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Abstract The notion that the Atlantic Meridional Overturning Circulation (AMOC) can have more than one stable equilibrium emerged in the 1980s as a powerful hypothesis to explain rapid climate variability during the Pleistocene. Ever since, the idea that a temporary perturbation of the AMOC—or a permanent change in its forcing—could trigger an irreversible collapse has remained a reason for concern. Here we review literature on the equilibrium stability of the AMOC and present a synthesis that puts our understanding of past and future AMOC behavior in a unifying framework. This framework is based on concepts from Dynamical Systems Theory, which has proven to be an important tool in interpreting a wide range of model behavior. We conclude that it cannot be ruled out that the AMOC in our current climate is in, or close to, a regime of multiple equilibria. But there is considerable uncertainty in the location of stability thresholds with respect to our current climate state, so we have no credible indications of where our present-day AMOC is located with respect to thresholds. We conclude by identifying gaps in our knowledge and proposing possible ways forward to address these gaps.

1. Introduction

1.1. AMOC Stability and Its Relevance for the Climate System

The climate of northwest Europe, and indeed the whole of the Northern Hemisphere (NH), is profoundly influenced by the oceanic transport of heat and salt from the tropics to the subpolar regions. The release of heat from the ocean to the atmosphere in the subpolar North Atlantic makes a major contribution to the relatively mild climate in northwest Europe, which is up to 6 °C warmer than similar maritime climates bordering the Pacific (see Palter, 2015, for a review). In the Atlantic the northward transport of heat extends through the whole ocean with the remarkable consequence that in the South Atlantic Ocean heat is transported toward the equator. This meridional heat transport is associated with the Atlantic Meridional Overturning Circulation (AMOC), an overturning cell that features northward transport of warm upper ocean water, balanced by a southward flow of cooler deep water.

The strength of the AMOC at 26.5°N has been observed by the RAPID array since 2004 and has an average of about 17.0 Sv (1 Sv $\equiv 10^6$ m^3/s). The time series shows variability on all timescales, including a weakening of 15% over the length of the record (Smeed et al., 2018). Many climate models display variability of the AMOC on decadal and multidecadal timescales (e.g., Muir & Fedorov, 2015, 2017). Most models also predict a gradual weakening of the AMOC over the 21st century in response to anthropogenic forcing (e.g., Cheng et al., 2013; Weaver et al., 2012). In fact, Dima and Lohmann (2010), Rahmstorf et al. (2015), and Caesar et al. (2018) argue that such a gradual slowdown has already started and is noticeable in proxy records of the AMOC from the midtwentieth century.

A key process in the maintenance of the AMOC is thought to be deep convection in the subpolar North Atlantic Ocean, in particular the Labrador and Nordic Seas (e.g., Marshall & Schott, 1999). Here, wintertime cooling of relatively salty surface waters leads to convective instabilities that drains heat from the water column and produces cold and salty North Atlantic Deep Water (NADW). The salinity of the upper North

Atlantic is significantly higher than in other basins, a salinity surplus that is maintained partly by a net excess of evaporation over precipitation, river runoff, and ice melt (Emile-Geay et al., 2003; Schmitt, 2008; Warren, 1983) and partly by oceanic transports, in particular, the import of salty Indian Ocean water through Agulhas Leakage (De Ruijter et al., 1999; Gordon, 1986). These processes precondition the Atlantic for deep convection (e.g., Marsh et al., 2007). Yet, an important factor is the fact that the AMOC *itself* transports high-salinity water from the subtropical to the subpolar North Atlantic, hence maintaining high-salinity conditions in the convection regions. This provides a powerful feedback—called the salt-advection feedback–that has kept the AMOC strong throughout the Holocene.

However, models across the full spectrum of complexity have identified the concerning possibility that the AMOC could be prone to collapse. The basic idea is that the salt-advection feedback that currently maintains a strong AMOC can also work against it: a weakened AMOC would transport less salt into the subpolar North Atlantic, leading to reduced convection and even more weakening of the AMOC. A state would result in which no NADW is formed, and the deep Atlantic is either stagnant or experiences basinwide upwelling of abyssal waters formed around Antarctica. An AMOC collapse would have global consequences and could lead to reduction in surface air temperatures of up to 10 °C in the North Atlantic (Barreiro et al., 2008; Jackson et al., 2015; Liu et al., 2017; Manabe & Stouffer, 1988; Vellinga & Wood, 2002). Evidence of temperature change of this magnitude has been identified in paleo-proxy records (Blunier & Brook, 2001; de Abreu et al., 2003; Dansgaard et al., 1993), giving credence to the hypothesis that AMOC collapse events were responsible for rapid climate change during the Pleistocene era of ice ages (Broecker et al., 1990). The Fifth Assessment Report of the Intergovernmental Panel on Climate Change identifies an AMOC collapse as one of the tipping points in the climate system (Collins et al., 2013), with a low probability of occurrence but potentially with a high impact.

The idea that the AMOC could have more than one stable equilibrium is referred to in this paper as the AMOC stability paradigm; in the next section we briefly review the development of this paradigm. It is important to stress that this stability paradigm deals with equilibrium (steady) states. In contrast, a large body of literature considers the stability of the AMOC in terms of its transient behavior, for instance, due to oscillatory modes (e.g., Huck & Vallis, 2001; Weijer & Dijkstra, 2003; Sévellec et al., 2006), which can either destabilize the underlying equilibrium or be excited by stochastic forcing, or to nonnormal behavior, which can cause certain perturbations of a stable equilibrium to grow initially due to differential decay of eigenmodes (e.g., Germe et al., 2018; Sévellec et al., 2008; Sévellec & Fedorov, 2013, 2015, 2017; Tziperman et al., 2008; Zanna et al., 2011, 2012). Also, our discussion mostly excludes instabilities associated with changes in deep convection, for instance, deep decoupling oscillations (e.g., Cessi, 1996; Winton & Sarachik, 1993) (except for a brief discussion in section 4.2), and multiple equilibria that reflect slight changes in the region of deep convection (Lenderink & Haarsma, 1994; Vellinga, 1998).

1.2. Emergence of the AMOC Stability Paradigm

The problem of equilibrium stability of the deep ocean circulation first arose in a seminal study of Stommel (1961) who developed a simple nonlinear two-box model to describe convection in a system whose density is controlled by both temperature and salinity. Stommel showed the existence of multiple equilibria in the system and found a range of parameters yielding a bistable regime with two types of stable steady-state solutions—strong circulation or the thermal mode (the on-state corresponding to an active AMOC in the modern terminology) and reversed circulation or the haline mode (with sinking in the subtropics).

It took over two decades before the climate science community started to realize the relevance of Stommel's work, as paleoclimatologists tried to find explanations for the rapid climate swings that were being deduced from ice cores from the Greenland Ice Sheet. Rooth (1982) extended the Stommel (1961) box model by adding an extra box to account for interhemispheric flow. His analysis showed that two circulation patterns were possible, and he speculated that transitions between these states could have been triggered by the discharge of glacial lakes, hence giving rise to the rapid climate transitions associated with the Younger Dryas cold event. This concept was further explored by Broecker et al. (1985).

At the same time, numerical ocean and climate models started to come of age, and these models unambiguously confirmed that the AMOC could indeed have multiple equilibria. Bryan (1986) was the first to show the existence of two stable circulation patterns under the same forcing conditions in a three-dimensional ocean model; a result that was confirmed by Marotzke et al. (1988) in a two-dimensional model of the thermohaline circulation, by Maier-Reimer and Mikolajewicz (1989) in a global ocean model, and by Manabe

and Stouffer (1988) in a global coupled ocean/atmosphere model. Suddenly, the concept of multiple equilibrium states of the overturning circulation provided a feasible explanation for the rapid climate changes of the past and became cause for concern during a time when scientists started to realize the impact of human behavior on the climate system (Broecker, 1987).

Rahmstorf (1996) took the next step in the development of the AMOC stability paradigm. He showed convincingly that the behavior of a comprehensive global ocean model was surprisingly well reflected by a simple box model, which resembled that of Rooth (1982). He pioneered the hysteresis analysis as an experimental strategy to infer the equilibrium structure of the AMOC, a procedure that is routinely used today to probe the stability thresholds of the system. Another breakthrough occurred in the midnoughties, when De Vries and Weber (2005) and Weber and Drijfhout (2007) realized that the monostable and bistable regimes of the AMOC in a climate model of intermediate complexity were surprisingly well predicted by the sign of F_{ovS} (defined as the freshwater transport induced by the overturning circulation F_{ov}, evaluated at the southern boundary of the Atlantic, usually taken as 34°S); this analysis of F_{ovS} confirmed the suggestion of Rahmstorf (1996) that this is a key control parameter. They were able to force the AMOC from a monostable to a bistable regime simply by imposing a wind-driven freshwater flux across 33°S that forces the AMOC to transition from importing freshwater to exporting freshwater.

The idea that a simple metric like the sign of F_{ovS} would be predictive of the dynamical regime of the AMOC spurred significant activity to determine the sign of F_{ovS} in available ocean climate models. It was found that a majority of climate models feature an AMOC that imports freshwater (Drijfhout et al., 2011; Weaver et al., 2012; Weber et al., 2007). This would put most models in a regime where the AMOC is monostable. This is in contrast to the freshwater *export* that is being inferred from observations (e.g., Bryden et al., 2011; Garzoli et al., 2013; McDonagh & King, 2005; Weijer et al., 1999), which would suggest instead that the current AMOC is in a bistable regime. This has led to the speculation that the current generation of climate models exhibit systematic biases that may stabilize the AMOC (e.g., Liu et al., 2017). So even though the current generation of Intergovernmental Panel on Climate Change models predicts that an abrupt transition or collapse is very unlikely in the 21st century (i.e., less than 10% probability; Collins et al., 2013; Weaver et al., 2012), the issue of AMOC stability cannot be simply ignored. The mere possibility that the AMOC could be subject to stability thresholds should be a strong reason for concern in an era where human activity is forcing significant changes to the Earth system. Persistent model biases, imperfect parameterizations, and unresolved physics give further ample reasons to continue investigating this problem.

1.3. Why This Review

The goals of the current review are to (i) provide a comprehensive overview of the existing literature, (ii) formulate a synthesis of our current understanding of the stability of the AMOC, and (iii) identify gaps in our knowledge and propose possible ways forward to address these gaps. We will explore AMOC stability through a hierarchy of models to demonstrate the robustness of the primary instability mechanism and discuss the findings in the context of Dynamical Systems Theory (DST), which provides a unifying framework for results from a variety of approaches.

We believe that this review is timely for several reasons. First of all, climate change continues at a relentless pace and strongly affects the freshwater budget of the Arctic and subpolar North Atlantic, for instance, through impacts on sea ice cover (e.g., Sévellec et al., 2017; Stroeve et al., 2007), the Greenland Ice Sheet mass balance (e.g., Enderlin et al., 2014), and enhanced river discharge in the Arctic watershed (e.g., Peterson et al., 2002). A rigorous assessment of the resilience of the AMOC to these changes is critical, given that the AMOC is thought to be most sensitive to perturbations in the freshwater budget of the subpolar North Atlantic, and such perturbations have been implicated in slowdowns, or even shutdowns, of the AMOC in the past (Lynch-Stieglitz, 2017). Second, our understanding of the AMOC and its controls continues to evolve based on new observations (e.g., Lozier, 2010, 2012; Lozier et al., 2019) and theoretical considerations (e.g., Gnanadesikan, 1999; Nikurashin & Vallis, 2012; Wolfe & Cessi, 2009). The paradigm of the AMOC has shifted from a circulation pattern where convection-driven downwelling in the North Atlantic is balanced by diffusive upwelling in the rest of the ocean basins to one where wind-driven upwelling and eddy-driven transports in the Southern Ocean play a dominant role (e.g., Haertel & Fedorov, 2012; Toggweiler & Samuels, 1998). Currently, a new class of medium- and high-resolution ocean and climate models are coming online, which will increase our ability to model the ocean circulation in the eddying and nondiffusive regime that better supports our evolving views of the AMOC and its controls (e.g., Jackson & Wood, 2018a, 2018b;

Weijer et al., 2012). Finally, we hope that this review will provide some context for the current debate about whether the current generation of climate models is overestimating the stability of the AMOC in the context of future global warming (e.g., Gent, 2017; Liu et al., 2017).

In addition, we hope that our ambition to provide a synthesis view of the AMOC stability problem will be a useful complement to several review papers that have recently appeared and which discuss different aspects of the AMOC. Buckley and Marshall (2016) present a sweeping overview of many aspects of the AMOC, including observations of the AMOC, its role in the climate system, and mechanisms and impacts of AMOC variability. AMOC stability is discussed briefly, touching on the current debate about AMOC stability in climate models. Gent (2017) provides a critical overview of the discussion on AMOC stability in climate models, questioning the value of simple criteria to determine whether the AMOC is in a stable or bistable regime. Ferreira et al. (2018) review literature on a topic that is closely related to the problem discussed here, namely, what causes the asymmetry in deep water formation between the Atlantic and Pacific basins? This can arguably be seen as a multiple equilibria problem where several factors, including basin geometry, atmospheric water vapor transport, and possibly a salt-advection feedback, conspire to make the Atlantic the dominant basin for downwelling, even though there is tentative evidence for deep water formation occurring in both basins during the warm Pliocene epoch (Burls et al., 2017). Our review focuses mostly on AMOC stability from an Atlantic perspective, but many of the arguments apply to both the Atlantic and the Atlantic/Pacific views.

1.4. Paper Overview

In section 2, we will delve deeper into the dynamics that lead to AMOC bistability, exploring the concept of the salt-advection feedback through a hierarchy of models. Section 3 will dive deeply into the different feedbacks in the Atlantic basin that compete with the salt-advection feedback for control on the AMOC. We will assess whether the AMOC stability paradigm is still relevant for our current climate, and we will discuss the search for the "other" stability threshold. In section 4 we will review what history tells us about AMOC stability, by interpreting the paleorecord in the context of the AMOC stability paradigm. In section 5 we will be looking ahead and discuss the implications of our current understanding for the future of the AMOC in a warming climate. We will also discuss what observations can tell us about AMOC stability or possibly even an AMOC collapse. Section 6 will summarize some of the main points from this literature study and list some areas where our understanding is lacking.

2. Multiple Equilibria of AMOC in a Hierarchy of Models

2.1. The Single-Hemispheric Case

The fundamental dynamics of the AMOC stability paradigm were first presented in the visionary paper by Stommel (1961). He introduced a simple model (Figure 1a) representing circulation between two boxes, one subject to warming and salinification (later interpreted as the "subtropical" box) and the other subject to cooling and freshening ("subpolar" box). The flow strength m is proportional to the density difference between the boxes, which is controlled by the external forcing (in the figure represented by a water vapor flux F from the low- to high-latitude box) and by the exchange of properties between the boxes. Stommel showed that for certain conditions, three equilibrium solutions can coexist. One stable solution displays sinking in the subpolar box, driven by the temperature difference between the subtropical and subpolar box, while the other stable circulation pattern is driven by salinity, with sinking in the salty subtropical box. A third unstable equilibrium represents an almost stagnant circulation, with large temperature and salinity differences delicately opposing each other; a small perturbation will tip the circulation to one of the two stable states.

Two elements are crucial for the existence of multiple equilibria in Stommel's box model: advective feedbacks couple the circulation strength to the density difference that drives the flow, while differential damping timescales for temperature and salinity regulate the relative effectiveness of these feedbacks. For instance, while the equilibrium with subpolar sinking is driven by the temperature difference between the cold and warm boxes, it is braked by the salinity contrast between the salty subtropical and fresh subpolar box, imposed by the external forcing. Yet, the transport of saltier water from the subtropical to the subpolar box (and vice versa) effectively reduces the salinity difference and hence its braking effect. This is a positive
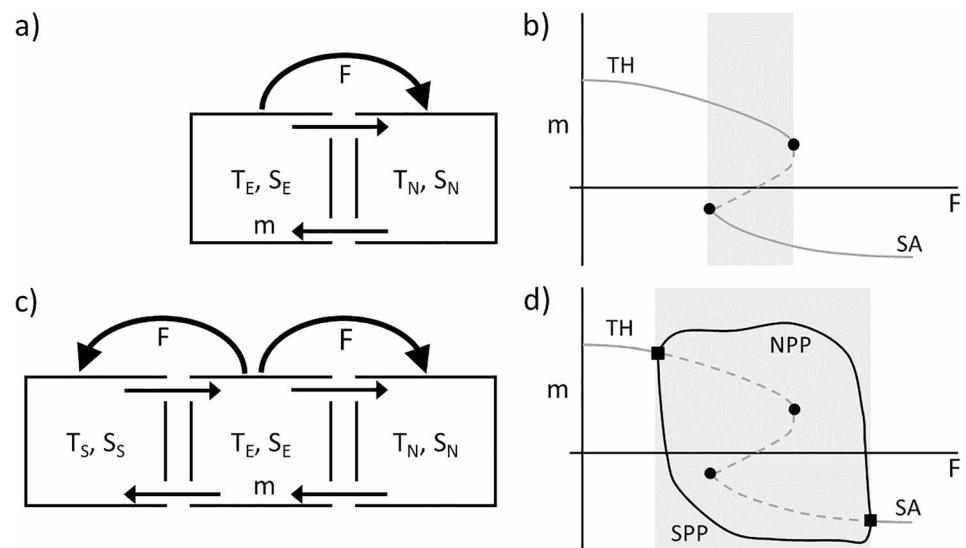
Figure 1. Schematic of the (a) Stommel (1961) two-box model and (c) the Welander (1986) three-box model, as prototypes for single-hemispheric and double-hemispheric systems, respectively. Flow strength m is proportional to the density difference between the boxes, which is determined by their salinity S and temperature T. Sketches of typical equilibrium diagrams are shown in panels (b) and (d) for single-hemispheric and double-hemispheric systems and are based on diagrams explicitly calculated in more comprehensive models by Dijkstra and Molemaker (1997) and Dijkstra et al. (2001). Control parameter is the atmospheric water vapor flux F from the low- to high-latitude box(es). Solid (dashed) lines indicate branches of stable (unstable) equilibria. Shaded areas are regions where multiple equilibria are stable. Dots represent limit points (or saddle-node bifurcations), while squares indicate pitchfork bifurcations. TH = thermally driven; SA = salinity driven; NPP = northern sinking; SPP = southern sinking.

feedback, as a stronger circulation reduces the braking effect of salinity, hence strengthening the circulation even more. The strength of this feedback is reduced to zero when the circulation is so strong that the salinity difference between the boxes is eliminated. For the same circulation, the temperature advection feedback is negative, as a stronger circulation advects more warm subtropical waters to the subpolar box, effectively reducing the temperature—and density—difference between the boxes. For the reversed circulation with sinking in the subtropical box, the feedbacks are of opposite sign, as a stronger circulation from the subpolar to the subtropical box advects fresher and colder water into the subtropical box.

If advection-induced temperature and salt anomalies are damped at the same rate, the circulation always advects lighter water into the sinking box, and the sum of the advective feedbacks is always negative. In this case, Stommel's box model only has one equilibrium, either with sinking in the subpolar or subtropical box, depending on the forcing (and the relative impact of salinity and temperature on density). However, when thermal anomalies are removed more quickly than salinity anomalies, then the positive salt-advection feedback can overcome the negative temperature advection feedback for a TH circulation. This would result in an additional densification of the subpolar box and a strengthening of the flow. But more importantly, this may make a TH circulation feasible even if external forcing conditions would suggest that only a reversed flow is possible. In other words, the salt-advection feedback extends the range of conditions for which a TH circulation can exist and causes this range to overlap with the existence of a SA circulation.

In many studies, differential damping of temperature and salinity is taken to the extreme case where temperature is strongly restored to a fixed pattern (or even prescribed), while salinity is forced by a prescribed flux and anomalies remain undamped. This combination is referred to as "mixed boundary conditions" and often used to force ocean stand-alone models (see Weaver & Sarachik, 1991, for a review). The use of a fixed surface salt flux recognizes the fact that sea surface salinity does not affect the surface freshwater fluxes (be it precipitation, evaporation, or runoff, although the salinity-dependent freezing point of sea water presents an exception in ice-covered regions). However, care needs to be taken when converting freshwater fluxes to so-called virtual salt fluxes, as is done in models that assume a fixed ocean volume. In some models, this conversion uses a fixed reference salinity, even though a proper conversion would use the local surface salinity. So even though surface freshwater fluxes are not affected by surface salinity in Nature, a proper conversion

to a virtual salt flux *would* make it dependent on surface salinity. The justification for strongly damped SSTs is the fact that SSTs are quickly communicated to the atmosphere through their impact on turbulent air/sea fluxes. However, when the ocean is coupled to an active atmosphere, the damping of thermal anomalies is additionally impacted by atmospheric feedbacks. Such feedbacks are discussed in section 3.

Many studies have expanded Stommel (1961)'s box model by adding more boxes or additional processes like rotation and wind stress (e.g., Guan & Huang, 2008; Huang & Stommel, 1992; Huang et al., 1992; Stommel & Rooth, 1968). Similarly, many studies have examined the single-hemispheric configuration in more comprehensive models (e.g., Dijkstra et al., 2001; Marotzke, 1989; Weaver et al., 1993). But the qualitative picture that stable TH and SA circulation patterns can coexist under the same forcing conditions is robust. This finding has several important implications: (i) The final equilibrium state reached by the system depends on its initial state and its history of forcing perturbations, (ii) a transition from one state to another can be triggered by a finite-amplitude perturbation, and (iii) a slight change in external conditions can move the system from the multiple-equilibria regime to a state where only one equilibrium is stable, possibly triggering a transition to another state.

2.2. The Double-Hemispheric Case With Symmetric Forcing

The most obvious problem when applying the Stommel model to the AMOC is the fact that it assumes a hemispheric circulation only, while clearly, the AMOC displays a strong cross-equatorial circulation component. Recognizing this deficiency, Welander (1986) placed two Stommel models back to back (Figure 1c). He applied forcing that is symmetric over both hemispheres, restoring temperature in both subpolar boxes to the same temperature while redistributing lower-latitude freshwater export equally over both subpolar boxes. His model allows for four stable equilibria. Two circulation patterns are symmetric and either TH with sinking in the subpolar boxes (in subsequent literature referred to as "TH") or salinity driven with sinking in the equatorial box ("SA"); these solutions are equivalent to the equilibria from Stommel's box model, albeit with a symmetric expansion toward two hemispheres. But in addition, two asymmetric circulation pole-to-pole cells exist, with either sinking in the northern subpolar box ("NPP") or southern subpolar box ("SPP"). So Welander's model allows *asymmetric* equilibria, even though the forcing is applied *symmetrically* across the hemispheres.

Rooth (1982) took a comparable approach as Welander, but his three-box model does not allow for vertical exchanges in the lower-latitude box, effectively allowing only pole-to-pole circulations. In his model (as well as in similar models of Rahmstorf, 1996; Scott et al., 1999, discussed below), the flow is determined by the density difference between the two high-latitude boxes. As in Stommel's model, the salt-advection feedback is responsible for the existence of multiple equilibria, but this feedback works differently. In the single-hemispheric framework, salt-advection from the salty subtropical box to the fresh subpolar box reduces the salinity difference between the boxes and lessens the braking effect of salinity on the TH flow. In the double-hemispheric case, salty subtropical waters are still advected into the subpolar box where sinking occurs, but this box is now saltier than the southern subpolar box with which it is being compared. So the salinity advection actually helps to *strengthen* the flow, leading to a circulation that is driven both by temperature and salinity.

The idea of spontaneous breaking of a system's symmetry, despite the application of symmetric forcing, was convincingly demonstrated by Bryan (1986), who showed that a three-dimensional, double-hemispheric ocean model, forced by symmetric boundary conditions, could exhibit asymmetric circulation patterns. These pole-to-pole circulation cells showed either sinking in the northern or southern polar region. In his model, even the symmetric circulation pattern was stable with respect to small perturbations. The symmetry breaking was subsequently shown by Marotzke et al. (1988) and Wright and Stocker (1991) in two-dimensional (latitude depth) models of the zonally averaged overturning circulation, although in these models the symmetric circulation was unstable.

The concept of AMOC stability is often cast in the framework of DST. DST studies the behavior of complex dynamical systems that are described by nonlinear differential equations. It analyzes the equilibrium states of these systems, often by mapping out branches of steady states as function of system parameters (maps known as equilibrium or bifurcation diagrams), characterizes the stability of these equilibrium states (or more generally, the transient behavior of perturbations around these steady states), and identifies the critical points where the equilibrium behavior changes qualitatively (bifurcations). The DST approach to

understanding AMOC behavior has helped tremendously in putting the results of a hierarchy of models in a comprehensive and unifying framework (see Dijkstra & Ghil, 2005; Dijkstra, 2016, for reviews).

Thual and McWilliams (1992) were among the first to explore the behavior seen in these models using concepts from DST. They systematically explored the equilibrium structure of a 2-D Boussinesq model forced with mixed boundary conditions, using the amplitudes of the temperature and salinity forcing as control parameters. Although their model could not determine unstable branches, they were able to deduce the rough outline of the equilibrium structure using heuristic arguments. Dijkstra and Molemaker (1997) confirmed the conclusions of Thual and McWilliams (1992) by explicitly calculating the branches of equilibria using a parameter continuation technique. This technique determines branches of (both stable and unstable) equilibria through iterative methods, without explicit time integration; the stability of the equilibria is determined through linear stability analysis. In fact, Dijkstra and Molemaker (1997) elegantly combined several previous studies (e.g., Cessi & Young, 1992; Marotzke et al., 1988; Quon & Ghil, 1992, 1995; Thual & McWilliams, 1992; Wright & Stocker, 1991) in a unifying framework by showing how changes in parameter values affect the bifurcation structure in subtle ways and hence the behavior of the transient responses in models that only allow forward time integration.

A typical bifurcation diagram of the single-hemispheric case is sketched in Figure 1b (after Dijkstra et al., 2001). The control parameter is the strength of the surface freshwater flux F, with a pattern that transports water vapor from low to high latitudes. For small values of F, only a TH state is stable, with sinking at the pole, and upwelling in the equatorial region. With increasing F, this branch of equilibria encounters a limit point (indicated by circles), where the branch "turns back" and loses stability. Along this branch, the solution transitions from a TH to a SA circulation, now with downwelling at the equator. The branch encounters another limit point, regaining stability as a SA state. These branches of equilibria reflect the equilibrium structure of the Stommel (1961) box model, with the limit points delimiting Stommel's region of multiple equilibria.

For the double-hemispheric case, a typical diagram is sketched in Figure 1d (after Dijkstra & Molemaker, 1997). Again, the control parameter is the strength of the surface freshwater flux F, with a pattern that symmetrically moves water vapor from low to high latitudes, to retain the hemispheric symmetry of the system. The branch of symmetric equilibria (indicated by the gray lines) is equivalent to the equilibria of the single-hemispheric case, sketched in Figure 1b. However, the branch of symmetric TH solutions becomes unstable through a so-called pitchfork bifurcation (indicated by squares). Such bifurcations are the points where two mirror-symmetric equilibria arise in a system that has reflective symmetry. Here, these circulation states are the NPP and SPP sinking pole-to-pole circulations. These NPP and SPP branches coexist for a significant range of F values and connect to the SA branch at another pitchfork bifurcation.

Even though the quantitative details of the equilbrium structure strongly depend on model details, the qualitative aspects have proven to be surprisingly robust. Vellinga (1996) showed equivalency between the equilibrium structure in a 2-D Boussinesq model and in a 2-D model in which the three-dimensional rotational momentum equations are zonally averaged (as in Wright & Stocker, 1992), while Weijer and Dijkstra (2001) showed equivalency between a two-dimensional model and a three-dimensional model, with planetary rotation and winds. What is more, Weijer and Dijkstra (2001) explicitly showed that it is the salt-advection feedback that is responsible for destabilizing the symmetric TH circulation in favor of the pole-to-pole circulation in both configurations, consistent with the dynamics represented by box models.

2.3. The Impact of Asymmetries on Equilibria in a Double-Hemispheric Basin

The study of the symmetric double-hemispheric case has been instrumental in our qualitative understanding of the multiple equilibria in more complex models. Once asymmetries are introduced to the double-hemispheric case, the pitchfork bifurcations break up, and the originally symmetric solutions connect to the preferred pole-to-pole solution (Figure 2). In particular, Dijkstra and Neelin (2000) showed that the greater continental area in the Northern Hemisphere and a slight asymmetry in the surface freshwater flux induce a strong preference for the northern sinking solution, while Weijer et al. (2001) showed that the buoyancy fluxes brought about by interocean exchange with the Southern Ocean also favor the NPP. Dijkstra et al. (2003) performed a similar analysis in a three-dimensional model of the double-hemispheric circulation in the Atlantic and concluded similarly that the presence of a circumpolar channel, Southern Ocean winds, and variable basin width favors the northern sinking solution. In fact, they argue that a likely source
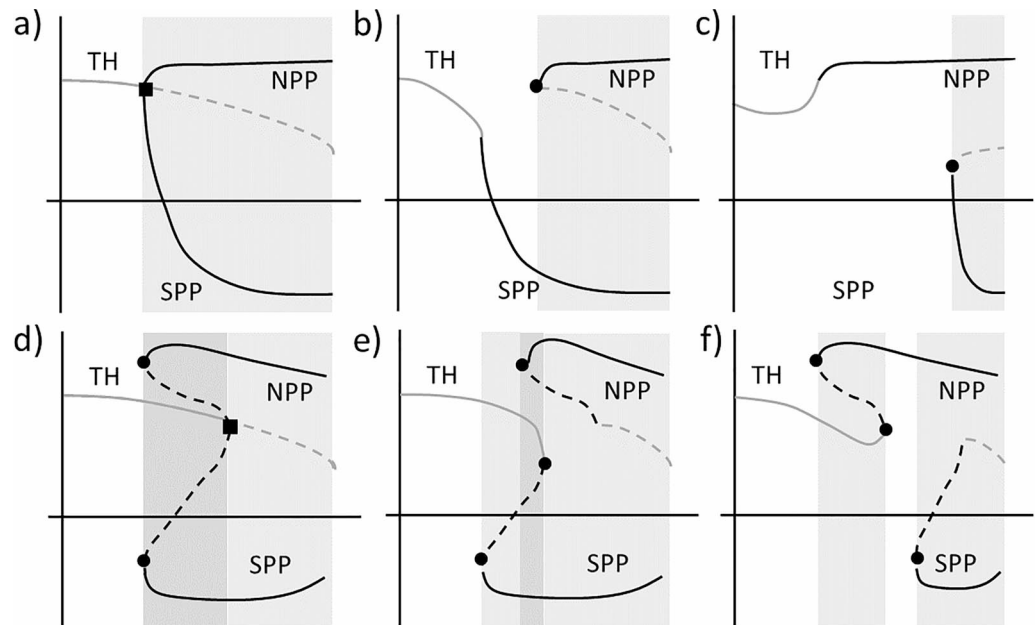
Figure 2. Schematic showing how an asymmetry in the model can break up a pitchfork bifurcation. (a, d) Tthe pitchfork bifurcation of an unperturbed, symmetric double-hemispheric model, with (a) a supercritical case (cf. Figure 1d) and (d) a subcritical case. The latter situation can lead to a regime with three stable equilibria (dark gray shading in panels d and e; e.g., Bryan, 1986). When the symmetry is broken, the TH branch connects to either (b, e) the SPP or (c, f) the NPP branch. Dependent on the degree of asymmetry, regimes with one, two, or three stable equilibria can be found. TH = thermally driven; NPP = northern sinking; SPP = southern sinking

of rapid climate change may have been a transition between a strong northern sinking pole to pole circulation and a weakly asymmetric version of the TH circulation, featuring downwelling at both poles. They claimed that the asymmetries were strong enough to make the southern sinking pole-to-pole circulation disappear, leaving the NPP state the unique stable state for a significant parameter regime.

Asymmetries have also been explored in the context of box models. In particular, Rahmstorf (1996) and Scott et al. (1999) extended the Rooth (1982) box model by allowing the thermal and haline forcing of the polar boxes to differ. As in the original model of Rooth (1982), the circulation strength is proportional to the density difference between the polar boxes, and symmetric circulation patterns (TH and SA) are excluded. Comparison between the single- and double-hemispheric cases is insightful. In Stommel (1961), the polar sinking solution is driven by temperature, and the salt-advection feedback acts to reduce the braking effect of the meridional salinity contrast. In the double-hemispheric case, on the other hand, the northern sinking solution is only thermally driven when the forcing temperature of the northern box is lower than that of the southern box. If the prescribed temperatures are equal, then the northern box may be warmer than the southern box, due to the advection of midlatitude heat. Either way, salt-advection drives the overturning, and the main control on the strength of the circulation is the atmospheric water vapor flux out of the low latitude and into the southern box. In the model of Scott et al. (1999) it is directed southward, to account for net freshwater export from the Atlantic north of 30°S. In Rahmstorf (1996), on the other hand, this flux is directed northward. Rahmstorf did not interpret this flux as the actual water vapor exchange between the Atlantic and the rest of the World Ocean but as an "active" flux that the AMOC sees when gyre-driven freshwater import is accounted for. Indeed, Weijer et al. (1999) showed that the overturning exports freshwater and that the South Atlantic gyre circulation compensate for both this AMOC-driven export and for the excess evaporation out of the Atlantic north of 30°S. Both studies argue that the intrabasin vapor transport from the low to northern high latitudes does not affect the strength of the AMOC (although it may affect the stability with respect to oscillatory modes; Scott et al., 1999), which is in clear contrast to the single-hemispheric case.

It is worth noting that pole-to-pole overturning circulation is also proposed for isopycnals that outcrop in the North Atlantic and the Southern Ocean even when the interior ocean diffusivity is small (Wolfe & Cessi, 2009). This adiabatic flow regime is supported by wind-driven upwelling, and in this case, the density difference between the North Atlantic and Southern Ocean reduces the shared isopycnal outcrop between

the poles and therefore weakens the circulation. To collapse such circulation, it is necessary that buoyancy fluxes no longer allow for shared isopycnals between northern subpolar waters and the Southern Ocean. For this type of circulation, it appears no longer obvious that Stommel's salt feedback and the resulting multiple equilibria would act the same way as for a circulation powered by buoyancy differences and diapycnal mixing. Yet, salt-advection feedback, whereby subtropical salty water is transported into the subpolar North Atlantic, is found to affect the strength of the overturning by changing the size of shared isopycnal window between the hemispheres (Wolfe & Cessi, 2014). Indeed, this salt-advection feedback, in combination with asymmetrical surface freshwater flux, could give rise to multiple overturning circulation states (Wolfe & Cessi, 2015). The conclusion is that even the turbulent quasi-adiabatic overturning can feature an on-state and off-state, which is also confirmed by Haertel and Fedorov (2012)'s computations.

2.4. Extension to Global Geometries

The Atlantic Ocean does not operate alone but is connected to the Pacific and Indian Oceans through the Arctic and the Southern Oceans. Multiple equilibria have been found in all multibasin configurations, ranging from models with two connected basins (Atlantic and Pacific only; e.g., Barreiro et al., 2008; Hughes & Weaver, 1994; Huisman et al., 2009; Jones & Cessi, 2016; Marotzke & Willebrand, 1991; Nilsson et al., 2013; Stocker & Wright, 1991a, 1991b) and three connected basins (Atlantic, Pacific, and Indian; e.g., Stocker et al., 1992a, 1992b; Wright & Stocker, 1992) to models with detailed representation of the global ocean geometry and bathymetry (e.g., Maier-Reimer & Mikolajewicz, 1989; Power & Kleeman, 1993; Rahmstorf, 1996). In general, in simplified model configurations, equilibria are found that are characterized by sinking in the Southern Ocean ("Southern Sinking"), downwelling in both the North Atlantic and North Pacific ("Northern Sinking"), downwelling in the North Atlantic only ('Conveyor'), or downwelling in the North Pacific only ('Inverse Conveyor'). Huisman et al. (2009) showed that these equilibria form two disconnected branches, with the conveyor belt circulation being connected to the southern sinking mode, while the inverse conveyor belt circulation is connected to the northern sinking solution. So in a sense, regardless of the situation in the Pacific, both branches represent transitions between NPP and SPP solutions in the Atlantic, in response to North Atlantic freshwater perturbations. Nilsson et al. (2013) argued that basin asymmetries may impede any circulation patterns with sinking in the North Pacific, while interbasin water vapor transport may play a role as well (e.g., Emile-Geay et al., 2003; Ferreira et al., 2010; Warren, 1983; Weyl, 1968; Zaucker et al., 1994). This would leave the Conveyor and Southern Sinking solutions as the only viable equilibria. It is not clear whether in a realistic setting, any equilibria exist that are dynamically linked to the equatorially symmetric TH and SA circulation patterns.

In an influential study, Rahmstorf (1996) pioneered the hysteresis approach to infer the equilibrium structure of the AMOC and its associated thresholds. In this approach, a perturbation in surface forcing is applied, and its amplitude is increased very slowly, so that the AMOC is assumed to remain close to the actual equilibrium for a given forcing state (Figure 3a). For some value of the forcing perturbation amplitude, the AMOC starts to collapse, marking its transition to the collapsed equilibrium state. When that equilibrium has been reached, the forcing perturbation is typically reduced in strength, until the AMOC starts to spin-up again, marking the transition from the collapsed to the active AMOC state. The surface forcing perturbation is usually a freshwater perturbation in the North Atlantic, an approach that is referred to as "hosing," although other forcing such as the Southern Ocean winds could also create a hysteresis of the overturning (Sévellec & Fedorov, 2011). Rahmstorf (1996) compared the resulting hysteresis structure with the equilibria of a simple Rooth-like box model (Rooth, 1982), showing that the AMOC response to North Atlantic freshwater perturbation in the full-blown ocean general circulation model was surprisingly similar to the behavior of the box model.

In a series of papers, Dijkstra and Weijer (2003), Weijer et al. (2003), and Dijkstra and Weijer (2005) explicitly determined the equilibrium structure of the global ocean circulation that was inferred by Rahmstorf (1996). Figure 4 shows that for $\gamma < L_2$ only one equilibrium exists, namely, a strong AMOC. For $L_2 < \gamma < L_1$ two stable equilibria exist: a strong AMOC state (equilibrium A) and a collapsed AMOC state or rather a state with downwelling in the Southern Ocean and upwelling in the Atlantic and Pacific Oceans (equilibrium B). For $\gamma > L_1$ only the collapsed state exists. The two thresholds of γ that separate the three regimes are the limit points introduced in section 2.2 (see Figures 1b and 1d). Dijkstra and Weijer (2003) showed, by analyzing a hierarchy of models, from box models to global ocean general circulation models (Dijkstra & Weijer, 2005; Weijer et al., 2003) that the hysteresis behavior, as found by Rahmstorf (1996), reflects a transition between
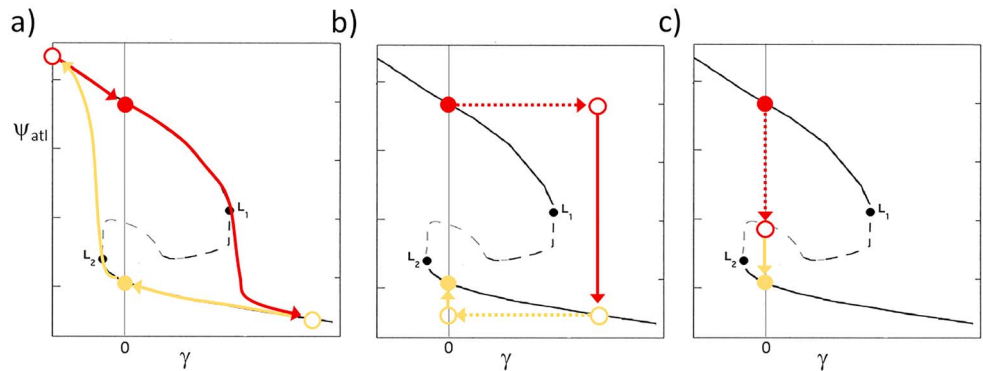
Figure 3. Schematic illustrating several strategies to probe for bistability, in the context of the equilibrium diagram in Figure 4. Suppose that for unperturbed conditions ($\gamma = 0$), the state with active Atlantic Meridional Overturning Circulation (AMOC; solid red dot) is in a bistable regime. Panel (a) shows the traditional hysteresis experiment (Rahmstorf, 1996), where the forcing γ is gradually increased, allowing the system's state (solid red curve) to follow the branch of equilibria. After passing the threshold defined by limit point L_1, the AMOC quickly collapses. After reaching the branch of "off-states," γ is slowly reduced again (solid orange curve), until it reaches the collapsed state under unperturbed conditions ($\gamma = 0$; solid orange dot). When the forcing perturbation is reduced even further, the threshold defined by limit point L_2 is passed, and the system transitions to the branch of active AMOC states. Panel (b) schematically shows the procedure of the classical hosing experiment (e.g., Jackson & Wood, 2018a, 2018b), where an instantaneous surface forcing perturbation is applied that moves the initial state into the monostable regime of off-states. Once the system reaches the stable branch of off-states (collapsed AMOC), the forcing perturbation is removed, putting the system back in the bistable regime of $\gamma = 0$. Panel (c) displays yet another strategy (e.g., Mecking et al., 2016), where the initial state is perturbed, that is, by redistributing salt in the ocean interior, leaving the external forcing unchanged ($\gamma = 0$). If the perturbation is large enough, the AMOC will collapse, and the system will transition to the off-state.
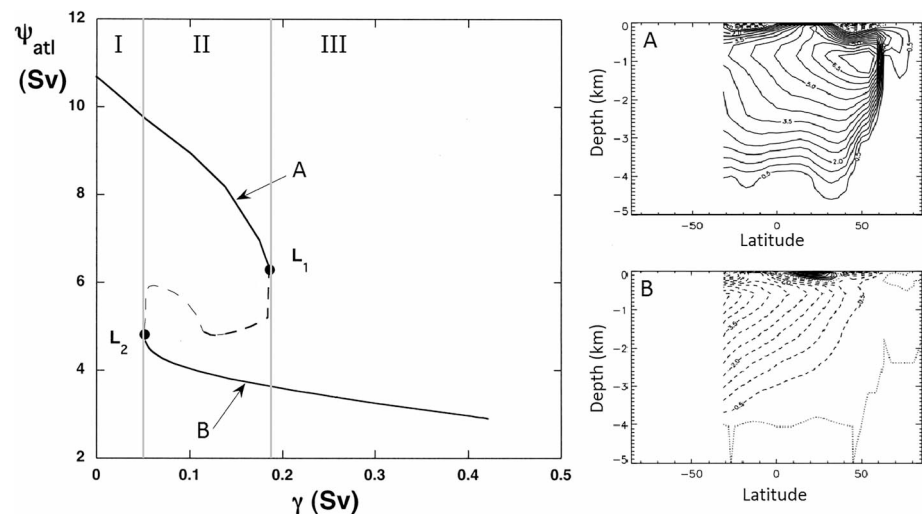
Figure 4. Equilibrium diagram of the Atlantic Meridional Overturning Circulation (AMOC) in a global ocean circulation model. Control parameter γ represents the amplitude of an excess freshwater flux perturbation in the subpolar North Atlantic, while Ψ_{atl} is the maximum of the overturning stream function in the Atlantic. Solid (dashed) lines indicate branches of stable (unstable) equilibria. L_1 and L_2 indicate the location of limit points that separate the three stability regimes. Regimes I and III are monostable, where only active (I) or collapsed (III) AMOC states are stable. Regime II is bistable, where both active (a) and collapsed (b) AMOC states are stable equilibria. The overturning stream function in the Atlantic for equilibria at (a) and (b) are shown in the right panels. Modified from Weijer et al. (2003), reproduced with permission from Elsevier, copyright 2003.

NPP and SPP states. The AMOC collapse is hence a transition toward a state with dominant sinking in the Southern Ocean and upwelling in the rest of the ocean.

2.5. The Global Fully Coupled Case

Multiple equilibria have also been found in climate models in which the ocean is coupled to an active atmosphere model. In fact, quickly after the discovery by Bryan (1986) that the AMOC in his double-hemispheric model allowed for multiple equilibria, Manabe and Stouffer (1988) reported the existence of multiple equilibria in a global AOGCM. The initial spin-up of their model resulted in an ocean circulation without deep water formation in the North Atlantic and only a weak upwelling cell ("off-state"). Realizing that the North Atlantic was too fresh, they applied a correction to the surface freshwater flux that was diagnosed from a simulation where the surface salinity was strongly restored to climatology. With this flux correction, the model simulated a strong AMOC, but they found that the off-state remained stable as well. They concluded that their fully coupled model—with flux correction—displayed two modes of operation, while without this correction, only the off-state was a stable equilibrium.

Subsequent studies have delivered mixed results, with several studies confirming the existence of multiple equilibria (e.g., Liu et al., 2013; Manabe & Stouffer, 1999), while others did not (e.g., Schiller et al., 1997). A set of 11 reduced-complexity Earth system models showed significant hysteresis loops with respect to North Atlantic freshening (Rahmstorf et al., 2005) and so did the FAMOUS (Hawkins et al., 2011) and CCSM3 (Hu et al., 2012a) models, which—to the best of our knowledge—are the most complex CGCMs on which a hysteresis analysis has been performed to date. Quasi-steady multiple equilibria have also been found in a recent generation, eddy-permitting CGCM (Jackson & Wood, 2018a, 2018b; Mecking et al., 2016; although the simulations are not sufficiently long to tell whether these are true equilibria). In contrast, the majority of 14 CGCMs of different complexity did not display an irreversible shutdown in response to coordinated hosing experiments (Stouffer et al., 2006). Still, absence of evidence for multiple equilibria should not be taken as evidence for their absence, as it is hard to make definite statements about AMOC equilibria and their stability based on transient model experimentation alone.

There may be several reasons for this diversity in model behavior. First, several studies suggest that the current generation of CGCMs may suffer from biases that hinder the models' ability to show AMOC multiple equilibria (Hawkins et al., 2011; Huisman et al., 2010; Liu et al., 2014, 2017; Mecking et al., 2017; Weaver et al., 2012). Several of these studies point out that GCMs often have a fresh bias in the South Atlantic that affects the amplitude, if not the sign, of F_{ovS} and hence the stability regime in which the modeled AMOC resides. This suggests that the stability characteristics of the modeled AMOC can be improved by improving model representation of processes that affect the salinity field in the South Atlantic (e.g., Agulhas Leakage; Weijer & van Sebille, 2013).

Second, several studies show how the AMOC moves between monostable and bistable regimes when these biases are corrected through flux adjustments. For example, Yin and Stouffer (2007) compared the stability of the AMOC in two consecutive releases of the GFDL model and found that only the older, flux-corrected configuration showed bistability. Similarly, Liu et al. (2017) moved the AMOC from a monostable to a bistable regime by applying flux adjustments to correct salinity biases in the Atlantic. However, flux adjustments should be applied with caution: Dijkstra and Neelin (1999) showed how the flux correction procedure, as applied by, for instance, Manabe and Stouffer (1988), could artificially affect the stability of the AMOC in models; the flux-corrected configuration could have multiple equilibria in a regime where the non-flux-corrected configuration would have a single equilibrium. That multiple equilibria can also be found in models that do *not* have flux adjustments have been shown in more recent studies (Hawkins et al., 2011; Hu et al., 2012a; Jackson & Wood, 2018a, 2018b; Mecking et al., 2016).

Third, the recent generation climate models have improved process representations, possibly introducing additional feedbacks that counteract the oceanic salt-advection feedback and modulate modeled AMOC stability (e.g., Vellinga & Wood, 2002; Yin et al., 2006). If indeed other feedbacks exist that dominate the salt-advection feedback, they can reduce the width of, or even eliminate altogether, the regime of multiple equilibria (e.g., Hofmann & Rahmstorf, 2009).

In addition to these reasons related to model structure, conclusions may depend on the experimental procedure that is used to probe AMOC stability. Apart from explicitly calculating the equilibrium structure (e.g., Weijer et al., 2003), which can be done only for a few specialized models, hysteresis analysis (Figure 3a)

may be the most reliable procedure. However, the procedure is expensive and can only be applied routinely to reduced-complexity models (e.g., Ganopolski & Rahmstorf, 2001; Rahmstorf, 1995; Rahmstorf et al., 2005). More comprehensive CGCMs, instead, have relied mostly on "hosing" (Figure 3b) or "perturbation" (Figure 3c) experiments. For these procedures it should be kept in mind that the AMOC may be quite resilient to finite-amplitude perturbations. Dijkstra et al. (2004) have shown that the AMOC can still recover after a perturbation is applied for hundreds of years, even if the off-state is known to be an equilibrium of the system (see also ; De Vries & Weber, 2005). So a recovery of a strong AMOC after a temporary hosing perturbation is not evidence that a steady alternative equilibrium does *not* exist. Similarly, a coupled model can maintain an off-state for a long time before eventually recovering an active AMOC (e.g., Stouffer & Manabe, 2003).

AMOC off-states, loosely defined as states without a significant Atlantic meridional overturning cell, can differ substantially between models and even scenarios, and some possibilities have already been discussed in the previous sections. Mecking et al. (2016) have shown that a quasi-steady off-state with no significant overturning circulation can be obtained in the latest generation of coupled climate models. However, theories suggest that in a steady state the global diapycnal mixing (Rahmstorf, 2003) and wind-driven upwelling in the Southern Ocean (de Boer et al., 2008; Kuhlbrodt et al., 2007) must be balanced by sinking somewhere. This would imply that a state without a significant overturning circulation for both the AMOC and the Antarctic Bottom Water cell cannot be an equilibrium and that at some point, possibly after millennia, a reduction in one overturning cell has to be compensated by an increase in the other. But even though a true off-state may not technically be an equilibrium, an off-state that is maintained for centuries or even millennia will have large climate impacts (see section 4.1).

Models have shown off-states where the Atlantic overturning is reversed with no North Atlantic sinking and with stronger Antarctic Bottom water or Antarctic Intermediate water cells (Gregory et al., 2003; Manabe & Stouffer, 1988, 1999; Marotzke & Willebrand, 1991; Weijer et al., 2003). Others have found sinking in the North Pacific forming a Pacific Meridional Overturning Circulation cell (Liu & Hu, 2015; Marotzke & Willebrand, 1991; Saenko et al., 2004). Yin and Stouffer (2007) suggested that a reverse Atlantic circulation maintained the off-state by importing freshwater needed to counteract the salinification in the North Atlantic due to the southward shift of the Intertropical Convergence Zone (ITCZ). Jackson et al. (2017) showed that the same model can achieve different off-states depending on how the forcing is applied. They suggested that the type of off-state might affect how stable it is.

To summarize, the literature review presented in this section has shown that (i) multiple equilibria have been found in the full hierarchy of models, ranging from box models to comprehensive climate models, (ii) the process responsible for these multiple equilibria is consistently found to be the salt-advection feedback, and (iii) the regime of multiple equilibria (i.e., its width and location in parameter space) is very sensitive to model details. In the following section we will discuss in more detail the different feedbacks that may affect the stability of the AMOC.

3. Feedbacks Controlling AMOC Stability

3.1. F_{ovS} as Stability Indicator

As discussed in the previous section, the stability of the AMOC has been extensively examined via freshwater hosing experiments (e.g., Stouffer et al., 2006), hysteresis diagrams (e.g., Hu et al., 2012a; Rahmstorf, 1996), and bifurcation diagrams (e.g., Dijkstra & Weijer, 2005; Dijkstra et al., 2004). Nevertheless, to evaluate the stability of the AMOC in a complex CGCM and more importantly in nature using available observations, one wishes to formulate a diagnostic indicator of AMOC stability. Rahmstorf (1996) suggested that F_{ovS}, the meridional overturning component of freshwater transport across the Atlantic southern boundary (around 34°S), may be such an indicator: he argued that, if the AMOC imports freshwater into the Atlantic ($F_{ovS} > 0$), the AMOC would be driven by both temperature and salinity and would be in a monostable regime. If, instead, the AMOC exports freshwater ($F_{ovS} < 0$), haline forcing would oppose the circulation, and a reverse SA circulation would exist as an alternative equilibrium. Studies with GCMs found the condition $F_{ovS} = 0$ to be a remarkably accurate indicator of the boundary between the monostable and bistable regimes (Cimatoribus et al., 2012; De Vries & Weber, 2005; Drijfhout et al., 2011; Hawkins et al., 2011; Jackson, 2013a; Mecking et al., 2017; Weber & Drijfhout, 2007). In fact, several studies with comprehensive GCMs showed

that the stability characteristics of the AMOC can be altered by constraining F_{ovS} to be positive or negative, as predicted by box models (Cimatoribus et al., 2012; De Vries & Weber, 2005; Jackson, 2013a; Liu et al., 2017).

The indicator was later refined by including the freshwater transport across the northern boundary to consider the effect from the Arctic, hence turning it into a divergence indicator ($\Delta F_{ov} = F_{ovS} - F_{ovN}$, where F_{ovN} is F_{ov} evaluated at the northern boundary of the Atlantic, often taken to be 60°N; Dijkstra, 2007; Huisman et al., 2010; Liu & Liu, 2013, 2014; Liu et al., 2014, 2017). Dijkstra (2007) and Huisman et al. (2010) in particular use parameter continuation techniques to show that the limit point that separates the monostable from the bistable regime corresponds exactly with the zero crossing of the divergence indicator ΔF_{ov}, while F_{ovS} by itself is a less accurate predictor. Why inclusion of the northern overturning-induced freshwater flux should improve the indicator is not obvious and may need some more scrutiny. Several studies show a connection between the freshwater transport through the Bering Strait and the strength of the AMOC (e.g., Hu et al., 2008, 2011), which would constitute a stabilizing feedback, as discussed in section 3.4 below. But the connection between this freshwater flux and F_{ovN} is not obvious.

The value of F_{ovS} as a stability indicator is the fact that it can readily be inferred from observations. Estimates of F_{ovS} in the South Atlantic made from WOCE hydrography (Weijer et al., 1999), GO-SHIP hydrography (Bryden et al., 2011; McDonagh & King, 2005), or a synthesis of XBT and Argo observations (Garzoli et al., 2013) all give negative values, ranging from −0.28 to −0.05 Sv, indicating that the salinity flux associated with the overturning circulation exports freshwater from the Atlantic. The overturning salinity flux is primarily a balance between the northward flowing upper limb and the southward flowing NADW in the lower limb. The southward flowing NADW is slightly more saline than the section average salinity of 34.864 PSU, whereas the northward flowing upper limb has an AAIW component (green in Figure 5) that is up to 0.5 PSU fresher than the NADW and a surface water component (red in Figure 5) that is up to 1.5 PSU more saline than NADW. It is the relative balance between the fresh and saline components of the upper limb that determine whether the transport weighted salinity in the upper limb is more or less saline than NADW and therefore whether F_{ovS} is negative or positive. Bryden et al. (2011) associate the fresh component of the upper limb with AAIW and water masses from the Drake Passage (often referred to as the cold water route; Gordon, 1986) and the more saline component with input from the Indian Ocean via the Agulhas System (warm water route).

It has been well established that in many current generation coupled climate models F_{ovS} tends to have a positive bias; hence, the AMOC is not exporting enough freshwater from the Atlantic (Cheng et al., 2018; Liu et al., 2014, 2017; Mecking et al., 2017; Weaver et al., 2012). The upper 800 m of the South Atlantic, where the transport is northward, tends to be too fresh, while at depth, where the transport is predominantly southward, the ocean is too salty leading to a positive bias in F_{ovS}. Through correction of the salinity bias, the majority of the fifth Coupled Model Intercomparison Project (CMIP5) models achieve an F_{ovS} at 34°S which falls within observational estimates (Mecking et al., 2017). The salinity bias has been shown to be related to the net precipitation in the Atlantic, with many CMIP5 models having too much evaporation/too little precipitation in the Atlantic resulting in too fresh waters outside the Atlantic; this leads to overly fresh inflow into the Atlantic in the upper ocean at 34°S (Mecking et al., 2017). It has been suggested that the bias in surface freshwater fluxes is caused by inaccurate representations of the Atlantic ITCZ (Liu et al., 2014). One method for reducing the salinity bias is through the use of surface flux adjustment. The use of fresh water flux adjustments causes F_{ovS} to become (more) negative (Jackson, 2013a; Liu et al., 2014, 2017). This allows the AMOC to exhibit bistable behavior (Liu et al., 2017; Yin & Stouffer, 2007) or to sustain an—ultimately unstable—off-state for a longer period of time (Jackson, 2013a). However, the use of such flux adjustments, although illustrative, should be considered with care, as it might unphysically impact AMOC stability (Dijkstra & Neelin, 1999; Gent, 2017). The focus should be on improving the physical processes that are responsible for the salinity biases in the South Atlantic.

3.2. The Basin-Scale Salt-Advection Feedback

The physics behind F_{ovS} and ΔF_{ov} is a salt-advection feedback that works on the scale of the entire Atlantic, rather than on the North Atlantic alone, as in Stommel (1961). Suppose an active AMOC produces a climate state with an Atlantic freshwater divergence ($\Delta F_{ov} < 0$). An initial perturbation in the North Atlantic (caused by either a freshwater discharge or a warming associated with an increase of atmospheric CO_2) will weaken the AMOC and, in turn, the associated freshwater divergence. The reduced freshwater divergence will then lead to an accumulation of freshwater within the North Atlantic, amplifying the initial perturbation (i.e.,
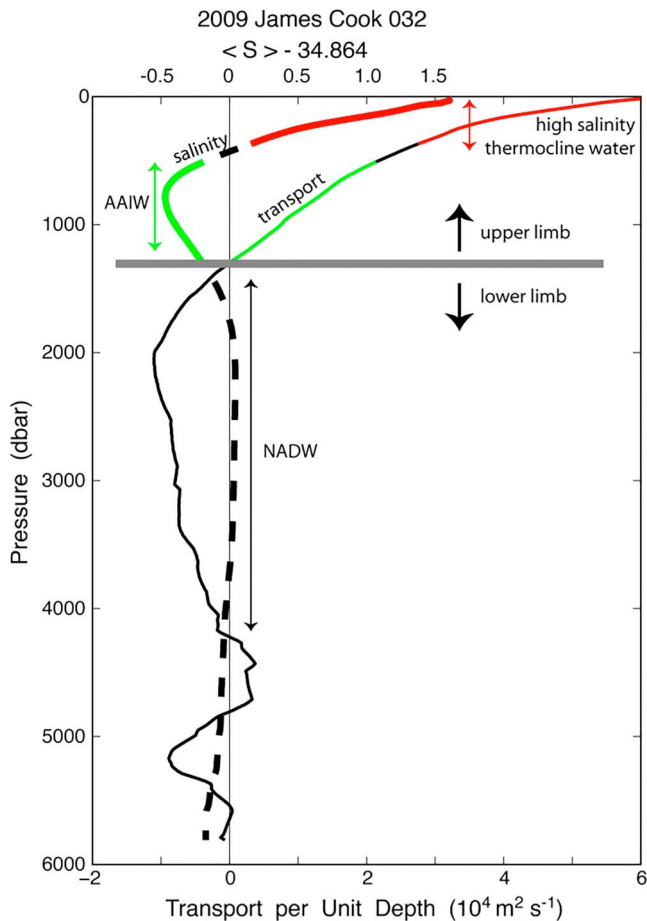
Figure 5. The zonally averaged salinity from the 2009 occupation of the 24S Atlantic section (dashed line) expressed as an anomaly from the section average salinity of 34.864. The zonally cumulated transport per unit depth for the same section (solid line). The green sections of the curves highlight the depth range occupied by Antarctic Intermediate Water (AAIW), and the red sections highlight the highest salinity thermocline water that has its source in the Indian Ocean. From the analysis of Bryden et al. (2011). NADW = North Atlantic Deep Water.

a positive feedback) and possibly leading to a collapse of the AMOC (Liu & Liu, 2013, 2014). Thereby, the AMOC can switch from the "on" to "off" state. In contrast, for an AMOC with freshwater convergence ($\Delta F_{ov} > 0$), the initial AMOC weakening will reduce freshwater convergence and increase the salinity of the Atlantic, facilitating deep mixing and preventing a further weakening of the AMOC (i.e., a negative feedback). In this case, the AMOC tends to recover to its original state. Hence, the sign of F_{ovS} or ΔF_{ov} determines the sign of the salt-advection feedback and therefore whether the AMOC can become unstable or not (Huisman et al., 2010). Still, it is unclear how this condition, which is a characteristic of the flow of the active AMOC (and not an externally imposed constraint), is able to predict the presence or absence of a collapsed state.

Several studies search for evidence of the salt-advection feedback in transient simulations. Jackson (2013a) explored the advective feedback during hosing experiments with a climate model with $F_{ovS} > 0$ and the same model flux adjusted to give $F_{ovS} < 0$. They found that, although the AMOC in both models recovered, the one with $F_{ovS} < 0$ recovered more slowly because the advective feedback freshened the South Atlantic and this anomaly propagated northward into the subpolar North Atlantic. However, this feedback was not sufficient to stop the recovery of the AMOC. De Vries and Weber (2005) also found that the AMOC recovered after hosing, even when $F_{ovS} < 0$; however, when they applied the hosing for longer, they found that other stabilizing feedbacks were overcome by the advective feedback. Hence, it seems that the advective feedback can become more important on longer timescales and with stronger perturbations.

Den Toom et al. (2014) compared the response to hosing in the North Atlantic in a low- and high-resolution version of the same ocean model for a period of 45 years. The anomalous freshwater flux leads to a comparable reduction in the AMOC in both configurations. In the low-resolution configuration, the basin-scale salt-advection feedback is clearly active, as F_{ovS} (which is positive in the reference state) becomes smaller in response to a weakening of the AMOC. In the high-resolution configuration F_{ovS} is also found to decrease from an initially positive value, but the interpretation is less straightforward than in the low-resolution case. den Toom et al. (2014) find that, despite the decline in F_{ovS}, the component associated with the weakening of the AMOC (acting on a fixed reference salinity profile) is slightly *increasing*, possibly reflecting a change in the vertical structure of the AMOC. Within the 45-year time span, the decline in F_{ovS} is therefore a consequence of a freshening of the deep ocean, not of the weakening of the AMOC.

Cheng et al. (2018) searched for the elements of the salt-advection feedback in internal variability of the AMOC in multicentennial simulations of two climate models. The two models show that, while AMOC variability controls F_{ovS} to a certain degree, this influence is much weaker than contributions from local salinity variation. But most significantly, no evidence was found that the natural variations in F_{ovS} had a significant impact on the stratification in the Atlantic, let alone the density in the subpolar North Atlantic. Cheng et al. (2018) concluded that stronger perturbations are required to activate and detect the salt-advection feedback (for instance those imposed by hosing experiments).

3.3. The Role of the Gyre Component F_{az}

In hosing studies where the North Atlantic is flushed by large amounts of freshwater, strong responses in F_{az} at the southern boundary of the Atlantic have been reported, sometimes larger than the response in F_{ovS} (Cimatoribus et al., 2012; Jackson, 2013a; Mecking et al., 2016). As a result it was questioned whether F_{ovS} was a correct stability indicator, as the change in freshwater transport by the gyre could be as large or even larger than the change in freshwater transport by the overturning circulation (Gent, 2017). Huisman et al. (2010), however, pointed out that the importance of F_{ovS} as stability indicator reflects its dynamical role as a
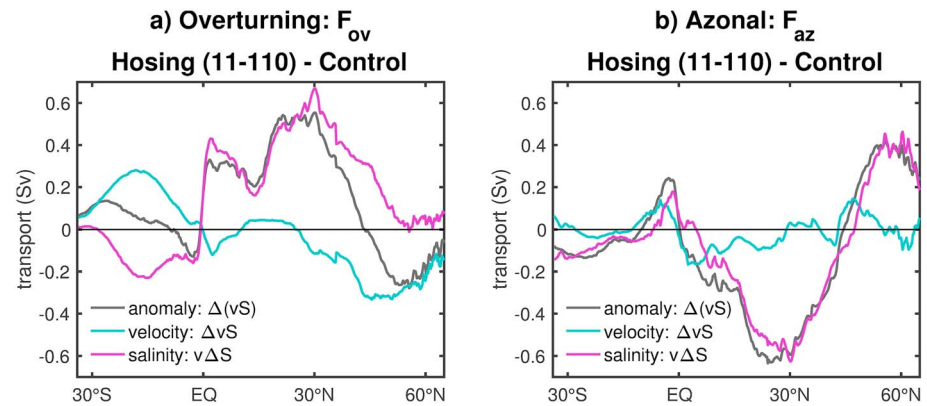
Figure 6. (a) The anomalous freshwater transport due to the Atlantic Meridional Overturning Circulation F_{ov}, here shown for all latitudes in the Atlantic and (b) azonal component, F_{az} at all latitudes in the Atlantic, in the first 100 years after hosing is stopped in a coupled climate model. The total anomalous freshwater transport, $\Delta(vS)$, is shown in gray, the velocity change contribution to the anomalous transport, ΔvS, in turquoise, and the change in salinity contribution to the anomalous transport, $v\Delta S$, in magenta. From the analysis of Mecking et al. (2016).

feedback which is either negative ($F_{ovS} > 0$) or positive ($F_{ovS} < 0$) and can destabilize the AMOC regardless of the other terms in the Atlantic freshwater budget.

The large anomalous salt transports in gyre transport mainly arise due to the anomalous salinity in the surface waters that quickly progresses southward (Jackson & Wood, 2018a, 2018b; Mecking et al., 2016). Such transport terms, however, do not control the stability of the AMOC as they tend to dampen out the initial salinity anomalies; that is, they are a negative feedback (Huisman et al., 2010). Without the presence of even larger transport anomalies due to changes in velocity, Huisman et al. (2010) showed that any transport anomaly associated with changes in salinity at 34°S (irrespective of whether it is part of F_{ovS} or F_{az}) always drives the Atlantic back to its original state until the salinity anomaly is zero again. A different state (with nonzero salinity anomaly in the Atlantic) can only be maintained if the salt transport anomaly due to changes in velocity eventually grows large enough to overcome the transport anomaly due to changes in salinity (which dominates the F_{az} term. Huisman et al. (2010) demonstrated that for small perturbations the anomalous salt transport associated with velocity changes (mainly F_{ovS}) is larger than the anomalous salt transport associated with salinity changes (mainly F_{az}) when F_{ovS} is (sufficiently) negative and that this term indeed is dominated by changes in overturning circulation. The more negative F_{ovS} becomes, the stronger the growth of the salt transport anomaly associated with velocity changes, until the AMOC can collapse without any type of hosing, due to stochastic perturbations in precipitation over the North Atlantic subpolar gyre (Cimatoribus et al., 2013).

However, during the first hundred years after strong hosing (the regime of a large perturbation not studied in Huisman et al., 2010), the transport anomaly associated with salinity changes (i.e., F_{az}) will always dominate the Atlantic freshwater/salt budget, removing part of the freshwater anomaly that resulted from the hosing. Because the anomaly is surface intensified and net volume transport at the southern boundary in the upper layers is northward, the freshwater transport anomaly that removes freshwater from the Atlantic to the Southern Ocean must be due to the gyre transport. It appears that this behavior is generic, regardless of the sign of F_{ovS} and does not imply F_{ovS} as an indicator for stability is invalid. This response in gyre transport is illustrated by Figure 6, which shows the change in F_{ovS} and F_{az} and their decomposition in contributions due to $S\Delta v$ (velocity changes) and $v\Delta S$ (salinity changes) at all latitudes in the Atlantic. During the first century after hosing, F_{ovS} freshens the basin further, helping to maintain the AMOC in its off-state, while F_{az} salinifies the basin, dampening the effect of hosing by transporting freshwater out of the basin and as a result, countering the AMOC off-state. F_{ovS} is indeed dominated by the change in velocity and F_{az} is dominated by the change in salinity and is slightly larger than F_{ovS}, so the total freshwater transport anomaly removes freshwater from the basin during the first 100 years. However, the resulting salinification is not enough to destabilize the AMOC off-state (Mecking et al., 2016).

In conclusion, a large response in gyre transport may occur after hosing that supersedes the response in salt transport by the overturning circulation. The gyre transport response F_{az} helps removing the initial
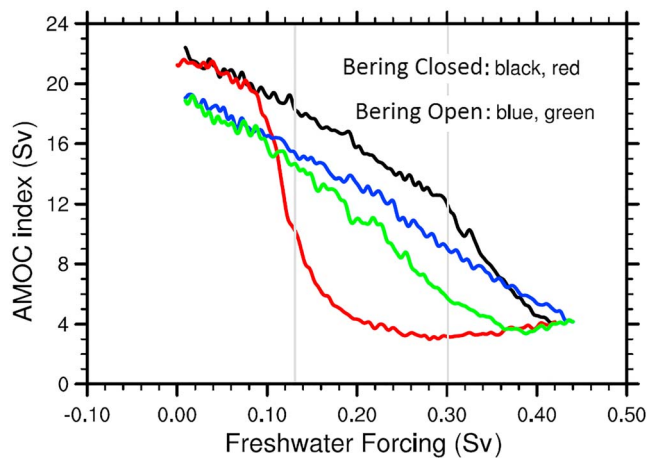
Figure 7. Hysteresis curves of the AMOC with respect to a freshwater flux perturbation F in the North Atlantic in CCSM3 (Hu et al., 2012a). The model with closed Bering Strait (black: increasing F, red: decreasing F) displays a significant hysteresis loop, while the configuration with open Bering Strait (blue, green) displays a linear relationship between AMOC F without hysteresis. Note that a change of the freshwater forcing by 0.1 Sv in this figure takes place over 500 model years. AMOC = Atlantic Meridional Overturning Circulation.

freshwater anomaly that resulted from freshwater hosing but is ineffective in changing the salinity of the Atlantic as a whole, or within the subpolar gyre, as it is dominated by, and scales with the salinity anomaly itself. Therefore, it does not control the stability of the AMOC. It is the salt transport anomaly associated with velocity changes that ultimately determines the fate of the AMOC, and this transport anomaly is dominated by F_{ovS}.

3.4. Bering Strait

Bering Strait throughflow is another important feedback on the AMOC. Bering Strait is a narrow and shallow strait connecting the Pacific and the Arctic, which opened roughly 5 million years (Ma) before present (BP; Marincovich Jr & Gladenkov, 2001). The Strait was closed during several glacial periods when global mean sea level was more than 50 m below the present day's Bering Strait sill depth. Today, there is a volume transport about 0.8 Sv from the North Pacific into the Arctic (Woodgate & Aagaard, 2005) through the Bering Strait and subsequently into the North Atlantic. Because the water in the North Pacific is fresher than the water in the Arctic and North Atlantic, this Bering Strait water transport acts as a freshwater source for the Arctic and North Atlantic. By using a three-box model, Shaffer and Bendtsen (1994) indicate that the Bering Strait throughflow may have played a role in determining the AMOC stability, such that with a deepened Bering Strait and an increased

throughflow, the AMOC would weaken and even collapse when the Bering Strait is 14 m deeper than it is now. De Boer & Nof (2004, 2004) suggested that due to strong Southern Ocean wind, about 4-Sv Southern Ocean water is pushed into the Atlantic basin. With an active AMOC, this water will return to the Southern Ocean via the NADW. With a collapsed AMOC, this water will exit at Bering Strait into the Pacific. By analyzing freshwater hosing experiments, Hu and Meehl (2005) and Hu et al. (2007, 2008) show that the Bering Strait transport can indeed reverse direction when the AMOC collapses. Hu et al. (2007, 2010, 2011, 2015) and Brierley and Fedorov (2016) further indicate that closing the Bering Strait can induce a strengthening of the AMOC. Moreover, the open Bering Strait can act as a control for AMOC long-term variability, such that when the AMOC weakens, less freshwater will flow through the Bering Strait into the Arctic and subpolar North Atlantic, leading to a strengthening of the AMOC and vice versa (Hu et al., 2011). Also, in terms of stability of the AMOC Bering Strait transport has an effect. When the Bering Strait is open, after freshwater hosing both (reversed) Bering Strait transport and the gyre circulation across the southern boundary of the Atlantic adjust to remove (part of) the freshwater anomaly from the Atlantic. Without an open Bering Strait the freshwater can no longer escape via this route and this tends to favor more the off-state once the AMOC is temporarily pushed in that state after hosing.

Building on these earlier works, Hu et al. (2012a) studied the effect of the Bering Strait opening or closure on the AMOC hysteresis. They found that with an open Bering Strait, AMOC weakens/strengthens linearly with the increasing/decreasing freshwater forcing in the North Atlantic due to changes in the Bering Strait freshwater transport (it is a freshwater source for a strong AMOC but a freshwater sink for a weak AMOC). With a closed Bering Strait this damping feedback is absent. As a result, the AMOC has significant hysteresis behavior with a closed Bering Strait (Figure 7; black and red curves). Under present-day condition with an open Bering Strait, the AMOC hysteresis is weaker or even absent depending on the background climate (Figure 7; blue and green curves). However, more recent model studies have shown multiple equilibria of the AMOC with an open Bering Strait (Jackson & Wood, 2018a, 2018b). They found that transport changes through the Bering Strait still salinified the North Atlantic, even though they did not lead to an AMOC recovery.

Bering Strait transport also affects a seesaw between Atlantic and Pacific deep convection. Saenko et al. (2004) found such seesaw-like changes of deep convection between the North Atlantic and North Pacific. When the deep convection weakens in the North Atlantic with a weakened AMOC, the deep convection in the North Pacific starts to strengthen and leads to the formation of a Pacific meridional overturning circulation. Based on new marine core data, Okazaki et al. (2010) indicated that seesaw-like climate change did occur between the North Pacific and North Atlantic during Heinrich 1 event, and their model simulation

showed that this seesaw-like climate change is due to the setup of North Pacific deep convection in association with a collapsed AMOC. Similar seesaw-like climate change during Heinrich 1 event has also been found in a transient simulation of the last deglaciation (Liu & Hu, 2015). Hu et al. (2012b) further demonstrated that the setup of the North Pacific deep convection with a collapsed AMOC can only occur during glacial times with a closed Bering Strait. With an open Bering Strait, the freshwater anomalies in the North Atlantic, which lead to the collapse of the AMOC, would be partly transported into the North Pacific via a reversed Bering Strait throughflow, resulting in freshening of the upper waters in the North Pacific preventing deep convection to occur. In contrast, Burls et al. (2017) simulated both AMOC and Pacific meridional overturning circulation active in a climate GCM with an open Bering strait in a Pliocene climate configuration.

3.5. Agulhas Leakage

Another gateway that has been hypothesized to affect the stability of the AMOC is Agulhas Leakage, the exchange of relatively warm and salty waters between the Indian and Atlantic Oceans, south of Africa. Gordon (1986) already speculated that this salty inflow salinifies the Atlantic Ocean and preconditions the North Atlantic for deep convection. Gordon et al. (1992) even speculated: "If the Indian Ocean salt input were severed, might the NADW thermohaline cell run down?"

While Weijer et al. (2002) showed with numerical experiments that Agulhas Leakage salt import may indeed strengthen the AMOC, as Gordon had speculated, the impact of Agulhas Leakage on the stability of the AMOC was explicitly addressed by Weijer et al. (2001). This study uses a simple 2-D Boussinesq model in which exchanges between the Atlantic and the Southern Ocean are represented by sources and sinks of heat and salt and Bering Strait freshwater import through a lateral salt flux. The study suggests that, within this schematized context, interocean exchange by Agulhas Leakage indeed enhances the stability of the AMOC. The main reason is that a realistic combination of heat and salt input from the subtropical thermocline of the South Indian Ocean represents a positive buoyancy source for the South Atlantic, strengthening the AMOC and inhibiting alternative equilibria with downwelling in the south. Note that this behavior is dominated by the heat input; Agulhas Leakage salt input was found to stimulate the AMOC only if at least part of it is removed at the levels of NADW (warm water route).

So how can this be reconciled by modern ideas, discussed in detail in section 3.1, that a saltier thermocline in the South Atlantic tends to make F_{ovS} more negative, possibly making the AMOC *less* stable instead? The main reason for this ambiguity is the fact that in the simple 2-D model of Weijer et al. (2001), the Atlantic freshwater budget north of 34°S, is controlled by net evaporation, so each equilibrium needs to satisfy the constraint of a fixed northward F_{ovS} (assuming that horizontal diffusion is a small contribution to the overall salt balance), which disfavors southern sinking solutions. In contrast, 3-D models allow F_{ovS} to adjust in response to any additional constraints on the salinity in the South Atlantic, hence enabling them to affect their AMOC stability behavior (e.g., De Vries & Weber, 2005). Note that in Weijer et al. (2001), the stabilizing effect of Agulhas leakage can be attributed mainly to the input of heat. As most studies have focused on the role of salt exchange, a proper evaluation of the impact of heat exchange on the AMOC may be due (e.g., Weijer et al., 2002).

Interestingly, Le Bars et al. (2013) suggest that Agulhas Leakage is influenced by the Indonesian Throughflow, hence adding another gateway transport to the potential controls on AMOC strength and stability.

3.6. The Effect of Mixing on the Stability of the AMOC

Several studies showed that vertical mixing enhances hysteresis width and the stability of the AMOC (Dijkstra, 2007; Nof et al., 2007; Prange et al., 2003; Schmittner & Weaver, 2001; Sijp & England, 2006). Prange et al. (2003) show that for weak mixing the AMOC becomes weak and hysteresis disappears, although other studies show that a regime of multiple equilibria continues to exist, even in the limit of small vertical diffusion (Hofmann & Rahmstorf, 2009; Sévellec & Fedorov, 2011; Wolfe & Cessi, 2014, 2015). The response to mixing seems dominated by the effect that the AMOC itself scales with vertical diffusivity and that a stronger AMOC needs larger perturbations to switch off. A similar change occurs for the reversed overturning circulation on the off-branch. Again, larger perturbations are needed to switch from the off-state to the on-state for increasing vertical diffusivity. Apart from the competition between increased advective salt (freshwater) feedbacks and freshwater perturbations, another mechanism might be that vertical diffusion dilutes the surface signal of the applied freshwater anomaly, preventing the formation of a halocline that leads to switch

from the on-state to the off-state. For the switch from the off-state to the on-state such argument is not applicable. Sijp and England (2006) confirmed the enhanced stability of the on-state for increased vertical mixing; however, when the vertical mixing changes are only applied inside the Atlantic basin, an opposite reaction occurs: the reversed overturning is also destabilized, and even stronger than the on-state, as it upwells solely within the Atlantic, while the on-state upwells for a large apart outside the Atlantic.

A few groups have investigated the stability of the AMOC with respect to horizontal or isopycnal mixing, but the results are inconsistent. Schmittner and Weaver (2001) found a narrowing of the hysteresis curve for increasing horizontal diffusion in a two-dimensional model and a reduction of the AMOC stability. In contrast, Dijkstra (2007) found that the regime of multiple equilibria does not change in width but moves to larger values of the freshwater perturbation, effectively stabilizing the AMOC. Sijp and England (2009) explored the sensitivity of the AMOC with respect to isopycnal, instead of horizontal diffusion, and showed that isopycnal diffusion enhances the stability of the on-state. They argue that enhanced isopycnal diffusion makes vertical tracer transport less dependent on convective mixing in areas of deep water formation; besides, isopycnal diffusion is less sensitive to freshwater anomalies than convective mixing, which can easily be shut down by freshwater capping. These factors prevent the formation of a halocline, making an AMOC collapse less plausible.

To summarize, these studies suggest that the stability of the AMOC is very sensitive to both vertical and horizontal (or isopycnal) mixing. Given our limited understanding of the processes that generate mixing in the ocean, caution is due when interpreting the stability of the AMOC in numerical models.

3.7. Primary Atmospheric Feedbacks

The ocean is strongly coupled to the atmosphere, and several feedbacks exist between the ocean circulation and the atmosphere that potentially affect the stability of the AMOC.

Arguably, the most important process is a temperature feedback that is absent in forced ocean runs and under mixed boundary conditions. A weakening AMOC will cool subpolar waters, but without an atmospheric temperature feedback, cooling will be counteracted by anomalous heat uptake by the ocean. At the same time, the weakening AMOC will freshen subpolar waters (the advective salt feedback) and without the atmospheric temperature feedback freshening will always dominate cooling and the reduced densities in the subpolar waters will further weaken the AMOC in the classical Stommel (1961) framework. In a coupled system, however, the atmosphere will develop a cold anomaly as well, which decreases the anomalous ocean heat uptake and counteracts the density loss due to reduced northward salt transport. As a result, the advective salt feedback is now counteracted by an opposite (atmospheric) temperature response in terms of density decrease, while this counteracting effect is absent in forced runs or under mixed boundary conditions. Consequently, the AMOC is oversensitive to the advective salt feedback in forced ocean-only runs, that is, an arbitrary decrease in the AMOC is amplified by the salt advective feedback, without a strong enough counteracting feedback being present to maintain the AMOC equilibrium implied by the initial conditions. Therefore, in ocean-only simulations the AMOC must be constrained by salinity restoring to maintain a reasonable AMOC (e.g., Behrens et al., 2013). Hence, coupling reduces, but does not take away, the AMOC's sensitivity to freshwater perturbations (Saravanan & McWilliams, 1995). Interestingly, den Toom et al. (2012), using the fully-implicit modeling framework of Weijer et al. (2003), determined that the temperature feedback had little impact on the equilibrium structure of the AMOC and, if anything, made the AMOC slightly less stable. The reason for this discrepancy has not been elucidated yet.

Atmospheric feedbacks also explain why the temperature response to an AMOC collapse in Northern Hemisphere (NH) high latitudes is so large, while ocean heat transport associated with the AMOC at these latitudes is relatively small compared to atmospheric heat transport (e.g., Trenberth & Fasullo, 2017). First, a robust response to decreasing convective activity in the subpolar North Atlantic is the southward expansion of sea ice cover (Drijfhout, 2014; Liu et al., 2017; Vellinga & Wood, 2002), which results in an increase of surface albedo in the NH. More important, however, is the subsequent reduction in latent heat release and evaporation, leading to a much colder and drier atmosphere (Drijfhout, 2015a; Laurian et al., 2009) and a much stronger response than can be inferred from Bjerknes compensation between ocean and atmospheric heat transport.

The wind response over the subpolar North Atlantic is less well examined, but a robust response appears to be an enhanced jet stream and increased storminess consistent with enhanced atmospheric heat transport

and a positive North Atlantic Oscillation index during the winter season (Jackson et al., 2015; Woollings et al., 2012). This response may be preceded in the first 20 years by increased blocking and a weaker jet stream associated with thermally forced high pressure in the subpolar North Atlantic and Arctic, where cooling is initiated (Drijfhout, 2015a). After spin-up of the NH jet stream the associated increased Ekman transport in the subpolar gyre might act as a negative feedback (promoting the on-state) in so far as it promotes water exchange between subtropical and subpolar gyres. This diffusive effect (Longworth et al., 2005) is supported by enhanced Ekman suction in the subpolar gyre increasing deep convection (Yang et al., 2016). The colder atmosphere also becomes drier (Drijfhout, 2015a; Jackson et al., 2015; Laurian et al., 2009; Liu et al., 2017), and net evaporation over the Atlantic decreases, being perfectly correlated with global mean surface temperature (Drijfhout et al., 2011). As a result, precipitation over the North Atlantic subpolar gyre as a whole decreases (Jackson et al., 2015; Woollings et al., 2012), which is a positive feedback, despite an increased northward atmospheric moisture transport and enhanced precipitation in some parts of the gyre (Schiller et al., 1997). The reduced evaporation and reduced ability of the colder air to hold moisture in the North Atlantic region overcompensate enhanced northward atmospheric moisture transport from the south with drying as a result. Another positive feedback is the subsequent increase in sea ice cover (Drijfhout, 2014; Liu et al., 2017; Vellinga et al., 2002). The problem of the two-way interaction between the AMOC and Arctic sea ice and its effect on the AMOC stability is an area of ongoing research (Liu & Fedorov, 2019; Liu et al., 2019; Sévellec et al., 2017; Sun et al., 2018).

3.8. The Role of ITCZ Shifts

When the AMOC collapses, a cross-equatorial SST anomaly arises (cold north of the equator and warm south of the equator), resulting from a collapse of the cross-equatorial northward heat transport by the AMOC (Stouffer et al., 2006; Timmermann et al., 2007). This anomaly is associated with an equatorward shift of the latitude of maximum heating, which, mainly due to the AMOC (Feulner et al., 2013), occurs in Northern Hemisphere. Furthermore, the cross-equatorial heat transport of the AMOC gives rise to a hemispheric heating imbalance which is (partly) remediated by a migration of the ITCZ toward the warmer hemisphere (Schneider et al., 2014), leading to moisture transport from the southern (SH) to the northern (NH) hemisphere and heat transport from the NH to the SH. The anomaly in thermal forcing due to an AMOC collapse is therefore equivalent to but smaller in amplitude than the solstitial shift from boreal summer to boreal winter. As a result, a southward shift in the ascending branch of the Hadley circulation occurs, dividing the NH and SH cells, manifested by a southward shift of the tropical rain belt (ITCZ) over the Atlantic (Manabe & Stouffer, 1988; Vellinga et al., 2002; Zhang & Delworth, 2005). In Fučkar et al. (2013) it was argued that the AMOC determines the hemispheric anomaly in zonal mean ITCZ position, further corroborated by, for example, Frierson et al. (2013) and Marshall et al. (2014). The southward shift of the ITCZ after an AMOC collapse is thus associated with a robust southward migration of the NH Hadley cell (Drijfhout, 2010; Liu et al., 2017; Yu & Pritchard, 2019). This occurs since changes in ocean northward heat transport by the AMOC are in the tropics partly compensated by changes in atmospheric heat transport by the Hadley circulation and the oceanic subtropical cells, which are connected to the atmospheric circulation through wind stress (Green & Marshall, 2017). Because of this, heat transport anomalies by subtropical cells suppress ITCZ meridional displacement especially in the tropical Pacific (Green & Marshall, 2017). When the NH Hadley cell shifts southward, its northern boundary contracts, while zonal velocities aloft increase, and the subtropical jet shifts equatorward, whereas in the SH cell the opposite occurs (Drijfhout, 2010). Further atmospheric teleconnections with the tropical and north Pacific (e.g., El Niño-Southern Oscillation, or ENSO) are discussed in Wu et al. (2005, 2008), Timmermann et al. (2007), and Dong and Sutton (2007) but do not appear as feedbacks on AMOC stability.

Changes in heat transport between ocean and atmosphere do not show (complete) Bjerknes compensation, and a top of atmosphere (TOA) radiation imbalance arises associated with net ocean heat uptake (Drijfhout, 2014; Vellinga & Wu, 2008). A rearrangement of the Earth's energy budget is necessary to accommodate the change in net meridional energy transport by the coupled ocean-atmosphere system (Cheng et al., 2007; Vellinga & Wu, 2008). At first sight this may seem somewhat surprising as the hemispheric radiation imbalance at the surface is not reflected at the TOA and thus is not a result from TOA radiative forcing but from asymmetric ocean heat transport (Frierson et al., 2013). Also, interannual changes in atmospheric heat transport are strongly coupled to variations in TOA radiative forcing but almost decoupled from variations in ocean heat transport that mainly cause changes in ocean heat storage (Donohoe et al., 2014). When variations in OHT become larger, or are sustained over longer timescales, the associated cooling of the NH and

warming of the SH with decreasing AMOC create a significant TOA imbalance associated with changes in outgoing longwave radiation, which in case of an AMOC collapse is dominated by NH cooling due to stronger fast feedbacks and reduced outgoing longwave radiation, giving a net global imbalance in surface temperature and TOA radiation. Without a reorganization of TOA radiation in terms of cloud responses, the net TOA imbalance and ocean heat uptake will then lead to warming of the subsurface North Atlantic waters and ultimately a destabilization of the halocline that formed over previous convective regions inhibiting deep convection, kicking in convective overturning after deep warming passed a threshold. Therefore, a stable off-state can only be maintained when the net TOA imbalance and ocean heat uptake quickly decrease or even change sign, as argued in Drijfhout (2015b). It is at present completely unclear how or whether this stabilization/destabilization of the AMOC off-state relates to the sign of F_{ovS} and how robust the associated reorganization of the TOA radiative imbalance is after an AMOC collapse for different AMOC stability characteristics. However, the relation between a stable AMOC off-state and the disappearance of the net TOA imbalance within a few decades was found in three models (SPEEDO, ECHAM5, and HadGEM3).

Apart from global considerations, a more regional feedback associated with ITCZ shifts may be important. In summer, the ITCZ is situated over the North Atlantic subtropical gyre (NASTG). When it shifts southward, the NASTG receives less precipitation and as a consequence salinifies through this atmospheric response, while the tropical belt in the Atlantic freshens. Coarse-resolution ocean models often feature weak freshwater convergence in the NASTG due to the AMOC (Mecking et al., 2016; Yin & Stouffer, 2007). This implies that an AMOC collapse even further salinifies the NASTG. Only a fast switch to a reversed THC that imports freshwater into the NASTG is able to counteract this tendency of salinification. Yin and Stouffer (2007) found that only flux-corrected models support such a fast switch to a reversed THC after freshwater hosing and that a newer model without flux correction was unable to balance the freshwater/salt budget in the NASTG as the reversed THC did not arise. The ever increasing salinity of the NASTG eventually gave rise to large-scale instabilities at the STG/SPG front, leading to advection of salt anomalies into the SPG that eventually kick started deep convection and subsequently the AMOC on-state.

This atmospheric feedback due to a southward shift of the ITCZ, however, was effectively counteracted in the Mecking et al. (2016) model study. The higher-resolution coupled model featuring an eddy-permitting ocean model allowed for eddy transports and swifter boundary currents transporting more freshwater into the NASTG, which was accompanied by an AMOC characterized by freshwater divergence into the area. As a result, the salinification of the NASTG associated with a southward shift of the ITCZ was effectively counteracted by freshening due to a collapsing AMOC. The equilibrium state after the AMOC collapse was indeed associated with a saltier NASTG and fresher NASPG, but the higher-resolution model did no longer support the development of large-scale instabilities at the STG/SPG front and enhanced southward eddy transport of freshwater and northward eddy transport of salt did not give rise to a kick-start of deep convection as they were counteracted by atmospheric changes in EPR and necessary to maintain the new North Atlantic equilibrium climate.

In general, we may conclude that the net result of all atmospheric feedbacks is to counteract the salt-advection feedback, but together, they are not powerful enough to incapacitate it. The situation will become different if we release the perspective of the pan-Atlantic freshwater budget and zoom in on the gyre scale.

3.9. Is the Freshwater Budget of the North Atlantic Subpolar Gyre More Crucial Than Over Whole Atlantic Budget

The original idea of the importance of F_{ovS} to the AMOC stability uses the sign of F_{ov} at the southern border of the Atlantic and assumes that the salinity of the whole Atlantic is a control on the AMOC. However, an advective feedback associated with the advection of fresh water locally within the Atlantic is also possible.

Mecking et al. (2016) showed that freshwater convergence within the North Atlantic is an important feedback for opposing the salinification caused by the ITCZ shift discussed in the previous section. Jackson and Wood (2018a, 2018b) also discuss this feedback in different experiments with the same GCM, showing that both collapsed and recovering AMOC states after hosing feature an increasing salinity in the subtropical gyre and decreasing salinity in the subpolar gyre. In both states the basin-integrated salinity is quite similar. Hence, it is the fresh water convergence by the AMOC across the subpolar gyre that is important, rather than the convergence across the whole Atlantic. This contradicts the findings of Sijp et al. (2012) who reported that the on-branch and off-branch for the AMOC are likewise coupled to two pan-Atlantic salinity states of high

and low salinity. However, in Mecking et al. (2016) and Jackson and Wood (2018a, 2018b) the collapsed state is characterized by a shallow overturning circulation in the same direction as the on-state AMOC, rather than a reversed circulation, and experiments are limited to a few hundred years. Hence, it may not be a true equilibrium and it is possible that F_{ov} at the southern boundary becomes relevant on longer timescales.

These results put the relevance of F_{ovS} as indicator for the stability of the AMOC into question. It appears that at least at shorter timescales, a large enough freshwater transport divergence by the overturning circulation into the North Atlantic subtropical gyre (ΔF_{ov}, but now taken between 45°N and 15°N) can counteract salinification by a southward shifting ITCZ when the circulation and resulting freshwater transport divergence collapse. This can only be achieved if both gyre and mesoscale eddies are associated with even larger freshwater transport convergence in the control state. Whether freshwater divergence by the overturning over the whole Atlantic (to be measured by F_{ov} at 34°S or ΔF_{ov} over the whole Atlantic) is a necessary condition for a stable off-state can only be answered when these high-resolution coupled models can be run for more than thousand years.

4. Implications of the AMOC Stability Paradigm for the Past Climate System

Millennial-scale climate changes, such as large-amplitude transitions between cold stadial and warm interstadial states known as "Dansgaard-Oeschger (D-O) oscillations" (Dansgaard et al., 1993) during the last glacial period, are the most relevant paleoclimate phenomena that are associated with pronounced expression of AMOC instability. Studies of millennial-scale climate changes have made great advances over the past decade, thanks to many recently constructed high-resolution isotope records that can determine the precise timing of major events occurred in the abrupt changes, as well as long-term simulations performed by EMICs and GCMs that can test the AMOC stability under past climate conditions. In this section, we review the joint efforts of paleo-observational and paleo-modeling communities in constructing an AMOC stability framework on millennial timescales, which can help to understand the role of the AMOC in D-O oscillations and, more importantly, can provide a link between glacial and modern AMOCs and inform its future changes.

4.1. Externally Forced Millennial Variability

A conceptual model (Figure 8a) that is widely utilized within the paleoclimate community to understand the glacial-interglacial AMOC behavior has three modes (Alley et al., 1999; Rahmstorf, 2002; Sarnthein et al., 1994): a "warm on" mode and an off mode, which correspond to the present-day on and off states described in section 2, as well as a "cold on" mode in which the upper cell is much shallower and reaches not as far north as the warm mode, leaving the deep North Atlantic dominated by Antarctic Bottom Water. A recent review of this conceptual model and its application to abrupt climate changes can be found in Lynch-Stieglitz (2017). However, a linkage between this three-mode model and the AMOC stability as discussed in section 2 is still lacking. Here, we suggest a framework that builds up this linkage by separating the dependence of the AMOC stability into two external forcing controls: a slow, long-term control represented by atmospheric CO_2 concentration and/or ice sheet level and a relatively fast, short-term control represented by a freshwater perturbation to the North Atlantic Ocean (Figure 8b).

Deep ocean proxies suggest that the AMOC transitioned from a warm on mode to a cold on when the background climate cooled down from interglacials to glacials and vice versa when the climate warmed up from glacials to interglacials (Burckel et al., 2015; Henry et al., 2016). The transitions in both directions are reported to be nonlinear in model simulations, both EMICs (Buizert & Schmittner, 2015; Ganopolski & Rahmstorf, 2001) and AOGCMs (Klockmann et al., 2018; Zhang et al., 2014a, 2017). A bistable regime, namely, the coexistence of a warm on and a cold on under a single climate condition, is further reported in some of them. This background-state dependency is conceptually represented by an equilibrium diagram in our framework (curve D-C in Figure 8b), with the forcing controls being atmospheric CO_2 concentration and/or ice sheet level. When CO_2 is high (e.g., the present level of 400 ppm or a future scenario with even higher concentrations) and there is no Laurentide Ice Sheet (LIS) present over the North America continent, the AMOC is strong and the warm on is the only possible on-state. When CO_2 is low, for example, at 185 ppm during the Last Glacial Maximum (LGM, ~22,000 years before present), and the LIS reaches its peak size, the cold on is the only possible on-state (Zhang et al., 2014a, 2017). It should be pointed out that it is debatable whether a low CO_2 corresponds to a strong or weak AMOC, or even an 'off' state of the AMOC. Reconstructions of the AMOC during the LGM are more certain about its structure but less certain about its
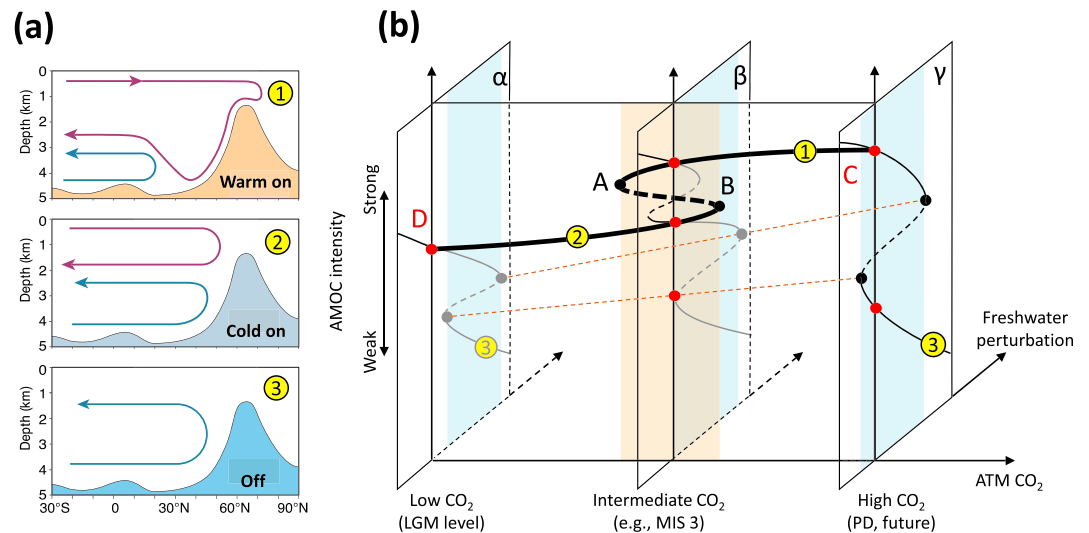
Figure 8. Synthesis diagram of AMOC stability and related states over the glacial-interglacial cycles. (a) Three modes of the Atlantic circulation that prevailed during the last glacial period. After Rahmstorf (2002). (b) Interplay between the AMOC equilibrium structure with respect to atmospheric CO_2 concentration (curve D-C) and with respect to freshwater perturbations in the subpolar North Atlantic (curves on planes α, β, and γ, similar to that in Figure 4a). Solid lines are (linearly) stable; dashed lines are unstable. Orange and blue shades indicate the bistable regimes of each curve. Numbers indicate the corresponding circulation modes in (a). Red dots indicate stable AMOC states under different climate conditions. LGM = Last Glacial Maximum; MIS = Marine Isotope Stage; PD = present day. Red dashed lines indicate the location of the limit points with respect to the freshwater perturbations. AMOC = Atlantic Meridional Overturning Circulation.

strength (Burckel et al., 2016; Gebbie, 2014; Lynch-Stieglitz et al., 2007; Peck et al., 2007), although existing evidence collectively suggests it likely to be comparable to or weaker than the 'warm on' (McManus et al., 1999), but unlikely to be 'off' (Otto-Bliesner et al., 2007).

When both CO_2 and the LIS are at their intermediate levels, such as Marine Isotope Stage (MIS) 3 between 60,000 and 25,000 years before present, a bistable AMOC regime with respect to CO_2 occurs. The location of the thresholds A and B is uncertain, probably somewhere near 210 and 230 ppm (Zhang et al., 2017)—a narrow band which indicates that a small change in atmospheric CO_2 of ~15–25 ppm is sufficient to alter the climate between cold stadials and warm interstadials (Ahn & Brook, 2014; Bereiter et al., 2015). It remains debatable whether ice sheets can act as an individual forcing that induces nonlinear responses of the AMOC besides CO_2, since in reality ice sheets and CO_2 usually go hand in hand and the effects of the two can hardly be separated. Some models exhibit a nonlinear AMOC evolution with abrupt changes in response to a linear change of the LIS elevation (Zhang et al., 2014a), while others show such responses in a more gradual and linear fashion (Zhu et al., 2014).

Our AMOC CO_2 equilibrium diagram indicates that there is no single normal mode of operation of the glacial climate system as previously thought (Buizert & Schmittner, 2015; Ganopolski & Rahmstorf, 2001). Instead, it supports the idea that there was a mode change from warm on to cold on with a decreasing CO_2 across the last glacial (Bereiter et al., 2012). Furthermore, the AMOC could frequently switch between warm on and cold on when CO_2 fell in range of ~210 and 230 ppm, possibly with temperate overshoots (e.g., with an extremely deepening NADW when the AMOC suddenly strengthened; Barker et al., 2010; Deaney et al., 2017; Knorr & Lohmann, 2003). Since gradual CO_2 change could induce abrupt AMOC changes (Barker et al., 2015; Knorr & Lohmann, 2003, 2007), and AMOC changes could in turn control atmospheric CO_2 (with slowdown of the AMOC gradually releasing CO_2 and vice versa; Schmittner & Galbraith, 2008), the AMOC-CO_2 system could be involved in self-sustained oscillations without a necessary involvement of freshwater perturbations. This could potentially explain why freshwater input from massive discharges of icebergs to the North Atlantic (the so-called "Heinrich events" or 'Heinrich stadials') occurred during some of the D-O cold stadials but not all of them.

As reviewed in previous sections, AMOC stability is often discussed in the context of freshwater perturbations to the subpolar North Atlantic for a certain background climate. To the extent that such freshwater

perturbations are not *directly* linked to CO_2-induced climate change, this sensitivity can be represented in planes roughly perpendicular to the AMOC-CO_2 plane (e.g., planes α, β, and γ in Figure 8b). So, for instance, while this direction can appropriately reflect anomalous freshwater input due to rapid ice sheet melt, changes in the strength of the hydrological cycle should be considered along the CO_2 axis instead, as they are directly linked to atmospheric CO_2 concentrations. With this interpretation, an AMOC-FW equilibrium diagram like Figure 4a can then be drawn on such a plane, and the intersection with the AMOC-CO_2 curve (e.g., C or D) indicates the position of the background climate. As discussed in previous sections, it is possible that the intersection C under the preindustrial climate conditions falls in a bistable regime (the blue band) on the γ plane, if indeed the negative value of F_{ovS} deduced from observations indicates bistability. Under extreme glacial conditions such as the LGM (α-plane), the intersection D is suggested to fall in a monostable regime by most observational (Peck et al., 2007) and modeling (Ganopolski & Rahmstorf, 2001; Liu et al., 2015; Prange et al., 2002; Romanova et al., 2004; Zhang et al., 2014a) studies. Nonetheless, a few models (e.g., Schmittner et al., 2002) indicate that the off mode could be a stable background state, so that for those models the bistable regime on plane α is shifted "forward" in the diagram Figure 8b, and the AMOC-FW curve intersects the AMOC-CO_2 plane three times. How sensitive the AMOC at the full glacial was to the freshwater perturbation compared with that at the interglacial (viz., how much freshwater is needed to bring the AMOC to collapse) has been intensively studied but the answer remains unclear. Some proxies indicate the cold on is less sensitive compared with the warm on with variant explanations (Galaasen et al., 2014; Lynch-Stieglitz et al., 2014), while other proxies (Mokeddem et al., 2014; Thornalley et al., 2013) and most models (Bitz et al., 2007; Swingedouw et al., 2009; Weber & Drijfhout, 2007) suggest the opposite.

The AMOC stability is much more complicated when freshwater perturbations are imposed on the climate system under intermediate glacial conditions, such as during MIS 3 when pronounced D-O oscillations occurred along with Heinrich events. We place an AMOC-FW equilibrium diagram curve on a β plane near the transition periods between warm-on's and cold-on's (the orange band) in Figure 8b to indicate such scenarios. Without freshwater perturbations, the AMOC is bistable. With freshwater perturbations, the AMOC could be tristable theoretically and the climate system would be highly sensitive to external forcing, which is indeed the case during MIS 3. This probably could explain why D-O oscillations during glacial times occurred most frequently under intermediate glacial conditions (Kawamura et al., 2017; McManus et al., 1999; Schulz et al., 1999; Sima et al., 2004). When a full glacial state such as the LGM is reached, D-O oscillations were less frequent or even absent. Recent model simulations further support such a highly sensitive AMOC under intermediate glacial conditions, which is more sensitive to freshwater perturbations compared with those under interglacial and full glaciated climates (Kawamura et al., 2017; Zhang et al., 2014).

It has long been hypothesized that the Heinrich events could have provided freshwater perturbations to interrupt the AMOC (Heinrich, 1988), and it has been a protocol to use freshwater perturbations to generate millennial-scale AMOC variations in models (Ganopolski & Rahmstorf, 2001, 2002; Schmittner et al., 2002). However, high-resolution proxies collectively suggest Heinrich events are unlikely the trigger, and the AMOC shutoff process is far more complicated than previously thought. Instead, in support of our framework, the AMOC transition from warm on to off is likely a two-step process (Ng et al., 2018): in response to gradual cooling associated with atmospheric CO_2 decrease (Barker et al., 2015), an early slowdown of circulation may have occurred. It was accompanied by an early external freshwater input (Chen et al., 2016; Ménot et al., 2006) mostly likely from Eurasian ice sheets (Grousset et al., 2001; Ng et al., 2018; Peck et al., 2006). As a second step, an off state persisted for quite a few hundreds to a thousand years, with substantial ice rafting events from the LIS (Heinrich events) as a consequence (Barker et al., 2015; Hodell et al., 2017; Marcott et al., 2011).

In the simplified view of our framework, AMOC stability should be considered in the context of equilibria on a warped surface, where both climate change associated with atmospheric CO_2 concentrations (possibly through modifying the hydrological cycle) and freshwater perturbations have the ability to induce regions of multiple equilibria. Whenever stability is discussed, one may need to specify along which axis they are considering. For example, the preindustrial AMOC is monostable with respect to CO_2-induced multiple equilibria but in a bistable regime with respect to freshwater perturbations, together suggesting a bistable AMOC. It is worth pointing out that we use the two axes to emphasize the AMOC variability over two different timescales, namely, glacial-interglacial timescales controlled by atmospheric CO_2 concentration and/or ice sheet level and relatively short, interannual to millennial timescales controlled by freshwater perturbations. It has to be further pointed out that our current knowledge is not yet sufficient to complete the

plot, and due to the transient nature of the abrupt events, it is not easy to search for evidence in observations to support or deny the hysteresis loop as in the models. Future efforts from both observations and models are needed to reduce uncertainties and reconcile conflicts in order to provide more information to complete the plot.

4.2. Intrinsically Generated Millennial Variability

Many of the modeling studies explore the generation of AMOC millennial variability in the exogenous context. That is, AMOC variations are considered to be driven by external freshwater perturbations as discussed in the previous section. Such simulations can reproduce a reasonable millennial climate variability, but this approach faces several problems. The most important is that one has to assume a sufficiently strong quasi-regular or stochastic variations in freshwater fluxes (e.g., Braun et al., 2005; Ganopolski & Rahmstorf, 2001; Stastna & Peltier, 2007; Timmermann et al., 2003) for which there is no sufficient observational evidence, at least in the case of D-O oscillations. Consequently, the possibility of self-sustained millennial oscillations driven by internal AMOC dynamics has become an important alternative to explain millennial variability.

AMOC variations related to a robust internal natural mode of the AMOC, associated with propagating ocean density anomalies, have been found in most comprehensive climate models configured for the present-day climate (e.g., Muir & Fedorov, 2015, 2017; Sévellec & Fedorov, 2013, 2015). However, the magnitude (a few Sverdrups) and typical periodicity (decadal to centennial) of such variations mismatch the observed millennial variability during the past glacial-interglacial interval. In contrast, a possibility of pronounced self-sustained oscillations on a millennial timescale has been shown by many models of lower complexity, from box models (e.g., Colin de Verdière, 2007; Colin de Verdière et al., 2006) and loop models (mapping the ocean overturning circulation onto a rotating wheel; e.g., Sévellec & Fedorov, 2014, 2015; Winton & Sarachik, 1993) to simplified coupled models (e.g., Sakai & Peltier, 1995; 1996, 1997; Weaver & Hughes, 1994; Weijer & Dijkstra, 2003), suggesting that one or more of AMOC steady states may be unstable. This appears to be especially true for the models configured for glacial conditions.

A plausible change in the AMOC behavior from the present-day climate to glacial conditions can be illustrated by a simple loop model of the AMOC (Sévellec & Fedorov, 2014, 2015; Figure 9). The model assumes that the overturning circulation can be represented as a simple rotation on a latitude-depth plane and the flow is driven by a buoyancy torque balanced by friction. In this loop model for present-day conditions the AMOC is monostable. Shifting toward glacial conditions, which in the model implies a southward shift of deep convection and hence the AMOC, leads after two successive bifurcations to the emergence of three unstable states. This allows for a quasiperiodic oscillation in which the system wanders around those unstable states. As a result, the AMOC experiences chaotic millennial variability (in the sense of deterministic chaos) reminiscent of D-O oscillations with a periodicity close to the observed 1,500 years wherein transition between different modes is driven by the salt-advection feedback. It is critical that such oscillations do not require any external forcing for the transition between weak and strong AMOC states. Whether the real system is better described by deterministic chaos or stochastic resonance, for example, (Ganopolski & Rahmstorf, 2002; Rahmstorf & Alley, 2002) needs further investigation.

Recent studies with comprehensive state-of-the-art GCMs show qualitatively similar self-sustained AMOC oscillations for glacial conditions. Peltier and Vettoretti (2014) and Vettoretti and Peltier (2018) used an LGM configuration of CESM1 with additional modifications to ocean diffusivity to obtain a D-O-like variability with a period of about 700 years and oscillatory behavior consistent with a salt relaxation oscillator. These oscillations require an initial "kick" that in nature could be provided by a Heinrich event. How robust these results are across a broader selection of coupled climate model remains to be seen as a previous version of this coupled model, CCSM3, did not show such oscillatory behavior. Nevertheless, these results demonstrate that the system can behave in a manner clearly different from the classical monostable or bistable regimes.

5. Implications of the AMOC Stability Paradigm for the Future Climate System
5.1. AMOC Stability Under Anthropogenic Forcing

In coupled climate models the AMOC slows down in response to anthropogenic global warming (e.g., Collins et al., 2013; Gregory et al., 2005; Schmittner et al., 2005), although there is a large uncertainty in the amount of decrease (Reintges et al., 2017). For a subset of the CMIP5 models Drijfhout et al. (2012) show that
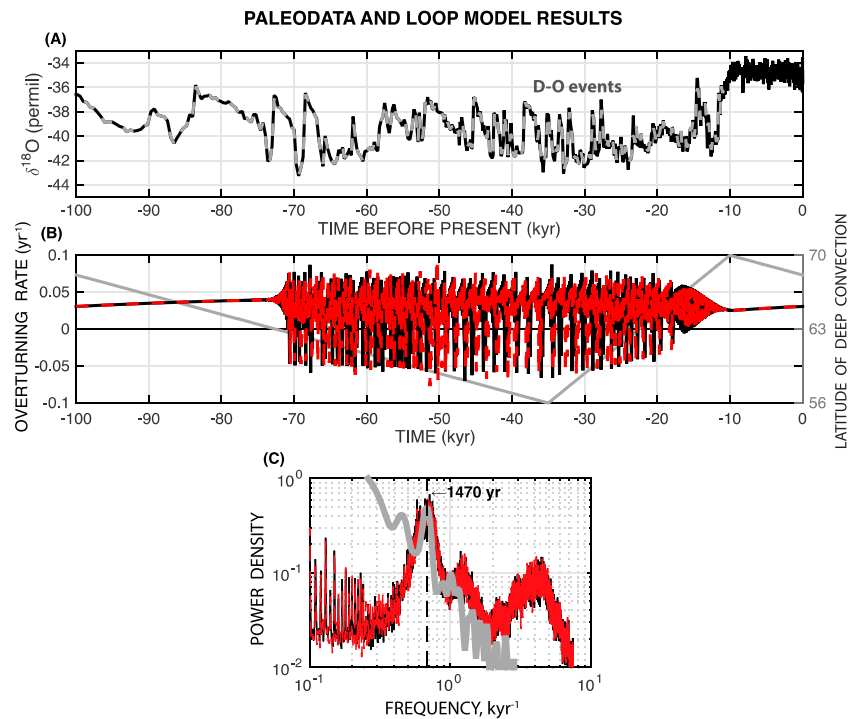
Figure 9. Greenland δ^{18}O paleodata and Atlantic Meridional Overturning Circulation variations in a loop model of the ocean overturning circulation. (a) δ^{18}O records from Greenland ice cores (solid black and gray dashed lines) over the last 100 kyr (Grootes & Stuiver, 1997) showing strong millennial variability including D-O events. (b) Variations in the Atlantic Meridional Overturning Circulation strength (overturning rate, year^{-1}) simulated by the loop model for two different sets of initial conditions (solid black and red dashed lines) within a single glacial-interglacial cycle. The gray sawtooth line indicates the imposed temporal changes in the position of deep convection that modify the system's stability characteristics. (c) The power spectra of the simulated (red) and observed (gray) records. The model captures well the 1,470-year spectral peak associated with the observed D-O oscillation but does not capture slower variability that may depend on other components of the climate system. After Sévellec and Fedorov (2015), reproduced under Creative Commons Attribution License (CC BY; https://creativecommons.org/licenses/by/4.0/).

the ensemble average AMOC weakened by 1.5 Sv/K in the scenarios with the greatest warming and 2 Sv/K with the least. However, there is a very large spread in these values across models. It should also be noted that the AMOC decline also scales with the strength of the AMOC itself under strong warming (Weaver et al., 2012), so a fractional change may be a better indicator.

This projected slowdown seems to be at odds with the positive relationship between AMOC strength and CO_2 in the past—as reflected in our synthesis diagram (Figure 8b). A possible reason for this discrepancy is that the equilibrium diagram shows the final equilibrium states, which usually take thousands of years to be reached. Over decadal to centennial timescales, the AMOC decreases in response to an CO_2 increase. This weakening may be exacerbated by potential positive freshwater perturbation from the Greenland Ice Sheet (Bakker et al., 2016; Böning et al., 2016; Luo et al., 2016; Rahmstorf et al., 2015) and the Arctic Ocean (Gelderloos et al., 2012; Giles et al., 2012). However, barring AMOC collapse, this initial phase can then be followed by a partial AMOC recovery that occur on centennial timescales. The cause of this partial recovery is slow subsurface ocean density changes that introduce a negative feedback eroding the stratification and partially reinvigorating convection and the AMOC (Thomas & Fedorov, 2019). Depending on the model and the forcing, the strengthening of the AMOC can continue for millennia (Jansen et al., 2018; Zhu et al., 2015).

Drijfhout et al. (2015) and Sgubin et al. (2017) investigated AMOC shut down within CMIP5 projections. In two climate models the AMOC shut down (reached a state with very weak AMOC) by 2100, while in seven models deep convection in the Labrador Sea/subpolar gyre collapsed, which affected the AMOC, but because deep convection was still present in the Nordic Seas, it did not lead to a full AMOC collapse. Sgubin et al. (2017) also found that those models which experienced a collapse of deep convection in the Labrador
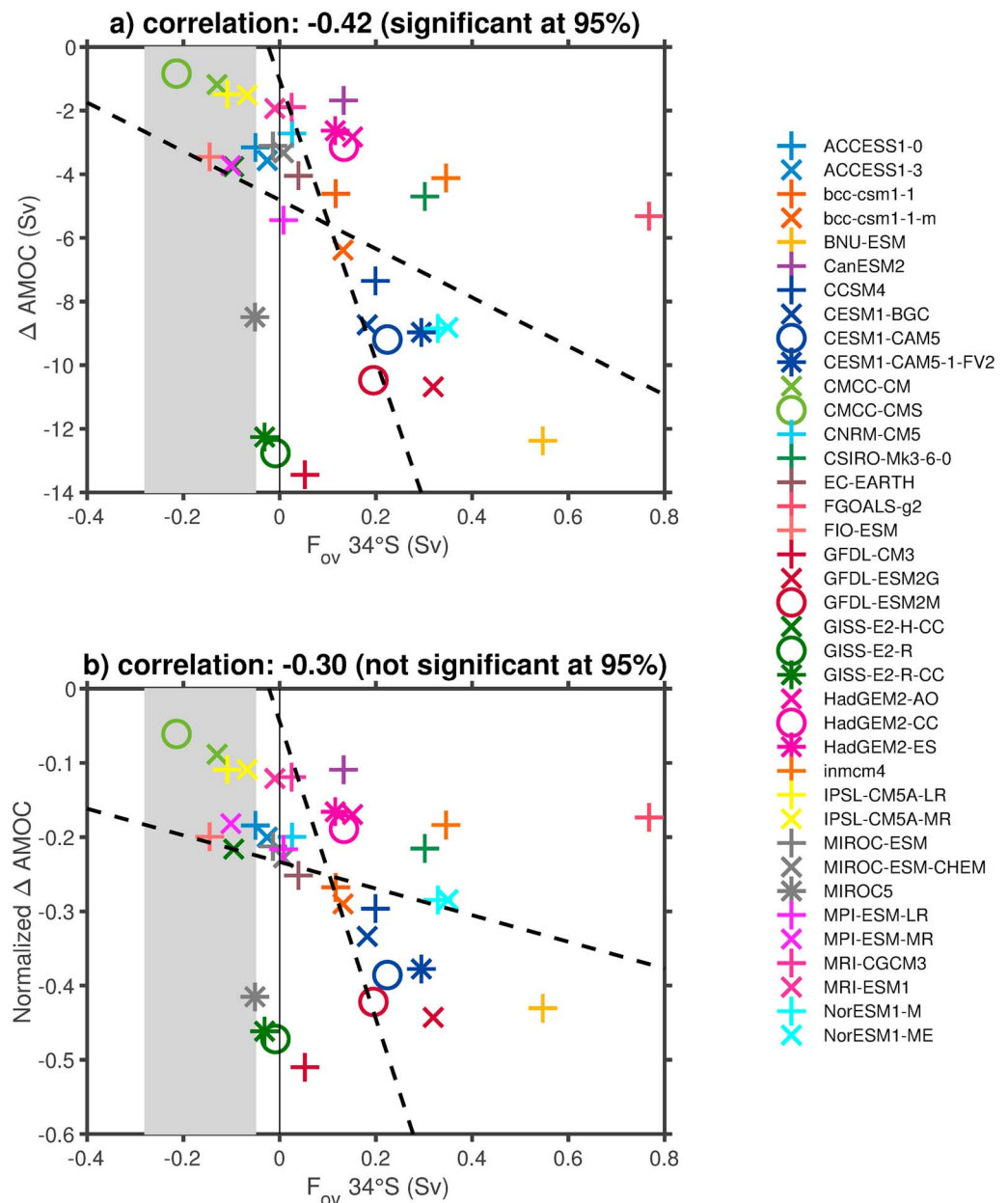
Figure 10. The relation between AMOC decline between 2070–2099 and 1850–1900 and F_{ovS} for the period 1850–1900 for the 38 models simulating the RCP8.5 scenario. The upper panel shows (a) the absolute AMOC decline in sievert, the lower panel (b) the fractional decline relative to the 1850–1900 climatology. The dashed lines on the figures show the lines of best fit using a least squares estimate; one line minimizes the error in the vertical direction and the other in the horizontal direction, and the shading indicates the observational range of F_{ovS} in the South Atlantic. From the analysis of Mecking et al. (2017).

Sea had weaker, more realistic, stratification in the Labrador Sea. Many CMIP5 models were shown to be too stratified making them less sensitive to buoyancy changes in the region.

The question of whether there is a relationship between the magnitude of the AMOC decline and F_{ovS} has also been considered. Since F_{ovS} is generally considered to indicate the presence of bistability, it might be thought more likely with negative F_{ovS} to experience a greater AMOC decline; however, it does not necessarily follow that anthropogenic forcing would result in a transition across the threshold to an off-state. It is also possible for models to experience a large weakening of the AMOC from a monostable state. Weaver et al. (2012) investigated the relationship in CMIP5 models of the change in AMOC and the change in F_{ovS}

between their 2081–2100 average and preindustrial average and found no relationship. The relationship between AMOC change and preindustrial F_{ovS} itself is shown in Figure 10 (from data shown in; Weaver et al., 2012 and Mecking et al., 2017). A relation between stronger AMOC decline and more negative F_{ovS} is not seen; on the contrary the opposite relationship is present. One possible explanation is that the models with stronger AMOC decline are the models with a stronger initial AMOC strength (Gregory et al., 2005; Weaver et al., 2007, 2012), which are shown by Mecking et al. (2017) to be too fresh in the South Atlantic and therefore have a more positive F_{ovS}. The fundamental cause of the relationship is unclear however.

Climate change will affect the salinity in the Atlantic by acceleration of the hydrological cycle (Collins et al., 2013; Held & Soden, 2006) making wet regions wetter and dry regions drier, increasing gradients in salinity. This could affect the AMOC and feedbacks on AMOC strength through advection, with a freshening of waters from polar and subpolar origin and salinification of waters from subtropical origin. South Atlantic salinity might also be affected by changes in Agulhas leakage (Beal et al., 2011). It is unclear how much of an impact this might have on North Atlantic salinity (Weijer & van Sebille, 2013); however, Weijer et al. (2001) showed that Agulhas leakage can have a stabilizing effect on the AMOC. Salinity changes in the South Atlantic can also affect F_{ovS}. In CMIP3 projections a general decrease in F_{ovS} during the next century was found, implying a tendency toward more bistability, followed by an increase in F_{ovS} during the 22nd century and afterward if the radiative forcing becomes strong enough (Drijfhout et al., 2011). Other factors associated with climate change could also affect the stability of the AMOC. Increasing Southern Ocean winds could enhance the northward Ekman flow, although increased southward eddy-transport might compensate for such increase. This can increase the freshwater transport not associated with the AMOC. In the long term, if there is a balance between freshwater in and out of the Atlantic, then an increase in net evaporation over the Atlantic with climate change may restrict how much F_{ovS} can decrease (Drijfhout et al., 2011).

The various feedbacks discussed could be altered by biases and inaccurate representation of processes in the models. These include problems with the location and magnitude of deep convection (Heuzé, 2017), subpolar stratification (Sgubin et al., 2017), biases in subpolar temperature and salinity (Menary et al., 2015), and representation of overflows (Zhang et al., 2011). Many models have also been shown to underestimate high-frequency variability of the AMOC compared to observations (Roberts et al., 2014) and to underestimate multidecadal Atlantic variability (Kim et al., 2018). It is possible that these factors could influence the response of the AMOC to anthropogenic forcing or the stability.

We conclude with the assertion that the relation between AMOC-decline and F_{ovS} is far from straightforward. It should be stressed that F_{ovS} was only introduced as an indicator for monostability/bistability of the equilibrium of an AMOC on-state, and it is unclear what the relationship is in a transient simulation. We also note that it is far from straightforward to deduce from the multimodel ensemble any reliable assessment of future AMOC behavior or changes in stability and that such assessment cannot be done without carefully taking model bias into account.

5.2. Climate Implications of an AMOC Collapse

A collapse or significant reduction of the AMOC would lead to reduced northward heat transport in the Atlantic Ocean and hence a cooling of the North Atlantic and surrounding areas; the direct cooling effect is amplified by sea ice expansion in response to reduced ocean heat transport into the Arctic (Vellinga & Wood, 2002). This is seen in many different model studies (Kageyama et al., 2013; Stouffer et al., 2006) although the geographical extent and magnitude of the cooling vary across models. Some also show a slight warming in the South Atlantic. Other impacts that can be considered robust in that they are seen in many different models are a reduction in evaporation and precipitation over the North Atlantic and a southward shift of the Atlantic ITCZ (Kageyama et al., 2013; Stouffer et al., 2006; see Figure 11). This shift in the ITCZ causes very large changes in seasonal rainfall locally, which could have seasonal impacts on the Amazon (Jackson et al., 2015; Parsons et al., 2014) and Africa (Chang et al., 2008). Outside the Atlantic, many models show large changes in tropical rainfall. Exact regions vary between models, but impacts have been shown on the Indian monsoon and on ENSO (Timmermann et al., 2007; Williamson et al., 2017; Zhang & Delworth, 2005).

Impacts over Europe have particularly been studied, with changes in weather patterns seen from shifts in atmospheric circulation. A shift to a more positive winter North Atlantic Oscillation (NAO) and a strengthening of the North Atlantic storm track is seen in many models (Brayshaw et al., 2009; Woollings et al., 2012). Other impacts over Europe seen in models include changes to summer precipitation through shifts in
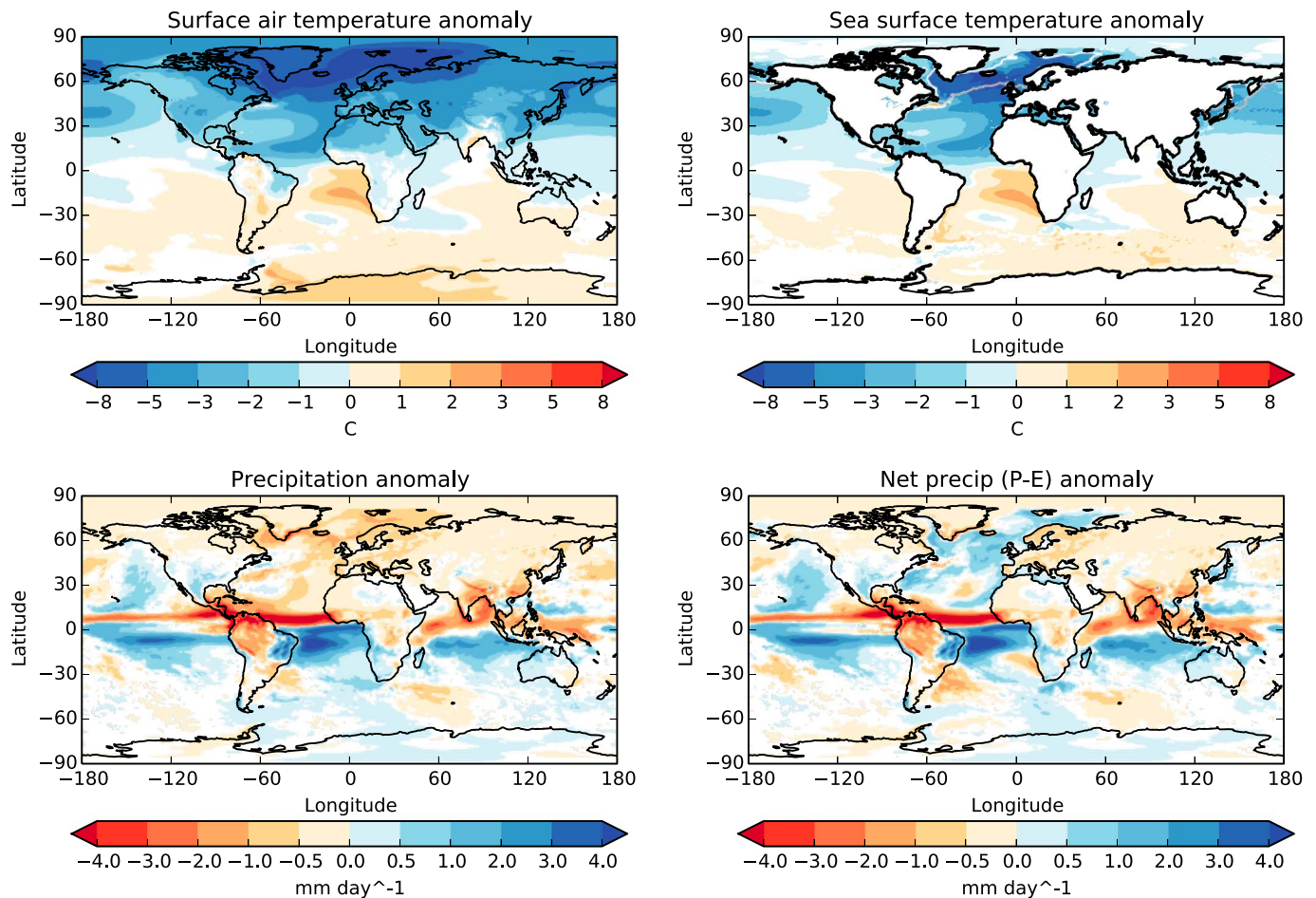
Figure 11. Anomalies of surface air temperature (top left, °C), sea surface temperature (top right, °C), precipitation (bottom left, mm/day), and net precipitation (precipitation-evapotranspiration, bottom right, mm/day). Anomalies that are not significant compared to the control variability (see text) are white. The top right panel also shows the extent of the March sea ice (concentrations >15%) for the control (white lines) and perturbed experiment (gray lines). Reproduced from Jackson et al. (2015), with permission from Springer, copyright 2015.

atmospheric circulation (Jackson et al., 2015; Haarsma et al., 2015) and changes in cloud cover, snow cover, and river runoff (Jackson et al., 2015; Jacob et al., 2005).

Several studies have shown how large changes in temperature and ocean circulation can impact sea level, particularly in the North Atlantic. This has potentially serious consequences for north east America and western Europe (Levermann et al., 2005; Little et al., 2017; Pardaens et al., 2011; Yin et al., 2009). As well as direct physical impacts, several studies have also examined how the carbon cycle, and terrestrial and marine ecosystems are impacted (Bozbiyik et al., 2011; Kuhlbrodt et al., 2009; Zickfeld et al., 2008).

If a collapse of the AMOC was associated with an increase in global temperatures, then impacts seen would be a combination of those from a warmer planet and those from a weakened AMOC. There is some evidence that at least the large-scale impacts are additive (Vellinga & Wood, 2008). Hence, whether North Atlantic temperatures cool, or just warm less than in other regions, would depend on the relative magnitudes and timescales of a weakening AMOC and a warming planet.

5.3. Early Warning Signs of an AMOC Collapse

The search for early warning signals of an impending AMOC collapse is an important but complicated endeavor. Even though methods might allow us to detect a change in the AMOC, the challenge is to distinguish the start of a collapse from longer-term variability or a gradual weakening. Models may help us to understand what natural variability looks like and help to detect signals of impending AMOC collapse (Roberts et al., 2014); however, there is a wide range of variability mechanisms, magnitudes, and timescales across models (Buckley & Marshall, 2016). At present no metrics have been found in models to distinguish

multidecadal variability from a systematic downward trend of the AMOC. There has also been no analysis to establish which properties would precede a collapse and at the same time do not precede a downward trend as, for instance, simulated by the CMIP5 multimodel ensemble (Reintges et al., 2017) in response to global warming.

One line of research focuses on physical precursors, that is, changes in properties (flow, water mass characteristics) that precede a collapse. Among the signals connected with a decreasing AMOC are the NADW masses getting lighter, a signal that has also been linked to a recent AMOC fluctuation (Robson et al., 2013). Long-term trends in this signal are, however, uncertain because temperature and salinity changes are strongly density compensated. Because of the larger variability in UNADW (Rhein et al., 2011), the clearest signal might come from LNADW, which is also the water mass in which the transport dropped associated with the reduced overturning after the 2011 relative to before 2007 (Smeed et al., 2014, 2018). Another precursor might come from measuring the amount of cooling and freshening of the East Greenland Current when circulating in the Labrador Sea, to assess the net downwelling in the boundary current that is now thought to dominate the sinking in the basin (contrary to the earlier hypothesized link with deep convection in the interior; Katsman et al., 2018). Finally, in one GCM study, Jackson and Wood (2018a, 2018b) find that differences in AMOC strength and mixed-layer depths indicate whether states recover or not after hosing. The AMOC is currently being monitored by the RAPID array (McCarthy et al., 2015), and the recently deployed OSNAP observing system (Lozier et al., 2017) will extend this capability. These networks will potentially be able to detect such early warning signs of AMOC changes.

Earlier indications of a change in the AMOC strength could also come from seasonal-decadal prediction systems which have shown some skill in predicting their own AMOC (Collins et al., 2006; Pohlmann et al., 2013), although some states might be easier to predict than others (Hermanson & Sutton, 2010). Observable indicators such as North Atlantic densities (Hawkins & Sutton, 2008; Roberts et al., 2013; Robson et al., 2013) or temperature patterns (Zhang, 2008) might precede changes in the AMOC or might have a greater signal-to-noise ratio making a significant change easier to detect (Baehr et al., 2007).

Time series analysis has also been proposed as method of early warning: this uses statistical properties of a time series to detect approaching a threshold (Boulton et al., 2014; Kleinen et al., 2003; Lenton et al., 2012). Although this method shows some promise, it also cannot show how far away the threshold is and requires hundreds of years of observations which are not yet available. However, if sufficiently robust, high-resolution paleo-proxies are available in the future; time series analysis may prove useful.

6. Discussion and Conclusions

In this paper we reviewed our current understanding of the (equilibrium) stability of the AMOC, in view of the possibility that the current climate system may allow other equilibria, with a different configuration of the ocean's overturning circulation. Based on our understanding of the available evidence, we cannot rule out the possibility that the AMOC in our current climate system is in, or close to, a regime of multiple equilibria. We base this conclusion on the following facts.

- Multiple equilibria have been robustly found in the full spectrum of models of the AMOC, including box models, two-dimensional or zonally averaged models, ocean general circulation models, coupled climate models, and quasi-adiabatic models of the overturning circulation.
- The dynamical process responsible for multiple equilibria, namely, the salt-advection feedback, is well understood and consistently found in this hierarchy of models.
- No feedback has been identified yet that would be strong enough to incapacitate the salt-advection feedback.

Still, there are many questions that need to be answered. We will discuss the most pressing research questions in the following subsections.

6.1. Does the AMOC Have A Stable Off-State in the Most Realistic Ocean Models

Climate models are achieving increasingly higher resolutions. This makes the representation of the dynamics more realistic, since more processes are explicitly resolved and fewer processes have to be parameterized. A more realistic representation of the Agulhas system, narrower western boundary currents, and changes in eddy transport can all affect the transport of freshwater within the Atlantic (Mecking et al., 2016; Treguier et al., 2012). High resolution reduces modeled temperature and salinity biases (Roberts et al., 2018; Sakamoto

et al., 2012; Small et al., 2014), and improvements in salinity biases can change the value of F_{ov} and its sign (Mecking et al., 2017). There is also evidence that the total transport of heat and freshwater within the Atlantic are stronger at higher resolution (Roberts et al., 2016; Treguier et al., 2012). Mecking et al. (2016) suggest that eddy transports in an eddy-permitting model allow for a stronger northward salt transport by the AMOC and therefore a stronger advective feedback, helping to maintain an AMOC off state.

However, higher spatial resolution makes these models more and more expensive to run. The current generation of eddy-resolving configurations (about 0.1° spatial resolution) are typically run for decades or centuries at most. This is well short of the typical millennial timescale that it takes for the ocean circulation to reach a global equilibrium, making it impossible to ascertain whether a certain AMOC state is stable in the limit of $t \to \infty$. Also, this makes it all but impossible to perform the systematic hysteresis studies that have been used to locate the limit points in lower-resolution models.

Recent experiments show that ocean models at so-called eddy-permitting resolution (typically 0.3–0.25°) can maintain a stable off-state for several hundreds of years (Jackson & Wood, 2018a, 2018b; Mecking et al., 2016). Yet, it is critical that these experiments are performed with global eddy-resolving, rather than just eddy-permitting, models to investigate whether multiple equilibria exist even in the most realistic models available. Hopefully, the ever-increasing performance of the World's supercomputers will allow us to routinely perform millennial timescale hysteresis experiments within the next decade. Also, the development of variable-resolution models may allow us to reduce the simulation cost by applying high resolution in regions where it matters (e.g., Southern Ocean and western boundary currents) while allowing for lower resolution in regions that are less critical (e.g., gyre interiors; Ringler et al., 2013). Finally, despite the higher cost of running high-resolution models, they will hopefully converge closer to the truth, as more processes are explicitly resolved and fewer processes are represented by inadequate parameterizations and subjective parameter choices. Case in point, Rahmstorf et al. (2005) show that the spread in hysteresis width is decidedly smaller for more complex models compared to EMICs.

6.2. Where Are the Stability Thresholds and How Close Is the Current AMOC to the Regime of Equilibria

The realization that the sign of a simple metric like F_{ovS} could tell us something about the stability regime of the AMOC was an important breakthrough, since F_{ovS} is reasonably well constrained by observations and a clear target for bias reduction in climate models. It is critical that the validity of this metric as stability indicator is tested in eddy-resolving models, by investigating whether a collapsed state can be reached (and maintained) in high-resolution model configurations that feature $F_{ovS} < 0$.

Yet, even though it is important to know whether the AMOC is in a monostable or bistable regime, the more relevant questions are as follows: What (temporary) perturbation or (permanent) change in forcing (or combination of both) would it take to trigger an irreversible shutdown of the AMOC? Would plausible rates of future Greenland Ice Sheet melt be sufficient to destabilize the AMOC? Could plausible scenarios of continued anthropogenic forcing change the global heat and freshwater cycles to such an extent that an active AMOC can no longer be sustained?

We are not aware of any successful attempts to formulate a simple criterion to define the *other* limit point, L_1 in Figure 4, that is, the magnitude of forcing perturbations beyond which the on-state no longer exists, and one can be forgiven to think that the community's focus on F_{ovS} has taken away the attention from the threshold that really matters. But characterizing the conditions for which a strong AMOC is no longer a feasible equilibrium should be a priority of the research community.

6.3. What Are the Main Feedbacks That Compete With the Salt-Advection Feedback

As discussed throughout the paper, the stability of the AMOC is strongly linked to the salt-advection feedback. There seems to be little doubt that the relatively high salinity of the subpolar North Atlantic reflects the import of salty subtropical waters by the AMOC nor that the strength of the AMOC is somehow linked to the formation of deep water that takes place in those regions. Yet, as discussed in section 3, many other feedbacks affect the strength and stability of the AMOC, including the role of shifts in the ITCZ, and many of these feedbacks have not been properly quantified.

For instance, how will receding Arctic sea ice affect the strength and stability of the AMOC? In fact, recent studies indicate that the ongoing Arctic sea ice decline can substantially weaken the AMOC (Liu & Fedorov, 2019; Liu et al., 2019; Sévellec et al., 2017). This weakening occurs as heat and freshwater anomalies due to

the contraction of sea ice cover accumulate in the Arctic and then spread to the North Atlantic, suppressing deep convection. The freshening is further amplified by the salt-advection feedback. In turn, the reduction of poleward heat transport due to a weaker AMOC partially stabilizes sea ice cover. Such two-way interaction between the AMOC and Arctic sea ice, involving both positive and negative feedbacks, is an important part of the AMOC response to external forcing, and the global impacts of sea ice decline critically depend on whether the AMOC is affected or not (Liu & Fedorov, 2019; Thomas & Fedorov, 2019). A greater understanding of the relative strengths and timescales of such different feedbacks across models is needed. We also need to understand the relative importance of feedbacks affecting the whole Atlantic to those affecting the subpolar North Atlantic in particular.

6.4. How Can We Detect or Predict an Impending AMOC Collapse

Many studies have found relationships between the AMOC and various observable indicators (Roberts et al., 2014; Robson et al., 2013). However, as discussed in section 5.3, there is little knowledge of indicators that could give an early warning of a collapse, as opposed to a more gradual weakening or long time-scale variability. This is a clear knowledge gap that needs to be addressed with some urgency, given the rapid pace of climate change and the observed or implied decline in the AMOC (e.g., Caesar et al., 2018; Rahmstorf et al., 2015; Smeed et al., 2014, 2018; Thornalley et al., 2018). In our view, addressing this problem requires a multifaceted approach.

First, the community needs to identify the metrics that robustly predict a pending collapse among a wide range of models (if such metrics exist), be it a measure of mixed-layer depth in the subpolar North Atlantic, density of NADW at certain key locations, or the rate of downwelling in the rim current of the Labrador Sea. This will require a generation of models that has reduced salinity biases in the Atlantic Ocean and more consistent stability properties, as well as high enough resolution to accurately represent small-scale processes like deep convection and narrow boundary currents.

Second, these metrics need to be routinely targeted by observational programs. The RAPID/MOCHA program has provided us with an unprecedented view of the state of the AMOC and its variability. Even though it is not clear whether its observations, or that of the newly deployed OSNAP system, observe metrics that are useful indicators of a pending AMOC collapse, the success of these programs demonstrates that deployment of a long-term operational monitoring system is feasible; we just need to know what we should be looking for.

Finally, these observations then need to be extrapolated by decadal prediction systems. In recent years, scientists have made tremendous progress in the development of prediction systems that display skill in predicting certain aspects of ocean evolution, up to a decade in advance (Yeager & Robson, 2017). It will be exciting to see these systems reach full maturity in the next decade or so; once they do, they will certainly be challenged to show that they are capable of predicting the crossing of a nonlinear threshold in the AMOC.

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