Assessing the Stability of the Atlantic Meridional Overturning Circulation of the Past, Present, and Future

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ABSTRACT

This paper is a review of the recent development of researches on the stability of the Atlantic meridional overturning circulation (AMOC). In particular, we will review recent studies that attempt to best assess the stability of the AMOC in the past, present, and future by using a stability indicator related to the freshwater transport by the AMOC. These studies further illustrate a potentially systematic bias in the state-of-the-art atmosphere-ocean general circulation models (AOGCMs), in which the AMOCs seem to be over-stabilized relative to that in the real world. This common model bias in the AMOC stability is contributed, partly, to a common tropical bias associated with the double intertropical convergence zone (ITCZ) in most state-of-the-art AOGCMs, casting doubts on future projection of abrupt climate changes in these climate models.

Key words: AMOC, stability indicator, freshwater transports, ITCZ, deglaciation

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1. Introduction

The Atlantic meridional overturning circulation (AMOC) consists of a lower limb of denser, colder southward return flow at depth, and an upper limb of returning flow of near-surface warm waters (Fig. 1). Presently, it is estimated that the AMOC has a mean magnitude of 18 $\pm$ 1.0 Sv (1 Sv = 10^6 m^3 s^{-1}), based on direct observations at 26.5°N by the RAPID/MOCHA array between April 2004 and April 2009 (Cunningham et al., 2007; Kanzow et al., 2007; Johns et al., 2011). The AMOC affects the climate by its significant northward oceanic heat transport, about 1.3 PW (1 PW = 10^{15} W) over 24°–26°N (Hall and Bryden, 1982; Trenberth and Caron, 2001; Johns et al., 2011). A collapse of the AMOC would induce significant climate impacts, such as a bipolar see-saw response with a pronounced cooling over the North Atlantic and warming in the Southern Hemisphere (Drijfhout et al., 2011; Shakun et al., 2012), the shift of the intertropical convergence zone (ITCZ) as well as the Atlantic rainfall (Vellinga and Wood, 2002), the modulation of North Atlantic sink for CO_2 (Schuster and Watson, 2007), and variations in marine ecosystems (Schmittner, 2005). Generally, a collapse of the AMOC has often been associated with the multiple equilibria of the AMOC, and has been used to indicate the observed abrupt changes in past climates (e.g., Clark et al., 2002). In the Holocene (around last 10000 years), the absence of rapid climate changes seems to suggest that the AMOC has remained in a mono-stable regime (Grootes et al., 1993). Nevertheless, the future climate response to global warming may induce an intensified hydrological cycle, which may destabilize the AMOC (Broecker et al., 1985; Rahmstorf, 2002) and cause potential abrupt climate changes. Therefore, it is important to investigate the AMOC stability under various climates. Herein, one issue arises, i.e., for the real climate system of the present, the stability of the AMOC cannot be assessed with sensitivity experi-

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Fig. 1. A simplified schematic of the AMOC showing both the overturning and gyre recirculation components. Warm water flows northward in the upper ocean (red), gives up heat to the atmosphere (atmospheric flow gaining heat represented by the changing color of broad arrows), sinks, and returns as a deep cold flow (blue). Latitude of the 26.5°N AMOC observations is indicated. Note that the actual flow is more complex. For example, see Bower et al. (2009; their Fig. 1) for the intermediate depth circulation in the vicinity of the Grand Banks and Biastoch et al. (2008; their Fig. 2) for the mid-depth circulation around South Africa, showing the importance of eddies in transferring heat and salt from the Indian Ocean to the Atlantic Ocean. [From Srokosz et al., 2012]

ments, but such a diagnostic indicator is highly needed. Besides the assessment, such an indicator also enables a stability intercomparison among climate models, which then can be used for selecting some validated models to predict the future changes of the AMOC. In this study, we will review the studies so far of the AMOC stability in the past, present, and future climates.

The rest of the paper is structured as follows. Section 2 describes some insights on the AMOC multiple equilibria from the paleoclimate perspective. From these insights, substantial theoretical and modeling studies of the AMOC stability are reviewed in Section 3, especially on the topic of the AMOC stability indicator. Then, based on the indicator, the stability of the AMOC under the present and future climates is assessed in Sections 4 and 5, respectively. In particular, a critical issue is posed that, due to a common double ITCZ problem, most state-of-the-art atmosphere-ocean general circulation models (AOGCMs) may exhibit a systematic bias in simulating the AMOC stability. This may distort the estimation of current AMOC stability or prediction of the AMOC behavior in future climate. Concluding remarks and further discussion are given in Section 6.

2. The AMOC stability in past climates

Past climate records have shown clear evidence of various types of abrupt climate changes on millennial timescales, notably the Dansgaard/Oeschger (D/O) cycles (Dansgaard et al., 1993) and Heinrich events (see reviews of Rahmstorf, 2002; Clark et al., 2002, 2007). These cycles are characterized by opposite temperature responses in the Northern and Southern Hemispheres in a so-called bipolar "see-saw" response (e.g., Stocker and Johnsen, 2003), as indicated in ice cores in Greenland and Antarctic. These abrupt changes have been speculated to link with abrupt changes in the AMOC and its associated heat transport (e.g., Ganopolski and Rahmstorf, 2001; Liu et al., 2009). In one hypothesis, the abrupt D/O events are proposed to be caused by the multiple equilibria or bi-stability of the AMOC, i.e., the switch of interstadial and stadial modes in the D/O cycles is induced by transitions between different equilibrium states of the AMOC (Broecker et al., 1985). This point of view is consistent with the reconstruction of the North Atlantic Deep Water (NADW) production. As shown in Sarnthein et al. (1994), the NADW production is reduced from the interstadial mode to the stadial mode in a D/O event, or interrupted in a Heinrich event, which represents a weaker or collapsed AMOC herein (Fig. 2).

Despite a strong body of evidence that associates the AMOC with the abrupt changes (e.g., Broecker et al., 1985; Clark et al., 2002; McManus et al., 2004), the trigger mechanism of these abrupt changes of the AMOC has remained yet unclear. One candidate is the meltwater pulse. For instance, one major meltwater pulse, the Meltwater Pulse 1A (MWP-1A), as

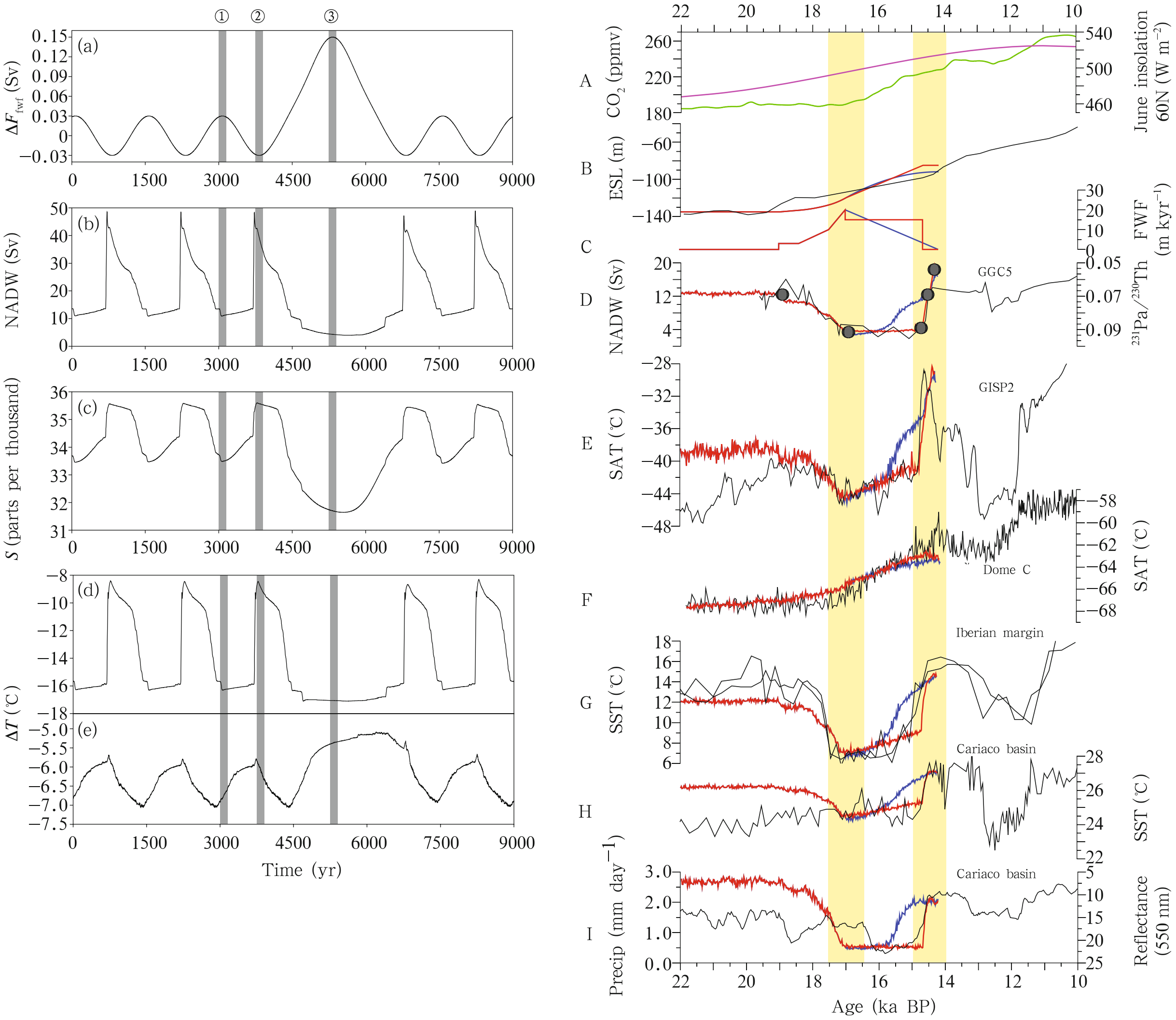

Fig. 2. (Left panel) Simulated D/O and Heinrich events. (a) Forcing, (b) Atlantic overturning, (c) Atlantic salinity (S) at 60°N, (d) air temperature in the northern North Atlantic sector (60°–70°N), and (e) temperature over Antarctica (temperature values are given as the difference from the present-day climate, ΔT) (from Ganopolski and Rahmstorf, 2001). (Right panel) Data-model comparison for several benchmark time series. (A) June insolation at 60°N (purple) (Berger, 1978) and atmospheric CO_2 concentration (green) (Joos and Spahni, 2008), ppmv means parts per million by volume. (B) Sea level from the reconstruction (gray) (Peltier, 2004) and model (meters of equivalent global sea level (ESL) for meltwater). (C) Freshwater fluxes (FWF) in the model. (D) Pa/Th ratio at Bermuda (GGC5 core) as a proxy for the AMOC strength (McManus et al., 2004), and model maximum AMOC transport (below 500 m). (E) Greenland surface air temperature (SAT) based on Greenland Ice Sheet Project 2 (GISP2) δ^{18}O reconstruction with borehole temperature calibration (Cuffey and Clow, 1997) and in the model (model offset by –3℃). (F) Antarctic surface air temperature based on Dome C δ^{18}O reconstruction (Jouzel et al., 2007) and in the model. (G) Sea surface temperature (SST) from the Iberian Margin from reconstructions (Waelbroeck et al., 1998; Bard et al., 2000) and model. (H) SST from the Cariaco Basin from reconstruction (Lea et al., 2003) and model (model offset by 4℃). (I) Rainfall in Cariaco Basin from reconstruction (Peterson et al., 2000) and model. In (B) to (I), gray is used for the reconstruction, and red and blue for experiments DGL-A and DGL-B, respectively. The five circles on DGL-A in (D) represent the glacial state (GLA; 19 ka), H1 (17 ka), PreBA (14.7 ka), Recovery (REC; 14.5 ka), and BA (14.35 ka). All model variables are annual means with a 20-yr running average. Overall, model simulations, especially DGL-A, are in good agreement with the proxy records, especially outside the tropical Atlantic. BP means before present. [From Liu et al., 2009]

regarded, triggers the transition from the Heinrich event 1 (H1) to Bølling-Allerød (BA). However, this candidate is challenged by uncertainties in the lead-lag relationship between the meltwater pulse and the AMOC change. Due to a poor chronology of the reconstructions, the meltwater history and even the location of meltwater pulses (such as MWP-1A) have remained controversial (Clark et al., 1996; Peltier, 2005; Stanford et al., 2006; Deschamps et al., 2012). As a result, it is difficult to establish a precise chronological order between the meltwater pulses and the AMOC changes. If an abrupt AMOC change does not follow a meltwater discharge within a self-adjustment scale (several hundred years), the AMOC change can be viewed as a response of a bi-stable AMOC to a smooth change of freshwater forcing (e.g., Ganopolski and Rahmstorf, 2001; Knorr and Lohmann, 2003; Weaver et al., 2003). Otherwise, the AMOC change is likely forced by the abrupt change of meltwater forcing due to the instability of ice sheet (e.g., MacAyeal, 1993), instead of related to the mono-stable AMOC itself (e.g., Liu et al., 2009). Because current observations are not sufficient to distinguish the lead-lag time between abrupt changes in the AMOC and corresponding meltwater pulses unambiguously, many abrupt climate events are still considered most likely to be caused by a bistable AMOC. This suggests that the instability of the AMOC could play a key role in past rapid climate transitions. Meanwhile, such changes in the AMOC are of great importance to either present or future climates, considering that the potential for an AMOC collapse is a key uncertainty in future climate projections. Thereby we will continue to review the AMOC multiple equilibria as well as its stability indicator.

3. The AMOC multiple equilibria and stability indictor

The nonlinear nature of the AMOC results in the existence of its multiple equilibria. One pioneering study was from Stommel (1961) who used a simple two-box model to propose a positive salinity advection feedback in destabilizing the AMOC and leading to a bi-stable circulation. The Stomnel box model was further examined (Mu et al., 2004) and extended to an inter-hemisphere box model (e.g., Rooth, 1982; Welander, 1986; Rahmstorf, 1996, see Fig. 3). Later on, the existence of the AMOC multiple equilibria has been demonstrated in ocean general circulation models (OGCMs), ranging from a three dimensional model of a single, flat basin (Bryan, 1986) in which a concept of "halocline catastrophe" was proposed, to zonally averaged global models (e.g., Marotzke et al., 1988; Stocker and Wright, 1991a, b; Hughes and Weaver, 1994) and global OGCMs (e.g., Marotzke and Willebrand, 1991; Power and Kleeman, 1993; Weaver et al., 1993). Furthermore, the AMOC multiple equilibria were explored in two types of more advanced models, the earth system models of intermediate complexity (EMICs) and atmosphere-ocean general circulation models (AOGCMs). Rahmstorf et al. (2005) compared 11 EMICs and found that all the models simulate AMOCs with significant multiple equilibria (Fig. 4). The bi-stability of the AMOC in EMICs was further demonstrated in the limit of low mixing (Hofmann and Rahmstorf, 2009). In contrast to the robust bi-stability of EMICs, there has been little evidence of the AMOC multiple equilibria in state-of-the-art AOGCMs. For example, in AOGCMs from Coupled Model Intercomparison Project phase 3 (CMIP3) (Stouffer et al., 2006), most AMOCs exhibit monostable behaviors. The circulations recover to their original conveyor states after a termination of freshwater perturbation (Fig. 5)①.

There is so far no consistent explanation why the AMOC tends to be bi-stable in intermediate models, but not in AOGCMs. The systematic lack of multiple equilibria in AOGCMs, however, seems to indicate that certain factors in common are inclined to overstabilize the AMOC in these models. Great endeavors have been put in seeking such factors in several aspects. From the ocean aspect, some studies suggested that increasing oceanic diapycnal diffusivity can inhibit the AMOC multiple equilibria by generating

① So far, two exceptions can be found among current AOGCMs: the GFDL R30 and FAMOUS models; see Manabe and Stouffer (1988) and Hawkins et al. (2011), respectively.

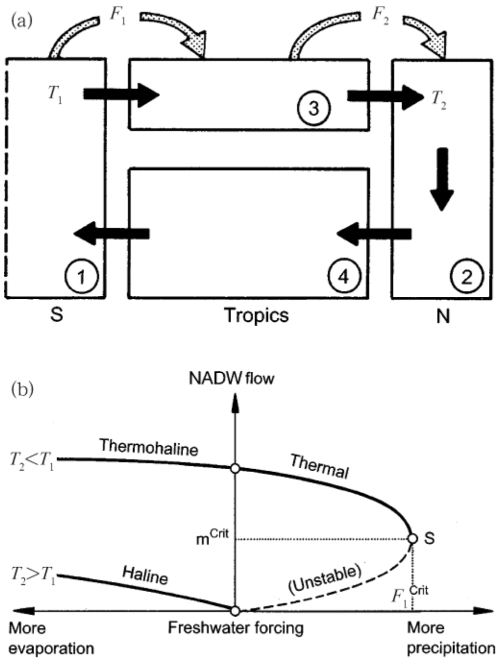

Fig. 3. (a) A simple 4-box model of cross-hemispheric thermohaline flow. NADW forms in box 2; its outflow towards box 1 is controlled by the density difference between boxes 2 and 1. Salinities in the boxes are determined by the flow and the surface freshwater fluxes entering boxes 1, 2, and 3. Only two of these three fluxes are independent, since their sum must vanish in a steady state. Therefore, the surface freshwater fluxes are portrayed as two atmospheric vapor transports F_1 and F_2. (b) The three flow regimes (solid) of the box model. The dashed line is an unconditionally unstable solution, and S is the saddle-node bifurcation point. [From Rahmstorf, 1996]

a more diffusive and linear circulation (Manabe and Stouffer, 1999), whereas others proposed that strong diffusivity would enhance the multiple equilibria via a strong oceanic upwelling (Prange et al., 2003; Nof et al., 2007). From the atmosphere aspect, strong internal atmospheric variability was found to set up a stochastic forcing in generating the bimodality of the AMOC (e.g., Cessi, 1994; Timmermann et al., 2003), which was further supported by results from several intermediate models. The wind stress feedback was argued to stabilize the AMOC in modern climate (Mikolajewicz, 1996; Schiller et al., 1997) but to destabilize the AMOC under glacial climate (Arzel et al., 2008). From the coupling aspect, the AMOC multiple equilibria were suggested to be suppressed by a strong ocean-atmosphere coupling and associated precipitation response (Yin et al., 2006; Yin and Stouffer, 2007). In brief, great divergences exist in the above model-based arguments, which provide a strong motivation to assess the AMOC stability against the real world.

To evaluate the stability of the AMOC offline in a complex climate model, and more importantly in the real world, one has to formulate a diagnostic indicator of the AMOC stability. By using a freshwater budget (discussed later) and treating the Atlantic and Arctic basin as a united "box" (Fig. 3a), Rahmstorf (1996) first proposed the freshwater export by the AMOC across the southern boundary (approximately 34°S) of the Atlantic, abbreviated as $F_{\rm OT}$, as a diagnostic indicator of the AMOC stability. This indicator can be used to evaluate the AMOC stability because it captures the salinity-advection positive feedback (Stommel, 1961) that is critical for the bistability of the AMOC. Physically, consider an AMOC state with a freshwater export. An initial pulse of perturbation freshwater flux in the North Atlantic weakens the AMOC and therefore reduces the export of freshwater. This will lead to an accumulation of freshwater that further enhances the initial pulse of freshwater perturbation, and eventually result in a collapse of the AMOC. In practice, the freshwater transport by the AMOC is calculated as the freshwater transport by the zonal mean overturning circulation.

This indicator was later adopted by Weber et al. (2007), who found that all the models in the Paleoclimate Modelling Intercomparison Project (PMIP) except ECBilt/CLIO exhibit a freshwater import across the southern border of the Atlantic basin, which correctly suggested a mono-stable AMOC in these models (Stouffer et al., 2006). However, the only exception, ECBilt/CLIO, though with a freshwater export, failed to indicate a bi-stable AMOC, since its AMOC was still apt to recover in a pulse hosing experiment (de Vries and Weber, 2005). Therefore, $F_{\rm OT}$ does not appear to be an accurate stability indicator for the AMOC, at least in some EMICs or AOGCMs. One

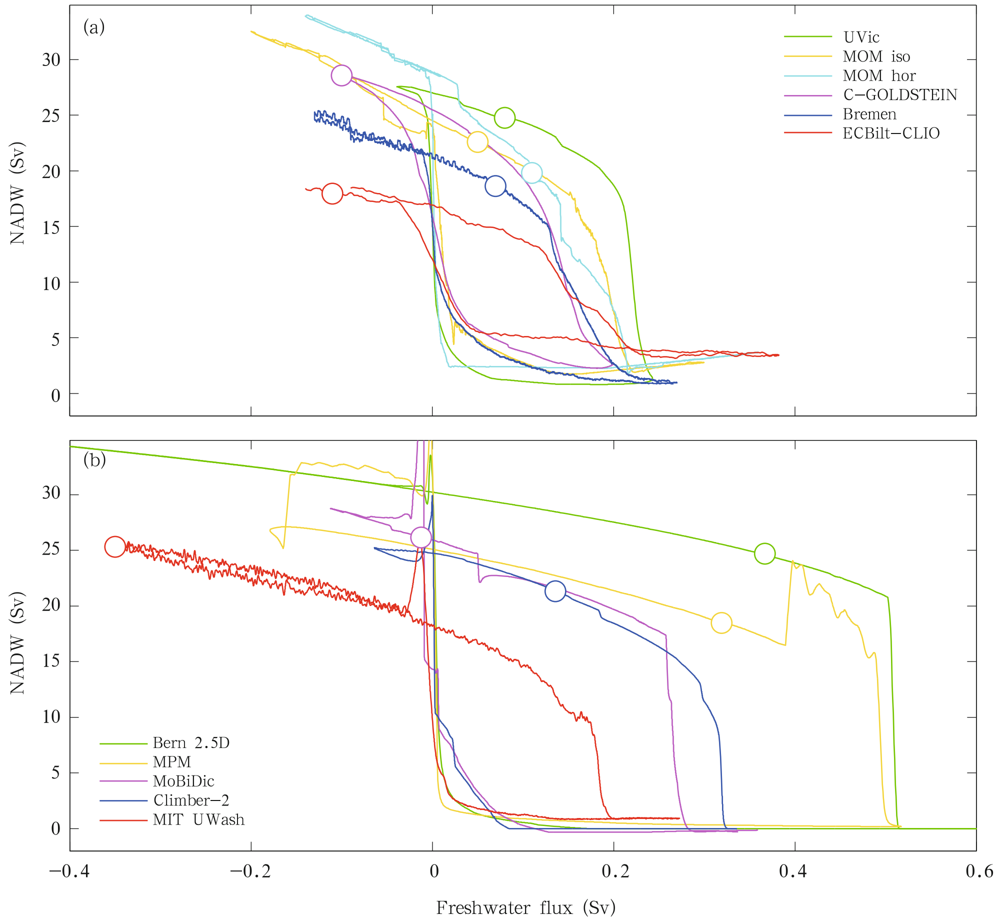

Fig. 4. Hysteresis curves found in the model intercomparison from (a) coupled models with 3-D global ocean models, and (b) those with simplified ocean models (zonally averaged or, in case of the MIT_UWash model, rectangular basins). Curves were slightly smoothed to remove the effect of short-term variability. Circles show the present-day climate state of each model. [From Rahmstorf et al., 2005]

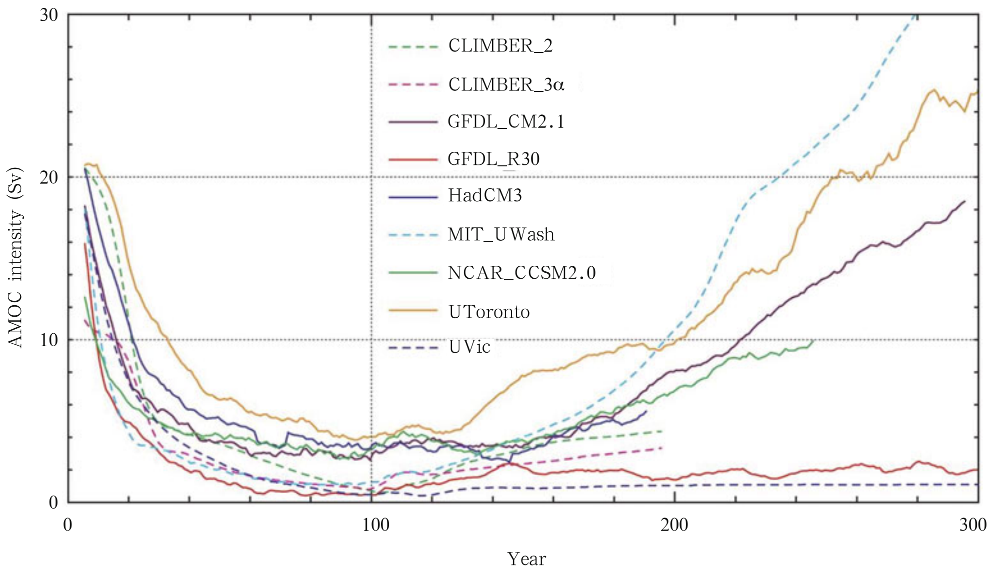

Fig. 5. Time series of the AMOC intensity evolution in the 1.0-Sv water-hosing experiments for the CMIP/PMIP models. [From Stouffer et al., 2006]

probable cause is that $F_{\rm OT}$ includes the Arctic freshwater budget and therefore does not represent the real net freshwater forcing exerted on the AMOC.

In a later study, Dijkstra (2007) proposed an alternative indicator $\sum$, which is defined as the freshwater transport convergence by the AMOC for the Atlantic basin, i.e., the net freshwater transport between the southern and northern boundaries (approximately 34°S and 60°N). This indicator was subsequently demonstrated largely valid in an OGCM coupled with an energy-balance atmosphere model (Huisman et al., 2010). Nevertheless, one concern of the definition of $\sum$ is that the northern boundary of the Atlantic basin is placed at 60°N. This definition excludes the GIN Seas (Greenland, Iceland, and Norwegian Seas) region, a major region for the NADW formation (e.g., Schiller et al., 1997; Holland et al., 2007; Renold et al., 2010). As related to the status of Bering Strait (e.g., Hu et al., 2008, 2012), the freshwater transport through this region is expected to have a significant effect on the AMOC stability (e.g., Holland et al., 2001; Komuro and Hasumi, 2005; Oka and Hasumi, 2006; Rennermalm et al., 2006), so this indicator may not correctly indicate the AMOC stability in some AOGCMs (Liu and Liu, 2013) due to the lack of the freshwater transport via the GIN Seas.

Liu and Liu (2013) proposed a refined indicator, $\Delta M_{\rm ov}$, which includes the GIN Seas region and is defined as the net freshwater transport between the southern and northern boundaries (approximately 34°S and 80°N). Same as $F_{\rm OT}$ and $\sum$, $\Delta M_{\rm ov}$ is derived from a decomposition of freshwater transport and a basin-integrated freshwater budget. In the Atlantic, the meridional freshwater transport can be divided into two parts: the meridional overturning part ($M_{\rm ov}$) that is thought to be associated with the AMOC and the azimuthally asymmetric part ($M_{\rm az}$) that is assumed to be associated with the wind-driven gyre circulation:

$$M_{\rm ov} = -\frac{1}{S_0}\int {\rm d}z\bar{v}(z)\{(\bar{s}) - S_0\}, \tag{1}$$

$$M_{\rm az} = -\frac{1}{S_0}\int {\rm d}z\overline{v'(z)s'(z)}. \tag{2}$$

In Eqs. (1) and (2), S_0 is reference salinity (unit: psu); the overbar and brackets denote the zonal integration and zonal averaging along a latitude circle, respectively; v is the velocity normal to the section and s is salinity; v' and s' are deviations from their along section means. Here, it is worth noting that $M_{\rm ov}$ and $M_{\rm az}$ are just simply geometric decompositions for representing the horizontal and overturning contributions to the freshwater transport, which do not serve as a neat division between the contributions from different physical processes such as the NADW formation and the gyre circulation. Thus, by neglecting the effects of diffusion, the freshwater budget over the Atlantic basin (roughly between 34°S and 80°N) can be approximately estimated as a balance between the net evaporation ($E_{\rm net}$) and the freshwater transport through the southern and northern boundaries, i.e.,

$$\begin{aligned} E_{\rm net} &= M_{\rm azS} + M_{\rm ovS} - M_{\rm azN} - M_{\rm ovN} \\ &= \Delta M_{\rm az} + \Delta M_{\rm ov}, \end{aligned} \tag{3}$$

where $E_{\rm net} = E - P - R - M + {\rm Br}$, i.e., the sum of evaporation E, precipitation $-P$, runoff $-R$, sea ice melting $-M$, and brine rejection Br due to sea ice melting. The subscripts S and N denote southern and northern boundary, respectively. $M_{\rm azS}$ and $M_{\rm azN}$ are associated with the gyre circulation, whilst $M_{\rm ovS}$ and $M_{\rm ovN}$ are associated with the overturning circulation. $\Delta M_{\rm az} = M_{\rm azS} - M_{\rm azN}$ and $\Delta M_{\rm ov} = M_{\rm ovS} - M_{\rm ovN}$ are the convergence due to gyre and overturning, respectively. Again, as mentioned before, the latter ($\Delta M_{\rm ov}$) is defined as the stability indicator of the AMOC since it denotes a basin-scale salinity advection feedback proposed by Stommel (1961). Particularly, if the active AMOC leads to a net freshwater divergence ($\Delta M_{\rm ov} < 0$), this indicator potentially implies a multiple equilibria behavior of the AMOC.

4. The AMOC stability in the modern climate

To assess the bi-stability of the AMOC of the present day, it is essential to examine $\Delta M_{\rm ov}$ and the freshwater budget in the real world. A net evaporation ($E_{\rm net} > 0$) currently exists in the Atlantic (e.g., Schmitt et al., 1989), which is primarily compensated by the oceanic freshwater import. At the southern boarder (around 34°S), the gyre circulation induces a freshwater import ($M_{\rm azS} > 0$) in that the Brazil Current transports saltier water southward in the western boundary of the gyre, whilst the interior flow, especially the Benguela Current, transports fresher water northward (Fig. 6b). On the other hand, the AMOC exports freshwater southward ($M_{\rm ovS} < 0$) due to the salinity stratification at nearly 34°S (Rahmstorf, 1996). Particularly, the surface and thermocline water (< 500 m) at 34°S are saltier than NADW (1000–3500 m) underneath, such that the upper/lower limb of the AMOC transports saltier/fresher water northward/southward to generate a freshwater export (Fig. 6a). Based on available instrumental observations, $M_{\rm ovS}$ has been estimated ranging from –0.34 to –0.1 Sv (Weijer et al., 1999; Huisman et al., 2010; Bryden et al., 2011; Hawkins et al., 2011; Garzoli et al., 2012). For example, Weijer et al. (1999) estimated $M_{\rm ovS}$ of –0.2 Sv using the "best estimate" solution of an inversion from Holfort (1994). Huisman et al. (2010) suggested that $M_{\rm ovS} \approx$ –0.1 Sv based on a dataset from Gouretski and Koltermann (2004). Bryden et al. (2011) estimated $M_{\rm ovS} \approx$ –0.34 to –0.1 Sv based on two transatlantic hydrographic cruises along 24°S in 1983 and 2009 and two different methods. Garzoli et al. (2012) reported several estimations, i.e., a mean value of $M_{\rm ovS}$ = –0.16 Sv from the expendable bathythermograph (XBT) data collected along 27 sections at nominally 35°S during the period 2002–2011; values of $M_{\rm ovS}$ as –0.15 and –0.14 Sv for the cruises conducted during 1993 and 2003; and $M_{\rm ovS}$ of –0.11 Sv from the Argo climatological section. Besides, based on oceanic reanalysis data, Hawkins et al. (2011) estimated that $M_{\rm ovS}$ is mostly within the range of –0.2 to –0.1 Sv. In summary, aforesaid observational values of $M_{\rm ovS}$ are between –0.34 and –0.1 Sv, which suggests a bi-stable AMOC in the modern climate if the transport indicator $F_{\rm OT}$ (here $F_{\rm OT} = M_{\rm ovS}$) is employed.

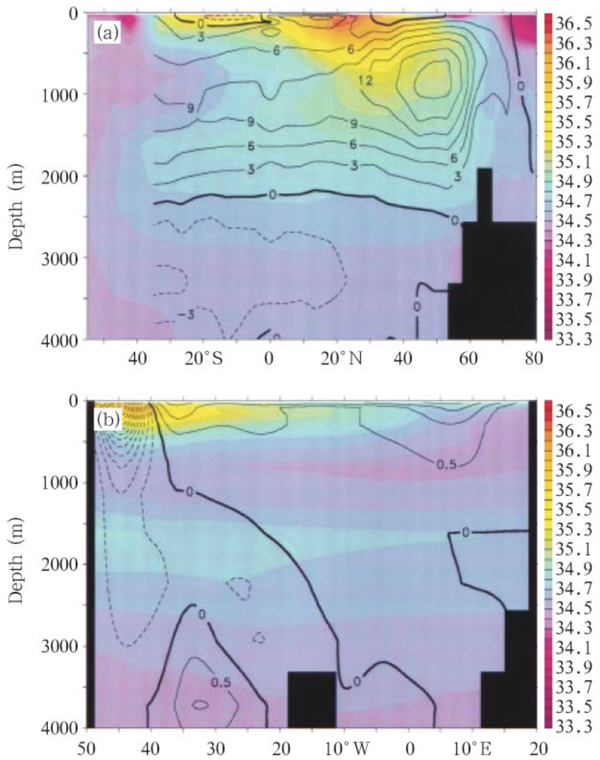

Fig. 6. (a) Atlantic overturning stream function (contoured in Sv) for the "present-day" equilibrium of the global model, superimposed on a plot of Atlantic zonal-mean salinity, and (b) zonal section of meridional flow (contoured in cm s^{-1}) and salinity (color) at 30°S in the global model. [From Rahmstorf, 1996]

At the northern boundary, freshwater transports entering the Atlantic are composed of three components, one via the Fram Strait, one via the western Barents Sea, and the other via the Canadian Arctic Archipelago (refer to Serreze et al., 2006 for more details). For each component, current observations are limited to the total freshwater transport that includes both AMOC ($M_{\rm ovN}$) and gyre ($M_{\rm azN}$) contributions. Individual contribution, especially $M_{\rm ovN}$, is unknown from observation so far. Herein, Liu et al. (2013) resorted to a relationship as diagnosed from a climate model that $M_{\rm ovN}$ is about 80% of the total freshwater import from the Arctic, such that $M_{\rm ovN}$ was estimated as an import of approximately –0.15 Sv. Therefore, by combining the observational values of $M_{\rm ovS}$ and $M_{\rm ovN}$, $\Delta M_{\rm ov}$ was estimated ranging from –0.2 to +0.05 Sv over current Atlantic, which indicates an AMOC close to neutral but with a tendency towards bi-stable under modern climate.

In short, in spite of significant uncertainties, available evidence from both paleo- and modern observations suggests that the AMOC in the real world is likely to be bi-stable. This sets a target to test by using climate models. Nevertheless, climate models, especially the state-of-the-art AOGCMs (see Weber et al., 2007 for the PMIP models; see Weaver et al., 2012 for the CMIP phase 5 (CMIP5) models), exhibit opposite results to observations. The AMOC freshwater transports in the models mostly appear as an import ($M_{\rm ovS} > 0$) across the southern boundary, which then leads to a freshwater convergence over the Atlantic ($\Delta M_{\rm ov} > 0$) (Liu et al., 2014a). As such, it may offer a partial explanation why these AOGCMs tend to simulate a mono-stable AMOC for the present. Particularly, as compared with the observation, AOGCMs generally exhibit a systematic bias in the AMOC freshwater transports at the southern boundary ($M_{\rm ovS}$). The export of $M_{\rm ovS}$ in the observation is biased as an import in AOGCMs (Weber et al., 2007), which then results in a systematic distortion of the AMOC stability. Further analyses show that the distortion of freshwater transport in AOGCMs originates mainly from the salinity bias in the models. In comparison with the observation, most AOGCMs simulate much freshening surface and thermocline waters and a slightly saltier NADW around 34°S, which then results in a freshwater import by the AMOC across the southern border of the Atlantic (Fig. 7). Meanwhile, they also simulate a somewhat stronger freshwater import $M_{\rm ovN}$

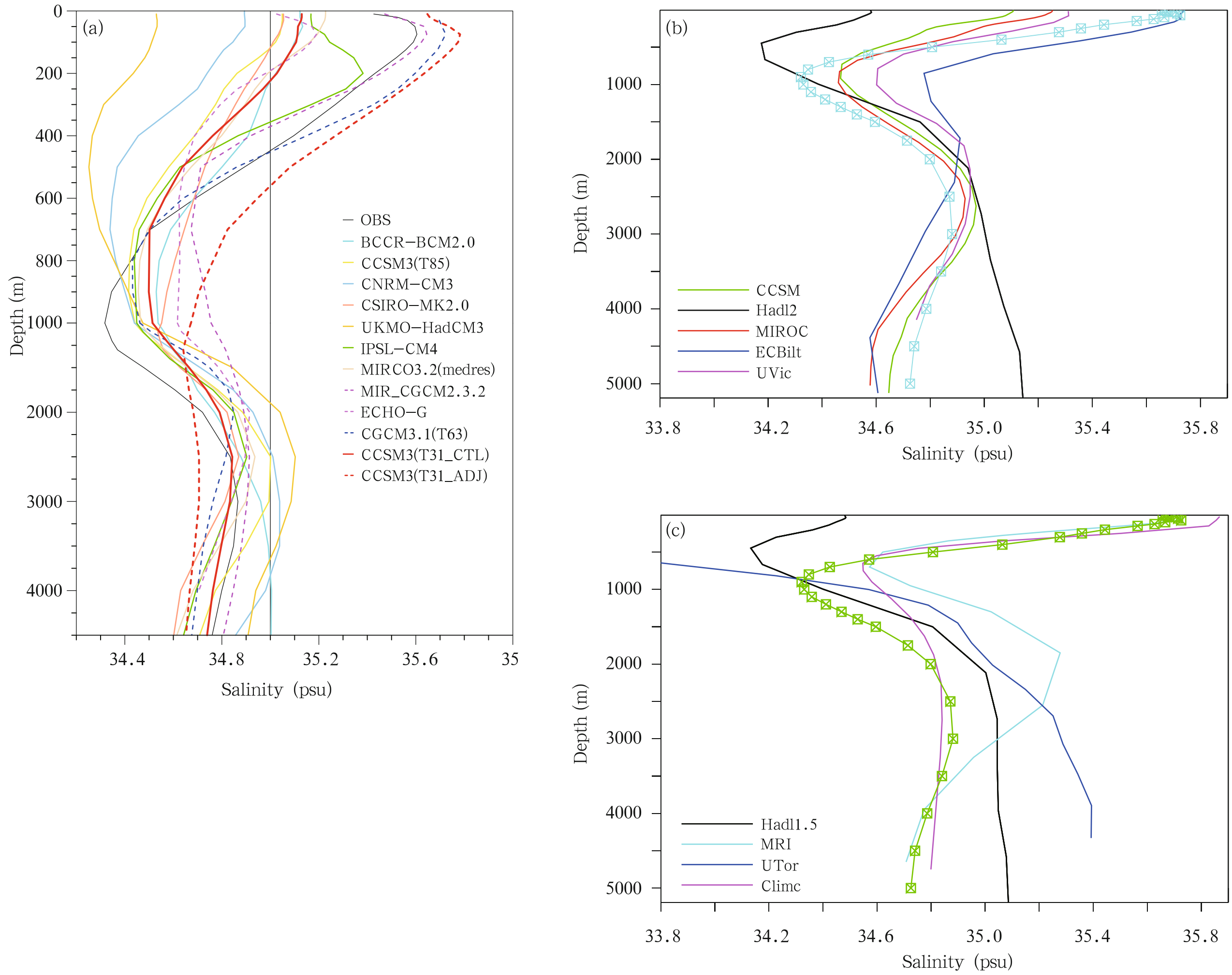

Fig. 7. (a) The zonal mean salinity along 34°S across the Atlantic in the IPCC AR4 models and CCSM3 T31. The observed salinity for WOA/P datasets (Levitus et al., 1998; Steele et al., 2001) is also shown in black curve (from Liu et al., 2013). (b) The zonal-mean salinity at the southern border of the Atlantic basin as a function of depth for the control state (0 ka) for the PMIP2 simulations. (c) As in (b) but for the PMIP1.5-type simulations. In (b) and (c), the observed Levitus salinity profiles are marked by squares. [From Weber et al., 2007]

from the Arctic. The biased imports, especially from the south, by the AMOC lead to a freshwater convergence across the Atlantic basin, which causes a mono-stable AMOC in these AOGCMs.

Further analyses show that this systematic bias in salinity, and in turn, in freshwater transport, is partly caused by the notorious tropical bias associated with the double ITCZ in AOGCMs (Liu et al., 2014a). In the tropical Atlantic (and eastern Pacific), AOGCMs generally suffer from a common bias in the annual mean climate, which is characterized by a double ITCZ straddling across the equator and an excessive cold tongue penetrating westward along the equator (e.g., Mechoso et al., 1995; Davey et al., 2002; Lin, 2007). This bias leads to excessive rainfall and therefore a negative sea surface salinity (SSS) bias in the South Atlantic, freshening the inflow across roughly 34°S via the upper branch of the AMOC. As a result, the southward freshwater export M_{ovS} is reduced greatly or even reversed to a northward import, which induces a freshwater convergence ($\Delta M_{\mathrm{ov}} > 0$) over the Atlantic in most AOGCMs and implies a mono-stable AMOC.

To eliminate the potential bias in the AMOC stability in AOGCMs, the first remedy is to correct the surface climate bias. However, the fix of the double ITCZ problem is beyond the reach of current climate model developers. Therefore, as a practical approach so far, Liu et al. (2014a) employed a global flux adjustment method (e.g., Manabe and Stouffer, 1988; Yin and Stouffer, 2007) to reduce the model climate bias of the NCAR CCSM3 T31, despite the fact that this method is known to have undesirable side effects on climate models (e.g., Marotzke and Stone, 1995; Neelin and Dijkstra, 1995). In the paper, they argued that the flux adjustment method is a useful first step, considering the impossibility of fixing the model bias in current stage and the primary goal of the AMOC stability only.

Figure 8 shows the results from Liu et al. (2014a). Similar to 7 models without flux adjustment in the 4th Assessment Report of the Intergovernmental Panel on Climate Change (IPCC AR4), the NCAR CCSM3 T31 (Collins et al., 2006; Yeager et al., 2006) exhibits a surface freshening bias at 34°S, which then leads to a

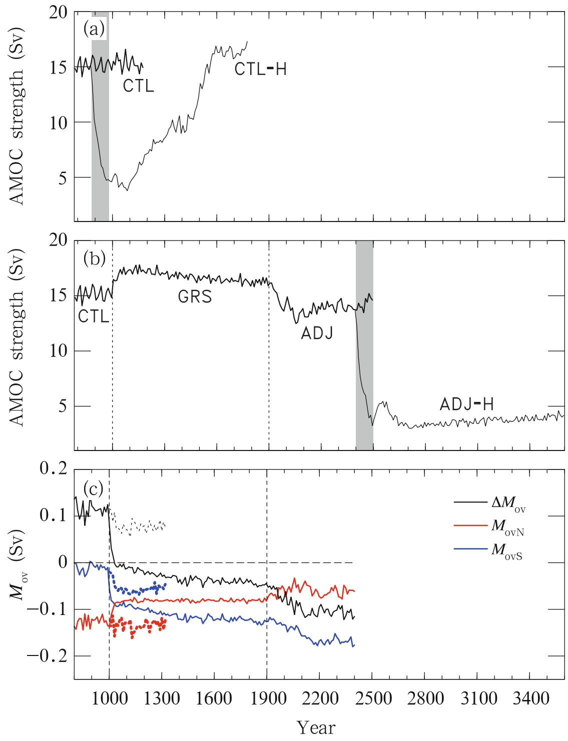

Fig. 8. Time evolutions of the decadal mean (a, b) AMOC strength and (c) AMOC freshwater transports. (a) CCSM3 T31 CTL run (black) and the hosing experiment CTL-H (gray). (b) CCSM3 T31 CTL run (< year 1000), GRS run (year 1000–1900), ADJ run (> year 1900) (black) in the transient period, and the hosing experiment ADJ-H (gray), with the vertical gray dashed lines representing the change time between CTL to GRS and from GRS to ADJ (AMOC strength is defined as the maximum streamfunction value below 500 m within the Atlantic basin; the 100-yr hosing period is shaded in light gray). (c) Evolution of the AMOC freshwater transport at the southern boundary (M_{ovS}; blue solid), northern boundary (M_{ovN}; red solid), and the divergence indicator (ΔM_{ov}; black solid) in the transition period from CTL (< year 1000) to GRS (year 1000–1900), and ADJ (> year 1900). The freshwater transports of the tropical restoring run (TRS; year 1000–1300) are also shown in dashed lines (M_{ovS} in blue, M_{ovN} in red, and ΔM_{ov} in black). Here, CTL denotes the present day control run of the NCAR CCSM3 T31, GRS denotes the run with global restoring SST and SSS, TRS denotes the run with tropical (15°S–15°N) restoring SST and SSS, and ADJ denotes the run with global heat and virtual salt flux adjustment. [From Liu et al., 2013]

weak freshwater export of M_{ovS} = –0.013 Sv. Accordingly, a freshwater convergence of $\Delta M_{\mathrm{ov}} = 0.114$ Sv is generated over the Atlantic basin, implying a mono-stable AMOC. This mono-stable AMOC was then confirmed explicitly in a freshwater hosing experiment. On the other hand, after adopting a global flux adjustment, the ITCZ in CCSM3 remains to the north of the equator and the upper ocean salinity bias is greatly reduced at 34°S. This results in a strong freshwater export of M_{ovS} = –0.185 Sv and in turn a freshwater divergence of ΔM_{ov} = –0.113 Sv across the Atlantic. Such a negative value of ΔM_{ov} indicates a bi-stable AMOC and is validated by a subsequent hosing experiment. Nevertheless, it merits attention that, besides the tropical bias related to the double ITCZ, salinity biases in other regions might also contribute to the distortion in the freshwater transports across the Atlantic and thus the AMOC stability. Seen from Fig. 8c, a simply correction of the tropical bias (TRS) by restoring the SSS and SST in the tropical Atlantic between 15°S and 15°N can only correct about half of the distortion in ΔM_{ov}, as compared with a parallel globally restoring experiment (GRS). This further indicates that the tropical bias related to the double ITCZ plays a major role but far from the whole story in distorting the AMOC stability in current AOGCMs.

5. The AMOC stability in the future climate

Studies related to the IPCC AR4 (e.g., Meehl et al., 2007) and CMIP5 model results (Weaver et al., 2012) show that the AMOC in the historical simulations matches more closely to observations than that in the CMIP3 (Cheng et al., 2013). Also, similar to the CMIP3, all the CMIP5 models predict a weakening of the AMOC by 22%–30% in response to the increase of atmospheric CO_2 in the 21st century; this weakening, however, will develop gradually with little chance of abrupt collapse (also see Delworth et al., 2008). Moreover, the global warming is most unlikely to result in an AMOC collapse beyond the end of the 21st century. Particularly, Weaver et al. (2012) explored the AMOC behavior under anthropogenic radiative forcing, greenhouse gas, and aerosol emission in various scenarios of the representative concentration pathways (RCPs; detailed in Moss et al., 2010). They adopted F_{OT} as a predictor of the transient, radiatively forced behavior of the AMOC and discovered that 40% of the CMIP5 models were in a bi-stable regime of the AMOC during the RCP integrations. In the strongest forcing scenario, RCP8.5, two CMIP5 models (the NCAR CCSM4 and Bern3D) eventually realized a slow shutdown of the AMOC. A further analysis (Jahn and Holland, 2013) showed that the AMOC collapse in the NCAR CCSM4 was caused by such a process: an enhanced Arctic sea-ice melting increased the liquid freshwater imports from the Arctic, freshened the surface layer in the northern North Atlantic, and finally shut down the deep convection there, the NADW formation, and the AMOC. It is interesting to find that almost all the CMIP5 AOGCMs abandon the flux adjustment; thus potentially, they have a tropical bias due to the double ITCZ, and in turn, a bias in the AMOC stability. Provided that such a bias in the AMOC stability is corrected in the CMIP5 models, it remains unclear how the model AMOC will behave in future RCP scenarios.

6. Conclusions and discussion

In this paper, we have reviewed the history of research on the AMOC stability and its significance to climate change in the past, present, and future. Of particular interest is the question if the AMOC will remain stable or change abruptly in the near future. To predict future abrupt changes of the AMOC, it is essential to build a state-of-the-art climate model with credible AMOC stability. Nevertheless, understanding and/or evaluating the AMOC stability in state-of-the-art AOGCMs is challenging. Recent results suggest that state-of-the-art AOGCMs may exhibit a systematic bias in the AMOC stability. This systematic bias, if true, is likely to distort future climate projections of abrupt climate change significantly. As a result, some approaches to correct this bias, such as a global flux adjustment, are needed prior to the conduction of climate simulations and predictions.

It should be pointed out that, in spite of significant progress, many important issues on the stability of the AMOC remains open. First, previous studies on the AMOC stability indicator are based on an active AMOC in equilibrium and may not be applicable for an evolving AMOC (Hawkins et al., 2011). Meanwhile, paleo-data analysis suggests that the past AMOC has never maintained a perfect equilibrium (e.g., Severinghaus and Brook, 1999; McManus et al., 2004). As such, a generalized stability indicator is proposed (Sijp, 2012; Sijp et al., 2012; Liu et al., 2013). It can be defined as $L = \partial\overline{\Delta M}_{\rm ov}/\partial\overline{\psi}$, where L denotes the indicator, $\overline{\psi}$ and $\overline{\Delta M}_{\rm ov}$ are the AMOC strength and the AMOC-induced freshwater transport convergence in the equilibrium state (Liu et al., 2013). In contrast to $\Delta M_{\rm ov}$, L is more generalized in showing how the Atlantic freshwater transport modulates as the AMOC transits from one equilibrium to another. It does not require a divergence-free freshwater transport in the Atlantic for a collapsed AMOC and therefore manages to correctly monitor the AMOC stability through a slow evolution, such as the last deglaciation (Liu et al., 2014b).

Second, the diagnostic indicator, either $F_{\rm OT}$ or $\Delta M_{\rm ov}$, is based on a hypothesis derived from the box model of Rahmstorf (1996), i.e., a zero net freshwater transport ($F_{\rm OT} = 0$ or $\Delta M_{\rm ov} = 0$) is induced by a collapsed AMOC due to the absence of mass transport. This hypothesis is usually taken for granted without validation (e.g., Hawkins et al., 2011; Weaver et al., 2012; Liu et al., 2013). Recently, Liu and Liu (2014) found that this hypothesis can still be achieved in an AOGCM, but by a compensation of non-zero mass and freshwater transports across $M_{\rm ovS}$ and $M_{\rm ovN}$. Therefore, one should be cautious in interpreting the AMOC stability in terms of the freshwater transport.

Third, by definition, $\Delta M_{\rm ov}$ indicates a basin-scale freshwater feedback associated with the NADW cell and is valid only when the AMOC collapses to a very weak NADW cell (e.g., Liu and Liu, 2013). However, some studies (e.g., Gregory et al., 2003; Saenko et al., 2003; Sijp and England, 2006; Sijp et al., 2012) showed that, for a bi-stable AMOC, the collapsed circulation appears as an Antarctic intermediate water (AAIW) reverse cell whose non-linear behaviors suppress the NADW formation and governs the collapsed state. The AAIW reverse cell has a strong effect on the Atlantic freshwater budget, which may render the indicator $\Delta M_{\rm ov}$ no longer valid.

Finally, one uncertainty on the AMOC stability is Agulhas leakage, a transport of warm and salty Indian Ocean waters into the Atlantic Ocean (Gordon et al., 1992; De Ruijter et al., 1999; Lutjeharms, 2006). Paleo-proxy records show substantial glacial-to-interglacial variations in the amount of Agulhas leakage (e.g., Biastoch et al., 2008, 2009; Beal et al., 2011; de Deckker et al., 2012), which is accompanied by modulations of the stratification at 34°S and thus the AMOC stability. However, many coarse resolution climate models fail to correctly simulate Agulhas leakage due to a poor resolving nonlinear dynamics (inertial mechanisms and ring formation) associated with Agulhas leakage. Thus, a correct simulation of Agulhas leakage using high-resolution climate models (e.g., Biastoch et al., 2008; Tsugawa and Hasumi, 2010) is needed in the future for assessing the stability of the AMOC.

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