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Review

Instability of Atlantic Meridional Overturning Circulation: Observations, Modelling and Relevance to Present and Future

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Abstract: The Atlantic Meridional Overturning Circulation (AMOC) has changed dramatically during the glacial–interglacial cycle. One leading hypothesis for these abrupt changes is thermohaline instability. Here, I review recent progress towards understanding thermohaline instability in both observations and modelling. Proxy records available seem to favor thermohaline instability as the cause of the abrupt climate changes during the glacial–deglacial period because the deep North Atlantic water mass and AMOC seemed to have changed before the North Atlantic climate. However, most fully Coupled General Circulation Models (CGCMs) so far seem to exhibit monostable AMOC, because (1) these models have failed to simulate abrupt AMOC changes unless they are forced by an abrupt change of external forcing and, (2) these models have shown opposite freshwater convergence from the current observations. This potential model bias in the AMOC stability leaves the model projection of the future AMOC change uncertain.

Keywords: thermohaline instability; model bias; abrupt changes

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1. Introduction

Paleoclimate proxies suggest that the Atlantic Meridional Overturning Circulation (AMOC) and global climate have experienced abrupt changes of millennial time scales [1,2]. Figure 1 shows multiple proxies relevant to the AMOC, its forcing and its climate impact, for the last 80,000 years [3]. The meltwater flux to the North Atlantic is characterized by strong millennial changes superimposed on the last glacial–deglacial cycle, as seen in the oxygen isotope ratio $\delta^{18}O$ (~$^{18}O/^{16}O$) over the North Atlantic surface (Figure 1c). A more depleted (lower) surface water $\delta^{18}O$ indicates a greater meltwater flux, due to the contribution of more meltwater from ice sheets that have all been very depleted $\delta^{18}O$ relative to the average ocean water (−30‰ vs. 0‰). Similarly, large variability is also evident in two proxies of deep water masses ($\delta^{13}C$, $\varepsilon_{\mathrm{Nd}}$ in Figure 1d,e) and three proxies of circulation ($^{231}Pa/^{230}Th$, $\delta^{18}O$, grain size, Figure 1f–h). The carbon isotope composition $^{13}C/^{12}C$, expressed as $\delta^{13}C$, and the neodymium isotopic composition $^{143}\mathrm{Nd}/^{144}\mathrm{Nd}$, expressed as $\varepsilon_{\mathrm{Nd}}$, from the shells of benthic foraminifera are two proxies for deep water masses, because their value in the deep water is determined by the competition between the high $\delta^{13}C$/low $\varepsilon_{\mathrm{Nd}}$ North Atlantic Deep Water (NADW) source water and the low $\delta^{13}C$/high $\varepsilon_{\mathrm{Nd}}$ of the Antarctic Bottom Water (AABW) source water [4–6]. The different behavior of uranium decay-series nuclides of protactinium and thorium, expressed as $^{231}Pa/^{230}Th$, from sediments in the North Atlantic is considered a proxy of the rate of deep Atlantic circulation, because ^{231}Pa has a longer residence time than ^{230}Th (111 years vs. 26 years) such that the sediment $^{231}Pa/^{230}Th$ decreases with increased export of Atlantic deep water [7]. The variability of benthic calcite $\delta^{18}O_c$ is related to sea water temperature and salinity, and is therefore related to density anomaly and, in turn, ocean current strength through the thermal wind relation [8]. The grain size in the deep western boundary current region (DWBC) is a proxy of DWBC strength because a coarser grain size is likely to be caused by a stronger current. Finally, large variability is also found accompanying deep ocean changes in the proxy for surface air temperature in Greenland and Antarctica ice cores as indicated in the stable water isotope composition in precipitation $\delta^{18}O$, which represents annual temperature via the "temperature effect" associated with Rayleigh distillation [9] (Figure 1a,b).

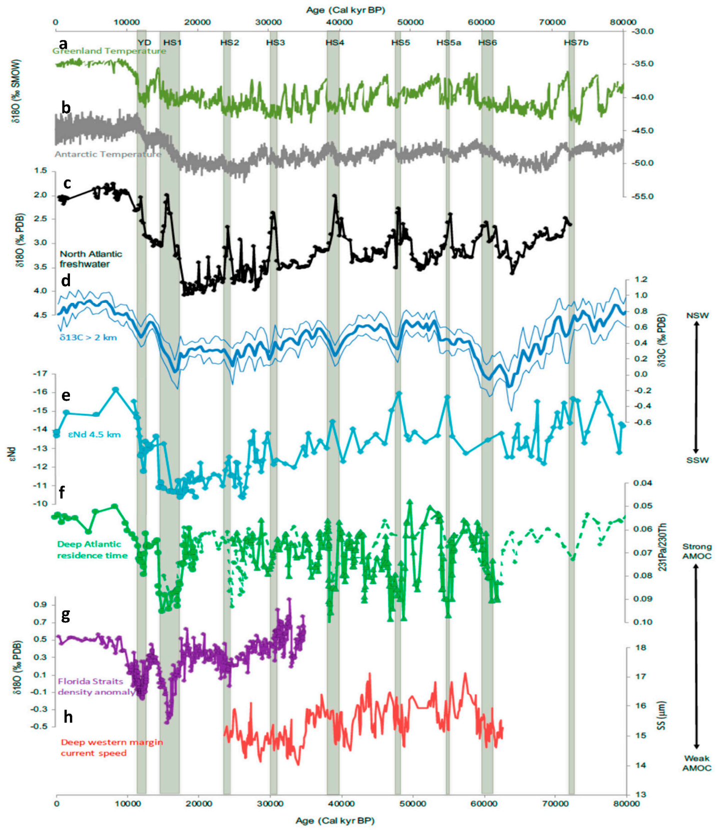

Figure 1. North Atlantic water-mass and circulation proxy records over the Heinrich events, along with temperature proxies over Greenland and Antarctica. (**a**) Greenland ice core $\delta^{18}O$. (**b**) Antarctica $\delta^{18}O$. (**c**) The oxygen isotope ratio in the planktonic foraminifer Neogloboquadrina pachyderma from the western North Atlantic; low values reflect the presence of glacial meltwater. (**d**) Mean and 2-sigma standard-error $\delta^{13}C$ values (thick and thin lines, respectively) for the deep North Atlantic (>2 km). (**e**) The deep (4.55 km) Nd isotope ratio (εNd) at the Bermuda Rise. (**f**) The Pa^{231}/Th^{230} ratio, which can reflect changes in deepwater residence time, from the deep North Atlantic at the Bermuda Rise. (**g**) The ice-volume-corrected oxygen isotope ratio of benthic foraminifera on the Florida Margin, which can reflect changes in the density structure in the Florida Straits and the strength of the upper branch of the Atlantic meridional overturning circulation (AMOC). (**h**) The mean sortable silt grain size along the western boundary of the North Atlantic at the depth of today's North Atlantic Deep Water, which can indicate the current speed on the deep western margin; a large mean size indicates a vigorous flow. See [3] Figure 4 for more details on the figure. (Courtesy of J. Lynch-Stieglitz).

After the discovery of abrupt changes in events of Dansgaard–Oeschger (DO) oscillation in Greenland ice cores and their links to North Atlantic water masses in the 1980s, AMOC instability associated with basin-wide positive salinity feedback [10,11] has been proposed as one leading mechanism for the millennial climate variability [12–14]. Greenland stadial and NADW reduction have been further found to be associated with massive ice rafting of the Heinrich events [15,16]. The direct link of abrupt climate changes with the AMOC intensity was further found in the proxies sensitive to the AMOC intensity, notably sediment $^{231}Pa/^{230}Th$ [7,17,18], $\delta^{18}O_c$ gradient for Florida Current [8] and sortable grain size for DWBC current speed [19]. Furthermore, the greatest AMOC reduction and the coldest stadial intervals were found to be concurrent with the largest iceberg discharges during Heinrich events [3,18,20,21] and the AMOC reductions associated with the Heinrich stadial also coincide with intervals of rising atmospheric CO_2 [22], suggesting a potential influence of deep ocean circulation on atmospheric CO_2 [23]. Within the chronology uncertainty, these AMOC changes were found to be concurrent with climate variability in most records. In the real world, abrupt climate change could have been caused by ice-sheet instability as well as thermohaline instability. If the thermohaline instability is the cause of abrupt changes, AMOC changes likely occurred before climate changes. On the other hand, if the ice-sheet instability is the cause, the climate likely has changed with ice sheet before the AMOC. This observed concurrence of AMOC variability and climate variability, therefore, has left the origin of the abrupt climate change still wide open: is it of the ocean origin from AMOC instability, or the ice-sheet origin from ice-sheet instability?

Observations of abrupt climate changes have stimulated intensified modelling studies on AMOC instability first in ocean-alone models with mixed boundary conditions [24–27] and then a Coupled General Circulation Model (CGCM) [28] and, most comprehensively, in earth system models of intermediate complexity (EMICs) [29,30]. The AMOCs in most EMICs have been found in a bistable regime [31], associated with the positive salinity feedback [10] in the Atlantic basin [32,33]. Therefore, abrupt climate change events similar to DO and Henrich events have been simulated in these models even under a gradual change in external forcing [27,29,30,34]. In the state-of-the-art CGCMs, however, what is the current status of AMOC instability and how is the model AMOC instability compared with observations?

In this paper, I will review recent progress in observations, modelling and understanding of AMOC instability. Given extensive reviews on this topic, especially related to the theoretical aspect [35–38], I will focus on some recent progress in paleo observations and potential bias in CGCMs.

The review is arranged as follows. In Section 3, I will review the current status of AMOC instability in CGCMs. In Section 4, I discuss some personal perspective on the relevance of AMOC instability to the present and future climate changes. A summary and perspective are given in Section 5.

2. Materials and Methods

All materials are derived from published papers where the detailed methods are described.

3. Results

3.1. Paleo Evidence of AMOC Instability

Evidence of an ocean origin of the abrupt change in AMOC, or thermohaline instability, requires a clear lead of the AMOC change over the ice sheet and the resulting meltwater flux and ice rafting. Similarly, an ice-sheet origin of abrupt change via ice-sheet instability requires a clear lead of the ice-sheet and meltwater changes over the AMOC. Recent paleo observations seemed to favor an ocean origin, although not yet unambiguously. These analyses were carried out on multiple proxies in the same sediment core, such that the relative phasing was accurate among different variables, saving the potentially different preservation times of different proxies. Ref. [21] analyzed ice rafting debris and AMOC strength in the subpolar North Atlantic and showed a lead of ~200–300 years of the AMOC weakening prior to ice rafting events for most Heinrich events, demonstrating clearly that AMOC collapses were not triggered by icebergs. This result was consistent with reconstructions of the subsurface temperature in the North Atlantic, which showed warming leading to ice rafting events by hundreds of years for the most recent Henrich events [39,40].

However, it remains possible that the AMOC weakening was triggered by the initial meltwater flux from ice-sheet instability [41] if the iceberg export occurred later than the initial melting. There are no records that can clarify this point unambiguously. However, ref. [18] so far likely provides the clearest evidence suggesting AMOC change preceding North Atlantic climate change. Ref. [18] performed a lagged correlation between multiple proxies and the Greenland temperature proxy ($\delta^{18}O_p$) in a sediment core over the Bermuda

Rise. Relative to the Greenland temperature, this analysis showed a clear lead of 200 years for the benthic $\delta^{13}C$ (purple), accompanied by an almost in phase AMOC strength proxy $-^{231}Pa/^{230}Th$ (green, the legend should be of a minus sign) and North Atlantic SST (red) (Figure 2). At face value, this analysis suggested that the NADW was enhanced (represented by a positive $\delta^{13}C$ anomaly) before the enhancement of the AMOC intensity (represented by more negative $^{231}Pa/^{230}Th$ or more positive $-^{231}Pa/^{230}Th$), North Atlantic SST and Greenland temperature. Allowing for a century scale response time of sediment $^{231}Pa/^{230}Th$ after the AMOC change, the AMOC strength likely has changed preceding Pa/Th with its phase close to the NADW [18]. Thus, the AMOC likely has changed before the temperatures over the North Atlantic region, precluding the AMOC change as a response to climate change, which would otherwise have led to an AMOC response lagging surface temperature. This analysis so far provides the strongest evidence favoring the oceanic origin of AMOC collapse. This analysis, however, has left the question open as to why the temperature changed after the AMOC, as opposed to synchronously (within data resolution time scale of decades) with the AMOC and the associated heat transport.

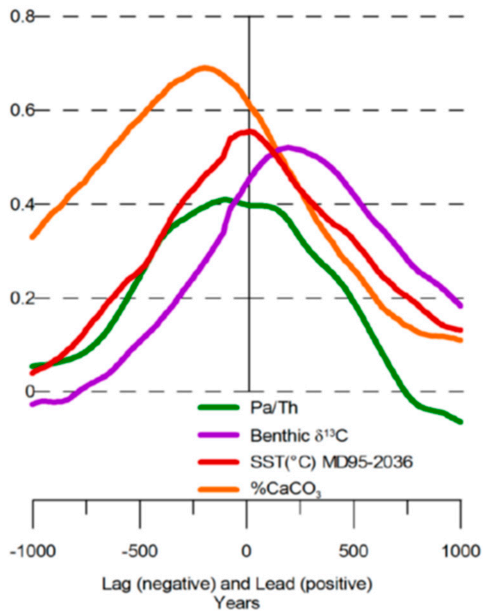

Figure 2. Cross-correlation of NGRIP ice core $\delta^{18}O$ with CDH[19] CaCO3 flux (orange), Pa/Th of bulk sediment from CDH[19] (green), $\delta^{13}C_{BF}$ from CDH19 (purple), Sea Surface Temperature (SST, °C) from Bermuda Rise sediment core MD95-2036 (red). A lead (positive) refers to the lead of the proxy relative to Northern Greenland Ice Project (NGRIP) ice core. See [25] for details. Adapted from [18]. (Note: the legend for Pa/Th should be of minus sign −Pa/Th, J. McManus, personal communication.)

Regardless of the AMOC instability, even the concurrent variability of the AMOC and climate implies a robust positive feedback between the AMOC and ice-sheet change. Proxy evidence [39,40] and modelling studies [42–44] suggested a subsurface warming accompanying the surface cooling in the subpolar North Atlantic in response to a meltwater flux, weakened deep convection and AMOC. This subsurface warming would promote ice-sheet calving and further melting and, in turn, a further weakening of the AMOC [44]. This positive feedback suggests a coupled AMOC–ice-sheet instability that depends on the coupling of both components, even if either component alone is stable. This is analogous to the case of El Nino, which is caused by a positive coupled ocean–atmosphere feedback, or instability, while the atmosphere or ocean alone is stable when the other component is fixed.

Finally, it should be noted that some abrupt transition can occur within decades. For example, the termination of the Younger Dryas event (YD, 12,900–11,600 years ago) has been observed to occur within 40 years in the stable water isotope signals in Chinese cave records [45,46]. It remains challenging to interpret such a rapid transition in terms of AMOC instability. Other mechanisms, notably those related to sea ice changes, may be responsible for such rapid transitions.

3.2. AMOC Instability in CGCMs

In most CGCMs so far, the AMOC seemed stable. This statement, however, remains highly uncertain, because, unlike EMICs, CGCMs are computationally too expensive to perform comprehensive sensitivity experiments over thousands of years under different perturbation forcing. Instead, so far, AMOC stability in most CGCMs has been assessed in two approaches. The first approach is the direct perturbation experiment, but mostly of a short duration of hundreds of years. In response to a meltwater pulse, the AMOCs in most CGCMs have been found to recover after the termination of the freshwater forcing for the Present Day (PD) [47] and LGM [48]. In response to long term CO_2 rise, CGCMs so far also show gradual responses [49]. Similarly, in response to a rapid rise of atmospheric CO_2, the AMOC also tended to decline gradually (instead of abruptly), including the simulations in IPCC reports [50], followed by slow recovery at millennial time scales [49,50]. These short experiments have led to the impression of a monostable AMOC, although it is possible that the perturbations might not have been strong or long enough to push the AMOC out of the stability attractor [32]. This impression was corroborated by a few long simulations of past climate changes in a small subset of models. Therefore, in most current CGCMs, abrupt climate changes could be generated only by abrupt changes in the forcing ([51–55]).

There are a few CGCMs (without flux adjustment) that exhibited AMOC instability in idealized paleo climate modelling. In a coarse resolution model, the AMOC seemed to be locked in a broad bistability regime such that the AMOC changed abruptly to another state in response to a gradual change in ice-sheet topography, atmospheric CO_2 or orbital forcing [56–58]. Furthermore, AMOC instability and the resulting millennial variability tended to occur for intermediate levels of glacial ice sheet and CO_2, consistent with the observations that the strongest millennial variability occurred in Stage 3 in the last glacial–interglacial cycle (Figure 1). Nevertheless, temperature variability over Greenland was underestimated in this model relative to reconstructions [59]. In several CGCMs, the AMOC seemed to transition to an unstable state of strong millennial variability in response to perturbation climate forcing [60–64]. In one version of the Community Earth System Model version 1 of the modified ocean diffusivity coefficient, the AMOC in the LGM setting exhibited DO-like variability of about 700 years of duration after being triggered by a strong meltwater pulse [60,61]. Why the change in diffusivity caused the change in AMOC stability, however, remains unclear. Oceanic diapycnal mixing has been suggested to enhance AMOC instability in some theoretical models and EMICs [65,66], but to suppress AMOC instability in a CGCM [67] and not to qualitatively affect the AMOC stability in an EMIC [68]. Overall, it remains unclear why these few CGCMs that do not exhibit monostable AMOC are different from most CGCMs that seem to exhibit monostable AMOC, given their comparable resolution and complexity of model physics.

The second approach was an indirect approach in which AMOC stability was inferred from a diagnostic indicator $M_{ov,S}$ that represents the southward freshwater export by the zonal mean overturning circulation across ~30° S in the Atlantic [11]. The AMOC is inferred bistable for freshwater exporting $M_{ov,S} < 0$, because a weakening AMOC leads to a freshwater pileup in the Atlantic, and then a further weakening of the AMOC. This indicator has been found to be remarkably successful in EMICs [32,33,69–71], and even in a few CGCMs [72,73], in the sense that its sign was consistent with the AMOC stability tested in direct simulations. It should also be noted that the fidelity of the indicator seemed to be improved substantially after a refinement to include the AMOC freshwater transport across the northern boundary of the North Atlantic $M_{ov,N}$ as the net freshwater import $\Delta M_{ov} = M_{ov,S} - M_{ov,N}$ [68,74]. For example, almost all CMIP5 models responded to increased CO_2 with a gradual slowdown, likely implying monostable AMOC, even though 40% of the models were inferred as bistable in terms of $M_{ov,S}(< 0)$ [75–78]. However, in terms of ΔM_{ov}, most models were indeed inferred as stable [79] (Figure 3), suggesting the non-negligible role of $M_{ov,N}$. An additional improvement can be further made to take into account the slow temporal evolution of the AMOC [76,77]. The most valuable aspect of this indicator was its applicability to present observations. All the observational data sets showed a net freshwater export $M_{ov,S} < 0$ or $\Delta M_{ov} < 0$, implying a bistable regime of the present AMOC [24,32,38,72,78–81] (Figure 3). If this indicator is indeed correct, the real-world AMOC would be unstable at the present and most CGCMs are too stable and thus might have underestimated the likelihood of abrupt climate changes [80,81].

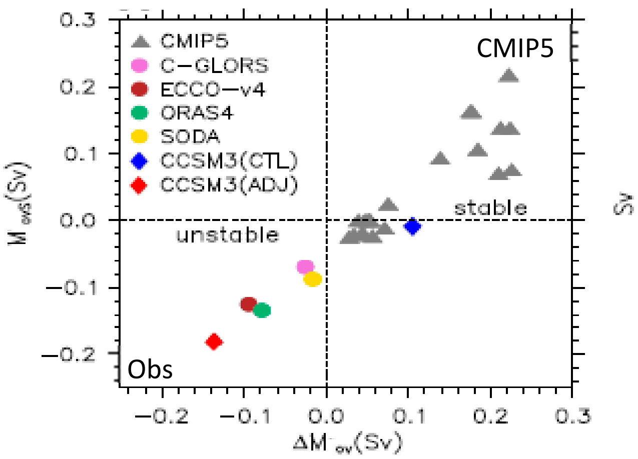

Figure 3. AMOC stability in the modern climate as assessed by stability indicators. The indicator values in the modern climate are shown in solid circles with color for 4 reanalysis data sets, in gray triangles for the CMIP5 model simulations, and in blue and red diamonds for the CCSM3 before (CTL) and after (ADJ) flux adjustment. A positive or negative indicator implies that the AMOC resides in a stable or bistable regime. Note that for CMIP models, about 40% shows $M_{ov,S} < 0$, but all show $\Delta M_{ov} > 0$. Adapted from [72].

Given the potential importance of this stability indicator, it is interesting to ask why most CGCMs have shown a net AMOC freshwater import, opposite to current observations, while EMICs have shown an AMOC freshwater export as in current observations. The biased freshwater transport in CGCMs has been found to be caused mainly by the salinity bias in the South Atlantic, especially the fresh bias of the surface subtropical water [38,82,83]. From the oceanic perspective, this surface fresh bias could be caused by the deficient ocean model, such as the coarse model resolution and the deficient Agulhas Retroflection [37,38,84]. Yet, even with similar coarse resolution ocean models, EMICs were able to simulate the freshwater export as in observations. This seemed to suggest that the salinity bias in the CGCMs were caused by certain model biases common in CGCMs, but not in EMICs. One such bias was suggested to be the tropical bias of the double Intertropical Convergence Zone (ITCZ) [81], which has been a persistent bias in almost all CGCMs [85,86], especially over the tropical Atlantic sector [87]. This southward bias of ITCZ would lead to excessive rainfall and surface freshening in the surface subtropical South Atlantic [81,83]. This hypothesis has been tested systematically in one CGCM using flux-adjustment sensitivity experiments [81]. While the original CCSM3 model was indeed monostable under direct meltwater perturbation, consistent with $\Delta M_{ov} > 0$, the model after flux adjustment is changed to being bistable in terms of both direct perturbation experiment and $\Delta M_{ov} < 0$ (Figure 3). This hypothesis appeared to also be consistent in another two CGCMs, in which the model without flux adjustment is monostable but the model with flux adjustment is bistable [88,89]. However, a systematic study is still lacking across models on the role of flux adjustment on AMOC stability.

The tropical bias hypothesis seemed to offer an explanation as to why EMICs tended to be more unstable than CGCMs. Due to the simplified atmospheric physics in EMICs, their model climatology is generated with some information on the present climatology of wind and/or SST. Thus, all EMICs could be considered as models after flux adjustment. Another hypothesis is that the salinity bias was contributed by the model bias of the Antarctica Intermediate Water (AAIW) [90]. However, for most CGCMs, the AAIW was more saline than observations, which would have favored freshwater export, instead of import, in these CGCMs [91].

The key question still remains open as to how correct the indicator is, in CGCMs and ultimately the real world. Physically, the freshwater budget of the North Atlantic and the final salinity feedback with the AMOC can be affected by the AMOC freshwater transport as well as other processes, such as the azonal gyre transport M_{az}, atmospheric feedback on evaporation minus precipitation and wind stress (and in turn gyre transport) and oceanic mixing processes (see [37,84] for comprehensive reviews). Indeed, there are at least two CGCMs showing AMOC instability, but without flux adjustment, they are inconsistent with this indicator ([72], but see an alternative explanation by [92,93]).

3.3. Relevance to the Present and Future

AMOC instability is highly relevant to the future climate state and abrupt changes [94], which remain highly uncertain [95]. AMOC instability can be state dependent such that the AMOC stability differs in different times, glacial periods, Holocene, the present and future. What is then needed is the understanding of how the AMOC stability changes with climate state. In spite of the uncertainty, the stability indicator can serve as a useful starting point. The indicator implies an unstable AMOC in observations at the present time. This unstable AMOC, however, seems to contradict the lack of abrupt events in the Holocene (except for the modest 8.2 ka event, [96]), if the Holocene is taken as an analogue of the present. Alternatively, this lack of large abrupt changes in the Holocene could be caused by the lack of a strong trigger in the Holocene.

If we take the evidence of a likely unstable AMOC in the glacial cycle and assume the indicator is correct for the present period, most current CGCMs would be too stable, implying an underestimation of the possibility of abrupt climate changes in the future [79]. This over-stable AMOC is consistent for most recent CGCMs without flux adjustment, in which the AMOC responded gradually to the future rise of CO_2 [49]. There are, nevertheless, three exceptions for CGCMs without flux adjustment that show abrupt collapse in hundreds of years, as presented in ref. [57,93,97]. The evolution of the AMOC in the future will be affected further by the melting of ice sheets in Greenland and Antarctica in the long run. It therefore remains highly uncertain how the AMOC will change in the future. It should be kept in mind that abrupt changes in models with flux adjustment should be treated with great caution [37], because of the potential distortion of the AMOC stability by flux adjustment, as analyzed in simple models [98]. Equally, however, it should be realized that there is no reason to trust the projections more from those current models without flux adjustment, as long as these models still suffer from severe salinity bias and, in turn, the AMOC freshwater transport, even if the stability indicator may not be perfectly correct.

4. Discussion

Recent progress in paleoceanographic proxies seem to favor the oceanic origin of AMOC instability as the cause of the abrupt climate changes during the glacial–interglacial period. Most CGCMs, however, seem to be over-stable, judging from the limited sensitivity experiments available, as well as the stability indicator ΔM_{ov}, although it remains uncertain how correct this indicator is across CGCMs and in the real world.

Further paleo proxy records, especially those with high-temporal resolution, are needed to distinguish the AMOC instability from the ice-sheet instability as the origin for abrupt climate changes. These records may further include those outside the North Atlantic, say, in the North Pacific [99], because of their potential links to the abrupt changes in the AMOC.

Even more challenging is the assessment of the AMOC instability in the real world for the present and future. For the present, it has remained difficult to detect the AMOC response to the global warming of rising CO_2. This is because the direct instrument measurement of AMOC transport has only been available for two decades. This short record can be significantly distorted by multidecadal variability and therefore is too short to detect the trend response to CO_2 rise. Observational evidence of deep warming in the Atlantic and Southern Ocean [100–104] is not good evidence of the AMOC response either. Besides also being too short, the deep warming could be caused simply by the advection of the mean circulation, notably, the deep western boundary current [105], instead of a change in the AMOC circulation. Nevertheless, a recent study of two AMOC fingerprints in the North Atlantic surface temperature [106,107] and South Atlantic surface salinity [108,109] seemed to provide the clearest evidence so far of the AMOC slowdown response to global warming. This slowdown response, if true, could be simply the forced response of the AMOC, even without instability.

Finally, it is certainly worrisome, to say the least, that the state-of-the-art CGCMs still show the opposite AMOC freshwater transport, which is potentially related to the salinity feedback and in turn, AMOC instability. A diagnostic indicator, even if imperfect, provides the only way that the AMOC stability can be assessed for the present day real world, which then can be compared with models. Given all the odds of potential feedback beyond a simple conceptual model, it is already surprising that the indicator ΔM_{ov} even works in many EMICs and some CGCMs. In EMICs, this indicator has been shown to represent the physical process of basin-wide salinity feedback associated with perturbation flow on mean salinity, while the gyre-induced freshwater transport is not sensitive to AMOC changes [32,33]. These feedback processes may be altered in CGCMs, especially in high-resolution models, leading to inconsistency between the indicator and AMOC stability [37,73,110]. Is it then possible to derive an improved stability indicator? For example, should the AMOC freshwater transport be calculated at a latitude other than 30° S, such as the intergyre boundary where the gyre transport change seems to be weak [73,90]?

5. Conclusions

Ultimately, AMOC instability, including any potential instability indicator, should be studied in the most realistic models without flux adjustment: high-resolution models with little bias in model climatology. This poses several challenges. First, the high-computational cost for the eddy resolving high-resolution models makes it difficult to perform extensive and long simulations that are needed to test any stability indicator. Second, if the indicator is related to a certain model bias, such as the tropical bias, these biases need to be significantly reduced in these CGCMs for a credible test of the indicator. The reduction of this bias, however, will be challenging because some biases are stubborn, notably the tropical bias which has been one of the most stubborn biases in CGCMs. Finally, AMOC instability may involve different feedback on different time scales, which may also be related to various transient behaviors of the AMOC responses, the latter being more relevant in the near future of climate change [93,97,110–112]. The different transient behaviors may be related to the basin-wide salinity feedback [35], as well as other feedback, such as the local convective feedback in the subpolar North Atlantic [113], feedback with atmosphere and sea ice [114], and the salinity feedback between the tropical and North Atlantic [89].

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