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Review

A commentary on the Atlantic meridional overturning circulation stability in climate models

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A B S T R A C T

The stability of the Atlantic meridional overturning circulation (AMOC) in ocean models depends quite strongly on the model formulation, especially the vertical mixing, and whether it is coupled to an atmosphere model. A hysteresis loop in AMOC strength with respect to freshwater forcing has been found in several intermediate complexity climate models and in one fully coupled climate model that has very coarse resolution. Over 40% of modern climate models are in a bistable AMOC state according to the very frequently used simple stability criterion which is based solely on the sign of the AMOC freshwater transport across 33° S. In a recent freshwater hosing experiment in a climate model with an eddy-permitting ocean component, the change in the gyre freshwater transport across 33° S is larger than the AMOC freshwater transport change. This casts very strong doubt on the usefulness of this simple AMOC stability criterion. If a climate model uses large surface flux adjustments, then these adjustments can interfere with the atmosphere–ocean feedbacks, and strongly change the AMOC stability properties. AMOC can be shut off for many hundreds of years in modern fully coupled climate models if the hosing or carbon dioxide forcing is strong enough. However, in one climate model the AMOC recovers after between 1000 and 1400 years. Recent 1% increasing carbon dioxide runs and RCP8.5 future scenario runs have shown that the AMOC reduction is smaller using an eddy-resolving ocean component than in the comparable standard 1° ocean climate models.

1. Introduction

Buckley and Marshall (2016) have recently written a comprehensive review of Atlantic meridional overturning circulation (AMOC) observations, inferences, and mechanisms. They include a brief discussion of AMOC stability in climate models, but a more in depth look is instructive in light of some recent modeling results and assumptions. The first assumption is that a climate model's AMOC is monostable or bistable depending solely on the sign of the AMOC freshwater transport across 33° S. Monostable means that the only stable AMOC configuration is a strong overturning in the North Atlantic which extends into the South Atlantic, often called an AMOC 'on' state, as in today's climate. Bistable means that in addition to this AMOC 'on' state, there is a stable AMOC 'off' state, in which there is a deep reverse cell in the tropical Atlantic and very weak overturning in the North Atlantic midlatitudes. Fig. 1 a and b shows AMOC 'on' states from control runs of the Geophysical Fluid Dynamics Laboratory (GFDL) R30 and CM2.1 climate models, and Fig. 1c and d shows typical AMOC 'off' states after 81–100 years of a strong North Atlantic water hosing experiment. This figure is from Yin and Stouffer (2007), which will be discussed at length in Section 5. The second assumption is the use of climate models with a large surface freshwater flux adjustment to study AMOC stability, under the assumption that the flux adjustment is justified because it improves the upper ocean salinity distribution in the Atlantic. The third assumption is that the AMOC 'on' state is too stable in nearly all modern climate models where the atmosphere and ocean resolution is about 1°. This paper is not a comprehensive review of the hundreds of papers on this subject, but concentrates on papers that highlight important results with respect to the three assumptions outlined above.

2. Stability in ocean alone models

The idea that the AMOC could have more than one stable state started with the paper by Stommel (1961). He used a simple two vessel experiment, and showed that, when the salinity restoring is very weak, there can be two stable states. The first has temperature mainly influencing the density and flow from one vessel to the other, whereas the second has reverse flow between the vessels and salinity mainly influencing the density.

The first study of AMOC stability in a numerical ocean model was by Bryan (1986). He used a 3.75° latitude × 4.5° longitude ocean sector model that spanned from pole to pole, and boundary conditions of

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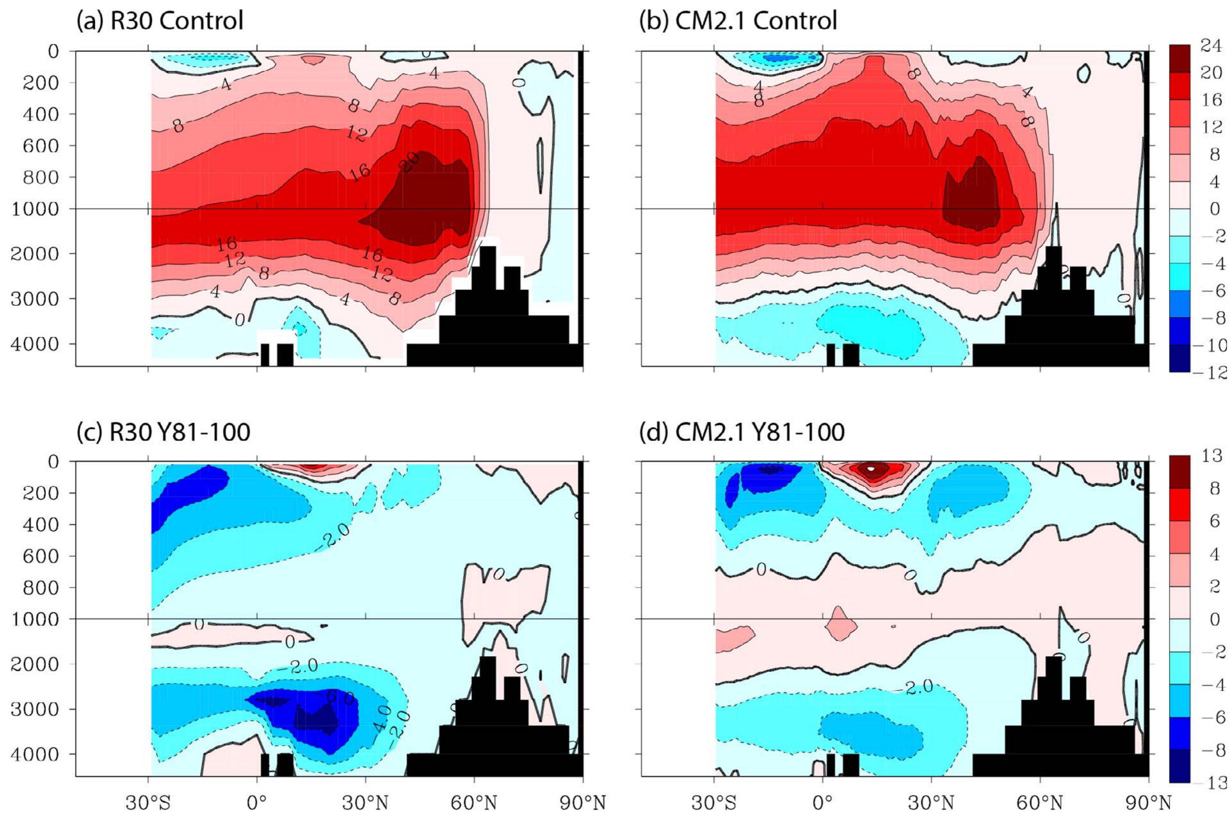

Fig. 1. Fig. 4a–d from Yin and Stouffer (2007). Atlantic meridional overturning streamfunction from control runs of a) the R30 and b) CM2.1 models, and after years 81–100 of a strong North Atlantic water hosing experiment in c) the R30 and d) CM2.1 models.

restoring on sea surface temperature (SST) but a virtual salt flux that is symmetric about the equator and independent of the ocean surface salinity. He showed that the model has three equilibrium solutions; one with the overturning symmetric about the equator, and two mirror-image pole to pole circulations with strong overturning in the different hemispheres. The model went to these different stable states depending on the initial state of the surface salinity.

The Bryan (1986) paper sparked many stability studies in ocean alone models, such as Weaver and Sarachik (1991). This paper nicely shows that an AMOC state which is stable using boundary conditions of restoring to both SST and surface salinity can become unstable under mixed boundary conditions, which are restoring to SST but a fixed freshwater flux. Prange et al. (2003) used a coarse resolution $3.5^\circ \times 3.5^\circ$ model forced by mixed boundary conditions. They used five different profiles of vertical mixing, and showed that the AMOC strength had a strong dependence on the vertical diffusivity chosen. They also changed the advection scheme in their model, and found that this could substantially change the AMOC stability properties. Sijp and England (2006) used a $1.8^\circ \times 3.6^\circ$ ocean model, and showed that changing the vertical mixing only in the Atlantic Ocean or elsewhere could quite strongly affect the AMOC stability properties.

More recently, Griffies et al. (2009) and Danabasoglu et al. (2014) found that the AMOC representation varied strongly in many different ocean components of climate models which were all forced using the same mixed boundary conditions. In order to prevent a drift in the salinity distribution, the forcing includes a weak restoring to surface salinity, and the AMOC maximum value depends quite strongly on the strength of this restoring term. The conclusion from this section is that the AMOC strength, and its stability, depend rather strongly on the ocean formulation, especially the vertical mixing, and the boundary conditions used to force ocean alone models.

3. Stability in coupled models

Manabe and Stouffer (1988) were the first to find a bistable AMOC state in a coupled model. They used the same $3.75^\circ \times 4.5^\circ$ global ocean model as Bryan (1986) coupled to a coarse resolution atmosphere component and large flux adjustments. The model has a stable AMOC 'on' state with strong overturning in the North Atlantic. However, if they used a different initial condition, then the model could maintain a stable AMOC 'off' steady state with very weak overturning in the North Atlantic. In a subsequent study using the same model, Manabe and Stouffer (1999) showed that the AMOC 'off' state could be reached in a hosing experiment by imposing a large freshwater flux between 50° N–70° N in the Atlantic and that the AMOC 'off' state remained stable for the remaining 7000 year run. They then repeated this run with a larger value of the upper ocean vertical mixing and found that the AMOC collapsed to an 'off' state during the hosing, but then slowly recovered to its original 'on' state over the next 1500 years. This again shows that AMOC stability properties strongly depend on ocean vertical mixing values.

Manabe and Stouffer (1993) also used their same model when the carbon dioxide (CO2) level was increased by 1% per year for 140 years and then held constant at 4 times the original value. The North Atlantic AMOC collapsed to an 'off' state, which persisted for the rest of the 500 year run. However, Stouffer and Manabe (2003) document that they later continued this run beyond year 500. After about 1500 years the North Atlantic AMOC did recover to its original 'on' state, which persisted for the remainder of the 5000 year run. The reason was that the warming near the surface of the North Atlantic diffused down slowly over the 1500 years, so that the upper 2–3 km of the ocean became less stratified and deep water formation started up again.

These results raise an interesting question about timescales. Is the Manabe and Stouffer model bistable because it maintains an AMOC 'off' state for 1500 years, or is it monostable because it eventually returns to a stable AMOC ‘on’ state? I will use a strict definition here, and say that this AMOC ‘off’ state is not a stable equilibrium. This model was run for several thousand years, which is long enough for the ocean to reach equilibrium. However, modern climate models are almost never run this long because of the computational expense, and the runs tend to be shorter as the model resolution becomes finer. If climate models are not run out to equilibrium, this raises a potential mismatch between AMOC stability properties in the climate models and those found in box models. By their nature, box models give the stability of equilibrium solutions, whereas nearly all climate model runs do not represent equilibrium solutions. Modern climate models are usually called bistable if they maintain an AMOC ‘off’ state for a few hundred years, but are not integrated long enough to see whether they would eventually return to a stable AMOC ‘on’ state, as in Stouffer and Manabe (2003). However, AMOC being ‘off’ for 1500 years is a very long time, and this situation can certainly drastically change the earth’s climate over that time period.

Saravanan and McWilliams (1995) coupled a two-level atmosphere with simplified physical parameterizations to a zonally averaged sector ocean thermohaline model. They compared AMOC stability in the coupled model to the stability found when the same ocean model is forced by mixed boundary conditions. They conclude, “Mixed boundary conditions may enable ocean-only models to capture the gross time-averaged features of the corresponding coupled equilibria, but they distort the stability characteristics of the equilibria.” Bjornsson et al. (1997) coupled an atmosphere energy moisture balance model to a zonally-averaged ocean thermohaline model and compared the results with those from an ocean-only model employing mixed boundary conditions. Their abstract states, “The authors conclude that due to the effects produced by the feedbacks in the coupled model, they must have serious reservations about the results concerning long-term climate variability obtained from ocean-alone models.”

Rahmstorf et al. (2005) show results from 11 coupled models of intermediate complexity, often called EMICs, which all have simplified atmosphere components, so that an EMIC does not have all the feedbacks of a full climate model. They show the AMOC hysteresis curves when freshwater was added uniformly to the latitude band 20° N–50° N across the Atlantic. For positive freshwater forcing, the AMOC will collapse and remains small when the freshwater forcing is reversed until the forcing becomes negative, forming a hysteresis loop in the AMOC strength. Fig. 2 is from Rahmstorf et al. (2005), and shows that all 11 models had a hysteresis loop in AMOC when the freshwater forcing was varied. Fig. 2 shows the width of the loop is larger in the models with a simplified ocean component than in the models which use a 3-D ocean component. In contrast to these EMICs, an AMOC hysteresis loop has only been found in one fully coupled model. This is documented in Hawkins et al. (2011), who used the FAMOUS model with very coarse resolution of 5° × 7.5° in the atmosphere and 2.5° × 3.75° in the ocean. Their Fig. 2 shows that the width of the hysteresis loop is much narrower than in the EMIC results shown in Fig. 2. Jackson et al. (2017) have very recently shown that the hysteresis loop in the FAMOUS model is even narrower if the hosing compensation is applied over the full ocean volume rather than just at the ocean surface. Note that existing computer power does not permit the large number of long integrations of a fine resolution climate model that are required to thoroughly check for hysteresis behavior.

The conclusion from this section is that the feedbacks contained in a fully coupled climate model frequently eliminate the stable AMOC ‘off’ state found in ocean alone models forced by mixed boundary conditions. These feedbacks include the effects of SST and sea ice edge location on the atmosphere circulation, which alters the surface winds and the heat and freshwater transports within the atmosphere, which in turn affect the fluxes. Sea ice strongly reduces the fluxes into the ocean and increases the surface albedo, which again changes the atmosphere circulation. Evaporation causes a latent heat flux, which links the heat and freshwater fluxes, and there is no restoring to surface salinity in a coupled model, whereas ocean alone models usually use a weak restoring. Section 3 of Griffies et al. (2009) contains a fuller discussion of this topic.

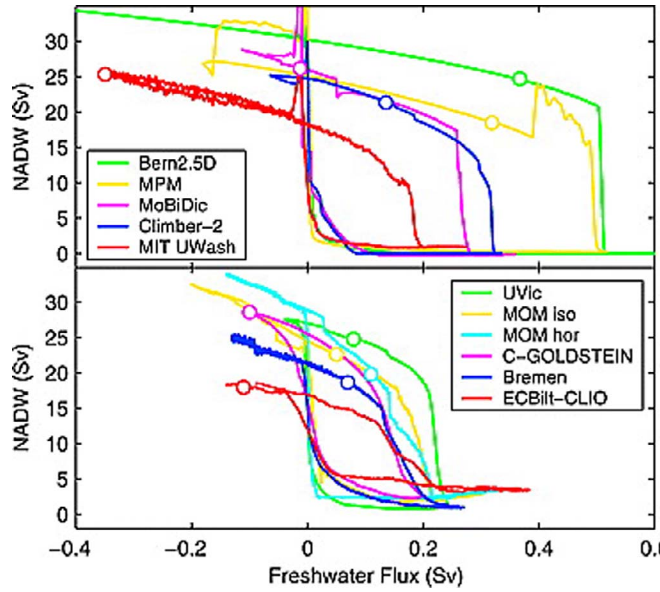

Fig. 2. Fig. 2 from Rahmstorf et al. (2005). AMOC strength versus the freshwater flux, showing the hysteresis curves from 11 EMICs. The top panel shows models with a simplified ocean component, and the lower panel shows models with a global 3-D ocean component.

4. Comments on the AMOC bistability criterion

Rahmstorf (1996) shows results from a 4-box model and an ocean model coupled to an energy balance atmosphere. He was the first to define the AMOC bistability criterion as the sign of the AMOC freshwater transport at the southern boundary of the Atlantic Ocean, F_{OT}, defined as

$$F_{OT} = -\int \bar{v} < S > dz/S_0, \tag{1}$$

where $\bar{v}$ is the zonally-integrated baroclinic meridional velocity, $< S >$ is the zonally-averaged salinity, and S_0 is a reference salinity. Rahmstorf (1996) states that, if F_{OT} is negative so that AMOC is transporting freshwater southwards at 33° S, then, “The freshwater forcing opposes the thermal driving, and equilibria with and without North Atlantic deep water formation exist.” The idea is that if F_{OT} is negative and AMOC weakens, then less freshwater is transported out of the basin, making the Atlantic fresher and AMOC weakens further; a positive feedback mechanism on AMOC strength. Conversely, if F_{OT} is positive, then a weakened AMOC will transport less freshwater into the basin making the Atlantic saltier; a negative feedback mechanism on AMOC strength. This criterion of a negative F_{OT} implying a bistable AMOC, whereas a positive F_{OT} implies a monostable AMOC, has since been very widely used to characterize AMOC stability in a whole range of models. For example, Huisman et al. (2010) state in their Abstract, “The sign of F_{OT} precisely shows whether this net anomalous freshwater transport is stabilizing or destabilizing the MOC. Therefore, it can indicate whether the MOC is in a single or multiple equilibrium regime.” Drijfhout et al. (2011) state in their Abstract, “Apart from atmospheric feedbacks, the sign of the salt flux into the Atlantic basin that is carried by the MOC determines whether the MOC is in the single or multiple equilibria regime.” Also, Deshayes et al. (2013) document ocean hindcasts using a 1/12° model, and their Conclusion starts with, “All four eddy-resolving simulations have an overturning freshwater transport negative on average, suggesting the present-day AMOC is in the bistable regime.”

In my opinion, this criterion of diagnosing only the AMOC freshwater transport can be misleading, because it neglects other freshwater transport processes, notably that by the gyre circulation. The original Rahmstorf (1996) paper points out that evaporation exceeds precipitation plus runoff over the tropical Atlantic Ocean, so that the total freshwater transport at 33° S must be positive or northwards. Table 10 in Talley (2008) suggests a value of 0.21 ± 0.04 Sv, estimated from observations. Rahmstorf's Fig. 11 shows that if F_{OT} is negative but the total is positive, then the gyre transport must be positive and larger than the AMOC transport. Rahmstorf (1996) says, "The wind-driven gyre transport is thus the largest term in the freshwater budget and balances not only the net freshwater loss from the Atlantic basin through the atmosphere, but also the additional freshwater export by the conveyor belt." However, using only the sign of F_{OT} as the stability criterion ignores changes in other processes, including the gyre freshwater transport, when AMOC weakens or strengthens. Buckley and Marshall (2016) state on page 21 that this assumption is "questionable", and some recent papers have shown that this assumption is wrong.

This has been demonstrated most clearly in a recent paper by Mecking et al. (2016). They show results from a version of HadGEM3 that uses eddy-permitting ocean resolution of 0.25° coupled to an atmosphere component with midlatitude resolution of 60 km. The salinity in the upper 350 m is reduced by 0.6 psu in the Arctic and Atlantic north of 20° N, which is an equivalent hosing of 10 Sv every year for 10 years. Fig. 3 is from Mecking et al. (2016), and shows the freshwater transport components in the Atlantic from their control run. The AMOC transport, F_{OT}, which they label M_{OV}, is negative throughout the Atlantic including 33° S and agrees well with some estimates from observations. The gyre transport, M_{AZ}, is positive at 33° S and is larger than M_{OV}, so that the total transport is positive, as required. The gyre transport is non-negligible throughout the Atlantic and dominates M_{OV} north of 45° N. Fig. 4 is also from Mecking et al. (2016), and shows the changes in M_{OV} and M_{AZ} compared to the control run averaged over years 311–410 after the hosing is stopped, during which time the AMOC remains in an 'off' state. Changes in M_{OV} and M_{AZ} are comparable in magnitude throughout the Atlantic, but at 33° S the change in M_{AZ} is over twice as large as the change in M_{OV}. Note from Fig. 4 that the change in M_{OV} is mainly due to the change in velocity, whereas the M_{AZ} change is mainly due to the change in salinity. Mecking and Drijfhout (personal communication) have shown that the changes in M_{OV} and M_{AZ} are comparable throughout the Atlantic at all times after the hosing is stopped, and confirmed that the M_{AZ} change is over twice the M_{OV} change at 33° S in years 11–110, as well as in years 311–410 shown in Fig. 4. Therefore, the assumption behind the simple stability criterion, that other freshwater transport processes, including the gyre transport, do not change when AMOC changes, is clearly incorrect.

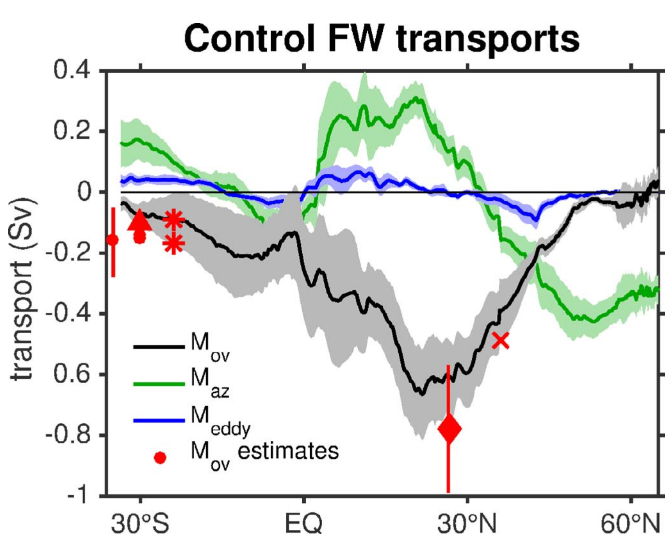

Fig. 3. Fig. 7 from Mecking et al. (2016). Control run freshwater transports: Mean and ± one standard deviation. Black – overturning M_{ov}, Green–gyre M_{az}, Blue – eddy M_{eddy}, Red – estimates from observations. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

A good number of climate models have a positive F_{OT}, and this has led some authors to write that climate models are biased in this regard. Hofmann and Rahmstorf (2009) write, "However, substantial evidence suggests that current models may be systematically too stable in that they are in a monostable regime far away from the threshold, in contrast to what observational data suggest." Liu et al. (2014) write, "In contrast, sensitivity experiments in CGCMs tend to show a monostable AMOC (Stouffer et al., 2006), indicating a model bias toward a monostable AMOC."

Tables 1 in Weaver et al. (2012), Liu et al. (2014) and Mecking et al. (2017) show that over 40% of the CMIP5 climate models have a negative F_{OT}, although most of these values are smaller than the negative value suggested by observations. Therefore, it is certainly incorrect to say that climate models show a common bias toward a stable AMOC (with a single equilibrium), as in Liu et al. (2017). Weaver et al. (2012) agree and write, "Previous criticism regarding a tendancy for models to be overly stable appears not to be the case in the CMIP5 models." They show results from climate models using the strong RCP8.5 future forcing scenario. The AMOC reduces in all models, but only collapses completely in one model which has a positive F_{OT}. In all 10 models that have a negative F_{OT} possibly indicating a bistable regime, the RCP8.5 forcing is not strong enough to collapse AMOC completely.

Liu and Liu (2013) propose that an improved AMOC bistability criterion is ΔM_{OV}, which is the difference between the overturning freshwater transport at the southern and northern boundaries of the Atlantic. Clearly the freshwater transport through Fram Strait, the Barents Sea and the Canadian Archipeligo can affect the deepwater formation regions that strongly influence North Atlantic AMOC strength much more quickly than that entering the South Atlantic. Table 1 in Liu et al. (2014) shows that in all 8 CMIP4 models studied the overturning is transporting freshwater into the Atlantic at the northern boundary, and 6 of the model transport values are quite close to the 0.16 Sv estimated from observations. Therefore, in these climate models the largest difference in freshwater transport compared to observations is in the value of F_{OT}. It would be interesting to know the value of the northern boundary overturning freshwater transport in all CMIP5 models, but as far as I am aware this has not been done. In fact, in most recent papers it is still F_{OT}, rather than ΔM_{OV}, which is quoted as the AMOC bistability criterion.

5. The dangers of using flux adjustments

Dijkstra and Neelin (1999) study a zonally-averaged ocean model of the thermohaline circulation forced by mixed boundary conditions. They use a flux adjustment procedure to maintain a realistic AMOC when parameters are changed, which represents running a coarse resolution general circulation model with too large a viscosity or diffusivity. This procedure produces an expanded multiple equilibria regime, and their Abstract states, "Furthermore, areas in parameter space exist with multiple equilibria in the flux-corrected case that have a unique state in the uncorrected case. Care should thus be used in drawing conclusions on the existence of multiple equilibria and the stability of the thermohaline circulation when a flux-correction procedure is used."

Yin and Stouffer (2007) compare a hosing experiment in two GFDL climate models. The R30 has an atmosphere resolution of 2.25° × 3.75° and an ocean resolution of 2.25° × 1.875°, and uses flux adjustments to maintain the modern day climate. The CM2.1 has an atmosphere resolution of 2° × 2.5° and an ocean resolution of 1°, and does not need any flux adjustments. The hosing is a freshwater perturbation of 1 Sv input uniformly over 50° N–70° N in the North Atlantic for 100 years.

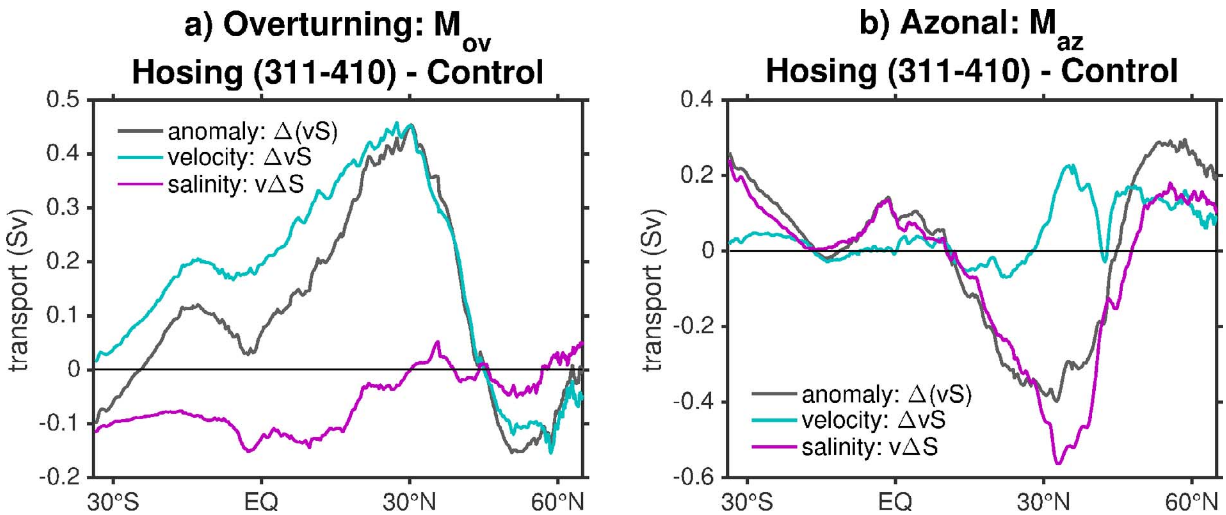

Fig. 4. Fig. 8b,d from Mecking et al. (2016). Changes in freshwater transport between years 311–410 of hosing run compared to the control run. a) Overturning transport M_{ov}, and b) Gyre transport M_{az}. The curves show the total anomaly, the contribution by the anomalous meridional velocity, and the contribution by the anomalous salinity.

The hosing is then switched off, and the models run for another 200 years. In R30 the AMOC collapses to the 'off' state shown in Fig. 1c, which persists for the remainder of the run. In contrast, in CM2.1 AMOC collapses to the 'off' state shown in Fig. 1d, but then starts to recover and the maximum AMOC has reached nearly 20 Sv by year 300, which is near the CM2.1 control run average of 24 Sv shown in Fig. 1b. Yin and Stouffer (2007) perform a comprehensive and incisive analysis of why these two models show such different results. They show that the weak reversed thermohaline circulation in the upper 1 km of the South Atlantic and the enhanced Antarctic Bottom Water cell below 3 km between the equator and 30° N shown in Fig. 1c and d are maintained over years 100–300 in R30, but they quickly reduce in strength after year 100 in the CM2.1. They conclude, "An additional factor, which we find important in our analysis of the R30 results, is the use of flux adjustments in this model. Due to the large flux adjustments employed by R30, the actual net surface water flux felt by the oceanic model is much smaller in R30 than in CM2.1. This leads to a much weaker oceanic freshwater transport and convergence necessary to stabilize the coupled system in R30. In this case, a stable off state of the THC is much more easily obtained. In contrast, the pronounced net surface water flux in CM2.1 has to be totally compensated by oceanic processes only." In other words, it is the freshwater flux adjustment in R30 that maintains the AMOC 'off' state, suggesting that AMOC is bistable, whereas CM2.1 with no flux adjustment suggests that AMOC is monostable.

Jackson (2013) shows results from the standard and flux adjusted versions of the HadCM3 climate model. The atmosphere resolution is 2.5° × 3.75°, and the ocean resolution is 1.25° × 1.25°. The rationale for the flux adjusted version is to improve the Atlantic surface salinity distribution and to change the sign of F_{OT} from positive in the standard version to negative in the flux adjusted version. Both versions are subjected to a hosing of 1 Sv for 150 years over 50° N – 70° N in the Atlantic. Fig. 5 is from Jackson (2013), and shows the AMOC in control and hosing runs from the two versions. In the standard version, AMOC quickly starts to recover after the hosing is stopped, by year 350 has recovered to stronger than in the control run, and then slowly decays towards the control run value. In contrast, in the flux adjusted version, AMOC does not start to recover until year 300, and is only just returning to the control run value by year 750. Thus, AMOC is less stable in the flux adjusted version than in the standard version. Jackson (2013) also shows similar results to the flux adjusted version when only the freshwater flux adjustment is applied.

Liu et al. (2017) show results from the standard and flux adjusted versions of the CCSM3 T31 × 3 climate model. The atmosphere resolution is 3.75° × 3.75°, and the ocean resolution is 3° × 3°. Fig. 6 is from Liu et al. (2017) and shows the AMOC in a control run and a continuation run after the CO_2 value has been instantaneously doubled. In the standard version, AMOC reduces by about 4 Sv but then slowly recovers towards the control run value. In contrast, in the flux adjusted version, AMOC collapses to 4 Sv and stays at that low value for the remainder of the 1000 year run. Thus, this shows an extreme change where AMOC changes from monostable in the standard version to bistable in the flux adjusted version.

Clearly the freshwater flux adjustment and the improved Atlantic salinity distribution occur together, so which of them is more important in affecting the AMOC behavior? The analysis of Yin and Stouffer (2007) discussed above clearly demonstrates that a large freshwater flux adjustment strongly changes AMOC stability properties in a climate model. Therefore, I think that there is much more evidence that the change in AMOC stability shown in Figs. 5 and 6 is mostly caused by the flux adjustments imposed, which strongly interfere with the stabilizing feedbacks between the atmosphere and ocean components. Therefore, my conclusion from these results using standard and flux adjusted versions of the same model is that AMOC stability conclusions from a climate model that uses large flux adjustments should be viewed with great suspicion.

6. AMOC collapse in modern climate models

Can a modern, 1° climate model simulate an AMOC collapse to a typical 'off' state, even though its bistability criterion says it is monostable? The answer is yes, if the perturbation is large enough. The CCSM4 was forced by the RCP8.5 scenario from 2005 to 2300, by which time the CO_2 level is near 1960 ppm, or about 5 times the 2005 level. The run was then extended until 2600 with the CO_2 level constant, and results are shown in Jahn and Holland (2013) and Hu et al. (2013). They show that the Arctic becomes ice free (defined as $< 10^6$ km^2) by 2300, and the Antarctic becomes ice free by 2400 because the surface temperature in these regions has increased by > 15 °C. Deep water formation has ceased in the Arctic and Nordic Seas and south of Iceland. Fig. 1 in Hu et al. (2013) shows the AMOC maximum between 40° N–80° N as a function of time, which reduces from > 20 Sv in 2005 to 3 Sv in 2200, and remains < 3 Sv for the remainder of the run to 2600.

An AMOC collapse has also been documented in Mecking et al. (2016), which was discussed in Section 4. They use a version of HadGEM3 with eddy-permitting ocean resolution of 0.25°. The hosing is very large and is the equivalent of 10 Sv every year for 10

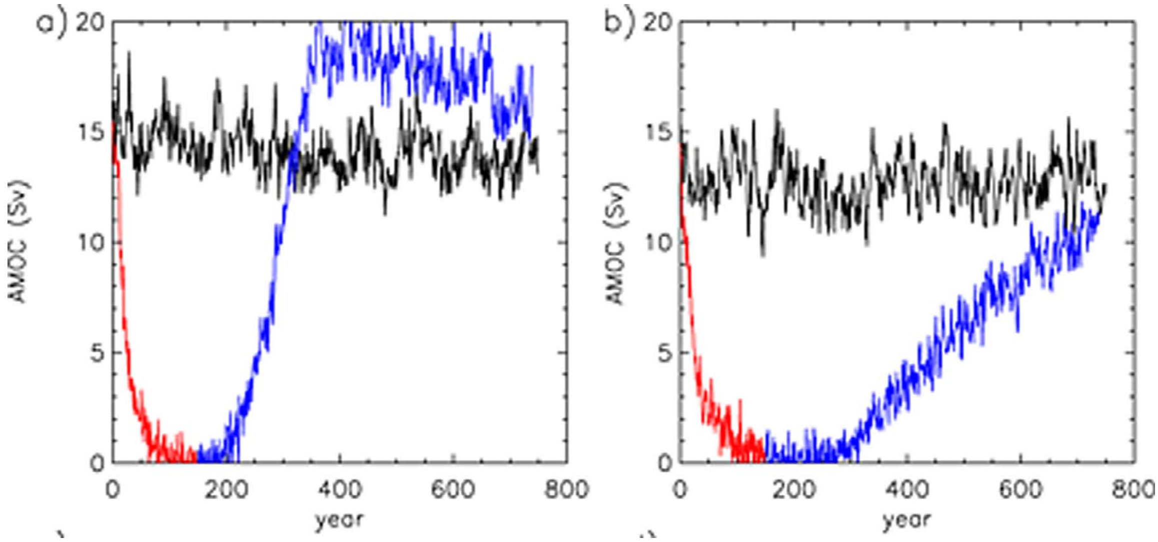

Fig. 5. Fig. 2a and b from Jackson (2013). AMOC maximum versus time. a) Non flux adjusted model, b) Flux adjusted model. Black -control run, Red –hosing run, Blue – hosing off recovery run. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

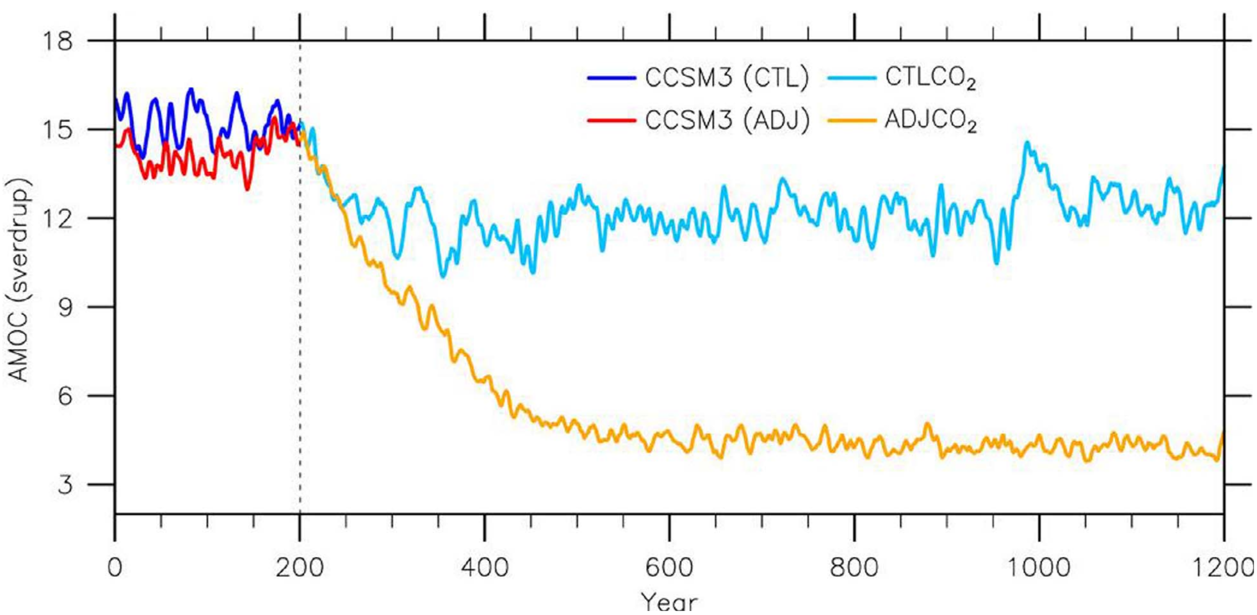

Fig. 6. Fig. 1c from Liu et al. (2017). AMOC maximum versus time, with CO_2 doubled at year 200. Dark and light blue – Non flux adjusted model, Red and Orange–Flux adjusted model. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

years. The AMOC maximum reduces to < 2 Sv during the hosing, and it remains shut off for the duration of the 450 year run. This demonstrates that, if modern climate models with non-eddy-resolving or eddy permitting resolution are given a very large perturbation, then AMOC can collapse and remain in an 'off' state for several hundred years. These two examples show this is true whether the simple stability criterion of the sign of F_{OT} at 33° S is monostable (CCSM4) or bistable (HadGEM3). It would be very interesting to know whether the AMOC in CCSM4 and HadGM3 would eventually recover after more than 1000 years, as in the Stouffer and Manabe (2003) model.

Very recent work by Rind et al. (2018) shows that AMOC can recover after being collapsed for a very long time in a modern climate model. They use the Goddard Institute for Space Studies E2-R model, which has an atmosphere resolution of 2° × 2.5° and an ocean resolution of 1° × 1.25°, and the version which has a calculated aerosol indirect effect (TCADI). Fig. 7 is from Rind et al. (2018), and shows the AMOC when the CO_2 level is either abruptly set to 4 times the control run level (upper panel), or is kept at 4 times the control run level after 140 years of a 1% per year increasing CO_2 run (lower panel). In all three runs, AMOC collapses to an 'off' state value of < 5 Sv after a few hundred years, but then recovers to stronger than its control run value after between 1000 and 1400 years and eventually settles down to near the control run value of 22 Sv. Rind et al. (2018) show that AMOC recovers because the high North Atlantic warms at depth, so that deep water formation there reactivates. This is similar to the Stouffer and Manabe (2003) result and is the first case I know of where this occurs in a climate model without flux adjustments. By my earlier strict definition, this TACDI model does not have a stable AMOC 'off' state.

In fact, I think it might be incorrect for a climate model to have a true long time stable AMOC 'off' state. The reason is that there have been large freshwater perturbations in the North Atlantic that can change AMOC from a stable 'on' state to an 'off' state, but it is more difficult to find large enough perturbations to return the climate from a truely stable AMOC 'off' state back to an AMOC 'on' state. However, Weaver et al. (2003) show that if the large meltwater pulse 1A is assumed to all come from Antarctica ice melt, then this freshwater input

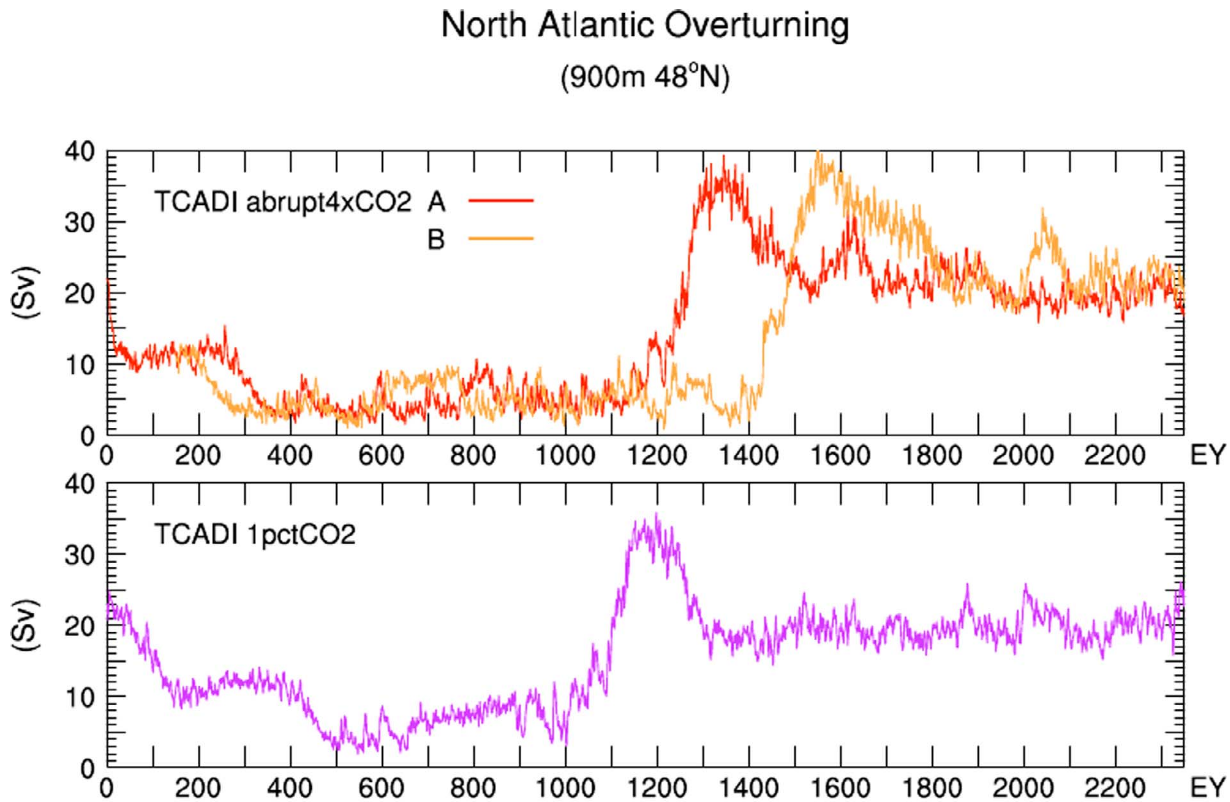

Fig. 7. Fig. 1a and b from Rind et al. (2018). AMOC maximum versus time from the GISS TCADI model for a) two abrupt 4 × CO_2 runs with slightly different initial conditions, and b) a 1%CO_2 run where the CO_2 level is kept constant after 140 years at 4 × the control run level.

can return the University of Victoria EMIC from a stable AMOC 'off' state back to an 'on' state.

7. Results from eddy-resolving climate models

Weijer et al. (2012) perform hosing experiments when the freshwater input is confined to around the coast of Greenland, as well as across 50° N–70° N in the North Atlantic. They apply a hosing of 0.1 Sv over 50 years in models where the ocean resolution is 0.1° and 1°. Their Abstract states, "We find that the overall decline of the AMOC on decadal time scales is quantitatively similar (< 10%) in the two configurations. Nonetheless, the transient response is significantly different, as the AMOC decline and reduction in wintertime convection is markedly more gradual and persistent in the strongly-eddying configuration." den Toom et al. (2014) use the same 0.1° and 1° models as Weijer et al. (2012), but use a five times stronger freshwater forcing around Greenland. Their Abstract states, "The freshwater input leads to a quantitatively comparable reduction of the overturning strength in the two models."

More recently there have been two comparisons that directly compare results using 0.1° and 1° ocean components of full climate models. Winton et al. (2014) run 1% per year increasing CO_2 experiments using an atmosphere resolution of 0.5° and ocean resolution of 0.1° (CM2.6), 0.25° (CM2.5) and 1° (CM2.1), and in seven other GFDL climate models with 1° ocean resolution. Fig. 8 is from Winton et al. (2014), and shows the reduction in AMOC at CO_2 doubling compared to the control run value. The AMOC reductions in the higher resolution models CM2.5 and CM2.6 are < 2 Sv and < 3 Sv respectively, although they both have a somewhat weak AMOC in their control runs of < 13 Sv. Table 1 shows the percentage AMOC reduction compared to the control run value and the ocean resolution for all 10 models shown in Fig. 8. It is clear that there is a rather wide range of percentage decreases in 1° models, and that the decreases in the 0.25° (CM2.5) and 0.1° (CM2.6) models are smaller than in many of the 1° models.

Small et al. (2014) show results from a 100 year control run of a very high resolution version of the CESM1, which has 0.25° atmosphere and 0.1° ocean resolution. Very recently, this model has been run between 2000 and 2050 forced by the RCP8.5 future scenario, which started from year 50 of the control run. Fig. 9 shows the maximum AMOC from both the control and RCP8.5 runs. There is quite large variability in this model, and the AMOC is generally increasing throughout the control run, but that trend is weak over years 80–100. AMOC increases for the first 13 years of the RCP8.5 run, but then decreases for the remainder of the run to 2050. Taking an average over the last 10 years of both runs, gives AMOC values of 26.2 Sv and 22.8 Sv, so that the decline is 3.4 Sv or 13% of the control run value. Fig. 8 of Meehl et al. (2013) shows the AMOC maximum in the standard 1° CESM1 version for several future scenarios including RCP8.5. The AMOC decline at 2050 is 5.3 Sv compared to 23.9 Sv at 2000; a decline of 22%, which is larger than in the very high resolution CESM1 version.

The Winton et al. (2014) and CESM1 results show there is quite a range of AMOC declines in current climate models using 1° ocean resolution, which indicates that ocean resolution is certainly not the dominant factor affecting AMOC stability. In addition, they show that AMOC declines using eddy-resolving resolution of 0.1° are smaller than when using a 1° ocean with the same atmosphere model. Therefore, I think that assuming AMOC is more unstable in eddy-resolving models using 0.1° compared to climate models using 1° ocean resolution is not justified.

8. Conclusions and discussion

There are a number of conclusions that can be drawn from the results highlighted in this review:

(1) Bistability in ocean alone models with mixed boundary conditions

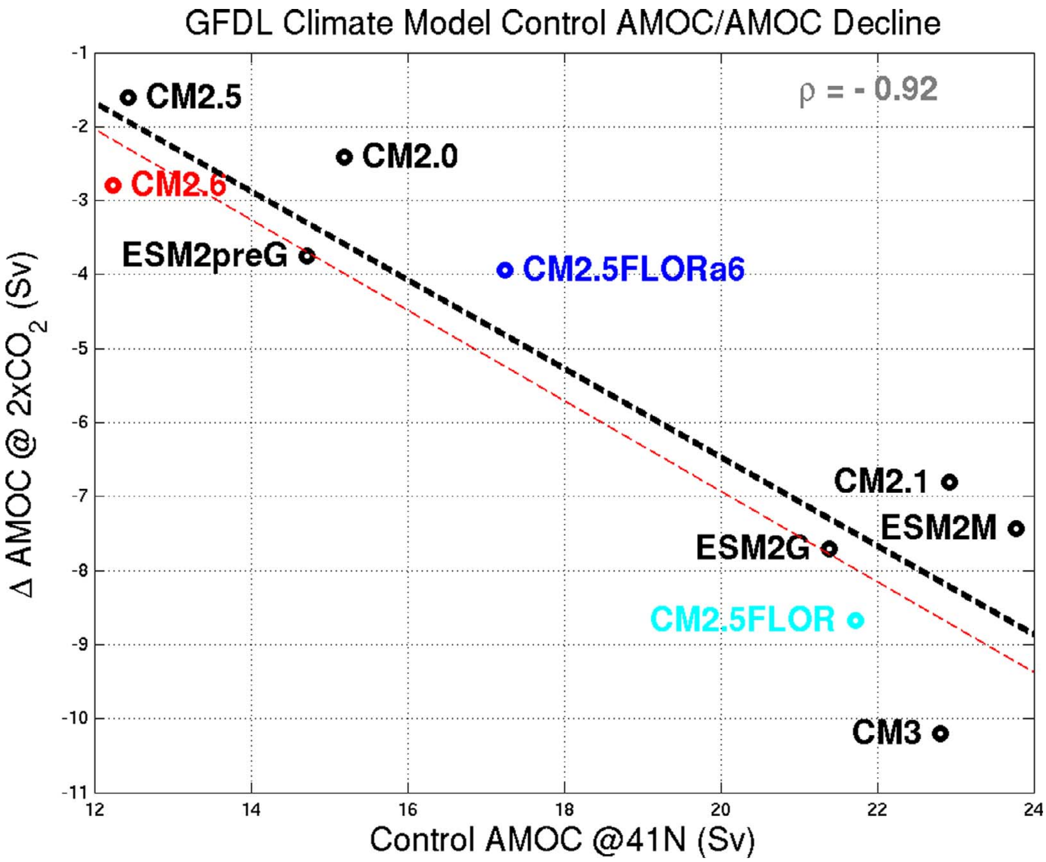

Fig. 8. Fig. 2b from Winton et al. (2014). AMOC reduction at CO_2 doubling in 1% CO_2 increasing runs versus AMOC control run maximum in a suite of GFDL climate models.

Table 1
Percentage reduction in AMOC at CO_2 doubling in 1% increasing CO_2 runs compared to the control run value in 10 GFDL climate models.

Model	*% AMOC reduction*	*Ocean resolution*
CM2.5	13%	0.25°
CM2.0	16%	1°
CM2.6	23%	0.1°
CM2.5FLORa6	23%	1°
ESM2preG	23%	1°
CM2.1	30%	1°
ESM2M	31%	1°
ESM2G	36%	1°
CM2.5FLOR	40%	1°
CM3	45%	1°

depends quite strongly on the model formulation; especially the vertical mixing scheme.

(2) Bistability is much reduced when going from ocean alone models to fully coupled climate models, which have several more air–sea interaction stabilizing feedbacks.

(3) Bistability has been found in several EMICs which have reduced air–sea interaction feedbacks, and in one fully coupled climate model that has very coarse resolution.

(4) Whether the AMOC freshwater transport across 33° S is positive or negative is very frequently used as a criterion of whether a climate model's AMOC is monostable or bistable. However, if the AMOC freshwater transport is negative, then the gyre freshwater transport must be positive and larger than the AMOC transport because the total freshwater transport across 33° S must be positive. A recent strong hosing experiment using an eddy-permitting 1/4° ocean has shown that the gyre circulation freshwater transport change is comparable in magnitude to the AMOC freshwater transport change throughout the Atlantic, and is larger at 33° S. This casts very strong doubt on the usefulness of the simple AMOC stability criterion, which assumes that other freshwater transport processes, including the gyre circulation, do not change when AMOC changes.

(5) Three recent surveys of this simple AMOC stability criterion show that over 40% of modern CMIP5 climate models have a negative AMOC freshwater transport across 33° S, which possibly indicates a bistable state. This is contrary to statements in two recent papers saying that nearly all climate models have a positive freshwater transport across 33° S, possibly indicating a monostable state, and therefore have a too stable AMOC.

(6) In a flux-adjusted climate model, the adjustments interfere with the air-sea interaction feedbacks and strongly change the AMOC stability properties. This has been shown in the GFDL R30, HadCM3 and CCSM3 T31 × 3 models. I conclude that AMOC stability results from a climate model that uses large flux adjustments should be viewed with great suspicion. A flux adjusted model may be useful for short-term predictions when an improved climate state is arguably more important than the air–sea feedbacks, but for AMOC the opposite is true because the air–sea feedbacks strongly affect its stability.

(7) AMOC can be shut off for many hundreds of years in modern, fully coupled climate models if the perturbation is large enough, such as a very large increase in CO_2 or very strong hosing in the North Atlantic. This suggests that a model does not need a stable 'off' state in order to have very long periods of weak AMOC strength. AMOC can then potentially recover if deep water formation resumes in the Labrador and Nordic Seas.

(8) Recent 1% increasing CO_2 runs using a suite of GFDL models and RCP8.5 scenario runs in very high and standard resolution CESM1 versions have shown that the reduction in AMOC is smaller using an eddy-resolving 0.1° ocean than the standard 1° ocean resolution. Therefore, assuming AMOC is more unstable in eddy-resolving models compared to standard 1° climate models is not justified.

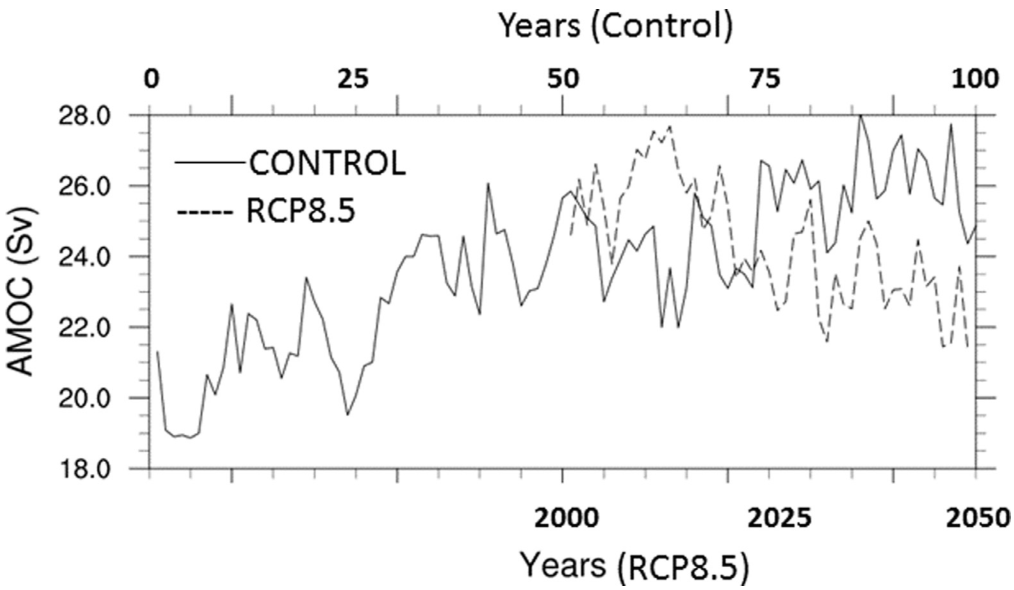

Fig. 9. AMOC maximum versus time from high resolution CESM control run, Small et al. (2014), and RCP8.5 scenario run from 2000 to 2050.

In summary, I think that obtaining the correct AMOC stability in climate models is very subtle and tricky and is clearly not settled at this point. Modern 1° climate models show a wide range of AMOC magnitudes in control runs and future declines when forced by the same increasing CO_2 scenario. However, recent evidence does not support statements that nearly all modern climate models have an AMOC that is too stable. A climate model does not need a stable AMOC 'off' state in order to simulate a rapid decline due to a large freshwater injection in the North Atlantic, followed by a long period of small AMOC before it recovers, and this cycle is very relevant on paleoclimate time scales. I think the way forward must be to improve the Atlantic salinity distribution, which is indeed quite unrealistic in many climate models. This can be achieved either by improving modern climate models or by using higher resolution than 1° and obtaining a much improved evaporation minus precipitation plus runoff distribution from the atmosphere component. It is not clear to me that using ocean eddy-resolving resolution is either required or optimal in the near future. Eddy length scales are partially resolved near the equator in 1° models, especially those with finer meridional resolution there. However, the real problem is that eddy-resolving resolution very strongly limits the length of coupled integrations. Runs of at least several hundred years, if not a thousand years, are required in order to make more reliable conclusions about AMOC stability, so that a resolution of near 0.5° in all components would allow adequately long runs. In my opinion, obtaining a realistic freshwater flux field from the atmosphere component is more important than higher ocean resolution in order to improve the Atlantic Ocean salinity distribution. There is an excellent discussion on this topic in Section 5 of Mecking et al. (2017).

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