A Diagnostic Indicator of the Stability of the Atlantic Meridional Overturning Circulation in CCSM3

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ABSTRACT

A diagnostic indicator $\Delta M_{ov}$ is proposed in this paper to monitor the stability of the Atlantic meridional overturning circulation (AMOC). The $\Delta M_{ov}$ is a diagnostic for a basinwide salt-advection feedback and defined as the difference between the freshwater transport induced by the AMOC across the southern border of the Atlantic Ocean and the overturning liquid freshwater transport from the Arctic Ocean to the North Atlantic. As validated in the Community Climate System Model, version 3 (CCSM3), for an AMOC in the conveyor state, a positive $\Delta M_{ov}$ (freshwater convergence) in the Atlantic basin indicates a monostable AMOC and a negative $\Delta M_{ov}$ (freshwater divergence) indicates a bistable AMOC. Based on $\Delta M_{ov}$, the authors investigate the AMOC stability in the Last Glacial Maximum (LGM) and analyze the modulation of the AMOC stability by an open/closed Bering Strait. Moreover, the authors estimate that the real AMOC is likely to be bistable in the present day, since some observations suggest a negative $\Delta M_{ov}$ (freshwater divergence) is currently in the Atlantic basin. However, this estimation is very sensitive to the choice of the observational data.

1. Introduction

The Atlantic meridional overturning circulation (AMOC) has been suggested to play a key role in rapid climate changes in the past (Clark et al. 2002; Rahmstorf 2002). The rapid change is associated with the nonlinear nature and, in turn, multiple equilibria of the AMOC. The multiple equilibria of the AMOC have been found in models of various complexity, including simple box models (Stommel 1961; Rooth 1982), ocean general circulation models (OGCMs; Bryan 1986; Marotzke et al. 1988; Marotzke and Willebrand 1991; Weaver and Sarachik 1991; Weaver and Hughes 1992; Weaver et al. 1993; Stocker and Wright 1991a,b; Power and Kleeman 1993; Zhang et al. 1993; Hughes and Weaver 1994), earth system models of intermediate complexity (EMICs; Rahmstorf et al. 2005; Hofmann and Rahmstorf 2009), and some coupled atmosphere–ocean general circulation models (AOGCMs; Manabe and Stouffer 1988; Yin and Stouffer 2007; Hawkins et al. 2011). It has, however, remained unclear if the bistable AMOC with the conveyor and collapsed states is a common characteristic of state-of-the-art AOGCMs. For example, Stouffer et al. (2006) dealt with the AMOC response to the freshwater forcing in AOGCMs of the Paleoclimate Modeling Intercomparison Project (PMIP) but found no evidence of multiple equilibria except for one AOGCM [Geophysical Fluid Dynamics Laboratory R30 (GFDL-R30); Yin and Stouffer 2007], because the AMOC restores to its original conveyor state after the termination of the freshwater hosing.

To better study the stability of the AMOC, especially in a complex climate model and in the real world, it is highly desirable to have a diagnostic indicator. In a box model, in which the Atlantic and Arctic Oceans are combined into a single box, Rahmstorf (1996) found that the meridional freshwater transport associated with the AMOC $F_{OT}$ across the southern boundary of the Atlantic basin can be used as a diagnostic indicator for AMOC stability, with a positive $F_{OT}$ (freshwater import) and negative $F_{OT}$ (freshwater export) indicating a monostable regime and a bistable regime, respectively. Weber et al. (2007) further applied this indicator to PMIP climate models and found that all models except for one
EMIC Climate deBilt-Coupled Large-Scale Ice Ocean (ECBilt/CLIO) show a freshwater import across the southern border of the Atlantic basin. This is consistent with the hosing experiments of Stouffer et al. (2006), which exhibit no evidence of bistability of the AMOC in these models.

Physically, the transport indicator can be understood as follows (Rahmstorf 1996): For an AMOC of negative $F_{OT}$ (freshwater exporting), a freshwater anomaly over the North Atlantic will reduce the AMOC and, in turn, the freshwater export. The latter will further lead to a basinwide freshening and in turn the suppression of deep convection in the North Atlantic and eventually the collapse of the AMOC. In this argument, it is clear that a better indicator should be the net convergence of the freshwater transport associated with the AMOC, with a net convergence for the monostable regime and a net divergence for the bistable regime. In general, the convergence indicator will not be the same as the transport indicator if the AMOC exchanges freshwater not only in the south with the Southern Ocean but also in the north with the Arctic. This motivates Dijkstra (2007) to modify the stability indicator as a convergence indicator $\Sigma$, which is associated with the convergence/divergence of the freshwater transport by the AMOC over the Atlantic basin. This convergence indicator has subsequently been shown valid in an OGCM coupled with an energy-balance atmosphere model (Huisman et al. 2010). However, to our knowledge, the convergence indicator has not been validated in any AOGCMs.

In this study, we propose an improved convergence indicator for the AMOC stability and for the first time validate it in a fully coupled AOGCM. As improved from $\Sigma$, our convergence indicator includes the meridional overturning freshwater transport across the Greenland, Iceland, and Norwegian (GIN) seas, since either in observation (e.g., Killworth 1983) or many AOGCMs (Schiller et al. 1997; Holland et al. 2001; Renold et al. 2010) the GIN seas are one of the major regions for the North Atlantic Deep Water (NADW) formation, and the freshwater transport via this region is very important in modulating the strength and stability of the AMOC (Holland et al. 2001; Curry and Mauritzen 2005; Komuro and Hasumi 2005; Oka and Hasumi 2006; Rennermalm et al. 2006, 2007). The remainder of the paper is structured as follows: Section 2 provides the information regarding the model and experimental design. In section 3, we propose and validate an improved AMOC stability indicator in the model. In section 4, we explore the usage of the indicator for other applications. Concluding remarks and further discussions are given in section 5.

2. Model and experiments

The AOGCM used in this study is the Community Climate System Model, version 3 (CCSM3), from the National Center for Atmospheric Research (NCAR). CCSM3 is a global, coupled ocean–atmosphere–sea ice–land surface climate model without flux adjustment (Collins et al. 2006). Unlike the EMICS, it includes a fully dynamic atmosphere model, the Community Atmosphere Model, version 2.0 (CAM2.0). All the simulations are performed in the version of T31_gx3v5 resolution (Yeager et al. 2006), which utilizes T31 spectral truncation (3.75° by 3.75° transform grid) for the atmosphere and adopts an x3ocn grid (100 × 116 points, nominally 3°) in the ocean and sea ice components, with 25 vertical levels in the ocean. Benefited from the x3ocn grid, the model resolution becomes significantly finer toward Greenland so that the model topography is well resolved in the Arctic and the North Atlantic. The Bering Strait is open, and the model resolution increases over the Canadian Archipelago so that it is possible to open a relatively realistic Northwest Passage between Baffin Bay and the Beaufort Sea.

The study is based on a control run of T31_gx3v5 in the perpetual AD 1990 scenario, which has been integrated for 1200 model years. By year 800, the model has reached a quasi equilibrium, except for some very slow adjustment associated with the abyssal water (Yeager et al. 2006). Thus, here, we start from year 780 (denoted here as new control year 0) and use the following 400-yr model integration as the control experiment (CTRL). To obtain a different state from CTRL for testing the AMOC stability indicator, we conduct an experiment
Dipole, the experiment with a dipole-like freshwater correction (DPOL) following the approach from De Vries and Weber (2005, hereafter, VW2005). Starting from year 100, a dipole-like freshwater correction is added into CTRL over the 17°–34°S belt in the South Atlantic gyre, with an anomalous freshwater flux of 20.25 Sv (1 Sv = 10^6 m^3 s^-1) added to the west of 15°W and an anomalous freshwater flux of 0.25 Sv added to the east of 15°W. The AMOC stability in the control and DPOL climates is tested by two parallel freshwater hosing experiments (CTRL-H and DPOL-H). A 100-yr pulse of 1.0-Sv freshwater flux is uniformly distributed into the North Atlantic between 50° and 70°N from year 100 in CTRL and from year 1100 in DPOL. The freshwater flux of 1.0 Sv is approximately equal to the total runoff over the world and is sufficiently large to shut down the AMOC in many AOGCMs (Stouffer et al. 2006). Here, it should be mentioned that a 100-yr transient hosing is just a common routine for testing the AMOC stability in AOGCMs, which may not enable the model to reach a different steady state. The integration of DPOL, CTRL-H, and DPOL-H lasts 1100, 900, and 1600 yr, respectively. The experimental designs are shown in Fig. 1.

3. Results

a. The AMOC stability indicator

First, we examine the validity of the indicators $F_{OT}$ and $\Sigma$ in the CCSM3 control run. In the Atlantic, the meridional freshwater transport to the south of 72°N can be divided into two parts: the meridional overturning part $M_{ov}$ associated with the AMOC and the azimuthally asymmetric part $M_{az}$ associated with the wind-driven gyre circulation. These two parts are defined as

$$M_{ov} = -\frac{1}{S_0} \int dz \overline{\nu}(z)[(\overline{\xi}) - S_0]$$

and

$$M_{az} = -\frac{1}{S_0} \int dz \overline{\nu}(z)\overline{s}(z).$$

Fig. 2. The meridional freshwater transport in the Atlantic from (a) CTRL and (b) DPOL with the azimuthal component $M_{az}$ (light gray; solid), the overturning component $M_{ov}$ (black; solid), and the total $M_{ov} + M_{az}$ (black; dotted). The $M_{ov}$ and $M_{az}$ are calculated from the monthly output by Eqs. (1) and (2) and shown as a 100-yr mean (years 1–100 in CTRL and years 1001–1100 in DPOL). Here, $\Sigma = M_{ov}(60°N) - M_{ov}(34°S)$ and is calculated as −0.134 Sv in CTRL and −0.256 Sv in DPOL, respectively. The zero line is drawn as a dashed line (light gray; long dashed). The southern (34°S) and northern (60°N) boundaries of $\Sigma$ are denoted as dotted lines (dark gray; dotted). In the figure, we limit $M_{ov}$ and $M_{az}$ to the south of around 72°N (the southernmost point of the western shelf of the Barents Sea) for strictly satisfying Eqs. (1) and (2).

Fig. 3. Time evolution of the AMOC strength in (a) CTRL (black) and CTRL-H (dark gray) and (b) DPOL (black) and DPOL-H (dark gray). The AMOC strength is defined as the maximum streamfunction value in the circulation below 500 m within the North Atlantic basin. It is calculated from the annual mean output and shown as a decadal average. The 100-yr hosing period is shaded in light gray.
where $S_0$ is a reference salinity from a global mean value of 34.7 psu; the overbar and the angle brackets $\langle \cdot \rangle$ denote zonal integration and zonal averaging along one latitude, respectively; and $\nu'$ and $s'$ are deviations from their zonal means. Here, $M_{az}$ and $M_{ov}$ are calculated from the monthly output of CCSM3. Figure 2a shows the distribution of $M_{ov}$, $M_{az}$, and their sum over the Atlantic basin in the equilibrium state of the CTRL. In the figure, the AMOC in CTRL generates a freshwater export across the southern boundary of the Atlantic basin: that is, $F_{OT} = M_{ov} = M_{ov}(34^°S) = -0.014$ Sv. At the same time, it induces a northward freshwater transport across 60°N, suggesting a freshwater divergence between 34°S and 60°N of $\Sigma = M_{ov}(34^°S) - M_{ov}(60^°N) = -0.134$ Sv. According to Rahmstorf (1996) and Dijkstra (2007), the negative $F_{OT}$ and $\Sigma$ indicate a bistable AMOC. However, as shown in Fig. 3a, the AMOC in CTRL is in a monostable regime because the circulation rapidly recovers after the removal of the freshwater forcing. Therefore, $F_{OT}$ and $\Sigma$ do not seem to be suitable for indicating the AMOC stability in CCSM3.

To find a desirable stability indicator for the AMOC in CCSM3, we diagnose the freshwater budget integrated over a generalized Atlantic basin, which is confined to

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34°S in the south and to about 80°N in the north. The northern boundary is to separate the Arctic Ocean in the north along the Canadian Arctic Archipelago (CAA), the Fram Strait, and the western shelf of the Barents Sea (Fig. 4). This extended northern boundary allows us to include all the major regions of deep-water formation within the generalized Atlantic, as indicated by the deep convection in the Labrador Sea and GIN seas. As shown in Fig. 4a, associated with deep convection in the late winter, the March mean mixed layer depth (MLD) in CTRL exceeds 1000 m in the Labrador Sea and reaches about 500 m in the GIN seas (Fig. 4a). Because of the deep convective mixing, the NADW forms in the Labrador Sea and GIN seas, making these regions the sinking sites of the AMOC. Therefore, this generalized Atlantic basin explicitly covers the AMOC pathway so that the freshwater transport across the basin is critical in regulating the AMOC stability.

The freshwater budget for the generalized Atlantic is calculated as in Eq. (3). In equilibrium, a net meridional overturning freshwater transport across the basin $\Delta M_{ov}$ is balanced by the basinwide net evaporation $[E_{net}]$, the azimuthal freshwater transport at the southern $M_{azS}$ and northern $M_{azN}$ boundaries as well as a residual term that includes the freshwater transport by diffusion: that is,

$$[E_{net}] = \Delta M_{ov} + M_{azS} - M_{azN} + \text{Res},$$

where $E_{net} = E - P - R - M + B_r$. Here, the sum of evaporation is $E$, precipitation is $P$, runoff is $R$, sea ice melting is $M$, brine rejection from sea ice melting is $B_r$, and $[\cdot]$ denotes a basinwide averaging.

The net meridional overturning freshwater transport $\Delta M_{ov}$ in Eq. (3) is defined as

$$\Delta M_{ov} = M_{ovS} - M_{ovN},$$

where $M_{ovS}$ and $M_{ovN}$ are the meridional overturning freshwater transports across the southern and northern boundaries of the Atlantic basin, respectively, and $M_{ovN}$ is equal to the overturning component of liquid freshwater import from the Arctic Ocean. Previous studies (e.g., Aagaard and Carmack 1989; Serreze et al. 2006; Holland et al. 2007; Jahn et al. 2010) show that the liquid Arctic freshwater enters the North Atlantic through three sections: the CAA, the Fram Strait, and the western shelf of the Barents Sea. Thus,

$$M_{ovN} = M_{ovCAA} + M_{ovFRA} + M_{ovBAR} \quad \text{and (5)}$$

$$M_{azN} = M_{azCAA} + M_{azFRA} + M_{azBAR}, \quad \text{(6)}$$

where $M_{ovCAA}, M_{ovFRA},$ and $M_{ovBAR}$ ($M_{azCAA}, M_{azFRA},$ and $M_{azBAR}$) are the overturning (azimuthal) liquid freshwater transports across the CAA, the Fram Strait, and the western shelf of the Barents Sea, respectively. Calculation of $M_{ovCAA}, M_{ovFRA}, M_{ovBAR}, M_{azCAA}, M_{azFRA},$ and $M_{azBAR}$ generally follows Eqs. (1) and (2), in which $\bar{v}$ and $\bar{u}$ are normal to the section while integration and averaging are along the direction of section (Fig. 4). It is worth mentioning that, besides $M_{ovN}$, the azimuthal component $M_{azN}$ also contributes to the Arctic freshwater sinks into the Atlantic, but with a much smaller partition (Table 1). For example, at the Fram Strait, the northward West Spitsbergen Current carries a warm and saline water into the Arctic Ocean while the southward East Greenland Current induces fresh Arctic water to the GIN seas in the surface layer (e.g., Aagaard and Carmack 1989; Serreze et al. 2006), so that this azimuthally asymmetric pattern leads to a liquid freshwater transport from the Arctic to the North Atlantic. Also, besides the liquid Arctic freshwater, a substantial amount of ice enters the North Atlantic via the Fram Strait, which has an input of freshwater in the North Atlantic and modulates the sea ice melting and the sea ice formation in the North Atlantic. This effect has been included in the surface flux $[E_{net}]$ by terms $M$ and $B_r$. We propose $\Delta M_{ov}$ as an improved convergence indicator of the AMOC stability for CCSM3 because it contains the full mechanism of the basin-scale salt-advection feedback (Stommel 1961) in this AOGCM. Consider the AMOC in its conveyor state; a freshwater discharge will weaken the AMOC. If $\Delta M_{ov}$ is positive (freshwater...
convergence), the initial weakening of the AMOC will reduce the freshwater convergence and therefore salinify the ocean, which then promotes deep mixing and in turn prevents a further weakening of the AMOC. Alternatively, the initial weakening of the AMOC will be amplified if $\Delta M_{ov}$ is negative. This is because the initial weakening of the AMOC reduces the divergence of the freshwater and therefore promotes a freshwater accumulation in the basin, which tends to suppress deep convection and therefore further amplify the initial weakening of the AMOC.

**b. Validation of the indicator**

To test the convergence indicator $\Delta M_{ov}$, we first conduct a DPOL experiment following VW2005 (see section 2). In the experiment, a dipole-like freshwater correction applies over the South Atlantic. This freshwater flux correction increases the longitudinal salinity contrast at the southern border of the Atlantic basin. The Brazil Current transports much saltier water southward, and the Benguela Current transports much fresher water northward (Fig. 5), resulting in an increase of the azimuthal freshwater transport $M_{azS}$ (Figs. 6b, 7d). At the same time, the Brazil Current has an eastward branch around 34°S, carrying much saltier surface and thermocline waters away from the west of the southern American continent (Fig. 5). As a result, the zonal mean salinity is largely enhanced (Fig. 6a). The “excess” salt flows out via the strong southward Brazil Current, which results in an enhanced freshwater export in the upper limb of the AMOC. Meanwhile, the deep southward-flowing NADW is mildly freshened, which leads to a small reduction in the freshwater export by the lower limb of the AMOC. With the combination of these two limbs, the AMOC produces a much stronger freshwater export $M_{ovS}$ in DPOL than in CTRL (Fig. 6c).

Figure 7 displays the evolution of the AMOC and its associated freshwater transport, as well as the components in a basinwide freshwater budget during the adjustment from CTRL to DPOL. The azimuthal component $M_{azS}$ and net evaporation $[E_{net}]$ are directly related to the rapid adjustments in the atmosphere and upper ocean as well as the interfacial atmosphere–ocean coupling, so that they enhance soon after the input of the freshwater correction and then keep steady within the whole integration period. The overturning component $M_{ovS}$ is associated with the relatively slow adjustment of the AMOC in which the circulation strength has a reduction of 4 Sv in the first 300 yr and then gradually recovers back to 13.5 Sv by year 1200. The input freshwater flux correction is confined within the belt of 17°–34°S. Therefore, it mainly modifies the salinity structure to the south of 10°N via the wind-driven gyre circulation in the southern Atlantic and equatorial regions (Fig. 8g) and therefore changes the overturning freshwater transports in the north ($M_{ovCAA}$, $M_{ovFRA}$, and $M_{ovBAR}$) only slightly. In contrast, the freshwater export $M_{ovS}$ increases significantly after the freshwater flux correction. As a result, the freshwater export $M_{ovS}$ exceeds the freshwater import $M_{ovN}$ after about year 500 (Fig. 7c), generating a freshwater divergence ($\Delta M_{ov} < 0$) in the Atlantic basin.

Table 1 summarizes the Atlantic freshwater budget, associated freshwater transports, and the AMOC strength...
in the equilibrium state of four experiments. The AMOC strength here is defined as the maximum in the streamfunction of the circulation below 500 m in the North Atlantic basin. From the table, a dipole-like freshwater flux correction slightly changes the AMOC strength (from 15.0 to 13.5 Sv) but significantly modulates the Atlantic freshwater budget. The azimuthal freshwater transport $M_{azS}$ is greatly enhanced to balance most of the net evaporation $E_{net}$ in the basin. More importantly, the net meridional overturning freshwater transport $M_{ov}$ switches from a convergence of 0.112 Sv in CTRL to a divergence of $-0.038$ Sv in DPOL, which indicates a transition of the AMOC stability from a monostable regime in CTRL to a bistable regime in DPOL.

We test the diagnostic indicator $\Delta M_{ov}$ by means of the two parallel freshwater hosing experiments (CTRL-H and DPOL-H). From the steady state in CTRL (DPOL), we impose a strong pulse of freshwater perturbation to test the AMOC stability in this state. As shown in Fig. 3, the AMOCs in CTRL-H and DPOL-H are shut down during the 100-yr hosing period. However, the cease of the AMOC in CTRL-H causes a net salinifying effect in the Atlantic basin. The excess salinity gets evolved in restarting the deep convection in the Labrador Sea and GIN seas (Figs. 9a,c, 4c). As a result, the AMOC rapidly recovers 600 yr after the termination of the freshwater perturbation (Fig. 3). The experiment CTRL-H is therefore consistent with a freshwater divergence ($\Delta M_{ov} < 0$) for a monostable AMOC.

On the other hand, the collapse of the AMOC in DPOL-H induces a net freshening effect in the Atlantic basin. The excessive freshwater accumulates in the upper 400 m in the North Atlantic (Fig. 8d), especially between 40° and 80°N (Fig. 10f), to stabilize the stratification in the Labrador Sea and GIN seas. As a result, deep convection is inhibited in the GIN seas, with a mean MLD shallower than 100 m in the late winter (Fig. 9b). Meanwhile, deep convection is also greatly suppressed in the Labrador Sea, with a March mean MLD reducing from 900 m in the prehosing period to 300 m in the posthosing period (Fig. 9b). The remaining deep convective mixing in the Labrador Sea can only induce a very limited NADW formation (Fig. 8d). As a result, the AMOC, instead of recovering, resides in a weak stable state after the termination of the freshwater perturbation. The experiment DPOL-H proves that a freshwater divergence ($\Delta M_{ov} > 0$) indicates a bistable AMOC.

4. Implications

a. The stability of real-world AMOC at the present day

An important usage of the AMOC stability indicator is to estimate the stability of the AMOC in the real world from the observations. Previous studies (Huisman et al. 2010; Hawkins et al. 2011) that are based on $F_{OT}$ suggest a potential existence of a bistable AMOC in the present day since most observations and reanalysis data (Weijer et al. 1999; Huisman et al. 2010; Hawkins et al. 2011) consistently support a negative $F_{OT}$, in the form of $M_{ovS}$ induced by the AMOC across the southern border of the Atlantic basin.

In this study, we estimate the present-day AMOC stability based on the indicator $\Delta M_{ov}$. From the observations,
Weijer et al. (1999) find that the AMOC is associated with a 0.2-Sv freshwater export across the southern border of the Atlantic: that is, $M_{ovS} = 0.2$ Sv. In the meantime, Serreze et al. (2006) report that, associated with the present-day AMOC, the liquid Arctic freshwater enters the North Atlantic with $2660$ km$^3$ yr$^{-1}$ via the Fram Strait, $110$ km$^3$ yr$^{-1}$ via the Barents Sea, and $3200$ km$^3$ yr$^{-1}$ via the CAA, which gives a total Arctic freshwater transport of $5970$ km$^3$ yr$^{-1}$ (0.189 Sv) into the North Atlantic. As discussed in section 3, the total liquid Arctic freshwater transport is accomplished by its azimuthal and overturning components so that $M_{ovN} > -0.189$ Sv. Therefore, the present-day AMOC is estimated to generate a freshwater divergence ($\Delta M_{ov} < 0$) across the Atlantic basin, which indicates that the circulation is in a bistable regime, consistent with the transport indicator $F_{OT}$.

However, this estimation has a great uncertainty. First, $M_{ovS}$ and $M_{ovN}$ adopted in the estimation are based on different reference salinities ($M_{ovS}$ based on 35.08 psu and $M_{ovN}$ based on 34.8 psu). The $\Delta M_{ov}$ is prone to be modified if a uniform reference salinity is applied in the calculation of $M_{ovS}$ and $M_{ovN}$. Second, $M_{ovS}$ (and, in turn, $\Delta M_{ov}$) has a great uncertainty as it is calculated from different observational data. For example, Huisman et al. (2010) propose a 0.1-Sv freshwater export of the AMOC across the southern boundary ($M_{ovS} = 0.1$ Sv) by using different observational data from Weijer et al. (1999). Hawkins et al. (2011) report that $M_{ovS}$ is generally negative in multiple ocean re-analysis data, but with large variations in magnitude, ranging from close to zero to over $-2$ Sv. Therefore, based on these studies, the net freshwater transport may be either convergent or divergent in the Atlantic basin so that the present-day AMOC is either monostable or bistable.

b. $\Delta M_{ov}$ and the AMOC stability in the LGM

Besides the present-day scenario, the indicator $\Delta M_{ov}$ can also correctly indicate the AMOC stability in the Last Glacial Maximum (LGM) scenario. From a recent simulation of transient climate evolution of the last 21.0 thousand years [ka; Transient Climate Evolution of the
Last 21,000 years (TraCE-21000); Liu et al. 2009; He 2011], we calculate that the global mean salinity equals 36.5 psu during the LGM (19.0 ka Before Present). Based on $S_0 = 36.5$ psu, we find a freshwater convergence of $0.202$ Sv across the Atlantic basin ($\Delta M_{ov} = 0.202$ Sv). This positive $\Delta M_{ov}$ indicates a monostable LGM AMOC, which has been demonstrated in Liu et al. (2009) and He (2011).

c. Role of the Bering Strait

The opening of the Bering Strait plays an important role in regulating the AMOC stability (De Boer and Nof 2004a,b), which can be explained by $\Delta M_{ov}$ since it connects the Arctic freshwater budget with the AMOC stability. In the present-day scenario, a closed Bering Strait cuts off the import of low-salinity water from the North.
Pacific, which modulates the freshwater budget over the Arctic Ocean (Hu and Meehl 2005; Hu et al. 2007) and leads to a decreasing liquid Arctic freshwater export (Hu et al. 2008). Correspondingly, the magnitude of $M_{\text{ovN}}$ is reduced. Therefore, in terms of $M_{\text{ovN}}$, a reduction in $M_{\text{ovN}}$ caused by a closed Bering Strait contributes an analogous divergence to $\Delta M_{\text{ov}}$, which tends to make the AMOC shift from a monostable regime toward a bistable regime.

5. Conclusions and discussion

In this paper, we propose an improved AMOC stability indicator $\Delta M_{\text{ov}}$, which is defined as the difference between the freshwater transport $M_{\text{ovS}}$ induced by the AMOC across the southern border of the Atlantic and the overturning liquid freshwater transport $M_{\text{ovN}}$ from the Arctic to the North Atlantic. The $\Delta M_{\text{ov}}$ is a diagnostic for the basinwide salt-advection feedback. Compared with previous indicators, $\Delta M_{\text{ov}}$ accurately indicates the AMOC stability in CCSM3, either for a monostable regime (CTRL) or a bistable regime (DPOL). In CTRL, a freshwater convergence ($\Delta M_{\text{ov}} > 0$) is induced by the AMOC in the Atlantic basin, which indicates a monostable circulation. In DPOL, the AMOC generates a freshwater divergence ($\Delta M_{\text{ov}} < 0$) in the Atlantic, and the circulation is within a bistable regime.

This improved indicator $\Delta M_{\text{ov}}$ can also be applied to diagnose the AMOC stability in other studies. In TraCE-21000, $\Delta M_{\text{ov}}$ correctly indicates a monostable AMOC in the LGM. Besides, in the present-day scenario, $\Delta M_{\text{ov}}$ tends to be more divergent with a closed Bering Strait, which induces a shift of the AMOC stability toward a bistable regime. It is worth mentioning that the opening of the Bering Strait is one of the essential deglacial processes in the paleoclimate studies (e.g., De Boer and Nof 2004a,b; He 2011); a thorough investigation of $\Delta M_{\text{ov}}$ will shed light on the evolution of the AMOC stability during the last deglaciation.

More importantly, we estimate the AMOC stability for the real world based on observations. The AMOC is likely to be bistable in the present day, as indicated by a freshwater divergence ($\Delta M_{\text{ov}} < 0$) in the Atlantic basin. However, this estimation is very sensitive to the choice of the observational data. Compared with other indicators, $\Delta M_{\text{ov}}$ includes the contribution from $M_{\text{ovN}}$, which make the estimation of a bistable present-day AMOC less robust.

In the following, we would like to discuss several associated issues. First, unlike the observation-based estimation, most state-of-the-art AOGCMs can only simulate a monostable AMOC in a preindustrial scenario. As we speculate, this is because $M_{\text{ovS}}$ is generally negative (Weber et al. 2007; Drijfhout et al. 2010) and in turn $\Delta M_{\text{ov}}$.
is positive (a freshwater convergence) in these models. Compared with observations, most AOGCMs simulate much fresher surface and thermocline waters but a slightly saltier NADW around 34°S, which leads to a freshwater import ($M_{ovS} > 0$, as in all the AOGCMs from PMIP) or a very weak freshwater export ($M_{ovS} < 0$, as in CTRL from this study) induced by the AMOC across the southern border of the Atlantic. Meanwhile, a strong freshwater import $M_{ovN}$ comes from the Arctic. Therefore, a freshwater convergence is generated within the Atlantic basin by the cooperation between $M_{ovN}$ and $M_{ovS}$ or a dominant contribution from $M_{ovN}$, which results in a monostable AMOC in these AOGCMs. In our future work, we will focus on correcting the $\Delta M_{ov}$ in CCSM3 toward an observational value, for a realistic simulation of the AMOC in the present day.

Second, it is worth mentioning that $\Delta M_{ov}$ indeed indicates a basin-scale freshwater feedback associated with the NADW cell so that it may not be valid when the Antarctic Intermediate Water (AAIW) reverse cell has a strong effect on the Atlantic freshwater budget. As shown in section 3, $\Delta M_{ov}$ is valid in this study because the collapsed state of the AMOC behaves as a very weak NADW cell, and the bistability of the AMOC is primarily determined by a basin-scale salinity advection related to the NADW cell (Fig. 8). On the other hand, several studies (e.g., Saenko et al. 2003; Gregory et al. 2003; Sijp and England 2006; Sijp et al. 2012) suggested that, for a bistable AMOC, the collapsed circulation appears as an AAIW reverse cell, and the nonlinear behavior of the AAIW reverse cell plays a critical role in suppressing the NADW formation and maintaining the collapsed state (Sijp et al. 2012). Therefore, we speculate that the indicator $\Delta M_{ov}$ may not be applicable to the AMOC in these studies.

Third, we should point out that we utilize a constant salinity $S_0$ rather than the 34°S section-average salinity
\[ S_m \left[ S_m = \int_{34^\circ S} dz(3) / \int_{34^\circ S} dz(1) \right] \] as the reference salinity in the calculation of \( M_{ov} \). This is because the usage of \( S_m \) can only guarantee a correct removal of the barotropic contribution in \( M_{ov} \) at the southern boundary of the Atlantic (Drijfhout et al. 2010) but not at the northern boundary as well. Hence, it is unsuitable to use \( S_m \) for calculating the convergence indicator \( \Delta M_{ov} \) in our study. On the other hand, by choosing a constant \( S_0 \), which is close to the basin-averaged salinity over the Atlantic Ocean, we can make sure that the barotropic transports across the southern and northern boundaries of the Atlantic mostly cancel with each other. As a result, \( \Delta M_{ov} \) has little contribution from the barotropic transport, which enables it to correctly represent the net meridional overturning freshwater transport across the Atlantic basin.

Actually, the barotropic transport induced by \( S_0 \) does not change our conclusion in section 3a. At 34°S, \( \delta S = 34.767 \) psu in CTRL and \( \delta S = 34.883 \) psu in DPOL, which are both very close to \( S_0 \). Based on \( S_m \), we calculate that \( M_{ovS} = -0.011 \) Sv in CTRL and \( M_{ovS} = -0.168 \) Sv in DPOL, which suggests that the barotropic transport induced by \( S_0 \) only slightly modifies \( M_{ovS} \) in magnitude but does not change the sign of \( M_{ovS} \) (Table 1). As a result, the transport indicator is still negative, incorrectly indicating the AMOC in CTRL.

Finally, we must clarify that the hosing experiment adopted in this study is a commonly used and perhaps, a currently applicable way to study the AMOC stability in state-of-art AOGCMs. Unlike simple models (Stommel 1961; Rooth 1982; Tziperman et al. 1994; Rahmstorf 1996), the complexity of AOGCMs does not allow us to calculate the analytical solutions of the AMOC and obtain all the unstable modes (multiple equilibria) of the circulation by means of stability analysis. Thus, for AOGCMs, a practical way to find the unstable modes of the AMOC is to add a perturbation and see which state the model will reside in. Among various perturbations, we choose one that is more physically plausible: the freshwater perturbation in the high latitude of the North Atlantic (also named the hosing experiment) to test the AMOC stability. This added freshwater perturbation follows the physical concept of Rahmstorf (1996), and helps us to get a seeming “optimal” unstable mode of the AMOC in the complex AOGCMs. The resulting unstable mode of the AMOC is associated with a basin-scale salinity-advection feedback (Stommel 1961), which is different from the convective instability in Lenderink and Haarsma (1994).

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