Global Oceanic Overturning Circulation Forced by the Competition between Greenhouse Gases and Continental Ice Sheets during the Last Deglaciation

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ABSTRACT: The deglacial change of Atlantic meridional overturning circulation (AMOC) since the Last Glacial Maximum (LGM; ~21 ka) has been studied extensively in both reconstructions and model simulations. While reconstructions suggest a shoaling of AMOC at the LGM, the strength of glacial AMOC relative to the modern day remains highly uncertain in both reconstructions and models. Using transient simulations of climate evolution forced by individual deglacial forcings since the LGM, this study shows that the uncertainties in glacial AMOC intensity can be caused by a competition between the elevated glacial Northern Hemisphere (NH) ice sheets and the reduced glacial greenhouse gases, in which the former tend to strengthen the AMOC while the latter play an opposite role. In spite of the dramatic difference of climate between the LGM and the present, the cancellation between the impacts of the two forcings leaves the strength of the glacial AMOC not too different from the modern day (0.5 Sv stronger in our study; 1 Sv = $10^6 \text{ m}^3 \text{ s}^{-1}$). Furthermore, consistent with theoretical analysis, the response of the AMOC return flow to either forcing is predominantly compensated by an interbasin exchange between the Indo-Pacific (including the Indo-Pacific sector of Southern Ocean) and Atlantic via the Agulhas Current.

KEYWORDS: Meridional overturning circulation; Greenhouse gases; Ice sheets; General circulation models

1. Introduction

As the middepth cell of the global overturning circulation confined to the Atlantic sector, the Atlantic meridional overturning circulation (AMOC) is of vital importance for the local and global climate (e.g., Buckley and Marshall 2016). Deglacial evolution of AMOC since the Last Glacial Maximum (LGM) has been studied extensively in both proxy records and model simulations. There are multiple pieces of evidence that the AMOC was shallower at the LGM. Reconstructions based on $\delta^{13}\text{C}$ and Cd/Ca have been interpreted to reflect a shoaling of AMOC by ~1000 m during the LGM, with more southern sourced waters filled the abyssal Atlantic (Duplessy et al. 1988; Curry and Oppo 2005; Marchal and Curry 2008). The shoaling is further confirmed in combining $\delta^{13}\text{C}$, Cd/Ca, and $\delta^{18}\text{O}$ measurements with a tracer transport model, but with smaller magnitude (Gebbie 2014; Oppo et al. 2018). Model simulations show that the shoaling of AMOC at the LGM can be attributed to changes in surface buoyancy forcing and deep ocean stratification (Shin et al. 2003; Liu et al. 2005; Ferrari et al. 2014; Jansen and Nadeau 2016; Jansen 2017; Sun et al. 2018). In contrast to the observed shoaling of glacial AMOC, the AMOC strength at the LGM still remains highly uncertain. In the geotracer proxy $^{231}\text{Pa}/^{230}\text{Th}$ (hereafter Pa/Th) that is

believed to be related to the export of North Atlantic Deep Water (NADW), the intensity of AMOC at the LGM ranges from somewhat weaker than (McManus et al. 2004) and indistinguishable from (Yu et al. 1996) to somewhat stronger than (Gherardi et al. 2009; Böhm et al. 2015) the present, depending on the location of the proxy. Combining Pa/Th reconstructions with a tracer model suggests that the AMOC during the LGM can be either at least as strong as that in the present (Lippold et al. 2012) or weaker than the present (Gu et al. 2020). PMIP (Paleoclimate Modeling Intercomparison Project) atmosphere–ocean general circulation models under glacial boundary conditions show agreement on neither the geometry nor the strength of the AMOC at the LGM (Otto-Bliesner et al. 2007; Weber et al. 2007; Muglia and Schmittner 2015), which has been attributed to different Antarctic sea ice formation (Marzocchi and Jansen 2017), model-specific bias (Galbraith and de Lavergne 2018), initial conditions, and insufficient equilibrium (Zhang et al. 2013; Galbraith and de Lavergne 2018). These seemingly inconsistent results in reconstructions and models nevertheless may imply that the AMOC intensity at the LGM is not too much different from the present, neither too much stronger nor too much weaker.

Physically, however, it is not obvious why AMOC intensity remains not too much different between the LGM and the modern day. During the last deglaciation, there are large changes in external climate forcing, notably a significant increase of atmospheric greenhouse gases (GHGs), including an increase in CO_2 of ~90 ppm (Monnin et al. 2001), and a large retreat of continental ice sheets, including a lowering in Laurentide Ice Sheet of over 3 km (Peltier 2004). One naturally wonders how AMOC intensity can remain not too much

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FIG. 1. Deglacial evolutions of (a) AMOC intensity (Sv), (b) AABW intensity (Sv), and (c) AMOC depth (m) in the ALL (black), GHG (red), ICE (blue), ORB (green), and MWT (purple) runs. (d)–(f) As in (a)–(c), but for deglacial changes relative to the LGM state in GHG (red), ICE (blue), and ICE+GHG (gray). Stars in (d)–(f) denote relative changes at the LGM and MOD. Normalized CO_2 concentration and continental ice sheets volume are also plotted in dashed red and blue lines in (d)–(f), respectively, corresponding to the right axis.

changed at the LGM in response to these large changes in climate forcing.

Several modeling studies have recently explored the effects of individual glacial forcings on glacial AMOC. In three independent coupled models, glacial reduction of CO_2 concentration reduces the intensity of AMOC (Zhu et al. 2015; Klockmann et al. 2016; Galbraith and de Lavergne 2018). In five independent coupled models, elevated glacial continental ice sheets, notably the Laurentide Ice Sheet, strengthen the AMOC (Zhu et al. 2014; Ullman et al. 2014; Gong et al. 2015; Klockmann et al. 2016; Galbraith and de Lavergne 2018). Reconstructions of GHG concentration (Bereiter et al. 2015) and global sea level (Spratt and Lisiecki 2016) show that the two forcing factors tend to coevolve with each other during glacial–interglacial cycles, with an increasing GHG concentration accompanying retreating ice sheets, and vice versa, such as in the last deglaciation (Fig. 1). If the AMOC response to GHG and ice sheets is correct in the model simulations discussed above, the coevolution of the two forcings will impose a competition effect on AMOC during glacial–interglacial cycles. For example, through the last deglaciation, the rising GHG concentration tends to strengthen the AMOC, while the retreating ice sheets tend to weaken the AMOC (Figs. 1a,d). This leads to a forcing-competition hypothesis that the impacts of the two major forcings compensate each other, leading to

an AMOC intensity that does not change much between the LGM and the modern day. A credible test of this competition hypothesis would require ultralong (thousands of years) simulations forced by the two forcing factors in the same fully coupled model and satisfy, at least, the prerequisite of the reproduction of the “correct” AMOC geometry: a shoaling of AMOC at the LGM, as observed across proxies. Using MPI-ESM, Klockmann et al. (2016) found the competition effect between GHG and ice sheets in the deep ocean, but their model produces no shoaling of the AMOC in their glacial reference simulation. Galbraith and de Lavergne (2018) integrated GFDL-ESM2 to equilibrium under different combinations of three external forcings (greenhouse gases, ice sheets, and orbital configuration), but as they noted, “none of the simulations is specifically designed to simulate the present interglacial (the Holocene) or the LGM” (p. 658).

In the Transient Climate Evolution (TraCE) experiment over the last 21 000 years (TraCE21; Liu et al. 2009; He 2011), several deglacial forcing sensitivity runs are performed in a coupled general circulation model (CGCM)—Community Climate System Model version 3 (CCSM3) to isolate the climate impact of individual climate forcings. Using TraCE21 simulations, Zhu et al. (2014) found the linear AMOC weakening in response to deglacial receding ice sheets. In their Fig. 1, Zhu et al. (2015) also showed that deglacial increase of GHG enhances the

AMOC. In addition, TraCE21 simulations also include the experiments forced by orbital forcing, meltwater forcing, and all the forcing combined. Therefore, this set of five TraCE21 simulations will allow us to study the full deglacial AMOC response and the relative contribution of each forcing. Taking advantage of TraCE21, we analyze the deglacial AMOC change under different external forcings and offer one plausible explanation to the ambiguous AMOC strength at the LGM as seen in both proxy records and models.

The other purpose of this paper is to understand the slow evolution of global overturning circulation in a simple theoretical framework. Forced by slowly evolving external forcings, such as GHG concentration and continental ice sheets, the climate response can be treated as a quasi-equilibrium solution. Therefore, the previously proposed theoretical models for overturning circulation can be applied to TraCE21 simulations. As a pioneer study, Gnanadesikan (1999) proposed a theoretical single basin (connected to a reentrant channel) model of overturning circulation, relating North Atlantic sinking to low-latitude upwelling and Southern Ocean winds and eddies. Despite its simplicity, it has been widely used to interpret the results from full three-dimensional simulations of global overturning circulation (e.g., Allison et al. 2011; Munday et al. 2011; Nikurashin and Vallis 2012). Recent studies further extended Gnanadesikan's framework to a two-layer and two-basin model and found an interbasin overturning regime (e.g., Jones and Cessi 2016; Ferrari et al. 2017; Sun et al. 2020; Nadeau and Jansen 2020), which has been shown to be critical for the interpretation of the global overturning circulation pathways observed in today's ocean (e.g., Schmitz 1995; Lumpkin and Speer 2007; Talley 2013). With more realistic water-mass transformations and ocean–land geometry, the global overturning circulation in CGCM should be more like the observed figure-eight pattern of Talley (2013) that spans the Atlantic, the Southern Ocean, and the Indo-Pacific. In Talley's framework, sinking in the North Atlantic is largely compensated by wind-driven upwelling in the Southern Ocean and interior diapycnal upwelling in the deep Indo-Pacific basin. The closure of global overturning circulation through multiple basins can be partly reproduced in the two-basin models (Jones and Cessi 2016; Cessi and Jones 2017; Ferrari et al. 2017). In our TraCE21 simulations, the circulation changes in the Indo-Pacific Ocean and the Southern Ocean will be seen playing an important role in deglacial global overturning circulation response, consistent with the theoretical two-basin model and observations.

The paper is organized as follows. The model and experiments are briefly described in section 2. In section 3, we first investigate the forcing-competition effect in deglacial AMOC and the related changes in local deep water formation; then we introduce the theoretical two-basin model of global overturning circulation and apply it to TraCE21 simulations. A summary and discussion are given in section 4.

2. Model and experiments

The National Center for Atmospheric Research (NCAR) CCSM3 (Yeager et al. 2006) is used in TraCE21 experiment to simulate the climate evolution from the LGM (Liu et al. 2009;

He 2011). The ocean model is the NCAR implementation of the Parallel Ocean Program (POP) in vertical z coordinates with 25 levels. The longitudinal resolution is uniformly 3.6° and the latitudinal resolution is nominally 3° with finer resolution near the equator ($\sim 0.9^\circ$). More details of the model setup can be found in He (2011). The TraCE21 experiment consists of five simulations corresponding to different deglacial forcings: greenhouse gas forcing (GHG; 22–0 ka, where ka indicates “thousand years ago”), continental ice sheets forcing (ICE; 19–0 ka), orbital forcing (ORB; 22–0 ka), meltwater forcing (MWT; 19–0 ka), and all forcings combined along with ocean bathymetry changes (ALL; 22–0 ka). TraCE21 simulations are performed at the Oak Ridge National Laboratory (ORNL) computing facility. The early part of ALL (22–12.9 ka) is performed on supercomputer Phoenix while the rest of ALL and all the single-forcing simulations are performed on supercomputer Jaguar. Greenhouse gas concentrations are prescribed following Monnin et al. (2001) and ice sheets elevations follow Peltier (2004). TraCE21 captures many important features of glacial AMOC in its all forcing experiment ALL, including a shoaling of NADW and a volumetric expansion of Antarctic Bottom Water (AABW; Figs. 1c and 2a,d; Liu et al. 2009). Effects of deglacial orbital forcing on ocean circulation are small compared with other forcings, and meltwater forcing is more dominant on a millennial scale (Fig. 1; He 2011); therefore, we will mainly focus on the ALL, GHG, and ICE runs in the following. The time resolution for all the variables used in this study is decadal mean. Climatologies for the LGM and modern (MOD) states are taken as the 1000-yr means of 21 and 1 ka, respectively. We note here that only Eulerian velocities [and thus the Eulerian meridional overturning circulation (MOC)] are considered in the present study since eddy velocities are not archived. For our analysis of overturning transports north of 34°S , this is unlikely to cause a big error. Indeed, we have checked another ocean-alone transient simulation C-iTraCE of comparable resolution (Gu et al. 2019) and found that the eddy-induced transport is small in most regions north of 34°S in both the mean state (less than 1 Sv ; $1\text{ Sv} \equiv 10^6\text{ m}^3\text{ s}^{-1}$) and deglacial response (less than 0.5 Sv) (not shown).

3. Results

a. Forcing-competition effect in deglacial AMOC

The forcing-competition effect can be detected directly from the time series of deglacial overturning circulation in TraCE21. AMOC intensity is diagnosed as the maximum in the Atlantic overturning streamfunction below 500 m over 20° – 50°N . The intensity of the lower cell associated with AABW is diagnosed as the minimum in the Atlantic overturning streamfunction below 2000 m over 10° – 34°S . The depth of AMOC is diagnosed as the depth of zero streamfunction between 1000 and 3500 m in the Atlantic. We note that the strength of the AMOC at the LGM is 12.6 Sv in ALL and 11.5 Sv in the single forcing runs (Fig. 1a). The most likely reason for the $\sim 1\text{ Sv}$ difference is the change of machine from Phoenix to Jaguar at ORNL. To make the comparison between the LGM and MOD reasonable, in the following, unless otherwise specified, both the LGM and MOD states refer to those simulated on Jaguar.

FIG. 2. Mean climatologies of global Eulerian overturning circulation (red/blue filled contour; Sv) and zonal mean isopycnals (σ_t ; potential density referenced to 2000 m; black contour; kg m^{-3}) at the (a)–(c) LGM and (d)–(f) MOD periods. Also shown are the MOD states obtained from two single forcing runs: (g)–(i) GHG and (j)–(l) ICE. Columns from left to right represent the Atlantic basin, Indo-Pacific basin, and global ocean basin, respectively. Zonal mean annual mixed layer depths are plotted in green lines for the Atlantic basin and Indo-Pacific basin. Note that to visualize the connection between the basins, the sense of circulation for the Atlantic basin (at left) is opposite to the one for the Indo-Pacific basin (at center) (i.e., red contour indicates counterclockwise flow in the left panels and clockwise flow in the center panels). Note the different latitudinal ranges for individual basins and global basin.

For deglacial change of AMOC, although it is punctuated by millennial events (caused predominantly by meltwater forcing; Fig. 1a), the final change between the LGM and MOD states is only 0.5 Sv (Fig. 1d). The strength of the AMOC tends to evolve monotonically and gradually in the two single forcing runs GHG and ICE, closely following the evolutions of the prescribed GHG concentration and ice sheets volume (Fig. 1d). The LGM–MOD AMOC change in TraCE21 can be well reproduced by the linear superposition of the impacts simulated in the GHG and ICE runs (Fig. 1d, stars vs gray line). The AMOC

weakening forced by retreating continental ice sheets (~ -4 Sv) and AMOC strengthening forced by increasing GHG concentration (~ 4.5 Sv) compete with each other, leaving little LGM–MOD AMOC change (Figs. 1b and 3a,d,g). For the abyssal Atlantic, the AABW is reduced by deglacial GHG forcing by ~ 1 Sv and continental ice sheets by ~ 0.5 Sv, such that the total Atlantic AABW weakening of ~ 1.5 Sv is also well reproduced by the linear superposition of GHG and ICE (Fig. 1e, stars vs gray line). The depth of AMOC deepens by ~ 500 m from the LGM to MOD, matching especially well with Gebbie (2014).

FIG. 3. Similar to Fig. 2, but for deglacial changes of overturning circulation (black contours with interval of 1 Sv) and potential density (σ_2 ; shading) for each run and each basin. Note the larger magnitude of the color bar for the ALL run.

About 50% of the deepening is reproduced by the linear superposition of GHG (deepening by ~ 350 m) and ICE (shoaling by ~ 100 m) while the remaining 50% is probably accounted for by the nonlinear interactions (Fig. 1f).

The competition between the impacts of GHG and ice sheets on AMOC strength in TraCE21, which has been shown in Zhu et al. (2014, 2015), agrees well with the previous equilibrium simulations (Klockmann et al. 2016), while the depth change of AMOC is captured in TraCE21 only. Furthermore, the linear strengthening (weakening) of the AMOC (AABW) in response to increasing GHG concentration is also consistent with the equilibrium simulations of Galbraith and de Lavergne (2018), in which the CO_2 concentration increases gradually from 147 (glacial value) to 270 ppm (preindustrial value). These consistencies add confidence in using TraCE21 simulations to investigate the mechanisms controlling the deglacial AMOC response.

b. Mechanisms for deglacial AMOC change in the Atlantic perspective

The deglacial change of deep water formation in the North Atlantic and the resulting AMOC change have been diagnosed using TraCE21 in previous studies. The linear weakening response

of AMOC to receding ice sheets in the ICE run has been fully discussed in Zhu et al. (2014) in terms of a wind-driven mechanism as follows: lowering the NH ice sheets shifts the westerly jet northward to the subpolar North Atlantic, as indicated by the dipole barotropic streamfunction anomaly (Fig. 4b); this wind shifting promotes eastward expansion of sea ice over deep convection region in the subpolar North Atlantic, which further reduces the oceanic heat loss and suppresses deep water formation (Fig. 4d). A similar influence of continental ice sheets on AMOC strength through wind changes has also been found in other models independently (Ullman et al. 2014; Gong et al. 2015; Klockmann et al. 2016).

Transient AMOC behavior under GHG forcing in glacial climate has also been investigated by Zhu et al. (2015) along with additional CCSM3 simulations forced by suddenly doubled CO_2 concentration. At the LGM, the Greenland, Iceland, and Norwegian Seas (hereafter referred to as the GIN Seas) are covered by permanent sea ice (Fig. 5a). The isolation effect of sea ice prohibits oceanic heat loss from the sea surface, shutting off the deep water formation there. Under deglacial GHG forcing, sea ice retreats significantly and the convection over GIN Seas is reactivated in response to enhanced oceanic heat loss (Fig. 5b). The GIN Seas overflow crosses the

FIG. 4. (a),(c) North Atlantic climatology at the LGM and (b),(d) changes at the MOD in the ICE run, shown (top) for barotropic streamfunction (BSF; shading; Sv) and annual mixed layer depth [MLD; contour; m, interval of 100 m with positive (negative) value in solid (dashed) lines] and (bottom) for sea surface heat flux [SHF; shading; W m^{-2} , negative (positive) value means oceanic heat loss (gain)] and sea ice margin (defined as 15% annual sea ice coverage). Blue, green, and red lines in (c) and (d) represent sea ice margin at the LGM, 13.5 ka, and MOD, respectively.

Greenland–Scotland Ridge through Denmark Strait (Fig. 5d) and feeds into the deep North Atlantic, strengthening the AMOC. The slow enhancement of the AMOC has also been partly attributed to the weakening deep stratification caused by the reduced AABW in the abyssal North Atlantic (Figs. 2g and 3d; Shin et al. 2003; Liu et al. 2005).

It is still under debate whether the GIN Seas overflow is present during the LGM. Early reconstructions of the LGM suggest permanent sea ice cover over the GIN Seas (Labeyrie et al. 1992), much weakened northward warm and salty water transport, and the absence of convection in the GIN Seas (Duplessy et al. 1988), while some more recent paleo-proxies suggest at least seasonally open waters at the GIN Seas and intermittent overflow (Meland et al. 2008; Crockett et al. 2011). Model simulations including TraCE21 support depressed convection over the GIN Seas and inactive overflow during the LGM (Figs. 5a,c; Liu et al. 2009; Klockmann et al. 2016). Furthermore, TraCE21 suggests that the low GHG conditions at the LGM accounts for the depressed GIN Seas convection and overflow.

c. A global perspective for deglacial AMOC change

The diagnostic analysis of the deglacial AMOC above is confined locally in the Atlantic basin, with no consideration of its connection with the overturning circulation in the

Indo-Pacific and the Southern Ocean. Here, we will further study the AMOC response from the global perspective. The simulated modern-day oceanic overturning circulation exhibits a strong asymmetry between the Atlantic basin and the Indo-Pacific basin (Fig. 2). In the Atlantic, the overturning streamfunction flows largely along isopycnals, indicating a quasi-adiabatic interhemispheric circulation (Fig. 2d). The quasi-adiabatic circulation in the North Atlantic is augmented by a modest diffusively driven positive cell north of 20°N ($\sim 2\text{ Sv}$). In the Indo-Pacific, however, the overturning streamfunction flows across the isopycnals from the abyss all the way to $\sim 1000\text{ m}$, indicative of a strong diapycnal (diabatic) upwelling (Fig. 2e). This interbasin asymmetry simulated in TraCE21 is consistent with the theoretical two-basin overturning models (Jones and Cessi 2016; Ferrari et al. 2017) and the observed figure-eight pattern of global overturning circulation (Talley et al. 2011; Talley 2013). The diapycnal upwelling in the Indo-Pacific, which is absent in the single Atlantic basin model, has been shown to be critical to the closure of global ocean heat, freshwater, and mass budgets in reality (Talley 2003, 2008, 2013).

As will be shown later, the deglacial change of AMOC is accomplished by an interbasin compensation between Atlantic and Indo-Pacific (including the Indo-Pacific sector of the Southern Ocean). In the following, we will first introduce the

FIG. 5. (a),(c) North Atlantic climatology at the LGM and (b),(d) changes at the MOD in the GHG run, showing (top) sea surface heat flux (SHF; shading; W m^{-2}) and annual mixed layer depth (MLD; contour; m, interval of 100 m) and (bottom) ocean flow at 326 m (vector; cm s^{-1}) and bathymetry (shading; m). Blue and red lines in (a) and (b) represent sea ice margin at the LGM and MOD, respectively.

theoretical two-basin overturning model and then apply it to GHG and ICE runs to better understand the deglacial AMOC change from the global perspective.

1) A THEORETICAL MODEL FOR GLOBAL OVERTURNING BUDGET

We now discuss our deglacial simulation of AMOC in terms of a theoretical overturning model. The geometry of the

theoretical model largely follows the two-basin model of Ferrari et al. (2017) (Fig. 6). The theoretical model is a reduced gravity model, consisting of two layers separated by an isopycnal (blue surface in Fig. 6). The interface separates NADW and Intermediate Water in the Atlantic. In our realistic model, the equivalent isopycnal for the LGM (MOD) state would therefore correspond to $38.0\sigma_2$ ($36.8\sigma_2$) (Fig. 2). Different from the previous two-basin models, in the present model we

FIG. 6. Theoretical two-basin overturning model configuration used in the present study. It largely follows Ferrari et al. (2017) with the addition of a weak North Pacific sinking T_{NP} . The budget for the overturning circulation above the isopycnal dividing the upper and lower branch of the MOC is examined. The terms h_a and h_p are the depths of interface along the eastern boundaries of Atlantic basin and Indo-Pacific basin, respectively. See main text and appendix for details on each term.

introduce a weak North Pacific sinking. As can be seen from Fig. 2e, besides the deep water formed in the North Atlantic, there is also a weak intermediate water formation in the North Pacific characterized by an overturning maximum of ~ 4 Sv located at $\sim 35^\circ\text{N}$, namely the North Pacific Intermediate Water (NPIW) (Fig. 2). This weak analog of NADW formation in the North Pacific is consistent with present observations (Talley 2013). Considering the mass conservation, similar to the single basin model of Gnanadesikan (1999), the volume budget for the upper branch of global overturning circulation takes the following form:

$$T_{\text{EK}}^{\text{so}} - T_{\text{ed}}^{\text{so}} = T_N - T_u^{\text{glob}}, \quad (1)$$

where T_N denotes the total northern sinking associated with NADW and NPIW, T_u^{glob} denotes the interior diapycnal upwelling within the global basin, and $T_{\text{EK}}^{\text{so}}$ and $T_{\text{ed}}^{\text{so}}$ are the Ekman and eddy transports across the northern boundary of the Southern Ocean (ϕ_s), respectively. Note here the diapycnal upwelling in the strip between southern tips of the model's South Africa (ϕ_p) and model's South America (ϕ_s) is ignored due to its limited area.

With our focus on the AMOC response and its connection with other basins, we further consider the volume budgets for individual subdomains. The budget for the L-shaped region covering the Atlantic basin and the southern strip between latitudes ϕ_p and ϕ_s is given by

$$T_{\text{EK}}^{\text{so}} - T_{\text{ed}}^{\text{so}} + T_g = T_{\text{NA}} - T_u^{\text{Atl}}, \quad (2)$$

where T_{NA} denotes the sinking in the North Atlantic (NADW), T_u^{Atl} denotes the diapycnal upwelling within the Atlantic basin, and T_g denotes geostrophically balanced transport from the Indo-Pacific basin to the Atlantic basin. A positive (negative) T_g means that seawater enters (exits) the Atlantic basin.

For simplicity, we will then ignore the net Ekman transport at the southern edge of the two basins (ϕ_p) since the zonal wind stress is close to its minimum at that latitude (Fig. 8; Ferrari et al. 2017). As mentioned before, the circulation pattern in the Indo-Pacific basin is characterized by a strong diapycnal upwelling (Fig. 2). According to Talley's figure-eight framework (Talley 2013), the basinwide upwelling transfers AABW to Indian Deep Water (IDW) and Pacific Deep Water (PDW). Part of these deep waters return to the Southern Ocean, upwelled adiabatically and blown northward by winds, feeding the northward Ekman transport. This part has been included in the net Ekman transport term $T_{\text{EK}}^{\text{so}} - T_{\text{ed}}^{\text{so}}$ in Eq. (2). The other part of these deep waters however, is continuously upwelled into the thermocline within the Indo-Pacific basin. Here in our theoretical model, the diapycnal upwelling in the Indo-Pacific (T_u^{IP}) refers to the latter part, of which a small amount feeds NPIW formation and the rest exits the Indian Ocean via the Agulhas Current and enters the South Atlantic geostrophically. Therefore, the volume budget for the Indo-Pacific overturning circulation is

$$T_g = T_u^{\text{IP}} - T_{\text{NP}}, \quad (3)$$

where T_u^{IP} denotes the diapycnal upwelling from deep waters to thermocline waters within the Indo-Pacific basin; T_{NP} denotes

the weak sinking in the North Pacific associated with NPIW, which is usually omitted in previous two-basin models.

Further dividing Southern Ocean into the Atlantic sector ("so_A") and Indo-Pacific sector ("so_IP") and substituting them into Eq. (2), we have

$$(T_{\text{EK}}^{\text{so}_A} - T_{\text{ed}}^{\text{so}_A}) + (T_{\text{EK}}^{\text{so}_{\text{IP}}} - T_{\text{ed}}^{\text{so}_{\text{IP}}}) + T_g = T_{\text{NA}} - T_u^{\text{Atl}}. \quad (4)$$

Similar to Cessi and Jones (2017), we further define the interbasin exchange term T_G as the sum of the net Ekman transport convergence within the Indo-Pacific sector of Southern Ocean between latitudes ϕ_p and ϕ_s (note that the net Ekman transport at latitude ϕ_p has been assumed to be zero) and the geostrophically balanced transport from the Indo-Pacific basin to the Atlantic basin:

$$T_G = (T_{\text{EK}}^{\text{so}_{\text{IP}}} - T_{\text{ed}}^{\text{so}_{\text{IP}}}) + T_g = (T_{\text{EK}}^{\text{so}_{\text{IP}}} - T_{\text{ed}}^{\text{so}_{\text{IP}}}) + (T_u^{\text{IP}} - T_{\text{NP}}). \quad (5)$$

The interbasin exchange is in geostrophic balance and proportional to the difference between interface depths along eastern boundaries of the Atlantic and Indo-Pacific Oceans (Jones and Cessi 2016; Cessi and Jones 2017; Ferrari et al. 2017). By introducing T_G , the budget for the L-shaped region [Eq. (1)] now reduces to the budget for Atlantic basin plus Atlantic sector of Southern Ocean (hereafter referred to as Atlantic domain):

$$(T_{\text{EK}}^{\text{so}_A} - T_{\text{ed}}^{\text{so}_A}) + T_G = T_{\text{NA}} - T_u^{\text{Atl}}. \quad (6)$$

With eddy-induced anomalous transport being neglectable as discussed in section 2, the overturning budgets for the global basin [Eq. (1)], Atlantic domain [Eq. (6)], and interbasin exchange [Eq. (5)] can now be written in perturbation forms, respectively:

$$\Delta(T_{\text{EK}}^{\text{so}} + T_u^{\text{glob}} - T_N) = 0, \quad (7)$$

$$\Delta(T_{\text{EK}}^{\text{so}_A} + T_G + T_u^{\text{Atl}} - T_{\text{NA}}) = 0, \quad (8)$$

$$\Delta T_G = \Delta T_{\text{EK}}^{\text{so}_{\text{IP}}} + \Delta T_g = \Delta T_{\text{EK}}^{\text{so}_{\text{IP}}} + \Delta T_u^{\text{IP}} - \Delta T_{\text{NP}}. \quad (9)$$

To put our CGCM overturning circulation into the theoretical framework, we first divide the Atlantic and global MOCs in TraCE21 into a pole-to-pole part (adiabatic part) and diapycnal upwelling part (diabatic part) (Fig. 7). Similar to the definition of AMOC intensity, the global MOC (GMOC) intensity is defined as the maximum in the global overturning streamfunction below 500 m over the northern combined basin. The equivalent terms for AMOC and GMOC in the theoretical model are T_{NA} and T_N , respectively. The pole-to-pole part is diagnosed as the maximum in the overturning streamfunction below 500 m at the southern edge of the two basins (ϕ_p ; $\sim 32^\circ\text{S}$), which represents the net wind-driven Ekman transport for GMOC [left-hand side of Eq. (1)] and the sum of the net Ekman transport and interbasin exchange for AMOC [left-hand side of Eq. (2)]. For GMOC (AMOC), the diapycnal upwelling is therefore the residual difference between the GMOC (AMOC) intensity and its pole-to-pole part,

FIG. 7. Decompositions of MOC intensity (black) into a pole-to-pole part (red) and diapycnal upwelling part (blue) for (top) the Atlantic and (bottom) the globe in the (a),(c) ALL, (b),(e) GHG, and (c),(f) ICE runs. For each part, the definition and the equivalent term in the theoretical model can be found in [section 3c\(1\)](#).

corresponding to T_u^{glob} (T_u^{Atl}). This decomposition of MOC has previously been used in a theoretical model and a general circulation model (MITgcm) with idealized geometry ([Wolfe and Cessi 2011](#); [Jones and Cessi 2016](#)) and is now applied to a realistic CGCM. In the mean climatology, both AMOC and GMOC are dominated by pole-to-pole circulations ([Fig. 7](#), black and red lines). However, their deglacial changes differ significantly for different forcings and different basins. In the following, we first analyze the GHG run and then the ICE run in the theoretical framework.

2) DEGLACIAL OVERTURNING BUDGET FOR GHG RUN

Forced by increasing GHG, the northern sinking associated with GMOC (T_N) is increased by ~ 4.5 Sv, which is perfectly compensated by an increase in diapycnal upwelling ([Fig. 7e](#)). Given the small change in Atlantic upwelling ($\Delta T_u^{\text{Atl}} \sim 0$; [Fig. 7b](#), blue line), the increasing diapycnal upwelling must happen predominantly in the Indo-Pacific basin. Furthermore, change in pole-to-pole part is neglectable ([Fig. 7e](#), red line), consistent with the minor change in deglacial Southern Hemisphere (SH) westerly in both strength and position ([Figs. 8a,b](#)). Therefore, the GMOC response under GHG forcing falls into a diabatic limit.

The anomalous overturning circulation in the Indo-Pacific is rather incoherent between the two hemispheres ([Fig. 3e](#)). The diapycnal upwelling in the South Indo-Pacific is significantly enhanced ($\Delta T_u^{\text{IP}} > 0$), while the sinking in the North Pacific is almost unchanged ($\Delta T_{\text{NP}} \sim 0$). The interhemispheric asymmetry implies that the enhanced diapycnal upwelling in the South Indo-Pacific is a pure oceanic compensation between overturning circulations of the Atlantic and Indo-Pacific. In the

diabatic limit and considering $\Delta T_{\text{NP}} \sim 0$, the deglacial global overturning balance in [Eq. \(7\)](#) for GHG run reduces to

$$\Delta T_u^{\text{IP}} \sim \Delta T_u^{\text{glob}} \sim \Delta T_N \sim \Delta T_{\text{NA}}. \quad (10)$$

The diapycnal upwelling in the deep Indo-Pacific transfers deep waters to lighter thermocline waters, which must be balanced by a downward diffusive flux of heat driven by turbulent diapycnal mixing ([Munk 1966](#)). These fluxes can be related to the interface depth D between the AABW cell and NPIW cell in the Indo-Pacific; that is,

$$T_u^{\text{IP}} = \frac{K_v A}{D}, \quad (11)$$

where K_v denotes the vertical diffusion coefficient and A is the interfacial area of the Indo-Pacific. Here we define D as the depth of zero meridional streamfunction between 600 and 1500 m at a representative latitude of 25°S in the Indo-Pacific, separating the AABW cell and NPIW cell. We note that the depth scale D is defined differently from the single basin case of [Gnanadesikan \(1999\)](#) in which the pycnocline depth is used (a discussion on various scale depths can be found in the [appendix](#)). We argue that our depth scale D may be more reasonable for the real global ocean, where the dominant upwelling and sinking happen in Indo-Pacific and Atlantic, respectively, instead of within one single basin. Consistent with the prediction from [Eq. \(11\)](#), a shoaling in D is observed, corresponding to the strengthening diapycnal upwelling in the southern Indo-Pacific ([Fig. 3](#)) and the strengthening sinking in the North Atlantic ([Fig. 9a](#)).

The excessive thermocline waters, supplied by the strengthening diapycnal upwelling in the southern Indo-Pacific, enter the

FIG. 8. Hovmöller diagrams of (left) deglacial circumglobal zonal mean and (right) Atlantic zonal mean zonal wind stress (dyn cm^{-2}) in the (a),(b) GHG, (c),(d) ICE, and (e),(f) ALL runs. In each panel, the gray line depicts the shift of maximum in zonal wind stress. Blue and green dots mark the model latitudes of the southern tips of South Africa and South America, respectively.

South Atlantic geostrophically, closing the Atlantic overturning budget:

$$\Delta T_G \sim \Delta T_u^{\text{IP}} \sim \Delta T_{\text{NA}}. \quad (12)$$

Under deglacial GHG forcing, the interbasin exchange between the Indo-Pacific and Atlantic Oceans is increased, ultimately feeding the strengthening deep water formation in the North Atlantic. The value of T_G can be quantitatively estimated by taking the difference of the vertically integrated zonal transports above the AMOC maximum along the two meridional sections: from the southern tip of South America to the coast of Antarctica and from the southern tip of South Africa to the coast of Antarctica (Cessi and Jones 2017). At the LGM, the interbasin exchange diagnosed as the horizontal convergence is about 8 Sv, transporting water from the Indian Ocean to the Atlantic Ocean through the Agulhas Current, corresponding to the systematically deeper isopycnal depth along eastern boundary of the Atlantic than that of the Indo-Pacific (see the appendix). The patterns of deglacial circulation change are shown in Fig. 10. The strengthening of AMOC is associated with a stronger northward flow from the South Atlantic to deep water formation region in the North Atlantic (Figs. 10a,c). This anomalous cross-equator transport is fed by a stronger Agulhas Current, supported by

increased diapycnal upwelling in the southern Indo-Pacific (Fig. 3e). The circulation change within the upper limb of AMOC is consistent with Saenko et al. (2004) and Cai et al. (2006). The enhanced interbasin exchange is accomplished by an increased contrast between isopycnal depths along eastern boundaries of the South Atlantic and the southern Indo-Pacific (Figs. 10a,c; also see the appendix). Not only for equilibrium states, this compensation relationship between changes in AMOC (ΔT_{NA}) and interbasin exchange (ΔT_G) also holds for the transient response (Fig. 11a), as has also been suggested by Sun et al. (2020) in an idealized simulation.

3) DEGLACIAL OVERTURNING BUDGET FOR THE ICE RUN

Different from the GHG run, the deglacial GMOC change in the ICE run follows an adiabatic limit. As response to receding continental ice sheets, the overturning circulation is reduced (Figs. 1 and 3), which leads to a surface warming at southern high latitudes by nearly 3°C (not shown). The reduced temperature gradients between mid and high latitudes in the SH weaken the circumpolar westerly (Fig. 8c), leading to a weakening pole-to-pole circulation. As can be seen from Figs. 7c and 7f, changes of AMOC and GMOC have only small diabatic components ($\Delta T_u^{\text{Atl}} \sim 0$, $\Delta T_u^{\text{glob}} \sim 0$); instead,

FIG. 9. Scatterplots for (a) AMOC (T_{NA}) and $1/D$ in the GHG run and (b) AMOC (T_{NA}) and weighted average of zonal wind stress between 65° and 40°S in the ICE run. (c),(d) As in (a) and (b), but for GMOC (T_N). Each dot represents a 250-yr mean, starting from the LGM to MOD with color from blue to red. The gray diagonal lines in (a) and (c) serve as references for the linear relationship, while the gray diagonal lines in (b) and (d) serve as references for the 1:1 relationship between the relative changes of the two variables.

they are predominantly in adiabatic limits. Therefore, the deglacial global overturning balance in Eq. (7) for the ICE run reduces to

$$\Delta T_{EK}^{so} \sim \Delta T_N. \quad (13)$$

The scaling of Ekman transport across the northern boundary of the channel is

$$T_{EK}^{so} \sim \frac{\tau_{so}}{\rho f_s} l_s, \quad (14)$$

where τ_{so} is the mean zonal wind stress blowing along the northern boundary of the channel, ρ is the density of seawater, and f_s and l_s are the Coriolis frequency and the circumpolar length at latitude ϕ_s , respectively.

Substituting Eq. (14) in Eq. (13), we have

$$\frac{\Delta \tau_{so}}{\rho f_s} l_s \sim \Delta T_N. \quad (15)$$

The theoretically linear response of GMOC (ΔT_N) to Southern Ocean winds is well reproduced in ICE (Fig. 9d). In the mean climatology, the net wind-driven Ekman transport accounts for 90% of the GMOC transport (Fig. 7f) and thus the 1:1 scaling relationship between their change ratios can be expected during the deglaciation. As is clearly seen in Fig. 9d, the deglacial weakening of 15% (2.3 Sv) in GMOC agrees excellently with the 15% reduction in Southern Ocean winds, confirming the dominant quasi-adiabatic response of GMOC as described in Eq. (15).

For the Atlantic, however, winds change is not the only driver for deglacial AMOC change in ICE. Given the wind contribution of -2.3 Sv and total AMOC change of -4.0 Sv, according to the perturbation form of Eq. (2), there must be a

FIG. 10. Deglacial changes (MOD minus LGM) of circulation (vector; cm s^{-1}) and isopycnal depth (shading; m) in (a),(c) GHG and (b),(d) ICE along $37.6\sigma_2$ and $37.8\sigma_2$, which are both within the upper branch of AMOC (see Figs. 2a,g,j).

~ 2 Sv decrease in geostrophically balanced transport (T_g). Under deglacial ice sheets forcing, there seems to be a seesaw response between the North Atlantic sinking and the North Pacific sinking. The ~ 2 Sv increase in NPIW (Fig. 3h) is likely triggered by atmospheric teleconnection associated with the weakening of AMOC (Hu et al. 2012). In the absence of change in the diapycnal upwelling (Figs. 7c,f), strengthening of NPIW will itself lead to a reduced transport from the Indo-Pacific basin to the South Atlantic, compensating the other ~ 2 Sv reduction in AMOC [according to the perturbation form of Eq. (3)]. Noting the change of SH winds occurs mostly in the Indo-Pacific sector (Figs. 8c,d) and substituting Eq. (13) in Eqs. (8) and (9), we find that

$$\Delta T_G \sim \Delta T_{\text{EK}}^{\text{so-IP}} - \Delta T_{\text{NP}} \sim \Delta T_{\text{NA}}. \quad (16)$$

Forced by deglacial receding ice sheets, both the enhanced NPIW formation (ΔT_{NP}) and the reduced northward Ekman transport ($\Delta T_{\text{EK}}^{\text{so-IP}}$) contribute to the reduced interbasin exchange, ultimately compensating the reduced deep water formation in the North Atlantic (ΔT_{NA}). Opposite to GHG run, there is a weaker cross-equator northward transport. The reduction in AMOC return flow is geostrophically compensated by a weaker Agulhas Current (Figs. 10b,d), accompanied by a decreased contrast between isopycnal depths along the eastern boundaries of the South Atlantic and the southern Indo-Pacific (Figs. 10b,d; see the appendix). The transient compensation

behavior between ΔT_{NA} and ΔT_G in ICE is also robust during the deglaciation (Fig. 11b).

4. Summary and discussion

In this study, we investigate the deglacial evolution of global overturning circulation using a set of transient simulations isolating effects of individual deglacial forcings. The TraCE21 simulation reproduces the shoaling of AMOC under glacial climate, matching especially well with the solution of Gebbie (2014). On the basis of the reproduction of AMOC geometry, we further offer one plausible explanation for the ambiguous AMOC strength difference between the LGM and the present, namely the forcing-competition effect: increasing GHG concentration leads to sea ice melting in the GIN Seas, activating the convection there through enhanced oceanic heat loss and finally causing a strengthening AMOC; lowering the NH ice sheets, on the other hand, forces a weakening AMOC through northward westerly shift and the following sea ice expansion and reduced oceanic heat loss in the North Atlantic. Deglacial AMOC change in the all-forcing run (ALL) can be well reproduced by the linear superposition of its responses to the two single-forcing runs GHG and ICE. This forcing-competition effect may have been operating during glacial-interglacial cycles in Earth history given the coevolution between ice sheets and

FIG. 11. Scatterplots of anomalous interbasin transport ΔT_G diagnosed as zonal divergence and anomalous AMOC transport $\Delta AMOC$ (ΔT_{NA}) in (a) GHG and (b) ICE (units are Sv). Both anomalies are relative to the LGM state. Each dot represents 250-yr averaged anomaly relative to the LGM (21 ka), starting from the LGM to MOD with color from blue to red. The gray diagonal line in each plot serves as a reference for the quantitative 1:1 relationship.

greenhouse gases as observed in reconstructions (Bereiter et al. 2015; Spratt and Lisiecki 2016).

The deglacial evolution of the oceanic overturning circulation under each forcing is also investigated in light of a simple theoretical two-basin overturning model. To our best knowledge, it is the first attempt applying such a theoretical model to realistic CGCM simulations. We show that the pattern of global overturning circulation in our simulations is qualitatively consistent with Talley's figure-eight framework and can be partly represented by the two-basin theoretical model. With some simplifications (and thus sacrificing some realism), the deglacial change of GMOC falls into diabatic (adiabatic) limit under deglacial greenhouse gas (ice sheet) forcing. The strengthening of AMOC in response to increasing greenhouse gases is dominantly compensated by enhanced diapycnal upwelling in the southern Indo-Pacific Ocean, while the weakening of AMOC in response to receding ice sheets is compensated by reduced Ekman transport in the Southern Ocean and enhanced NPIW formation in the North Pacific. The circulation changes outside the Atlantic basin finally feed into AMOC return flow through anomalous interbasin exchange within the Agulhas Current.

Our general ideas for deglacial response of oceanic overturning circulation are robust, as supported independently by previous literature (Zhu et al. 2014, 2015; Ullman et al. 2014; Gong et al. 2015; Klockmann et al. 2016; Galbraith and de Lavergne 2018). However, there are also caveats in the present study. For example, the simulated modern AMOC (~ 11.5 Sv) is considerably weaker in our model compared with present observations (~ 18 Sv; e.g., Talley 2013). This underestimation is expected in the low-resolution model (Yeager et al. 2006; Shields et al. 2012; Jackson et al. 2020; Hirschi et al. 2020) and likely caused by the poor representation of eddies and the bias in ocean currents and properties (Danabasoglu et al. 2012; Jackson et al. 2020). The SH westerly in our deglacial simulation is slightly decreased ($\sim 15\%$) without significant shifting during

the last deglaciation (Fig. 8e), while neither models nor proxies agree on the magnitude, or even the sign, of the deglacial SH westerly change in both strength and latitude (Kohfeld et al. 2013; Sime et al. 2013). Also, tidal mixing due to the deglacial sea level change is not considered in TraCE21, which has been shown to affect the strength and depth of AMOC (Wilmes et al. 2019). These limitations add uncertainties in the interpretation of deglacial AMOC change. For example, given the considerable underestimation of the modern-day AMOC in our model, one might expect a ~ 1 Sv (rather than 0.5 Sv) stronger AMOC at the LGM in reality. Further coupled simulations with higher resolution and more realistic glacial boundary conditions are needed to better understand the oceanic circulation change in the past.

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Data availability statement. TraCE21 data used in this study are publicly available at <https://www.earthsystemgrid.org/project/trace.html>.

APPENDIX

Scaling for AMOC and Interbasin Exchange

So far in the present study, we take North Atlantic convective transport T_{NA} as a prescribed term since the scaling laws

for this transport are not well established as those for other processes. By assuming various balance associated the meridional flows at the North Atlantic (Gnanadesikan 1999; Johnson et al. 2007; Wolfe and Cessi 2010; Nikurashin and Vallis 2012), the scaling laws for T_{NA} follow a similar form as

$$T_{NA} \sim \frac{\Delta\rho H^2}{f_n}, \quad (A1)$$

where H denotes the scale depth defined as the interface depth in two-layer theoretical models and is commonly interpreted as either depth of pycnocline/AMOC maximum (Gnanadesikan 1999; Johnson et al. 2007; de Boer et al. 2010; Ferrari et al. 2017) or the depth of cell interface between the NADW cell and the AABW cell (Wolfe and Cessi 2010, 2011; Nikurashin and Vallis 2012); f_n is the Coriolis frequency at the latitude where convection occurs in North Atlantic. The definition of $\Delta\rho$ ranges from meridional density gradients in North Atlantic (Gnanadesikan 1999) and density difference between the thermocline and deep boxes (reduced gravity; Johnson et al. 2007) to the shared density range between the circumpolar channel and the isopycnal outcrop region in the North Atlantic (Wolfe and Cessi 2010; Nikurashin and Vallis 2012). Despite the diversity, T_{NA} is always proportional to $\Delta\rho H^2$.

As mentioned in section 3a, the interface depth between the upper and lower cells of the overturning circulation is deepening ($\sim +350$ m) under GHG forcing and shallowing (~ -100 m) under ice sheets forcing during the deglacial. We also calculate the deglacial change in depth of the AMOC maximum and find a shallowing (~ -100 m) under deglacial ice sheet forcing and deepening (~ 100 m) under deglacial GHG forcing. In response to GHG (ice sheet) forcing, $\Delta\rho$ is increased (decreased), accompanying the strengthening (weakening) of AMOC no matter which definition of $\Delta\rho$ is applied. While the signs of deglacial changes of $\Delta\rho$ are consistent among the three definitions, the magnitude for each depends heavily on the domain choice. For example, given the anomalous density distributions shown in Fig. 3, the meridional density gradients can be very different when different latitude ranges are considered. Model studies show that the depth and strength of the AMOC are influenced strongly by remote forcing such as Southern Ocean winds and AABW (de Boer et al. 2010). Besides, our analysis suggests that changes in Indo-Pacific circulation can also exert a strong impact on AMOC. Therefore, we note that AMOC in our simulations qualitatively follows the scaling law of Eq. (16), and a quantitative scaling investigation is beyond the scope of the present study.

Theoretical studies suggest that the transport of surface and thermocline water from the Indo-Pacific to the Atlantic is geostrophically balanced at the southern boundary (Jones and Cessi 2016; Cessi and Jones 2017; Ferrari et al. 2017), which requires a difference in the isopycnal depths along eastern boundaries of the two basins. In the mean climate in TraCE21, the depth of isopycnal within NADW upper limb along eastern boundary is systematically deeper (by ~ 200 m) in Indo-Pacific than that in Atlantic (not shown; also can be seen from the zonal mean depth in Fig. 2), consistent with observations (Fig. 20 of Wolfe and Cessi 2010), indicating an interbasin transport from Indo-Pacific to Atlantic. In two-layer and

two-basin models as proposed by Ferrari et al. (2017), the interbasin exchange T_G in equilibrium state can be scaled as

$$T_G \sim g' h_p \delta h, \quad (A2)$$

where g' denotes the density difference between the two layers, h_p is the depth of interface depth along the model's Pacific eastern boundary, and δh is the interface depth difference between model's Pacific eastern boundary (h_p) and Atlantic eastern boundary (h_a , $h_a \approx H$). Furthermore, the change ratio of h_p is much less than that of δh and thus can also be omitted. Under the above premises and considering changes between two climate states, we have

$$\Delta T_G \sim \Delta(g' \delta h). \quad (A3)$$

Under deglacial greenhouse gas (ice sheet) forcing, we observe a relatively shallowing (deepening) isopycnal in the eastern South Atlantic relative to the eastern South Pacific especially along $37.6\sigma_2$, corresponding to the stronger (weaker) interbasin exchange and NADW return flow (Fig. 10). Different from a simple theoretical model, there is strong dependence of g' and δh on the domain and isopycnal choices in CGCM (not shown).

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