

MASTER THESIS

Stability of the Atlantic Meridional Overturning Circulation

Author:

Anton WIMMER

Supervisors:

Prof. Dr. Gerrit LOHMANN

Prof. Dr. Joke LÜBBECKE

June 19, 2026

Fachbereich 01 - Physik / Elektrotechnik
Institut für Umweltphysik

Contents

1	Introduction	6
2	Methodology	11
2.1	Introduction to Box Models	11
2.1.1	Stommel Model	11
2.1.2	Rahmstorf Model	13
2.2	Nine-Box Atmosphere-Ocean-Coupled Box Model (9-BM)	16
2.3	Experimental Design	18
2.4	Data Analysis of AWI-ESM	19
2.4.1	Definition of Relevant Parameters	20
2.4.2	ESM Data	22
3	Results	23
3.1	Results of the 9-BM	23
3.1.1	Results of Warming for Different Variants of the 9-BM	25
3.1.2	Hysteresis of the 9-BM	27
3.2	Results of AWI-ESM	27
3.2.1	Atlantic Meridional Overturning Circulation	29
3.2.2	Salinities	30
3.2.3	Oceanic Freshwater Fluxes	33
3.2.4	Oceanic Heat Fluxes	37
3.2.5	Atmospheric Freshwater Fluxes	38
3.3	Comparison of the Freshwater Fluxes of AWI-ESM and the 9-BM	38
4	Discussion	43
4.1	Comparison of the 9-BM with AWI-ESM	43
4.2	Stability Mechanisms of the 9-BM and AWI-ESM on AMOC	50
4.2.1	Stability of AMOC in the 9-BM	50
4.2.2	Stability of AMOC in AWI-ESM	54
4.3	Is the Overturning Freshwater Flux a good indicator of AMOC stability?	55

5	Conclusions	57
	Sources	59
6	Appendix	65

List of Figures

2.1	Sketch: Stommel and Rahmstorf BM	12
2.2	9-Box Coupled Atmosphere-Ocean Model (9-BM)	16
2.3	Warming Scenarios 9-BM	19
2.4	Precipitation Areas	20
2.5	Salinity and Velocity Distribution	20
3.1	Prognostic Variables 9-BM	24
3.2	Stability Map: Warming With Varying Parameters	26
3.3	Hysteresis of 9-BM	28
3.4	Streamfunction in AWI-ESM	29
3.5	AMOC strength AWI-ESM	30
3.6	Salinity of Ocean Basins, per Scenario and Depth	34
3.7	Flux Decomposition	35
3.8	Overturning Freshwater Flux Over Depth and Time	36
3.9	Heat Fluxes	37
3.10	Net Precipitation over Basins	39
3.11	Comparison of the Freshwater Fluxes in the ESM and the BM	40
4.1	Salinity Comparison ESM and BM	48
4.2	F_{ovS} Decomposition	48
4.3	Stability Map: Oceanic Interaction	52
4.4	Density Anomaly Transport	53

List of Abbreviations

AMOC	Atlantic Meridional Overturning Circulation
PMOC	Pacific Meridional Overturning Circulation
SSP	Shared Socioeconomic Pathway
SSP5-8.5_{ice}	SSP5-8.5 with the Greenland ice sheet activated
GHG	Greenhouse Gas
CO₂ (eq)	Carbon Dioxide (Equivalent)
GIS	Greenland ice sheet
NA	North Atlantic
NADW	North Atlantic Deep Water
ESM	Earth System Model
NO	Northern Ocean
AUPO	Atlantic's Upper Part
ALPO	Atlantic's Lower Part
SO	Southern Ocean
IPO	Indo-Pacific Ocean
BM	Box Model
9-BM	Box Model used in this thesis, with 9 boxes
NB	Northern Box
MLB	Mid-Latitude Box
DWB	Deep Water Box
SB	Southern Box
IPB	Indo-Pacific Box
NB-A	Northern Atmospheric Box
MLB-A	Mid-Latitudinal Atmospheric Box
SB-A	Southern Atmospheric Box
IPB-A	Indo-Pacific Atmospheric Box

Abstract

The Atlantic Meridional Overturning Circulation (AMOC) plays a critical role in global ocean heat transport. Research suggests that the AMOC is weakening due to anthropogenic global warming, which is projected to disrupt global precipitation patterns, intensify climate extremes, and accelerate sea-level rise. Consequently, a comprehensive understanding of the mechanisms driving these changes in AMOC is crucial. While complex Earth System Models (ESMs) are indispensable for projections, they are computationally expensive. Conceptual box models provide an interesting alternative, enabling the investigation of fundamental climate feedback mechanisms.

In this thesis, a coupled 9-box atmosphere-ocean model (9-BM) that includes the Atlantic and the Indo-Pacific is introduced. Using this framework, AMOC stability was probed under different emission scenarios and coupling strengths of the Atlantic to the Indo-Pacific. The conceptual model's behavior was compared against the comprehensive Alfred Wegener Institute Earth System Model (AWI-ESM) to analyze the impact of different drivers.

The conceptual model suggests that the typical freshwater overturning flux at the southern boundary of the Atlantic may not have the definitive significance on the AMOC stability, as is often attributed to it. The Indo-Pacific has a destabilizing effect on the AMOC, particularly so in high-emission scenarios. Finally, the model indicates that following a severe perturbation induced by global warming, the AMOC may require several millennia to relax into a new equilibrium state, undergoing a number of transition states first.

Chapter 1

Introduction

Global Warming and System Stability

Global warming is one of the biggest challenges facing humanity today. The anthropogenic emission of Carbon Dioxide (CO_2) leads to a reduction in Earth's thermal emission to space, and therefore to a general warming of the Earth system (Pierrehumbert, 2011). This general warming is accompanied by a general change of circulation patterns, which are changing mutually dependent on each other, and are therefore notoriously hard to predict. A better understanding of the potential changes is a necessary prerequisite to take purposeful actions in preventing harmful pattern shifts or preparing society for them.

If the Earth system is not disturbed over a sufficiently long period, stable equilibria are reached for many subsystems. A rapid change in an external forcing may perturb these systems, which in turn may cause a perturbation in other systems. Because of their mutual dependence, a mutual positive feedback may occur. If system states then change qualitatively (David Tàbara et al., 2018), an even stronger perturbation of other systems is caused, potentially leading to a tipping cascade (Wunderling et al., 2021).

A general warming of the Earth system has a number of interactions of subsystems, which may change their behavior, which in turn can amplify or dampen the perturbation caused by warming. Different systems that may change their states by warming are the Amazon forest (Nobre & Borma, 2009), the permafrost environment near the north pole (Anisimov, 2007), or the Atlantic Meridional Overturning Current (AMOC). Both the Amazon forest and the arctic permafrost may turn from a stable system that stores CO_2_{eq} in the trees or ground, to another state, where they emit CO_2_{eq} into the atmosphere, amplifying the initial warming (David Tàbara et al., 2018).

The consequences of a change of AMOC due to global warming are more complex.

Description of AMOC

The AMOC spans from the Southern Ocean (SO) to the North Atlantic (NA). On its upper limb, it transports warm, salty water northwards, and on its lower limb, it transports cooler, fresher water southward. In the NA, the water cools, increases its density, sinks, and flows back on the lower limb as North Atlantic Deep Water (NADW). In the SO, this NADW is up-welled by wind-driven gyres (Jackson et al., 2022).

Consequences of a Potential AMOC Slowdown or Collapse

The global increase of global temperature will lead to a slowdown of AMOC, or even its sudden collapse (Liu et al., 2017). Indeed, observational data indicates that a significant weakening of the circulation is already underway (Caesar et al., 2018). A further slowdown of AMOC will have a cooling effect in the northern hemisphere, and especially so in Europe, where, according to climate model analysis, it would lead to a cooling of several degrees (Jackson et al., 2015). A slowing of AMOC is projected to affect regional extreme weather events (Ma et al., 2024). If AMOC collapses, a wide array of climate models predict a strong decrease in precipitation of the West African Monsoon (-29 %) or the Indian Summer Monsoon (-19 %) (Ben-Yami et al., 2024). The reduction of temperature in Europe and the reduced precipitation in other regions have severe consequences for agriculture.

External Drivers of a Potential AMOC Slowdown or Collapse

The weakening of AMOC due to an increase in global temperatures is primarily driven by changes in the North Atlantic, largely due to Arctic Amplification, which results in disproportionately high warming in the Arctic region compared to the rest of the globe, in particular to the Southern Ocean. Meltwater influx from the Greenland ice sheet and glaciers causes the surface waters in the NA to freshen (Dai, 2022). Both effects, the warming and the freshening of the water, lead to an increase in vertical stratification, which inhibits deep convection in the NA. Consequently, the formation of NADW is weakened, slowing down the overall circulation (Levang & Schmitt, 2020).

Internal Drivers of a Potential AMOC Slowdown or Collapse

The NADW-flow and hence circulation strength of AMOC is theorized to be proportional to density gradients in the NA convection regions and the SO up-welling regions, as was indicated by analysis of General Circulation Model (GCM) results (Rahmstorf, 1996)(Johnson et al., 2019)(Weber et al., 2006). Atlantic water in the upper limb is transported by AMOC into the deep convection regions. This transported water is warmer and saltier than the water in the deep convection regions. The temperature increase in

the deep convection regions increases the buoyancy of the water, while the salinity increase lowers it. Since these transports scale with, and are themselves influencing AMOC strength, they are causing feedbacks. The heat transport into the deep convection regions weakens the NADW formation, therefore slowing the AMOC, hence slowing the heat transport. The feedback of heat transport is therefore slowing itself, called a *negative feedback*. A salinity transport into the sinking region, on the other hand, amounts to a buoyancy loss, an increase in NADW formation, and hence a strengthening of circulation, increasing the salinity transport into the deep convection regions further. This salinity transport is then said to have a *positive* or *amplifying feedback*.

On the other hand, a slowing down of AMOC will cause a decrease in the strength of these feedbacks. If the salinity feedback is in some way dominant over the temperature feedback, the reduction of salinity transport into the deep convection regions will cause a further reduction of AMOC, hence salinity transport. This reduction of salinity transport and its feedback transports then might lead from a weakened AMOC to a further weakened AMOC and possibly to its collapse (Rahmstorf, 1996). Simulations and observational data tend to disagree on the strength, and sometimes even on the *direction* of this salinity transport (van Westen & Dijkstra, 2024).

AMOC Box Models and the Derived Quantity " F_{ovs} "

For a better understanding of these feedback mechanisms and of AMOC in general, a number of conceptual models were formulated. These range from simple two-box models (Stommel, 1961), only taking into account mixing between the boxes and a freshwater forcing, to more complex models that take into account different parts of the ocean (Wood et al., 2019)(Rahmstorf, 1996). Even in the simple two-box models, which can be exactly solved (Lohmann & Schneider, 1999), multiple stable states exist (see Fig.2.1 a) (Stommel, 1961). Starting from this fairly abstract view, additional boxes were added to capture the dynamics of AMOC better. The model proposed by Rahmstorf (1996) not only consists of an up-welling Southern Box and convecting Northern Box, but also contains a box for the upper and lower limb of AMOC (see Fig.2.1 b). An introduction to these box-models is given in Chapter 2.1.

Inside the framework of the Rahmstorf model, the oceanic freshwater overturning flux entering the Atlantic from the Southern Ocean (F_{ovS}) is of importance for the stability of the AMOC. If this flux is negative ($F_{ovS} < 0$), meaning AMOC transports saltwater into Atlantic, two stable AMOC-states, one positive and one reversed, are possible (Drijfhout et al., 2011)(Rahmstorf, 1996). Starting from this theoretical assumption, the freshwater overturning flux was calculated at the southern boundary of the Atlantic basin at about 34°S from observational data (Bryden et al., 2011) (Arumí-Planas et al., 2024) and in GCMs (de Vries & Weber, 2005). A reanalysis of the observational data suggests that

F_{ovS} is negative in the real ocean (Rousselet et al., 2021). This behavior of F_{ovS} is not replicated by most GCMs (van Westen & Dijkstra, 2024). Following the argument, most GCMs underestimate the risk of a possible AMOC collapse.

In this thesis, a new Box Model (9-BM) is presented (Sec. 2.2). The model consists of nine boxes, five oceanic and four atmospheric boxes. Processes that were considered relevant for the AMOC system were added.

Important added features, relative to the Rahmstorf model, include an Indo-Pacific Box that is coupled to the AMOC-boxes. Atmospheric boxes that are coupled to their corresponding oceanic boxes (the atmosphere above their oceanic region), as well as to their adjacent atmospheric boxes. The atmospheric freshwater fluxes are not fixed, as in Stommel’s or Rahmstorf’s model, but are assumed to be dependent on the atmospheric temperatures and the gradient of the temperature between the atmospheric boxes. This is done to get a more realistic understanding of the behavior of AMOC and the relevance of features not captured by a simpler framework. Another aspect of this thesis will be the analysis of data from the Earth System Model (ESM) used at the Alfred-Wegener-Institut, AWI-ESM. Starting from the framework of the *9-BM* presented, different oceanic and atmospheric fluxes in the ESM between defined oceanic regions are calculated (Sec. 2.4). To capture the dynamics of the ESM at different warming rates, two different extended warming scenarios (Shared Socioeconomic Pathways) SSP2-4.5 and SSP5-8.5 were analyzed (Meinshausen et al., 2020). Additionally, runs with an interactive Greenland ice sheet were compared to runs with a fixed Greenland ice sheet. Since the melting of Greenland ice is introducing freshwater into the NA, and therefore possibly causing a slowdown or even a collapse of the AMOC, it is important to understand whether a more realistic behavior of the Greenland ice sheet is changing the overall behavior of the Earth system in general, and the AMOC in particular.

Research Questions

In this thesis, a box model is introduced to project AMOC behavior and is subsequently compared to AWI-ESM output. To evaluate these different modeling approaches, this work addresses the following research questions:

RQ1: How do the 9-BM and AWI-ESM AMOC mechanisms compare?

The 9-BM will be analyzed to identify the underlying mechanisms driving either the stability or the collapse of the AMOC. This leads to the second research question:

RQ2: Which mechanisms are important for AMOC stability in the 9-BM (and AWI-ESM)?

Compared to the foundational box model introduced by Rahmstorf (1996), the 9-BM incorporates additional physical processes to yield more realistic results. Consequently,

the third question addresses a key diagnostic parameter of this system:

RQ3: How relevant is the parameter F_{ov} for AMOC stability in the 9-BM and AWI-ESM?

Chapter 2

Methodology

2.1 Introduction to Box Models

In climate science, models are used to get an understanding of the complex interaction of different Earth system elements. In terms of complexity, there exists a hierarchy of models. The complexity is determined by the dimensionality of the model, spatial and temporal, and the number of processes used. While simple box models or conceptual models are used to understand simple mechanisms and interactions, more complex models allow for making projections, or to compare the current understanding of the mechanisms, with observational data (Dijkstra, 2024).

Box Models (BM) are used to gain a simple, qualitative understanding of the mechanism of a complex dynamic system. Each individual box contains a set of parameters. External forcings are defined that change the parameters in an individual box. The connection between the boxes is established via parameter fluxes, mimicking transport, which might depend on the parameters and change the parameters of both boxes.

In an oceanic BM, the main parameters of the individual boxes are salinity and temperature. The associated forcings are salinity, or freshwater forcings, as well as heat forcings on the individual boxes. The fluxes between the oceanic boxes transport both salt and heat, changing the salinity and temperature of both boxes.

2.1.1 Stommel Model

The Stommel Model is a simple BM of a dynamic ocean system (Stommel, 1961). Using its very simple formulation, it can be shown that for a set of forcings, the same external forcings allow multiple possible stable states. Outside of this set of forcings, only one state is allowed. At the boundary of this set, a bifurcation point exists, where the system is changing from one stable state to another.

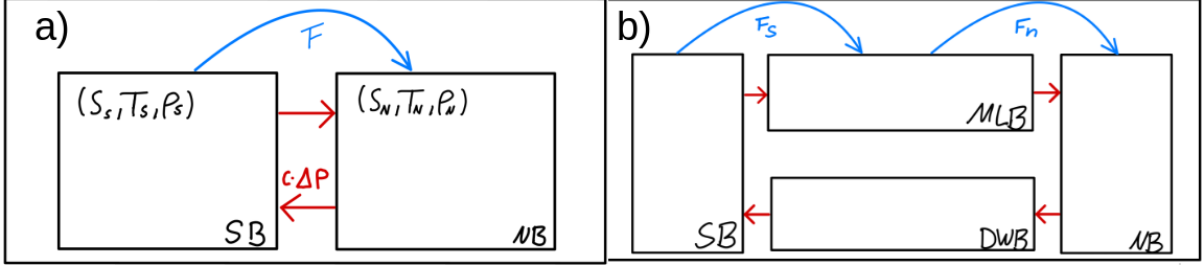


Figure 2.1: Box models are used to simulate basic ocean mechanisms. Each ocean box has the prognostic parameters salinity (S_i) and temperature (T_i), and therefore a density ($\rho_i(S_i, T_i)$). The transport between the ocean boxes, indicated by red arrows, is dependent on the density difference between the Northern Box and the Southern Box. The atmospheric freshwater transport is indicated by blue arrows, with the fixed strength F . The left box model a) is the *Stommel Model* (Stommel, 1961) and the right box model b) is the *Rahmstorf Model* (Rahmstorf, 1996) of the AMOC. The box *MLB* represents the Atlantic's upper part, and *DWB* the lower part of the Atlantic.

Definition of the Model

A sketch of the Stommel model is shown in Figure 2.1 a). In the formulation of Stommel's model, there are two boxes containing the parameters temperature, salinity, and, dependent on these, density.

Both boxes experience heat forcings. The so called Northern Box (NB) is externally forced by a lower temperature than the so called Southern Box (SB). Additionally, a freshwater forcing acts on the individual boxes. In the warmer SB, salinity is increased, while in the colder NB, the salinity is decreased by the same amount. This forcing, acts as an external freshwater flux between the boxes, which can be interpreted as an atmospheric freshwater flux.

The overturning flux (Φ) between the boxes is determined by the density difference. The density (ρ) is a function of salinity (S) and temperature (T) of the individual boxes. ΔT and ΔS denote the parameter difference between the NB and SB. The thermal expansion coefficient is denoted by α and the haline contraction coefficient by β . T_0 and S_0 are reference values.

$$\rho = \rho_0 - \alpha(T - T_0) + \beta(S - S_0) \quad (2.1)$$

$$\Delta\rho = -\alpha\Delta T + \beta\Delta S \quad (2.2)$$

$$\Phi = \text{cons} \cdot \Delta\rho \quad (2.3)$$

This definition mimics the physical process that denser water masses displace lighter water masses on the bottom of the ocean, as they expand into another region. To balance this flow of denser water at the bottom, upper water masses of the other region flow

back into the region, where the denser water masses originated. The flux Φ exchanges, depending on $\Delta\rho$, salinity, and temperature, between the boxes. In this formulation, the motion of water is kept, but the stratification of the water masses corresponding to the boxes is neglected.

Dynamics of the Model

If the initial parameters of the boxes are the same, and the freshwater flux is neglected, the water will sink in the NB, indicated by a positive Φ , since the heat forcing at the NB is smaller than at the SB, therefore denser.

An increase in the northward freshwater flux (F) will lead to a salinity difference between the boxes. This differential is weakened by the water flux, which mixes the water masses. A further increase in the northward freshwater flux increases the salinity difference, thereby lowering the density difference, weakening the water flux. Since the strength of the water flux determines the reduction of the salinity difference, this poses a feedback: A higher ΔS leads to a weaker Φ , which in turn leads to a weaker reduction of ΔS ; an introduction of a higher F therefore leads to a non-linear increase of ΔS .

If Φ is weakened such that it can not reduce a slight increase ΔS anymore, the system changes modes. Φ becomes negative, which in the physical interpretation means that the water sinks in the SB. In this *haline mode*, the dampening ΔT is not reduced, as is the dampening ΔS , stemming from F in the thermal mode. Therefore, the overturning flux is weaker. Since the overturning flux is weaker, the reduction of ΔS is too. ΔS , which now drives the circulation, can not be reduced as fast as in the thermal mode.

A small decrease of F therefore does not lead to a switch of modes, until a threshold value is reached, when the salinification of the SB can not fight the temperature forcings anymore.

Starting from the freshwater flux, where the thermal mode collapses, until the freshwater flux, where the haline mode collapses, two stable equilibria are possible, with the same forcing parameters.

2.1.2 Rahmstorf Model

Stefan Rahmstorf created his own box model, taking inspiration from Stommel's model, and further developed it to incorporate more realistic ocean processes. Between the Southern and the Northern Box, a Mid-Latitude Box (MLB) is introduced for the upper water mass, and a Deep Water Box (DWB) for the lower water mass (Rahmstorf, 1996).

Definition of the Model

The Rahmstorf model boxes (see Fig. 2.1 b) represent different regions of the Atlantic Ocean. The NB represents the convection region in the NA, the SB the upwelling region

in the SO, the MLB represents Atlantics Upper Part, the Atlantic Surface Water and the Antarctic Intermediate Water, and the DWB represents the Atlantics Lower Part.

The temperatures of the boxes are fixed, though the qualitative behavior is not influenced, if instead a heat forcing is introduced on the NB, MLB, and SB. The southern and MLB are forced by a freshwater flux (F_s), as well as the MLB and the NB (F_n).

A water flux representing the AMOC connects the neighboring boxes. In the positive mode, water is advected from the SB to the MLB, from the MLB to the NB, from the NB to the DWB, and from the MLB to the SB.

The water flux representing the strength of the AMOC is again set proportional to the density difference of the NB and SB. In the case of an equilibrium, the salinity transport via the AMOC and the freshwater flux must cancel each other out. A positive freshwater flux in the atmosphere corresponds to a negative freshwater flux in the water transport of the same strength.

Dynamics of the model

The sign of the external freshwater fluxes is allowed to be negative. This implies that the system is symmetric for positive and negative AMOC strength (Φ). Without loss of generality, a positive Φ is assumed.

Assuming that the AMOC-mode is positive ($\Phi > 0$), it can be shown that the strength of the AMOC is independent of the freshwater flux F_n . For fixed temperatures, the strength is then only dependent on F_s . If F_s is negative, the salinity in the SB decreases. This makes a flow regime possible for $T_n > T_s$, which is haline driven. If $T_n < T_s$, both the temperature and the salinity forcing work to strengthen the circulation. If F_s is increased, such that $F_s > 0$, the thermal and haline forcing act in opposite directions. A salinity increase in the SB, which in first order weakens AMOC, gets advected proportionally to the AMOC strength to the MLB. The AMOC strength acts against the increase in salinity in the SB, by importing less saline deep water, as well as indirectly increasing the salinity in the NB and DWB. This reduction effect on the salinity difference is weaker, if due to a further increase of F_s , the salinity increase weakens the AMOC strength. The effect of the AMOC strength on the salinity difference is reduced, increasing the density difference between the NB and SB further, which in turn decreases the AMOC strength further. At a critical value for F_s , the system collapses and enters a reverse mode, or, depending on F_n , an off-mode.

If the AMOC transports saltwater into the NB, it is amplifying itself. In the model, this feedback is set in motion by an increase in salinity in the SB.

More generally, if the AMOC is dependent on stability on this self-stabilizing feedback, a freshwater perturbation in the north, which lowers the density gradient and weakens AMOC, leads to a further weakening.

If instead AMOC was stable with a northward freshwater advection, a weakening would lead to a lower freshwater import into the north by AMOC, strengthening it again.

Therefore, the northward transport of salinity that is driven by AMOC, and if the transport acts as freshening or salinification, is an important characteristic in terms of AMOC-stability.

2.2 Nine-Box Atmosphere-Ocean-Coupled Box Model (9-BM)

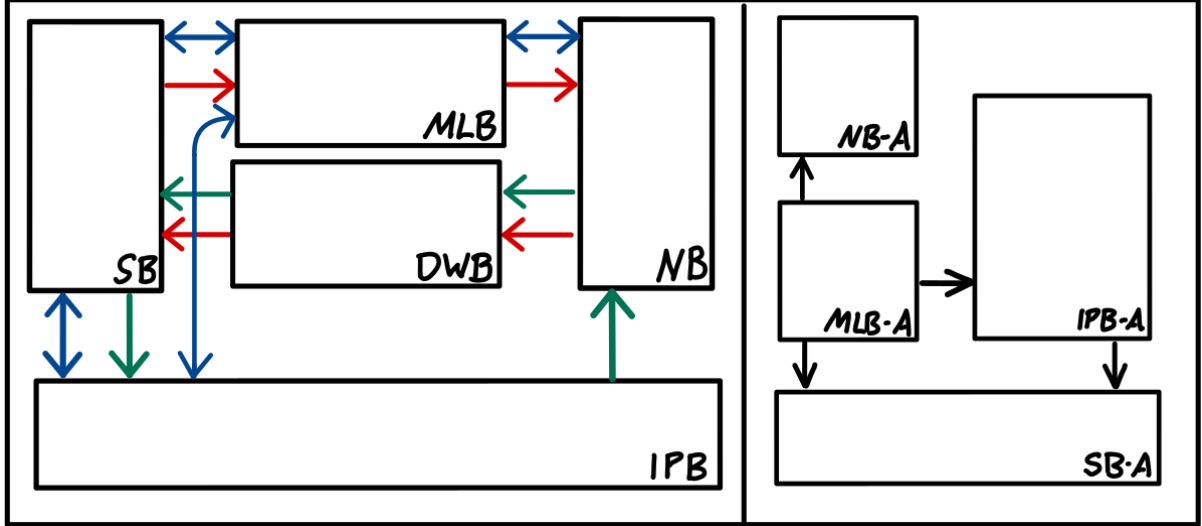


Figure 2.2: On the left-hand side, the ocean model and the coupling between the boxes are shown. The letters in the bottom right designate the box. The red arrows indicate the flow by the AMOC, driven by the density gradient from the NB to the SB; the blue arrow indicates the mixing, driven by gyres, the green arrow the Bering Strait transport (and salt conservation). On the right-hand side, the atmosphere boxes are shown, labeled the abbreviation for the corresponding ocean box, to which it is coupled and then an A. The arrows on the right-hand side indicate the atmospheric freshwater and heat transport.

The 9-Box Coupled Atmosphere-Ocean Model (9-BM) used in this study (Fig. 2.2) to analyze the AMOC is a modified version of the 7-Box Model developed by Lohmann (2012). Ocean boxes are coupled with atmospheric boxes. The ocean boxes correspond to water masses in the North Atlantic, the Atlantic, the Southern Ocean, and the Indo-Pacific. The Atlantic is divided into a Mid-Latitudinal Box and a Deep Water Box.

The boxes are coupled by advection and mixing terms. The advection terms correspond to the meridional overturning circulation (red arrow), and the Bering Strait (green arrow) and its corresponding terms, to yield over all salt-conservation. The mixing terms (blue arrow) are related to the ocean gyres. Both quantities, salinity and temperature, are advected and mixed. Mass and salinity conservation is prescribed.

The circulation term connects, in the positive phase of an active circulation, the MLB to the NA box, the NA box to the deep-water box, the deep-water box to the SB, and the SB to the MLB. In the negative phase, this transport is reversed in direction.

A Bering Strait term was added to account for water transport from the IPB into the NB. In order to conserve salt, a transport of the same strength connects the NB to the MLB, from there to the SB, and from the SB back to the IPB, closing the loop.

The advection terms are shown in Eqs. (2.4) for the positive phase. The indices denote the corresponding box. The salinity is denoted by S , the volume of the box by V . Φ is the circulation variable associated with the circulation strength, and is proportional to the density difference between the northern and southern boxes. The density $\rho(T, S)$ is calculated using a linear approximation. Be is the transport through the Bering Strait, and its corresponding conservation fluxes.

The added mixing terms correspond to mixing of the salinities at the boundaries of the defined regions, related to ocean gyres (Eqs. (2.5)). The mixing strength is denoted by M . The indices of M indicate which basins are mixing. A parameter M_{scal} is defined, which globally scales the mixing strength such that $m_{idx} \cdot M_{scal} = M_{idx}$.

The terms related to temperature advection and mixing are defined analogously.

$$\begin{aligned}
V_n \frac{dS_n}{dt} \Big|_{adv} &= (S_m - S_n)\Phi + (S_{ip} - S_n)Be & V_n \frac{dS_n}{dt} \Big|_{mix} &= M_{nm}(S_m - S_n) \\
V_d \frac{dS_d}{dt} \Big|_{adv} &= (S_n - S_d)(\Phi + Be) & V_d \frac{dS_d}{dt} \Big|_{mix} &= 0 \\
V_s \frac{dS_s}{dt} \Big|_{adv} &= (S_d - S_s)(\Phi + Be) & V_s \frac{dS_s}{dt} \Big|_{mix} &= M_{ms}(S_m - S_s) + M_{sp}(S_{IPO} - S_s) \\
V_m \frac{dS_m}{dt} \Big|_{adv} &= (S_s - S_m)\Phi & V_m \frac{dS_m}{dt} \Big|_{mix} &= M_{ms}(S_s - S_m) + M_{nm}(S_n - S_m) \\
& & & + M_{Al}(S_s - S_m + 2) \\
V_{ip} \frac{dS_{ip}}{dt} \Big|_{adv} &= (S_s - S_{ip})Be & V_{ip} \frac{dS_{ip}}{dt} \Big|_{mix} &= M_{sp}(S_s - S_{ip}) + M_{Al}(S_m - S_{ip} - 2)
\end{aligned}
\tag{2.4}
\tag{2.5}$$

Except for the Deep Water Box, all oceanic boxes are connected to corresponding atmosphere boxes. The atmospheric box model described in Prange et al. (1997) was modified to fit to the specified oceanic boxes of the 9-BM. The oceanic and atmospheric boxes interchange heat, based on their temperature difference and predefined heat flux constants. The incoming short-wave radiation, taking into account the albedo, is heating the atmospheric box. The atmospheric boxes exchange latent and sensible heat, changing the temperatures of the atmospheric boxes.

The sensible heat flux between the atmospheric boxes is proportional to the temperature gradient, except for the transport from the MLB to the IPB. Since the transport distance is fixed by definition, the sensible heat flux is proportional to the temperature difference between the adjacent boxes.

The latent heat flux is calculated using an approximation of the Clausius-Clapeyron equation, taking as input a derived temperature at the box-boundaries, corresponding to the vapor transport from the MLB northward, southward and westward, and from the IPO southward. The transport distance and the relative humidity are assumed to be constant.

The transport of latent heat between atmospheric boxes is proportional to the atmo-

spheric freshwater transport. It depends on the temperatures of the atmospheric boxes and acts directly on the oceanic boxes through evaporation or precipitation.

The Agulhas leakage, which brings warm and saline water from the IPB into the MLB, was added. Since the IPB is less warm and saline compared to the MLB, a simple fixed transport would not correspond well to the physical reality. Therefore, the salinity and temperature of the water transported into the MLB were artificially increased, each by adding (subtracting) a fixed value from temperature (salinity).

To better capture the dynamics of global warming, a "melting Greenland" was added. A meltwater flux proportional to the Northern boxes Atmospheric (NB-A) temperature was included. Along with the meltwater flux, cooling of the NB-A (mimicking heat loss caused by the phase change of melting ice) and cooling of the NB (mimicking heat loss from 0°C water running into the ocean) were added.

The sum of the freshwater stemming from meltwater was fixed. The warming period was set to 150 simulation years.

In order to investigate the influence of the IPO, a variable that defines the strength of the coupling of the AMOC-boxes with the IPO was added and kept fixed for the run. This variable then influences the horizontal atmospheric freshwater transport, the atmospheric freshwater transport from the IPO, and also the volume and area of the SO and Southern Atmospheric Box (SB-A). Additionally, the strength of the Agulhas leakage, the Bering Strait transport, and the oceanic mixing of the Southern and the IPB can be decreased.

2.3 Experimental Design

The 9-BM was forced by a decrease in the thermal emission into space by the atmospheric boxes for 150 years. This approach was implemented to model the warming effect on AMOC caused by the anthropogenic CO_2 emission into the atmosphere (Fig. 2.3). This was repeated for different variations of model parameters, accounting for uncertainties in the parametrization of the model, in particular for the influence of the Indo-Pacific Box on the AMOC system.

The parameters varied account for different coupling strengths of the IPB to the AMOC boxes, divided into atmospheric and oceanic parameters. Each coupling strength was varied by multiplying it by a scalar from zero to one. The atmospheric interaction comprises the variation of the atmospheric freshwater and heat transport of the MLB to the IPB, and of the IPB to the SB. Additionally, in order to force the model into a positive state, the volume of the Southern Ocean was increased, with an increase in the atmospheric interaction.

The oceanic interaction is comprised of the coupling of the SB with the IPB via the mixing, the Bering Strait, and the Agulhas leakage term. Each of these terms was multiplied by a scalar from zero to one.

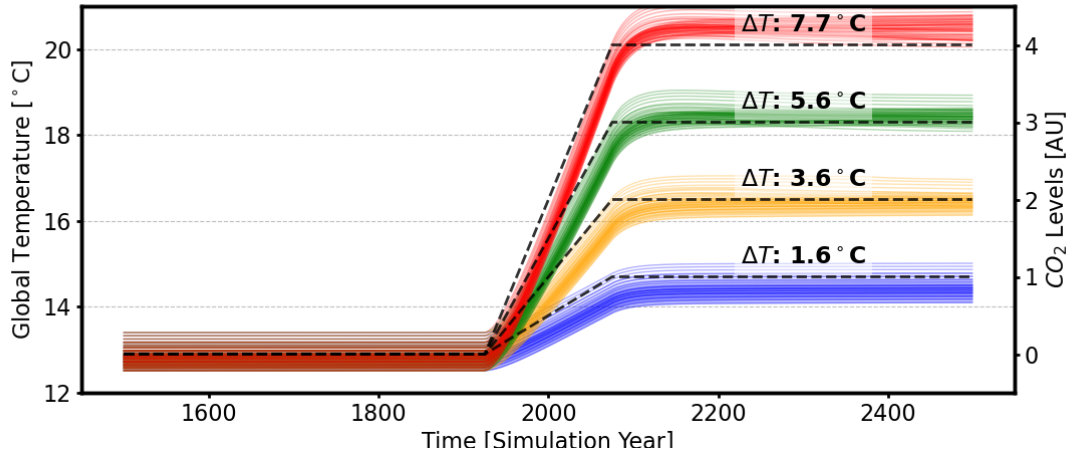


Figure 2.3: The 9-BM was forced by a decrease in the thermal emission, simulating an increase in the CO_2 -level for 150 years, shown in the black dashed line. The different runs consist of different parameter variations of the 9-BM, as well as four different thermal emission forcings. The global temperature graphs of each forcing scenario share the same color. Above each dotted line, the average temperature increases for the runs of different parameter variations are shown.

The standalone model run, analyzed more in depth, resembling the AMOC of the ESM most closely (Fig. 2.2), is the parameter variation (5,5), marked by the black box in Fig. 3.2, forced by the decrease in thermal emission with an average global warming of about 7.6 K.

Additionally, a hysteresis experiment was done for the run with the 9-BM parameter variation (5,5), with a fixed volume of Greenland ice (or, equivalently, with no reduction in meltwater influx, when the temperature melting condition was met). In 40 steps of incremental CO_2 increase, each allowed to equilibrate for 2500 years, the system was first warmed, and then in 40 steps incrementally cooled back to its initial state.

2.4 Data Analysis of AWI-ESM

In order to fit the 9-BM, data was extracted from AWI-ESM, and useful parameters were calculated. The global ocean was divided into four basins, and different salinity fluxes were calculated. These basins are the Northern Ocean (NO), defined as the Atlantic north of $45^\circ N$; the Southern Ocean (SO), defined as the Global Ocean (GO) south of $34^\circ S$; the Atlantic Ocean, defined by the Atlantic between $34^\circ S$ and $45^\circ N$; and the Indo-Pacific Ocean (IPO), essentially the difference of the Global Ocean and the other basins. In Figure 2.4, the selected ocean basins are shown on a world map. The Atlantic was divided in an Upper Part (AUPO), up to a depth of 1000 m and a Lower Part (ALPO), from a depth of 1000 m to 2500 m

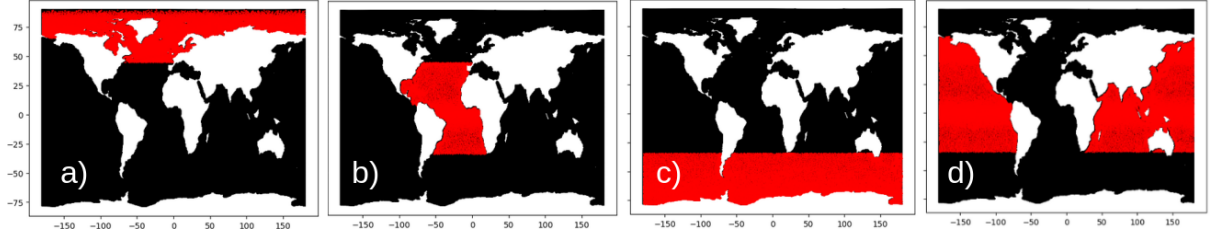


Figure 2.4: The ocean was divided into four basins. a) Northern Ocean, b) Atlantic, c) Southern Ocean, d) Indo-Pacific. At the boundaries of these basins, oceanic water fluxes were calculated. The net precipitation on the areas of the basins was calculated, too.

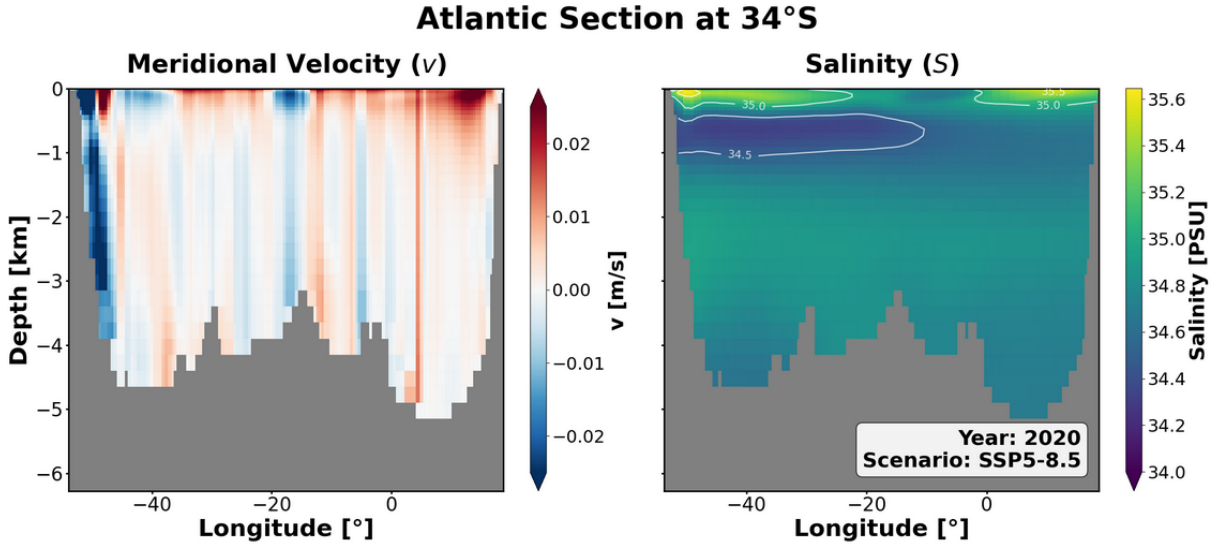


Figure 2.5: The southern edge of the Atlantic at 34°S. On the left side, the meridional velocity distribution, and on the right side, the salinity distribution by depth are plotted.

2.4.1 Definition of Relevant Parameters

For each basin, the salinity fluxes were calculated. The total salinity flux is captured in this equation:

$$F_{salt}^{tot} = \frac{1}{S_0} \int v(x, z) \cdot S(x, z) dx dz - (P - E + R) \quad (2.6)$$

The salinity flux is measured in units of Sv. The integral of the first term runs at the (zonal) boundaries of the basin, v indicates the meridional velocity, and S the salinity at the intersection of the basins. The second term represents the difference between precipitation and evaporation, added by the continental runoff from rivers. S_0 denotes a reference salinity, chosen to be 35 PSU. By employing the definition of a reference salinity, the volume change of the basin, introduced by net water import, can be neglected.

One can then define the freshwater flux (F_{fresh}^{tot}) by subtracting a reference salt flux from the actual salt flux.

$$F_{\text{fresh}}^{\text{tot}} = \left(\frac{1}{S_0} \int v \cdot S_0 \, dA - \frac{1}{S_0} \int v \cdot S \, dA \right) + (P - E + R) \quad (2.7)$$

$$F_{\text{fresh}}^{\text{tot}} = \frac{-1}{S_0} \int v \cdot (S - S_0) \, dA + (P - E + R) \quad (2.8)$$

$$F_{\text{fresh}}^{\text{tot}} = F_{\text{fresh}}^{\text{oc}} + F_{\text{fresh}}^{\text{atm}} \quad (2.9)$$

While the definition of a reference salinity is useful and will be used in this thesis, one should note that the value of the reference salinity influences non-linearly the value of the diagnostic equation, especially the freshwater flux equations (Schauer & Losch, 2019).

Oceanic Freshwater Fluxes

The oceanic freshwater flux can then be further decomposed. This is done because the different decomposed fluxes are caused by different physical mechanisms. The azonal flux component (F_{az}) captures the freshwater exchange caused by oceanic gyres at the boundaries, and the overturning flux (F_{ov}) captures the freshwater exchange caused by the overturning. The barotropic component captures the freshwater transport caused by a net motion of the water at the boundary (F_{bt}), and the eddy flux, which is caused by the transport of eddies (F_{ed}). This decomposition is done as in Jüling et al. (2021).

$$F_{\text{az}} = -\frac{1}{S_0} \int_{z_1}^{z_0} \int_{x_w}^{x_e} S' \cdot v' \, dx \, dz \quad (2.10)$$

$$F_{\text{ov}} = -\frac{1}{S_0} \int_{z_1}^{z_0} \left[\int_{x_w}^{x_e} (v - \hat{v}) \, dx \right] [\bar{S} - S_0] \, dz \quad (2.11)$$

$$F_{\text{bt}} = -\hat{v} \frac{\hat{S} - S_0}{S_0} \quad (2.12)$$

$$F_{\text{ed}} = F_{\text{fresh}}^{\text{oc}} - F_{\text{bt}} - F_{\text{ov}} - F_{\text{az}} \quad (2.13)$$

The hat symbol ($\hat{\cdot}$) indicates the average value of the whole ocean section. The prime symbol ($\cdot'$) indicates the deviation of the value, relative to the zonal average of the depth layer. This zonal average of the depth layer is indicated by the bar symbol ($\bar{\cdot}$). The eddy flux is calculated here as the residual of the other fluxes, since the dataset used in this thesis has an annual time resolution, which is not sufficient for calculating it directly. The basis for these calculations is the distributions of velocity and salinity at the boundary of the basin. An example section at 34°S at the southern boundary of the Atlantic for the year 2020 is shown in Figure 2.5.

Atmospheric Freshwater Fluxes

The atmospheric freshwater fluxes are divided into precipitation, evaporation, and continental runoff. By defining the oceanic shapefiles as in Fig. 2.4, the net precipitation can be extracted directly from the oceanic module FESOM2 (Danilov et al., 2017) of AWI-ESM, by using the FESOM plotting tool tripyview (Scholz, 2025).

Overturning Stream Function

The stream function calculates the meridional transport integrated from the surface down to a given layer depth. It is defined here as follows:

$$\Phi_y(z, t) = \int_z^0 \int_{x_w}^{x_e} v(x', y, z', t) dx' dz' \quad (2.14)$$

The maximum of the stream function of the Atlantic at 27°N is then defined to be the AMOC strength.

$$\text{AMOC}(t) = \max_{z \in [-H, 0]} (\Phi_{27^\circ\text{N}}(z, t)) \quad (2.15)$$

Oceanic Heat Transport

The meridional oceanic heat transport (F_{heat}) at a given section is defined analogously to the oceanic salt transport at a single boundary (Eq.2.6). The density (ρ) and the heat capacity (c_p) are treated as constant.

$$F_{\text{heat}} = c_p \cdot \rho \int T \cdot v dA \quad (2.16)$$

2.4.2 ESM Data

Two different warming scenarios using AWI-ESM run by Ackermann et al. (2026) were analyzed. Each warming scenario was run twice – once with an interactive ice sheet module activated, and once deactivated. The warming scenarios analyzed are the shared socioeconomic pathways SSP2 and SSP5, with a radiative forcing of 4.5 W m^{-2} and 8.5 W m^{-2} (Lee et al., 2023). The model output analyzed spans the years from 1850 to 3000. For each year, average annual values of different quantities, such as salinity, velocities, and temperature, were saved. The underlying ocean module FESOM2's (Danilov et al., 2017) grid is formed in a prism structure, which makes diagnostic calculations complicated. In order to rotate the velocity vectors in the meridional direction, as well as for other operations on the data, the plotting tool tripyview was used (Scholz, 2025). The majority of the programming in this thesis was done in Python (Python Software Foundation, 2024).

Chapter 3

Results

In this chapter, the results of the Earth System Model and box model runs will be presented. First, the 9-BM runs will be shown, both for the model most closely fitted to the output data of the AWI-ESM in the warming scenario SSP5-8.5 with the interactive ice sheet of Greenland being used (SSP5-8.5_{ice}). Additionally, stability maps were created for 9-BM runs with varied parametrization and warming.

For four different AWI-ESM runs, relevant diagnostics and parameters for the strength and stability of AMOC, such as the freshwater fluxes, were calculated. The model runs include the warming scenarios SSP2-4.5 and SSP5-8.5. For each warming scenario, the runs were done once with an interactive Greenland Ice Sheet (GIS) and once with a prescribed one.

In the last section of this chapter, the freshwater fluxes acting on the defined ocean basins and their corresponding oceanic boxes will be compared.

3.1 Results of the 9-BM

In this section, the results of different 9-BM runs are presented. In the first section, the results of the model that best fits the calculated parameters of the ESM are shown, which undergoes a warming of about 8 K. In the next section, the interaction of a weakened interaction of the AMOC cells with the Indo-Pacific is simulated, in the edge case of the Agulhas leakage set to zero, no mixing of the Southern with the IPB, no Bering Strait transport, and no atmospheric freshwater transport into the IPO, this model is conceptually resembling the Rahmstorf model, except for the interactive atmospheric transport still being intact.

The values of the 14 state variables (blue) and of the AMOC strength (black) are shown in Figure 3.1. The system starts at an unstable state, and is allowed to reach an equilibrium state for the first 3425 years. Then a warming is applied for 150 years, by decreasing the thermal loss of the atmospheric boxes to space, by tuning the CO₂ value.

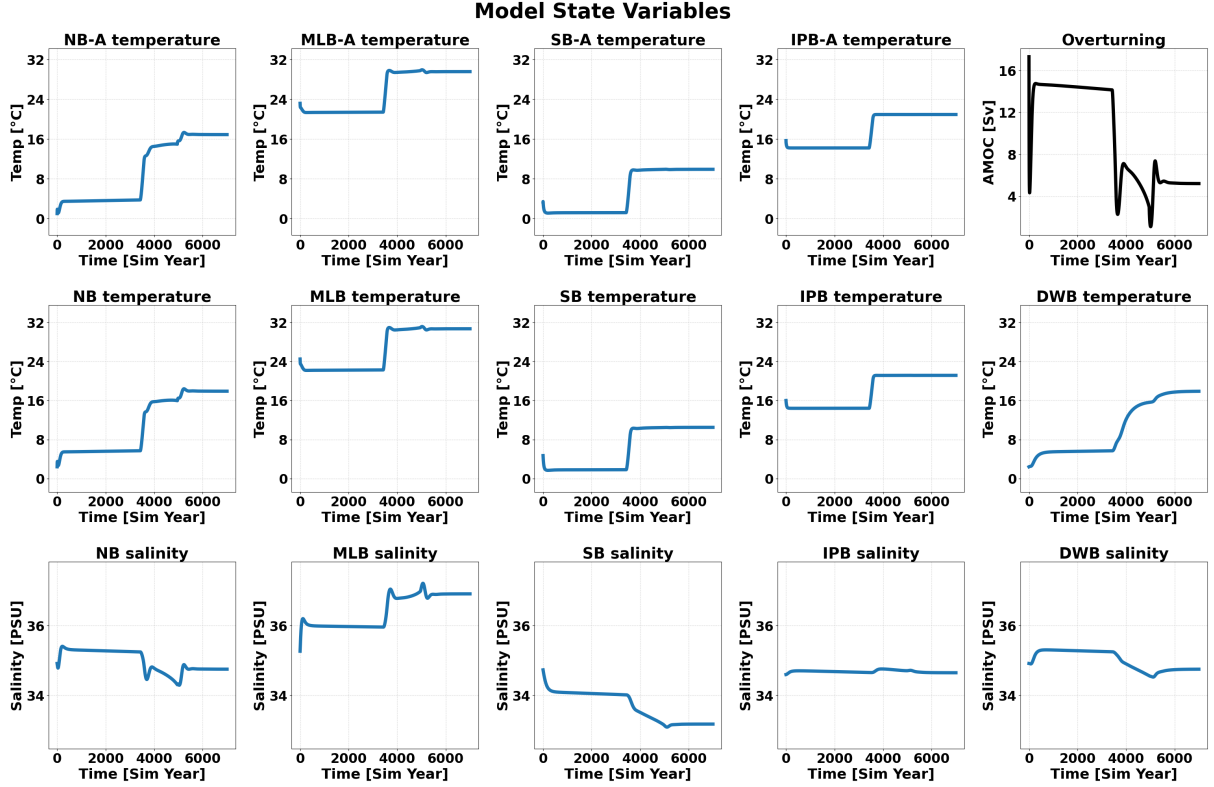


Figure 3.1: The results of a run of the BM are shown. The model was given 3425 simulation years to reach equilibrium. Starting in 3425, an introduction of CO_2 into the atmosphere was simulated, stopping in the simulation year 3575. After that, the simulation ran for another 3425 years to reach equilibrium again. In the top row, the atmospheric temperatures are shown, as well as the overturning strength (black). In the second row, the temperatures, and in the third row, the salinities of the oceanic boxes are plotted.

For the remaining 3425 years, the CO_2 level is constant.

The temperature of the atmospheric boxes in the first equilibrium state is 4°C for the northern atmospheric box (NB-A), 22°C for the mid-latitudinal atmospheric box (MLB-A), 2°C for the southern atmospheric box (SB-A), and 14.5°C for the Indo-Pacific atmospheric box (IPB-A).

In the case of the oceanic boxes, the temperatures are 5°C for the NB, 22°C for the MLB, 2°C for the SB, 15°C for the IPB, and 5°C for the DWB. The salinities reach an equilibrium at 35.25 PSU (NB), 36 PSU (MLB), 34 PSU (SB), 34.75 PSU (IPB) and 35.25 PSU (DWB).

As the warming starts in the simulation year 3425, the temperatures of all oceanic and atmospheric boxes increase abruptly, except for the Deep Water Box, where the warming effect takes more time. The salinity of the NB decreases, and the salinity of the MLB increases abruptly.

In the time after the forced warming, the temperatures of the boxes reach fairly stable values after the simulation year 5200. The temperatures of all boxes are higher than before

the warming event. The atmospheric boxes reach temperatures of 17°C (NB-A), 30°C (MLB-A), 10°C (SB-A), and 21°C (IPB-A) until the simulation year 5200. During the same time, the oceanic temperatures increase to 16°C (NB), 31°C (MLB), 10°C (SB), 21°C (IPB), and 16°C (DWB). After 5200, the DWB increases in temperature to 18°C.

In the year 5200, which has a lower net salinity due to melting, the salinities of the northern, southern, and the Deep Water Box dropped to 34.25 PSU, 33.5 PSU, and 34.25 PSU, respectively. The MLB increased in salinity to about 36.5 PSU, and the salinity of the IPB stayed roughly constant.

In between the warming start and the year 5200, AMOC drops in strength for about 200 years to about 3 Sv, then recovers for another 200 years to about 7 Sv. AMOC loses strength, and the strength loss is accelerating until 5200, reaching a minimum strength of 2 Sv, when it suddenly recovers. After 5200, it reaches a stable equilibrium at about 6 Sv.

In addition to this run, further parametrization of the 9-BM and different warming scenarios were tested to assess the stability of the AMOC in this model. The model configuration corresponding to the run discussed above is denoted by the parametrization (5,5), which is highlighted with a black box in Figure 3.2. The forth column uses the same warming scenario as the run presented above.

3.1.1 Results of Warming for Different Variants of the 9-BM

In Figure 3.2, a stability map for AMOC is shown, for different parameterizations of the 9-BM and for different warming scenarios. Each column represents a different warming scenario, with the same CO_2 increase. The axes of the plots in the first two rows represent the strength of the coupling of the AMOC boxes to the Indo-Pacific oceanic and atmospheric boxes. The index for each axis runs from zero to six, scaling the strength of the coupling linearly; the index pair (0,0) represents the run, where the IPB and IPB-A do not influence the AMOC and its coupled atmospheric boxes. The CO_2 increase started after 1925 years, and increased until 2075 in simulation time. For the next 1925 simulation years, the CO_2 level stayed constant.

The first row of Figure 3.2 shows the minimum value of the overturning circulation after warming is applied. A square is colored gray when the circulation collapses ($\Phi < 0$). The second row shows the value of the final state of the circulation. If the square corresponding to a parametrization of the model is red, then the simulation remained, or returned to a positive AMOC state. If the color is blue, then the circulation changed into a negative AMOC state. In the line plots of the third row of Figure 3.2, the trend of the overturning strength is shown, referencing the model run indicated by the cyan squares at index (3,3) in the first two rows.

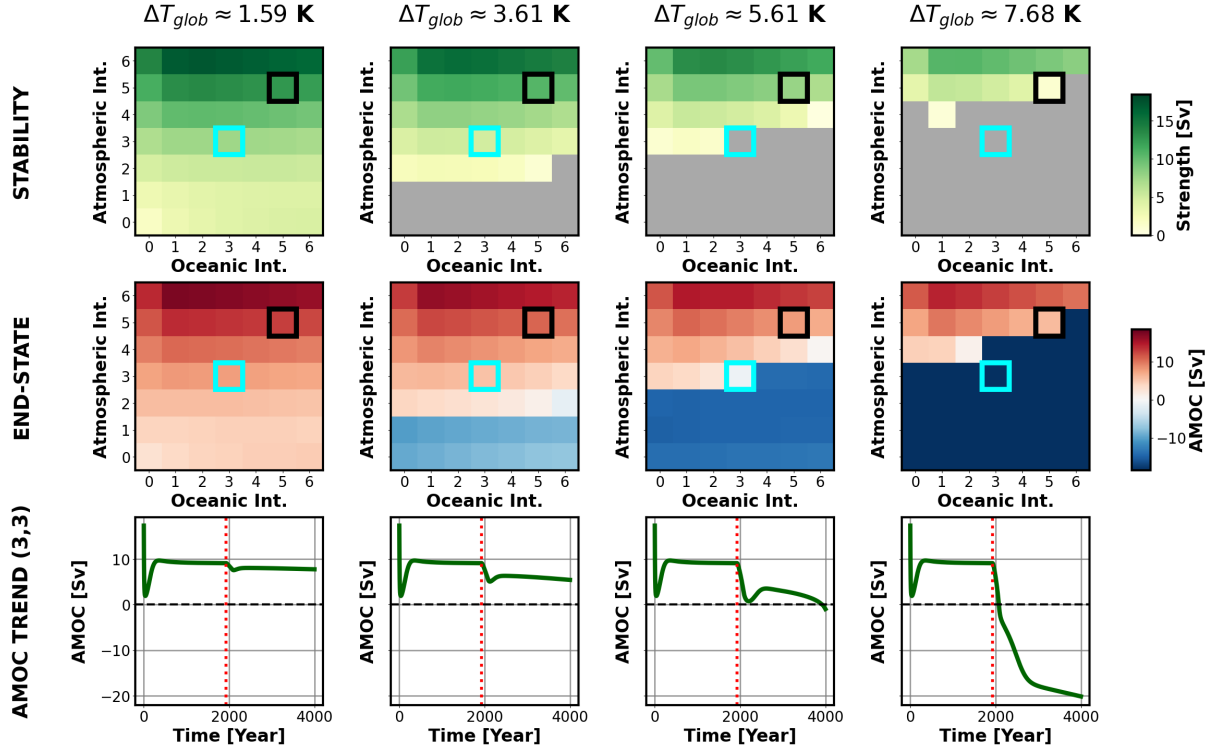


Figure 3.2: Global warming experiments were done for different parameterizations of the 9-BM. The oceanic and atmospheric interactions of the AMOC with the IPB were varied. Lower indices indicate weaker coupling. The columns are divided into four regimes of global warming. The top row shows in color the minimum Overturning Strength after warming was applied; if the circulation strength dropped below zero, the square was colored gray. The black box at index (5, 5) uses the model parametrization shown in Section 3.1. The cyan box at index (3, 3) represents the graphs in the third row. The second row shows the colors representing the state at the end of the simulation time. Blue indicates a reversal of circulation, while red indicates positive circulation. The darkness of the squares indicates the strength of the circulation.

With rising temperatures, the system gets more volatile. While for a warming of about 1.59 K all simulations remain in a positive state, 15 out of 49 collapsed for a warming of about 3.61 K. A further increase of temperature to 5.61 K leads to an AMOC-collapse in additional ten runs. The simulation of a warming of 7.68 K leads to a collapse of ten further positive AMOC-states, totaling at 35 collapsed AMOC states out of 49 runs.

The positive AMOC states that collapsed, switched into a stable negative AMOC state, with two exceptions: at the warming of 7.68 K at index (0,4) and (2,4), where the AMOC strength dropped below zero, and then recovered into a positive state.

The graphs representing the runs with the model parameter of index (3,3) show different behavior for an increase in temperature. Before the warming, the overturning equilibrates at about 10 Sv. For a warming of 1.59 K, the overturning gets slightly weaker after the warming starts, then equilibrates at a strength of about 9 Sv. In the case of a warming of 3.61 K the drop of AMOC strength is more pronounced. After the decrease

in AMOC strength, it stays roughly at 8 Sv, while decreasing slightly until the end of the run. In the case of a warming of 5.61 K, the overturning drops to about 0 Sv after the warming is applied, to then recover to about 3 Sv in 2200, then drops again and finishes at just below 0 Sv at the end of the run. The run for the model parameter (3,3) in the case of a warming of 7.68 K shows a sudden decrease in AMOC strength after the warming was applied, to then switch into a negative phase, where AMOC is reversed with a higher strength than in the positive case at a circulation strength of -20 Sv.

3.1.2 Hysteresis of the 9-BM

The 9-BM with the same parameter configuration as in Figure 2.2, with the modification that the Greenland ice sheet volume was fixed, was subjected to an incremental increase in CO₂ forcing. After each incremental increase, the system was allowed for 2500 years to reach a new equilibrium under the increased forcing.

In Figure 3.3, the AMOC strength of the system is plotted against the global temperature. The violet color indicates the starting state of the hysteresis experiment, and it changed towards yellow, as the experiment progressed. First, the system was warmed, starting at 13°C, to about 23°C, and then was cooled back to 13°C.

After an increase of about 9 K the positive AMOC state switches abruptly into a negative state, after previously weakening slowly under the warming to about 5 Sv. Further warming does not change its state qualitatively, with a slight decrease in AMOC strength to about -22 Sv. After the CO₂ forcing was reduced to below 22°C, the system remained in the negative state, switching back into the positive state after it was cooled to below 17°C at an AMOC strength of about -10 Sv. The further cooling to 13°C leads to a fairly close match of the AMOC strength of the initial state.

Between the temperatures of 17.5°C and 22°C, two distinct AMOC states exist.

The 9-BM was manually fitted to the output data of AWI-ESM, in order to recreate the box model as close to reality as possible, and to compare the reactions of the AMOC system to an increase in global warming. The analyzed data will be the content of the next section.

3.2 Results of AWI-ESM

In this section, the runs of AWI-ESM will be analyzed for the warming scenarios SSP2-4.5 and SSP5-8.5. The Streamfunctions were plotted, along with the trend in AMOC strength. Different freshwater fluxes and their effects on the specified ocean regions were calculated, as well as their salinity content. The ocean sections were divided into three

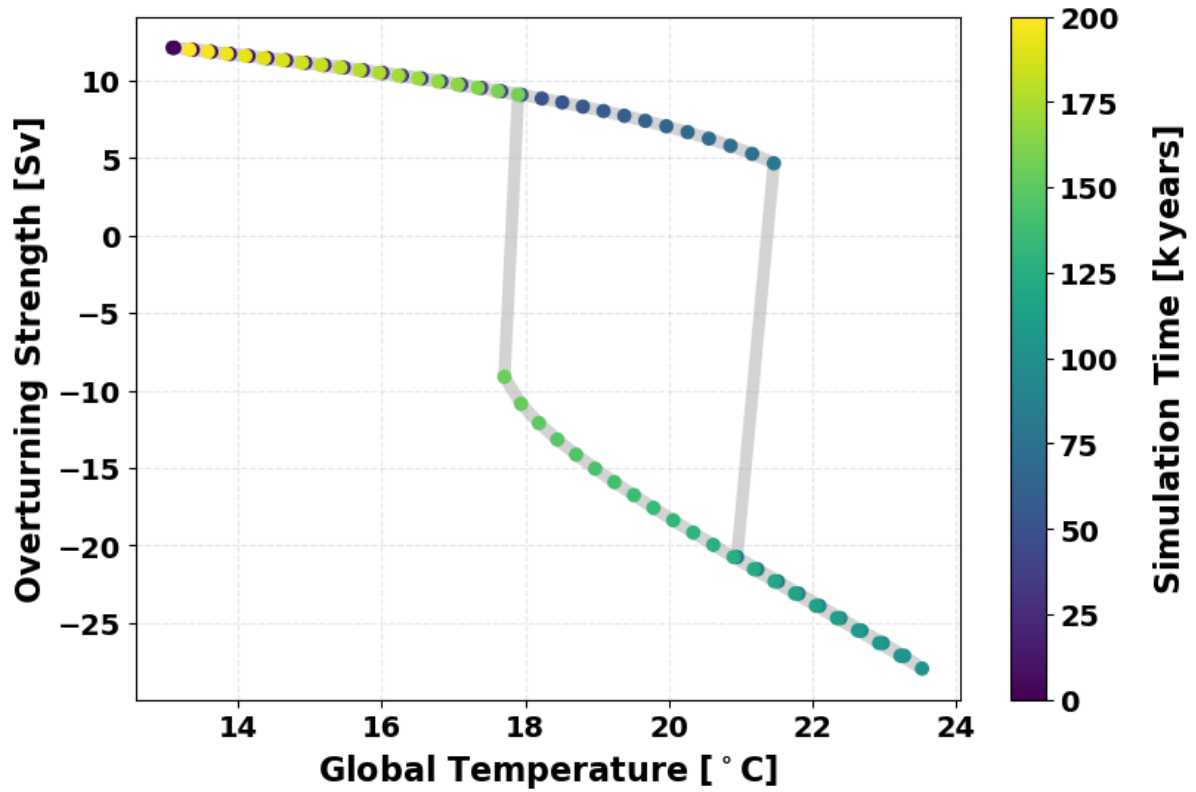


Figure 3.3: The model was chosen with the parameter configuration of Fig. 2.2 with the index configuration (5,5). The CO₂ forcing was stepwise increased, then the system was allowed to reach equilibrium for 2500 years, and after that, decreased again, until the CO₂ forcing was the same as at the start of the simulation. The Hysteresis shows the path dependence of the system. With the same forcing, different AMOC states are possible.

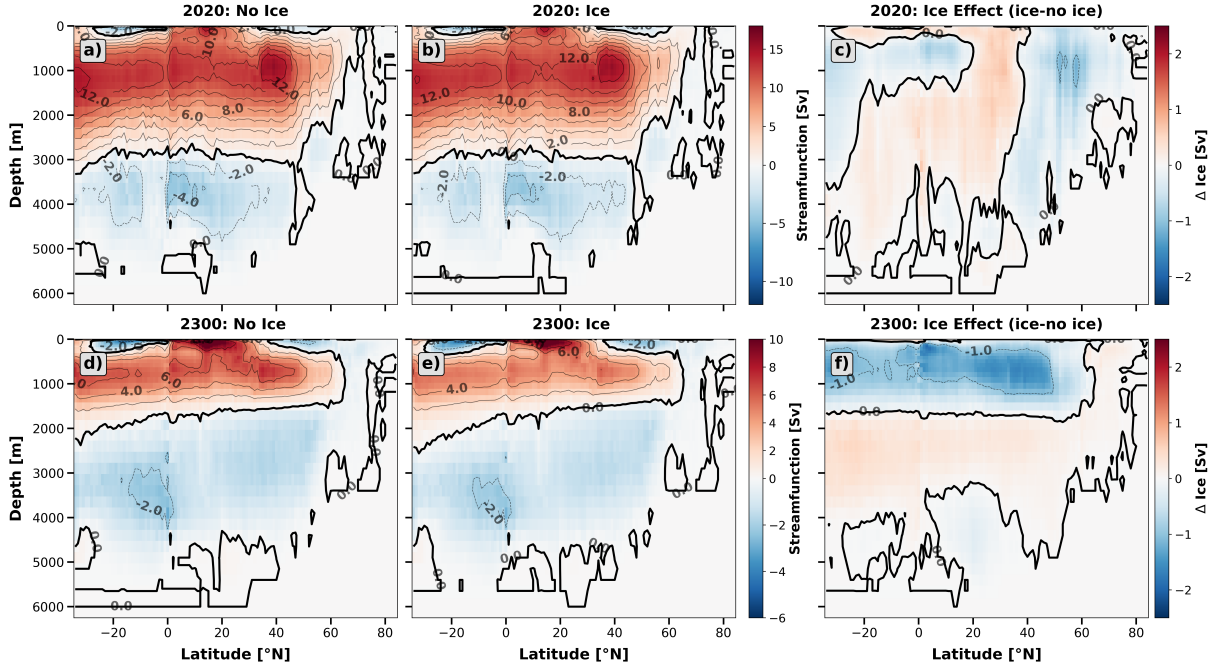


Figure 3.4: The value of the overturning stream function was calculated for every point in the Atlantic. Red indicates a clockwise overturning, if viewed from the east of the Atlantic. In the top row the overturning plot for the year 2020 in the SSP5-8.5 scenario a), and the SSP5-8.5_{ice} b) is shown. The plot in c) shows the difference in overturning strength, between both model versions. In the second row the overturning strength is shown for the year 2300 for both SSP5-8.5 d) and SSP5-8.5_{ice} e). In f) the difference for both model versions is plotted.

basins: the upper part, from 0 - 1000 m, from 1000 - 2500 m, and the whole ocean basin from top to bottom. More focus was given to the freshwater fluxes crossing the southern boundary of the Atlantic at 34°S, since the stability parameter F_{ov} is often considered relevant for the stability of AMOC (Weijer et al., 2019).

3.2.1 Atlantic Meridional Overturning Circulation

The streamfunction of AMOC, calculated from 34°S to 80°N is shown in Figure 3.4 for the warming scenario runs SSP5-8.5 and both model versions, once with GIS active and once without. In the right column the difference effect of the inclusion of GIS is shown, for the years 2020 and 2300.

The red area indicates a clockwise rotation, if the observer is on the eastern side of the Atlantic, the blue area a counter-clockwise rotation. The overturning plot in the year 2020 shows a depth of the AMOC system of about 3000 m, spanning from 34°S to about 65°N, with a maximum overturning strength of about 15 Sv. The difference of the streamfunction of the different model version in this year is negligible.

The weaker AMOC in the year 2300 has a depth of about 2000 m, spanning the Atlantic from 34°S to 65°N, with a maximum strength of about 8 Sv in both model

versions. The difference plot between the version of the model using GIS and the one without in the year 2300 shows a weaker AMOC in the GIS version, with a difference of 1 Sv to 2 Sv.

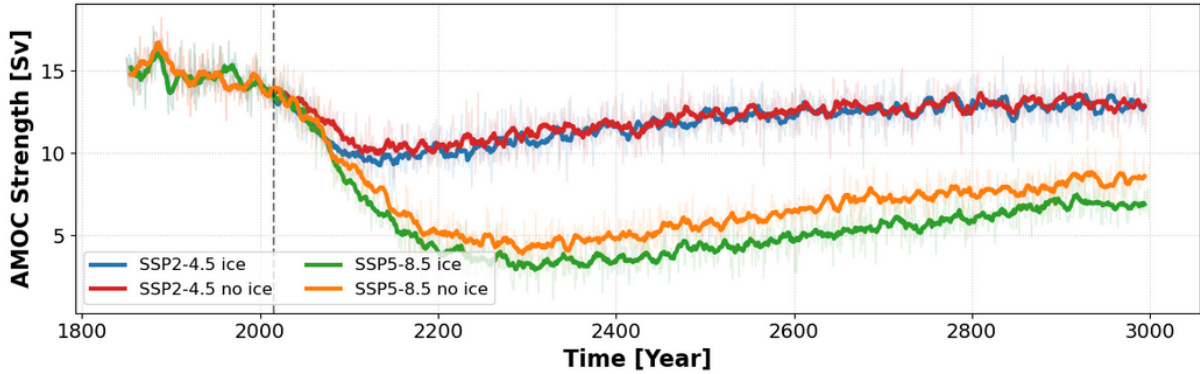


Figure 3.5: The graphs shows the trend of the maximum overturning strength at 27°N for the warming scenarios SSP2-4.5 and SSP5-8.5, each scenario once with and once without GIS being activated. The values of the graph show an averaged value, averaging over the five years before and five years ahead. the yearly averages are shown in faint color.

The strength of the overturning was calculated at 27°N, using Eq. 2.15. The graphs of the two SSP scenarios, for each once with GIS being active and once with a prescribed Greenland ice sheet, can be seen in Figure 3.5.

The four runs follow a close path for the first 200 years and begin to diverge around 2050. Beginning at about 15 Sv at the start of the simulation in 1850, the overturning weakens to 13.5 Sv in 2050. The overturning strengths of both runs of the scenario SSP2-4.5 follow each other closely, experiencing a minimum at about 10 Sv in 2150, and regaining strength slowly until the end of the simulation in 3000, at a value of 12.5 Sv.

After diverging from the other runs in 2050, the overturning strengths of the runs simulating SSP5-8.5 experience minima in the year 2300. The run with the interactive Greenland ice sheet activated has a minimum of the overturning at about 3 Sv, and the other run a minimum of about 4 Sv. Both runs show a growing overturning strength after that, with the run without an interactive Greenland ice sheet terminating at about 8.5 Sv and the run with the ice sheet module activated at about 7 Sv.

3.2.2 Salinities

The salinity of the different regions were analyzed, in order to understand the mechanisms of the freshwater transport better, since each oceanic transport is dependent on the salinity of its originating region. The depth dependence is of importance, since gyre mixing takes place in the upper part of the ocean, whereas the overturning transport takes place in both the upper and lower part of the section.

The ocean basins defined earlier were separated into three different depth regimes. The first one starts at the surface until 1000 m, the second one reaches from 1000 m to 2500 m and the last one contains the whole basin. The upper two parts resemble, in the case of the Atlantic, the cells of the AMOC. The warming scenarios SSP2-4.5 and SSP5-8.5 are shown in Figure 3.6.

First the focus will be laid on the scenario SSP5-8.5_{ice}, and then it will be compared to the other runs. Until the year 2000, the salinity in the upper part of the ocean remains fairly constant. The Atlantic's upper part is the saltiest with a salinity of 34.40 PSU. The upper part of the Northern and of the Global Ocean show the similar salinities for the first 150 years, at about 34.75 PSU. The salinity of the Indo-Pacific's upper part is lower, at about 34.55 PSU – the Southern Ocean has the freshest upper part at about 34.45 PSU.

Starting in the year 2050, the Atlantic's salinity in the upper part increases fast until 2150 to 35.6 PSU, and then more slowly for the time until 2700, when it reaches its peak at 35.7 PSU, then it declines until the end of the simulation run, ending at about 35.6 PSU. The Northern and the Global ocean's salinity start to diverge after the year 2000. While the salinity in the upper part of the Global Ocean stays roughly the same, the salinity of the upper NO drops for 100 years, reaching a salinity of 34.1 PSU. Then it slowly increases again until the end of the simulation run in the year 3000, equilibrating at 34.50 PSU. The IPO's upper part freshens slowly over time, reaching 35.5 PSU in the year 3000. The SOs upper part freshens until it reaches a salinity of 34.20 PSU in the year 3000.

In the deeper part of the ocean, ranging from 1000 m - 2500 m, the salinities for the first 150 years of the simulation, until the year 2000, stay roughly the same. The NO's deeper part is the saltiest at 35.05 PSU, slightly fresher is the Atlantic's deeper part at 35.00 PSU. The Global Ocean's deeper part has a salinity of 35.50 PSU, the SO a salinity of just above 34.60 PSU and the IPO's salinity is just below 34.60 PSU for the first 150 years.

The deeper part of the NO then loses salinity until 2550 at 34.80 PSU, and then gains salinity until the end of the run, reaching a salinity of 34.85 PSU. The Atlantic Oceans lower part (ALPO) loses salinity until about 2600 when it reaches about 34.95 PSU, to then return to a salinity of slightly above 35.00 PSU in the year 3000. The salinities at this depth range decrease for the GO, the SO and the IPO steadily after the year 2200, until the end of the run. the Global Ocean reaches a value of about 34.60 PSU, the Southern Ocean at 34.50 PSU and the Indo-Pacific at 34.40 PSU.

The full depth salinity of the basins in the SSP5-8.5_{ice} remains constant for the first 150 years in all ocean basins. The Atlantic's salinity is about 35.1 PSU, the NO just below 35.00 PSU. The GO's salinity is at 34.70 PSU, and the IPO and the SO salinity at 34.60 PSU.

After the year 2000 the salinity of the Atlantic increases until 2150 to a value of just below 35.2 PSU, and then remains roughly the same. The NO loses salinity until the year 2400, when it reaches a salinity of 34.60 PSU. Then its salinity increases until the end of the run, finishing at 34.70 PSU. The GO's, the IPO's and the SO's salinities drop continuously until the end of the run. The GO reaching a final salinity of 34.60 PSU and the IPO and SO a salinity of just above 34.40 PSU.

The SSP5-8.5 run with the prescribed Greenland ice sheet (SSP5-8.5_{no ice}) diverges from the SSP5-8.5_{ice} in the upper part in the year 2150. After 2150 the salinity of the salinity decrease of the NO is less pronounced, with a minimum of 34.4 PSU in 2350, when it recovers later in 2900 to its initial state at about 34.75 PSU. After the salinity of the AUPO diverges in 2150, it increases compared to the GIS run. It reaches a salinity of 36.0 PSU in 2600, and drops again to reach the same level as the other run again in 3000, at about 35.6 PSU.

The salinities of the lower part of the ocean diverge after the year 2150 too for the NO and ALPO. The NO is about gets about 0.3 PSU more saline then in the GIS run, and the ALPO about 0.2 PSU more saline. The SO is at the end of the run about 0.1 PSU more saline than the other run.

Over the whole ocean depth, the NO is notably saltier than in the GIS run, after diverging in 2150 it reaches a value 0.4 PSU higher. The Atlantic is slightly less saline, with a salinity difference of -0.1 PSU.

The deviations between the two SSP2-4.5 runs is rather small. In the upper 1000 m of the ocean, the runs of the SSP2-4.5 differ mainly in three ways, when compared to the SSP5-8.5 scenario runs. 1) The salinity decrease of the NO is shortly lived and far less pronounced. The salinity drops in the year 2000 by about 0.1 PSU and recovers in about 2250 to its former level. 2) Compared to the SSP5-8.5 runs the salinity increase in the AUPO is weaker. Until the year 2100 the SSP2-4.5 and SSP5-8.5 runs match each other closely, but the high emission scenario runs converge to a higher value of 35.75 PSU, while the lower emission scenario converges to a value of 35.5 PSU in the year 3000. 3) While the salinity of the upper SO stays roughly constant in the SSP2-4.5 runs, it decreases after 2200 steadily in the warmer scenario runs.

In the part of the ocean ranging from 1000 m - 2500 m, the behavior of the salinities from the SSP2-4.5 scenario in the Atlantic and the NO differ quite clearly from the SSP5-8.5 runs. While the salinities begin to drop in the year 2100 for the latter, they increase continuously in the lower warming scenarios. The freshening in the IPO and the SO is less pronounced. The Global Ocean's salinity stays roughly the same.

Compared over the whole depth of the basin, the salinities for the NO differ quite clearly for the lower warming runs, compared to the runs of the SSP5-8.5 scenario. The salinity of the NO drops in both runs of the scenario SSP2-4.5, but starts to rise again at about 2200, when it continues to fall in the warmer scenario runs. The minimum of the

warmer runs is about 34.90 PSU, when for the warming run SSP5-8.5_{ice}, the minimum is at 34.60 PSU. The minimum, when in the warmer scenario the ice sheet module is not activated, is at about 37.70 PSU.

The salinity changes are driven by water transport of different salinities. These salinity transports are decomposed in oceanic and atmospheric transports. In the following sections, the transports, expressed in terms of freshwater fluxes, are presented.

3.2.3 Oceanic Freshwater Fluxes

Freshwater Fluxes at 34°S

The oceanic freshwater transport is divided in four parts: the azonal (or gyre) component, the overturning component, the barotropic and the eddy component (see Eqs. 2.12 - 2.13). These freshwater fluxes were computed for the different runs, and for different depth ranges (see Fig. 3.7), at the southern boundary at 34°S of the Atlantic ocean. Again, first the focus will be laid on the run simulating the scenario SSP5-8.5 with the interactive Greenland ice sheet module activated, and then it will be compared to the other runs. The barotropic component is about ten orders of magnitude smaller than the other computed fluxes, and is therefore neglected.

In the upper 1000 m of the ocean and in the first 150 years the oceanic freshwater flux is dominated by the overturning component. The freshwater import into the Atlantic basin is at about 0.1 Sv for both this component (F_{ov}) and the component sum (F_{fresh}). After the year 2000, the azonal component increases from around 0 Sv to about 0.4 Sv in 2300, and is, starting from the time of its increase, the dominant component of F_{fresh} . Then F_{az} decreases over the next 700 years to about 0.3 Sv, while still being the major contributor of freshwater transport. The overturning component decreases, and becomes negative at 2050, reaching a minimum at 2125 at about -0.75 Sv. Then it increases until 2300, when it is – and continues to stay – at about 0 Sv. The eddy component stays at about 0 Sv.

At the lower depth range of 1000 - 2500 m the overall freshwater transport is smaller, but more complex. The azonal component increases over time, from about 0.05 Sv to, at the beginning, to 0.5 Sv. The overturning component is negative at about -0.05 Sv in the beginning, slightly increasing to about -0.025 Sv in 2350, then dropping until reaching a value of about -0.1 Sv in 3000. The eddy component remains small, compared to the other component fluxes. Overall, the freshwater flux into the Atlantic is at this depth negative. Starting at about -0.025 Sv, it reaches about 0.0 Sv in 2300, and then drops again, terminating at about -0.035 Sv.

While at the depth range of the upper 1000 m of the ocean, the overturning component converges to zero in the year 2300, it stays negative, when integrated over the whole

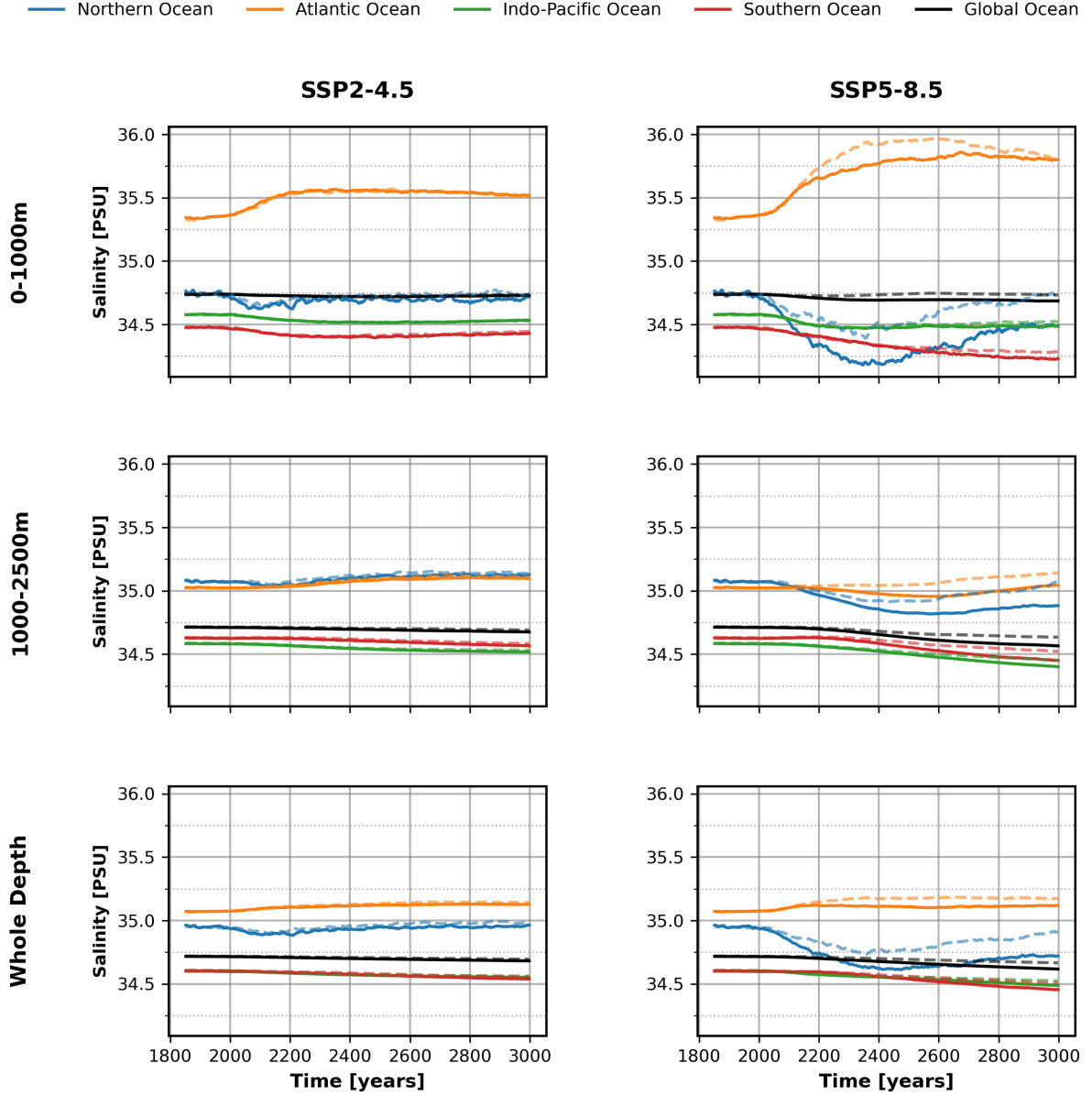


Figure 3.6: The salinities of the ocean basins is plotted for the upper part (0-1000 m) and the lower part (1000-2500 m) of the AMOC cell, and once for the whole depth. The left column shows the salinities of the SSP2-4.5 and the right column of the SSP5-8.5 runs. The solid line indicates the scenarios with GIS active, and the dashed lines with the prescribed Greenland ice sheet. The data was averaged for 15 years

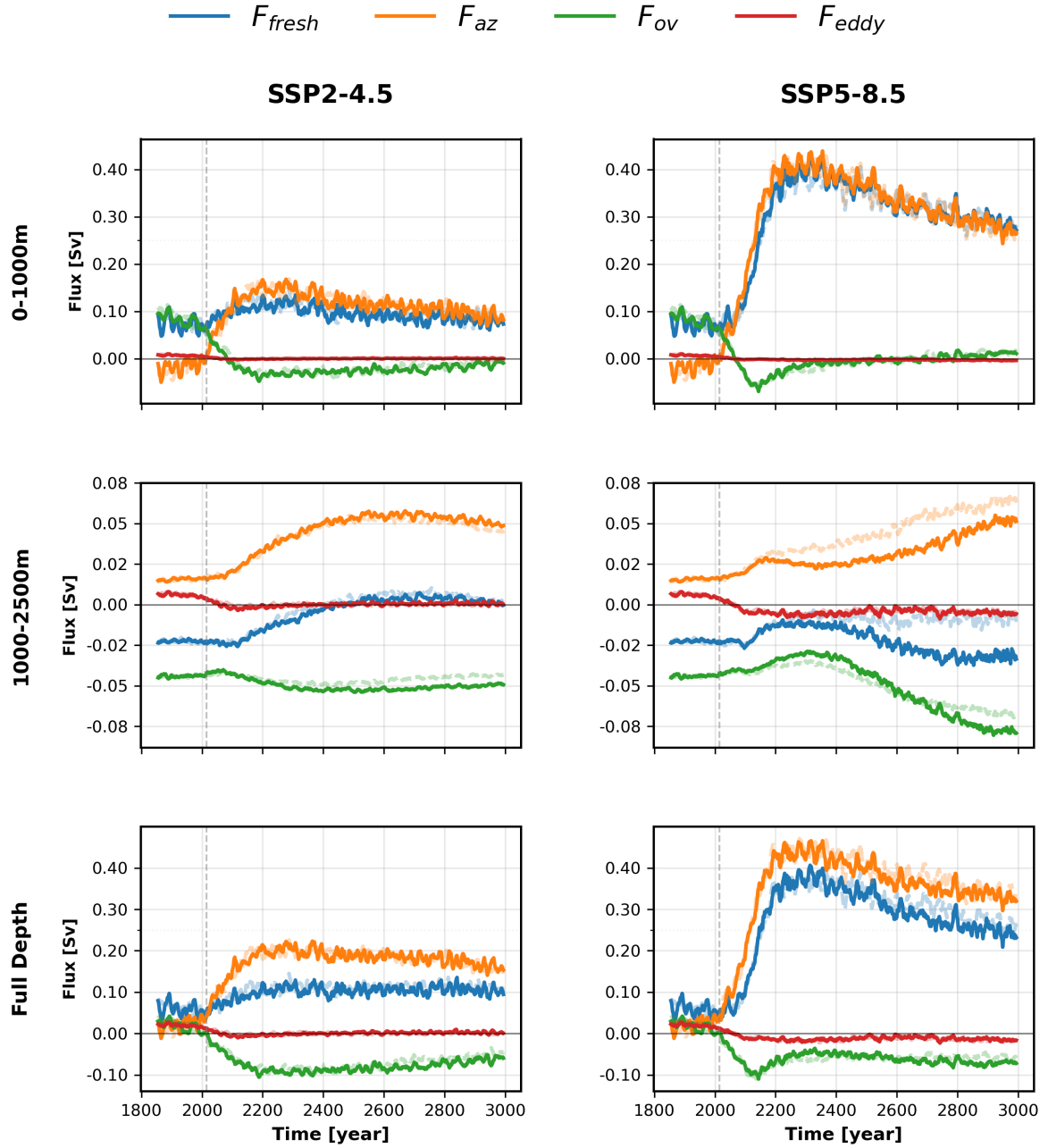


Figure 3.7: The freshwater flux into the Atlantic at the southern boundary (34°S), and its decomposition into a gyre, an overturning and an eddy component are plotted (see Eqs. 2.12 - 2.13) for the different runs and different depth ranges. The solid lines indicate the runs with the GIS, the see-through lines with the prescribed Greenland ice sheet. The data was averaged for eleven years.

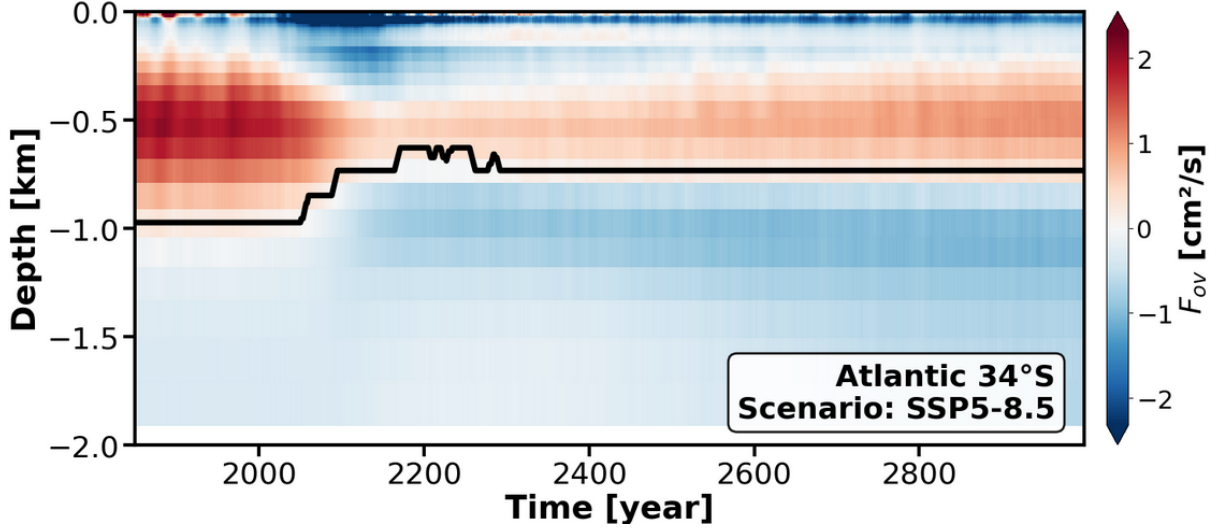


Figure 3.8: Shown is the overturning freshwater flux F_{ov} , before being integrated by depth, as was done in Eq. 2.11. The red areas indicate a positive overturning freshwater flux, and the blue areas indicate a negative one. The depth, where the integral of F_{ov} is biggest for a given year, is indicated by the black line.

depth range. The azonal flux component still dominates the freshwater transport, but the freshwater transport is consistently below the azonal flux.

In the upper box the SSP5-8.5 run is similar to the other runs, the main difference being, that the increase in the azonal, and therefore the increase of the total freshwater flux is less pronounced, barely reaching 0.20 Sv in the SSP2-4.5 runs. In the colder scenarios, the overturning component stays negative after the year 2300, slowly converging to 0.0 Sv until 3000. There is no notable difference between the interactive Greenland ice sheet run, and without it.

Comparing the fluxes of the SSP5-8.5 run at the depth range 1000-2500 m to the other runs, a notable difference is, that after behaving the same as the reference run until 2300, the other runs diverge from it, and the overturning and azonal component cancel each other out, yielding a net zero freshwater transport.

The fluxes integrated over the whole depth differ for the colder scenario, compared to the SSP5-8.5 scenario. The influence of the ice sheet module is small. The total freshwater flux stays at roughly 0.10 Sv for the whole simulation, but with varying contributions of the azonal and the overturning flux.

F_{ov} -density over time

The overturning freshwater flux changes over time. In Figure 3.8 the F_{ov} -density change over depth and time is shown. At the start of the SSP5-8.5 run, using the interactive Greenland ice sheet at the southern boundary of the Atlantic at 34°S, the upper 200 m transport saltwater from the SO to the Atlantic. At 200 m up to 1000 m freshwater is transported. Below 1000 m the transport is negative, transporting saltwater into the

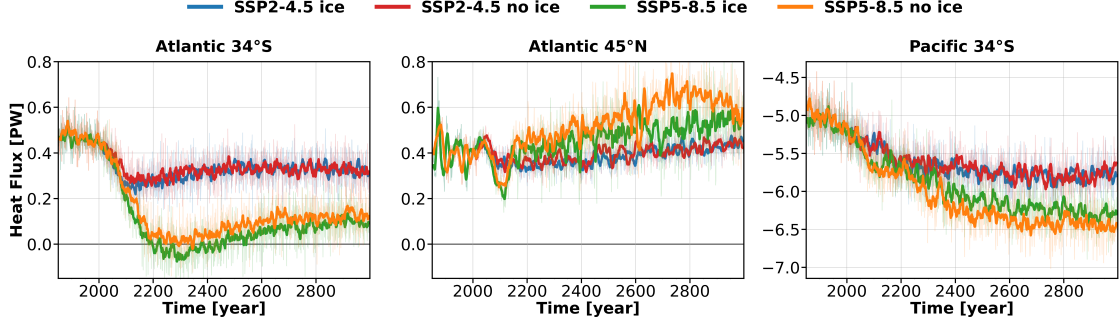


Figure 3.9: The northward heat flux trend is shown for the defined Atlantic boundaries at 34°S and 45°N, as well as for the IPO boundary with the SO at 34°S, for each of the four runs. The data was averaged over 15 years.

Atlantic. The range of the freshwater transport declines over time, at both the upper and lower edge of the freshwater transport. In the year 2200 the cell transporting freshwater is the smallest, at about 400 m to 700 m of depth. The cell then continues to grow again, reaching at the end of the run a range from about 250 m to 800 m. While the range recovers over time, the amplitude of the positive part of the F_{ov} -density at the end of the run is lower.

3.2.4 Oceanic Heat Fluxes

The trends of the northward oceanic heat flux at the boundaries of the Atlantic basin and the boundary of the SO with the IPO is captured for the four runs in Figure 3.9.

In the SSP5-8.5_{ice} run the heat flux at the Atlantic’s southern boundary starts out at about 0.5 PW, and continues to fall to about -0.05 PW in 2300. After that it continues to grow in strength continuously, until it reaches a strength of about 0.1 PW at the end of the run. The SSP5-8.5_{no ice} run is slightly less affected by the drop of strength. After 2200 it has a higher strength of about 0.15 PW, relative to the SSP5-8.5_{no ice} scenario. The scenario runs for SSP2-4.5 start out at the same strength of 0.5 PW, but continue to drop less strong. These runs drop to about 0.3 PW at 2175, and slightly increase to about 0.35 PW in 2500, and then stay roughly constant for the remaining runtime.

At the northern boundary of the Atlantic the heat transport starts at about 0.4 PW for all runs. The heat transport is volatile, even though it is averaged for 15 years, using a rolling mean. In the SSP5-8.5_{ice} the heat flux increases on average, reaching a maximum at about 2750 at 0.7 PW., then it drops to about 0.5 PW in 3000. The other run of the warmer scenario continues to drop from about 0.4 PW to 0.5 PW in 3000. The runs for the SSP2-4.5 show a drop in heat flux to about 0.3 PW in 2200, and continue to grow to 0.4 PW at the end of the runs.

While the northward heat transports at the Atlantic’s boundaries were in general positive, the heat flux from the SO into the Indo-Pacific were negative. Starting out at

a strength of -0.5 PW at the beginning of the runs in 1850, the SSP5-8.5_{ice} run weakens continuously to -6.5 PW in 3000. The SSP5-8.5_{no ice} run shows a slightly lower decline in heat transport, after the year 2300 it is stronger than the other run by about 0.2 PW until the end of the run. The runs for the SSP2-4.5 scenarios match closely. They weaken continuously down to a value of -5.75 PW at 3000.

3.2.5 Atmospheric Freshwater Fluxes

In Figure 3.10 the net precipitation over the four defined ocean basins and the global ocean for the four runs are shown. First the SSP5-8.5_{ice} run is described.

The precipitation pattern does not change much until 2000. The net precipitation being 1 Sv over the SO, 0.5 Sv over the NO, -0.5 Sv over the IPO and -0.75 Sv over the Atlantic Ocean.

After that, the pattern changes for about 100 years. Then the precipitation over the SO increases by 0.2 Sv, the precipitation over the NO increases by 0.25 Sv and the precipitation of the IPO by about 0.1 Sv. The net evaporation of the Atlantic increases by about 0.6 Sv. The Global Ocean experiences a net precipitation by about 0.05 Sv after the year 2100 (continental runoff is included in the net precipitation). After this change, the net precipitation stays roughly constant.

The difference to the SSP5-8.5_{no ice} scenario can be seen in the plot in the third row of the second column of Figure 3.10. The global freshwater flux is positive, roughly aligning with an additional freshwater influx into the NO of about 0.1 Sv.

In the fourth row of the first column of Figure 3.10, the difference between the SSP5-8.5 and the colder SSP2-4.5 scenario runs are plotted. The additional radiative forcing in the warmer scenario leads to a substantial relative increase of the net evaporation in the Atlantic basin. Beginning in 2050 the net evaporation begins to increase relative to the colder scenario, until 2200 it increases linearly to a value of about 0.4 Sv, after which it increases slower and reaches its maximum of 0.5 Sv more evaporation in the Atlantic, compared to the colder scenario. The NO and SO both experience an increased net precipitation, equilibrating at about 0.2 Sv each. The precipitation of the IPO increases relative to the colder scenarios. Its increase reaches an average value of 0.1 Sv. The freshening of the Global Ocean also reaches a value of about 0.1 Sv.

3.3 Comparison of the Freshwater Fluxes of AWI-ESM and the 9-BM

The 9-BM is supposed to characterize important aspects of the global circulation, that contribute to the stability of the AMOC. Computed freshwater fluxes calculated with the

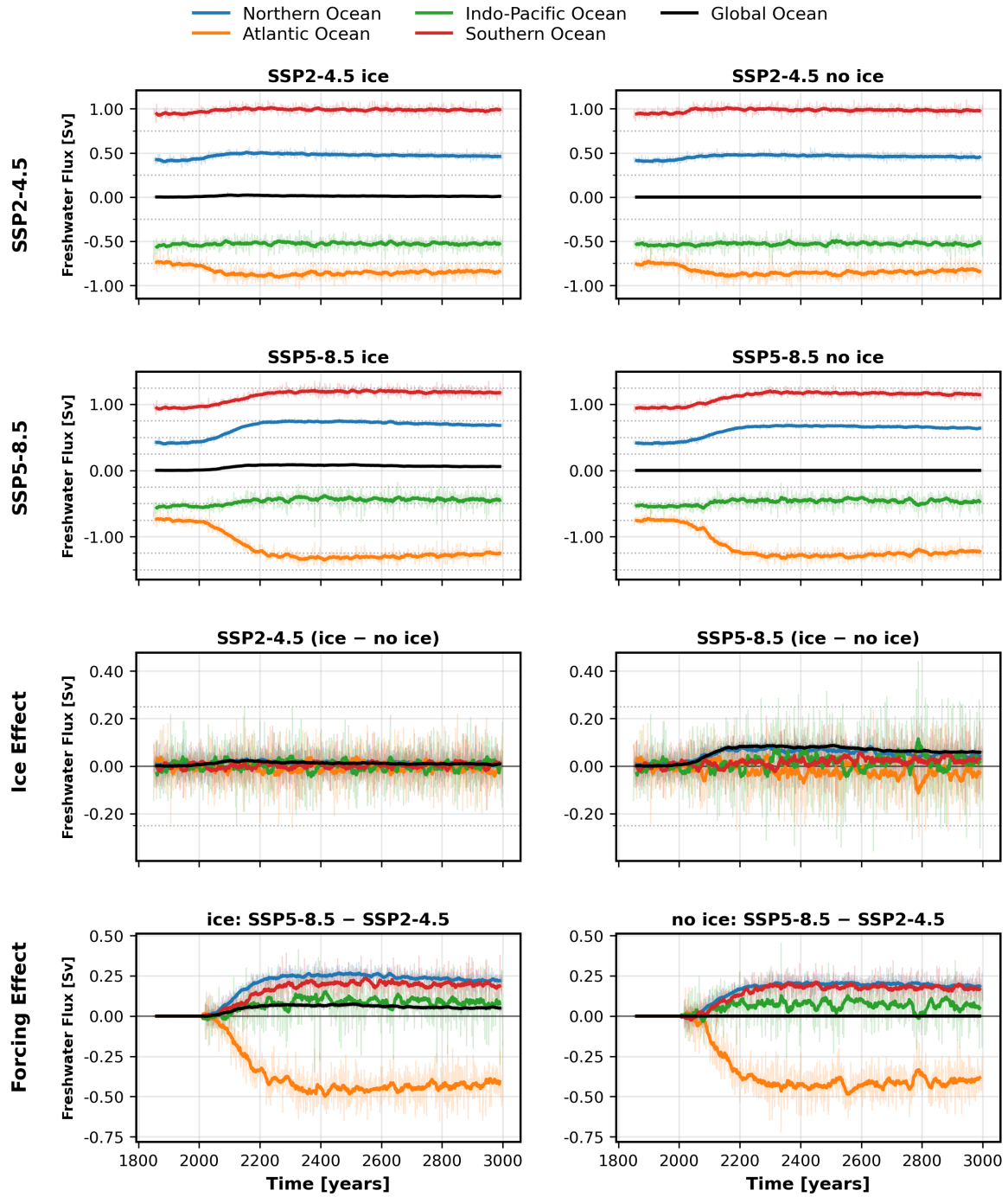


Figure 3.10: The first four plots show the net atmospheric freshwater flux into the ocean basin for the four runs. In the third row, the difference of the same scenario, but once with an active interactive Greenland ice sheet, and one with it being deactivated is plotted. In the fourth row, the difference of the warming scenarios is shown, once for an interactive Greenland ice sheet, and for the prescribed one.

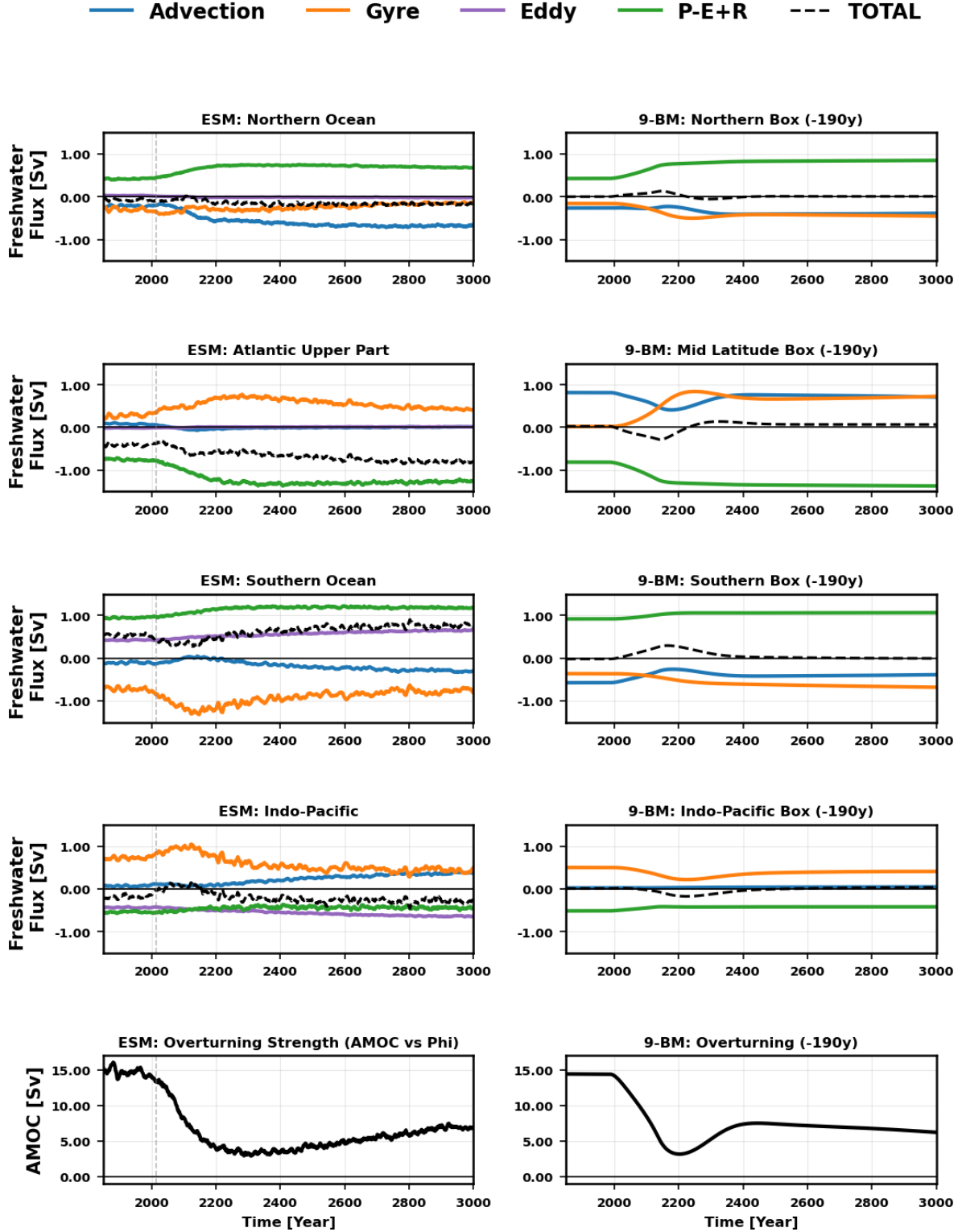


Figure 3.11: In the left column the net freshwater fluxes into a given oceanic basin, defined as in Figure 2.4. For the Atlantic only the upper part is considered. On the right column the freshwater fluxes acting on the given boxes are shown. The last row shows the AMOC trend for both the AWI-ESM and the 9-BM runs. The AWI-ESM run used the scenario SSP5-8.5 with the interactive Greenland ice sheet. The 9-BM run used the parameter configuration (5,5) and the warming of about 7.6 K. The BM run was shifted by 190 years, to fit temporally the ESM run.

output of AWI-ESM for SSP5-8.5_{ice}, and the 9-BM run, shown in Figure 3.1, will now be compared (Fig. 3.11).

The ocean of AWI-ESM was split in four different basins (Fig. 2.4). The Atlantic was split into an upper part (0-1000 m) and a lower part (1000-2500 m), corresponding to the upper and lower AMOC cells (Fig. 3.4). The other basin areas were integrated down to a depth of 2500 m. To match the time of the warming events, the results of the box model simulation were shifted by 190 years.

Between all the boundaries of these defined ESM-ocean-boxes, shown in the left column, the freshwater fluxes were calculated, as was done for 34°S at the Atlantic in Sec. 3.2.3. The fluxes were divided in incoming and outgoing freshwater fluxes. The outgoing freshwater fluxes of these boxes were subtracted and the incoming freshwater fluxes were added. In the Atlantic' upper and lower part, the freshwater fluxes incoming from the boundary at 34°S were added, and then subtracted by the corresponding outgoing freshwater fluxes at 45°N. Only in the AUPO net precipitation was added. In the SO the outgoing freshwater fluxes into the IPO and the Atlantic were subtracted, and the net precipitation added. The IPO experienced freshwater influx from the SO, and net precipitation. The Bering strait freshwater transport and the barotropic freshwater transports were small, and therefore neglected.

The last row represents the computed AMOC in both AWI-ESM and 9-BM.

In the right column the results of the 9-BM output are shown. The advection component is composed of the overturning flux and the Bering Strait flux, entering the box (Eqs. 2.4). The gyre part includes all the mixing terms (Eqs. 2.5), relevant for the box, including the Agulhas leakage transport. The net precipitation consists of the sum of all atmospheric freshwater fluxes acting on each box, including the melting term for the Northern Box.

In the first row the freshwater terms of the northern ocean and the Northern Box are placed side by side. The net precipitation for both models is about the same. It starts out with a net precipitation of 0.5 Sv, converges to a value of about 0.8 Sv in the ESM and 0.9 Sv in the 9-BM. The gyre term exports over all time for both model freshwater. The increase of this export from 2100-2200 in the 9-BM can not be seen in the ESM output. The advection exports freshwater (imports saltwater) from the NO/NB in both models. While this increase of the freshwater export is more pronounced in the NO, were it rises from about 0.0 Sv to about 0.8 Sv, the freshwater export from the NB rises from 0.2 Sv to about 0.3 Sv in 3000.

The net precipitation at the AUPO/MLB is in both models negative. In the ESM it starts out at -0.8 Sv and converges to a value of about -1.7 Sv, and in the BM it starts out at about -0.8 Sv too, but increases to about -1.8 Sv. In both models the gyre part increases over time. While the in the AUPO the the gyre freshwater flux starts out at 0.3 Sv and converges at 0.4 Sv, in the BM the upper part of the Atlantic starts out at

about 0.0 Sv, it increases over time to about 0.8 Sv. The advection term for the AUPO is roughly zero, while in the 9-BM it starts out at 0.9 Sv and converges to a strength of 0.8 Sv

In the Southern Ocean the eddy term plays an important part, that is not considered in the Southern Box of the 9-BM. It has a value of about 0.5 Sv of freshwater transport and increases slightly over time to about 0.6 Sv. The precipitation starts out at about 1.0 Sv and reaches a value of about 1.1 Sv in both the SO and the SB. The gyre component acting on SO/SB is negative in both models. While in 1850 it has a value of -0.8 Sv in the ESM, it increases until 2150 and drops back to about the same as the starting value in 3000. In the 9-BM the gyre term of the oceanic mixing starts out at about -0.3 Sv and converges to about -0.6 Sv. The advection part for the SO is roughly zero, increasing to about -0.2 Sv after 2200, while in the SB it is strictly negative, starting at a freshwater import of -0.7 Sv and reaching a strength of 0.4 Sv in 3000.

In the IPO of the ESM the eddy term is again important, with a freshwater flux of about -0.5 Sv. The precipitation in the IPO is increases slightly from -0.5 Sv to -0.4 Sv, similar values for the IPB. The gyre freshwater flux in the ESM starts out at about 0.8 Sv and reaches a final value of 0.5 Sv and in the BM it starts at 0.4 Sv and reaches a value of 0.3 Sv in the simulation year 3000. In the IPO an advection flux was calculated, which increases over time from 0.0 Sv to about 0.3 Sv in 3000. In the 9-BM the advection flux of the IPB is only related to the Bering Strait freshwater flux, which is negligible.

The advection and gyre fluxes are negligible or zero for both the ESM and the BM in the ALPO/DWB, and not shown here.

In the last row the AMOC trends for both AWI-ESM and the 9-BM run are shown. Both decrease from a strength of 15 Sv to about 3 Sv. The AMOC in the ESM reaches its minimum in 2300, while the BM reaches it in 2200. In the ESM the AMOC strength recovers over the duration of the run, but in the BM it starts decreasing again after a local maximum of 8 Sv in 2400.

Chapter 4

Discussion

In this chapter, the research questions are discussed. First, the 9-BM results will be compared to the results of the AWI-ESM. Then, the stability mechanism important for AMOC in the 9-BM will be analyzed, and the limitations of the model approach will be discussed. After that, it will be discussed whether the stability parameter F_{ovS} is a useful indicator of AMOC stability.

4.1 Comparison of the 9-BM with AWI-ESM

In Figure 3.11 freshwater fluxes going northward are shown for both AWI-ESM and 9-BM. Although they share similarities, both models differ clearly. Some of the differences are caused directly by the framework difference of the model, while other differences indicate different physical interactions.

Difference of the Freshwater Fluxes caused by the Different Model Framework

Limitations of the 9-BM

The logic of BMs and ESMs differs. While the goal of an ESM is to match nature and its processes as closely as possible, Box Models simplify nature using a set of assumptions to deepen the understanding of certain processes (Dijkstra, 2024). The simplifications used in the construction of the 9-BM cause deviations with AWI-ESM, which was used as a reference point in the model development.

The simplifications include that the ocean was assumed to be separable into given homogeneous volumes. This does not match reality or ESMs, since the oceanic water is stratified by density, and therefore are not homogeneous in terms of salinity and temperature. The selection and borders of the boxes were chosen, following assumptions of the underlying physics. In the case of the 9-BM, there was an important assumption that

the density difference of the northern and southern oceans drives the AMOC (Rahmstorf, 1996). This implies a high focus on freshwater and heat fluxes between the boxes. Other theoretical assumptions, which would value the importance of isopycnal outcrop in the SO higher for the AMOC strength, would yield a different design (such as Cimadoribus et al. (2012)). The effect that an enhanced stratification of the deep convection decreases the amount of deep convection and therefore reduces the AMOC strength (Munk & Wunsch, 1998) is also neglected.

The exact boundaries of the oceanic boxes are somewhat arbitrary. The depth of the boxes were chosen to be from 0 - 1000 m and from 1000 - 2500 m depth, since the overturning streamfunction for the year 2000 showed in the upper part a northward and below a southward flux, but the depth ranges are changing over time. The latitudinal boundaries also possess some arbitrary elements. The southern boundary of the Atlantic could be chosen closer to the equator, or the definition of the Northern Box, which is treated as the deep convecting area in the North Atlantic, could be chosen to not contain all the water masses above 45°N.

The fluxes that connect the boxes are also simplified. Simple formulas were used to approximate the exchanges of vast water masses, where information is necessarily lost. This, for example, is the case for the overturning flux, where the depth stratification at the boundary is not considered in the BM. The mechanisms that were deemed important were added to the system. Still, others were neglected: for example, a change of the latitude of the Intertropical Convergence Zone, which would change wind and precipitation patterns, and which itself is theorized to be dependent on the AMOC heat transport (Weijer et al., 2019).

Since not every process is calculated by computing the interaction of all of the parts with each other, but by simplifying the real processes, a parametrization is necessary. This parametrization is chosen to match the AMOC system as closely as possible to the ESM output. But since not every interaction can be considered, the parametrization cannot perfectly mirror the real processes, and the scaling strength of the parameters possesses an element of arbitrariness.

The oceanic freshwater fluxes are of the highest importance for the AMOC stability; the differences in the model calculations will be discussed now.

Difference in Calculation of Freshwater Fluxes between BM and ESM

A notable difference is that the ESM contains an Eddy freshwater flux, which has no equivalent in the 9-BM.

Overturning Freshwater Flux

A less notable difference is that the oceanic overturning freshwater fluxes are calculated differently, and hence, a systematic difference occurs. The formula for the calculation of the overturning freshwater flux in the ESM at the southern boundary of the Atlantic is defined as:

$$F_{\text{ovS}}^{ESM} = -\frac{1}{S_0} \int_{-2500}^0 \left[\int_{x_w}^{x_e} (v - \hat{v}) dx \right] [\bar{S} - S_0] dz \quad (4.1)$$

In Eq. 4.1, the $(v - \hat{v})$ term, which is integrated by the width of the basin and the depth of 2500 m, represents the influence of the circulation strength on the freshwater transport. The difference of the average basin velocity is subtracted to exclude non-circulation terms, such as the Bering Strait precipitation influence, that cause a net flux across the boundary. The depth integral of the upper 1000 m conceptually represents the AMOC strength. Because of the volume conservation of the water in a given basin, the reflux below 1000 m has about the same value.

The overturning freshwater flux of the 9-BM can be rewritten as:

$$F_{\text{ovS}}^{9BM} = -\frac{1}{S_0} (S_s - S_d) \Phi \quad (4.2)$$

$$F_{\text{ovS}}^{9BM} = -\frac{1}{S_0} ([S_s - S_0] - [S_d - S_0]) \Phi \quad (4.3)$$

$$= -\frac{\Phi}{S_0} [S_s - S_0] \Big|_{-1000}^0 + \frac{\Phi}{S_0} [S_d - S_0] \Big|_{-2500}^{-1000} \quad (4.4)$$

$$\approx -\frac{\Phi}{S_0} [\bar{S} - S_0] \Big|_{-2500}^0 \quad (4.5)$$

The upper water mass (0 m – -1000 m), that is transported via AMOC into the MLB, has the salinity of the Southern Box, while the incoming water mass from the DWB has the salinity of the DWB.

A systematic difference between the models arises here because the ESM models the salinity of the whole section, where the salinity is non-homogeneously distributed. Therefore, the incoming water mass from the lower part of the Atlantic does not just replace some average water mass of the Southern Ocean, as is the case in the 9-BM.

Additionally, the salinities in the 9-BM, after it reaches its equilibrium, do not match up with the salinities in AWI-ESM, which leads to a different average salinity of each basin, too (Fig. 4.1).

Gyre Freshwater Flux

At the southern boundary of the Atlantic the gyre flux is defined like this for each model:

$$F_{azS}^{ESM} = -\frac{1}{S_0} \int_{z_1}^{z_0} \int_{x_w}^{x_e} (S' \cdot v')(y = 34^\circ S) dx dz \quad (4.6)$$

$$F_{azS}^{9BM} = -\frac{1}{S_0} M_{ms}(S_s - S_m) \quad (4.7)$$

In the ESM, the deviation of the average velocity of the section v' is, is formally defined as the velocity deviation of the whole section, but is negligible for the lower part of the ocean. In this regard the definition of F_{azS}^{9BM} matches the logic of the freshwater flux in the ESM, when it only takes into account the gyre mixing in the upper part of the Atlantic with the Southern Ocean ($S_s - S_m$). Again, the SB and the MLB are treated as homogeneous boxes in the 9-BM, which leads to a systematic error when compared to the ESM and the average salinity of the boxes differs in AWI-ESM and 9-BM. Additionally, in the box model, the gyre strength (M_{ms}) is constant, while in the ESM, it is dependent on the wind patterns in the gyre region, which changes with global warming (Pontes et al., 2016).

In the ESM, the deviation of the average velocity of the section v' is formally defined as the velocity deviation of the whole section, but it is negligible for the lower part of the ocean. In this regard, the definition of F_{azS}^{9BM} matches the logic of the freshwater flux in the ESM, when it only takes into account the gyre mixing in the upper part of the Atlantic with the Southern Ocean ($S_s - S_m$). Again, the SB and the MLB are treated as homogeneous boxes in the 9-BM, which leads to a systematic error when compared to the ESM, and again the average salinity of the basins/boxes differs in AWI-ESM and 9-BM. Additionally, in the box model, the gyre strength (M_{ms}) is constant, while in the ESM, it is dependent on the wind patterns in the gyre region, which changes with global warming.

After this general analysis of the difference in calculation of the freshwater fluxes in BM and ESM, now a more in-depth focus will be taken on the runs of the 9-BM, and ESM runs in Figure 3.11.

Atmospheric Freshwater fluxes

In the ESM, the atmospheric processes were modeled using ECHAM6 (Popke et al., 2013), a complex GCM, while the atmospheric freshwater fluxes were modeled in the 9-BM by fairly simple linear approximations, assuming that the temperature of and the temperature gradient between the boxes are driving the freshwater exchange. In the case of the wind transport from the MLB to the IPB, a fixed scalar was chosen, which was multiplied by the moisture of the MLB, which is dependent on temperature. These

interactions were scaled such that the atmospheric fluxes in the BM correspond closely to the ESM.

Comparison of the Freshwater Fluxes

Three main mechanisms that cause a difference in the freshwater fluxes in AWI-ESM and 9-BM were identified. 1) The average salinity of the defined basins/boxes differs, 2) the salinity is stratified in the ESM, while it is treated as homogeneous in the BM, 3) the gyre strength is assumed to be constant in the BM.

Overturning Freshwater Flux

Starting at the SO/SB, it is clear that the freshwater overturning flux for the SO is close to zero, while in the SB it is strictly exporting freshwater. The F_{ov} in Figure 3.11 includes both the freshwater fluxes into the IPO and the AUPO, which add up to a negative freshwater flux into the SO. Strictly for the freshwater overturning flux into the AMOC system, the overturning is negative after the year 2000, at about -0.05 Sv (Fig. 3.7). On the other hand, the SB experiences a salinification of about 0.5 Sv by the overturning, such that it has a value for F_{ov} into the Atlantic of 0.5 Sv.

The first difference arises from the fact that the salinity difference is smaller in the ESM than in the BM. In the SSP5-8.5 run, the salinity difference of the ALPO and SO is about 0.4 PSU in the year 2000, while in the BM, the difference of DWB and SB is about 1.3 PSU. Therefore, it is expected that the SB exports more freshwater compared to the SO, since it is relatively fresher. That the northward freshwater transport of the ESM is negative at the southern boundary of the Atlantic cannot be an effect of the average salinity difference of SO and Atlantic water, since the SO is fresher, and therefore must be because of the stratification of the water, exporting saltier than the sectional-average water northward. As is shown in Figure 4.2, especially in the top 300 m, where the water is saltiest, the northward velocity is highest.

The stratification of the overturning freshwater flux changes over time (Fig. 3.8). The section depth, where negative freshwater flux takes place, is increasing while the AMOC strength is decreasing. This was not taken into account in the 9-BM.

Since the SO exports negative freshwater into the AUPO after 2000, while the SB exports freshwater into the MLB, the imported freshwater of the AUPO and MLB differ significantly. Therefore, the 9-BM produces a freshening of the MLB, instead of a salinification as in the AUPO. This has implications for the stability mechanisms of both of these systems, which will be discussed in the next section 4.2.

The increase in negative freshwater transport into the NO and NB matches fairly closely. Since the AMOC strength is decreasing in both models, the increase of average salinity due to precipitation changes in the MLB/AUPO is counteracting the effect of a

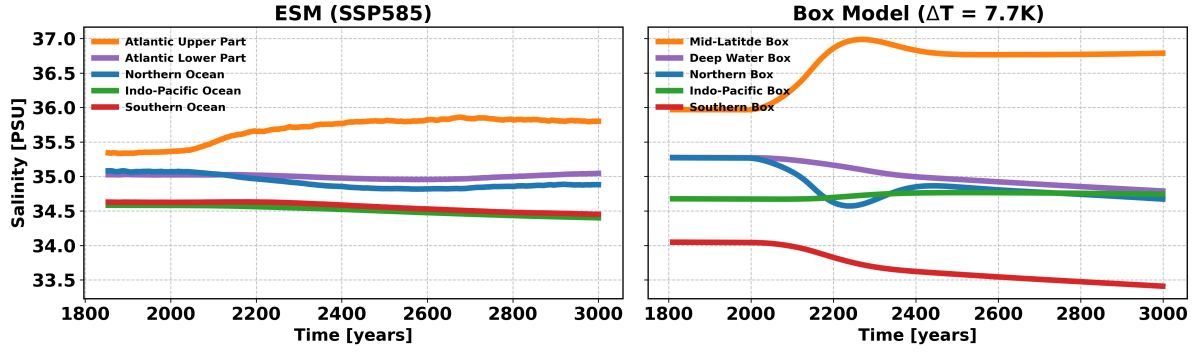


Figure 4.1: The average salinities of the run in SSP5-8.5 with AWI-ESM and the 9-BM in the warming run of 7.67 K differ. This leads to different oceanic freshwater fluxes, which are important for the stability of AMOC, even when the water volume transport is the same.

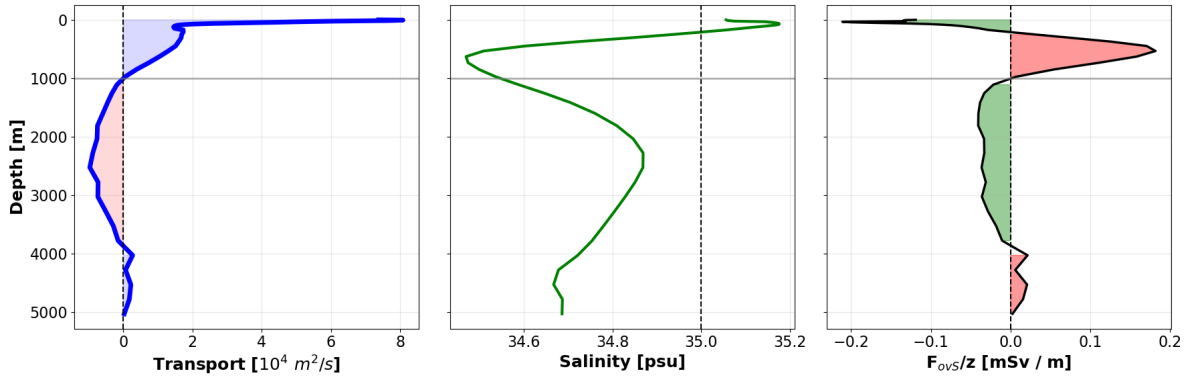


Figure 4.2: The freshwater transport via the overturning (right) is dependent on the stratification of the average velocity multiplied by the basin width at that depth (left) and salinity (middle) of the section. The plots were created for the simulation year 2000 in the SSP5-8.5_{ice} scenario, at the southern boundary of the Atlantic at 34°S.

decreased overturning transport, and is the cause for a net decrease in the overturning freshwater transport, in both NO and NB.

The overturning flux in the IPO is increasing over time, while in the IPB, this flux is negligible. This difference is caused by a different behavior of the overturning system in AWI-ESM, and may be the effect of the Pacific Meridional Overturning Circulation (PMOC), which is not accounted for in the 9-BM.

Gyre Freshwater Flux

The gyre freshwater fluxes decrease in both the SO and SB after the warming event. This is driven in both cases by an increase in the difference of the average salinity of the water in the AUPO/MLB and the SO/SB. In the case of the ESM, this is mainly due to the salinification on the AUPO, while in the BM, both MLB increases and the SB decreases in salinity. In the ESM, even though the difference of salinity is increasing over time (the difference in salinity of SO and IPO stays roughly the same), the gyre negative freshwater flux decreases over time. This is probably due to a change in the wind and therefore gyre patterns.

In the NO, the gyre freshwater flux stays roughly constant, even though the difference of average salinity increases between the AUPO and the NO. This again is an indicator of changing wind patterns. The analogous flux in the BM is increasing, due to the difference of the salinity difference.

The freshwater flux into the AUPO is decreasing over time, after an increase until 2300 due to the increase in salinity difference, which follows from the same gyre changes between AUP and NO. The Agulhas leakage, which is essentially forced to transport saline water into the MLB, cancels the freshwater gyre fluxes at the beginning of the simulation of the BM, after which the gyre freshwater flux turns positive, due to the increase in MLB salinity and a decrease in SB and NB salinity, mainly due to precipitation changes.

Atmospheric Freshwater fluxes

The net precipitation, which is a sum of the precipitation, evaporation, and melt water influx, is closely related for the model runs. This is partly due to a fitting of scaling parameters, such that both system experience similar changes in precipitation, following a change in the warming of the different regions.

Melting Freshwater Flux

The meltwater influx in the ESM can be gathered from the plot on the third row on the right in Figure 3.10. Since the total difference of freshwater influx in the SSP2-4.5 scenario with GIS and without is negligible, the black graph in the SSP5-8.5 with GIS and without is representing the freshwater influx into the northern ocean due to the melting

of Greenland. In the ESM, this value is about 0.1 Sv. The melting of the 9-BM has, in the warmest scenario, about the same strength.

4.2 Stability Mechanisms of the 9-BM and AWI-ESM on AMOC

4.2.1 Stability of AMOC in the 9-BM

Even though the 9-BM run, which is modeled the closest to the AWI-ESM run of SSP5-8.5_{ice} in the highest emission scenario, remains in a positive phase, and recovers after a weakening fairly quickly (although it is not in equilibrium yet; see Fig. 2.2), other choices of parameters in the 9-BM lead to a negative AMOC state (Fig. 3.2). This is important, since the model parameters have an intrinsic uncertainty or error when put in relation to the real AMOC system.

The increase of the atmospheric interaction between the Indo-Pacific Box and the AMOC boxes leads to a more stable AMOC state, while the increase of the interaction between the oceanic boxes of the Indo-Pacific and the Atlantic boxes leads to a more unstable AMOC. The increase in global temperature leads to more collapsed states.

Atmospheric Interaction

The atmospheric interaction was decomposed into three parts. The "wind strength" of the MLB to the IPB, which acts as a scalar of the wind transport multiplied by the humidity of the MLB-A, determines the precipitation and evaporation of IPB and MLB, respectively. The other wind-related interaction is the transport of water from the IPB to the SB, which is multiplied by a scalar. Additionally, the area of the SO and its atmosphere were scaled up, when the atmospheric interaction constant increased, to ensure a stable initial equilibrium.

The wind transport from the MLB to the IPB acts as a stabilizer of AMOC. Its freshwater transport in the IPB is salinifying the MLB, as well as cooling it. Since, if AMOC is in a positive phase, the salinity transport northward is bigger than southward, this increases the salinity in the NB, and therefore its density and AMOC strength.

The atmospheric interaction between the IPB-A and SB-A leads to a relative freshening of the SB. This leads to a less dense SB and, therefore, a stronger AMOC.

Oceanic Interaction

The oceanic interaction index modulates the Bering Strait transport, the mixing of the IPB with the SB and the Agulhas leakage flux, modeled as a mixing of the IPB with the MLB. The Bering Strait influence is small compared to the other fluxes, but if the

AMOC is self-amplifying, the Bering Strait term, acting in the same direction as AMOC, is strengthening AMOC. Both mixing terms together act as destabilizers on AMOC, since after 2100, the salt influx into the NB is higher than in the SB (Fig. 4.4). The effect on AMOC by the mixing of the SB and IPB and the Agulhas leakage individually is less clear, since the mixing of the SB with the IPB increases the salinity of the SB and increases its temperature; so both a weakening of AMOC by the increase of the haline density and a strengthening by the decrease of the thermal density. Most of the water introduced into the MLB by the Agulhas leakage is transported into the north. Its influence on the NB is a salinification and a warming, increasing the haline density, therefore strengthening, and decreasing its thermal density, therefore weakening AMOC.

In Figure 4.3, the influence of the mixing of IPB and SB and the Agulhas leakage on the circulation is shown. The cyan box resembles the same run as the cyan box in Figure 3.2. Instead of atmospheric and oceanic interaction strength, only the two oceanic terms were varied. For the two lower emission and the highest emission scenario, these oceanic terms show no influence on the stability of AMOC, but in the warming case of 5.61 K the oceanic terms matter. The increase of each term leads to a weakening of the circulation. For the stability, therefore, the warming of the Agulhas leakage of the NB and the warming of the SB by the mixing term dominate each over their haline part in terms of stability.

Comparison of the 9-BM to Rahmstorf's Model

In Rahmstorf's Box Model (Rahmstorf, 1996), the feedback of AMOC is determined by a net evaporation on the SB, therefore, a negative freshwater flux into the MLB. This negative freshwater flux then increases the salinity of the NB, counteracting the initial salinity increase by evaporation in the SB. In the 9-BM, instead of the changing the atmospheric freshwater transport from the MLB into the SB, the parameter changed is the CO₂-level. The atmospheric freshwater transport into the SB is always positive; the feedback mechanism described in Rahmstorf's model does not take place. Still, the warming in different model runs leads not just to a weakening of AMOC, but to a collapse, indicating another feedback mechanism that, when AMOC gets weaker, leads to a further weakening. The negative freshwater overturning flux at the southern boundary of the Atlantic is not necessary for this kind of collapse or reversal of AMOC in the 9-BM. In both BMs exist for certain control parameters multiple stable states, the feedback that keeps the 9-BM stable, is of another nature, since both the SB but also the NB import freshwater.

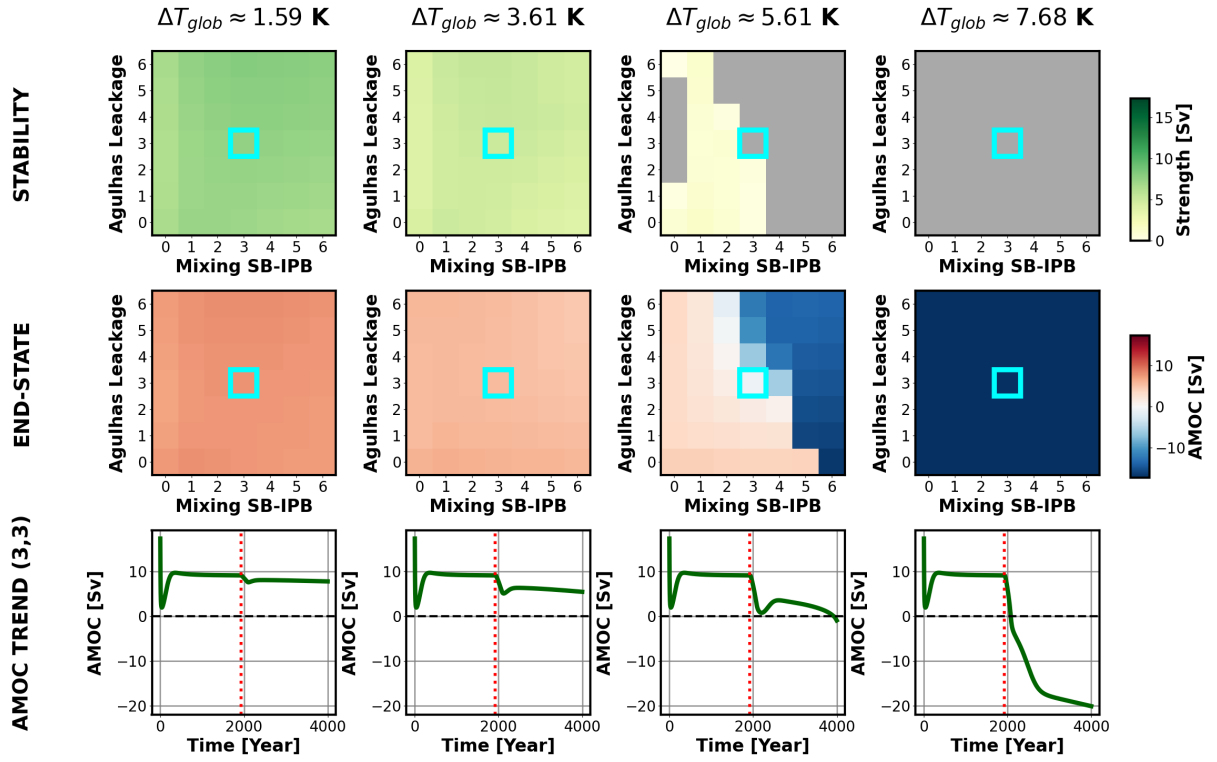


Figure 4.3: This stability map was created like in Figure 3.2, but only the oceanic interaction of the AMOC system with the Indo-Pacific Box was considered. On the horizontal axes, the interaction strength of the mixing of IPB and SB, and on the vertical axes, the interaction strength of the Agulhas leakage transporting water between IPB and MLB are shown. The cyan box resembles the cyan box in Figure 3.2. The last row shows the AMOC-trend for the cyan boxes.

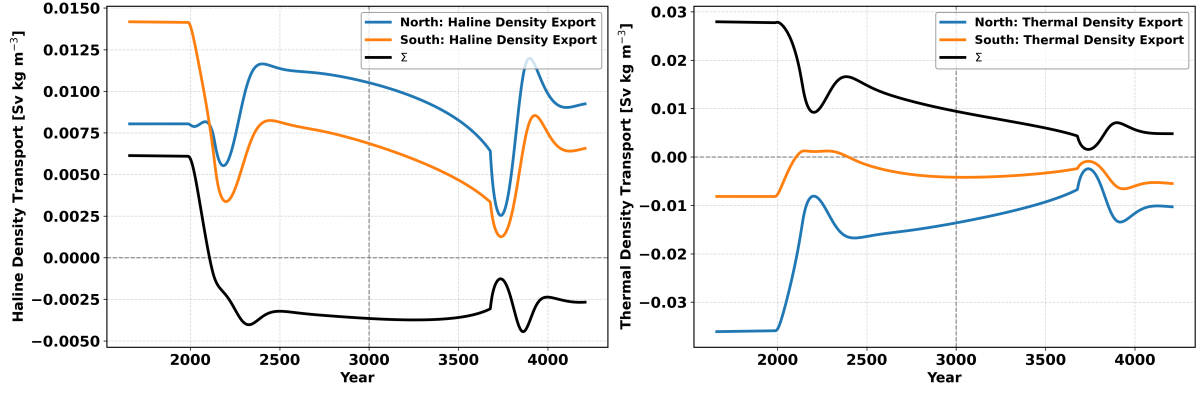


Figure 4.4: The density anomaly transport of the NB (blue) and SB (orange) is shown. On the left side, the haline density export and on the right side the thermal density export of a given box are plotted. The black graphs represent the difference between the southern and northern density export, essentially showing the Σ value of the water transport in units of density transport.

Relevance of Heat and Freshwater Fluxes in the 9-BM

In the case of the mixing of the IPB with the SB, and the Agulhas leakage between IPB and MLB, the warming terms matter in terms of stability over the freshwater terms. The feedback mechanism in the 9-BM has to be different from that in the Rahmstorf model. Therefore now a close look at the haline and thermal density transports is taken.

The SB is net exporting freshwater, but the density difference between the northern and southern boxes is also dependent on the freshwater flux into the NB. The parameter that is used to describe the difference of the freshwater overturning flux into the north and south is called Σ (Huisman et al., 2010)(Weijer et al., 2019).

$$\Sigma = F_{ovS} - F_{ovN} \quad (4.8)$$

If Σ is negative, the meridional overturning is increasing the density difference, and could therefore be another stability indicator. On the left side of Figure 4.4, the values of the freshwater overturning flux in units of the haline density transport at the northern and southern boundary and Σ are shown for the run duration from 1850 to 4500. In the 9-BM run, the initial equilibrium Σ has a positive value. After the warming is applied, the value turns negative, which is caused by the increased freshwater influx caused by the melting of the term in the north, and by the increase of salinity in the MLB, which is caused by the increased evaporation due to global warming. AMOC itself continues to stay in the negative Σ -state, which is conceptually self-amplifying. A decrease of AMOC strength should then lead to a decrease of salt-advection into the north, although the value is roughly constant during the decline of AMOC from 2500 to 3500.

The heat flux, which is ignored in the Rahmstorf model, has a positive sigma value over the whole duration of the run. It is shown, in units of thermal density transport on

the right side of Figure 4.8. The NB is importing heat, relative to the SB. This reduces the density difference between the boxes, and is therefore amounting to a weakening of AMOC relative to the AMOC strength itself, which is a dampening or stabilizing feedback.

The heat advection feedback in the 9-BM is stronger than the corresponding salinity feedback and dominates it.

Synthesis of Stability Mechanism in the 9-BM

The oceanic and atmospheric parameters of the model determine the possibility of an AMOC collapse. The dampening or amplifying feedback mechanisms of the heat and freshwater transport the reaction of AMOC on a perturbation of the AMOC strength. Even though the heat transport transports a bigger density change in the north, the effect on AMOC is two-fold. After an initial perturbation, the weakening of AMOC is dampened. But since the heat transport is faster than the freshwater transport (Cessi, 1994), the heat gradients are saturated faster, especially with the inclusion of atmospheric heat transports (Marotzke & Stone, 1995), after which the freshwater transport matters for the existence of multiple stable equilibria (which do exist in the 9-BM; Fig. 3.3).

Here, the timescale of interest is of importance. Over longer time scales the positive sigma value can increase the density difference between north and south, which over a long time scale stabilizes AMOC. A decrease of AMOC strength would weaken the AMOC, and the positive feedback would cease to exist, but the weakening would take a large amount of time, and the more short term heat feedback would work to stabilize the AMOC, and hence lead to a slower slowdown.

4.2.2 Stability of AMOC in AWI-ESM

In AWI-ESM, the changes of the freshwater fluxes are in first degree effects of the change in precipitation. Especially visible is this in the azonal freshwater mixing term, where the increase of precipitation in the SO and the increase of evaporation in the AUPO leads to a sudden increase in freshwater transport northward. This then dominates the oceanic freshwater transport for a long time via overturning.

Before the warming takes place, the overturning freshwater transport is positive, hence in the Rahmstorf framework, a negative, and hence stable feedback would be concluded. The latter change of sign would not lead to two widely different stability regimes, since the AMOC did not have the time to equilibrate yet.

After the warming took place, the SO transports saltwater northward, and the NO imports saltwater. This is amplifying AMOC, so it then enters a regime where the salt-advection feedback is driving AMOC. Since both parameter inequalities $F_{ovS} < 0$ and $\Sigma < 0$ are met, this indicates a convergence to a bistable regime. The runtime of the

SSP5-8.5_{ice} run is not sufficient to see if this is then a new bistable regime, or if it then recovers to another state where it equilibrates with $F_{ovS} \geq 0$ or $\Sigma \geq 0$, which should then be stable.

On a short timescale, after the initial perturbation due to warming and glacier melting, AMOC weakens significantly. But the northward advective transport of heat slows the weakening on a short timescale. Additionally, the salinity increases in the AUPO, which leads to a salinifying effect of AMOC on the Northern Ocean. Both the reduced heat transport and the increased salinity transported by AMOC lead to a stabilization. If the heating of the AUPO, which then warms the NB, is not sufficient to drive the AMOC to a sudden collapse after an initial weakening, this additional heat would diffuse relatively quickly, with the slower salt effect remaining.

After a longer but not yet equilibrium timescale, the salt would also distribute via the NB to the ALPO, then to the SO, and to the IPO. If equilibrium then is reached, the change of the salinity due to a change in precipitation patterns of the ALPO would possibly not matter anymore in terms of stability. This can be studied in the Rahmstorf Model, where the salinity of the MLB does not matter for the stability regime (Rahmstorf, 1996).

This increases the interest of the question, whether the sign of F_{ovS} or Σ before the warming takes place, is correct. If the real ocean has a negative sign of F_{ovS} , this does not necessarily mean that there are two clearly distinct equilibrium states, since the time to develop the difference between the two states may not have been sufficient to clearly diverge. Data of F_{ovS} and Σ before the warming took place would be necessary to draw better conclusions.

When the freshwater fluxes and AMOC strength were compared in Figure 3.11, only a shorter time range than was calculated was shown for the 9-BM run. The complete run shows that the AMOC actually takes a much longer time to equilibrate (Fig. 2.2). Before that, it experienced severe fluctuations in its strength on a timescale of about 1000 years. In the runtime of the ESM, the system has not reached equilibrium yet, and the 9-BM suggests that AMOC may undergo some severe changes until the equilibrium state is reached, even when external forcings are constant.

4.3 Is the Overturning Freshwater Flux a good indicator of AMOC stability?

The argument that F_{ovS} is a good stability indicator presupposes an equilibrium state (Vanderborght et al., 2025)(van Westen & Dijkstra, 2024). In that equilibrium state, the negative overturning freshwater flux at the southern boundary of the Atlantic was causing an increase of the salinity difference of the Northern and Southern Ocean areas, where

deep convection and up-welling are taking place. When the freshwater overturning flux recently changed its sign, it loses its importance. Additionally, on shorter timescales, the effect of other feedbacks has an additional importance, since they influence and change the salinity of the different ocean areas more quickly, which leads to a complex AMOC behavior, if it has not reached a stable equilibrium yet.

In the ESM, the precipitation changes influence the salinity change of the oceanic areas faster by at least an order of magnitude. These changes then influence the strength of the azonal mixing term, which is stronger than the freshwater overturning flux by about a factor of five. On shorter timescales, of a few hundred years, these effects dominate over the advective feedback, which might take a few thousand years to establish its significance (Haines et al., 2022).

The advective salt feedback in the ESM seems to be mainly driven in the time after the warming by the increase in salinity in the AUPO, which is transported northward (Mignac et al., 2019). Here, the salt-advective feedback may be a significant amplifier of weakening perturbations on AMOC, but is largely decoupled by the influx of saltwater at the southern boundary, since the precipitation changes dominate the change of salinity at the AUPO and NO on these timescales (Haines et al., 2022).

In the 9-BM, a collapse of AMOC is possible, even though the indicator F_{ovs} is positive. Additionally, it can be shown that multiple equilibria are possible in that model. Therefore, the indicator F_{ovs} is not a necessary condition for multiple stable equilibria in the 9-BM, and may not be in the real ocean either.

Chapter 5

Conclusions

Although there are clear differences from the AWI-ESM, the 9-BM successfully captures the key dynamics of the AMOC. The most important difference between the AWI-ESM and the real ocean is the clearly positive freshwater overturning transport (F_{ovS}) from the Southern Ocean to the Atlantic.

The finding that the 9-BM exhibits multiple stable states despite a positive F_{ovS} demonstrates that a negative F_{ovS} is not a strict prerequisite for AMOC stability. Therefore, other feedback mechanisms must exist to help stabilize the system in one state over another. As demonstrated in the AWI-ESM, different feedback mechanisms—such as heat and salt transport—act on different timescales. On shorter timescales, which for the AMOC might mean a few hundred years, heat transport likely plays a more dominant role than salinity transport.

While F_{ovS} may be relevant in determining the long-term bistability of the AMOC, common theoretical frameworks assume an equilibrium state with a negative F_{ovS} for bistability to exist. Consequently, it is important to investigate whether this sign has changed in the recent past. The AWI-ESM suggests that this sign may already be influenced by global warming.

Furthermore, the box model formulation demonstrates that Indo-Pacific influences can significantly impact AMOC stability, especially in high-emission scenarios. The atmospheric water transport in the 9-BM closely matched the precipitation changes calculated by the AWI-ESM. This is a crucial prerequisite for understanding the AMOC, as precipitation changes driven by global warming are central to its transient state transitions.

Because the 9-BM exhibits multiple stable states even with a positive F_{ovS} , the AMOC may need to maintain a certain threshold of strength to remain self-sustaining. Consequently, changes in AMOC strength of just a few Sverdrups could be critical. Therefore, the influence of melting Greenland and continental glaciers, which decrease deep convection in the North Atlantic, is highly relevant to overall system stability.

Finally, the 9-BM simulations indicate that following an initial warming perturbation, the AMOC does not necessarily simply converge to a new equilibrium state. Instead, un-

der constant forcing, it may undergo several transient state switches before stabilizing. The 9-BM suggests that these transitions occur on timescales ranging from several hundred to a few thousand years.

Sources

- Ackermann, L., Willeit, M., Knorr, G., et al. (2026, February). *Timescale competition controls tipping behaviour under climate overshoot* [Preprint (Version 1) available at Research Square]. <https://doi.org/10.21203/rs.3.rs-8941071/v1>
- Anisimov, O. A. (2007). Potential feedback of thawing permafrost to the global climate system through methane emission. *Environmental Research Letters*, 2(4), 045016. <https://doi.org/10.1088/1748-9326/2/4/045016>
- Arumí-Planas, C., Dong, S., Perez, R., Harrison, M. J., Farneti, R., & Hernández-Guerra, A. (2024). A multi-data set analysis of the freshwater transport by the atlantic meridional overturning circulation at nominally 34.5°s [e2023JC020558 2023JC020558]. *Journal of Geophysical Research: Oceans*, 129(6), e2023JC020558. <https://doi.org/https://doi.org/10.1029/2023JC020558>
- Ben-Yami, M., Good, P., Jackson, L. C., Crucifix, M., Hu, A., Saenko, O., Swingedouw, D., & Boers, N. (2024). Impacts of amoc collapse on monsoon rainfall: A multi-model comparison [e2023EF003959 2023EF003959]. *Earth's Future*, 12(9), e2023EF003959. <https://doi.org/https://doi.org/10.1029/2023EF003959>
- Bryden, H. L., King, B. A., & McCarthy, G. D. (2011). South atlantic overturning circulation at 24°s. *Journal of Marine Research*, 69(1), 38–55. https://elischolar.library.yale.edu/journal_of_marine_research/294
- Caesar, L., Rahmstorf, S., Robinson, A., Feulner, G., & Saba, V. (2018). Observed fingerprint of a weakening atlantic ocean overturning circulation [Epub 2018 Apr 11]. *Nature*, 556(7700), 191–196. <https://doi.org/10.1038/s41586-018-0006-5>
- Cessi, P. (1994). A simple box model of stochastically forced thermohaline flow. *Journal of Physical Oceanography*, 24(9), 1911–1920. [https://doi.org/10.1175/1520-0485\(1994\)024<1911:ASBMOS>2.0.CO;2](https://doi.org/10.1175/1520-0485(1994)024<1911:ASBMOS>2.0.CO;2)
- Cimatoribus, A. A., Drijfhout, S. S., & Dijkstra, H. A. (2012). Meridional overturning circulation: Stability and ocean feedbacks in a box model. *Climate Dynamics*, 42(1-2), 311–328. <https://doi.org/10.1007/s00382-012-1576-9>
- Dai, A. (2022). Arctic amplification is the main cause of the Atlantic meridional overturning circulation weakening under large CO₂ increases. *Climate Dynamics*, 58(11), 3243–3259. <https://doi.org/10.1007/s00382-021-06096-x>

- Danilov, S., Sidorenko, D., Wang, Q., & Jung, T. (2017). The finite-volume sea ice–ocean model (fesom2). *Geoscientific Model Development*, 10(2), 765–789. <https://doi.org/10.5194/gmd-10-765-2017>
- David Tàbara, J., Frantzeskaki, N., Hölscher, K., Pedde, S., Kok, K., Lamperti, F., Christensen, J. H., Jäger, J., & Berry, P. (2018). Positive tipping points in a rapidly warming world [Sustainability governance and transformation 2018]. *Current Opinion in Environmental Sustainability*, 31, 120–129. <https://doi.org/https://doi.org/10.1016/j.cosust.2018.01.012>
- de Vries, P., & Weber, S. L. (2005). The atlantic freshwater budget as a diagnostic for the existence of a stable shut down of the meridional overturning circulation. *Geophysical Research Letters*, 32(9). <https://doi.org/https://doi.org/10.1029/2004GL021450>
- Dijkstra, H. (2024). The role of conceptual models in climate research. *Physica D: Non-linear Phenomena*, 457, 133984. <https://doi.org/https://doi.org/10.1016/j.physd.2023.133984>
- Drijfhout, S. S., Weber, S. L., & van der Swaluw, E. (2011). The stability of the moc as diagnosed from model projections for pre-industrial, present and future climates. *Climate Dynamics*, 37(7), 1575–1586. <https://doi.org/10.1007/s00382-010-0930-z>
- Haines, K., Ferreira, D., & Mignac, D. (2022). Variability and feedbacks in the atlantic freshwater budget of cmip5 models with reference to atlantic meridional overturning circulation stability. *Frontiers in Marine Science*, Volume 9 - 2022. <https://doi.org/10.3389/fmars.2022.830821>
- Huisman, S. E., den Toom, M., Dijkstra, H. A., & Drijfhout, S. (2010). An indicator of the multiple equilibria regime of the atlantic meridional overturning circulation. *Journal of Physical Oceanography*, 40(3), 551–567. <https://doi.org/10.1175/2009JPO4215.1>
- Jackson, L. C., Biastoch, A., Buckley, M. W., Desbruyères, D. G., Frajka-Williams, E., Moat, B., & Robson, J. (2022). The evolution of the North Atlantic Meridional Overturning Circulation since 1980. *Nature Reviews Earth & Environment*, 3(4), 241–254. <https://doi.org/10.1038/s43017-022-00263-2>
- Jackson, L. C., Kahana, R., Graham, T., Ringer, M. A., Woollings, T., Mecking, J. V., & Wood, R. A. (2015). Global and european climate impacts of a slowdown of the AMOC in a high resolution GCM. *Climate Dynamics*, 45(11), 3299–3316. <https://doi.org/10.1007/s00382-015-2540-2>
- Johnson, H., Cessi, P., Marshall, D., Schloesser, F., & Spall, M. (2019). Recent contributions of theory to our understanding of the atlantic meridional overturning circulation. *Journal of Geophysical Research: Oceans*. <https://doi.org/10.1029/2019jc015330>

- Jüling, A., Zhang, X., Castellana, D., Von Der Heydt, A. S., & Dijkstra, H. A. (2021). The atlantic’s freshwater budget under climate change in the community earth system model with strongly eddying oceans. *Ocean Science*, 17(3), 729–754.
- Lee, H., Calvin, K., Dasgupta, D., Krinmer, G., Mukherji, A., Thorne, P., Trisos, C., Romero, J., Aldunce, P., Barret, K., et al. (2023). Synthesis report of the ipcc sixth assessment report (ar6), longer report. ipcc.
- Levang, S. J., & Schmitt, R. W. (2020). What causes the amoc to weaken in cmip5? *Journal of Climate*, 33(4), 1535–1545. <https://doi.org/10.1175/JCLI-D-19-0547.1>
- Liu, W., Xie, S.-P., Liu, Z., & Zhu, J. (2017). Overlooked possibility of a collapsed atlantic meridional overturning circulation in warming climate. *Science Advances*, 3(1), e1601666. <https://doi.org/10.1126/sciadv.1601666>
- Lohmann, G. (2012). Description of the 7-box-model. https://paleodyn.uni-bremen.de/gl/lecture/Schedule_2012_June_Borno.pdf
- Lohmann, G., & Schneider, J. (1999). Dynamics and predictability of stommel’s box model. a phase-space perspective with implications for decadal climate variability. *Tellus A: Dynamic Meteorology and Oceanography*, 51(2), 326–336. <https://doi.org/10.3402/tellusa.v51i2.12314>
- Ma, Q., Shi, X., Scholz, P., Sidorenko, D., Lohmann, G., & Ionita, M. (2024). Revisiting climate impacts of an amoc slowdown: Dependence on freshwater locations in the north atlantic. *Science Advances*, 10(47), eadr3243. <https://doi.org/10.1126/sciadv.adr3243>
- Marotzke, J., & Stone, P. H. (1995). Atmospheric transports, the thermohaline circulation, and flux adjustments in a simple coupled model. *Journal of Physical Oceanography*, 25(6), 1350–1364. [https://doi.org/10.1175/1520-0485\(1995\)025<1350:ATTTCA>2.0.CO;2](https://doi.org/10.1175/1520-0485(1995)025<1350:ATTTCA>2.0.CO;2)
- Meinshausen, M., Nicholls, Z. R. J., Lewis, J., Gidden, M. J., Vogel, E., Freund, M., Beyerle, U., Gessner, C., Nauels, A., Bauer, N., Canadell, J. G., Daniel, J. S., John, A., Krummel, P. B., Luderer, G., Meinshausen, N., Montzka, S. A., Rayner, P. J., Reimann, S., . . . Wang, R. H. J. (2020). The shared socio-economic pathway (ssp) greenhouse gas concentrations and their extensions to 2500. *Geoscientific Model Development*, 13(8), 3571–3605. <https://doi.org/10.5194/gmd-13-3571-2020>
- Mignac, D., Ferreira, D., & Haines, K. (2019). Decoupled freshwater transport and meridional overturning in the south atlantic. *Geophysical Research Letters*, 46(4), 2178–2186. <https://doi.org/https://doi.org/10.1029/2018GL081328>
- Munk, W., & Wunsch, C. (1998). Abyssal recipes ii: Energetics of tidal and wind mixing. *Deep Sea Research Part I: Oceanographic Research Papers*, 45(12), 1977–2010. [https://doi.org/https://doi.org/10.1016/S0967-0637\(98\)00070-3](https://doi.org/https://doi.org/10.1016/S0967-0637(98)00070-3)

- Nobre, C. A., & Borma, L. D. S. (2009). ‘tipping points’ for the amazon forest. *Current Opinion in Environmental Sustainability*, 1(1), 28–36. <https://doi.org/https://doi.org/10.1016/j.cosust.2009.07.003>
- Pierrehumbert, R. T. (2011). Infrared radiation and planetary temperature. *Physics Today*, 64(1), 33–38.
- Pontes, G. M., Sen Gupta, A., & Taschetto, A. S. (2016). Projected changes to south atlantic boundary currents and confluence region in the cmip5 models: The role of wind and deep ocean changes. *Environmental Research Letters*, 11(9), 094013. <https://doi.org/10.1088/1748-9326/11/9/094013>
- Popke, D., Stevens, B., & Voigt, A. (2013). Climate and climate change in a radiative-convective equilibrium version of echam6. *Journal of Advances in Modeling Earth Systems*, 5(1), 1–14.
- Prange, M., Lohmann, G., & Gerdes, R. (1997). Sensitivity of the thermohaline circulation for different climates – investigations with a simple atmosphere-ocean model. *Palaeoclimates*, 2(2), 71–99.
- Python Software Foundation. (2024). *Python language reference, version 3.12*. <https://www.python.org>
- Rahmstorf, S. (1996). On the freshwater forcing and transport of the atlantic thermohaline circulation. *Climate Dynamics*, 12(12), 799–811. <https://doi.org/10.1007/s003820050144>
- Rousselet, L., Cessi, P., & Forget, G. (2021). Coupling of the mid-depth and abyssal components of the global overturning circulation according to a state estimate. *Science Advances*, 7(21), eabf5478. <https://doi.org/10.1126/sciadv.abf5478>
- Schauer, U., & Losch, M. (2019). “freshwater” in the ocean is not a useful parameter in climate research. *Journal of Physical Oceanography*, 49(9), 2309–2321. <https://doi.org/10.1175/JPO-D-19-0102.1>
- Scholz, P. (2025). *Tripyview* [Accessed: 2025-07-19]. <https://github.com/FESOM/tripyview>
- Stommel, H. (1961). Thermohaline convection with two stable regimes of flow. *Tellus*, 13(2), 224–230. <https://doi.org/https://doi.org/10.1111/j.2153-3490.1961.tb00079.x>
- van Westen, R. M., & Dijkstra, H. A. (2024). Persistent climate model biases in the atlantic ocean’s freshwater transport. *Ocean Science*, 20(2), 549–567. <https://doi.org/10.5194/os-20-549-2024>
- Vanderborght, E., van Westen, R. M., & Dijkstra, H. A. (2025). Feedback processes causing an amoc collapse in the community earth system model. *Journal of Climate*, 38(19), 5083–5102. <https://doi.org/10.1175/JCLI-D-24-0570.1>
- Weber, S., Drijfhout, S., Abe-Ouchi, A., Crucifix, M., Eby, M., Ganopolski, A., Murakami, S., Otto-Bliesner, B., & Peltier, W. (2006). The modern and glacial overturning

- circulation in the atlantic ocean in pmip coupled model simulations. *Climate of The Past*, 3, 51–64. <https://doi.org/10.5194/cp-3-51-2007>
- Weijer, W., Cheng, W., Drijfhout, S. S., Fedorov, A. V., Hu, A., Jackson, L. C., Liu, W., McDonagh, E. L., Mecking, J. V., & Zhang, J. (2019). Stability of the atlantic meridional overturning circulation: A review and synthesis. *Journal of Geophysical Research: Oceans*, 124(8), 5336–5375. <https://doi.org/https://doi.org/10.1029/2019JC015083>
- Wood, R. A., Rodríguez, J. M., Smith, R. S., Jackson, L. C., & Hawkins, E. (2019). Observable, low-order dynamical controls on thresholds of the atlantic meridional overturning circulation. *Climate Dynamics*, 53(11), 6815–6834. <https://doi.org/10.1007/s00382-019-04956-1>
- Wunderling, N., Donges, J. F., Kurths, J., & Winkelmann, R. (2021). Interacting tipping elements increase risk of climate domino effects under global warming. *Earth System Dynamics*, 12(2), 601–619. <https://doi.org/10.5194/esd-12-601-2021>

Acknowledgments

I would like to thank Dr. Lars Ackermann especially for his invaluable guidance, patience, and support throughout this project.

Chapter 6

Appendix

Use of AI-based applications

I hereby declare that in the creation of this Master's thesis, generative AI systems and AI-assisted software tools were used as aids as follows:

Google Gemini was used selectively to support troubleshooting and fixing programming code bugs within the scope of modeling, as well as for the linguistic revision and refinement of individual sentence formulations.

Grammarly was used for the software-supported checking of English grammar and spelling.

At no point were content-related, conceptual, or scientific personal achievements replaced or generated by these systems. All results, interpretations, and conclusions of the work originate entirely from myself.