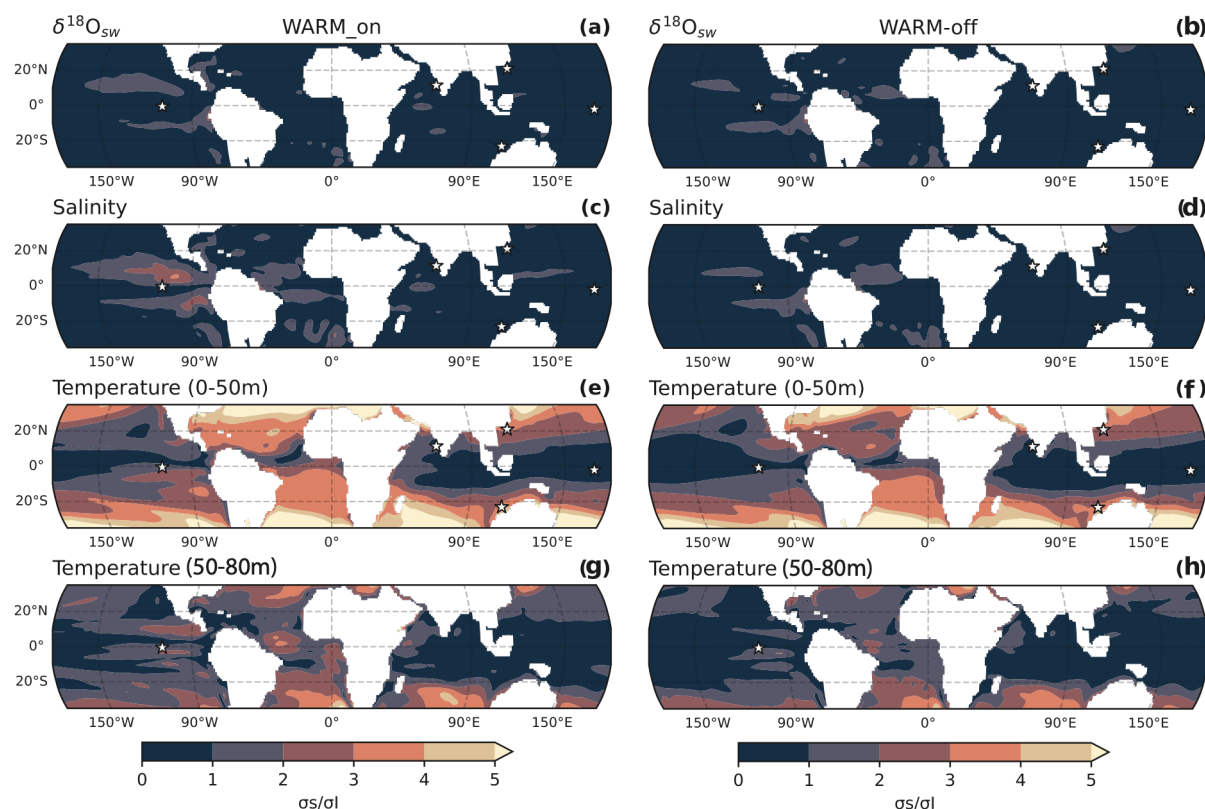
Figure 2. Mean change in simulated temperature (°C), averaged between 0 and 50 meters (a-c) and 50 and 80 meters (d-f). Absolute for WARM-on (right); Difference for ΔGOC (middle) and ΔCO_2 (right) scenarios. Stars indicate the location of drilling sites with published records (see Sect. 2.3) – ODP 806, ODP 1146, ODP 761, NGHP-01-01A and IODP 1336.

Temperature responses to ΔGOC are spatially heterogeneous and vary with depth (Fig. 2a-b, d-e), with patterns that differ from the $\delta^{18}\text{O}_{\text{sw}}$ and salinity changes. In the upper 0-50 m, temperatures increase by 2-3 °C in the tropical North Atlantic and by < 0.5 °C in the Indo-Pacific Warm Pool (IPWP) but decrease by 1-2 °C in the central and southern tropical East Pacific (Fig.



260 2b). At 50–80 m depth, temperature changes are more intense, with patterns diverging from those at the surface outside the tropical Atlantic and the southern tropical Pacific (Fig. 2b, e). A strong cooling is simulated in the EPWP ($>5^{\circ}\text{C}$), and the IPWP warms by 1 to 2°C (Fig. 2e), indicating a change in water column stratification in those regions.

The response to radiative cooling (ΔCO_2) is distinctly different. Changes in upper ocean $\delta^{18}\text{O}_{\text{sw}}$ are small, generally less than 0.05‰ (Fig. 1c), while salinity decreases modestly across much of the tropics. The largest freshening (0.4 to 0.6 psu) occurs in the south-central tropical Pacific, in the same region that exhibits a strong salinity response under ΔGOC (Fig. 1f). In contrast to ΔGOC , ΔCO_2 produces more vertically coherent cooling. Temperatures decrease by approximately $0.5\text{--}2^{\circ}\text{C}$, with similar magnitudes of cooling in the 0–50 m and 50–80 m intervals, implying comparatively little change in upper-ocean thermal stratification (Fig. 2c, f).



270

Figure 3. Ratio of seasonal variability (σS) over inter-annual variability (σI) for $\delta^{18}\text{O}_{\text{sw}}$ (‰) (a–b) and salinity (c–d), averaged between 0–50m, and temperature ($^{\circ}\text{C}$), averaged between 0 and 50 meters (e–f) and 50 and 80 meters (g–h) for WARM-on and WARM-off. Stars indicate the location of drilling sites with published bulk foraminifera records (see Sect. 2.3).



275 Beyond its effects on the mean state, a change in AMOC equilibrium also reorganizes seasonal and interannual variability of upper-ocean properties, providing an additional means of distinguishing ΔGOC and ΔCO_2 . Both modes of variability are governed by the background thermal structure of the tropical ocean, including zonal and meridional SST gradients, the depth and slope of the thermocline, and the coupled ocean–atmosphere feedback these support (An and Jin, 2001; An and Wang, 2000; Fedorov and Philander, 2000). Changes in the mean state can therefore alter the relative importance of seasonal and interannual variability. We quantify this effect using the ratio of seasonal to interannual standard deviations ($\sigma\text{S}/\sigma\text{I}$). Values near or below unity indicate that interannual variability is comparable to or greater than seasonal variability. In the upper 0–50 m, variability in both $\delta^{18}\text{O}_{\text{sw}}$ and salinity is dominated by interannual variability across a substantial portion of the tropics ($\sigma\text{S}/\sigma\text{I} \leq 1$) (Fig. 3a-d). At 50–80 m, the hydrographic pattern is largely unchanged.

285 Upper-ocean temperature variability shows a similar first-order pattern. Temperatures are dominated by interannual variations across the western and central tropical Pacific and the Indian Ocean, with a larger seasonal contribution in the tropical Atlantic and the Pacific cold tongue (Fig. 3e-f). Notably, seasonal variability increases when the AMOC is active (*-on* cases), as is apparent in the Atlantic basin (Fig. 3e). At 50–80 m, the temperature variability is more strongly interannual, with year-to-year fluctuations accounting for nearly the entire spread (Fig. 3g-h). This behaviour highlights how thermocline-depth variability can drive subsurface temperature changes. When the thermocline is dynamically active, its interannual shifts dominate temperature variance, reducing the influence of seasonal changes at depth.

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3.2 Fingerprint on synthetic bulk versus individual foraminiferal analyses

3.2.1 Bulk foraminiferal records

295 Under the ΔGOC scenario, mean upper-ocean $\delta^{18}\text{O}_\text{c}$ decreases by as much as 0.7 ‰ across much of the tropical region, with the main exceptions occurring in the Pacific Cold Tongue and South-Central Pacific, where $\delta^{18}\text{O}_\text{c}$ either slightly increases or remains stable (Fig. 4a-b). Local changes in $\delta^{18}\text{O}_\text{c}$ reflect a combination of temperature and $\delta^{18}\text{O}_{\text{sw}}$ changes, but the relative contribution of these factors varies regionally (Fig. 1b and 4e). In the Indian Ocean and EPWP, for example, the change in $\delta^{18}\text{O}$ reflects seawater ^{18}O depletion (Fig. 1b and 4b) associated with changes in atmospheric freshwater fluxes. In other regions, temperature changes act to either amplify the $\delta^{18}\text{O}_\text{c}$ signal, as in the North Equatorial Atlantic; partially offset it, as in the South Equatorial Atlantic; or dominate it altogether, as in the Pacific Tongue (Fig. 1b, d). Changes in $\delta^{18}\text{O}_\text{c}$ at 50-80 m are almost similar to 0-50 m, except in the EPWP where the influence from temperature increases (Fig. 4c, f). Because surface $\delta^{18}\text{O}_\text{c}$ at the surface retains a strong hydrographic imprint under the ΔGOC scenario, its spatial pattern differs substantially from that of Mg/Ca (Fig. 4g-h). The latter increases at the surface (0-50 m) by 0.5 mmol mol⁻¹ in the IPWP and by >1 mmol mol⁻¹ in the tropical North Atlantic, while decreasing by up to 0.5 mmol mol⁻¹ in other parts of the tropical Pacific and Atlantic basins. These changes largely track the simulated temperature response (Fig. 2b). At 50-80 m depth, Mg/Ca anomalies are larger and continue to closely follow the subsurface temperature changes (Fig. 2b).

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The ΔCO_2 scenario generates a drastically different $\delta^{18}\text{O}_c$ response, characterized by an increase across all the tropical basins (Fig. S4) and only minor changes in the vertical $\delta^{18}\text{O}_c$ gradient. In this case, the isotopic response is driven almost entirely by ocean cooling (Fig. 2c, f and S4), as changes in $\delta^{18}\text{O}_{sw}$ are approximately an order of magnitude smaller than under ΔGOC (Fig. 1c). Consequently, $\delta^{18}\text{O}_c$ and Mg/Ca exhibit largely coherent spatial patterns, with both proxy responses primarily reflecting the underlying temperature change (Fig. S4).

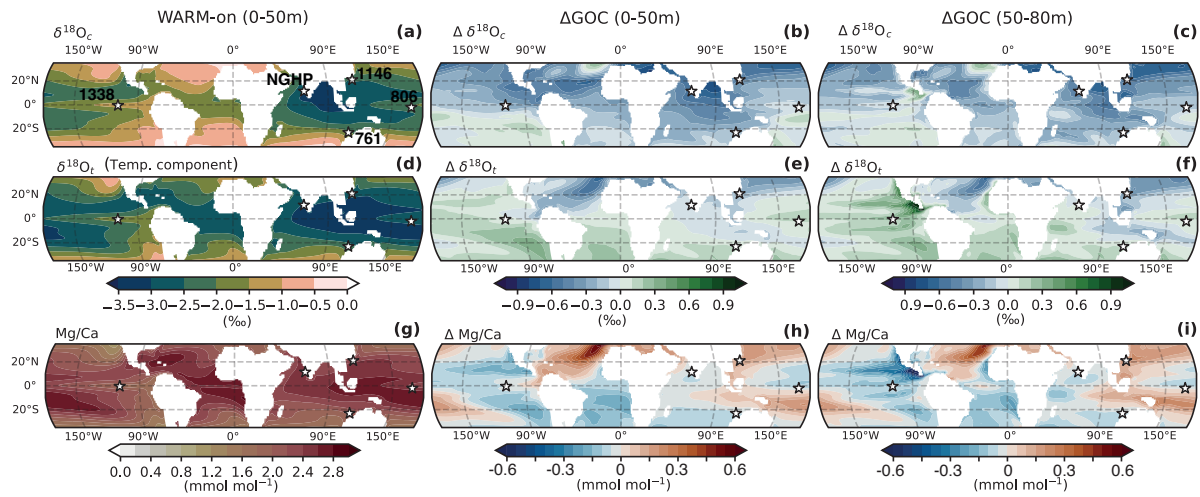


Figure 4. Mean forward-modeled $\delta^{18}\text{O}_c$ (‰) (a) and Mg/Ca (mmol mol^{-1}) (g) for WARM-on (averaged between 0-50m, left). Mean forward-modeled $\delta^{18}\text{O}_c$ (a-c), temperature contribution to $\delta^{18}\text{O}_c$ ($\delta^{18}\text{O}_t$) (d-f), and Mg/Ca (g-i) for the ΔGOC scenario, averaged between 0 and 50 meters (middle) and 50 and 80 meters (right). Note that Mg/Ca values have been corrected for Miocene Mg/Ca_{sw} (see sect. 2.2). Stars indicate the location of drilling sites with published bulk foraminifera records (see Sect. 2.3) – ODP806, ODP 1146, ODP 761, NGHP-01 and IODP 1336.

3.2.2 GOC-dependent changes in simulated IFA distributions

The ΔGOC scenario alters not only mean upper-ocean conditions but also their seasonal and interannual variability, changes that could potentially be captured by IFA records (Fig. 5). Because no IFA records spanning the MMCT are presently available, we use forward-modelled monthly $\delta^{18}\text{O}_c$ distributions to identify the regions and depth habitats where this variability fingerprint is most clearly expressed (Sect. 2.4). The spatial patterns of $\sigma\text{S}/\sigma\text{I}$ are broadly similar for Mg/Ca and $\delta^{18}\text{O}_c$, consistent with changes in temperature variability providing the dominant contribution to the modelled proxy variance (Fig. 3). In both GOC states, $\sigma\text{S}/\sigma\text{I}$ remains at or below unity across much of the Indo-Pacific Warm Pool (IPWP), indicating a comparatively large interannual contribution, whereas values greater than unity in the southern tropical Atlantic indicate a stronger seasonal control (Fig. 5). The synthetic $\delta^{18}\text{O}_c$ distributions (Fig. 6) illustrate how these contrasting regimes of variance would appear in IFA populations at representative IPWP (4°S, 122°E) and southern tropical Atlantic (7°S, 3°W) sites (Fig. 5).



At both sites and depth intervals, WARM-off values are shifted toward lower $\delta^{18}\text{O}_c$ relative to WARM-on, producing detectable changes in the centres of the distributions (Fig. 6a-b, e-f). Additionally, the response of the interquartile ranges, tails, and degree of asymmetry also differs between the two GOC states. In the IPWP, particularly at 50–80 m, the WARM-off distribution develops a pronounced tail toward higher $\delta^{18}\text{O}_c$, whereas the WARM-on distribution is more compact and comparatively symmetric (Fig. 6b). The southern tropical Atlantic distributions are broader, consistent with the stronger seasonal contribution at this location, and the subsurface WARM-off case displays an extended tail toward lower $\delta^{18}\text{O}_c$ (Fig. 6f). Depth therefore modifies the distributional response, with the 50–80 m interval generally showing more prominent tail behaviour than the 0–50 m interval (Fig. 6a-b, e-f). These results indicate that synthetic IFA can distinguish the modelled GOC states through changes in distribution shape as well as through changes in mean $\delta^{18}\text{O}_c$, although the diagnostic expression is region- and habitat-depth dependent.

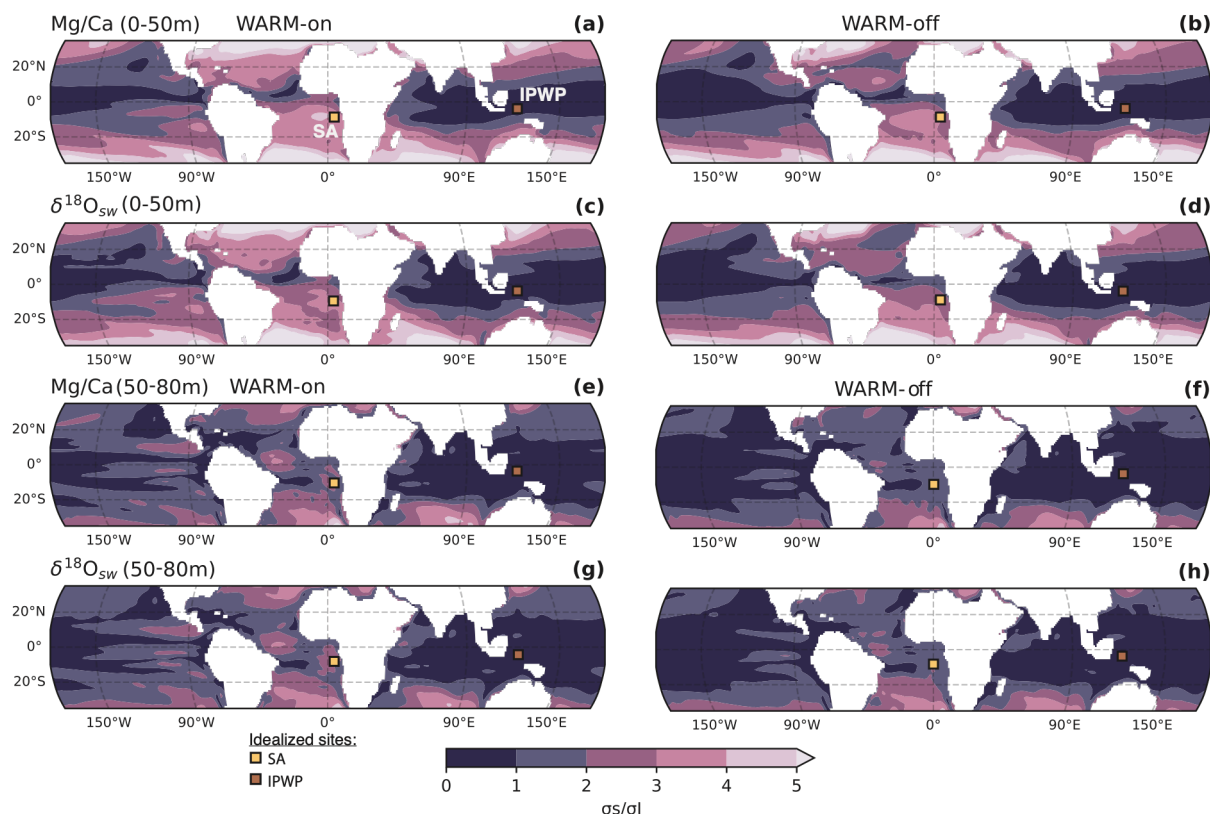


Figure 5. $\sigma\delta/\sigma I$ for forward-modelled Mg/Ca (a-b, e-f) and $\delta^{18}\text{O}_c$ (c-d, g-h) (averaged between 0 and 50 meters (top) and 50 and 80 meters (bottom)) for WARM-on (left) and WARM-off (right). Squares indicate the location of sampled grid cells for the synthetic records shown in Fig. 6.

We investigate this contrast explicitly for both proxy systems in Fig. 6 ($\delta^{18}\text{O}_c$) and Fig. S5 (Mg/Ca), which compare the WARM-on and WARM-off IFA populations at the IPWP and SA using quantile–quantile (Q–Q) plots (Ford et al., 2015). In



the IFA- $\delta^{18}\text{O}_c$ distributions, the WARM-on populations are consistently lower than WARM-off at both sites and both depths, so that the mean offset dominates the Q-Q structure (Fig. 6c-d, g-h). In IFA-Mg/Ca, the mean offset is minor and reverses depending on the location (Fig. S5c-d, g-h); WARM-on is marginally warmer than WARM-off at the IPWP but cooler at SA. On the other hand, the change in population statistics is similar to that of IFA- $\delta^{18}\text{O}_c$ distributions at each site and depth. The $\delta^{18}\text{O}_c$ Q-Q relationships are quasi-parallel to the 1:1 line at the surface-ocean depths, indicating a near-uniform mean offset across the distributions. For 50–80 m in the IPWP, the WARM-on distribution is narrower and less cold-tailed than the WARM-off distribution (Fig. 6d). The Mg/Ca Q-Q relationships take on similar shapes at each site and depth; however, they lie on the opposite side of the 1:1 line at SA relative to the IPWP, consistent with a cooler WARM-on state there (Fig. 2d). As a result, the bulk (population-mean) difference between GOC states, approximating a conventional pooled measurement, is larger and of consistent sign for $\delta^{18}\text{O}_c$ but small and reverses sign regionally for Mg/Ca. Accordingly, bulk $\delta^{18}\text{O}_c$ and bulk Mg/Ca records would imply different—and at the SA site even opposite—changes across GOC states.

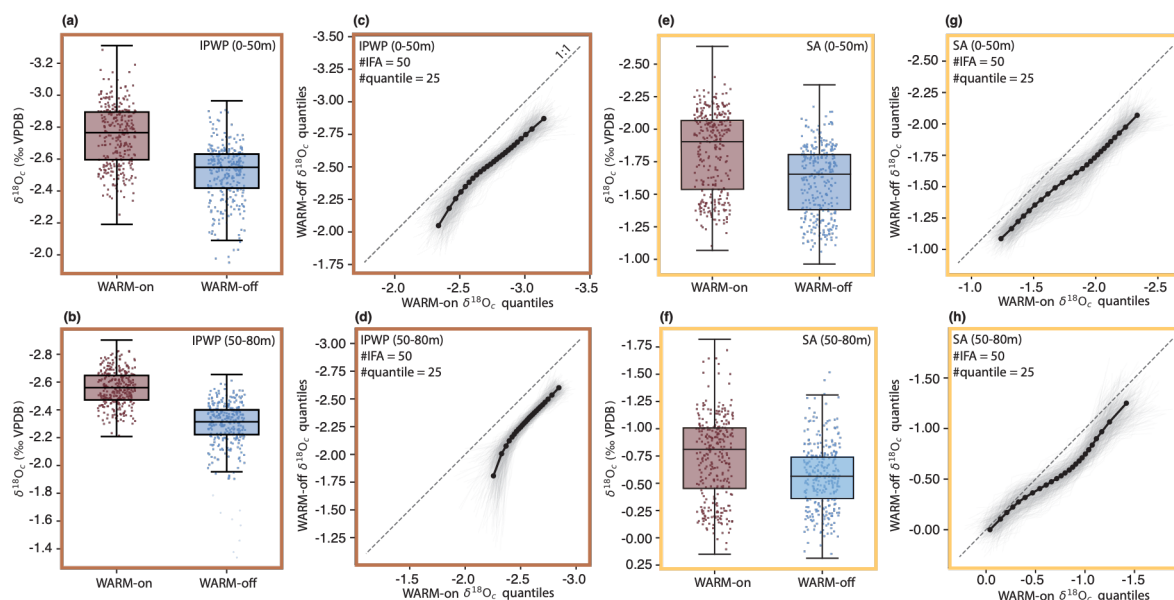


Figure 6. Forward-modelled individual foraminiferal analyses of $\delta^{18}\text{O}$ (IFA- $\delta^{18}\text{O}_c$) distributions for WARM-on versus WARM-off scenarios at sites in the Indo-Pacific Warm Pool (IPWP—4°S, 122°E; a–d) and the southern tropical Atlantic (SA—7°S, 3°W; e–h), for 0–50 m (a, c, e, g) and 50–80 m (b, d, f, h) depth intervals. $\delta^{18}\text{O}_c$ is forward-modelled from monthly iCESM1.2 temperature and $\delta^{18}\text{O}_{\text{sw}}$ values with the Bemis et al. (1998) low-light equation (Sect. 2.2). Box-and-swarm panels (a, b, e, f) show monthly distributions (box: median and interquartile range; points: a 300-value subsample of the 4,800 monthly values). Quantile–quantile (Q-Q) panels (c, d, g, h) compare WARM-on against WARM-off from a bootstrap IFA procedure (1,000 iterations, 50 “picked” IFA per iteration, 25 quantiles, ± 0.08 ‰ analytical uncertainty—see Methods for details): the black line depicts the median Q-Q relationship, grey



370 lines the individual bootstrap realizations, and the grey dashed line is the 1:1 line. All $\delta^{18}\text{O}_c$ axes are inverted so that warmer/lower values plot upward (box plots) and toward the upper right (Q-Q plots). Coloured contours refer to the site locations in Fig. 5.

3.3 Mechanisms for regional response to GOC perturbation

Climate simulations also enable mechanistic analysis of simulated ocean characteristic changes, which can provide insights into regional mechanisms at play. The heterogeneity of mean temperature and hydrography changes (Fig. 1 and 2) responsible for the patterns in proxy systems in ΔGOC (Fig. 4) reflects the restructuring of tropical ocean-atmosphere circulation and thermocline structure in response to change in AMOC state, as already well described in the literature for more recent time periods (e.g., Broccoli et al., 2006; Chang et al., 2008; DiNezio et al., 2025; Zhang and Delworth, 2005). In this scenario, AMOC invigoration (*-on* cases) strongly increases northward oceanic heat transport, resulting in North Atlantic warming (Lee et al., 2025), which triggers a readjustment of the tropical circulation system. As a response to changes in sea level pressure gradients, easterlies weaken in the northern tropical Atlantic and east of 120°W in the northern tropical Pacific and strengthen at other tropical locations (Fig. 7a). The weaker wind stress also decreases evaporative heat loss in the tropical North Atlantic, contributing to warmer SST locally (Fig. 2b and S6 and S7a-b), which maintains weak winds, consistent with the SST-wind-evaporation feedback (Chiang and Bitz, 2005). The wind-stress anomaly expands across Central America (Fig. 7a), in line with the Atlantic-Pacific atmospheric bridge (Xie et al., 2008) and explains the warmer SST also at the surface of the Eastern Pacific Warm Pool (EPWP) (Fig. 2h). The stronger easterlies west of 120°W (Fig. 5a), on the other side, push the warm tropical water westward, shifting the IPWP westward (Figure 2 and S7a-b), and deepening the thermocline in the Western Pacific. This contributes to warmer temperatures both at the surface and sub-surface (Fig. 2b, e). In the Western Pacific, the change in surface circulation (Fig. 7b) also modifies the proportion of (cooler) extratropical waters that enter the western boundary current around 25°S (Fig. S8), while the circulation in the Indonesian Throughflow intensifies, similar to Hu et al. (2021). As a result, more warm water is transported from the Pacific to the Indian Ocean and contributes to warming the equatorial region, around the Maritime Continent (Fig. 2b, e). At the same time, increased easterlies south of the equator trigger stronger upwelling along South American coastlines (Fig. 7a), resulting in cooler temperatures in the East Equatorial Pacific (EEP) cold tongue (Fig. 2b, e).

395 In response to the increased zonal SST gradient in the equatorial Pacific in WARM-*on*, the Indo-Pacific Walker cell is shifted westward, and convection is more intense over the eastern Indian Ocean and the Maritime Continent (Fig. 7c), increasing precipitation over the ocean (Fig. 7c). Differences in atmospheric circulation result in more ^{18}O -depleted precipitation ($\delta^{18}\text{O}_p$) (Fig. 7d), which explains the lower surface $\delta^{18}\text{O}_{sw}$ in the northern Indian Ocean (Bay of Bengal) (-0.5 to -0.8‰ , Fig. 1b). Precipitation is also higher over land due to a more intense monsoon (Fig. 7c), which likely increases coastal runoff with an ^{18}O -depleted signature that is then redistributed by surface currents. As the atmospheric circulation shifts westward, the convection weakens over the central Pacific, resulting in lower freshwater flux and a more saline water mass (Fig. 1e & 7c). In the EPWP, the great change in temperature and hydrography arises from a combination of regional ocean-atmosphere



circulation and atmosphere fluxes that modify the upper ocean structure (Fig. S7g-h). The region lies below the ITCZ (Fig. 7c) and therefore already has a lower salinity relative to adjacent places when the AMOC is active (*-on* cases) (~ 34 psu) (Fig. 1d and S7d). In the *WARM-off* case, atmospheric convection is weakened over the EPWP (Fig. 7c), likely due to decreased convergence (Duffy et al., 2020; Liu et al., 2024), resulting in decreased precipitation over the ocean and an enrichment in ^{18}O (Fig. 7d). This difference in ocean-atmosphere dynamics is associated to a salinity maximum (~ 37 psu) around 20°N , mirroring the southern hemisphere (Fig. S7c). The presence of this warm and salty water mass along the Central American coastlines, which extends to depth, and the thick mixed layer (Fig. S7g) prevent cold water from upwelling near the surface. This difference in upper ocean dynamics explains the strong temperature signal below the surface in ΔGOC (Fig. 2e), but also why in this region the anomaly in $\delta^{18}\text{O}_{\text{sw}}$ and salinity expand at depth, with pattern that mimics that of temperature (Fig. 1b, d, e).

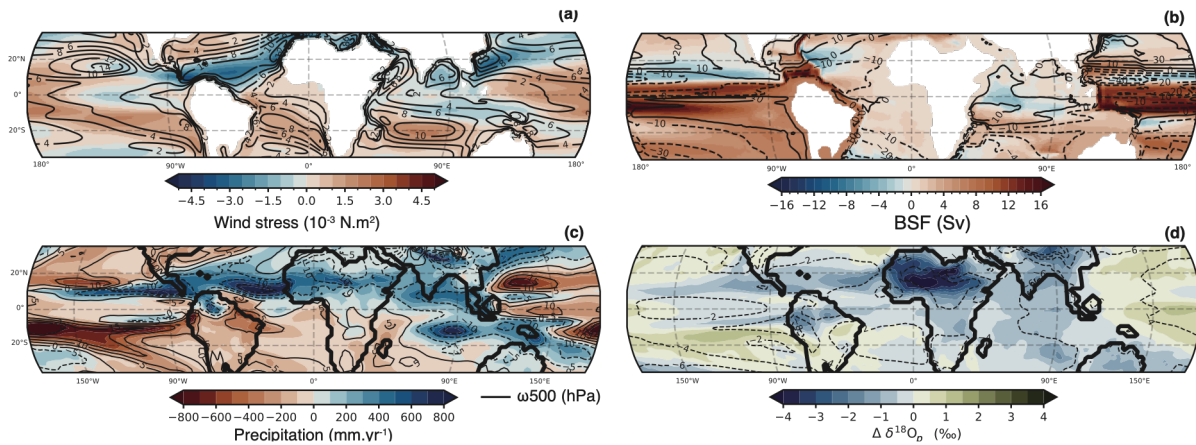


Figure 7. Ocean-atmosphere dynamics difference between *WARM-on* and *WARM-off* (ΔGOC). (a) wind stress anomaly at the surface (m.s^{-1} , coloured background) and absolute value for *WARM-off* (contour), (b) barotropic stream function anomaly (Sv, background colour) and absolute value for *WARM-off* (contour), (c) precipitation (mm.day^{-1} , background colour) and $\omega 500$ (hPa, contour) anomalies, and (d) amount-weighted isotopic composition of precipitation anomaly ($\delta^{18}\text{O}_p$, ‰) and absolute $\delta^{18}\text{O}_p$ for *WARM-off* (contour). Model outputs are averaged over the last 100 years of each simulation.

4 Discussion

4.1 On the importance of reconstructing $\delta^{18}\text{O}_{\text{sw}}$

Our simulations show that AMOC state change (“dynamic effect”) generates a perturbation of $\delta^{18}\text{O}_{\text{sw}}$ in the upper tropical ocean that is an order of magnitude higher than the one generated by the change in radiative forcing due to pCO_2 decrease (ΔCO_2) (“thermodynamic” effect) (Fig. 1b-c). Importantly, in the ΔCO_2 scenario, the intensity of the intermediate cell of the AMOC increases (Fig. S2d, f) without noticeable changes in the $\delta^{18}\text{O}_{\text{sw}}$ (Fig. 1c). In this scenario, the large-scale AMOC pattern remains stable (Figure S2a, c-d, f), which confirms that significant changes in upper ocean $\delta^{18}\text{O}_{\text{sw}}$ are tracking the restructuring of GOC rather than the linear intensification of the pre-existing pattern. This is because the upper ocean-atmosphere tropical circulation readjusts to drastic changes in global heat transport in ΔGOC , leading to a new mean state,



while cooling alone does not strongly disrupt the first-order circulation patterns. The tropical current system, in ΔCO_2 , efficiently redistributes the anomaly in the atmospheric isotope fluxes ($\delta^{18}\text{O}_p$) related to the intensified hydrological cycling, coherent with results obtained for $p\text{CO}_2$ sensitivity experiments for other deep-time periods (e.g., Sarr et al., 2025; Zhu et al., 2020). On the other hand, the regional temperature is similarly sensitive to both “dynamic” and “thermodynamic” changes (albeit that the patterns in ΔGOC are much more spatially heterogeneous (Fig. 2b-c). This has important implications for understanding climate forcing using surface temperature proxy systems, especially when sampling is sparse, which could lead to biases in assessing the heterogeneity of temperature patterns, which we illustrate using our scenarios (see Sect. 2.1). For example, interpreting the decrease in $\delta^{18}\text{O}_c$ signal from ΔGOC scenario (Fig. 4b-c) without accounting for the change in $\delta^{18}\text{O}_{\text{sw}}$ (Fig. 1b-c) would suggest a low-latitude warming in most of the basins. Such interpretations would not be coherent with other temperature proxy systems like Mg/Ca (Fig. 4h), $\Delta 47$ or TEX_{86} that would likely show a cooling; providing those are not impacted by other non-thermal factors.

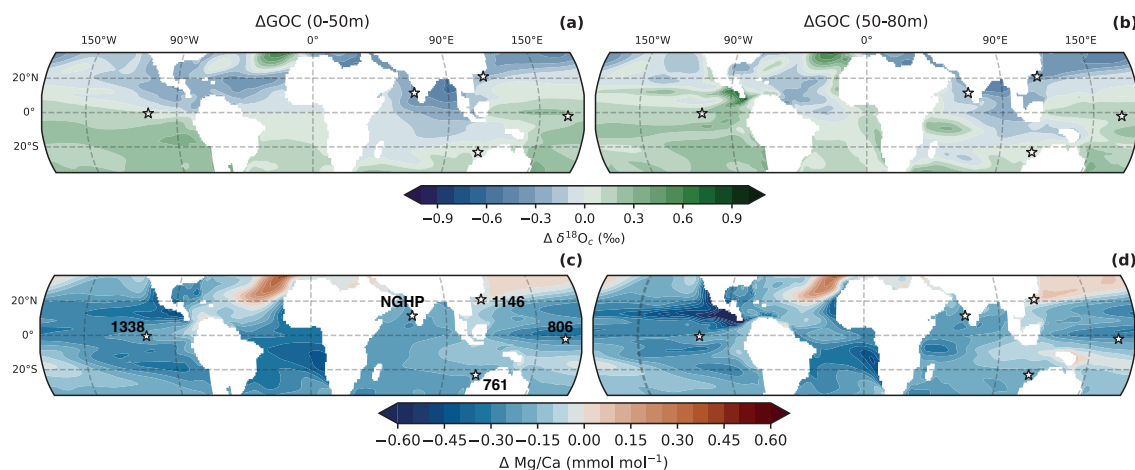


Figure 8. Mean forward-modelled $\delta^{18}\text{O}_c$ (‰) (a-b) and Mg/Ca (mmol mol^{-1}) (c-d) for $\Delta\text{GOC_CO}_2$, averaged between 0 and 50 meters (left) and 50 and 80 meters (right). Note that Mg/Ca values have been corrected for Miocene Mg/Ca_{sw} (see sect. 2.2). Stars indicate the location of drilling sites with published bulk foraminifera records (see Sect. 2.3) – ODP 806, ODP 1146, ODP 761, NGHP-01-01A and IODP U1336.

Another case could be that AMOC reorganization is coeval with carbon cycle perturbation, either because a change in $p\text{CO}_2$ would be triggering the ocean circulation change (e.g., Zhang et al., 2017) or because of carbon cycle feedback (e.g., Nian et al., 2026). To explore this, we generated a $\Delta\text{GOC_CO}_2$ scenario that compares COLD-on with WARM-off simulations (Sect. 2.1, Table 1), mimicking a 120 ppm drop on $p\text{CO}_2$ alongside GOC switching-on. In that case, the 0.5 to 2 °C cooling effect due to $p\text{CO}_2$ forcing (that enriched $^{18}\text{O}_c$) acts to compensate the change in $\delta^{18}\text{O}_{\text{sw}}$ associated to the change in AMOC mean state (Fig. 8a-b). The only locations where the $\delta^{18}\text{O}_{\text{sw}}$ change is still clearly visible as a local depletion in $^{18}\text{O}_c$ are the Bay of Bengal

and the EPWP (Fig. 8a-b); and at such places the $\delta^{18}\text{O}_c$ signal could be interpreted as a local warming. In that case, the cooling inferred from Mg/Ca reconstruction could, on the other hand, be over- (e.g., EPWP) or under-estimated (e.g., IPWP) depending on site location (Figure 8c-d), as the “dynamic” effect on upper ocean temperature would either overwhelm or over-amplify the “thermodynamic” effect.

Overall, our results show that long-term $\delta^{18}\text{O}_{sw}$ is a key parameter to reconstruct that allows for the deconvolution of “dynamic” global forcing vs “thermodynamic” forcing associated with $p\text{CO}_2$ variations, which paleo-thermometers alone cannot do. Relying on $\delta^{18}\text{O}_{sw}$ reconstructions would both prevent misinterpretation of global radiative forcing effects on temperature and climate sensitivity and help identify major transitions of the global ocean circulation that could support the interpretation of paleo-circulation tracers.

4.2 Statistical data–model comparison and targets for future records

We use the simulations to evaluate whether a change in the AMOC equilibrium state could explain trends depicted in previously published MMCT tropical bulk foraminifera records. As our modelling results suggest that $\delta^{18}\text{O}_{sw}$ responds more strongly to a reorganization of the AMOC than to the imposed decrease in $p\text{CO}_2$, we focus on reconstructed $\delta^{18}\text{O}_{sw}$ rather than raw $\delta^{18}\text{O}_c$ to enable the identification of “dynamic” versus “thermodynamic” drivers for regional changes (Fig. 1 and 4). A conventional point-to-point comparison is not appropriate here because the records differ in duration, sampling resolution, age uncertainty, and sedimentary reworking (Bienzobas Montávez et al., 2024), whereas the simulations represent equilibrated climate states rather than a transient MMCT chronology. Averaging every record over a common fixed interval would therefore make the comparison sensitive to record resolution and age-model selection. Instead, we compare the range and standard deviation of each reconstruction with those of the three modelled scenarios (Table 1). For each scenario, the monthly time series from its two constituent simulations were concatenated, producing an 800-year series containing 9,600 monthly values (Sect. 2.3). This approach tests whether the magnitude of reconstructed hydrographic variability is consistent with the contrast between modelled climate states, without requiring exact temporal alignment between the records and simulations.

Table 2. $\delta^{18}\text{O}_{sw}$ statistics for existing tropical site locations (Figure 1a and Sect. 2.3). Values are extracted for proxy records (Sect. 2.3) and ΔGOC , ΔCO_2 and $\Delta\text{GOC_CO}_2$ scenarios (Sect. 2.1). r = range, std = standard deviation. LS06 : using SSS- $\delta^{18}\text{O}_{sw}$ relationship from LeGrande and Schmidt (2006); CESM : using simulated SSS- $\delta^{18}\text{O}_{sw}$ relationships; OrbF : 100-kyr high-pass filtered.

Scenario	ODP 1146		ODP 806		ODP 761		IODP U1338		NGHP-01	
	r	std	r	std	r	std	r	std	r	std
Data LS06	1.18	0.21	1.20	0.28	1.66	0.41	1.28	0.22	1.84	0.47
Data CESM	1.10	0.18	1.07	0.25	1.64	0.41	1.48	0.26	2.04	0.53



Data OrbF	0.18	0.18	-	-	0.32	0.06	0.56	0.12	-	-
Model ΔCO_2	0.10	0.05	0.05	0.03	0.12	0.07	0.07	0.03	0.28	0.14
Model ΔGOC	0.29	0.13	0.20	0.1	0.33	0.16	0.3	0.14	0.64	0.28
Model $\Delta\text{GOC_CO}_2$	0.27	0.12	0.22	0.1	0.28	0.13	0.25	0.12	0.60	0.27

At all five site locations, simulated $\delta^{18}\text{O}_{\text{sw}}$ variability in ΔGOC and $\Delta\text{GOC_CO}_2$ scenarios is systematically higher than in the ΔCO_2 scenario (Table 2), with standard deviation almost doubling in the former cases. For example, the simulated standard deviation increases from 0.05 ‰ under ΔCO_2 to 0.12 ‰ under ΔGOC at ODP 1146, from 0.07 ‰ to 0.16 ‰ at ODP 761, from 0.07 ‰ to 0.30 ‰ at U1338, from 0.14 ‰ to 0.28 ‰ at NGHP-01-01A, and from 0.05 ‰ to 0.20 ‰ at ODP 806. The reconstructed standard deviations are generally larger, ranging from 0.21‰ at ODP 1146 to 0.47‰ at NGHP-01-01A (Table 2, LS06). Thus, a change in GOC state produces a substantial fraction of the observed $\delta^{18}\text{O}_{\text{sw}}$ spread and consistently yields more variability than radiative cooling alone, but it does not reproduce the full reconstructed variability at every site. The remaining difference is expected to include variability absent from the equilibrium simulations, including orbital forcing (e.g., Holbourn et al., 2005), shorter-timescale climate fluctuations (e.g., Schmidt et al., 2007), proxy, age, and calibration uncertainty (Thirumalai et al., 2016) and sedimentary processes (Bienzobas Montávez et al., 2024). Moreover, concatenating two equilibrated simulations represents the contrast between end-member states rather than the rate or trajectory of the transition. The comparison therefore supports major GOC reorganization as a contributor to the reconstructed hydrographic variability but does not uniquely attribute all observed variability to that mechanism.

The choice of $\delta^{18}\text{O}_{\text{sw}}$ –salinity relationship affects the reconstructed variability most strongly at ODP 1146, U1338, and NGHP-01-01A (Table 1, CESM), where the model-derived relationships differ substantially from modern basin-scale calibrations (Table S1). We also evaluated how the choice of $\delta^{18}\text{O}_{\text{ice_volume}}$ reconstructions affects the variability depicted in the record, by removing the ice-volume correction and note that $\delta^{18}\text{O}_{\text{sw}}$ inversion without correcting for $\delta^{18}\text{O}_{\text{ice_volume}}$ have a larger standard deviation. Uncertainty in the evolution and isotopic composition of continental ice therefore limits the fraction of local $\delta^{18}\text{O}_{\text{sw}}$ variability that can be assigned to ocean–atmosphere circulation. Improved constraints on $\delta^{18}\text{O}_{\text{ice_volume}}$, seawater Mg/Ca, and the chronology of individual records will provide for more quantitative discrimination among the simulated scenarios. After removing variability at eccentricity and longer periods with the 100-kyr high-pass filter (Table 2, OrbF), the reconstructed standard deviations at ODP 1146 and U1338 more closely approach those simulated for the ΔMOC experiment. ODP 761 is an important exception: its filtered variability is smaller and more consistent with the ΔCO_2 experiment, whereas ODP 806 and NGHP-01-01A are too sparsely sampled for a reliable, filtered comparison. Thus, the agreement with a circulation-driven response is spatially heterogeneous rather than universal. A change in the equilibrium state of the AMOC could generate this pattern by reorganizing tropical water-mass pathways, upper-ocean stratification, and atmosphere–ocean moisture transport (Sect 3.3). These changes alter regional precipitation, evaporation, runoff, and thermocline structure, producing relatively large $\delta^{18}\text{O}_{\text{sw}}$ responses at some sites but muted responses at others. The closer match between the eccentricity-filtered reconstructions



and ΔMOC at ODP 1146 and U1338 supports that such GOC-threshold mechanisms may in part explain the observed MMCT hydrographic variability.

515 Our results can also be used to target future study sites for investigating middle Miocene ocean-atmosphere evolution. Figure 1b shows that reconstructing long-term $\delta^{18}\text{O}_{\text{sw}}$ in areas located below atmospheric convection centres, in the northern tropics like the Northern Indian Ocean or the EPWP (Fig. 7c), could yield crucial information on abnormal hydrographic changes that could be associated with drastic restructuring of ocean-atmosphere circulation patterns. On the other hand, reconstructing surface temperature at sites located both in the northern and southern tropical Atlantic would help inform temperature changes
 520 that are characteristic of AMOC perturbation (Fig. 4h-i). Generating a reconstruction from multiple species with different dwelling depths jointly in the IPWP or the Equatorial Atlantic and the EPWP to reconstruct vertical gradients would enable tracking long-term change in the upper ocean structure (Fig. 2b,e). Idealized synthetic IFA distributions further indicate that individual-shell measurements could complement bulk reconstructions by testing changes in variance and distribution shape (Fig. 6). Sites in the Indo-Pacific warm pool are suited to examining signals with a large interannual contribution, whereas the
 525 southern tropical Atlantic provides a contrasting regime with stronger seasonal variability. These predictions should be tested using species-specific forward models that account for habitat depth, seasonal production, finite shell counts, analytical uncertainty, bioturbation, and preservation (Bienzobas Montávez et al., 2024; Glaubke et al., 2021). With these qualifications, coordinated bulk, multispecies, and individual-foraminiferal records offer a means to evaluate both the mean-state and variability fingerprints of a middle Miocene GOC reorganization.

530 The two proxy systems assimilate GOC reorganizations differently, which is also reflected distinctly in the bulk-versus-IFA approaches (Fig. 6; Fig. S5). For IFA- $\delta^{18}\text{O}_{\text{c}}$ distributions, the population means (i.e., approximating a bulk measurement) are more influenced by $\delta^{18}\text{O}_{\text{sw}}$ rather than thermal changes. Accordingly, the distributions are lower under WARM-on, consistently in sign across both sites and depths and most pronounced in the IPWP, where the subsurface signal is stronger because the
 535 hydrographic signal is more pronounced than temperature change (Fig. 1b, 2b). The mean offset, rather than changes in distributional statistics, dominates the Q-Q separation. This suggests that even a modest IFA population is capable of registering changes in local hydrography caused by ocean circulation states. The mean IFA-Mg/Ca offset, by contrast, is driven by temperature changes, even though we incorporate salinity sensitivity for Mg/Ca (see Sect. 2.2). The Mg/Ca-temperature signal is minor and regionally variable, with the two proxy systems agreeing in direction at the IPWP (both warmer
 540 under WARM-on) but diverging at SA, where WARM-on is lower in $\delta^{18}\text{O}_{\text{c}}$ (Fig. 6) yet cooler in Mg/Ca (Fig. S5). On the other hand, the distributional statistics of the population variance behave opposite to the mean: the Q-Q structure across states is similar in both proxy systems, reflecting the local balance of simulated seasonal and interannual temperature variability rather than being driven by the circulation state. We contend that the expectation for IFA is therefore potentially twofold, wherein the mean, bulk-equivalent $\delta^{18}\text{O}_{\text{c}}$ of IFA carries the $\delta^{18}\text{O}_{\text{sw}}$ fingerprint of GOC reorganization, whereas the spread and shape
 545 of that population, like those of co-measured Mg/Ca, record the thermal regime of the site. Paired $\delta^{18}\text{O}_{\text{c}}$ -Mg/Ca IFA



measurements could thus potentially separate the dynamic from the thermodynamic imprint of circulation-driven $\delta^{18}\text{O}_{\text{sw}}$ relative to thermal changes.

550 4.3 Implication for understanding forcing of Miocene climate patterns

Our simulation dataset also informs on mechanisms linking global changes to regional climate patterns (Fig. 7 and Sect. 3.3) and can be used to discuss forcing and feedback in the system. Most of the published studies explain the temperature and $\delta^{18}\text{O}_{\text{sw}}$ signals as well as other biogeochemical changes across the MMCT by different mechanisms such as 1) perturbation of the tropical current system forced by change in trade wind intensity (Fox et al., 2021; Holbourn et al., 2013b, 2014; Nathan and Leckie, 2009; Tian et al., 2014), 2) modification of the thermocline structure across the tropical Pacific (Auderset et al., 2022; Hess et al., 2023; Matsui et al., 2017), or 3) northward migration of the ITCZ and intensification of northern hemisphere tropical monsoons (Fox et al., 2021; Holbourn et al., 2010; Podder et al., 2024; Yang et al., 2020b; Zou et al., 2024). Such regional mechanisms are commonly attributed to Southern Ocean cooling (Leutert et al., 2020, 2021; Shevenell et al., 2004) and Antarctica cryosphere expansion (Levy et al., 2016, 2019; Sangiorgi et al., 2018) forced by $p\text{CO}_2$ drawdown (Raitzsch et al., 2021; Sosdian et al., 2018) that would have increased the meridional temperature gradient, intensified the trade winds and pushed the ITCZ northward. This is by analogy with the mechanism described for the Northern Hemisphere ice sheet during the Pleistocene (Chiang and Bitz, 2005), although the ice area increase related to the Laurentide ice sheet is incomparable to the Antarctic ice sheet, questioning the relevance of such comparison. Miocene simulations with realistic change in Antarctic ice-sheet configuration in addition do not show significant changes in tropical climate dynamics (e.g., Sarr et al., 2022). Another hypothesis brought out by Sosdian and Lear (2020) is that changes in upper ocean characteristics were triggered by seaway bathymetric changes forced by sea-level fall. The latter would have led to restriction of the Indonesian Throughflow, which would have modified the amount of Pacific water entering the Indian Ocean and shifted the IPWP westward, hence modifying regional ocean characteristics. Change in the Indonesian Throughflow circulation has already been used to explain ϵ_{Nd} trends in the tropical Indian ocean around 14 Ma (Gourlan et al., 2008), although there is yet no strong independent evidence of the seaway constriction at that time (e.g., Gallagher et al., 2024; Nathan and Leckie, 2009).

Interestingly, all the regional mechanisms hypothesized in the literature and described in the above paragraph are at play in our ΔGOC scenario. They are, however, not arising as a response to global cooling and growing icesheet (as both are fixed in our experiments), nor are they the result of change in paleogeography (see Sect. 2.1, Table 1). Instead, they represent the regional expression of ocean-atmosphere dynamics reorganization in response to dramatic change in global heat transport and energetic balance related to a shift in AMOC equilibrium state. Importantly, in the ΔGOC scenario, the intensification of Southern Hemisphere meridional gradient arises as a consequence of dynamics changes rather than a cause (Fig. S9), which itself would represent a paradigm shift in the understanding of MMCT forcing and feedback. We note here that a significant shift in AMOC pattern offers a complementary mechanism to help sustain glaciation and expansion of marine-based ice sheet



580 and sea-ice as recorded from the MMCT onward (Chorley et al., 2022; Hoem et al., 2024; Hou et al., 2023; Levy et al., 2016, 2019), by maintaining a cooler Southern Ocean. Growth of the Southern hemisphere cryosphere could in turn have contributed to stabilize circulation patterns *via* sea-ice feedback, implying a Southern Hemisphere control on the overturning circulation.

Our results also invite reinterpretation of Miocene records beyond the MMCT. Changes in ocean circulation are known to
 585 impact ocean biogeochemistry, including oxygenation patterns (e.g., Berg et al., 2026; Duteil et al., 2014, 2020; Schmittner et al., 2007) and nutrient distribution and primary productivity (e.g., Boot et al., 2025). GOC reorganization may provide a plausible mechanisms for some of the long-term changes in OMZ and productivity documented across the middle Miocene (e.g., Auderset et al., 2022; Cortese et al., 2004; Hess et al., 2023, 2026; Holbourn et al., 2014; Lyle and Baldauf, 2015; Nilsson-Kerr et al., 2025). Changes in ocean circulation also reorganize tropical hydroclimate. In our simulations, the
 590 intensification of tropical hydrology and shift in convection centres (Fig. 7c) potentially offer a mechanism to explain abrupt precipitation changes or freshwater influx recorded in monsoon regions (e.g., John et al., 2003; Yang et al., 2020b; Zammit et al., 2022). In addition, sustained change in the mean precipitation field, as simulated in our dataset (Fig. 7c), would affect weathering and should also be better investigated, as the latter is a potential mechanism for $p\text{CO}_2$ drawdown during the Miocene (e.g., Stewart et al., 2017; Wu et al., 2025). Lastly, our simulations show substantial changes in circulation through
 595 the two major tropical seaways, the Central American Seaway and Indonesian Gateway (Fig. S8), even though their geographic and bathymetric configurations remain fixed. This result cautions against attributing changes in gateway circulation solely to tectonic or bathymetric forcing.

4 Conclusions

600 Changes in the equilibrium state of the global ocean overturning circulation are difficult to identify in the geologic record because paleo-circulation tracers are challenging to interpret. By combining isotope-enabled climate simulations with proxy system modeling, we show that GOC reorganization can produce detectable signatures in upper-ocean temperature, hydrography, and variability that are identifiable using upper ocean $\delta^{18}\text{O}_{\text{sw}}$ reconstructions and $\delta^{18}\text{O}_\text{c}$ IFAs and are distinct from signatures from radiatively forced cooling. Our simulations are idealized middle Miocene cases and sample only two
 605 possible equilibrium states of the ocean circulation. Nevertheless, the upper ocean response we simulate is consistent with the general behaviour found in previous studies of overturning-circulation change and is similar, at large scale, to that obtained by in an independent set of Miocene simulations. More broadly, our study highlights that the representation of global ocean circulation is fundamental to simulating regional climate accurately. Given the large spread in ocean circulation in the MioMIP1 ensemble (Naik et al., 2025), more efforts are needed to better understand Miocene circulation patterns and
 610 accordingly refine boundary conditions. Failure to do so will strongly affect the reliability of data-model comparison, preventing an accurate understanding of past climate dynamics. Ultimately, one of the main limitations remains the limited availability of long-term records that are necessary to understand how the system evolved from the warm early Miocene to the



post-MMCT world. High-quality, well-dated, multi-proxy records are currently biased toward the Indo-Pacific region, and further work in the Atlantic basin is now necessary to provide a global view of the ocean.

615 **Code and data availability**

Model outputs, Python scripts used to generate figures, and PSU solver code and inversion are available in Sarr et al. (2026). Code for the water isotope-enabled CESM is available at <https://github.com/NCAR/iCESM1.2>. Raw planktic foraminifera $\delta^{18}\text{O}$ and Mg/Ca data used for the analysis have been extracted from previously published records and can be access directly from following publications supplementary material or data repositories : ODP Sites 761 and 806 (Sosdian, S. and Lear, C., 2021; Sosdian et al., 2020), Site NGHP-01-01A (Steinke et al., 2024; Yang et al., 2020a), IODP Site U1338 and ODP Site 1146 (Fox et al., 2021). Colored figures in this paper were made with perceptually uniform, color-vision-deficiency-friendly scientific color maps, developed and distributed by Fabio Crameri (Crameri et al., 2020). Analysis and figures were performed using Python 3.1 and the xarray, pandas, matplotlib, numpy, xarray, and pyGPlates packages. Claude (Anthropic) was used to help translate the PSU Solver forward model from its original MATLAB implementation (Thirumalai et al. 2016) to Python and to streamline the Python plotting routines; all code and outputs were checked and validated by the authors.

Supplement link

The link to the supplement will be included by Copernicus, if applicable.

Author contributions

ACS and KT conceptualized the study. ACS analyzed the simulation results and designed the figures. KT updated the PSU solver code, re-analyzed the published proxy records, and performed statistical analysis on simulations outputs. RPA and DL provided simulation outputs. CJP acquired funding and computation time. ACS prepared the manuscript with significant contribution from KT. CJP and RPA reviewed and edited the manuscript.

Competing interests

635 The contact author has declared that neither they nor their co-authors have any competing interests.

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