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AQ1 Isolated Northern Hemisphere Cooling in an Idealized Model Promotes Ocean Carbon Storage and Glacial-like Ocean Features

Running Head: SHANKLE ET AL.

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Abstract

Earth's past glacial cycles are paced by orbital insolation changes, but internal climate feedbacks modulating atmospheric carbon dioxide (CO₂) are necessary to explain the full magnitude of glacial–interglacial climate variability. Identifying such feedbacks and linking them to orbitally  paced forcing remains a challenge, impeding

a comprehensive understanding of the glacial cycles. Orbital variations, determining the latitudinal distribution and seasonality of insolation, are primarily important for allowing the buildup of large Northern Hemisphere ice sheets. Meanwhile, the processes canonically understood to drive CO₂ drawdown and mediate ocean–atmosphere carbon partitioning are largely hosted in the far-flung Southern Ocean. This study explores the role of remote Northern Hemisphere temperature forcing in influencing Southern Ocean conditions that impact global ocean circulation and carbon content. In a set of idealized ocean–sea ice model simulations, we find that cooling air temperatures to glacial-like levels at the high northern latitudes (emulating the thermal effect of a large Northern Hemisphere ice sheet) drives a substantial ~~achieves ~74% of the atmospheric~~ CO₂ drawdown of (32 ppm) – or 74% the magnitude of a ~~as a glacial-like~~ simulation forced by globally glacial-like temperatures everywhere (43 ppm). Northern cooling is found to cool not only northern-sourced middepth waters but ultimately also global deep waters, resulting in a strong solubility-driven CO₂ drawdown component. It also drives several characteristic features of glacial periods, including expanded southern sea ice and a colder, saltier, and better stratified deep ocean, which contributes to air–sea CO₂ disequilibrium. These results demonstrate an interhemispheric connection between Northern Hemisphere conditions paced by orbital variations and Southern Hemisphere processes that modulate deep-ocean carbon content.

SIGNIFICANCE STATEMENT

Linking the orbital pacing of the ice ages—primarily expressed by the growth and decay of Northern Hemisphere ice sheets—to processes hosted in the Southern Ocean that can regulate ocean carbon storage has long posed a challenge. Here, we demonstrate using idealized simulations a clear influence of Northern Hemisphere cooling on Southern Ocean conditions and global ocean circulation and carbon content, which in turn draws down carbon dioxide (CO₂) from the atmosphere. This result underscores the importance of conditions in the high northern latitudes in shaping the ocean–atmosphere carbon balance, offering lessons about the interconnectedness of remote regions that are relevant to both past and contemporary climate change.

KEYWORDS:

Southern Ocean; Ocean circulation; Ice age; Carbon dioxide; Paleoclimate; [AQ2](#) Idealized models

1. Introduction

Earth's past glacial cycles are known to be paced by orbitally-induced variations in insolation (Croll 1875) known as Milankovitch cycles (Milankovitch 1930), and yet orbital forcing alone cannot explain the full magnitude of glacial–interglacial climate variation (Hays et al. 1976). Internal feedbacks within the climate system—specifically those regulating atmospheric carbon dioxide (CO₂) concentration—are required to amplify the orbital signal and explain the full magnitude of temperature fluctuations recorded across glacial cycles. Identifying these feedbacks, understanding their interactions, and quantifying their contributions to glacial CO₂ drawdown remain key areas of ongoing research.

Such feedbacks likely involve the sequestration of carbon in the deep ocean, the largest carbon reservoir that exchanges with the atmosphere on glacial time scales (Sigman and Boyle 2000). The Southern Ocean in particular is implicated in these dynamics for ventilating significant volume of the global deep ocean (Talley 2013; Rae and Broecker 2018) and hosting a relatively inefficient biological pump that may have strengthened during glacial periods (Sigman and Boyle 2000, 2001). However, the mechanism(s) linking orbitally-based insolation forcing (primarily influencing the growth and decay of Northern Hemisphere ice sheets) to processes hosted in the far-flung Southern Ocean understood to govern ocean carbon storage remains unclear. A comprehensive theory of glacial–interglacial change must link Northern Hemisphere-based insolation to Southern Ocean carbon dynamics.

Proxy records over the last glacial cycle (over the last ~120 000 years before the present) suggest such an interhemispheric connection (Fig. 1). Northern Hemisphere ice sheet growth is paced by variations in local summer insolation (Figs. 1a,b), expanding with each insolation minima when reduced summer insolation limits melt back. Once grown to sufficient altitude and area, ice sheets maintain positive mass balance through cold temperatures at elevation and increased albedo, providing a positive feedbacks with which to maintain mass through subsequent insolation maxima and producing the characteristic “sawtooth” pattern of the decline into glacial conditions (Fig. 1b). Southern Ocean changes are also documented in glacial proxy data, such as changing sea ice conditions (Fig. 1d) which would influence air–sea gas exchange and alter bottom-water properties via brine rejection during ice formation. However, such changes (alongside atmospheric CO₂; Fig. 1c) more closely

resemble the canonical sawtooth pattern of northern ice sheet growth than local Southern Hemisphere insolation (Fig. 1e), despite sea ice lacking the thermal inertia and feedbacks of large continental ice sheets to survive through local insolation maxima. This points to some mechanism linking the hemispheres, imposing the sawtooth evolution of Northern Hemisphere ice sheets onto Southern Ocean processes regulating ocean–atmosphere carbon exchange and, in turn, atmospheric CO₂.

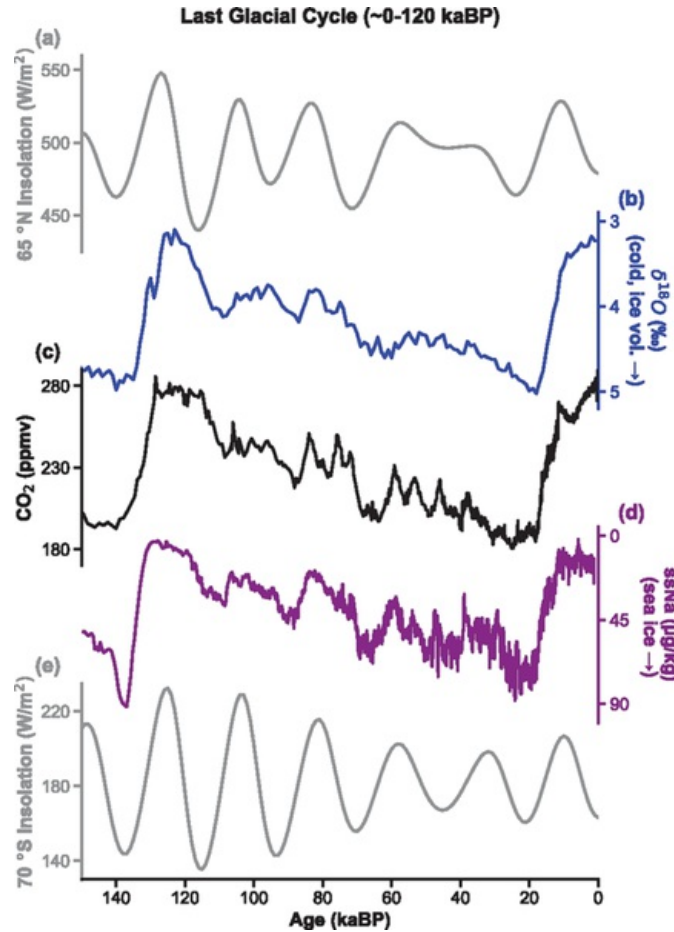


Fig. 1. Canonical proxy records of the last glacial cycle [~120 kaBP (1000 years before the present) to the present]. (a) Average daily Northern Hemisphere summertime (21 June) insolation at 65°N (W m^{-2}), important for determining the growth or decay of Northern Hemisphere ice sheets. (b) Global stack of benthic foraminiferal oxygen isotope composition ($\delta^{18}\text{O}$, per mille), reflecting deep-ocean temperature and continental ice sheet volume (Lisiecki and Raymo 2005). (c) Ice core gas bubble atmospheric CO_2 concentration (ppmv) (Bereiter et al. 2015). (d) Antarctic ice core (EPICA Dronning Maud Land) sea salt sodium concentration (ssNa; $\mu\text{g kg}^{-1}$), reflecting sea salt aerosols and sea ice production (Fischer et al. 2007). (e) Average daily Southern Hemisphere springtime (27 Sep) insolation at 70°S (W m^{-2}), reflecting the season of maximum ice sea ice extent. (a),(e) Average daily insolation was calculated using a MATLAB script (see data availability statement).

One prevailing hypothesis for such a link considers the influence of North Atlantic Deep Water (NADW) and its properties on Antarctic sea ice (Adkins 2013; Gildor and Tziperman 2001). It postulates that Northern

Hemisphere glaciation would cool NADW and downstream Circumpolar Deep Water (CDW) upwelling in the Southern Ocean. This in turn would promote southern sea ice growth and initiate a series of associated changes facilitating ocean carbon storage, including expanded sea ice cover, saltier southern-sourced bottom water, and enhanced stratification and reduced vertical mixing between the surface and deep waters. Gildor and Tziperman (2001) originally tested this in a coupled physical–biogeochemical box model, showing ~75-ppm CO₂ drawdown in response to orbitally–forced northern ice sheet growth (Gildor and Tziperman 2001).

We revisit this idea in idealized ocean–sea ice simulations, emulating the thermal effect of Northern Hemisphere ice sheets by cooling the high northern latitudes and examining resulting changes in ocean circulation and carbon content. Idealized simulations are particularly well suited to this question, as they allow us to isolate and quantify the specific effect of cooling the northern deep-water formation region—the central feature of the Gildor and Tziperman hypothesis—without the confounding influences that may arise in more complex model setups. The present setup also naturally builds upon the work of previous studies which investigated the ocean’s response to a global cooling of air temperatures to glacial-like levels, which reproduced several key features of the glacial ocean such as shoaled NADW and more isolated stratified deep ocean (Jansen 2017) and also enhanced deep-ocean carbon storage (Marzocchi and Jansen 2019). We build upon this work and revisit the Gildor and Tziperman (2001) hypothesis by testing how much of this signal is reproduced by cooling only the high northern latitudes.

2. Methods

a. Model description

Here, we use a configuration of the Massachusetts Institute of Technology General Circulation Model (MITgcm) (Marshall et al. 1997) that was previously employed to explore key features of ocean circulation and carbon storage under preindustrial and glacial-like temperature forcing (Jansen 2017; Marzocchi and Jansen 2019). The model includes coupled ocean, sea ice (Losch et al. 2010), and biogeochemistry (Dutkiewicz et al. 2005) modules. While the model does not simulate a Northern Hemisphere ice sheet directly, the thermal effect of an ice sheet is emulated by adjusting atmospheric temperatures. The model spans 70°S–70°N over a single basin (1° × 1° resolution), with a reentrant channel at 64°–69°S. It has 29 vertical layers (varying from 20 to 200 m thick) and a uniform depth of 4 km, except for a 3-km sill at the channel’s entrance. The ocean is coupled to a well-mixed atmospheric box (Parekh et al. 2006) such that the total mass of carbon is conserved between the atmosphere and ocean; however, atmospheric CO₂ is not radiatively active.

Diapycnal mixing follows a vertical profile varying from $\sim 2 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ in the thermocline to $\sim 2 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ in the deep ocean. Mesoscale eddies are represented using Gent–McWilliams and Redi parameterizations (Gent and McWilliams 1990; Redi 1982) with a variable eddy diffusivity (Visbeck et al. 1997). CO_2 air–sea exchange is parameterized with a gas transfer coefficient proportional to the square of the local wind speed (Wanninkhof 1992).

The model is forced by static 2D fields of surface (2-m) air temperature, 10-m zonal and meridional wind speed, evaporation, precipitation (rain and snow), and longwave radiation. Wind speed fields are as given in Jansen (2017) plus 6.5 m s^{-1} gustiness (Jansen 2017). Sea surface temperatures are restored via surface heat exchange to the prescribed air temperature profiles. Salt fluxes are determined from a prescribed zonally symmetric evaporation-minus-precipitation profile. A no-flux bottom boundary condition is applied.

b. Experimental design

Here, we explore the potential impact of a large Northern Hemisphere ice sheet on ocean circulation and biogeochemistry by comparing model responses to different idealized air temperature forcings. Specifically, we impose cool temperatures in the high northern latitudes to emulate the thermal effect of an ice sheet. We compare two “northern cooling” simulations against an interglacial-like control simulation and also include a simulation run under global glacial-like cooling to provide a benchmark of glacial-like changes.

The control simulation (“IG_Ctrl”) was forced with a zonally symmetric, preindustrial-like air temperature profile (Fig. 2a). Then, two experimental “northern cooling” simulations were forced by interglacial-like temperature profiles except with cooler temperatures imposed at the high northern latitudes, emulating the thermal impact of a large ice sheet (Figs. 2c,d). In one, temperatures north of 40.5°N were fully cooled (100%) to glacial-like levels (“IG_N100”). In the other, temperatures north of 40.5°N were set to be exactly halfway between interglacial- and glacial-like temperatures (“IG_N50,” representing a cooling of the northern latitudes “50%” of the way to glacial-like levels). Finally, a fully glacial-like reference simulation (“Glac_Ref”) was run, forced by global glacial-like temperatures (Fig. 2b). The threshold latitude of 40.5°N was selected to ensure full coverage of the region of northern dense water formation. In all simulations, the forcing is static in time; thus, there is no seasonal cycle, and the air temperature profiles in Fig. 2 represent annual-mean profiles.

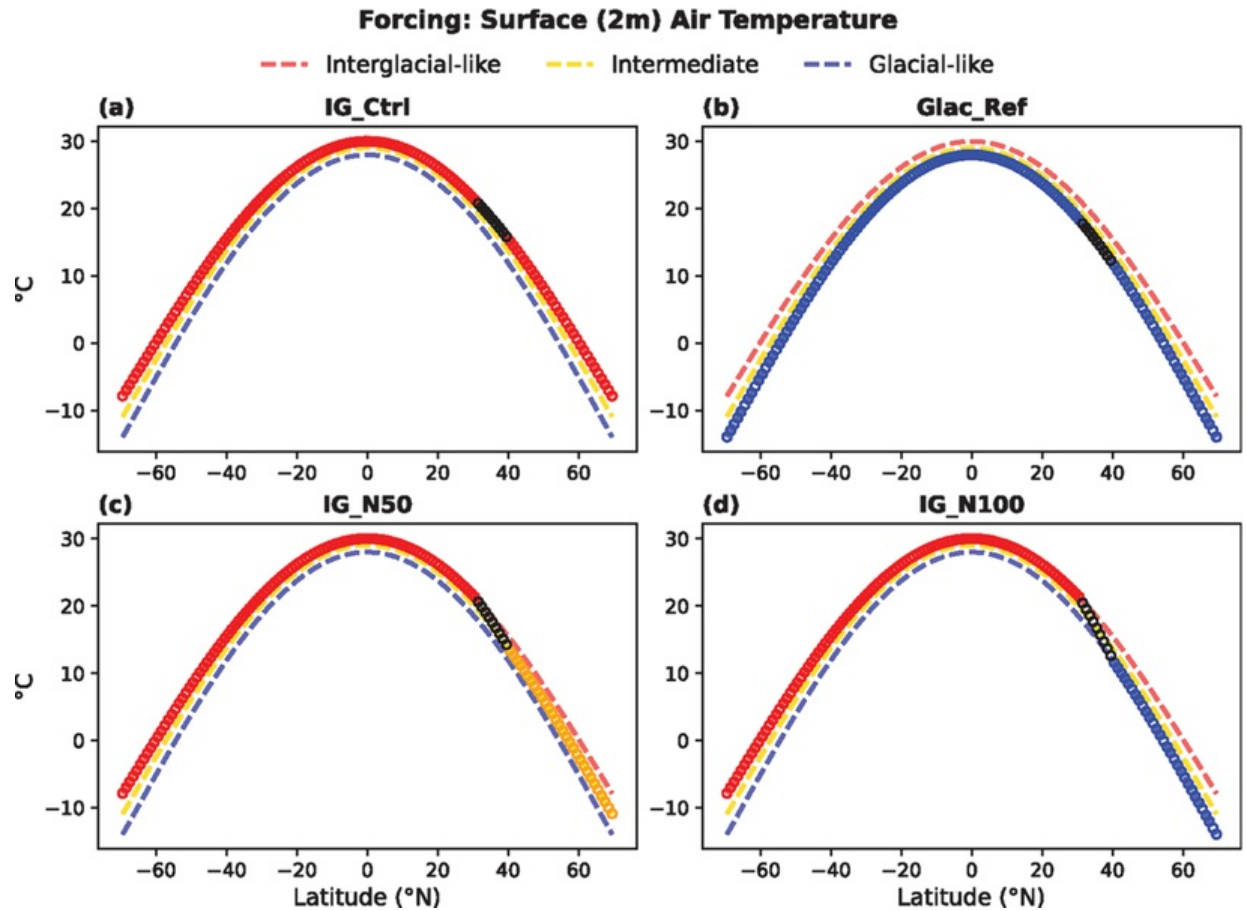


Fig. 2. Surface (2-m) air temperature forcing. Latitudinal profiles of surface (2-m) air temperature (°C) used to force (a) the interglacial-like control simulation (IG_Ctrl), (b) the glacial-like reference simulation (Glac_Ref), and (c),(d) the two northern cooling simulations with varying degrees of cooling in the north (IG_N50 and IG_N100). Temperatures are zonally symmetric and static in time. In (c) IG_N50 and (d) IG_N100, cooler temperatures are applied above 40.5°N; temperatures transition smoothly between glacial- and interglacial-like temperature profiles between 30.5° and 40.5°N (shown by black points).

This glacial-like profile is 2°C cooler at the equator and 6°C cooler at the poles than the interglacial-like profile, as per [Marzocchi and Jansen \(2019\)](#) and in line with recent estimates of glacial cooling from paleoproxy data-assimilated reanalysis products ([Annan et al. 2022](#); [Osman et al. 2021](#)). This glacial-like reference simulation provides a baseline against which to quantify the impact of northern cooling alone on ocean circulation and biogeochemistry. All simulations were run to equilibrium (15 000 years of simulation time), and averages over the last 500 years are compared between simulations.

c. Carbon pump decomposition

Simulated tracers of preformed phosphate and alkalinity (i.e., the phosphate/alkalinity concentration transported into the ocean interior from surface waters, without modification by remineralization) are used to decompose the ocean dissolved inorganic carbon (DIC) inventory into its various components (Ito and Follows 2013; Lauderdale et al. 2013; Ito and Follows 2005)—i.e., carbon associated with the potential saturation of surface waters with the atmosphere (C_{sat}), the soft- and hard-tissue (carbonate) biological pumps (C_{soft} , C_{carb}), and the disequilibrium pump (C_{dis}). These components are calculated as in Marzocchi and Jansen (2019)— C_{sat} from carbonate chemistry equations, C_{soft} and C_{carb} from remineralized nutrient and alkalinity levels, and C_{dis} as the residual of these from total DIC.

The change in atmospheric CO_2 associated with each carbon pump (i.e., associated with changes in the ocean inventory of each of the four DIC components described above) is derived as follows. First, the total ocean–atmosphere inventory of inorganic carbon is computed from the whole-ocean mean concentration of each DIC component (mol m^{-3}) and the model’s molar atmospheric volume (M , 4.9×10^{19} mol) and ocean volume (V , $3.6 \times 10^{17} \text{ m}^3$):

$$\Sigma C = Mp_{\text{CO}_2} + V [C_{\text{sat}} + C_{\text{dis}} + C_{\text{soft}} + C_{\text{carb}}], \quad (1)$$

where ΣC is approximately constant (ignoring the negligible contribution of organic carbon to the total carbon budget). Note that, given the idealized nature of our configuration, we omit the reduction in ocean volume and associated salinity changes caused by sea-level fall during glacial ice sheet expansion (Adkins et al. 2002).

Changes in atmospheric $p\text{CO}_2$ (e.g., between two simulations) can then be expressed in terms of the changes in the total ocean inventory of each DIC component:

$$\delta p_{\text{CO}_2} = -\frac{V}{M} [\delta C_{\text{sat}} + \delta C_{\text{dis}} + \delta C_{\text{soft}} + \delta C_{\text{carb}}]. \quad (2)$$

Positive changes (accumulation) in any component of DIC in the ocean thus drive negative change in $p\text{CO}_2$ (drawdown from the atmosphere), while negative changes (depletion) drive increases in $p\text{CO}_2$.

The δC_{sat} between two simulations can be further decomposed (ignoring nonlinear behavior) into the

contributions made from changes in temperature, salinity, preformed alkalinity, and atmospheric $p\text{CO}_2$ between simulations as:

$$\delta C_{\text{sat}} \approx \delta C_T + \delta C_S + \delta C_A + \delta C_{\text{CO}_2}, \quad (3)$$

where:

$$\delta C_T = \frac{\partial C_{\text{sat}}}{\partial T} \delta T, \delta C_S = \frac{\partial C_{\text{sat}}}{\partial S} \delta S, \delta C_A = \frac{\partial C_{\text{sat}}}{\partial A} \delta A, \delta C_{\text{CO}_2} = \frac{\partial C_{\text{sat}}}{\partial p\text{CO}_2} \delta p\text{CO}_2. \quad (4)$$

The partial derivatives in these terms represent the sensitivity of C_{sat} to each of the stated variables. In our simulations, these sensitivities vary minimally across the simulations and compare well with literature values (Williams and Follows 2011), giving us confidence in the model's representation of the carbonate chemistry system. In practice, we compute δC_T , and δC_S , etc. directly from finite differences between simulation end-members, which gives equivalent results. The δC_T , δC_S , δC_A , or δC_{CO_2} between two simulations is computed as the difference between the whole-ocean mean C_{sat} in a given simulation (e.g., the IG_Ctrl simulation) and that when C_{sat} has been calculated replacing one variable (temperature, salinity, alkalinity, or atmospheric $p\text{CO}_2$) with the field from a second simulation of interest (e.g., any of the cooling simulations).

Combining Eqs. (2) and (3) gives:

$$\delta p\text{CO}_2 = \alpha [\delta C_{\text{sol}} + \delta C_{\text{dis}} + \delta C_{\text{bio}}], \quad (5)$$

where $\alpha = [M/V + \partial C_{\text{sat}} / \partial p\text{CO}_2]^{-1}$, $\delta C_{\text{sol}} = \delta C_T + \delta C_S$ denotes the change associated with the solubility pump via changes in temperature and salinity, δC_{dis} denotes the change associated with the disequilibrium pump, and $\delta C_{\text{bio}} = \delta C_{\text{soft}} + \delta C_{\text{carb}} + \delta C_A$ denotes the net change associated with the biological pump. While δC_A reflects change in solubility (via changes in preformed alkalinity), it is assigned to the biological pump since, in our model setup (which lacks for example sediment fluxes and river fluxes of alkalinity), only biological production can alter the whole-ocean balance of preformed-versus-regenerated alkalinity and drive a significant signal in the whole-ocean mean δC_A . Dilution/concentration associated with salinity changes can modify the spatial distribution of preformed alkalinity but do not affect the global ocean preformed alkalinity inventory.

The contributions of each of the solubility, disequilibrium, and biological pumps to atmospheric $p\text{CO}_2$ change in our simulations can then be computed as:

$$(\delta p\text{CO}_2)_{\text{sol}} = \alpha \delta C_{\text{sol}}, (\delta p\text{CO}_2)_{\text{dis}} = \alpha \delta C_{\text{dis}}, (\delta p\text{CO}_2)_{\text{bio}} = \alpha \delta C_{\text{bio}}. \quad (6)$$

From Eq. (5), α can be estimated as $\alpha \sim \delta p\text{CO}_2 / (\delta C_{\text{sol}} + \delta C_{\text{dis}} + \delta C_{\text{bio}})$ [values of α vary minimally between this and its formulation in Eq. (5), and this choice does not meaningfully change the results]. Thus, $p\text{CO}_2$ drawdown driven by a specific pump is proportional to the total $p\text{CO}_2$ change between two simulations ($\delta p\text{CO}_2$) scaled by that DIC component's contribution (δC_{sol} , δC_{dis} , or δC_{bio}) to the total ocean carbon change ($\delta C_{\text{sol}} + \delta C_{\text{dis}} + \delta C_{\text{bio}}$). An overview of all terms and their changes across the simulations is given in Table S1 in the online supplemental material.

3. Results

The model reproduces interglacial- and glacial-like ocean states under the respective temperature forcings (IG_Ctrl and Glac_Ref), with realistic sea surface temperatures and sea ice latitudinal extents (Fig. S1) compared to present-day observational products (Huang et al. 2025; Fetterer et al. 2025) and proxy reconstructions (Osman et al. 2021; Crosta et al. 2022). Glac_Ref also captures key glacial ocean features, including salinity-driven deep stratification (Adkins et al. 2002) (Figs. S2 and S3) and a weakened Atlantic meridional overturning circulation (AMOC) (Pöppelmeier et al. 2023a; Blaser et al. 2025) (Fig. 3, Table S2). Simulations where only high northern latitudes were cooled (IG_N50, IG_N100) also show partial development of these glacial-like features. We thus compare simulations with and without high northern-latitude cooling to study the impact of Northern Hemisphere ice sheet-related cooling on basinwide ocean circulation and biogeochemistry. Anomalies are presented relative to the baseline interglacial control (i.e., IG_Ctrl – IG_N50 or IG_Ctrl – IG_N100) to isolate the effects of northern cooling. We also provide anomalies of a full glacial-like temperature state (i.e., IG_Ctrl – Glac_Ref) to serve as a benchmark for comparison.

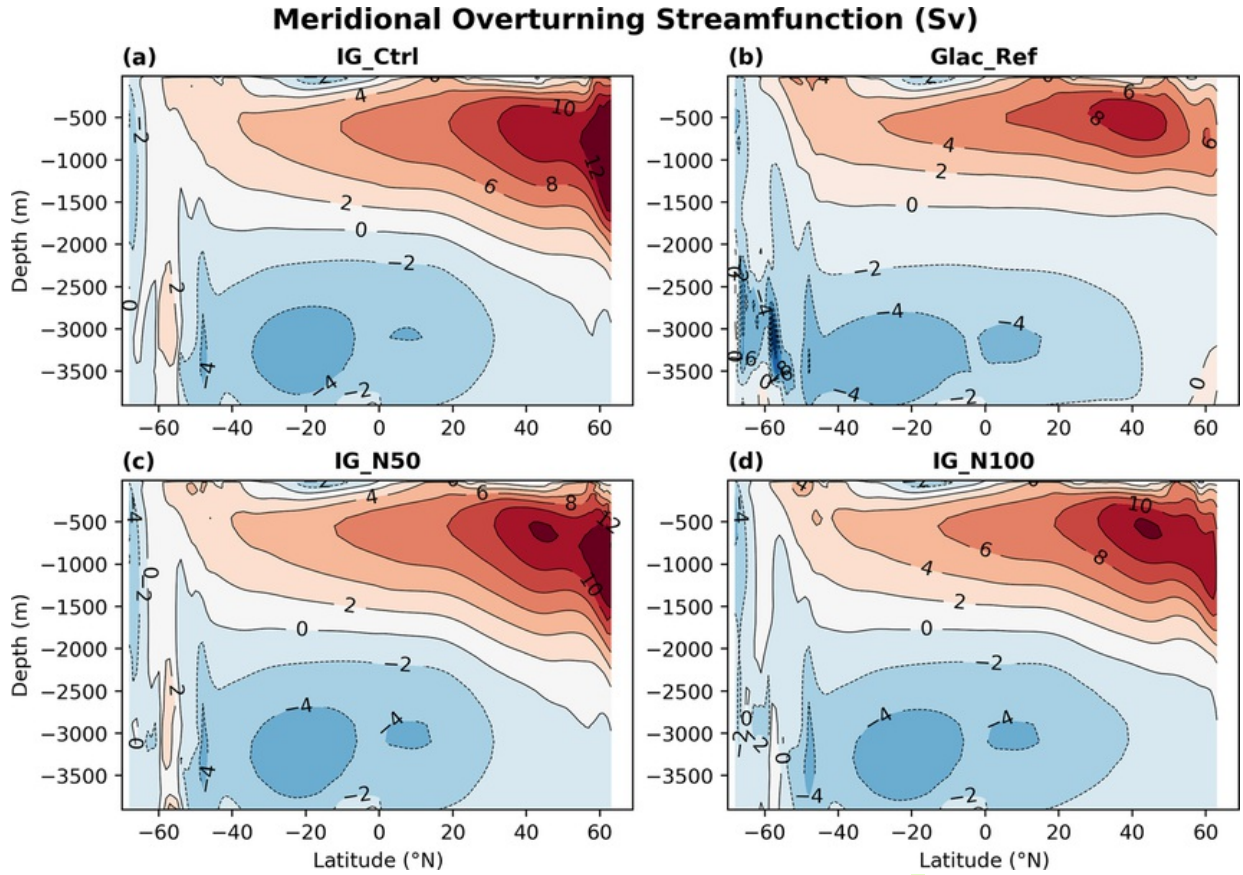


Fig. 3. Meridional overturning streamfunction (Ψ). Meridional sections of zonally-integrated streamfunction (Sverdrups, $1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) in the (a) IG_Ctrl, (b) Glac_Ref, (c) IG_N50, and (d) IG_N100 simulations. Positive (red) values denote clockwise overturning, while negative (blue) values denote counterclockwise overturning. The apparent diapycnal transport at $\sim 60^\circ\text{S}$ in (a) IG_Ctrl and (c) IG_N50 may be related to the effect of standing meanders [see Jansen (2017)] but does not impinge upon our results.

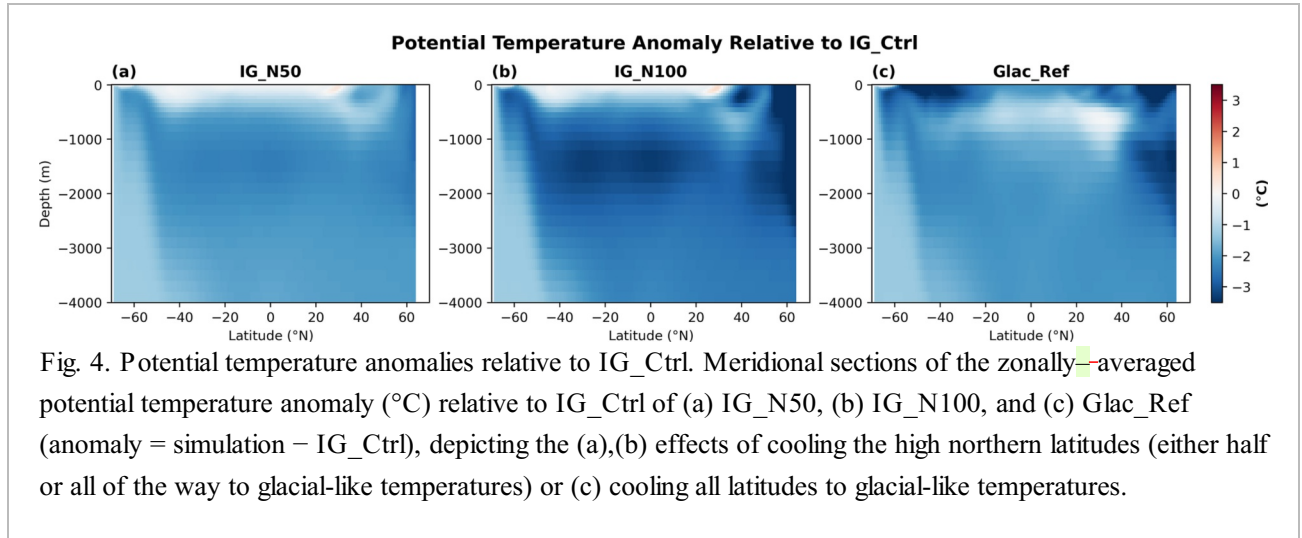
a. Circulation and physical changes

1) Basin-scale changes

The strength of the upper overturning cell (analogous to the AMOC) is modulated by the temperature forcings (Fig. 3, Table S2). Cooling the high northern latitudes reduces the maximum overturning streamfunction by 15% and 33% in IG_N50 and IG_N100, respectively, relative to IG_Ctrl, compared to a 47% weakening in the fully glacially-cooled case (Glac_Ref) (Table S2). In all cases, this change is primarily expressed as a contraction of the strongest overturning away from the northern boundary (only observable in Glac_Ref and IG_N100 in Fig.

3), while overturning throughout the basin remains relatively strong. Unlike Glac_Ref, the northern cooling simulations show little to no shoaling of the upper cell. Thus, northern cooling mainly weakens overturning strength at the far northern latitudes, with IG_N100 reproducing a comparable magnitude of weakening to that much of Glac_Ref's signal (a 33% vs 47% weakening of the upper cell).

Despite weakening, persistent overturning in IG_N50 and IG_N100 transports anomalously cold waters southward throughout the basin. Cooling the high northern latitudes thus proves highly effective at cooling the entire basin, with even modest cooling in IG_N50 producing temperature anomalies comparable to those of Glac_Ref (Fig. 4). The strongest cooling is found at middepths (~1000–2000 m) (Figs. 4a,b), where temperature anomalies in IG_N100 surpass even those of Glac_Ref. This is attributable differences in northern sea ice and downwelling between these two simulations IG_N100 and Glac_Ref. With limited northern sea ice, waters in IG_N100 are subject to a greater area and magnitudes of surface heat loss (Fig. S4). Sea ice extent also modulates the location of downwelling: Sinking in Glac_Ref is shifted south relative to IG_N100, which exhibits strong downwelling near the northern boundary. Consequently, the high northern-latitude waters feeding the upper overturning cell are colder in IG_N100 ($\sim -1.15^{\circ}\text{C}$ on average) than in Glac_Ref ($\sim -0.91^{\circ}\text{C}$) [changes in high northern- and southern-latitude (60° – 70°N , 60° – 70°S) average surface properties can be found in Tables S3 and S4]. The most important feature, however, is that cool anomalies stemming from the high northern latitudes in IG_N50 and IG_N100 reach the southern end of the domain, where they contribute to the cooling of global deep waters (as well as regional sea ice changes). Cooling the high northern latitudes thus ultimately cools the entire deep ocean.



Southward transport of anomalously cool water is also evident in changes to meridional heat flux (Fig. 5). All simulations exhibit a basinwide northward (positive) heat flux driven by the upper overturning cell. The IG_Ctrl exhibits greater heat flux than Glac_Ref, consistent with the stronger overturning in IG_Ctrl. Despite IG_N50 and IG_N100 maintaining weaker overturning than IG_Ctrl, these simulations show even greater meridional heat fluxes than IG_Ctrl. This arises from the pronounced negative middepth temperature anomalies which are transported southward by the lower overturning branch. This export of anomalously cold waters from the north is reflected as enhanced positive meridional heat fluxes in IG_N50 and IG_N100. This in turn leads to an important feature at the high southern latitudes. Southward heat flux (i.e., negative meridional heat flux, only present south of ~55°S in any of the simulations) is greatest in IG_Ctrl but reduced in the two northern cooling simulations. The IG_N100 shows near-zero southward heat flux in the southern sea ice region (Fig. 5), similar to Glac_Ref. Northern cooling in IG_N100 (and IG_N50) thus limits heat entering the high south latitudes. This in turn triggers changes in sea ice growth, surface buoyancy, abyssal temperature, and stratification—similar to the results of Jansen (2017), only driven here by Northern Hemisphere cooling rather than global cooling.

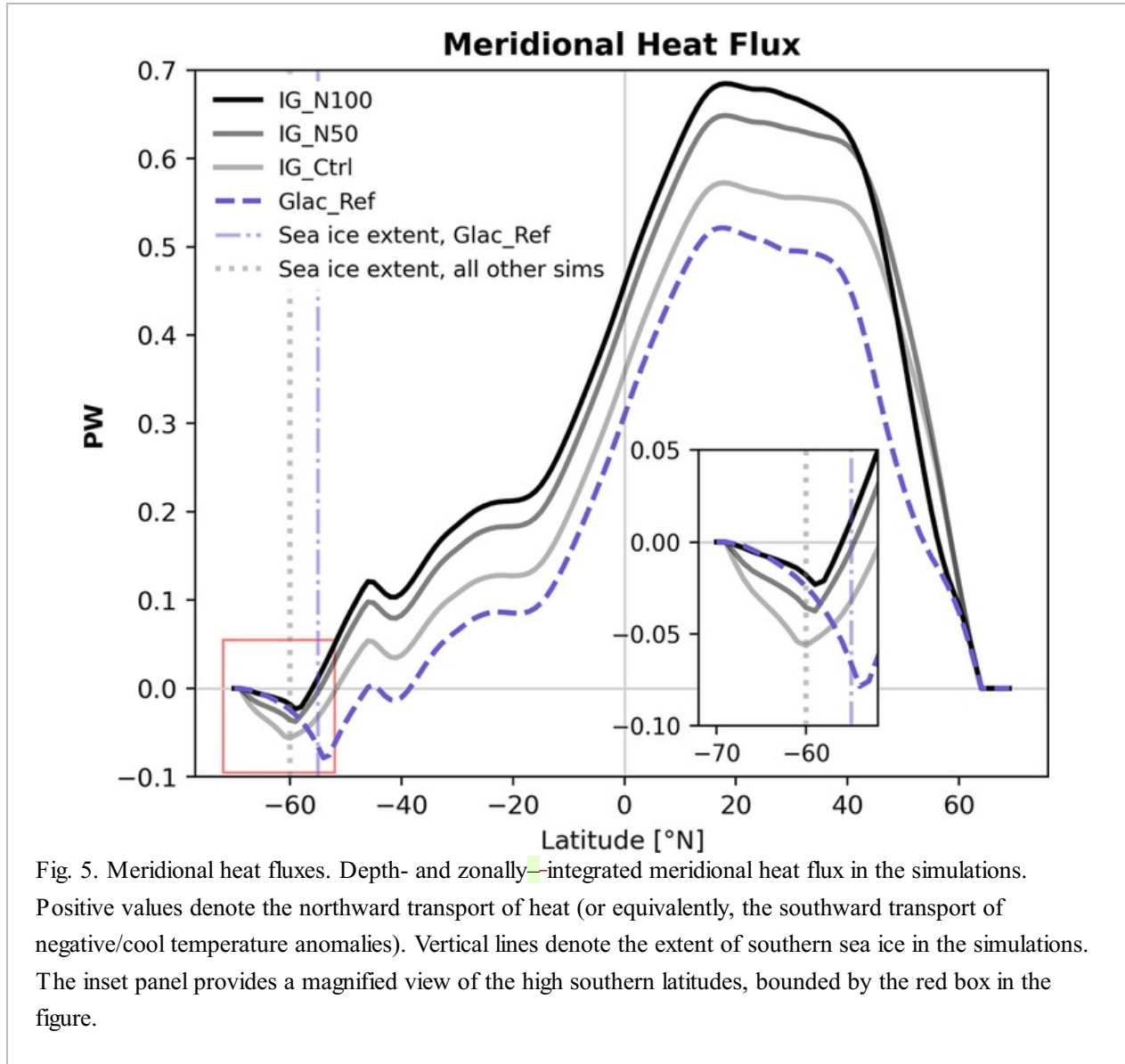
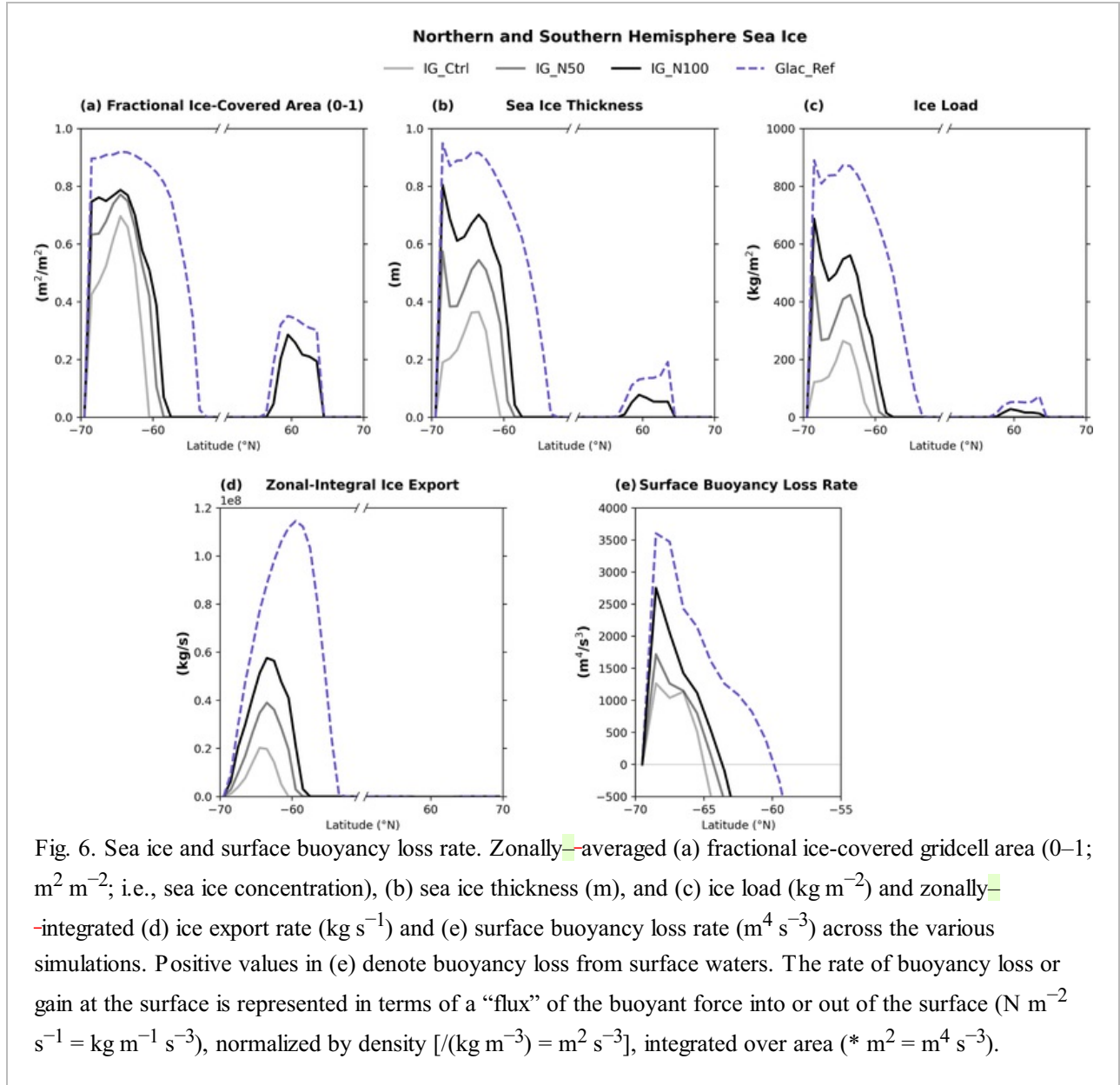


Fig. 5. Meridional heat fluxes. Depth- and zonally-integrated meridional heat flux in the simulations. Positive values denote the northward transport of heat (or equivalently, the southward transport of negative/cool temperature anomalies). Vertical lines denote the extent of southern sea ice in the simulations. The inset panel provides a magnified view of the high southern latitudes, bounded by the red box in the figure.

2) Changes in the high southern latitudes

Basinwide changes in meridional heat flux in IG_N50 and IG_N100 promote increased southern sea ice production (Fig. 6). Regions of expanded sea ice in IG_N50 and IG_N100, relative to IG_Ctrl, correspond to negative sea surface temperature anomalies (Fig. S5), consistent with the reduced heat flux into the southern latitudes in these simulations (Fig. 5) and illustrating the cool northern signal reaching the southern sea ice zone. It is thus cooler waters, in response to northern cooling, which are responsible for promoting sea ice growth in these simulations (as overlying air temperatures are held constant at interglacial-like levels; Fig. 2).



While latitudinal sea ice extent and fractional ice-covered gridcell area ($\text{m}^2 \text{m}^{-2}$) increase only modestly relative to IG_Ctrl (Fig. 6a), this effect is likely underestimated due to the model’s restoring of surface air temperatures to interglacial-like levels at high southern latitudes (Fig. 2). Cooling in the high northern latitudes does lead to significantly thicker southern sea ice and greater ice load, however, compared to IG_Ctrl (Figs. 6b,c). Mean sea ice thickness and load south of 60°S both roughly double in IG_N50 and triple in IG_N100 relative to IG_Ctrl

(Table S5), a sizable change compared to that ~~accounting for much of the change~~ seen in Glac_Ref (which shows a fourfold and sixfold increase in thickness and load, respectively, over IG_Ctrl). This greater sea ice mass drives greater sea ice export (Fig. 6d), with maximum export south of 50°S again roughly doubling and tripling in IG_N50 and IG_N100, respectively, compared to IG_Ctrl (Table S6). This again represents a ~~substantial-considerable change compared to~~ ~~portion of~~ the roughly sixfold increase in sea ice export observed in Glac_Ref. Notably, enhanced southern sea ice, through its effects on freshwater fluxes and surface buoyancy transformations, plays an important role in setting bottom-water properties—a feature of the glacial ocean supported by both proxy data and modeling insights (Adkins et al. 2002; Bouttes et al. 2010; Lhardy et al. 2021; Ferrari et al. 2014). While sea ice latitudinal expansion plays a major role in this in Glac_Ref, it is primarily driven by increased sea ice concentration and especially thickness in IG_N50 and IG_N100, though a modest expansion in extent in these simulations also occurs.

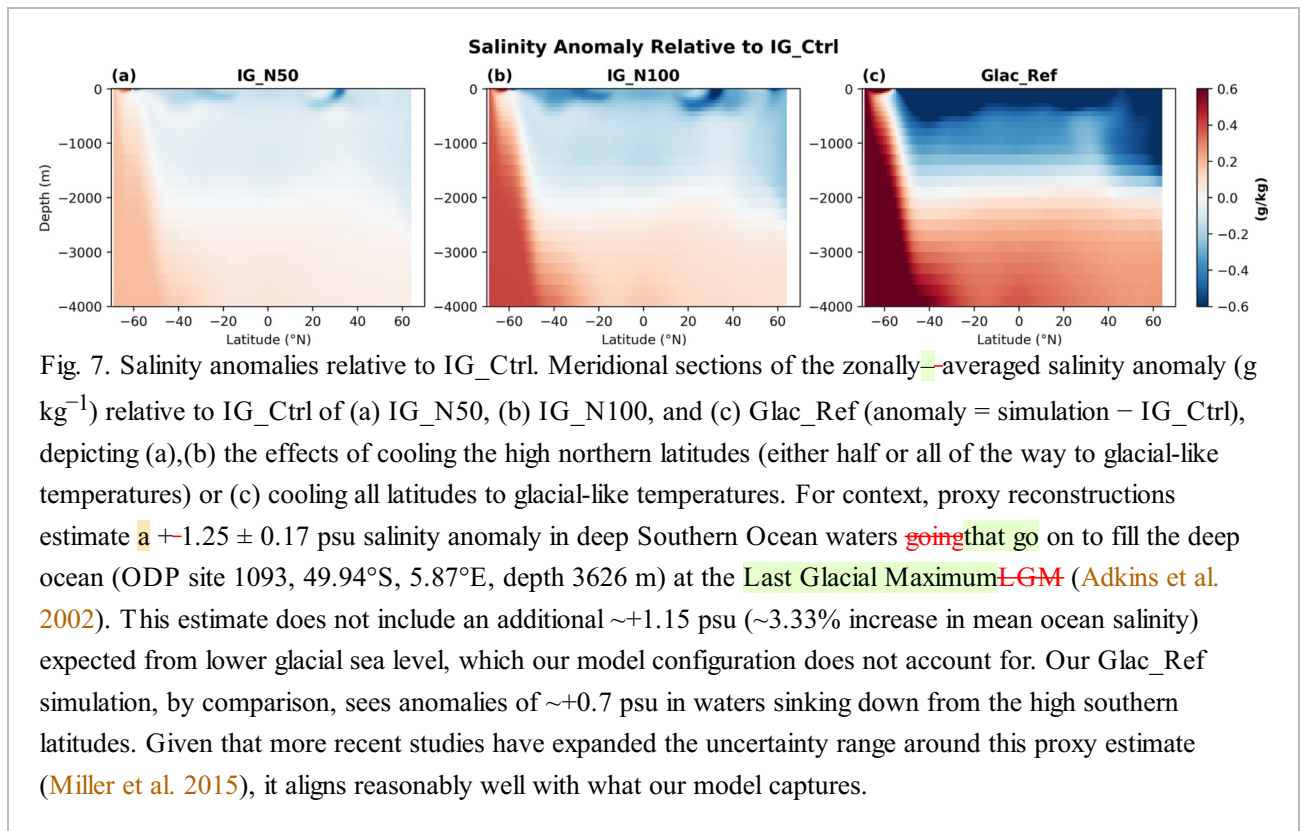
These changes in sea ice dynamics affect surface buoyancy transformations. In all simulations, positive surface buoyancy loss in high southern latitudes (Fig. 6e) reflects the densification of surface waters due to brine rejection during sea ice formation and the associated freshwater export via ice export to lower latitudes. Cooling the high northern latitudes serves to increase southern surface buoyancy loss rates. While IG_N50 only shows a modest ~28% increase in net buoyancy loss south of 55°S relative to IG_Ref, buoyancy loss rates are roughly doubled in IG_N100 (Table S7). This difference between the two northern cooling simulations arises from two competing effects of the cooler incoming waters: cooler waters simultaneously 1) reduce heat loss to the atmosphere, reducing surface buoyancy loss rates, and 2) enhance sea ice growth, increasing surface buoyancy loss rates via brine rejection. Heat-driven buoyancy loss decreases progressively from IG_Ctrl to IG_N50 to IG_N100 (Fig. S6a), consistent with cooler incoming waters. In contrast, buoyancy loss rates due to freshwater fluxes increase, moderately in IG_N50 and strongly in IG_N100 (Fig. S6b). In IG_N50, reduced heat-driven buoyancy loss rates (relative to IG_Ctrl) are roughly offset by increased rates due to freshwater fluxes, resulting in little net change over IG_Ctrl at the high southern latitudes. In IG_N100, however, greater sea ice growth drives especially strong freshwater-flux-driven buoyancy losses, outweighing the effect of reduced heat fluxes and resulting in greater overall buoyancy loss compared to IG_Ctrl.

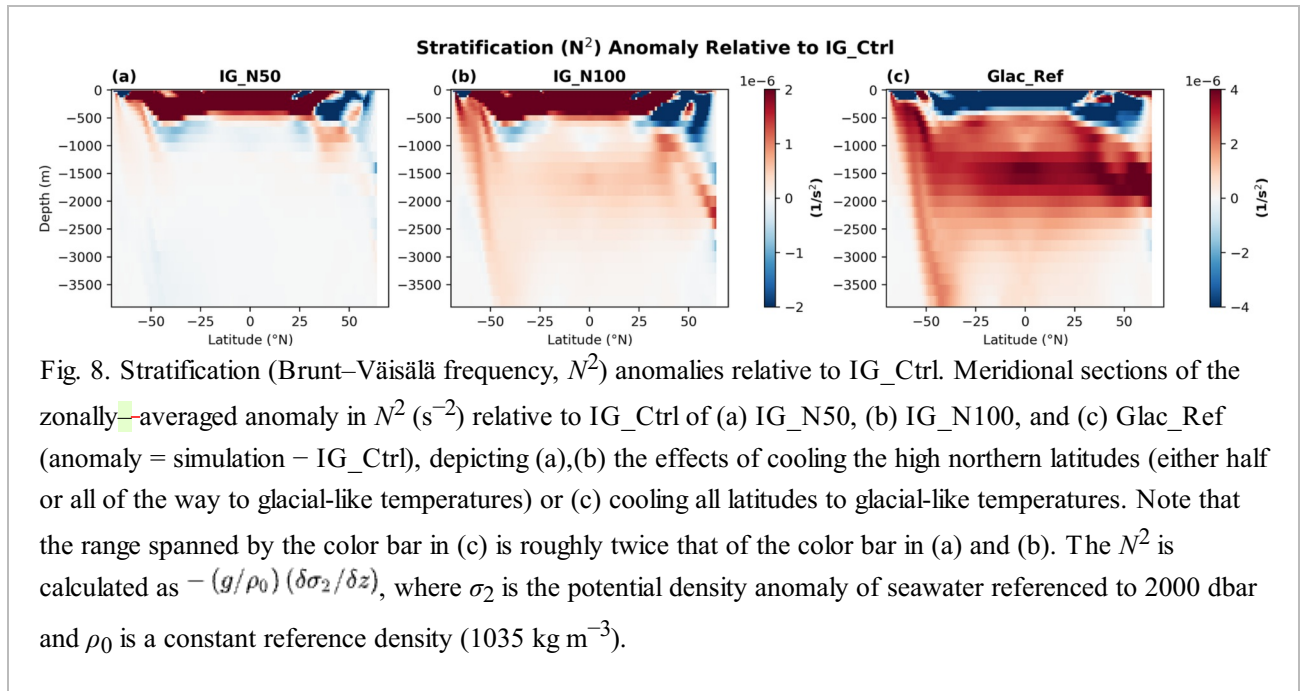
Altered freshwater fluxes driving strong surface buoyancy loss rates are also key in Glac_Ref. Northern cooling (at least in the case of IG_N100) thus enhances southern surface buoyancy transformations in a similar way to

Glac_Ref, though to a lesser extent. Northern cooling in IG_N100 roughly doubles the integrated buoyancy loss rate south of 55°S relative to IG_Ctrl versus a fourfold increase in Glac_Ref (Table S7). Again, this effect is likely underestimated due to the interglacial-like surface air temperatures imposed over southern latitudes in IG_N50 and IG_N100 (Fig. 2).

3) Deep-ocean properties and stratification

Finally, greater sea ice production and associated surface buoyancy loss rates in IG_N50 and IG_N100 relative to IG_Ctrl lead to saltier bottom water being formed at high southern latitudes (for a comparison of salinity changes in the high southern latitudes, see Table S3). This creates a distinct gradient in salinity anomalies between the upper and lower overturning cells, with fresher upper-cell waters and saltier bottom-cell waters (Fig. 7). Saltier bottom waters also appear in Glac_Ref and glacial proxy records (Adkins et al. 2002), supporting northern cooling's role in contributing to this key characteristic of the glacial ocean. Increased bottom-water salinity enhances stratification and isolation of the bottom overturning cell relative to IG_Ctrl (Fig. 8), consistent with glacial $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ proxy data (Curry and Oppo 2005; Lund et al. 2011).





While IG_N50 shows little net change in deep ocean stratification compared to IG_Ctrl, IG_N100 exhibits increased stratification near the boundary of the upper and lower overturning cells (Fig. 8). This difference between IG_N50 and IG_N100 stems from competing thermal- versus salinity-driven effects, with stronger salinity stratification winning out in IG_N100. In both cases, salinity anomalies, transitioning from freshening upper waters to salinifying deep waters (Fig. 7), enhance stratification throughout the water column and particularly at the boundary between the upper and lower cells (Fig. S7). Thermal stratification, by contrast, increases above ~ 1500 m (above the core of the coolest northern-sourced waters; Fig. 4) but mainly weakens below (Fig. S8), partially offsetting the salinity stratification. While IG_N50 shows minimal net change (Fig. 8a), strong salinity stratification wins out in IG_N100 with its greater southern sea ice growth and buoyancy loss, maintaining positive stratification anomalies at ~ 1500 m (Fig. 8b). This competition between salinity- and thermally driven stratification speaks to salinity's ability to counteract reduced thermal stratification expected to accompany a cooler glacial deep ocean (Chang and Jansen 2022; Adkins 2013). Even with the competing effects of thermal- versus salinity-driven stratification, IG_N100 still sees a $\sim 27\%$ increase in mean stratification over 1000–2000 m compared to IG_Ctrl. While this is a modest change compared to Glac_Ref [which sees a $\sim 200\%$ increase, due to a weaker competing thermal stratification signal (Fig. S8c)], the driving mechanism is the same,

with the two simulations sharing strikingly similar patterns of salinity stratification (Fig. S7). The magnitudes of salinity stratification are comparable but stronger in Glac_Ref, consistent with the greater sea ice formation and export rates in that simulation (Fig. 6). Northern cooling thus contributes to a glacial-like salinity-driven stratification and isolation of the bottom overturning cell.

4) Summary

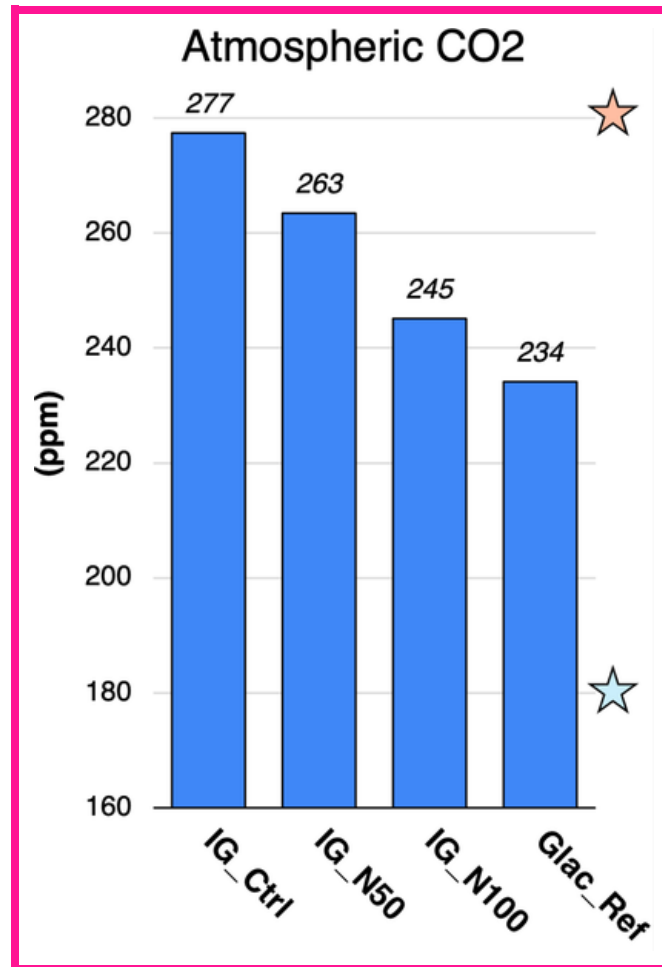
Cooling the high northern latitudes drives substantial changes in ocean properties and circulation, with impacts extending to the southern end of the domain. Northern cooling drives cooling not just of middepths but of the entire water column, and colder upwelling waters in the south promotes sea ice growth and export ($\sim 2\text{--}3\times$ greater than IG_Ctrl). In IG_N100, this drives $\sim 2\times$ greater surface buoyancy loss rates, producing saltier bottom waters and greater stratification between the overturning cells. These changes mark a shift toward important features of glacial ocean circulation: cooler global and deep waters, a weakened upper overturning cell, enhanced sea ice growth, saltier bottom waters, and increased bottom cell isolation. We next show how these features promote ocean carbon storage.

b. Ocean carbon content

1) Total ocean carbon inventory and atmospheric CO₂ change

The temperature and circulation changes induced by northern cooling enhance ocean carbon storage and reduce atmospheric CO₂ levels relative to IG_Ctrl. In response to northern cooling, atmospheric CO₂ decreases by 14 ppm in IG_N50 and 32 ppm in IG_N100 relative to IG_Ctrl (Fig. 9), representing $\sim 33\%$ and $\sim 74\%$ the magnitude of the drawdown achieved by a full glacial-like temperature forcing (Glac_Ref, -43 ppm), respectively. It is important to note that our model's idealized configuration is not expected to replicate the full magnitude of glacial CO₂ change (~ 100 ppm) (Bereiter et al. 2015). In particular, the -43 ppm CO₂ drawdown in Glac_Ref reflects only the impact of altered temperature forcing, without including other known glacial processes such as iron fertilization or carbonate compensation via interactive sediments. The absence of these processes limits the total magnitude of change but does not reduce the model's utility for our focus here: isolating and quantifying the impacts of northern cooling. The Glac_Ref simulation is thus not intended to serve as an exact analog for glacial conditions and instead provides a useful benchmark against which to measure the impacts of northern cooling alone. Northern cooling alone can thus strongly modulate CO₂, with the CO₂ change signal in

IG_N100 approaching the magnitude of that ~~full response~~ under full glacial-like temperatures. This is achieved by mechanisms promoting carbon storage in the ocean; total ocean carbon inventory increases by 8.2 PgC (petagrams carbon) in IG_N50 and 18.8 PgC in IG_N100 relative to IG_Ref, compared to 26.2 PgC in Glac_Ref.



Note:

 [Updated_Fig9_AtmsCO2_bars_F009.png](#)

Figure Replacement

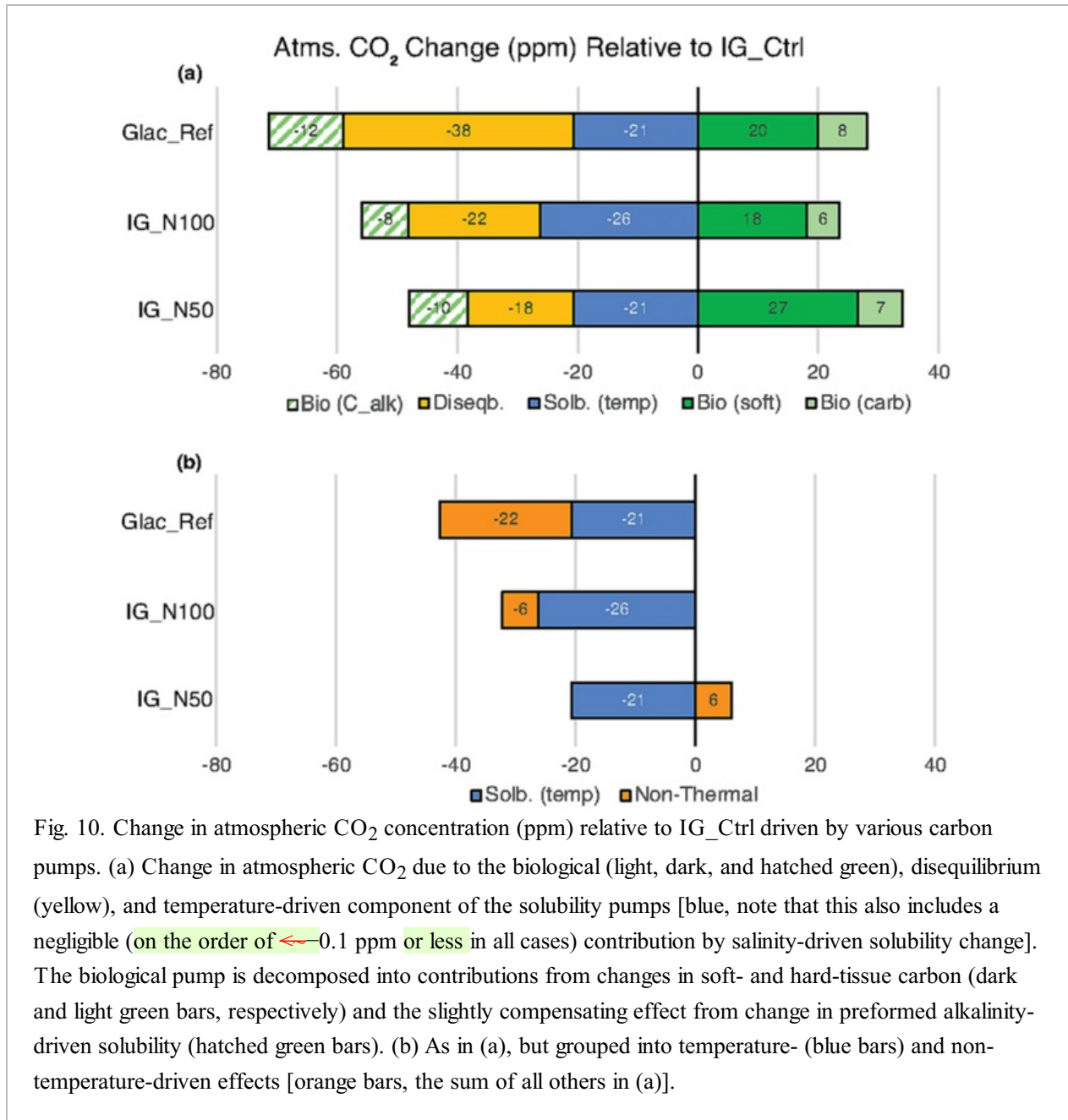
Requested. Fig. 9. Atmospheric CO₂ concentration. Equilibrated atmospheric CO₂ concentrations (ppm) across the simulations. The pale orange and pale blue stars denote preindustrial (~280 ppm) and glacial (~180 ppm) CO₂ as documented in ice cores (Bereiter et al. 2015). These are provided for context, with the understanding that the Glac_Ref simulation is not intended to serve as an exact analog for past glacial periods.

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2) Carbon pumps

To investigate drivers of increased ocean carbon storage, we decompose DIC into its components (C_{sat} , C_{dis} , C_{soft} , and C_{carb} ; see methods). This allows us to attribute atmospheric CO_2 changes to different carbon “pumps.” The conceptualization of ocean carbon storage into the action of various pumps was originally developed by Volk and Hoffert (1985) (~~Volk and Hoffert 1985~~) and has since been expanded to include the influence of air–sea disequilibrium (Williams and Follows 2011; Ito and Follows 2005, 2013). It has become a commonly used framework in studies of glacial ocean carbon storage (Eggleston and Galbraith 2018; Khatiwala et al. 2019; Marzocchi and Jansen 2019).

The solubility pump depends on the solubility of CO_2 in seawater (a factor of temperature, salinity, and preformed alkalinity) and determines the mass of C_{sat} at depth. The solubility pump can be decomposed into effects due to changes in temperature and salinity, but we combine these effects into a single metric as the temperature signal dominates (salinity changes drive changes in the atmosphere on the order of ≤ -0.1 ppm ~~change in the atmosphere~~ or less in all simulations). Meanwhile, biological production determines the preformed alkalinity inventory of the ocean and thus alkalinity-driven solubility, and for this reason, the preformed alkalinity term of C_{sat} is included under the biological—not solubility—pump (see methods). The disequilibrium pump reflects air–sea gas exchange imbalances, influencing C_{dis} which represents carbon over- or undersaturation. The biological pump transfers carbon to depth via carbon uptake by productivity in surface waters and the subsequent sinking and remineralization of soft- (C_{soft}) and hard-tissue (C_{carb} or carbonate) carbon. Note that the indirect effect of preformed alkalinity-driven solubility changes mentioned previously is also part of the biological pump. Here, we quantify each pump’s contribution to atmospheric CO_2 changes across the simulations (Fig. 10). In all simulations, CO_2 reductions are mainly driven by the solubility and disequilibrium pumps, partly offset by a weaker biological pump (Fig. 10a).



(i) Solubility pump

Both IG_N50 and IG_N100 show strong thermally-driven solubility effects (dark blue bars in Fig. 10a). Cooler waters throughout the water column (Fig. 4) enhance their solubility and uptake potential, increasing C_{sat} relative to IG_Ctrl (Figs. S9d–f and S10g–i). This drives CO₂ drawdowns of –21 and –26 ppm in IG_N50 and

IG_N100, respectively, relative to IG_Ctrl. Cooling the high northern latitudes thus very effectively strengthens the solubility pump, with both simulations matching the magnitude of the full thermal solubility effect seen in Glac_Ref (−21 ppm). ~~Indeed~~In fact, IG_N100 even exceeds Glac_Ref due to its cooler middepth anomalies (Fig. 4b). The effect of cooling the northern latitudes extends even beyond middepth northern-sourced waters, however, and ultimately cools the entire deep ocean, driving glacial-like levels of solubility drawdown. This result underscores the high northern latitudes as a key region for the solubility pump, with even modest cooling in IG_N50 achieving near-glacial-like levels of solubility-driven CO₂ reduction.

Alongside thermally-driven solubility enhancement, there is also a clear imprint of preformed alkalinity-driven changes on the distribution of C_{sat} anomalies (Figs. S10d–f). While temperature-driven solubility increases occur throughout the water column (supplemental Figs. 10g–i), the distribution of negative/positive C_{sat} anomalies in the upper/lower overturning cells is driven by preformed alkalinity changes (Figs. S10d–f). This distribution of preformed alkalinity anomalies in turn largely reflects the influence of salinity changes (Fig. S11) modulated by southern sea ice. This salinity influence arises from freshwater export associated with greater southern sea ice, which concentrates salinity and alkalinity in the southern-sourced waters and lower overturning cell while diluting it in the upper cell. These salinity-driven anomalies (Figs. S11d–f) sum to virtually zero in the global integral (only redistributing preformed alkalinity without driving any net change in the whole-ocean preformed alkalinity inventory). Consequently, they do not contribute to the net drawdown depicted by the hatched green bars in Fig. 10, which is entirely attributable to nonsalinity changes (Figs. S11a–c), i.e., the biological pump’s modulation of the preformed alkalinity inventory. Nonetheless, the strong control of salinity on the spatial distribution of preformed alkalinity and associated C_{sat} anomalies remains a significant result, even having no net effect on carbon. It demonstrates/exemplifies a dynamical southern-latitude change, induced by remote northern forcing, of sufficient magnitude to influence biogeochemistry and tracer distribution. The same pattern of anomalies in preformed alkalinity-driven solubility also appears in Glac_Ref (Fig. S10f) and (weakly) in IG_N50 (Fig. S10d).

(ii) Disequilibrium pump

The disequilibrium pump drives considerable change between IG_N50/IG_N100 and IG_Ctrl, capturing a substantial part-of-the disequilibrium-driven drawdown compared to seen-in-Glac_Ref (yellow bars in Fig. 10a). Both northern cooling simulations see an accumulation of C_{dis} at depth originating from high southern latitudes

(Figs. S9g,h)—a region where sea ice limits air–sea gas exchange and short residence times prevent full equilibration with the atmosphere. This result aligns with the expansion in sea ice extent and concentration (Fig. 6) in IG_N50 and IG_N100 over IG_Ref, with Glac_Ref achieving even stronger disequilibrium drawdown with its still greater expansion of sea ice. Enhanced stratification between the upper and lower overturning cells in IG_N100 (Fig. 8) may also facilitate C_{dis} accumulation at depth. Disequilibrium-driven CO_2 reductions are -18 and -22 ppm in IG_N50 and IG_N100, respectively ($\sim 47\%$ and $\sim 58\%$ the magnitude of that in Glac_Ref (of the -38 ppm) disequilibrium drawdown in Glac_Ref). While smaller in magnitude than Glac_Ref, this disequilibrium-driven drawdown in IG_N50 and IG_N100 is significant; emerging from the southern end of the domain, it demonstrates that remote northern cooling alone can trigger dynamical changes in far southern latitudes salient to ocean carbon storage.

(iii) Biological pump

Finally, in all three simulations, less biological carbon (C_{soft} , C_{carb}) in the ocean relative to IG_Ctrl raises atmospheric CO_2 , competing with but not exceeding solubility and disequilibrium drawdowns. Negative C_{soft} and C_{carb} anomalies throughout the water column in IG_N50 and IG_N100 (Fig. S9) indicate an overall less efficiently operating biological pump. Here, we define a “less efficient” global biological pump as one in which more nutrients go unused by production and ultimately are lost to the deep ocean; this accumulation of preformed nutrients in the deep in turn reduces the capacity of biological production to take up carbon in the surface, leading to less remineralized carbon (C_{soft} and C_{carb}) in the ocean interior (Ito and Follows 2005). We also clarify that anomalies in C_{soft} and C_{carb} are strictly a measure of the moles of carbon lost or gained by a parcel as a result of these pumps. This represents only one component of the biological pump, which also includes preformed alkalinity-modulated changes in solubility (i.e., the C_{alk} term of C_{sat}) discussed previously.

In our northern cooling simulations, the loss of soft- and hard-tissue regenerated carbon from the ocean appears to be driven by changes in the Southern Hemisphere, with the largest anomalies in C_{soft} and C_{carb} stemming from high southern-latitude surface waters. This is a region where in the present day production already tends to be inefficient and nutrients already tend to be lost to the deep. Northern cooling evidently enhances this effect, with more preformed nutrients being lost to depth in these southern-sourced waters and thus lowering the overall efficiency of the global biological pump. (For a table of changes in southern-latitude surface properties, including

preformed nutrient content, see Table S3.)

The same effect occurs in the Glac_Ref simulation (Figs. S9l,o). Importantly, this does not conflict with proxy evidence of a more efficient biological pump during glacial times (Studer et al. 2015; Ai et al. 2020). Rather, it shows that the effect of cooling global—or northern—air temperatures would be to reduce biological pump efficiency. In reality, the biological pump was likely enhanced under glacial conditions, due to other factors such as iron fertilization (Martínez-García et al. 2014; Lambert et al. 2015; Thöle et al. 2019) and changes in wind-driven upwelling (Ai et al. 2020). We, therefore, emphasize that the northern cooling discussed in this study is only one mechanism potentially contributing to glacial CO₂ change, operating alongside other canonical glacial processes. It should also be noted that the present model configuration does not account for the temperature dependence of metabolic processes determining production and remineralization rates. With cooler waters, the effect of slower remineralization rates (allowing organic detritus to sink deeper before remineralizing) outweighs the effect of slower export production (Matsumoto 2007), with a net effect of greater remineralized carbon storage at depth (Kwon et al. 2009). If this effect was simulated in our model, the resultant carbon storage would present a competing effect ~~that may partly compensate~~ to the loss of soft- and hard-tissue carbon observed here.

The loss of soft- and hard-tissue carbon (C_{soft} , C_{carb}) due to a less efficient biological pump thus drives +34-ppm CO₂ change in IG_N50 relative to IG_Ref, though this is slightly mitigated in IG_N100 and Glac_Ref (+24 and +28 ppm, respectively) due to positive anomalies and the accumulation of biologically-mediated remineralized carbon in the zone of enhanced middepth stratification (Figs. S9k,l,n,o). In all simulations, this effect is partly offset by preformed alkalinity-driven changes in solubility (the C_{alk} term in the biological pump; see also Figs. S11a–c), which drive drawdowns from ~–8 to –12 ppm (shown by the hatched green bars in Fig. 10). The net effect of biological pump changes is thus to raise CO₂ by +24 ppm in IG_N50 and +16 ppm in IG_N100 and Glac_Ref. These results again illustrate northern cooling impacting southern-latitude processes, which in turn alter the biogeochemistry of southern-sourced waters.

3) Distinguishing between temperature- and non-temperature-driven effects on atmospheric CO₂

Atmospheric CO₂ change relative to IG_Ctrl thus results from combined effects. In IG_N50 and IG_N100, this includes (i) enhanced temperature-driven solubility from northern cooling and (ii) changes in other pumps largely linked to changes in the high southern latitudes. Both results are significant. First, it is striking that cooling the

northern latitudes alone reproduces the full glacial-scale temperature-driven solubility drawdown. This is achieved by the impact of northern cooling extending beyond local middepth waters and ultimately cooling the entire deep ocean. Global deep waters cool as much under northern-latitude cooling as under global cooling (Glac_Ref), producing a strong overall solubility effect. This highlights the broad influence of northern-latitude cooling on the solubility pump. Second, it is significant that additional contributions are made by the remaining pumps, as these largely emerge from downstream changes in the high southern latitudes triggered by the northern cooling.

Expanded southern sea ice enhances disequilibrium carbon in the ocean. Similarly, a loss of biological carbon, reflecting an accumulation of preformed nutrients in the ocean interior, appears to stem from the high southern latitudes.

Separate from the solubility pump, the disequilibrium and biological pumps are partly interrelated by their diagnostic definitions. Since the disequilibrium pump is calculated as a residual from the other pumps, any changes in the biological pump will induce compensating change in the disequilibrium pump. Given this interaction between the biological and disequilibrium pumps in this diagnostic framework, it can be informative to interpret them collectively—grouping them as a set of interrelated processes separate from the temperature-driven solubility signal (Fig. 10b). This approach has the added benefit of distinguishing between northern- and southern-based drivers: The temperature-driven drawdown originates in the high northern latitudes, while the remaining components—hereafter “non-thermal effects”—primarily reflect changes occurring in the high southern latitudes.

In this framework, the net impact of non-thermal changes on atmospheric CO₂ in IG_N50 and IG_N100 (orange bars, Fig. 10b) is small compared to the temperature-driven solubility signal (blue bars) and is smaller ($\sim\pm 6$ ppm) than the non-thermal drawdown achieved in Glac_Ref ($\sim\pm 6$ ppm vs -22 ppm). However, a clear directional shift occurs in the non-thermal effects between the two northern cooling simulations, shifting from driving net increase in IG_N50 (+6 ppm) to a net drawdown in IG_N100 (-6 ppm). The intensified northern cooling in IG_N100 thus drives an extra 12 ppm of drawdown compared to IG_N50. This shift reflects both a slight mitigation of the weakened biological pump (driving only +16 ppm in IG_N100 vs +24 ppm in IG_N50, an 8-ppm difference) and a stronger disequilibrium pump (-22 vs -18 ppm, a 4-ppm difference). Stronger northern cooling in IG_N100 thus aligns biological pump effects more closely with Glac_Ref-like values (green bars, Fig. 10a) and contributes to a stronger disequilibrium pump.

Strong northern cooling in IG_N100 thus effectively reproduces the full glacial-like signal in both the biological- and temperature-driven solubility pumps. The primary reason it falls short of full glacial-scale CO₂ drawdown is the weaker disequilibrium pump response, likely due to limited southern sea ice expansion under the overlying preindustrial-like air temperatures (Fig. 2). Even so, the disequilibrium-driven drawdown in IG_N100 (and even IG_N50) captures is roughly half that the disequilibrium-driven drawdown observed in Glac_Ref, showing that northern cooling can substantially impact southern-latitude processes influencing the disequilibrium (and biological) pumps (this effect of northern cooling inducing changes in the high southern latitudes is illustrated well by the anomalies in Figs. S9 and S10). While the net effect of such changes on CO₂ is nuanced and partly compensatory, they nevertheless arise from dynamical changes in the high southern latitudes initiated by remote northern cooling, representing impacts on ocean carbon storage beyond the temperature-driven solubility signal. That northern cooling triggers such strong temperature-driven solubility drawdown that is also significant, however. By cooling not only middepth northern-sourced waters but also the entire deep ocean, northern cooling alone achieves nearly the full glacial-scale solubility drawdown. Together, these results highlight the powerful downstream effects of northern cooling on the Southern Ocean and global interior.

4. Discussion

A mechanism linking orbitally-paced Northern Hemisphere ice sheet changes to Southern Hemisphere processes modulating ocean carbon storage is key to explaining glacial–interglacial climate change. Gildor and Tziperman (2001) posited one such mechanism in which a large northern ice sheet cools NADW reaching the Southern Ocean, triggering southern sea ice expansion and related changes. Where they demonstrated the ability of this interhemispheric mechanism to drawdown CO₂ in a coupled box model (Gildor and Tziperman 2001), we extend this result using a dynamic coupled general circulation model in an idealized setup.

We find that cooling the northern latitudes alone (emulating the thermal effect of a large northern ice sheet) alters ocean circulation and enhances ocean carbon content, achieving up to 32 ppm of CO₂ drawdown or ~74% the magnitude of that of the CO₂ drawdown seen in a full glacial-like simulation (43 ppm). Notably, in addition to a thermal solubility-driven signal from the north, substantial changes in ocean carbon content also appear largely to originate from the high southern latitudes (Figs. S9 and S10). Cooling the high northern

latitudes thus emerges as a highly effective means of modulating atmospheric CO₂, supporting Gildor and Tziperman's (2001) hypothesis of northern ice sheets' role in glacial change.

This result is striking for several reasons. First, it highlights the high northern latitudes as a key region shaping ocean circulation and carbon storage; with global ocean cooling, enhanced ocean carbon uptake, and substantial CO₂ drawdown (74% the magnitude of that of the total CO₂ change achieved by the Glac_Ref simulation) all being triggered by temperature forcing over just 13% of the model domain. The effect of cooling the northern latitudes extends beyond locally formed middepth waters and also cools deep waters formed in the south, significantly amplifying solubility-driven carbon storage. Interestingly, isolated northern cooling induces greater mean whole-ocean temperature change (−2.5°C in IG_N100) than cooling of identical magnitude restricted to the Southern Hemisphere (−0.6°C), performed in a simulation with the same configuration used here (Jansen 2017) (Fig. S12). The disproportionate control over basinwide temperature held by the northern latitudes (and the evident nonlinearity between northern and southern cooling, with global cooling in our Glac_Ref simulation driving only −2.0°C mean-ocean change), while not a crucial feature of the Gildor and Tziperman hypothesis, is an intriguing result deserving of further study. The impact of northern cooling is also likely to be robust to the weakened glacial overturning evidenced by proxy data (Liu 2023; Pöppelmeier et al. 2023a; Blaser et al. 2025), with cool waters still reaching the south under the slightly weakened overturning rates of IG_N50 and IG_N100. Second, northern cooling drives changes beyond the temperature-driven solubility effect. While enhanced solubility drives the majority of the CO₂ drawdown, we also observe major impacts on the disequilibrium and biological pumps. Though partly compensatory in their net effect on CO₂, they nevertheless reflect dynamical rather than thermal changes downstream of northern cooling—comprising additional impacts outside the temperature-driven solubility signal. Finally, the central role of southern-based processes (e.g., expanded sea ice) in facilitating this ocean carbon storage (Figs. S9 and S10) highlights how strongly northern forcing can influence changes in the south. This result is significant for interhemispheric synergy and for linking orbitally paced Northern Hemisphere changes to Southern Hemisphere processes that can regulate ocean carbon storage.

More broadly this adds to a growing understanding of the tight coupling between remote hemispheric climate processes, such as the bipolar seesaw, etc. (Testorf et al. 2026).

While disequilibrium carbon storage associated with southern sea ice changes is a key outcome of northern

cooling, it represents only one possible mechanism operating under glacial conditions. Other processes are recognized to be central to glacial CO₂ drawdown, a key one being a more efficient biological pump. Enhanced nutrient uptake in the Southern Ocean under glacial periods is well supported by $\delta^{15}\text{N}$ proxy data (François et al. 1997; Studer et al. 2015; Ai et al. 2020) and would have led to a reduced preformed nutrient inventory and greater biological carbon inventory in the ocean interior. This contrasts with our results, in which northern cooling promotes the export of preformed nutrients sent to depth (evident in widespread reductions in biological carbon; Figs. S9j–o). Northern cooling increases the preformed nutrient content of the southern-sourced waters that feed the lower overturning cell (Table S3) and of the mean-ocean preformed nutrient content overall (by +12% in IG_N50 and by +8% in IG_N100, the latter of which exhibits a slight compensating decrease in northern-sourced waters’ preformed nutrient content; Table S4). We emphasize again that this outcome—reduced nutrient uptake in the southern endmember—reflects only the effect of northern cooling and in reality was likely offset by other processes enhancing nutrient utilization, such as iron fertilization (Martínez-García et al. 2014; Lambert et al. 2015; Thöle et al. 2019), changes in wind-driven upwelling (Ai et al. 2020), and altered nutrient content of source waters entering the high southern latitudes (Shankle et al. 2025).

a. Constraining the impact of northern cooling

Interestingly, our results differ from those of a similar study using MITgcm in a coupled ocean–sea ice–biogeochemistry setup, though in a non-idealized two-basin configuration and employing a different biogeochemical model (Du et al. 2022). In this study, an isolated cooling of North Atlantic sea surface temperatures under otherwise interglacial-like boundary conditions led to a slight increase (+6 ppm) in atmospheric CO₂. Cooler upper-cell waters and a stronger deeper AMOC reduced thermal stratification and enhanced vertical mixing, bringing deep carbon into the upper cell and promoting outgassing. This effect outweighed that of the enhanced solubility of cooler North Atlantic waters, leading to the slight CO₂ increase. Notably, this result in Du et al.’s simulation is driven largely by the response of the overturning, which strengthens and extends all the way to the seafloor, precluding deep-ocean carbon storage. Because such an overturning structure runs directly counter to proxy evidence of a glacial AMOC that was likely slightly shoaled and either unchanged or only modestly reduced relative to modern (Muglia and Schmittner 2021; Pöppelmeier et al. 2023a; Liu 2023; Blaser et al. 2025), we do not believe the results of Du et al. (2022) preclude the viability of the Gildor and Tziperman mechanism operating in glacial times.

~~Consistent with our findings,~~ Du et al. (2022) also found, consistent with our results, expanded southern sea ice and associated Antarctic Bottom Water (AABW) changes (resulting from an isolated Southern Ocean cooling) being key to enhanced ocean carbon storage. This drove enhanced stratification between the overturning cells and shortened Southern Ocean surface residence times, limiting air–sea gas exchange. (The capping effect of sea ice was minimal since, under preindustrial-like conditions, atmospheric CO₂ was relatively high and air–sea CO₂ disequilibrium was low, driving minimal uptake.) These effects produced a –22-ppm drawdown from isolated Southern Ocean cooling and –26 ppm from combined northern and southern cooling. Thus, effective carbon storage was dependent on southern sea ice changes and deep stratification, which in their simulations only ever emerged from southern cooling. By contrast, we observe such southern changes occurring in response to remote northern cooling.

It is unclear why Du et al. (2022)’s northern cooling simulation did not trigger Southern Ocean sea ice growth and related changes. One contributing factor may be the relatively warm NADW in their interglacial control simulation [$\sim 6^{\circ}$ – 8° C, comparable to modern NADW (Locarnini et al. 2024), see their supplemental Fig. 10, vs $\sim 2^{\circ}$ – 4° C in our IG_Ctrl, Fig. S2a]. With anomalies of only $\sim 1.5^{\circ}$ – 2° C (see their supplemental Fig. 4), NADW in their simulation may not have cooled sufficiently to drive southern sea ice expansion. Other differences include their non-idealized two-basin setup and differing forcing method (prescribing heat fluxes where we prescribe air temperature). While a detailed comparison is beyond this study’s scope, both studies support the role of expanded southern sea ice and associated changes in enhancing ocean carbon storage. Whether such changes can result from northern cooling alone—while supported by our results—remains a key question for future research and model intercomparison.

b. The influence of Antarctic ice melt

Gildor and Tziperman (2001) originally demonstrated this interhemispheric mechanism of glacial change in a coupled box model, achieving CO₂ drawdown through expanded southern sea ice and enhanced thermal stratification in the Southern Ocean due to cooler incoming waters from the Northern Hemisphere (Gildor and Tziperman 2001). Our results not only support this interhemispheric coupling but also reveal a role for salinity-driven changes linked to sea ice. Such downstream effects of sea ice on deep-ocean salinity are not captured in Gildor and Tziperman’s (2001) model due to the configuration of their boxes: Water upwelled into the Southern Ocean surface box is advected northward into a midlatitude box, with no return flow from the Southern Ocean

surface back to the deep ocean (see their Fig. 1). Thus, their model does not capture the salinification of AABW via the mechanism of cooler NADW and subsequent enhanced southern brine rejection that we observe.

Our results may even underestimate this effect, as cooler NADW would also mitigate the “prefreshening” effect of warm NADW on waters destined to become AABW—a process not simulated in our model. Today, warm NADW and downstream CDW upwelling in the Southern Ocean contributes to melting Antarctic ice shelves, prefreshening waters that ultimately become AABW (Hillenbrand et al. 2017; Nakayama et al. 2018). Cooler incoming waters during glacial periods may have reduced this melt, contributing to saltier AABW (Adkins 2013; Miller et al. 2012). Modeling studies support the role of cooler NADW in modifying bottom-water properties. For example, topographic barriers that block warm NADW from reaching southern ice shelves lead to colder, saltier bottom waters (Hellmer 2004), as does cooling incoming waters at the boundary of a Weddell Sea regional model (Miller et al. 2012). While our model does not simulate Antarctic ice sheets or the prefreshening effect of NADW, including these processes would likely reinforce our results and contribute to bottom-water salinification and deep-ocean stratification.

c. Southern Ocean surface conditions and ocean circulation

Southern Ocean processes—like expanded sea ice and surface buoyancy fluxes—are central to our northern cooling simulations, aligning with the known role of Antarctic sea ice and high-latitude conditions in shaping ocean circulation and stratification. The link between southern sea ice and the distribution of deep-water masses has been demonstrated in theory, idealized models, and more complex general circulation models (Ferrari et al. 2014; Jansen and Nadeau 2016; Marzocchi and Jansen 2017; Nadeau et al. 2019). Negative surface buoyancy fluxes applied in idealized simulations, for example, produce denser, more stratified AABW and a shoaling of the NADW–AABW boundary, resulting in a better-isolated bottom overturning cell (Jansen and Nadeau 2016; Nadeau et al. 2019). Our northern cooling simulations replicate, though to a lesser degree, this densification of AABW and deep stratification. They also capture a weakened upper overturning cell, consistent with glacial proxy data (Pöppelmeier et al. 2023a; Blaser et al. 2025).

However, northern cooling alone does not produce any shoaling of the boundary between the upper and lower cells, another feature supported by glacial modeling and proxy data (Pöppelmeier et al. 2023a). Our glacial-like simulation (Glac_Ref) achieves this shoaling due to extensive southern sea ice and associated surface buoyancy changes. By contrast, while IG_N50 and IG_N100 show considerable enhancement of sea ice and surface

buoyancy fluxes over IG_Ctrl, these remain well below the levels of Glac_Ref.

Strong sea ice formation and export thus appear central to driving major deep water-mass reorganization.

Supporting this idea, studies using intermediate complexity models identify brine rejection and shelf convection as key processes to shoaling the upper overturning cell (Klockmann et al. 2016; Lhardy et al. 2021); such processes may be inadequately represented in our model or insufficiently triggered by remote northern cooling.

Nevertheless, northern cooling clearly affects southern surface conditions in our simulations and contributes to glacial-like features such as a weakened upper overturning cell and enhanced deep-ocean stratification (Jansen and Nadeau 2016; Nadeau et al. 2019).

d. Ocean carbon pumps

Our results also shed light on how physical circulation affects ocean carbon and carbon pump efficiency. We find expanded southern sea ice enhancing the disequilibrium pump, agreeing with some studies (Marzocchi and Jansen 2019; Stephens and Keeling 2000) but challenging others (Khatiwala et al. 2019). Khatiwala et al. (2019) found expanded sea ice having minimal impact on disequilibrium carbon, simultaneously blocking both ingassing and outgassing of carbon by the ocean. Instead, increased sea ice in their model increased atmospheric CO₂ by ~16 ppm due to the greater light limitation on biological productivity. However, to study this effect, Khatiwala et al. (2019) imposed an Last Glacial Maximum (LGM)-like ice field on a preindustrial-like ocean—an approach which has been criticized (Marzocchi and Jansen 2019; Stein et al. 2020) for neglecting the coupling between sea ice and ocean circulation and stratification (Ferrari et al. 2014; Jansen 2017). In contrast, our results support sea ice’s role in shaping deep-ocean stratification and trapping disequilibrium carbon.

Our findings on disequilibrium carbon also align with a study using the more complex coupled climate-biogeochemical Climatecoupled Model version 2 with the Modular Ocean Model, at coarse gridresolution (CM2Mc)model (Eggleston and Galbraith 2018), which showed both disequilibrium (C_{dis}) and soft-tissue (C_{soft}) carbon responding sensitively to astronomical, ice sheet, and radiative forcings. This study also found an inverse relationship between C_{dis} (primarily governed by AABW volume) and C_{soft} (dependent on the product of export production and ideal age), driven by the inverse link between AABW volume and ideal age. In their simulations, increased AABW volume increases C_{dis} but ventilates and reduces deep-ocean ideal age, thus reducing C_{soft} at depth. A similar interplay between the disequilibrium and biological pumps appears in our simulations. Such

relationships partly reflect diagnostic artifacts from how the C_{soft} and C_{dis} tracers are defined (i.e., with C_{dis} being calculated as a residual from the other components of DIC). Eggleston and Galbraith (2018) noted that processes like iron fertilization, for example, could decouple C_{soft} changes from AABW volume and C_{dis} . These considerations highlight the potential for processes not simulated in our model to modulate the relationships and mechanisms we observe. Future work may refine our results through the inclusion of more complex representations of ocean biogeochemistry.

e. Limitations and future work

Given the idealized nature of our study, a number of glacial features and processes have not been simulated. This does not impinge upon the results presented here, which were designed to isolate and quantify the impact of the specific mechanism of cooler northern temperatures. Nevertheless, we hope that discussing the model's simplifications and unrepresented processes will provide useful context for future studies aiming to extend these results into more comprehensive glacial scenarios. Key idealizations in our setup include its single-basin configuration, flat-bottomed bathymetry, coupling of only ocean and sea ice components, and idealized, time-invariant forcing fields. We also did not set out to simulate a number of other important glacial features and processes, such as ocean volume change with ice sheet growth (Adkins et al. 2002), interactive sediments with carbonate compensation (Kobayashi and Oka 2018; Broecker and Peng 1987), enhanced dust fluxes and iron fertilization (Martínez-García et al. 2014; Lambert et al. 2015; Thöle et al. 2019), the temperature dependence of remineralization and export rates (Matsumoto 2007; Kwon et al. 2009), changes in winds and upwelling (Ai et al. 2020), and greater nutrient drawdown (Studer et al. 2015; François et al. 1997). While all these processes are important to include in comprehensive glacial simulations, we can highlight a few in particular that would be valuable to further validating the Gildor and Tziperman mechanism presented here.

One promising future direction would be to move beyond the idealized representation of the effects of Northern Hemisphere ice sheets. Incorporating a fully coupled ice sheet module in this model thus presents a logical next step for a follow-up study. Beyond thermal impacts, large ice sheets would have altered atmospheric circulation. For instance, stronger westerlies and increased wind stress (Broccoli and Manabe 1987) may have intensified wind-driven gyre circulation, increasing poleward salt transport and potentially strengthening the AMOC (Muglia and Schmittner 2015; Oka et al. 2012). Such effects may partly counteract the AMOC weakening we observe. Such complex and potentially competing interactions of ice sheet effects warrant exploration in more advanced

models. Models with interactive ice sheets (Bahadory and Tarasov 2018; Pöppelmeier et al. 2023b; Smith et al. 2021; Testorf et al. 2026) would be ideally suited to these investigations. Idealized models could also implement wind forcing based on outputs from such models.

Another improvement on this study would be to implement radiative feedbacks from carbon cycle changes. In our setup, carbon fluxes into and out of the atmosphere do not affect the radiative energy balance. Radiatively active CO₂ may introduce a positive feedback, with declining CO₂ cooling global temperatures, further cooling NADW and promoting southern sea ice. Currently, imposing warm preindustrial-like air temperatures (Fig. 2) over the Southern Hemisphere limits Southern Ocean cooling and sea ice growth in response to the northern cooling, potentially muting its full effect. Understanding such interactions and nonlinear behaviors requires further study. Outside these limitations, the value of our approach lies in isolating the thermal effects of Northern Hemisphere ice sheets on ocean carbon, providing a foundational understanding of this component's role in the broader climate system.

Finally, even given the idealized nature of our configuration, our results provide several recommendations relevant to broader glacial modeling efforts employing more complex Earth system models. First, our results support the Gildor and Tziperman mechanism and an important role for northern ice sheets in glacial change. The further development and use of models with coupled terrestrial ice sheets (e.g., Pöppelmeier et al. 2023b) is thus highly recommended for both follow-up studies and broader glacial modeling efforts. Next, in the absence of a terrestrial ice sheet module, it is important for models to capture North Atlantic conditions accurately, given their evident influence on Southern Ocean conditions. Data-matching model output to regional proxy records could provide a valuable benchmark against which to compare model output, for example. Finally, our results align with existing understanding of the importance of southern sea ice in modulating bottom-water properties (and associated stratification, etc.). This provides renewed motivation to simulate or parameterize sea ice processes accurately, a continuing area of focus in ongoing model development. Outside of glacial–interglacial CO₂ change, our results also encourage consideration of the spatial distributions of tracer anomalies alongside the conventional DIC decomposition based on mean-ocean changes. Specifically, the sea ice/salinity-driven influence over preformed alkalinity distribution, while not driving net CO₂ drawdown, reflects a major reorganization of ocean biogeochemistry induced by the dynamical changes in the high southern latitudes. This is an example of a process that is masked in the carbon pump decomposition because it does not produce a net change in CO₂ or

mean-ocean carbon, yet it remains a dynamically and biogeochemically significant feature of the ocean. This highlights the value of spatially resolved analyses, which can reveal processes shaping ocean biogeochemistry not captured by mean-ocean metrics.

Beyond the details of the Gildor and Tziperman mechanism, however, our results encourage an interhemispheric perspective to be taken on glacial change. Southern Ocean processes have been extensively studied and recognized as key drivers of glacial–interglacial carbon dynamics, but our results motivate a need to consider remote regions and factors—such as the northern ocean basins and northern ice sheets—that may exert substantial influence on Southern Ocean conditions (e.g., [Shankle et al. 2025](#)). Investigating such influences could reshape current understanding of glacial processes and reveal novel processes linking disparate parts of the climate system.

5. Conclusions

Our study offers insight into the interplay between Northern Hemisphere glaciation and Southern Hemisphere carbon dynamics. Idealized simulations show that high-latitude northern cooling—emulating the thermal effect of a large Northern Hemisphere ice sheet—significantly modulates ocean circulation and carbon storage and achieves drives substantial CO₂ drawdown, up to ~74% the magnitude (32 ppm) of the CO₂ drawdown that realized in a simulation forced by global glacial-like temperatures (43 ppm). This occurs through glacial-like ocean features such as widespread cooling, enhanced southern sea ice, saltier bottom waters, and stronger stratification between the upper and lower overturning cells. These changes enhance ocean carbon storage via the disequilibrium and particularly solubility pumps, partly offset by a weaker biological pump.

Though changes in the disequilibrium and biological pumps are partly compensatory, their emergence from the high southern latitudes (Figs. S9 and S10) is significant for showing that northern cooling can drive dynamical changes in remote southern regions salient to carbon cycling, representing an effect beyond temperature-driven solubility. The temperature-driven solubility response is itself also significant, however, with the impact of northern cooling extending beyond middepth waters to affect the full water column. This leads to a CO₂ drawdown equal to or greater than that from globally applied glacial-like cooling, highlighting the far-reaching impact of northern cooling.

These results support the [Gildor and Tziperman \(2001\)](#) hypothesis of glacial CO₂ drawdown and warrant further

testing in more comprehensive models. They underscore the importance of an interhemispheric perspective on glacial–interglacial carbon dynamics and the need to study feedbacks between distant parts of the climate system. Future work should continue to explore such interactions to build a unified theory of glacial change integrating orbital forcing, ice sheet changes, and ocean carbon cycling.

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Data availability statement.

The MITgcm source code is freely available at <http://mitgcm.org/>. The configuration files used for this study and scripts analyzing output are available at https://github.com/Maddie-Sh/ShankleEtAl2026_JClim_NHCool. [Please can you update this text from "https://github.com/Maddie-Sh/" to "https://github.com/Maddie-Sh/ShankleEtAl2026_JClim_NHCool" (and update its hyperlink too). Thank you.] Simulation output is available at <https://zenodo.org/records/21650059>.

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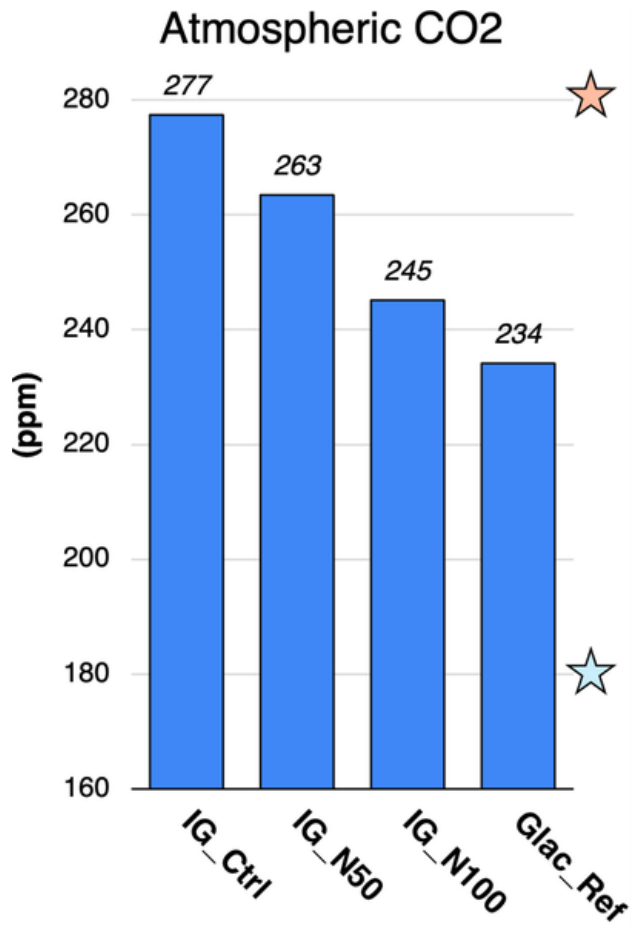
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