Why is the AMOC Monostable in Coupled General Circulation Models?

WEI LIU

Center for Climatic Research, and Department of Atmospheric and Oceanic Sciences, University of Wisconsin–Madison, Madison, Wisconsin

ZHENGYU LIU

Laboratory of Climate, Ocean and Atmosphere Studies, Peking University, Beijing, China, and Center for Climatic Research, and Department of Atmospheric and Oceanic Sciences, University of Wisconsin–Madison, Madison, Wisconsin

ESTHER C. BRADY

Climate and Global Dynamics Division, National Center for Atmospheric Research, Boulder, Colorado

(Manuscript received 2 May 2013, in final form 10 December 2013)

ABSTRACT

This paper is concerned with the question: why do coupled general circulation models (CGCM) seem to be biased toward a monostable Atlantic meridional overturning circulation (AMOC)? In particular, the authors investigate whether the monostable behavior of the CGCMs is caused by a bias of model surface climatology. First observational literature is reviewed, and it is suggested that the AMOC is likely to be bistable in the real world in the past and present. Then the stability of the AMOC in the NCAR Community Climate System Model, version 3 (CCSM3) is studied by comparing the present-day control simulation (without flux adjustment) with a sensitivity experiment with flux adjustment. It is found that the monostable AMOC in the control simulation is altered to a bistable AMOC in the flux-adjustment experiment because a reduction of the surface salinity biases in the tropical and northern North Atlantic leads to a reduction of the bias of freshwater transport in the Atlantic. In particular, the tropical bias associated with the double ITCZ reduces salinity in the upper South Atlantic Ocean and, in turn, the AMOC freshwater export, which tends to overstabilize the AMOC and therefore biases the AMOC from bistable toward monostable state. This conclusion is consistent with a further analysis of the stability indicator of two groups of IPCC Fourth Assessment Report (AR4) CGCMs: one without and the other with flux adjustment. Because the tropical bias is a common feature among all CGCMs without flux adjustment, the authors propose that the surface climate bias, notably the tropical bias in the Atlantic, may contribute significantly to the monostability of AMOC behavior in current CGCMs.

1. Introduction

Theoretical studies have suggested that the Atlantic meridional overturning circulation (AMOC) can possess multiple equilibrium states (e.g., Stommel 1961; Bryan 1986) because of the positive feedback between the AMOC and basinwide salinity transport (Marotzke 1996). As such, the AMOC can jump between ‘‘on’’ and ‘‘off’’ states under natural perturbation forcings, such as the deglacial meltwater pulses, and result in abrupt climate changes as observed in past climate records (e.g., Broecker et al. 1985; Sarnthein et al. 1994; Clark et al. 2002). This mechanism of abrupt climate change may also be relevant for future climate changes in response to large anthropogenic perturbations (Kuhlbrodt et al. 2007). However, in climate models, so far, the bistability of the AMOC has been demonstrated mostly in simpler climate models, either conceptual climate models (e.g., Stommel 1961; Marotzke and Stone 1995; Rahmstorf 1996; Marotzke 2000) or earth system models of intermediate complexity (EMICs) (e.g., Dijkstra and Neelin 1999; Ganopolski and Rahmstorf 2001; Weaver et al. 2003; Knorr and Lohmann 2003; Timmermann et al. 2009). In the state-of-the-art coupled climate models,

Corresponding author address: Z. Liu, 1225 W. Dayton St., Madison, WI 53706.
E-mail: zliu3@wisc.edu

DOI: 10.1175/JCLI-D-13-00264.1

or the so-called coupled general circulation models (CGCMs), except for two cases (Manabe and Stouffer 1988; Hawkins et al. 2011a,b), the AMOC seems to exhibit monostable behavior under meltwater perturbations, with the AMOC eventually recovering after the termination of the meltwater forcing (Stouffer et al. 2006). In those CGCMs with a monostable AMOC, abrupt climate changes can be generated only by an abrupt change in external forcing, such as the meltwater forcing (e.g., Liu et al. 2009). This raises several questions. First, why does the AMOC tend to be monostable in CGCMs? Second, is the AMOC stability behavior correct in the CGCMs, compared to the real world? This second question naturally brings out a third question: What is the stability behavior of the AMOC in the real world? These questions are important. If the stability behavior of the AMOC turns out to be incorrect in the CGCMs, it will compromise our confidence of the projection of the response of the AMOC to future climate change (Gregory et al. 2005; Schmittner et al. 2005) because these CGCMs are the major models that are now being used for future projections (Meehl et al. 2007).

Previous studies on AMOC stability have left many questions. Extensive paleoclimate records have shown strong evidence of abrupt climate changes in the North Atlantic region—notably the Dansgaard–Oeschger (D/O) events and the Heinrich events—suggesting the possibility of a bistable AMOC during the deglaciation (Broecker et al. 1985; Rahmstorf 2002; Clark et al. 2002). Using oceanic freshwater transport associated with the overturning circulation as an indicator of the AMOC bistability (Rahmstorf 1996), analyses of present-day observations also indicate a bistable AMOC (Weijer et al. 1999; Huisman et al. 2010; Hawkins et al. 2011a,b; Bryden et al. 2011; Garzoli et al. 2012; see more details in section 2). These observational studies suggest a potentially bistable AMOC in the real world. In contrast, sensitivity experiments in CGCMs tend to show a monostable AMOC (Stouffer et al. 2006), indicating a model bias toward a monostable AMOC. This monostable bias of the AMOC in CGCMs, as first pointed out by Weber et al. (2007) and later confirmed by Drijfhout et al. (2011), could be related to a bias in the northward freshwater transport in the South Atlantic by the meridional overturning circulation. The CGCM studies of Weber et al. (2007) and Drijfhout et al. (2011) are important, because they suggest that the monostable AMOC in current CGCMs is likely to be caused by some common errors in the model climate state. However, these studies have left a key question unanswered: What is the cause of the bias of the freshwater transport? Is it caused by a bias in the AMOC circulation, the salinity distribution, or ocean–atmosphere feedback? A further fundamental question is: why do all the CGCMs tend to exhibit the same bias? These questions are the focus of this study.

Here, we will combine the sensitivity experiments on the AMOC stability in the National Center for Atmospheric Research (NCAR) CCSM3 (Yeager et al. 2006) with the analyses of a suite of state-of-the-art CGCMs in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) (see Table 1 for a list of the 12 CGCMs models) (Meehl et al. 2007). The bistability of the AMOC is analyzed in light of recent developments, including a refined stability indicator. Our study suggests that the AMOCs in most CGCMs are likely to be overstabilized significantly by the bias in their oceanic surface climate. In particular, we found that the common tropical bias associated with the double ITCZ in current CGCMs (Mechoso et al. 1995; Davey et al. 2002; Lin 2007) contributes significantly and systematically to an excessively freshwater in the upper ocean in the tropical South Atlantic; this surface fresh bias reduces the freshwater export by the AMOC and eventually overstabilizes the AMOC in the models.

This paper is arranged as follows. First, in section 2, we briefly review current literature regarding the real-world AMOC bistability in observational studies of the past and present climate. In section 3, we study why the AMOC tends to be monostable by focusing on one particular CGCM, the NCAR CCSM3. In section 4, we further analyze the potential AMOC stability behavior in state-of-the-art CGCMs included in the IPCC AR4. A summary and further discussion are given in section 5.

2. The AMOC bistability in the real world

To establish a real-world benchmark for model evaluation, we first briefly review observational literature. The ultimate test of the AMOC stability is the real world itself. This test, however, is difficult to perform for the present-day world. The AMOC has remained rather stable in the last ~6000 years after the last deglaciation (McManus et al. 2004), suggesting that the present AMOC is stable to small natural perturbations, such as the fluctuations of rainfall or wind associated with internal atmospheric variability and short-term coupled ocean–atmosphere variability. As such, whether the AMOC is bistable can only be tested with experiments using large, yet artificial, perturbations. Since this type of artificial experiment is impractical in the real world, we are left with two options: one is to rely on records of past climate evolution and the other is to develop diagnostic indicators that can be used to infer the AMOC stability based on current observations. In

TABLE 1. Freshwater transports in 12 IPCC AR4 CGCMs (Meehl et al. 2007) (including CCSM3 T31 CTL and ADJ). These models include eight models without flux adjustment and four models with flux adjustment (with heat and freshwater fluxes) (MRI-CGCM2.3.2 applies adjustments, including the momentum, in the tropics between 12°S and 12°N). The observational estimation is adopted from Weijer et al. (1999), Huisman et al. (2010), Hawkins et al. (2011a,b), Bryden et al. (2011), and Garzoli et al. (2012) for M_{ovS} and from Serreze et al. (2006) for M_{ovN}. For each model, the freshwater transport uses 100 yr of model output. The IPCC model estimation is based on the twentieth-century simulation. For most IPCC models, our estimation of freshwater transport is consistent with a previous estimation on the simulations of the preindustrial control and present-day control (Drijfhout et al. 2011).

Model ID (country)		M_{ovS}	M_{ovN}	ΔM_{ov}
Observation	Model	−0.34 to −0.1	−0.16	−0.18 to +0.06
	No flux adjustment			
BCCR-BCM2.0 (Norway)	Bjerknes Centre for Climate Research Bergen Climate Model, version 2.0	0.023	−0.127	0.150
CCSM3(T85) (United States)	Community Climate System Model, version 3	0.078	−0.185	0.263
CNRM-CM3 (France)	Centre National de Recherches Météorologiques Coupled Global Climate Model, version 3	0.290	−0.097	0.387
CSIRO MK2.0 (Australia)	Commonwealth Scientific and Industrial Research Organisation Mark, version 3.0	−0.030	−0.465	0.435
HadCM3 (United Kingdom)	Hadley Centre Coupled Model, version 3	0.359	−0.013	0.372
IPSL-CM4 (France)	L'Institut Pierre-Simon Laplace Coupled Model, version 4	−0.008	−0.128	0.120
MIROC3.2 (medres) (Japan)	Model for Interdisciplinary Research on Climate, version 3.2 (medium resolution)	−0.004	−0.110	0.106
CCSM3(T31) (United States)	Community Climate System Model, version 3	−0.013	−0.127	0.114
Ensemble mean		0.087	−0.156	0.243
	Flux adjustment			
CGCM3.1(T63) (Canada)	Canadian Centre for Climate Modelling and Analysis (CCCma) Coupled Global Climate Model, version 3.1	−0.118	−0.082	−0.036
MRI-CGCM2.3.2 (Japan)	Meteorological Research Institute Coupled Atmosphere–Ocean General Circulation Model, version 2.3.2	−0.080	−0.160	0.080
ECHO-G (Germany and South Korea)	ECHAM and the global Hamburg Ocean Primitive Equation	0.046	−0.009	0.055
CCSM3(T31_ADJ) (United States)	Community Climate System Model, version 3	−0.170	−0.062	−0.108
Ensemble mean		−0.081	−0.078	−0.003

spite of substantial uncertainties, both past and present analyses suggest the potential presence of a bistable AMOC in the real world.

a. Paleoclimate analyses

Paleoclimate reconstructions in the last 120 000 years show two major types of abrupt climate changes likely associated with the AMOC: the Dansgaard–Oeschger events and the Heinrich events (see reviews of, e.g., Rahmstorf 2002; Clark et al. 2002, 2007). Both D/O events and Heinrich events are abrupt changes with the maximum temperature response in the North Atlantic region and an opposite response in the Southern Hemisphere. This so-called bipolar "see saw" response is a fingerprint of the AMOC change associated with its heat transport (Crowley 1992; Stocker and Johnsen 2003). The succeeding D/O events tend to occur in a multiple of ~1500 yr, with an abrupt warming in the North Atlantic occurring in decades. The Heinrich events tend to occur irregularly ~7000–10 000 yr and are indicated by distinct layers of ice-rafted debris in the North Atlantic (Bond et al. 1992). The North Atlantic Deep Water (NADW) production is reduced from the interstadial mode to the stadial mode in a D/O event and is interrupted in a Heinrich event (Sarnthein et al. 1994)—the former corresponding to a weaker AMOC while the latter a collapsed AMOC. However, in spite of 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; Shakun et al. 2012), the origin of these abrupt changes of the AMOC has remained unclear.

The abrupt D/O events have been proposed as permitted by the bistability of the AMOC, which enables the Atlantic climate to switch between the interstadial mode and stadial mode. Nevertheless, the trigger mechanism for these switches still remains controversial. Since there is no evidence of abrupt changes in the major forcings accompanying the D/O events (such as the meltwater pulses), these abrupt D/O events are likely to be associated with the bistability of the AMOC. When the AMOC is in or near the margin of a bistable state, stochastic climate variability of finite amplitude may drive the AMOC, deviating significantly away from the its long-term mean state (e.g., Ganopolski and Rahmstorf 2001; Dijkstra et al. 2004) or even switching between different states (Cessi 1994).

The Heinrich events are accompanied by major meltwater pulses. However, the lead–lag relationship between the meltwater pulse and the AMOC change remains unclear because of the poor chronology of the reconstructions, especially for the meltwater history. For example, the relative chronology, and even the location, of the most dramatic and well-documented meltwater event, the meltwater pulse 1A, have remained controversial (Clark et al. 1996; Peltier 2005; Stanford et al. 2006; Deschamps et al. 2012). The lead–lag relationship is crucial for the understanding of the bistability of the AMOC. If the abrupt AMOC change is not preceded immediately (within the time scale of the AMOC adjustment, e.g., a few hundreds of years) by an abrupt change in the meltwater flux, this abrupt event is likely to be caused by a bistable AMOC in response to a smooth change of the freshwater forcing (e.g., Ganopolski and Rahmstorf 2001; Weaver et al. 2003; Knorr and Lohmann 2003). Otherwise, this event is likely forced by the instability of the ice sheet (e.g., MacAyeal 1993) and the resulting abrupt change of the meltwater flux on a monostable AMOC (e.g., Liu et al. 2009).[1] Since there is no strong evidence supporting a short (or no) lag time between all AMOC abrupt changes and the corresponding meltwater pulses, the Heinrich events are also likely to be caused by a bistable AMOC.

b. Present-day analyses

To assess the bistability of the present-day AMOC, it is essential to develop a credible diagnostic indicator for the AMOC bistability only based on the observed climate state. Rahmstorf (1996) first proposed that the AMOC freshwater transport at the southern boundary of the South Atlantic at ~30°S (M_{ovS}) can be used as a diagnostic indicator of the AMOC bistability, as it is associated with the salinity advection feedback and in turn the multiple equilibrium proposed by Stommel (1961). In spite of its simplicity, this indicator has been largely validated later by sensitivity experiments in EMICs (de Vries and Weber 2005; Drijfhout et al. 2011). In these modeling studies, the equivalent freshwater transport associated with the AMOC along a zonal section is calculated approximately as

$$M_{\mathrm{ov}} = -\frac{1}{S_0}\int \overline{v}(z)[\overline{s}(z) - S_0]\,dz, \tag{1}$$

where v is the velocity normal to the section; s is the salinity; an overbar indicates the along-section mean; a square bracket denotes the along-section integration; and $S_0 \sim 35$ psu is a reference salinity, which can be taken as the basin mean salinity in the Atlantic. It should be kept in mind that the freshwater transport in Eq. (1) is not a truly freshwater transport associated with the AMOC; rather, it is only the freshwater transport carried by the zonal mean overturning circulation. This overturning transport of freshwater is usually considered a good approximation of the freshwater transport associated with the AMOC, although the accuracy of this approximation has not been studied carefully, a point to be returned later.

In addition to the AMOC freshwater transport in the South Atlantic, we should also consider the freshwater transport at the northern boundary of the North Atlantic. This follows because, in the real world or in a model with a realistic setting, a perturbation on the AMOC may also induce anomalous freshwater exchange between the North Atlantic and the Arctic. Therefore, the AMOC-induced freshwater transport at the northern boundary[2] (M_{ovN}) can also affect the freshwater budget and, in turn, the stability indicator—a point to be returned later. Therefore, the transport indicator was further refined to a divergence indicator ($\Delta M_{\mathrm{ov}} = M_{\mathrm{ovS}} - M_{\mathrm{ovN}}$), which is the difference of the AMOC freshwater transports across the southern (M_{ovS}) and northern (M_{ovN}) boundaries. This divergence indicator was first proposed in an EMIC study (Dijkstra 2007) and then validated in a CGCM after some further

[1] The relatively weaker 8.2-ka event (Alley et al. 1997) seems to be generated by abrupt meltwater pulses on a monostable AMOC (Renssen et al. 2001).

[2] The freshwater transport for the cross-sectional mean overturning circulation at the northern boundary is calculated as the cross-sectional mean overturning circulation across each section [the Canadian Arctic Archipelago (CAA), Fram Strait, and the western shelf of the Barents Sea] separately first before summed (as in Liu and Liu 2013).

refinement (Liu 2012; Liu and Liu 2013, 2014). The physical mechanism relating the freshwater transport and AMOC bistability can be understood as follows. A divergence of the AMOC freshwater transport ($\Delta M_{ov} < 0$), which can be caused by a freshwater export across the southern ($M_{ovS} < 0$) or northern ($M_{ovN} > 0$) boundary, indicates a bistable AMOC in response to freshwater perturbation owing to the positive feedback with the salinity advection (Stommel 1961; Rahmstorf 1996; Marotzke 1996). Suppose an active AMOC produces a climate state with a freshwater divergence. An initial freshwater perturbation in the North Atlantic weakens the AMOC and, in turn, the associated freshwater divergence. The reduced freshwater divergence leads to an anomalous convergence, or accumulation, of freshwater in the North Atlantic, which amplifies the initial freshwater perturbation and results in the collapse of the AMOC.

The freshwater budget in the Atlantic basin can be estimated approximately (neglecting the effect of oceanic mixing-induced transport) as the balance between the net evaporation E (evaporation minus precipitation and river runoff) and the freshwater transport through the southern and northern boundaries as

$$E = M_{azS} + M_{ovS} - M_{azN} - M_{ovN} = \Delta M_{az} + \Delta M_{ov}, \tag{2}$$

where M_{azS} and M_{azN} are associated with the gyre circulation; M_{ovS} and M_{ovN} are associated with the overturning circulation; and $\Delta M_{az} = M_{azS} - M_{azN}$ and $\Delta M_{ov} = M_{ovS} - M_{ovN}$ are the convergences due to gyre and overturning, respectively. As discussed on the overturning transport in Eq. (1), this separation of the total freshwater transport into the ''overturning'' and ''gyre'' parts is a convenient approximation and has not been fully justified dynamically. Strictly speaking, this separation only informs about the horizontal and overturning contributions to the freshwater transport but does not clearly distinguish between the contributions from different physical processes (the buoyance-induced AMOC versus the wind-driven gyre)—a point to be returned later. At present, the Atlantic is a net evaporation basin ($E > 0$) (Baumgartner and Reichel 1975; Wijffels et al. 1992). This net evaporation, ignoring the northern boundary transport for the time being, should be supplied by the freshwater transport from the southern boundary. This freshwater import is currently carried by the gyre circulation ($M_{azS} > 0$), with the Brazil Current transporting saltier water southward at the western boundary and the interior flow, especially the Benguela Current, transports fresher water northward. Opposite to the gyre transport, the AMOC overturning circulation exports freshwater southward ($M_{ovS} < 0$) because the southward outflow of the North Atlantic Deep Water (1000–3500 m) is on average fresher than the return flow in the upper ocean (<1000 m), especially in the thermocline (<500 m), as seen in the zonal mean salinity at the southern boundary in the observation (Rahmstorf 1996; also see Fig. 3a). The freshwater export at the southern boundary has been estimated in the range of ~ -0.34 to -0.1 Sv (Sv $\equiv 10^6\,\mathrm{m^3\,s^{-1}}$) (Weijer et al. 1999; Huisman et al. 2010; Hawkins et al. 2011a,b; Bryden et al. 2011; Garzoli et al. 2012). Weijer et al. (1999) calculated that $M_{ovS} = -0.2$ Sv by using the ''best estimate'' solution of an inversion discussed by Holfort (1994). Huisman et al. (2010) calculated $M_{ovS} \approx -0.1$ Sv based on a dataset from Gouretski and Koltermann (2004). Bryden et al. (2011) obtained a range of $M_{ovS} \approx -0.1$ Sv to -0.34 Sv from two transatlantic hydrographic cruises along 24°S in 1983 and 2009. Garzoli et al. (2012) obtained several estimations of M_{ovS}, ranging from -0.16 Sv (XBT data along 35°S for the period 2002–11) and -0.15 Sv and -0.14 Sv [CDT data along 30°S during two World Ocean Circulation Experiment (WOCE) realizations in 1993 and 2003] to -0.11 Sv (Argo climatological section along 30°S). In addition, Hawkins et al. (2011a) estimated M_{ovS} from six sets of ocean reanalyses, with the M_{ovS} in the range of -0.2 to -0.1 Sv. In summary, present-day observational results showed that M_{ovS} is between -0.34 Sv and -0.1 Sv, favoring a bistable AMOC if the transport M_{ovS} is adopted as the bistability indicator.

Now we take into account of the northern boundary exchange with the Arctic (Liu and Liu 2013). The stability indicator should adopt the convergence indicator ΔM_{ov} accordingly. Current observational analyses suggest an import of fresher Arctic water into the North Atlantic from the northern boundary ($M_{ovN} < 0$) (Serreze et al. 2006). This freshwater import compensates part of the southern boundary export and therefore acts as a stabilizing factor (Liu and Liu 2013). This observational transport consists of the transports through Fram Strait ($M_{ovFRA} \sim -0.084$ Sv), the western Barents Sea ($M_{ovBAR} \sim -0.005$ Sv), and the Canadian Arctic Archipelago ($M_{ovCAA} \sim -0.10$ Sv) (Serreze et al. 2006). It should be noted, however, that this transport is the total freshwater transport that consists of both AMOC (M_{ovN}) and gyre (M_{azN}) transports. We are unaware of direct observational estimates of the freshwater transport associated with the AMOC (M_{ovN}). We found through diagnosing climate models, for example, CCSM3(T31) (Liu and Liu 2013), that M_{azN} is usually much smaller than the total transport such that the overturning transport M_{ovN} is about 80% of the total transport. Therefore, the observational M_{ovN} was estimated as an import of ~ -0.15 Sv here (Table 1). These estimates suggest a ΔM_{ov} ranging from ~ -0.2 Sv to $+0.05$ Sv (Table 1), which indicates a present-day AMOC close to neutral but with a tendency toward bistable.

In short, in spite of significant uncertainty, available evidence from both paleo and modern sources suggest that the AMOC in the real world is likely to be bistable. This sets a target to be tested by climate models. Without confusion, we will often refer to the AMOC stability by its convergence indicator in the following discussion.

3. Climate biases and the AMOC bistability in CCSM3

a. *The control experiment*

To understand the cause for the monostable AMOC in CGCMs, we will first study in detail one model, the NCAR CCSM3. The CCSM3 is a global, coupled ocean–atmosphere–sea ice–land surface climate model without flux adjustment (Collins et al. 2006) and is one of the models in IPCC AR4 (Meehl et al. 2007). We use the version of T31_gx3v5 resolution (Yeager et al. 2006), which, although not included in the IPCC AR4, simulates a model climate similar to the standard version. The control experiment (CTL) is run for 1180 years forced by the greenhouse gases of the present day (1990) (Liu and Liu 2013). The overturning transport of the AMOC is 15 Sv. The AMOC exports a small amount of freshwater transport to the south ($M_{ovS} \sim -0.03$ Sv) and imports a significant amount of fresher Arctic water from the north ($M_{ovN} \sim -0.13$ Sv) (Fig. 1, before year ~1000), leading to a freshwater convergence [$\Delta M_{ov} = -0.03\,\text{Sv} - (-0.13\,\text{Sv}) = 0.1$ Sv], which implies a monostable AMOC. The monostable AMOC is confirmed with a hosing experiment (CTL-H) in which a strong freshwater perturbation of 1 Sv is imposed over the high-latitude North Atlantic (50°–70°N) for 100 years (Fig. 1a). The AMOC transport first collapses rapidly in the first 100 yr, in direct response to the meltwater forcing, and then recovers in 500 yr after the termination of the freshwater pulse (Fig. 1a). Were the southern boundary transport M_{ovS} used as the indicator, the small export of 0.03 Sv would have implied a bistable AMOC. Here, this southern boundary export is overwhelmed by the freshwater import from the Arctic, leading to a convergence, suggesting a monostable AMOC in the divergence indicator. This gives an example of the convergence ΔM_{ov} being a more precise indicator than the transport M_{ovS} for the AMOC bistability (Liu and Liu 2013).

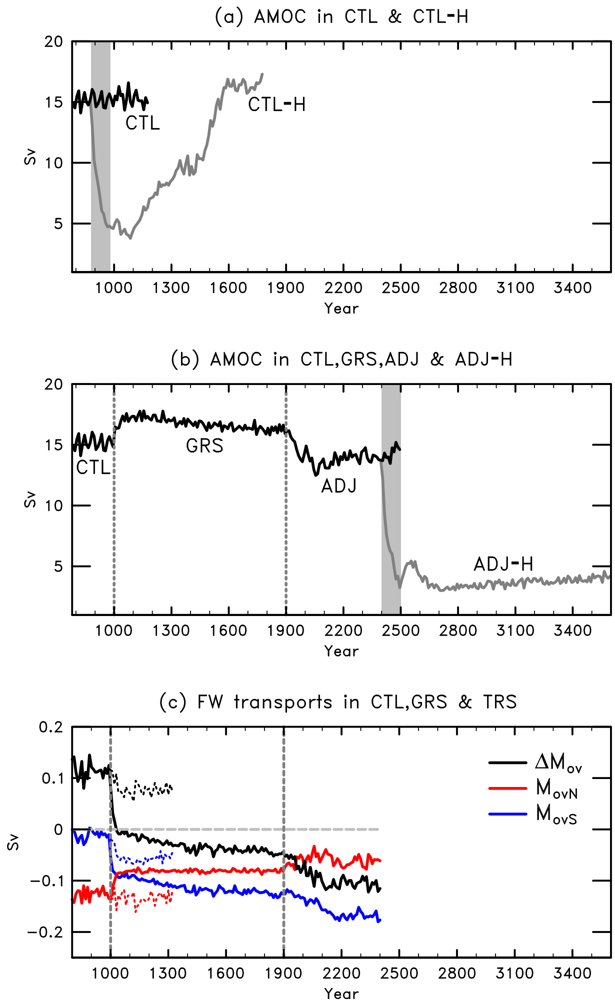

FIG. 1. Time evolution of the decadal mean AMOC strength and AMOC freshwater transports. The AMOC strength is shown in (a) CCSM3 T31 CTL run (black) and the hosing experiment CTL-H (gray) and (b) CCSM3 T31 CTL run (<year 1000), GRS run (years 1000–1900), and ADJ run (>year 1900) (black) in the transient period and the hosing experiment ADJ-H (gray), with the vertical gray dashed lines representing the change time from CTL to GRS and from GRS to ADJ. The 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 band. (c) The 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). It also shows the freshwater transports of the tropical restoring run (TRS, year 1000–1300) in dashed lines (M_{ovS} in blue dash, M_{ovN} in red dash, and ΔM_{ov} in black dash).

b. *The flux-adjustment experiment*

Given the freshwater import from the north (M_{ovN}) in the model is largely consistent with the observations (Table 1), CCSM3 fails mainly in simulating a significant freshwater export in the south (M_{ovS}), which then biases the model to a monostable AMOC. Therefore, we speculate, if the model could be altered such that M_{ovS} becomes a significant export, the AMOC should also become bistable, more consistent with the observation. One way to reverse the AMOC freshwater to export ($M_{ovS} < 0$) is to increase the gyre import ($M_{azS} > 0$), as implied in the freshwater budget (2) (assuming little change in the net evaporation and freshwater transport in the north). This approach has been tested successfully by imposing a west-saltier–east-fresher dipole of surface freshwater flux anomaly in the South Atlantic in an EMIC (de Vries and Weber 2005) and in a CGCM (Liu and Liu 2013, 2014). This artificial dipole flux anomaly, however, is useful mostly for theoretical purposes (e.g., Cimatoribus et al. 2012) since it is of little relevance to the real world.

Here, we go one step further to understand the key aspect of the model bias (from the real world) that leads to the biased freshwater transport and, in turn, the AMOC stability. Current CGCMs are constructed in two steps: the atmospheric and oceanic component models are first tuned separately under the observed surface climate forcing and then the two component models are coupled. Usually, each component model is able to achieve realistic climate before coupling, but significant climate bias (or the so-called climate drift) emerges after they are coupled. Therefore, we hypothesize that the bias in the freshwater transport in the ocean, and in turn the AMOC stability, is largely caused by the bias in the model surface climate. To test this hypothesis, we use the classical flux adjustment approach, in which the surface model climatology is adjusted toward the observation using a seasonal cycle of anomalous flux (Manabe and Stouffer 1988; Sausen et al. 1988). Starting from year 1001 of CTL, a global restoring run (GRS) is integrated for 1000 years with the surface temperature (SST) and surface salinity (SSS) restored (with a time of 10 days) toward their respective observed climatological seasonal cycle over the globe (Levitus et al. 1998; Steele et al. 2001). Then, the seasonal cycles of the restoring terms in the SST and SSS equations are diagnosed from the last 100 yr of the restoring run and are converted to anomalous heat and freshwater (or virtual salt) fluxes as flux adjustments. The flux adjustment run (ADJ) is then started from year 1901 of the restoring run and integrated for 700 years (years 1901–2600), with the restoring terms replaced by the flux adjustments. In spite of the flux adjustment, the transport of the AMOC remains little changed (Fig. 1b) because the AMOC intensity seems to be determined mainly by oceanic temperature, which is not changed significantly by the imposed small anomalous heat flux (relative to the mean, not shown). In contrast, the freshwater transport of the AMOC is altered significantly. Figure 1c displays the AMOC freshwater transports at the southern and northern boundaries as well as the net transport in the transition period from CTL to GRS and eventually to ADJ. The correction of surface climate bias in the restoring run (at year 1000) is a dramatic increase in the freshwater export at the southern boundary (M_{ovS} from ~ -0.01 Sv in CTL to ~ -0.14 Sv in GRS to ~ -0.17 Sv in ADJ). The freshwater import from the northern boundary is reduced, only modestly (M_{ovN} from ~ -0.12 Sv in CTL to ~ -0.09 Sv in GRS to ~ -0.06 Sv in ADJ). As a result, the net transport reverses sign from a convergence ($\Delta M_{ov} > 0$) to divergence ($\Delta M_{ov} < 0$) (Table 1), implying a change of the monostable AMOC in CTL to a bistable AMOC in ADJ. The bistable AMOC in ADJ is further confirmed in a hosing experiment (ADJ-H) parallel to CTL-H. It is seen that the AMOC transport remains in the collapsed state even 1100 yr after termination of the freshwater pulse (Fig. 1b), in contrast to CTL-H (Fig. 1a). These experiments demonstrate that the surface climate bias is the key element that leads to the AMOC import of freshwater M_{ovS} and hence a monostable AMOC in CTL. This experiment, and some previous experiments in CCSM3 (Liu and Liu 2013, 2014), are the first set of experiments in a CGCM that are successful in using the bistability indicator predictively.

c. *The role of the SSS bias*

The biased AMOC freshwater transport M_{ovS} is likely caused mainly by the SSS biases in the tropical and northern North Atlantic since the AMOC has changed little from CTL to ADJ in its transport (Fig. 1b) and the overturning flow field (Fig. 5a). As in most current CGCMs without flux adjustment (Mechoso et al. 1995; Davey et al. 2002; Lin 2007), the CCSM3 suffers from the tropical bias associated with a double ITCZ in the Atlantic–eastern Pacific sector (Collins et al. 2006; Yeager et al. 2006). In the Atlantic–eastern Pacific sector, present observations show an annual mean maximum SST and the corresponding ITCZ north of the equator. This equatorially asymmetric climate is caused by positive ocean–atmosphere thermodynamic feedbacks and the asymmetric continent distribution (Philander et al. 1996). However, in CCSM3 CTL, the maximum SST and the ITCZ straddle across the equator, causing a double ITCZ and, in turn, excessive rainfall south of the equator. This can be seen clearly in the zonal mean Atlantic surface climate bias from the observation (Figs. 2a–c, red solid lines). The CTL exhibits a dipole bias in precipitation around the equator, with an excessive (deficient) rainfall south (north) of the equator (Fig. 2a, red solid). This dipole bias is caused by the excessive shift of the ITCZ rain belt into the Southern

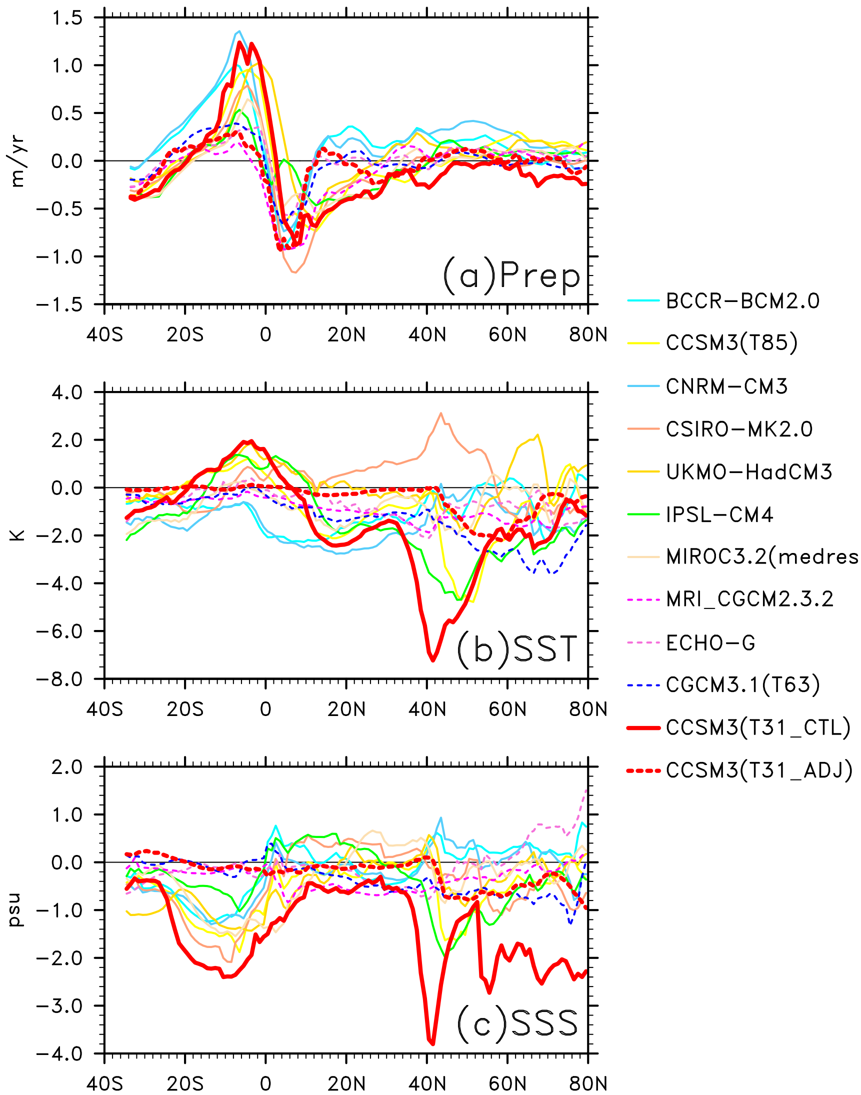

FIG. 2. The latitudinal distribution of zonal mean biases of (a) precipitation (m yr^{-1}), (b) SST (K), and (c) SSS (psu) over the Atlantic in IPCC AR4 models and the CCSM3 T31: BCCR-BCM2.0 (cyan), CCSM3 (T85) (yellow), CNRM-CM3 (light sky blue), CSIRO MK2.0 (light salmon), HadCM3 (gold), IPSL-CM4 (green), MIROC3.2(medres) (wheat), MRI-CGCM2.3.2 (magenta), ECHO-G (orchid), CGCM3.1 (T63) (blue), CCSM3 T31 CTL (heavy red solid), and CCSM3 T31 ADJ (heavy red dash). The biases are calculated as the difference between the model and the observation: results from all models with flux adjustment (dashed lines); the others (solid lines). The observation datasets are *World Ocean Atlas* (*WOA*) and Polar Science Center Hydrographic Climatology (PHC) (Levitus et al. 1998; Steele et al. 2001) for SST and SSS and the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) (Xie and Arkin 1997) for precipitation. In the figure, variables from IPCC AR4 models are calculated as the twentieth-century means. Variables from CCSM3 T31 CTL and CCSM3 T31ADJ are calculated from the 100-yr average of a steady state (year 1–100 in CTL and years 2301–2400 in ADJ).

Hemisphere, causing excessive rainfall south of the equator of up to ~1 m yr^{-1}. The rainfall bias is forced by a southward anomalous SST gradient, with the warming (cooling) south (north) of the equator (Fig. 2b, red solid). One direct consequence of the excessive rainfall is a fresh bias in the SSS south of the equator (Fig. 2c, red solid). This surface fresh bias is advected southward by the surface Ekman flow and is then subducted into the thermocline and flows northward in the subtropical Atlantic, before recirculated back in the southward North

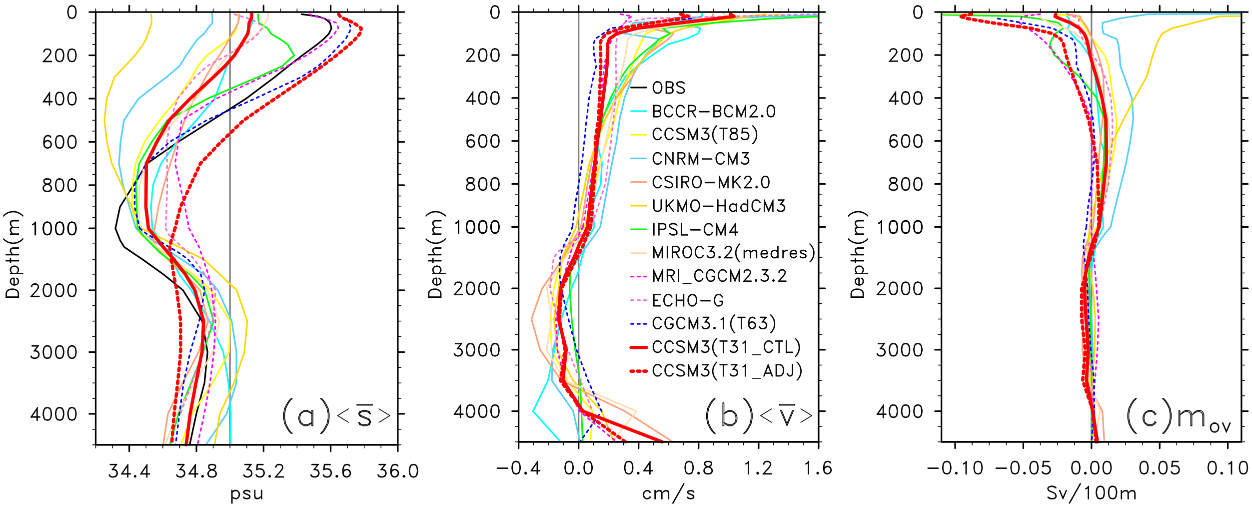

FIG. 3. Vertical profiles of (a) zonal mean salinity (psu), (b) zonal mean meridional velocity (cm s^{-1}), and (c) meridional overturning freshwater transport (m_{ov}) [nonvertically integrated M_{ov} transport (Sv (100 m)$^{-1}$] along 34°S across the Atlantic in the IPCC AR4 models and CCSM3 T31. The observed salinity is also shown in (a) (black curve). The reference salinity in (c) is 35 psu for all the models. The data used are as in Fig. 2.

Brazil Current (Liu et al. 1994; Inui et al. 2002), finally generating a fresh bias in the upper (500 m) ocean (Fig. 3a, red solid compared with the observation in black). The upper-ocean fresh bias leads to an anomalous import of freshwater in the thermocline (Fig. 3c, red solid) via the northward limb of the AMOC (Fig. 3b, red solid). With the flux adjustment in ADJ, the surface bias is reduced significantly in SST (Fig. 2b, red dash) and rainfall (Fig. 2a, red dash), as well as SSS (Fig. 2c, red dash), and in turn the upper-ocean (500 m) salinity (Fig. 3a, red dash).

We now examine how ADJ changes the SSS and eventually reverses M_{ovS} to export. Figure 4 shows the surface freshwater fluxes of ADJ and its difference from CTL. Immediately south of the equator (10°S–0°), the difference between ADJ and CTL (Fig. 4b) is a strongly negative surface freshwater flux (black solid), associated mainly with a decreased rainfall there (blue solid), which dominates over the freshwater flux adjustment (red dash). This suggests that the SST bias and in turn the forced atmospheric rainfall bias in the ITCZ contributes significantly to the SSS bias there. The changed freshwater flux in ADJ reduces the fresh bias in the South Atlantic significantly (red dash relative to red solid, Fig. 2c), creating a saline anomaly on the surface and then down to the upper ocean (upper 1000 m) in the South Atlantic through thermocline ventilation, as seen in the zonal mean salinity changes in the Atlantic (Fig. 5b). It is this increased upper-ocean salinity that is transported northward by the upper limb of the mean AMOC (upper 1000 m, Fig. 5b), corresponding to an equivalent freshwater export to the south and therefore enhanced M_{ovS} export in ADJ (relative to CTL). This relationship between the tropical bias and the M_{ovS} export should be robust because it is determined by the thermocline circulation of tropical–extratropical exchange (Liu et al. 1994) in the South Atlantic (Inui et al. 2002).

In addition to the saline upper ocean associated with the correction of the tropical bias, the M_{ovS} export in ADJ is also enhanced by a deep fresh tongue in the outgoing limb of the NADW (Fig. 5b). The cause of this deep fresh tongue seems to be contributed partly by the positive freshwater flux anomaly in high-latitude deep convection region in the North Atlantic (black line between 50° and 60°N in Fig. 4b), which is caused almost entirely by the flux adjustment in this specific model (red dash in Fig. 4b).

At the southern boundary, the salinity anomaly further shows that the saltier thermocline water is transported northward by the interior circulation in the eastern basin, while the fresher NADW is transported southward mainly in the deep western boundary current (Fig. 6b). Therefore, both reductions of the tropical fresh bias and high-latitude saline bias enhance the M_{ovS} export significantly, reversing the net AMOC freshwater transport from convergence to divergence ($\Delta M_{ov} < 0$), establishing a bistable AMOC in ADJ.

A further sensitivity experiment suggests that the tropical SSS bias contributes to about half of the enhanced M_{ovS} export. We performed a tropical restoring run (TRS) with the restoring applied only to the SSS in the tropical Atlantic between 15°S and 15°N. In comparison to the global restoring run GRS (whose model climate is similar to ADJ), the TRS generates a similar saline water in the upper South Atlantic (relative to CTL) but fails to produce the deep fresh tongue in the

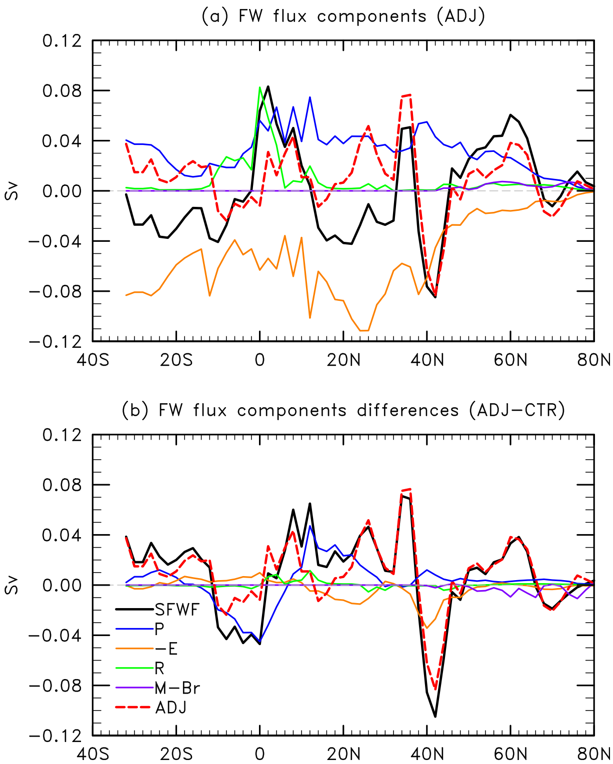

FIG. 4. Latitudinal variation of the components of surface freshwater fluxes zonally averaged across the Atlantic in CCSM3 T31 experiments for (a) the flux adjustment model (ADJ) and (b) the difference between ADJ and the control (CTL). The line plots are for total surface freshwater flux (black), precipitation (blue), river runoff (green), minus evaporation (orange), and flux adjustment (red dashed). The flux (Sv) is integrated for each 2° bin of latitude belt.

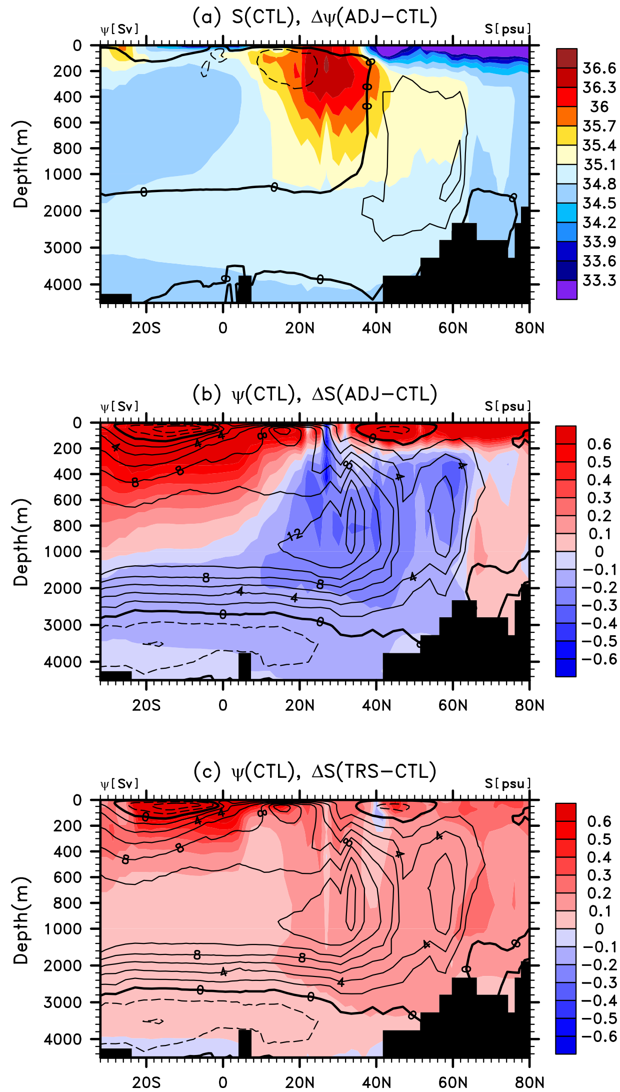

FIG. 5. Zonal mean mass transport streamfunction and salinity in the Atlantic in CCSM3 T31 experiments: (a) salinity in the control CTL (shading, psu) and the difference of streamfunction between ADJ and CTL [ADJ − CTL, contour interval (CI) = 1 Sv]; (b) streamfunction in CTL (CI = 2 Sv), and the difference of salinity between ADJ and CTL (ADJ − CTL) (shading, psu); and (c) as in (b), but the salinity difference is between the tropical restoring run (TRS) and CTL (TRS − CTL).

NADW (Fig. 5c, also see the southern section in Fig. 6c). This saline thermocline water, as well as the associated enhancement of the freshwater export, is indeed contributed to by the tropical SSS bias. More quantitatively, the freshwater export at the southern boundary is increased from $M_{ovS} \sim -0.01$ Sv to -0.07 Sv (Fig. 1, blue dash; years 1000–1320), accounting for about 50% of the M_{ovS} in GRS ($\sim$0.015 Sv at the end time of the TRS, year 1300). However, the net transport ΔM_{ov} is still a convergence of $\sim$0.06 Sv (Fig. 1c, black dash), so the AMOC is likely to remain monostable in the corresponding flux-adjustment run. This experiment suggests that, although important, tropical bias alone is insufficient to change the monostable AMOC in CCSM3.

In short, our CCSM3 study suggests that model SSS biases in both the tropics and northern North Atlantic reduce the AMOC export of freshwater at the southern boundary and then bias the model toward a monostable AMOC. In particular, the tropical bias freshens the thermocline water in the South Atlantic, reducing the equivalent freshwater export at the southern boundary in the upper limb of the AMOC.

4. The AMOC bistability in IPCC AR4 CGCMs

The CCSM3 study above has implications to other CGCMs. This is because at least a major part of the SSS bias, the tropical bias, is a common feature in many state-of-the-art CGCMs without flux adjustment (Mechoso et al. 1995; Davey et al. 2002; Lin 2007). We therefore hypothesize that the AMOC will be biased toward monostable in these CGCMs because the surface fresh bias in the tropical South Atlantic reduces the thermocline salinity and therefore the overturning freshwater export in

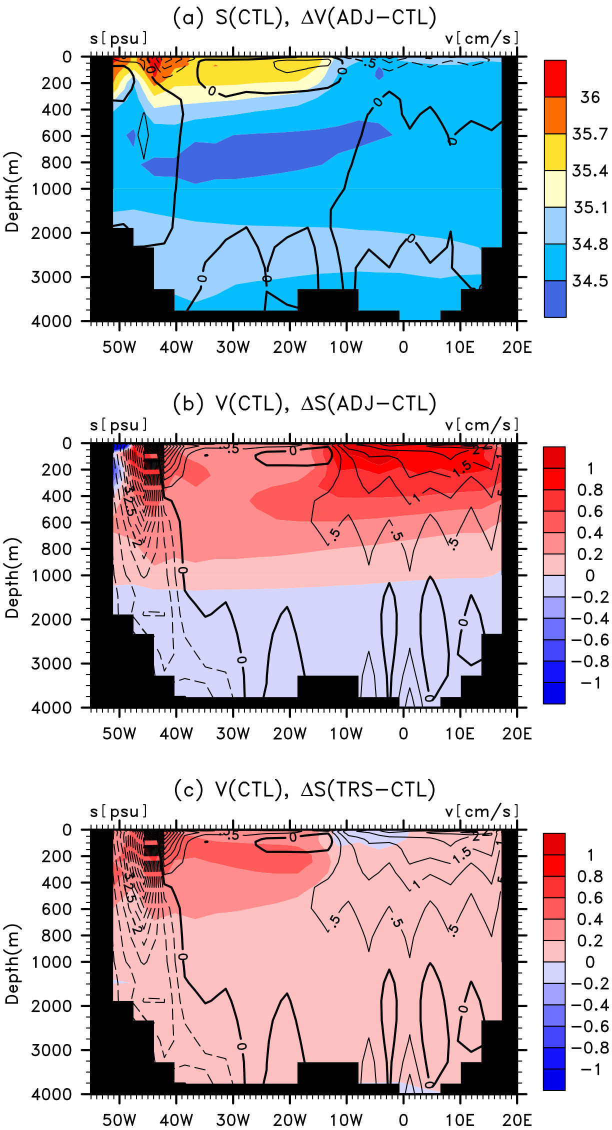

FIG. 6. Meridional velocity and salinity along the zonal section 34°S in the South Atlantic in CCSM3(T31) experiments: (a) salinity in CTL (shading, psu) and meridional velocity difference between ADJ and CTL (CI = 1 cm s^{-1}); (b) meridional velocity in CTL (CI = 1 cm s^{-1}) and salinity difference between ADJ and CTL (shading); and (c) as in (b), but for the salinity difference between TRS and CTL.

the South Atlantic M_{ovS}. If this hypothesis is correct, we further hypothesize that this bias on the M_{ovS} export will be reduced if surface flux adjustment is applied to these models, as in the case of CCSM3 studied above.

We test our hypotheses by analyzing the state-of-the-art CGCMs in the IPCC AR4 (Meehl et al. 2007). These CGCMs, now including our CTL and ADJ, consist of two groups of models (Table 1): the first group (eight models including CTL) are not subject to flux adjustment, while the second group (four models including ADJ) are subject to flux adjustment. In Fig. 2, in addition to CCSM3 CTL, we also show the zonal mean surface climate bias across the Atlantic Ocean in seven CGCMs that are not subject to flux adjustment (solid lines). All these models exhibit the same tropical bias as the CCSM3 CTL over the tropical–subtropical North and South Atlantic Ocean: a southward gradient of anomalous SST (warming south/cooling north) (Fig. 2b) forces an excessive southward shift of the ITCZ and the associated dipole bias in precipitation (Fig. 2a), causing excessive rainfall and, in turn the surface fresh bias, south of the equator (Fig. 2c). In comparison, the second group of four CGCMs with flux adjustment (dash lines), as discussed for CCSM3 ADJ, shows a much reduced bias in rainfall (Fig. 2a), SST (Fig. 2b), and SSS (Fig. 2c) in the tropical–subtropical Atlantic. In spite of the systematic improvement in the tropics in the flux adjustment models, there is no systematic improvement of the surface biases in the northern North Atlantic in these models relative to those nonflux adjustment models (Figs. 2a–c) (except for perhaps CCSM3 CTL, which seems to exhibit a too large SSS bias in the northern North Atlantic, Fig. 2c). This lack of improvement in the high latitudes may have resulted in the lack of systematic improvement of the salinity at depth along the southern boundary, in AAIW and NADW (Fig. 3a). Nevertheless, the systematic improvement of the tropical bias and the associated upper-ocean salinity is still expected to enhance the freshwater export to the south (M_{ovS}) and, in turn, the divergence of the freshwater transport ΔM_{ov} associated with the AMOC.

This expectation is consistent with the freshwater transport calculation (Table 1). For the first group of models that have no flux adjustment, the M_{ovS} ranges from nearly zero (−0.03 Sv) to strongly positive (0.36 Sv) with the ensemble mean as a weak import (0.087 Sv). This ensemble mean M_{ovS}, when combined with a freshwater import from the Arctic (ensemble mean M_{ovN} = −0.156 Sv), leads to a strong convergence ΔM_{ov} = 0.24 Sv, implying a monostable AMOC in the ensemble mean sense. Indeed, each individual model shows a significant convergence (all $\Delta M_{ov} > 0.1$ Sv) and therefore is monostable. In contrast, for the second group of models with flux adjustment, each model shows a southern boundary export (except ECHO-G showing a weak import), with the ensemble mean M_{ovS} reversed to a weak export of −0.081 Sv. This export largely cancels the import from the north M_{ovN} of −0.078 Sv and leads to a nearly nondivergent ΔM_{ov}, implying a neutral AMOC in the ensemble mean sense. Individually, two models show divergence ($\Delta M_{ov} < 0$) and are likely bistable, while the other two are only marginally convergent ($\Delta M_{ov} <$ 0.08 Sv) and are therefore near neutral.

Admittedly, the comparison between the models without (group 1) and with (group 2) flux adjustment does not present a direct estimation of the impact of flux adjustment because they are not compared on the same models [except for CCSM3(T31)]. Nevertheless, it is clear that, overall, the models with flux adjustment are systematically more divergent than those without flux adjustment and are therefore more likely to be bistable. Since the major common bias in the models without flux adjustment is the tropical bias, we suggest that the tropical bias plays a significant role in overstabilizing the CGCMs toward a monostable AMOC. A more quantitative estimation of the contribution of the tropical bias, however, requires further sensitivity experiments and therefore is beyond the scope of this study.

It should also be pointed out that there is no guarantee that a flux adjustment will change the bistability the AMOC completely for all CGCMs. For example, a recent study of the Hadley Centre Global Environment Model (HadGEM) in experiments without and with flux adjustment shows that the AMOC remains monostable even in the model with global flux adjustment (Jackson 2013). Nevertheless, in the model with flux adjustment, salinity is indeed increased significantly in the South Atlantic thermocline, as in other flux adjustment models. Furthermore, the AMOC takes a much longer time to return to the previous equilibrium state than it does in the model without flux adjustment, indicating an enhanced positive feedback associated with salinity advection in the former. This is consistent with the role of the correction of the tropical bias in reducing the freshwater import and, in turn, the AMOC monostability. Furthermore, tropical bias is unlikely the only cause of the biased freshwater transport. Indeed, the salinity profiles along the southern boundary (Fig. 3a) indicate that all CGCMs, regardless of flux adjustment, fail to simulate the salinity minimum in the very fresh Antarctic Intermediate Water (AAIW), because of, perhaps, the too diffusive nature of the current ocean models. Further studies are still needed.

5. Summary and discussion

This paper is motivated by the question why the Atlantic meridional overturning circulation appears to be biased toward a monostable state in current CGCMs. To establish a benchmark for models, we first explored the question what the AMOC bistability state is for the real world. We reviewed the available studies regarding the proxy records of the past climate evolution and the bistability indicator calculated using the present observations. These observational studies suggest that the AMOC in the real world is likely to be in or near a bistable state in the past and at present. To explicitly understand the potential source of the model bias, we next investigated the AMOC bistability in the NCAR CCSM3 with further analysis and sensitivity experiments. Our experiments show that the surface climate bias distorts the subsurface salinity distribution such that the AMOC freshwater export is diminished or even reversed, leading to a model biased toward a monostable AMOC. In particular, the tropical bias associated with the double ITCZ produces excessive freshwater in the South Atlantic thermocline, reducing the freshwater export through the upper limb of the AMOC. Finally, to substantiate the conclusion from the CCSM3 study, we analyzed the AMOC bistability indicator in two groups of state-of-the-art IPCC AR4 CGCMs, one without flux adjustment and the other with flux adjustment. Our analysis suggests that the group of CGCMs without flux adjustment exhibit a significant fresh bias in the upper ocean of the South Atlantic owing to the common tropical bias; this fresh bias reduces the freshwater export by the AMOC and then biases the models toward a monostable AMOC. In the group with flux adjustment, flux adjustment tends to reduce this surface fresh bias and, in turn, the freshwater export by the AMOC, restoring the model toward a bistable AMOC. In particular, since the tropical bias is the most robust bias common to all current CGCMs, we suggest that the tropical bias is a major error source that biases current CGCMs toward a monostable AMOC.

a. *Relation to previous work*

Previous studies have recognized that CGCMs exhibit a systematic bias of the freshwater transport at the southern boundary, which may be responsible for the seemingly biased monostable AMOC in the models (Weber et al. 2007; Drijfhout et al. 2011). In a recent analysis of the upcoming IPCC AR5 models (Weaver et al. 2012), there is also a strong preference of AMOC freshwater import in the CGCMs, although some CGCMs without flux adjustment show weak export, as in AR4 (Table 1). (Unfortunately, they did not calculate the AMOC freshwater transport in the north and, in turn, the convergence indicator.) Our major contribution here is to track the cause of the biased model freshwater transport associated with the AMOC further to the surface climate bias. In particular, we propose the tropical bias as a major cause that leads to the overstabilized AMOC in a CGCM.

Our study is consistent with previous CGCM studies of the AMOC bistability. One CGCM that consistently shows a bistable AMOC is an earlier version of the Geophysical Fluid Dynamics Laboratory (GFDL) model that employs flux adjustment (Manabe and Stouffer 1988;

Yin and Stouffer 2007). In a later version of the GFDL model that is not subject to flux adjustment, the model only possesses a monostable AMOC (Yin and Stouffer 2007). Although air–sea interaction has been suggested as a possible explanation for the different stability behaviors between the two models, our study here offers another explanation: it is the flux adjustment in the early version of the model that corrects the salinity bias and then leads to the bistable AMOC. The bistability of AMOC has been tested in two CGCMs in long hysteresis experiments, CCSM3 (Liu et al. 2009; Hu et al. 2012) and Fast Met Office/UK Universities Simulator (FAMOUS; Hawkins et al. 2011a,b). The FAMOUS model does show a bistable AMOC when the model climatology is altered by a finite and persistent freshwater forcing over the North Atlantic and its bistability behavior is largely consistent with the freshwater transport at the southern boundary M_{ovS}. In contrast, the CCSM3 shows a net convergence of freshwater transport associated with the AMOC (ΔM_{ov}), consistent with its monostable behavior that has been tested thoroughly in sensitivity experiments (Liu and Liu 2013, 2014; Liu et al. 2013).

b. Comparison with EMICs

In a systematic study of the bistability of AMOC in earth system models of intermediate complexity (EMICs), Rahmstorf et al. (2005) performed long hysteresis experiments in 11 EMICs, all forced by freshwater forcing over the North Atlantic. All 11 EMICs show strong hysteresis and therefore bistable AMOC, giving an impression that AMOCs tend to be bistable in EMICs. This impression is reinforced by many EMIC studies on abrupt climate changes (e.g., Ganopolski and Rahmstorf 2001; Weaver et al. 2003) in which abrupt climate changes can be induced by with the bistable AMOC under a gradual change of meltwater forcing. In comparison, AMOC seems to be monostable in many CGCMs because the AMOC tends to recover after the termination of a hosing pulse (e.g., Stouffer et al. 2006) and therefore is unable to simulate abrupt climate changes during deglaciation under a gradual climate forcing (e.g., Liu et al. 2009).

Our study leads to one speculation as to why AMOC tends to be bistable more in EMICs than in CGCMs without flux adjustment. In these nonflux-corrected CGCMs, the model climatology is generated through coupled processes internally and exhibits significant surface climate biases—notably, the common tropical bias, as discussed in section 4. In contrast, an EMIC, owing to its simplified physical processes, usually tunes its present-day climatology with certain type of "flux adjustment", explicitly or implicitly, to best resemble the observations. This adjustment might have reduced the surface climate bias, including the tropical bias, significantly, which then leads to the reduction of the biases in the SSS, the freshwater export, and eventually the bistability of the AMOC.

It should be pointed out that there is another reason why the AMOC appears to be bistable more in EMICs than in CGCMs. In most CGCM studies so far, owing to the constraint of computational resources, the AMOC bistability has not been tested systematically in truly equilibrium experiments, or in slow hysteresis experiments as in EMICs (Rahmstorf et al. 2005). Rather, the AMOC bistability has been judged only from a few sensitivity experiments perturbed by short hosing pulses on the default model climatology (e.g., Stouffer et al. 2006). Since a monostable AMOC at the default climatology state does not exclude a bistable AMOC at other states, this type of short hosing pulse experiments do not explore the bistability of the AMOC as exhaustively as the hysteresis experiments. Indeed, in the most thoroughly tested CGCM, FAMOUS, the bistable AMOC exists only for climate states forced by a finite and persistent freshwater forcing in the North Atlantic (exceeding ~0.1 Sv). In the absence of this persistent freshwater forcing, the AMOC remains monostable (Hawkins et al. 2011a,b). In some sense, this persistent freshwater forcing can be considered as a flux adjustment to the climatological freshwater forcing, although the target climate of the adjustment is an idealized state instead of a realistic world. It is not surprising that a persistent freshwater forcing can change the mean climate and in turn AMOC stability, as shown in many previous studies (e.g., Cimatoribus et al. 2012; Liu and Liu 2014). Similar behavior can also be found in some EMICs. Indeed, although all 11 EMICs in Rahmstorf et al. (2005) exhibit strong hysteresis behavior, 4 of them [the Massachusetts Institute of Technology–University of Washington model (MIT–UWash), Louvain-La-Neuve two-dimensional climate model (MoBiDic), Climate de Bilt-coupled large-scale ice–ocean model (ECBilt-Clio), and Coupled Global Ocean-Linear Drag Salt and Temperature Equation Integrator (C-GOLDSTEIN)] show hysteresis behavior only for a climate state forced by a persistent freshwater forcing over the North Atlantic (with positive freshwater flux forcing in their Fig. 2). At their default climate (with zero freshwater forcing), there is no hysteresis, implying a monostable AMOC. The recent analysis of Weaver et al. (2012) on IPCC AR5 models also shows that, in the five EMICs used in AR5, two export AMOC freshwater out of the southern boundary ($M_{ovS} < 0$), favoring a monostable AMOC.

c. The role of northern boundary transport M_{ovN}

As discussed by Liu and Liu (2013), in a CGCM of realistic setting, the freshwater transport by the overturning circulation across the northern boundary M_{ovN} is important for the correct assessment of the AMOC bistability. As recognized by Dijkstra (2007), the basinwide salinity advection feedback should be determined by the convergence of the freshwater transport by the AMOC, which depends on the transport from not only the south but also the north. It is known that the North Atlantic imports a significant amount of freshwater from the Arctic (Serreze et al. 2006). What is not clear is how much of this freshwater transport is associated with the AMOC or can be changed by a change of the AMOC. Since most of the AMOC mass transport flows southward, one may think that the freshwater exchange with the Arctic will not be influenced by a change of the AMOC. This simple thinking is flawed, especially for the setting of our realistic world. Indeed, this question involves a general question that has not been well studied: what is the response and dynamics of the freshwater transport associated with the AMOC? The freshwater transport induced by the (zonal mean) overturning circulation, as defined in Eq. (1), and the complementary horizontal gyre circulation, only represent a convenient separation, instead of a truly dynamical separation. It is known that both the AMOC and salinity distribution exhibit complex three-dimensional structure. Therefore, the freshwater transport induced by the change of the AMOC may differ from the freshwater transport by the zonal mean overturning circulation in Eq. (1). In particular, along the northern boundary, the exchange flow and the freshwater distribution are strongly controlled by the complex ocean topography, which can conceivably lead to a complex response of the circulation (e.g., the coupling between baroclinic and barotropic flows) and salinity field and eventually a change of the freshwater transport, in response to a change of the AMOC. This response can be seen, for example, in sensitivity experiments, such as hosing experiments. With a strong pulse of freshwater forcing over the North Atlantic, the AMOC collapses. This leads to the accumulation of freshwater in the upper North Atlantic, which can reduce or even reverse the import of freshwater from the Arctic (see, e.g., Figs. 3b,d,f in Liu and Liu 2014). The effect of Bering Strait provides another example illustrating the role of the freshwater transport from the northern boundary. In spite of a lack of impact on the mass transport of the AMOC, a closed Bering Strait can affect the freshwater transport into the North Atlantic, which then enhances the salinity advection positive feedback and can slow down the recovery of the AMOC in response to a freshwater forcing (Hu et al. 2012). Given the importance of the bistability of the AMOC and its close relation to the freshwater transport associated with the AMOC, it is highly desirable to study the dynamics that control the freshwater transport change associated with the AMOC. More generally, since the salinity feedback is determined ultimately by the basinwide freshwater budget, it is desirable to study the response of each major term in the freshwater budget for the North Atlantic, including the freshwater transport by the overturning circulation M_{ov} and the horizontal gyre circulation M_{az} as well as the surface flux of $E - P$ (Jackson 2013).

d. *The role of flux adjustment*

Ideally, it is always desirable to improve the coupled model climatology first, before applying the model for the study of abrupt climate changes. This approach, however, has not been practical, because some major model biases have remained the most unyielding obstacles in climate modeling: notably, the tropical bias. Given that these model biases cannot be eliminated in the foreseen future, we are faced with a practical question: Which of the two imperfect models should we trust more for climate change sensitivity experiments, the original model without flux correction but with climate bias or the model with flux adjustment but with a reduced climate bias? In general, a proper scheme of flux adjustment helps the model to improve certain aspects of the model simulations, especially those not of strongly nonlinear nature, and therefore is a useful approach (Sausen et al. 1988; Weaver and Hughes 1996; Dijkstra and Neelin 1999). However, a flux adjustment could lead to unexpected consequences, including the distortion of the bistability of the AMOC (Marotzke and Stone 1995; Marotzke 1996; Dijkstra and Neelin 1999). Therefore, one should always be cautious about the results from a flux-adjusted model. One should note, however, that those unexpected consequences of flux adjustment are often associated with poor understanding of the physical mechanism and consequence of the adjustment.

Our study above suggests that, for some CGCMs, the flux adjustment may provide a useful "bandage" for improving the capability of the CGCM to simulate abrupt climate changes associated with the AMOC. Therefore, we propose that, for climate sensitivity experiments that are relevant to the AMOC bistability, the flux-adjustment model is better suited than the original CGCM, if the latter is already diagnosed as having an AMOC freshwater convergence of the wrong sign. First, our objective of flux adjustment differs from those in previous idealized model studies. In those idealized studies (e.g., Dijkstra and Neelin 1999; Marotzke and Stone 1995; Marotzke 1996), the flux-adjustment scheme has a single purpose: that is, to correct the surface climate bias. Here, our objective of flux adjustment to the surface climate is not limited to the reduction of surface climate bias. Instead, our flux adjustment is evaluated by its correction of a freshwater transport parameter that is the key to the AMOC bistability: that is, the AMOC stability indicator associated with the freshwater transport. Because of a strong restoring forcing in the preparation of flux adjustment, it is usually trivial for a surface flux adjustment to correct the surface climate. However, it is not trivial to know if a surface flux adjustment can also correct the freshwater transport parameter and, in turn, the AMOC stability. In other words, our flux adjustment is evaluated by a more stringent set of criterion and therefore is more likely to be useful. Second, our flux adjustment and the diagnosis of AMOC bistability are based on the understanding of the physical mechanism of the AMOC bistability. Therefore, our flux adjustment corrects the bias of model AMOC bistability for the ''right'' reason. In contrast, the freshwater transport M_{ovS} could also be corrected for the ''wrong'' reason. For example, the freshwater transport M_{ovS} could be corrected using a west–east dipole of surface freshwater flux anomaly over the South Atlantic (de Vries and Weber 2005; Serreze et al. 2006; Cimatoribus et al. 2012; Liu and Liu 2013, 2014). This surface flux correction is able to switch the M_{ovS} to an export by increasing the import of the gyre transport M_{azS} to maintain the freshwater budget in Eq. (2). However, this surface correction produces an erroneous upper-ocean salinity structure. Along the southern boundary, the lower thermocline water is more saline than the upper-thermocline water, and there is virtually no minimum salinity in the intermediate water, opposite to the observation (not shown). Therefore, the dipole surface is not relevant to the real-world problem, even though it can change the AMOC bistability. In other words, a realistic surface flux adjustment corrects both the surface climate and freshwater export (and, in turn, AMOC bistability), while the dipole flux only corrects the latter.

Much work is needed to better understand the AMOC bistability in the real world, in particular, for the present day. From the observational perspective, there are still large uncertainties in the estimation of the bistability indicator of the real ocean associated with the AMOC freshwater transport, because of different methods and datasets and limited quality of data. For example, although most studies suggest an AMOC export of freshwater to the south ($M_{\mathrm{ovS}} < 0$, as summarized in Table 1), the calculation of Talley (2008), as an outlier, gives an M_{ovS} value of nearly zero (0.02 Sv) at 32°S. The temporal variability of the ocean, especially in the North Brazil Current, can cause significant changes of the freshwater transport between different hydrographic observations (Bryden et al. 2011). The neglect of Agulhas rings in the sparse hydrographic sections also leads to an underestimation of the freshwater export to the south (Weijer et al. 1999). The different approaches to remove the freshwater transport in the subtropical cell by the wind-driven Ekman transport can also cause significant differences in the calculation. Small errors in estimating the relative amount of saline Benguela waters and fresher Antarctic Intermediate Water in the northward branch of the AMOC also affect the AMOC freshwater transport significantly (de Ruijter et al. 1999). Further observational studies are needed to reduce the uncertainty of the estimated freshwater import from the northern boundary. From the dynamics perspective, it remains to be further studied how well the divergence indicator Eq. (2) represents the bistability of the AMOC and how well the classical zonal mean overturning transport calculation as in Eq. (1) represents the freshwater transport by the AMOC. Finally, the AMOC stability needs to be studied in improved climate models. The current generation of CGCMs is still deficient in representing some key physical processes related to the AMOC and its transport of freshwater water. For example, the coarse resolution in current ocean models prevents explicit simulation of oceanic eddies. This may lead to deficient simulations of oceanic mixing and, in turn, freshwater transport processes, affecting the salinity advection feedback and in turn AMOC stability.

Acknowledgments. We thank Dr. B. Otto-Bliesner for discussion and coordination throughout the work. We thank Drs. G. Danabasoglu, N. Norton, S. Bates, K. Lindsay, and F. He for helping the model setup. We also thank Drs. I. Eisenman, L. Talley, and R. Keeling for discussions. Comments from two anonymous reviewers and Dr. A. Gnanadesikan have improved the paper significantly. This works is supported by NSF, NSFC41130105, MOST 2012CB955203, and DOE. The computation is carried out at Oak Ridge National Lab and the NCAR supercomputing facility.

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