Air-sea interactions during glacial Heinrich events

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Abstract

'Heinrich events' - massive iceberg discharges from Northern Hemisphere ice sheets during the last ice age - coincided with cold periods that were followed by abrupt warmings in the Northern Hemisphere. Climate reconstructions suggest that the associated freshwater pulses caused a temporary collapse of the Atlantic Meridional Overturning Circulation (AMOC) by stabilizing the stratification in the regions of deep water formation.

In the present work a coupled atmosphere-ocean-sea ice model is employed under glacial boundary conditions to assess climate feedbacks after a simulated Heinrich event that lead to a fast recovery of the AMOC. Two main mechanisms have been identified. Initially, mixing and thermal processes weaken the stratification in the northern North Atlantic. Additionally, 300-400 years after the main collapse of the AMOC, the stratification is further destabilized by mean horizontal advection of anomalous saline waters within the subpolar gyre. In consequence the large-scale meridional overturning is re-initiated.

The positive salinity anomaly originates from the tropical Atlantic and relies on air-sea coupling. Reduced poleward heat transport in the North Atlantic leads to a cooling north of the thermal equator. Due to advection of cold air and intensification of the northeasterly trade winds the Intertropical Convergence Zone is shifted southward and north equatorial precipitation is reduced. A dilution of the arising positive salinity anomaly is prevented because cross-equatorial oceanic surface flow is halted during the shut-down of the AMOC. Experiments with suppressed tropical air-sea coupling reveal that the recovery time of the AMOC is almost twice as long as in the coupled case. The impact of a shut-down of the AMOC on the Indian and Pacific Oceans can be decomposed into atmospheric and oceanic contributions. Temperature anomalies in the northern hemisphere are largely controlled by atmospheric teleconnections, whereas southern hemispheric ones mainly rely on ocean dynamical changes.

Vertical diffusion is considered a key factor controlling the stability of the AMOC. This however may not be so after a shut-down. Whereas model simulations without air-sea coupling in the tropical Atlantic still show a strong sensitivity to vertical diffusion, this behaviour cannot be found in fully coupled simulations. Thus, after a Heinrich event the formation of a tropical salinity anomaly due to air-sea fluxes appears to be a more efficient negative feedback for the resumption of the AMOC than density homogenisation due to vertical diffusion.

Zusammenfassung

Heinrich Ereignisse - Synonyme massiven Eisexports nordhemisphärischer Eisschilde in den Nordatlantik während des letzten Glazials - stimmten zeitlich mit extremen Kaltphasen überein, denen abrupte Erwärmungen folgten. Klimarekonstruktionen legen nahe, dass die damit einhergehenden Süßwasserpulse einen vorübergehenden Stillstand der Atlantischen Meridionalen Umwälzzirkulation (AMOC) zur Folge hatten, indem sie die Schichtung in den Tiefenwasserbildungsgebieten stabilisierten.
In der vorliegenden Arbeit wird ein gekoppeltes Atmosphären - Ozean - Meereis - Modell unter glazialen Randbedingungen angetrieben, um Klimarückkopplungen nach einem simulierten Heinrich Ereignis zu untersuchen, die zu einer schnellen Erholung der AMOC führen. Zwei hauptsächliche Mechanismen konnten identifiziert werden. Anfänglich schwächen Vermischungs- und thermische Prozesse die Schichtung im nördlichen Nordatlantik. 300 bis 400 Jahre nach nach dem AMOC Kollaps wird die Schichtung dann zusätzlich durch mittlere horizontale Advektion außergewöhnlich salzigen Wassers innerhalb des Subpolarwirbels destabilisiert. Daraufhin setzt die großskalige meridionale Umwälzzirkulation wieder ein.
Die positive Salzgehaltsanomalie entsteht im tropischen Atlantik aufgrund von Ozean-Atmosphären-Kopplung. Verringerter polwärtiger Wärmetransport in den Nordatlantik bewirkt eine Abkühlung nördlich des thermischen Äquators. Durch Advektion kalter Luft und durch Verstärkung der nordöstlichen Passatwinde verschiebt sich die intertropische Konvergenzzone nach Süden, wodurch nordäquatoriale Niederschläge abnehmen. Die Unterbrechung äquatorüberschreitender Oberflächenströmungen infolge des AMOC Zusammenbruchs verhindert eine Abschwächung der Salzgehaltsanomalie. In Experimenten mit unterdrückter Ozean-Atmosphären-Kopplung verdoppelt sich die Erholungszeit der AMOC. Auswirkungen eines AMOC Zusammenbruchs im Pazifik und Indik können in atmosphärische und ozeanische Beiträge unterteilt werden. Während Temperaturanomalien der Südhemisphäre auf dynamischen Änderungen im Ozean beruhen, werden nordhemisphärische Anomalien hauptsächlich von atmosphärischen Telekonnektionen kontrolliert.
Vertikale Diffusion gilt als ein entscheidender stabilitätsbestimmender Faktor für die AMOC. Im Fall einer kollabierten AMOC ist dies aber möglicherweise unzutreffend. Zwar zeigen Modellsimulationen mit enkoppelter Atmosphäre im tropischen Atlantik eine starke Sensitivität gegenüber vertikaler Diffusion, jedoch verschwindet dieses Verhalten in vollständig gekoppelten Simulationen. Folglich könnte die Erzeugung tropischer Salzgehaltsanomalien durch Ozean-Atmosphärenflüsse eine wirksamere negative

Rückkopplung für eine Erholung der AMOC darstellen als Dichtehomogenisierung durch vertikale Diffusion.

Contents

Chapter 1

Introduction

1.1 Background and Motivation

1.1.1 The role of the AMOC in the present climate system

Oceanic waters cover about 71% of the earth's surface. Due to various interactions with the atmosphere the oceans exert a strong influence on the global climate. Compared to the atmosphere the ocean has a long term memory. Due to its huge heat capacity the ocean stores and redistributes heat over many years before it is released to the atmosphere. The large scale redistribution of heat and salt is commonly sketched as the oceanic "conveyor belt" (Broecker, 1991) (Fig. 1.1): Warm and increasingly salty near-surface water is exported from the upper Pacific via Indian Ocean into the Atlantic Ocean. The upper limb of the interhemispheric Atlantic Meridional Overturning Circulation (AMOC) carries these relatively salty waters to the subpolar North Atlantic, where North Atlantic Deep Water (NADW) is formed by cooling induced deep convection. The lower limb of the AMOC transports NADW southward into the Southern Ocean - mixing with the underlying cold Antarctic Bottom Water along its flow path- from where it is carried into the abyssal Indian and Pacific Ocean. Abyssal waters eventually resurface either by large scale upwelling or diapycnal mixing. This global overturning of oceanic water masses is commonly referred to as the thermohaline circulation (THC), because its spatial structure is controlled by surface heat- and freshwater fluxes and large-scale density gradients.

At present, the largest fraction of oceanic northern hemisphere poleward heat transport is carried by the AMOC (Bryden and Imawaki, 2001). The northward heat transport of the AMOC is responsible for the mild western European climate and

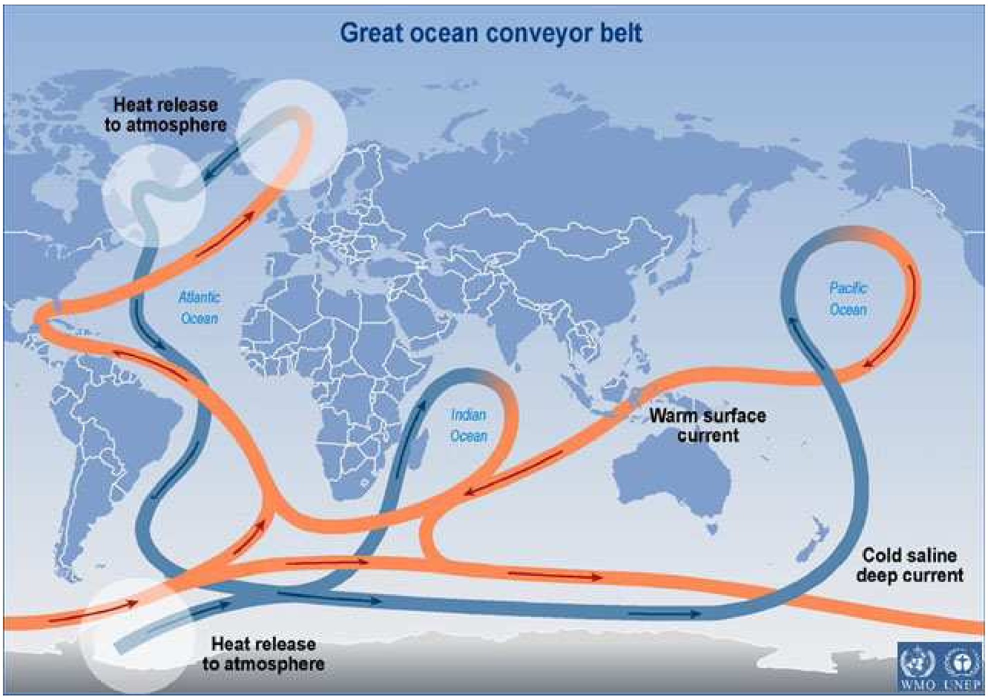

Figure 1.1: Schematics of the global ocean circulation as "conveyor belt" after W. Broecker . The graphics has been obtained from http://www.ipcc.ch

considerably contributes to the total heat exchanged between low and high latitudes. WUNSCH (2005a) roughly estimates the amount of heat transported poleward to be of O(0.9 PW) ($Petawatt = 10^{15}W$) at midlatitudes of the North Atlantic compared to a maximum of about $5PW$ in the atmosphere (TRENBERTH AND CARON (2001)).

1.1.2 The role of the AMOC in a changing climate

In a changing climate, the AMOC is of particular interest, as it appears to be highly sensitive to changes in its thermohaline forcing. This sensitivity relates to the process of NADW formation. Under present-day climate NADW is mainly formed in the Labrador and Greenland Seas after wintertime cooling of the relative salty surface waters. When the surface layer becomes denser than the water below, vertical mixing occurs in localized convective plumes of more than 1000 m depth (MARSHALL AND SCHOTT, 1999). Freshening of North Atlantic surface waters reduces convection depths and North Atlantic Deep Water formation rates and, as model simulations suggest, weakens the meridional overturning (e.g. OTTERA ET AL., 2004). Strong freshwater perturbations in the upper subpolar North Atlantic may even cause a complete breakdown of the AMOC with severe impact on global climate (SCHILLER ET AL., 1997). In response to a freshwater induced AMOC collapse simulated by VELLINGA ET AL. (2002) a temperature decrease of more than $1°C$ is observed over largest parts of the northern hemispheric continents, while several locations on the southern hemisphere exhibit a weak but significant warming.

Most climate simulations predict a weakening of the AMOC during the next century due to rising atmospheric greenhouse gas concentration and resulting warming and freshening of North Atlantic surface waters (IPCC, 2001; GREGORY ET AL., 2005) (Fig. 1.2). However, the rate of AMOC reduction is subject to large uncertainties. To some extent this can be attributed to the model-dependent response of North Atlantic thermohaline forcing to rising greenhouse gas concentration (e.g. atmospheric moisture transport and sea-ice response); but even standardized "waterhosing" experiments, recently compiled by STOUFFER ET AL. (2006) yield very different results in terms of the rate of weakening and the persistence of the weakened or collapsed state. State of the art climate models exhibit a range of 10% to 60% AMOC reduction in response to a freshwater perturbation of 0.1 Sv ($1Sv = 10^6 m^3/s$) applied over 100 years to the northern North Atlantic. For an analogous 1 Sv perturbation experiment, an AMOC collapse is found for all participating climate models, but while some show a rapid reintensification after the termination of the freshwater pertubation, other remain in

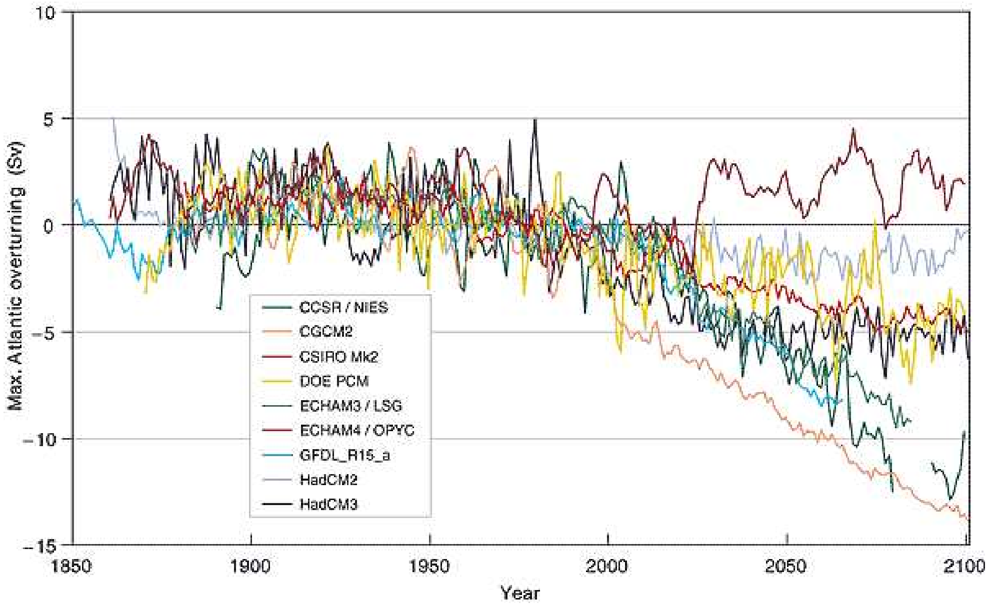

Figure 1.2: Evolution of maximum strength of the Atlantic meridional overturning [Sv] in a range of global warming scenarios. Shown is the annual mean relative to the mean of 1961 - 1990. Past forcing with greenhouse gas and aerosol only, future forcing scenario is the IS92a (IPCC, 2001).

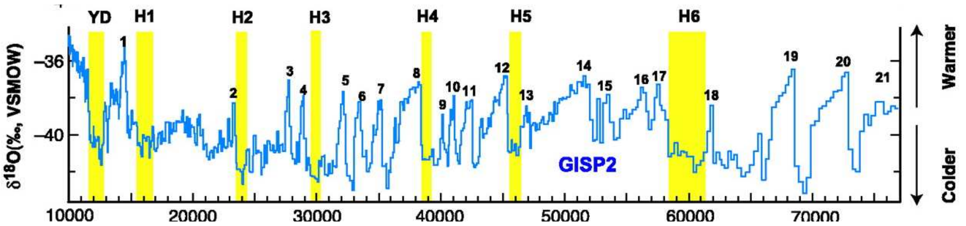

Figure 1.3: $\delta^{18}O$ of Greenland ice (GISP2, 1997). The figure was obtained from Wang et al. (2001). The Younger Dryas event and Heinrich events H1 - H6 are depicted with vertical bars (Bond et al., 1993)

the collapsed state for the remaining 200 years of the experiment. While the 0.1 Sv perturbation is within the range of realistic CO_2-scenarios, the 1.0 SV perturbation rather corresponds to estimates of meltwater releases associated with climate transitions of the last glacial period and the deglaciation (Clarke et al., 2003).

1.1.3 Glacial Heinrich events

Uncertainties in future climate predictions are not least related to the poorly constraint response of the AMOC to changes in North Atlantic freshwater forcing. The investigation of historic climate transitions related to variations of the AMOC might provide additional information about mechanisms and feedbacks controlling the AMOC. The last glacial period is prominent in this respect, as it has been punctuated by a series of abrupt climate changes known as "Dansgaard-Oeschger events" and "Heinrich events" (H1-H6 in Fig. 1.3). These events are well documented in climate reconstructions based on Greenland ice cores (Dansgaard et al., 1993) (Fig. 1.3) and marine sediment cores from the North Atlantic (e.g. Sarnthein et al., 1994; Bard et al., 2000).

In particular Heinrich events are commonly associated with a freshwater induced weakening of the AMOC. They are identified by pronounced layers of ice-rafted detritus (IRD) occuring every 7000 - 10000 years in sediment cores of the North Atlantic between $40°N$ and $55°N$ (the so-called IRD-belt) (Heinrich, 1988; Broecker et al., 1992). These "Heinrich layers" have been attributed to massive surges of ice into the North Atlantic, originating from the adjacent continental ice sheets. In consequence enormous amounts of freshwater must have been released into the North Atlantic, generating global sea-level anomalies in the order of 2-20m (Yokoyama et al., 2001; Sidall et al., 2003). The chronology of Heinrich events is subject to large uncertain-

ties. Particularly radiocarbon dating yields uncertainties of several 100 years due to potential reservoir age problems during drastic ocean circulation changes (Waelbroeck et al., 2001). Reviewing a great number of paleoclimatic archives, Hemming (2004) estimates the duration of the ice surging to be 495 ± 255 years, while the associated flux of meltwater may range between $3x10^4 km^3$ (and 1.6 Sv over one year) to $5x10^6 km^3$ (0.3 Sv over a 500 year interval).
There is evidence from various sites in the Atlantic and Antarctica that support a substantial weakening of the AMOC: Changes of deep water properties (Oppo and Lehmann, 1995; Vidal et al., 1997; Elliot et al., 2002) indicate a fundamental change of circulation in the abyssal Atlantic. Additionally, lowered air temperatures of about $10°C$, as documented in Greenland ice cores (Johnsen et al. (1995)), coincide with negative sea surface temperature (SST) anomalies of several degrees throughout the extratropical North Atlantic (Bond et al., 1993; Maslin et al., 1995; Cortijo et al., 1997; Bard et al., 2000), whereas concurrent warming appears in some locations of the southern hemisphere (Blunier and Brook, 2001; Mix et al., 2001)). This pattern is in agreement with the "bipolar see-saw effect" of reduced northward heat transport due to a weakened AMOC as depicted by Crowley (1992). Paillard and Cortijo (1999) used simple model experiments to show that the documented changes of sea surface temperature and salinity during Heinrich event H4 is predicted to result in complete shutdown of NADW formation.
Some Heinrich events can be associated with remote atmospheric changes such as the weakening of the Asian summer monsoon (Wang et al., 2004; Ivanochko et al., 2005) or meridional shifts of the ITCZ in northeastern South America (Peterson et al., 2000).
Remarkably, the cold episodes observed throughout the Atlantic sector during Heinrich events are terminated after few centuries by an abrupt warming on inter-decadal time scale (Bond et al., 1993). Heinrich events may thus involve a collapsed but unstable AMOC state that recovers on very short timescales. The abruptness of the warming indicates strong negative feedbacks, which, however, have not been identified satisfactorily, yet.

1.1.4 The stability of the AMOC

While a shutdown of the AMOC appears to have been a reversible process during glacial times, model studies yield conflicting results regarding the stability of the present-day climate (Stouffer et al., 2006). Stommel (1961) first highlighted

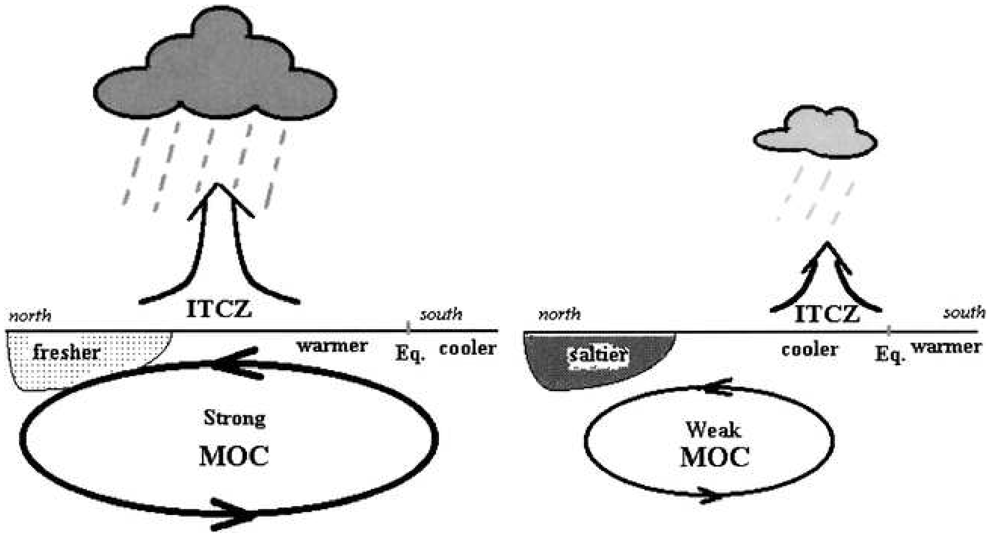

Figure 1.4: Schematic of mechanism responsible for centennial AMOC fluctuation in HadCM3. When the AMOC is (left) strong ITCZ shifts northward, in response to enhanced SST gradient across equator. Fresh anomaly in the upper-ocean propagate northward and weaken the overturning. This results in the (right) weak phase (Vellinga and Wu, 2004)

the possibility of abrupt transitions between multiple (stable) ocean circulation regimes in a box model driven by surface heat and salt fluxes. Stommel (1961) explictly linked the strength of the overturning circulation to the meridional density gradient, which in turn depends on the strength of the circulation. The nonlinear nature of this thermohaline-driven circulation allows for two stable equilibria which resemble off-state and on-state of the AMOC. This simple model provides a valuable approach to understand rapid climate transitions and has inspired many works until today (Greatbatch and Lu, 2003; Wunsch, 2005b). In box models as well as zonally integrated models (Marotzke et al., 1988; Stocker et al., 1992) the thermohaline-driven circulation is considered to be seperated from the wind-driven circulation . However, as thermohaline circulation is intrinsically tied to wind and tidal forcing in many ways (see e.g. Oka et al. (2001); Wunsch (2002); Timmermann and Goosse (2003)), this seperation is problematic and the analogy of the "off-state" in Stommel's box model to a collapsed AMOC is limited.
Furthermore, various air-sea interactions may influence the stability of the circulation regime. A recent study by Yin et al. (2006) demonstrates that an irreversible freshwater induced shutdown of the AMOC in an ocean-only model becomes a reversible process, when air-sea interactions are represented. When the ocean is coupled to a comprehensive atmospheric model a crucial negative feedback leads to a recovery of the AMOC. After the shutdown of the AMOC, the salinity of the upper ocean in the low-latitude North Atlantic increases as a result of reduced freshwater flux from the tropical atmosphere. Surface water with higher salinity is transported northward by the wind-driven circulation and facilitates the resumption of deep convection in the regions of deep water formation. A similar negative feedback of tropical freshwater fluxes is associated with centennial oscillations of the AMOC (Vellinga and Wu, 2004). In a long unperturbed control simulation of a coupled atmosphere-ocean model internal variations of the strength of the AMOC generate cross-equatorial SST gradients, which involve northward/southward shifts of the intertropical convergence zone (ITCZ) for a strong/weak AMOC, respectively. In consequence, an anomalous equatorial precipitation dipole is generated, which results in positive north equatorial salinity anomalies for the case of weak overturning. The positive salinity anomaly propagates to the subpolar North Atlantic at a lag of 5-6 decades, where it enhances NADW formation and accelerates the meridional overturning. The oscillation then enters the opposite phase. The mechanism can be summarized in the schematic picture of Fig. 1.4.

Another important process which influences the stability of the AMOC is vertical mixing. The overturning circulation in the three-dimensional ocean relys on tidal and wind-driven diapycnal mixing to convert cold deep waters into water of the warm water sphere (Sandström, 1908; Munk and Wunsch, 1998). On the global scale the sources of mechanical energy to sustain the mixing are mainly wind forcing and tidal dissipation, whereas geothermal heating, buoyancy forcing and atmospheric pressure loading appear to be of almost negligible importance (Wunsch and Ferrari (2004)). Consistently, several numerical studies confirm that enhanced vertical diffusion (representing subgrid scale vertical mixing) results in stronger and more stable meridional overturning (e.g. Bryan (1987),Wright and Stocker (1992),Marotzke (1997), Zhang et al. (1999)). Stability studies disagree regarding the influence of vertical diffusion on the stability of a collapsed "off-state" of the AMOC. Enhanced vertical diffusion was found to destabilize the off-state in coupled climate models with two-dimensional ocean components (Ganopolski et al. (1998), Schmittner and Weaver (2001)). Prange et al. (2003), by contrast, find vertical diffusivity to be stabilizing the off-state in an OGCM (ocean general circulation model). The authors attribute this to the transport of the horizontal wind-driven circulation, which is not represented in zonally integrated models.
On the other hand, Weber (1998) demonstrated that the sensitivity to vertical diffusivity of ocean only models might considerably change when coupled to an interactive atmosphere model. Since surface fluxes of heat, freshwater and momentum globally change in response to variations of the AMOC strength (e.g. Schiller et al. (1997),Timmermann et al. (2005b),Yin et al. (2006)) the watermass transformation and the stability of the AMOC can be modified by atmospheric feedbacks.

1.1.5 Objective

Given the diversity of feedback processes which regulate the strength of the AMOC, it is not surprising that different ocean models and coupled atmosphere-ocean-sea ice models exhibit different stability characteristics of the AMOC. While many studies investigate processes responsible for a collapse of the AMOC, little work has been done to understand what processes may lead to the recovery of a collapsed AMOC. The relative role of atmospheric interactions and vertical mixing is particular uncertain, as most fundamental studies regarding the stability of AMOC rely on highly simplified

models.
As a result of the previous considerations the ojective of the present work is

- to identify the processes responsible for the fast recovery of the climate system after glacial meltwater events,
- to obtain a better understanding of the evolution and climatic impact of Heinrich events as detected in geological records and to particularly assess the importance of atmospheric feedbacks and
- to investigate relative roles of tropical air-sea coupling and vertical diffusion on the stability of a collapsed AMOC.

1.2 Methods and Design

In order to investigate the climate dynamics associated with glacial freshwater induced shutdowns of the AMOC, model simulations with the earth system model of intermediate complexity ECBilt-Clio (Opseegh et al., 1998; Goosse et al., 1999) are conducted under glacial conditions. The coupled ocean - sea ice model Clio (Goosse et al., 1999) comprises a coarse resolution ocean general circulation model and a thermodynamic-dynamic sea ice model. This is efficient for the investigation of processes associated with changes in deep-ocean circulation on centennial time-scale. The atmospheric component ECBilt (Opseegh et al., 1998) is a three-layer model with a quasi-geostrophic adiabatic core and a partly linearized radiation code. All simulations of Heinrich events utilize a climate state which was equilibriated under Last Glacial Maximum (LGM) boundary conditions consisting of the Peltier (1994) ice sheet topography of about 21,000 years ago, an ice sheet albedo mask, reduced CO2 concentrations (200 ppm), modified orbital forcing, and a LGM vegetation index (Crowley and Baum, 1997). Comparison with a control simulation (CTR) under respective preindustrial boundary conditions reveals that the differences between LGM and CTR in equilibrated North Atlantic sea surface temperatures (SST) are smaller than in recent SST reconstructions (Sarnthein et al., 2003) (Fig. 1.5). A detailed analysis of a equilibrated LGM climate which was obtained in a very similar way is given in Justino et al. (2005).
The structure of this thesis follows the objectives formulated above.
Chapter 2 is aimed at identifying the processes responsible for the observed fast recovery of the climate system after glacial meltwater events. Particular attention is focussed

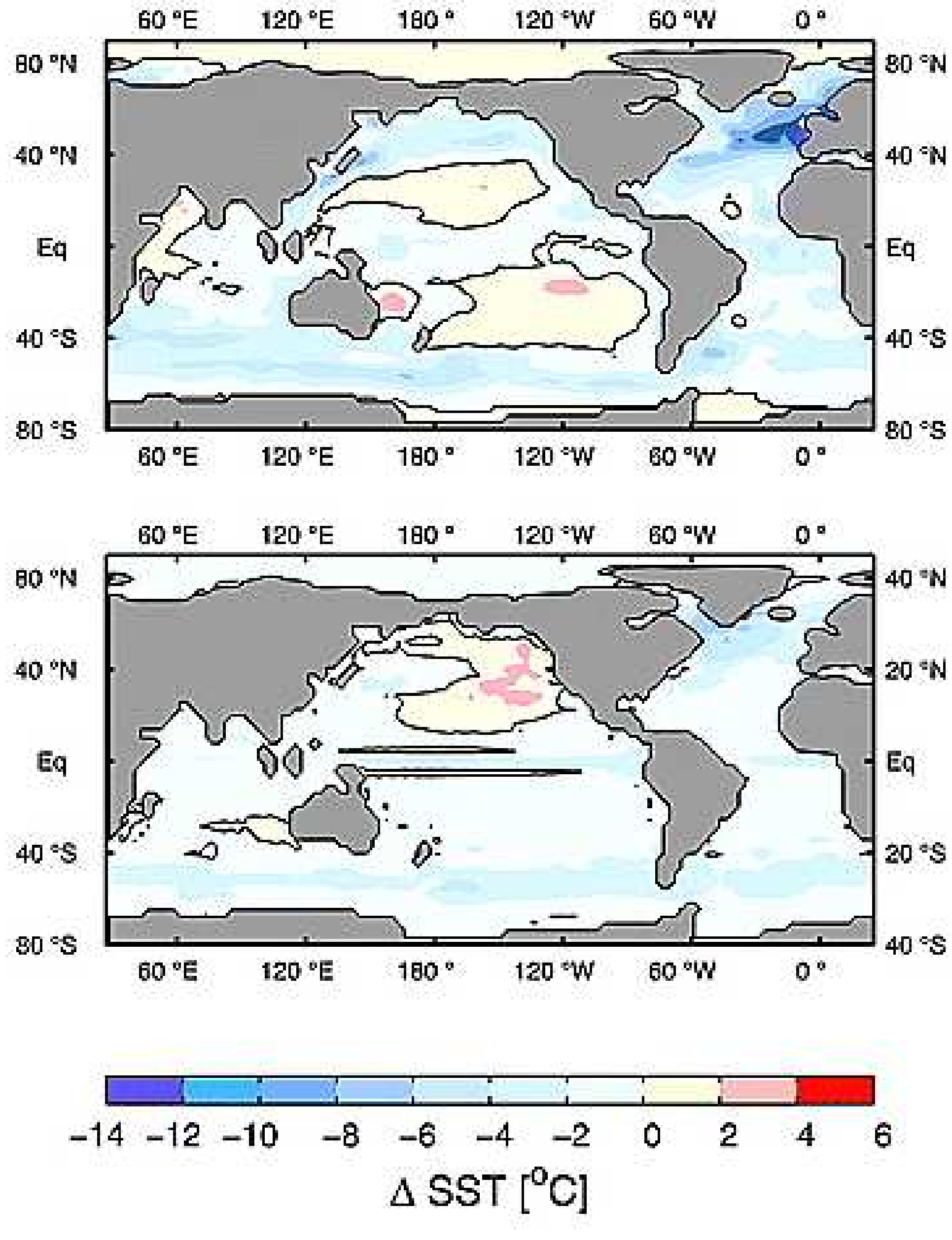

Figure 1.5: Reconstructed and simulated glacial sea surface temperature (SST) anomalies. (top) Difference between reconstructed North Atlantic GLAMAP (Sarnthein et al., 2003) and Pacific and Indian oceans (CLIMAP, 1981) sea surface temperature (time average of February and August) for the Last Glacial Maximum (LGM) and the present-day SST (CLIMAP, 1981). (middle) Sensitivity of the simulated annual mean SST to glacial boundary conditions as quantified by the difference of the equilibrated time-mean SST of the LGM experiment and the preindustrial CTR experiment

upon processes controlling the density stratification in regions of North Atlantic deep water formation.
In chapter 3, the role of atmospheric feedbacks is explored with respect to the evolution and climatic impact of Heinrich events. The atmospheric response to an AMOC collapse is assessed by suppression of air-sea coupling in certain key areas.
Chapter 4 investigates the relative roles of tropical air-sea coupling and vertical diffusion on the stability characteristics of the AMOC.
Conclusions and a comprehensive summary are given in chapter 5.

1.3 Publications

Chapter 2 to 4 are based on manuscripts, which have been submitted as stand-alone papers to refereed scientific journals.

Chapter 2:

Krebs, U. and A. Timmermann (2006), Fast advective recovery of the Atlantic meridional overturning circulation after a Heinrich event, *Paleoceanography*, accepted for publication.

Chapter 3:

Krebs, U. and A. Timmermann (2006), Tropical air-sea interactions accelerate the recovery of the Atlantic Meridional Overturning Circulation after a major shutdown, *Journal of Climate*, submitted.

Chapter 4:

Krebs, U. and A. Timmermann (2006), The relative effects of vertical diffusion and tropical air-sea coupling on the recovery of the AMOC, *Geophysical research Letters*, submitted.

Chapter 2

Fast advective recovery of the Atlantic meridional overturning circulation after a Heinrich event

Abstract Ice-core reconstructions and ocean sediment analysis have revealed that the climate of the last glacial period was highly variable with rapid stadial-interstadial transitions and glacial meltwater pulses (Heinrich events) modulating the climate evolution in the Northern and Southern Hemisphere. Heinrich events had the potential to weaken the Atlantic meridional overturning circulation (AMOC) substantially. Mechanisms which led to the resumption of the AMOC after such events have not been fully disentangled yet. Here a coupled atmosphere-ocean-sea ice model of intermediate complexity is employed to identify important negative climate feedbacks which contribute to a fast recovery of the glacial AMOC. Shortly after the AMOC collapse, thermal processes weaken the stratification in the northern North Atlantic making it more vulnerable to perturbations. Eventually 300-400 years after the main collapse of the AMOC the mean advection of salinity anomalies within the horizontal gyres generates an unstable stratification which will be homogenised through the resumption of convective activity. Eventually isopycnal slopes in the North Atlantic are readjusted, thereby re-initiating the large-scale meridional overturning flow.

2.1 Introduction

The Atlantic meridional overturning circulation (AMOC) plays an important role in transporting heat from the tropics to the northern North Atlantic. The driving mechanisms for this large-scale oceanic circulation cell still remain somewhat elusive. While energy considerations (Huang, 1999) suggest that both wind and tidal mixing are important mechanical energy sources for the AMOC, buoyancy forcing may contribute to the available potential energy and the formation of North Atlantic Deep Water (NADW). Indeed, ocean circulation model studies suggest a high sensitivity of the AMOC to local density perturbations of high-latitude surface waters: an anomalous input of freshwater into the North Atlantic can cause a weakening, and even collapse of the AMOC (Stocker and Wright, 1991; Manabe and Stouffer, 1999; Dong and Sutton, 2002; Knutti et al., 2004). The buoyancy forcing in the North Atlantic has contributions from lateral wind-driven density transports (see e.g. Oka et al. (2001); Timmermann and Goosse (2003)), from density transports provided by the AMOC, and from air-sea fluxes (Stommel, 1961) .
Using a two-box model Stommel (1961) suggested that the nonlinear interaction of lateral density transport and buoyancy forcing gives rise to multiple thermohaline equilibria. The behavior of general circulation models under North Atlantic freshwater forcing has been mapped to this simple paradigm (Manabe and Stouffer, 1999; Rahmstorf and Ganopolski, 1999), although the exact notion of a stable off-state of the AMOC might be problematic in a diffusive limit (see Timmermann and Goosse (2003)). In fact, in the absence of thermohaline *and* wind driven lateral density advection, vertical diffusion plays a key role in destablising the water column by warming the deep ocean gradually. Subsequently, densities in the deep ocean decrease to a point when convective mixing has to readjust the interior stratification. This intense mixing can result in rapid resumptions (flushes) of the AMOC (Winton and Sarachik, 1993; Weaver et al., 1993). In addition to diffusive processes other negative feedbacks may accelerate the resumption of the AMOC. Among them are wind-driven density buoyance transports (Schiller et al., 1997), tropical air-sea coupling (Vellinga et al., 2002; Yin et al., 2006), changes of the atmospheric heat and moisture transports (Nakamura et al., 1994) and sea-ice dynamics (Jayne and Marotzke, 1999).
Given the diversity of feedback processes which regulate the strength of the AMOC, it is not surprising that different ocean models and coupled atmosphere-ocean-sea ice models exhibit different stability characteristics of the AMOC. While models employing 2-dimensional ocean models (Ganopolski et al., 1998) have recovery timescales of

typically more than 1000 years, 3-dimensional coupled general circulation model simulations typically exhibit recovery timescales in the order of a few decades to centuries (e.g. Schiller et al. (1997), Vellinga et al. (2002), Knutti et al. (2004),Stouffer et al. (2006)).

Glacial Heinrich events identified as layers of ice-rafted detritus (IRD) in sediment cores of the North Atlantic (Heinrich, 1988; Broecker et al., 1992) have been attributed to instabilities of the northern hemispheric ice-sheets. Heinrich events released substantial amounts of freshwater into the North Atlantic, generating global sea-level anomalies in the order of 2-20m (Yokoyama et al., 2001; Sidall et al., 2003). Low temperatures and subsequent abrupt warmings in the North Atlantic (Broecker, 1994; Bond et al., 1993), interhemispheric temperature signals (Blunier and Brook, 2001) and substantial changes of deep water properties (Oppo and Lehmann, 1995; Vidal et al., 1997; Elliot et al., 2002) support the notion of a substantial freshwater-induced weakening and subsequent recovery of the AMOC.

Large efforts have been made to obtain an accurate chronology of Heinrich events. While ice core data show stadial/interstadial transitions associated with the AMOC recovery after Heinrich events on time scales of a few centuries, the correlation of these transitions to the IRD-layers of North Atlantic sediment cores is problematic. Radiocarbon dating, commonly used to derive calendar dates for marine sediments, proves to be particular inaccurate during periods of large scale circulation changes such as Heinrich events, resulting in dating uncertainties of several centuries (see Waelbroeck et al. (2001)). Studies of chemical characteristics of the deep Atlantic ocean succeeded in linking the abrupt changes in the Greenland ice cores with millenial AMOC variations, but do not allow conclusions about decadal to centennial variability (Boyle (2000)). Moreover, low sedimentation rates in the IRD-belt yield large uncertainties in the duration of typical glacial meltwater pulses: Synthesizing published estimates for the duration of the meltwater events related to Heinrich events H1 and H2 (ranging from 208 to 2280 years) Hemming (2004) proposed a typical duration of 495±255 years. In order to gain a better understanding of the chronology of Heinrich events, it is necessary to investigate the recovery mechanisms for a collapsed, glacial AMOC state. In fact there may be several different recovery scenarios for a Heinrich event, depending on the duration of the anomalous meltwater discharge: one for which the recovery is happening (e.g. due to diffusive processes in the deep ocean, like in Winton and Sarachik (1993)) while anomalous freshwater forcing continues to perturb high latitude salinities; another one for which the resumption is triggered (e.g. due to advective processes) during a phase when the anomalous freshwater forcing is absent. Our study explores the latter scenario. Here an attempt is made to understand glacial recovery mechanisms of the AMOC from a freshwater-induced shutdown by studying the density flux budget in the northern North Atlantic. The spatio-temporal signatures of the anomalous freshwater forcing are chosen such as to mimic a typical glacial meltwater pulse. The paper is organised as follows: In section 2 the model of intermediate complexity used in this study is described. Section 3 explains the experimental design chosen here and studies the climate response to the freshwater perturbation. In section 4 the mechanisms are disentangled which lead to a resumption of deep ocean convection after the AMOC collapse. The main results are summarised and discussed in section 5.

2.2 The Model

Our study is based on multi-century long simulations conducted with the three-dimensional atmosphere-sea ice-ocean model ECBilt-Clio. The atmospheric component is version2 of ECBilt (Opseegh et al., 1998), a spectral T21, three-level, based on quasi-geostrophic equations extended by estimates of the neglected ageostrophic terms in order to close the equations at the equator. The model contains a full hydrological cycle which is closed over land by a bucket model for soil moisture. Synoptic variability associated with weather patterns is explicitly computed. Diabatic heating due to radiative fluxes, the release of latent heat and the exchange of sensible heat with the surface are parametrised and cloudiness is prescribed.

The sea ice-ocean component Clio (Goosse et al., 1999; Goosse and Fichefet, 1999; Campin and Goosse, 1999) consists of a free-surface primitive equation model with 3°x3° resolution coupled to a thermodynamic-dynamic sea ice model. To avoid a singularity at the North Pole the oceanic component makes use of two subgrids: The first one is based on classic longitude and latitude coordinates and covers the whole ocean except the North Atlantic and Arctic. These are covered by the second spherical subgrid, which is rotated and has its poles at the equator in the Pacific (111° W) and Indian Ocean (69° E). In this work all analyses are conducted on these original grids. For the simulations conduced here we also use an implicit diffusion convective adjustment scheme (like in Marotzke (1991) or Hirst and Cai (1994)), which increases the vertical diffusivity whenever the density profile is unstable. Experiments with a complete explicit scheme yield very similar results.

Under present-day conditions, ECbilt-Clio exhibits a systematic underestimation in the

atmospheric moisture transport from the Atlantic to the Pacific. This can be partly attributed to the relatively weak trade winds in the tropical Atlantic. As a consequence, the northern North Atlantic and Arctic ocean are too fresh with implications for the AMOC and the Arctic snow-sea-ice pack. To compensate this model bias, a freshwater flux-adjustment is introduced which removes freshwater from the North Atlantic and dumps excess water into the Pacific Ocean, where the simulated precipitation is generally too weak. We employ the same adjustment for our glacial boundary conditions.
For a 5000 year glacial spin-up and the subsequent simulation of a glacial Heinrich event we use the glacial ice sheet topography of Peltier (1994), a vegetation index representing the Last Glacial Maximum (LGM) (Crowley and Baum, 1997), a corresponding ice albedo and reduced atmospheric CO_2 concentrations (200 ppm). Glacial ice-sheet effects are included only as topographic and diabatic forcings. The ice-sheet is not directly coupled to the ECBilt, nor is the sea-level in our ocean model adjusted to glacial levels. With these settings we obtain the LGM scenario which serves as initial condition for all further experiments.
Similar LGM set-ups for the ECBilt-Clio model were used in Justino et al. (2005); Timmermann et al. (2004, 2005a,b).

2.3 Simulation of glacial Heinrich events

During Heinrich events the melting of icebergs in the region of the IRD-belt caused an anomalous freshwater input into the north Atlantic between 40°N and 60°N (Ruddiman, 1977). Both duration and strength of the freshwater perturbation can hardly be assessed directly and are subject to large uncertainties. Estimates for the equivalent sea level rise provide values from 2m (Roche et al., 2004) up to 20 m ((Yokoyama et al., 2001), (Sidall et al., 2003)) for different Heinrich events. Using information from various sediment cores, Hemming (2004) specifies the duration of the meltwater events to 500 ± 250 years while model calculations (Roche et al., 2004) yield values of 250 ± 150 years. In our experiments Heinrich events are simulated by applying a pulse of additional freshwater flux of 1.3 Sv amplitude (see Figure 1), evenly distributed over the North Atlantic between 40°N and 55°N (see Figure 2, red lines), which corresponds to an integrated global sea level rise of about 8 m. Our pulse has a duration of 100 year only, which is significantly shorter than estimated from a comparison between a 2-dimensional climate model and paleo-data (Roche et al., 2004). It should be noted here that the model runs were performed without global-freshwater compensation and use freshwater fluxes, rather than virtual salt fluxes. The freshwater input and the advection of relatively fresh water in the subpolar gyre cause a considerable decline of northern North Atlantic sea surface salinities, thereby stabilising the water column in the convection regions and terminating deep convection (see Figure 3). Consequently North Atlantic Deep Water (NADW) formation, as measured by the transport at 20°S ceases and the AMOC collapses 50 years after the freshwater perturbation reached its maximum value (Figure 1 a). As shown in Knutti et al. (2004) and Timmermann et al. (2005b) the collapse of the AMOC in the ECBilt-Clio model has a great impact on Atlantic sea surface temperatures and velocities. While the North Atlantic exhibits a cooling of up to 10°C, nearly the whole southern hemisphere experiences a weak warming due to the respective changes of the meridional heat transport. This is in agreement with recent ice-core reconstructions from the Northern and Southern Hemisphere (e.g. (Blunier and Brook, 2001)) and previous Coupled General Circulation Model (CGCM) simulations (e.g. (Thorpe et al., 2001; Vellinga and Wu, 2004)). Due to colder sea surface temperatures (SST) and reduced sea surface salinities (SSS) large parts of the North Atlantic deep water formation regions are covered with sea-ice during the AMOC collapse (Figure 2), which, in addition to the already existing stable stratification, prevents further deep water formation.

2.4 Recovery of the AMOC

Horizontal and vertical density gradients in the North Atlantic are strongly modified by deep convection in the northern North Atlantic. These density gradients are partly responsible for driving the large-scale meridional overturning circulation due to geostrophic and frictional balances. Deep convection occurs when the surface layer becomes denser than the water below, and results in highly variable mixing depth as a function of space and time (Marshall and Schott (1999)). In hydrostatic ocean models this process is commonly represented by "convective adjustment" which is usually restricted to just a few grid points. This crude parametrisation does not capture the process of deep water formation in detail. In consequence its local characteristics can be highly model dependent. In the following we make an attempt to understand the AMOC recovery by assessing only spatially integrated and annually averaged physical properties in the convection region, which could be of particular interest for model intercomparison projects such as Stouffer et al. (2006). Figure 4 shows the temporal evolution of temperatures, salinities and densities in the subpolar North Atlantic area

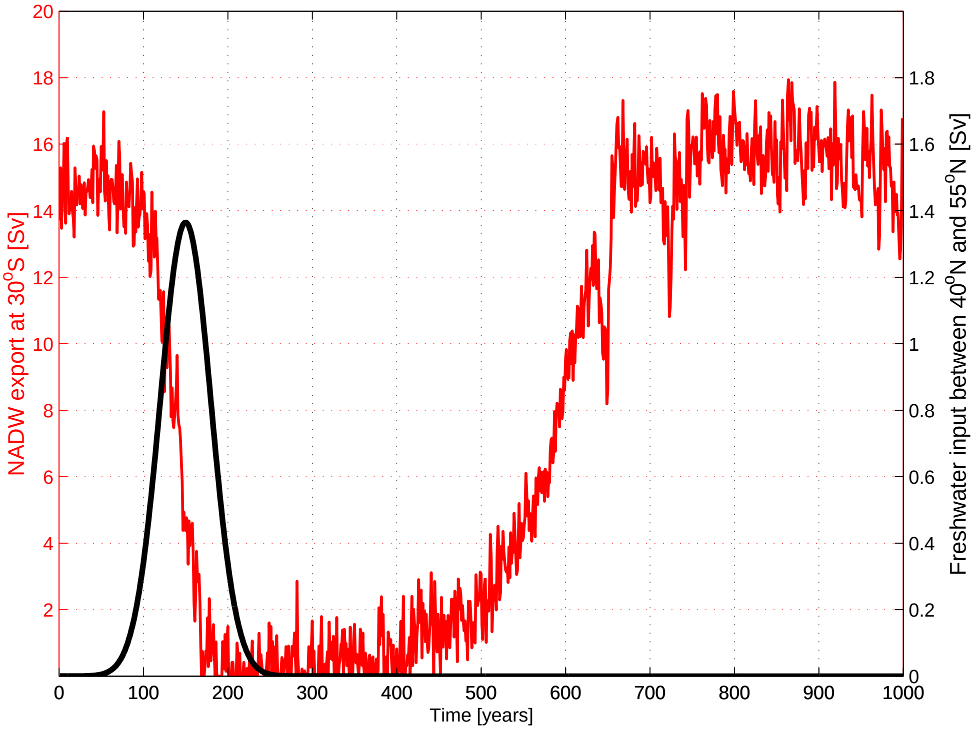

Figure 2.1: a: Annual mean of southward transport below 1000m depth across 30°S [Sv] (red) and freshwater perturbation [Sv] (black).

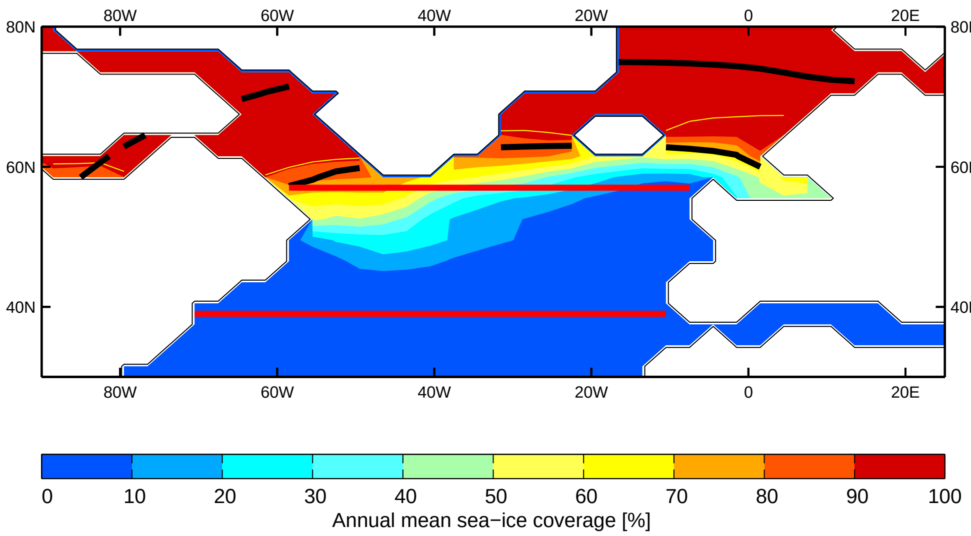

Figure 2.2: Mean sea ice coverage (%) for years 200-250. Yellow lines indicate the margin of 90% mean sea ice coverage for years 400-450. Red lines indicate the region of the freshwater perturbation, black lines the North Atlantic sinking region under consideration

averaged between 60°N and 80°N of the rotated subgrid (see Figure 2, black line, for region of analysis), in the following referred to as the North Atlantic sinking regions. Around model year 150 a cold water anomaly develops in the upper 500m due to reduced heat transport from the south. This is clearly illustrated in Figure 5, which shows the zonally averaged meridional velocity averaged meridionally over the North Atlantic sinking regions. In response to the AMOC collapse as well as to a weakening of the wind-driven transports due to increased sea-ice coverage, the poleward upper ocean transport almost ceases. In contrast, water at intermediate depth maintains relative warm temperatures in the absence of deep convection. After model year 100 temperatures even increase at depths of about 400m, which can be explained in terms of a temporary change of sign in the Denmark Strait overflow. This leaves colder water on top of warmer (and saltier) water. Due to the reduced flow of the AMOC and the reduced efficiency of the wind-driven circulation mostly vertical small-scale processes and vertical heat diffusion operate in the North Atlantic sinking regions to homogenize this vertical temperature gradient (years 350-400 in Figure 4).

During this thermal homogenisation process which ends around model years 400 the heat of the relatively warm watermasses at intermediate depth is partly released to

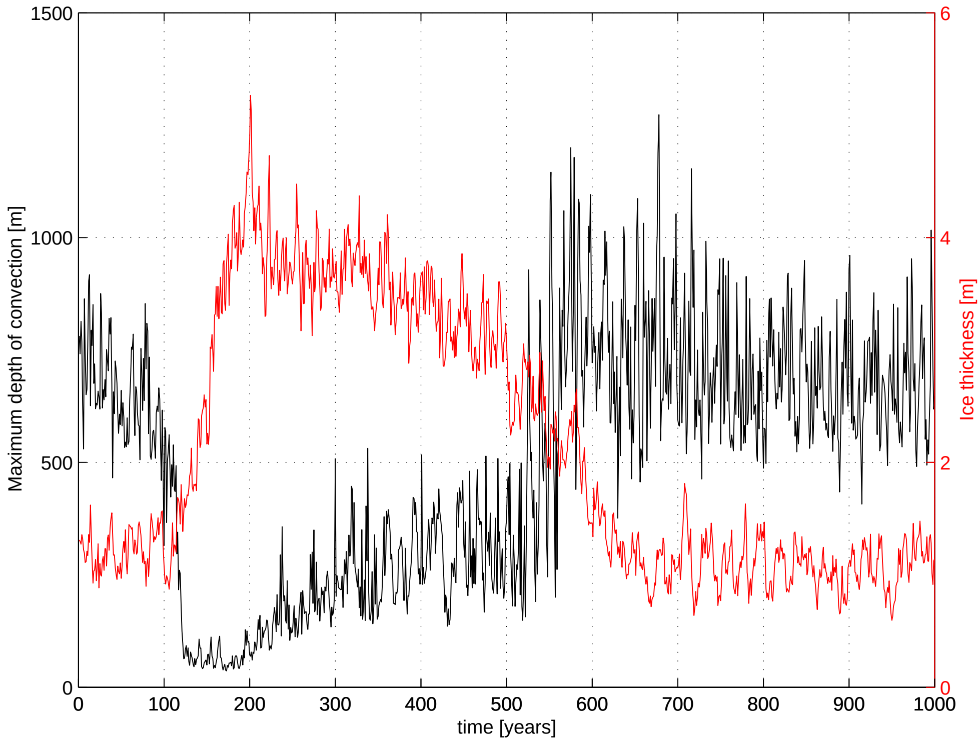

Figure 2.3: Annual mean of maximum depth of convection [m] in the North Atlantic sinking regions (black) and annual mean of ice thickness [m] averaged over the North Atlantic sinking regions (red)

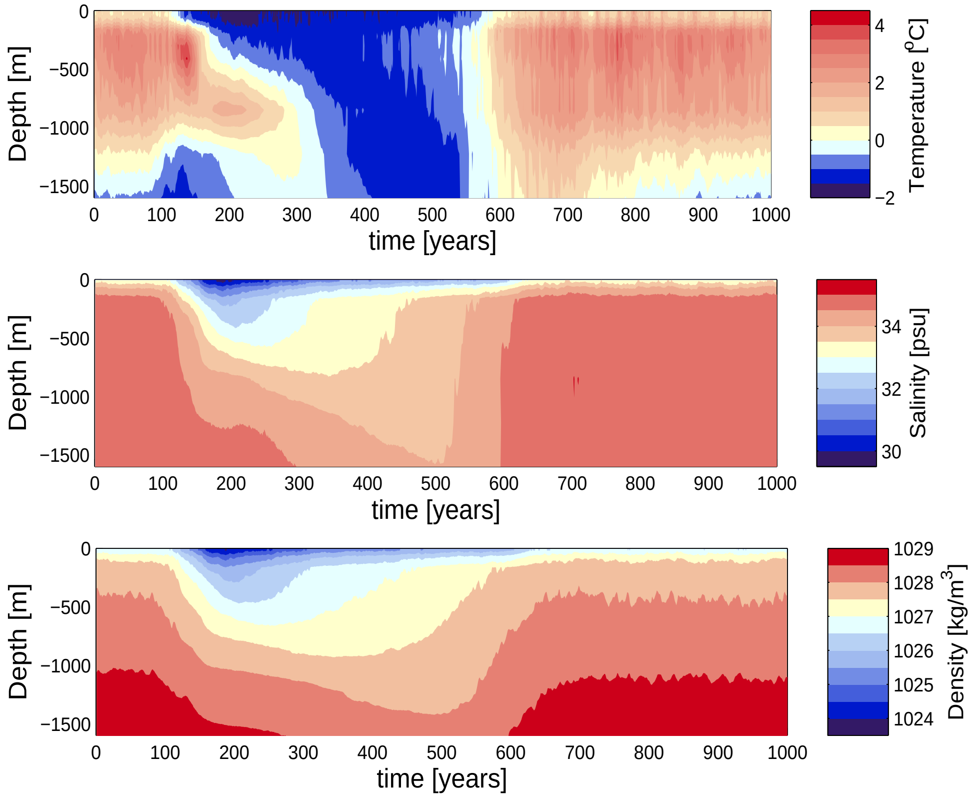

Figure 2.4: Hovmoeller diagram of anual mean potential temperatures [°C] (upper panel), salinities [psu] (middle panel), and potential densities [kg/m^3] (lower panel) averaged between 60°N and 80°N

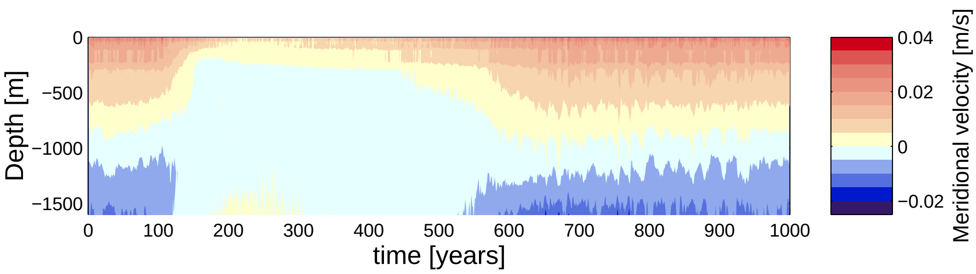

Figure 2.5: Hovmoeller diagram of anual mean meridional zonally averaged velocity [m/s], averaged meridionally between 60°N and 80°N

the surface and reduces the ice thickness (Figure 2). Sea-ice particularly reduces in the southern Norwegian Sea, leaving it ice covered only intermittently (Figure 2, yellow line). The associated buoyancy loss to the sea ice cover results in shallow convection (years 300-500 in Figure 3). However, this convection cannot be directly associated with North Atlantic deep water formation and remains shallow, as long as sea-ice insulates the ocean from the cold polar atmosphere. Thick sea-ice reduces the oceanic heat loss by more than 90% compared to an otherwise open ocean (not shown). Due to the sea-ice reduction, the net-surface freshwater flux into the ocean increases. Furthermore, during model years 300-500 the whole North Atlantic sinking region above 700 m is getting more saline. This feature can be attributed to two processes. While in the first stages of the collapse (years 0-150) the northward upper ocean flow (see Figure 5) transports part of the perturbation freshwater lens (which originated from the forcing at 40°N-55°N) into the Nordic seas, this freshening is significantly reduced during the stages of very weak meridional flow (years 200-400). Eventually, the poleward advection of anomalously salty water originating from the tropical North Atlantic (not shown) leads to a weakening of the halocline and re-establishment of the original vertical salinity gradients.

In order to quantify which processes contribute to the resumption of deep-ocean convection in the Nordic Seas, we consider processes potentially capable of destabilizing the stratification in the North Atlantic sinking region. In particular we analyse the density flow into the top 187m of this region due to horizontal advection (D_m) across its lateral boundaries, vertical advection (D_v) across its lower boundary and fluxes through its surface (D_S), all subdivided into thermal and haline contributions.

It should be noted here that our density budget is not entirely closed, because vertical density fluxes due to convective adjustment, vertical diffusion and horizontal density fluxes due to horizontal diffusion and subannual fluctuations of T,S and v are not captured in our analysis.

To compute changes of mass in the sinking area (depth interval $[-187m, 0m]$) due to meridional advection of the perturbation densities we introduce a time-dependent and depth-independent reference temperature $T_0(t)$ and reference salinity $S_0(t)$. The appropriate choice of T_0, S_0 are the respective volume averaged means in the sinking region, as discussed by Weijer et al. (1999). The buoyancy changes due to density flow across latitude y are then given by

$$D_m(y,t) = \int_{0m}^{-187m} \int_{east}^{west} v\,[-\alpha(T - T_0(t)) + \beta(S - S_0(t))]dxdz$$

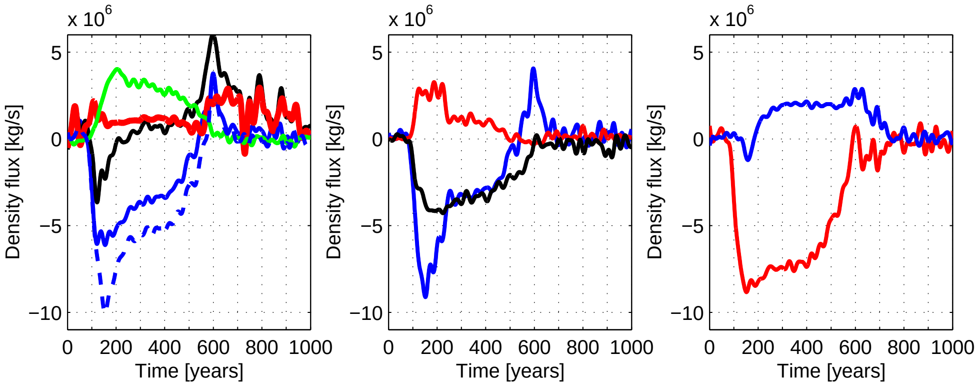

Figure 2.6: Spatially integrated density fluxes into the top 187m of the North Atlantic sinking regions (30 year low pass filtered) [kg/m^3]: *left panel*: horizontal haline component (solid blue), horizontal haline component through southern boundary (dashed blue), sum of horizontal thermal and surface heat flux component (red), surface freshwater flux component (green), and sum of surface density fluxes and density fluxes due to horizontal advections (black); *middle panel*: density fluxes due to salinity advection from the south: mean circulation anomalous salinity component $S^* < v >$ (blue), anomalous circulation mean salinity component $< S > v^*$ (black), and anomalous circulation anomalous salinity component $S^* v^*$ (red); *right panel*: density fluxes due to salinity advection from the south: gyre component originating from $S'v'$ (red) and overturning component due to $\overline{S}\overline{v}$ (blue).

where α, and β are the (T and S dependent) thermal and haline expansion coefficients for water respectively. Accordingly, the buoyancy changes due to vertical density flow at depth $-187m$ are given by

$$D_v(t) = \int_{60^\circ N}^{80^\circ N} \int_{east}^{west} w\,[-\alpha(T - T_0(t)) + \beta(S - S_0(t))]dxdy.$$

The spatially integrated mass flux through the surface ((Stern, 1975)) can be obtained from

$$D_s(t) = \int_{east}^{west} dx \int_{60^\circ N}^{80^\circ N} dy \left[\alpha \frac{Q_{net}}{c_p \rho} + \beta S \frac{E - P}{1 - S/1000}\right]$$

where Q_{net} is the net heat flux out of the ocean, $E - P$ evaporation-precipitation and c_p is the specific heat capacity of water. Here, S is the sea surface salinity in psu.
Figure 6 (left panel) shows the haline and thermal contributions to the density changes in the sinking region via surface buoyance forcing ($D_s(t)$) and due to the meridional

density flux divergence ($D_m(60°N, z, t) - D_m(80°N, z, t)$). Anomalies in our density budget analysis are computed with respect to the first (unpertubed) 20 years of the simulation. If the temporal derivatives of the density fluxes are positive/negative they tend to destabilise/stabilise the water column in the sinking regions and hence accelerate/delay the recovery of the AMOC via convective resumption.

Lateral and surface heat fluxes are in close balance, not only for the first decades of the experiment, but also during the collapsed phase and after the recovery of the AMOC. During the "on-state" both, lateral and surface heat fluxes take values of $10^7 kg/s$ and reduce to less than $5 \times 10^5 kg/s$ during the collapsed state (not shown). As surface heat flux and lateral temperature advection are strongly connected, we only show the sum of the respective density fluxes (Figure 6, left panel, red curve). During the "on-states" the sum of both fluxes shows greater variability, but no tendency to increase or decrease the given stratification. This indicates that, even though the meridional heat transport and the associated release of heat to the atmosphere are essential features in the formation of deep water, they do not control the recovery time scale of the collapsed AMOC. Figure 6 (left, red curve) shows that the balance between surface and meridional thermal fluxes remains almost unchanged throughout the experiment.

The haline surface density flux (green) is dominated by the freshwater flux related to the formation and melting of sea ice (not shown)[1]. Between model years 100 to 200 the ice cover is considerably increased and shields the ocean from the atmospheric freshwater fluxes. In the subsequent years the sea-ice cover reduces again due to the local temperature homogenisation processes discussed above. The associated freshwater flux anomalies delay the recovery of the AMOC. The haline component of the meridional density flux divergence (solid blue) is dominated by salinity advection across the southern boundary of the convection area (dashed blue). During the initial shutdown phase, which is triggered by the poleward advection of the subpolar freshwater lens, this component is reduced by $8 \times 10^6 kg/s$ within a few decades. Lateral haline density fluxes increase after year 200 and a strong rise of $1 \times 10^7 kg/s$ can be observed between years 400 and 600, supporting idea that deep water formation resumes due to the extraction of buoyancy from the upper ocean.

The buoyancy change due to vertical density advection (not shown) takes values of $O(10^4 kg/s)$ and does not contribute significantly to the recovery of the AMOC.

Our consideration of density fluxes into the deep water formation region reveals that

[1]Note that the freshwater input is distributed between 40°N and 55°N, whereas the density flux analysis focuses on the area north of this.

advection of saline water from lower latitudes is the dominant process controlling the recovery of the AMOC. Therefore we further analyse the salinity advection across the southern boundary of the convection area. First we split meridional velocity $v(t)$ and salinity $S(t) - S_0$ into a temporal mean $<>$ (the mean of the first 20 years of the experiment) representing the AMOC state prior to the collapse and its deviation *, representing the anomalies with respect to the uncollapsed LGM background climate:

$$\zeta(x, y, z, t) =< \zeta(x, y, z) > +\zeta(x, y, z, t)^*.$$

Furthermore, the lateral density fluxes can be decomposed into a zonal mean (overturning component) and the deviation from the zonal mean (gyre component)[2] (Delworth et al., 1993). This is abbreviated as

$$\zeta(x, y, z, t) = \overline{\zeta}(y, z, t) + \zeta(x, y, z, t)'$$

In our case ζ is chosen to represent v and $S - S_0(t)$ for $y = 60°N$ and the depth interval $z = [-187m, 0m]$, respectively. S_0 is again the time dependent basin mean salinity. Before the freshwater perturbation reaches the convection area the salinity difference between $60°N$ and the basin mean is mostly positive where flow is northward and negative where flow is southward (not shown). Figure 6, middle panel illustrates that the main contribution to the positive tendencies of the density flux after year 200 originates from the time mean transport of anomalous salinities $S - S_0(t)$ into the convection area from the south. In contrast, the anomalous transport (southward as compared to the LGM state) of anomalous (fresh) salinities at first counteracts the AMOC collapse, but with rising salinities counteracts the recovery. The anomalous transport of mean salinity accelerates the AMOC recovery somewhat, but is not one of the main driving forces because its positive tendencies after year 400 are a direct consequence of the recovering circulation. In Figure 6, right panel we observe that after year 400 the recovery feedback is provided by the gyre component while the contribution of the overturning component is of minor importance. After year 620, when the AMOC transport is restored to 80% of its pre-collapse values, the overturning component counteracts the recovery. This can be associated with the influence of the North Atlantic Current which increases the basin mean salinity S_0 faster than the zonal mean salinity at $60°N$, so that $\int_{west}^{east}[S - S_0(t)]dx$ becomes negative. Likewise, the positive tendency in the overturning component between years 160 and 300 can be

[2] This type of separation is not always physically meaningful.

associated with the salinity difference becoming positive when low salinities accumulate in the arctic basin.
As will be shown in a forthcoming study in more detail the salinity anomalies are generated in the tropical North Atlantic via positive coupled air-sea interactions. The mean wind-driven circulation transports these anomalies to the North Atlantic, where they trigger the recovery of the AMOC, once the subsurface temperature preconditioning has established a favourable stratification.

2.5 Summary and discussion

This study explored the recovery mechanism of the AMOC under glacial conditions using the earth system model of intermediate complexity ECBilt-Clio. A freshwater perturbation experiment has been performed and it was shown that the relatively fast recovery of the AMOC is due to two important processes:

- During the collapsed overturning stage the transport of warm upper ocean water into the sinking regions is strongly reduced. Relatively warm water at intermediate depths of about 800-1000m mixes with the cold water above. The associated upper ocean heat-flux is used to reduce the sea-ice thickness and coverage. Overall, in the absence of meridional transports, mixing and heat diffusion homogenise the vertical temperature gradients in the water column during years 200-400, thereby pre-conditioning the system for buoyancy instabilities.

- during the collapsed AMOC state, the mean wind-driven gyre circulation transports positive surface salinity anomalies into the nordic seas and the convection regions.

  This leads to a destabilisation of the water column and a resumption of deep ocean convection. Eventually the reorganised vertical density profiles lead to a readjustment of the entire ocean circulation via Kelvin and Rossby wave adjustment (Kawase, 1987; Huang et al., 2000). As has been discussed in Timmermann et al. (2005b) and as will be shown in a more detailed forthcoming study these positive salinity anomalies originate from the positive air-sea interactions and in particular the change of the strength and position of the ITCZ in the tropical North Atlantic.

The tropical origin of the salinity anomalies is in agreement with an AMOC stabilizing feedback of tropical freshwater fluxes in accordace with the results of Vellinga and

WU (2004). Low latitude air-sea interactions might considerably influence the recovery timescale of a collapsed AMOC and differences in ocean-atmosphere coupling might be responsible for the strong model sensitivity of freshwater pertubation experiments (e.g. STOUFFER ET AL. (2006)). The recovery time of the AMOC depends not only on the processes discussed here, but also on the length and strength of the anomalous freshwater forcing (TIMMERMANN ET AL., 2003). This raises the question as to whether the duration of Heinrich events was in the order of hundreds or thousands of years and whether the transitions from stadial (cold) to interstadial (warm) conditions under glacial boundary conditions were triggered by anomalous salinity advection or rather by diffusive processes. Apparently the transition from stadials to interstadials took place within a decade or so (SEVERINGHAUS ET AL., 1998). Climatically, interstadials were similar to present-day interglacial conditions. Virulent climatic reorganisations associated with stadial-interstadial transitions might have been triggered by the release of accumulated heat in the deep ocean to the surface (ADKINS ET AL., 2005), changes of sea-ice cover and the atmospheric circulation. Our study suggests that horizontal salinity advection may have played a significant role for shorter centennial-scale meltwater pulses. The role of horizontal diffusion could not be assessed in this study and should be subject of future research.

Other simulations conducted with the ECBilt-Clio model (not shown here) have revealed that the fast advective recovery mechanism does not only play a key role under glacial but also under present-day conditions. Relative to glacial conditions the recovery of the AMOC is delayed a few decades under present-day conditions. However, compared to the CLIMBER model simulation (GANOPOLSKI ET AL., 1998) which exhibits a strong change of the hysteresis under glacial conditions relative to present-day climate, our fully coupled 3-d atmosphere-ocean-sea ice model exhibits only a relatively weak sensitivity to the glacial boundary-conditions, even for very slow freshwater forcing changes.

Hence, the robustness of the proposed mechanism should be further explored by similar analysis of freshwater perturbation or water-hosing experiments using different coupled general circulation models. Such experiments have been conducted as part of the CMIP model data comparison. An in depth analysis of these experiments will help to understand the crucial negative feedbacks which lead to the destabilisation of the AMOC after a Heinrich event and may help to assess the sensitivity of the AMOC to future climate change.

Chapter 3

Tropical air-sea interactions accelerate the recovery of the Atlantic Meridional Overturning Circulation after a major shutdown

Abstract Using a coupled ocean sea-ice atmosphere model of intermediate complexity we study the influence of air-sea interactions on the stability of the Atlantic Meridional Overturning Circulation (AMOC). Mimicking glacial Heinrich events, we trigger a complete shut-down of the AMOC by delivery of anomalous freshwater forcing to the northern North Atlantic. Analyzing fully and partially coupled freshwater perturbation experiments under glacial conditions it is shown that associated changes of the heat transport in the North Atlantic lead to a cooling north of the thermal equator and an associated strengthening of the northeasterly trade winds. Due to advection of cold-air and an intensidication of the trade winds the Intertropical Convergence Zone (ITCZ) is shifted southward. Changes of the accumulated precipitation lead to generation of a positive salinity anomaly in the northern tropical Atlantic and a negative anomaly in the southern tropical Atlantic. During the shut-down phase of the AMOC, cross-equatorial oceanic surface flow is halted, preventing a dilution of the positive salinity anomaly in the North Atlantic. Advected northward by the wind driven ocean

circulation the positive salinity anomaly increases the upper ocean density in the deep water formation regions, thereby accelerating the recovery of the AMOC considerably. Partially coupled experiments which neglect tropical air-sea coupling reveal that the recovery time of the AMOC is almost twice as long as in the fully coupled case.

The impact of a shut-down of the AMOC on the Indian and Pacific ocean can be decomposed into atmospheric and oceanic contributions. Temperature anomalies in the northern hemisphere are largely controlled by atmospheric circulation anomalies, whereas those in the southern hemisphere are strongly determined by ocean dynamical changes. An intensification of the Pacific meridional overturning cell in the northern North Pacific during the AMOC shut-down can be explained in terms of wind-driven ocean circulation changes acting in concert with global ocean adjustment processes.

3.1 Introduction

The AMOC carries an enormous amount of heat northward (Ganachaud and Wunsch, 2000), thereby altering climates in northern Europe substantially. The thermal energy transported poleward exceeds by far the mechanical energy provided by wind and tidal forcing which is required to maintain this type of circulation (Huang, 1999). The AMOC is known to exhibit a strong sensitivity to thermohaline perturbations. Numerical studies reveal that a strong anomalous input of freshwater into the northern North Atlantic is likely to cause a shutdown of the AMOC (Stocker and Wright, 1991; Manabe and Stouffer, 1999; Dong and Sutton, 2002; Knutti et al., 2004; Timmermann et al., 2005a). Deep convection in the northern North Atlantic can be halted as a result of increased freshwater forcing. An associated cessation of North Atlantic Deep Water (NADW) formation may lead to a reduction of the large-scale overturning circulation and of the associated poleward heat transport. This will lead to a northern hemispheric cooling of up to $-6°$C (Vellinga and Wood (2002)) and eventually to a reorganization of the global ocean circulation (Goodman (2001); Huang et al. (2000); Cessi et al. (2004)) via wave-adjustment and advective processes.

Our understanding of the processes involved in driving the AMOC is still in its infancy. Model simulations of the AMOC sensitivity often neglect mixing energy constraints (Nilsson et al., 2003), tropical air-sea coupling and non-hydrostatic processes. Moreover, the response of coupled atmosphere-ocean models to a prescribed freshwater forcing is highly model-dependent (Rahmstorf et al., 2005; Stouffer et al., 2006), suggesting that the amplitude of relevant feedbacks is not very well constrained.

Present estimates of potential longterm trends of the observed AMOC strength Bryden et al. (2005); Kanzow et al. (2006) are inconclusive because of the sparseness of the available ocean observations and a lack of temporal statistics.

Indirect proxy evidence for the sensitivity of the AMOC stems from climate reconstructions of the last glacial period. As evidenced in marine sediment cores (Heinrich, 1988; Broecker, 1994; Bond et al., 1993) layers of largely enhanced concentrations of ice-rafted debris (IRD) in the North Atlantic occurred together with widespread sea surface cooling. These so-called Heinrich events occurred about every 6,000-10,000 years and are often interpreted as the result of major glacial ice-sheet/ice-shelf instabilities, which caused an influx of meltwater into the northern North Atlantic. This may have led to considerable changes in the formation of NADW (Oppo and Lehmann, 1995; Vidal et al., 1997; Elliot et al., 2002; McManus et al., 2004) and hence a weakening and subsequent recovery of the AMOC on timescales of centuries to millennia. These dramatic events had a widespread global impact, encompassing the tropical eastern Pacific (Kienast,2006 personal communication) western Pacific (Stott et al., 2003). They were associated with changing temperatures over Antarctica (Blunier and Brook, 2001), weakening of the Asian summer monsoon (Wang et al., 2004; Ivanochko et al., 2005) as well as meridional shifts of the ITCZ in northeastern South America (Peterson et al., 2000).

Previous modeling studies (Seidov and Maslin, 2001; Vellinga et al., 2002; Timmermann et al., 2005a; Zhang and Delworth, 2005; Dahl et al., 2005; Broccoli et al., 2006; Timmermann et al., 2006) have consistently simulated the characteristics of these reconstructed global climate changes in response to a freshwater-induced shut-down of the AMOC. An important conclusion from these studies is that the large-scale atmospheric circulation changes considerably in response to a weakening of the AMOC.

Moreover, the atmosphere may also provide important feedbacks which in turn alter the state of the AMOC. During a shutdown of the AMOC, the large-scale atmospheric cooling in the North Atlantic leads to a local enhancement of the surface water density and provides hence a negative feedback for the AMOC. On the other hand an enhancement of the meridional temperature gradient in the North Atlantic can lead to an enhancement of transient atmospheric eddy activity and hence an increased poleward moisture flux Nakamura et al. (1994) , thereby providing a positive feedback for the AMOC. Vellinga and Wu (2004) recently proposed that low-latitude negative freshwater anomalies, resulting from a southward shift of the ITCZ, might enhance a weakened AMOC. This result was recently confirmed by Yin et al. (2006).

Another interesting feature which has been found in several freshwater-perturbation experiments is the Atlantic-Pacific seesaw (Saenko et al., 2004; Timmermann et al., 2005a). A weakening of the AMOC can lead to an intensification of intermediate water formation in the northern North Pacific and to the generation of a relatively deep but relatively weak overturning circulation in the Pacific. Still unclear is the role of atmospheric and oceanic teleconnections in establishing this interbasin-wide seesaw.

In this study we investigate the atmospheric response to a freshwater-induced AMOC collapse and its effect on the stability and in particular on the recovery of the AMOC. The spatio-temporal signature of the anomalous freshwater forcing is chosen such as to mimic a typical glacial meltwater pulse. We are able to quantify the influence of certain types of air-sea interactions on the stability of the AMOC by applying the partial basin coupling technique (Wu et al., 2003).

The paper is organized as follows: In section 2 we describe the model of intermediate complexity which is used in this study. We also provide a detailed description of the experimental design applied to mimic glacial Heinrich events. The strategy to decouple the oceanic component from the atmospheric response is explained and an overview of all model simulations is given. Section 3 focuses on the oceanic and atmospheric response to a shut-down of the AMOC as well as on the processes that initiate the recovery of the overturning circulation. In section 4 we discuss the results from the partially-coupled freshwater perturbation experiments, focusing on the origin of global teleconnections and the role of tropical Atlantic air-sea coupling on the AMOC. The main results are summarized and discussed in section 5.

3.2 The Model

We use the three-dimensional atmosphere-sea ice-ocean model ECBilt-Clio. The atmospheric component is version 2 of ECBilt (Opseegh et al., 1998), a spectral T21, three-level, based on quasi-geostrophic equations extended by estimates of the neglected ageostrophic terms in order to close the equations at the equator. ECBilt s response to tropical sea-surface temperature (SST) anomalies is underestimated by a factor of 2-3, mostly due to the low resolution and partly due to the quasi-geostrophic approximation. Synoptic variability associated with transient eddies is explicitly computed. Diabatic heating due to radiative fluxes, the release of latent heat and the exchange of sensible heat with the surface are parameterized. In our model version cloudiness is prescribed. The model contains a full hydrological cycle which is closed over land by a bucket model for soil moisture. The atmosphere is assumed to be completely dry above 500 hPa. Precipitation occurs when moisture is vertically advected through the 500 hPa plane or when relative humidities exceed 80%.

The ocean-sea ice component Clio (Goosse et al., 1999; Goosse and Fichefet, 1999; Campin and Goosse, 1999) consists of a free-surface primitive equation ocean model with 3°x3° resolution coupled to a thermodynamic-dynamic sea ice model. Unlike Knutti et al. (2004) we do not apply freshwater flux compensation, which has resulted previously in artificial responses of the AMOC.

To avoid a singularity at the North Pole the oceanic component makes use of two subgrids: The first one is based on classic longitude and latitude coordinates and covers the whole ocean except the North Atlantic and Arctic. These are covered by the second spherical subgrid, which is rotated and has its poles at the equator in the Pacific (111°W) and Indian Ocean (69°E). In this work all analysis is conducted on these original grids but interpolated onto normal geographic coordinates for visualization purposes. In contrast to the pre-industrial setup used by Renssen et al. (2002), our meltwater experiment are conducted under glacial conditions (Timmermann et al., 2004, 2005a; Justino et al., 2005).

Our model boundary conditions include the Last Glacial Maximum (LGM) ice sheet topography of Peltier (1994), an LGM vegetation index (Crowley and Baum, 1997), a corresponding albedo and reduced CO_2 concentration (200 ppm). All experiments described here start from the last year of a 5000-year coupled control simulation which was run under glacial boundary conditions.

The model setup used here differs from the experiments presented in Timmermann et al. (2004) and Justino et al. (2005) in using a central differences advection scheme instead of an up-stream scheme for the oceanic component. The model setup is identical to the one employed in Timmermann et al. (2005a), Timmermann et al. (2005b). In our experiments Heinrich events are simulated by delivering additional freshwater flux into the northern North Atlantic. The maximum amplitude of the forcing amounts to 1.3 Sv and the duration of the Gaussian-shaped pulse is 200 years (Figure 3.1). The additional freshwater fluxes are evenly distributed over the North Atlantic between 45 °N and 60°N. This corresponds to an integrated global sea level

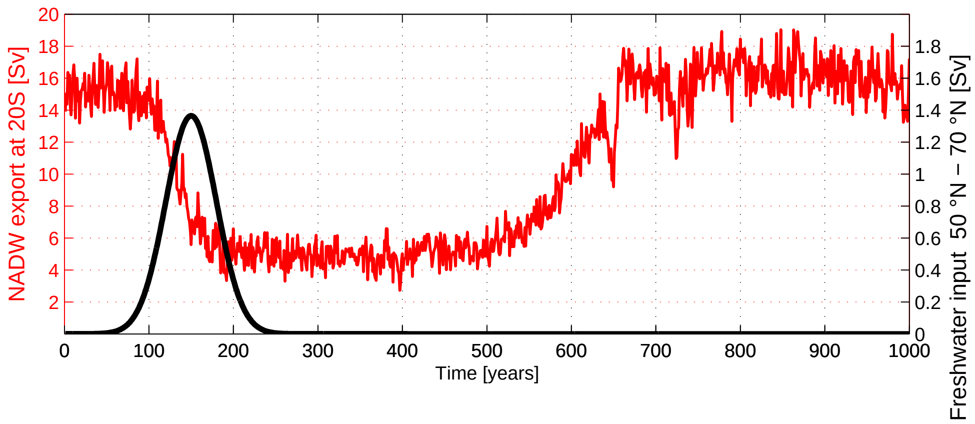

Figure 3.1: Black: Time evolution of the anomalous freshwater forcing [Sv] delivered to the northern North Atlantic between 50°N-70°N; Red: Time evolution of the annual mean maximum export of North Atlantic Deep Water at 20°S.

rise of about 8 m. As shown in Figure 3.1, the AMOC, as measured here by the southward export at 20°S, decreases rapidly to 4 Sv (1 Sv=10^6 m^3/s) and receovers after about 600-700 years

To estimate the climate impact of air-sea interactions related to a North Atlantic meltwater event we compare fully coupled Heinrich-event simulations to experiments where the atmospheric response to SST anomalies is suppressed in selected areas. We use a decoupling technique which is very similar to the partial coupling (PC)-approach proposed by Wu et al. (2003). In decoupled regions the atmospheric component is forced by the climatological seasonal cycle of sea surface conditions from a control experiment (i.e. SST, albedo and sea ice coverage). Atmospheric heat and freshwater fluxes are then computed within the atmospheric model based on this climatological boundary forcing. For the oceanic model component the actual sea surface conditions (including anomalies) predicted by the ocean component are taken to diagnose freshwater-, heat- and momentum forcing. As a consequence, the forcing for the oceanic and atmospheric model components will disagree in the decoupled regions. This particularly applies for the heat fluxes, which are strongly dependent on SST and to a minor degree for evaporation. The main goal of this decoupling approach is to suppress the response of the atmosphere to SST changes in certain regions, whithout damping the internal variability of the atmospheric circulation too much. This simulation strategy will be referred to as "blind atmosphere" further on. This technique differs from forcing the ocean model with prescribed near-climatological fluxes, because in the former case surface air-temperature is fixed, whereas in the latter case atmosphere-ocean fluxes are prescribed. Consequently the blind atmosphere experiments will always tend to damp sea surface temperature anomalies and largely correspond to ocean-only simulations with mixed boundary conditions in the decoupled areas (restored SSTs and fixed freshwater fluxes). However, the SST-dependent evaporation term and moisture transport from fully coupled regions can generate weak freshwater flux anomalies on top of the seasonal cycle of the freshwater fluxes. The following table lists all experiments presented in this study.

Exp. name	**decoupled**	**perturbed**	**duration** $[yrs]$
LGM	–	no	5000
BlindLGM	globally	no	5000
preHE	–	no	50
HE	–	yes	1000
BlindAtlHE	Atlantic north of 30°S	yes	1000
BlindNAtlHE	Atlantic north of 30°N	yes	1000
BlindTrAtlHE	tropical Atlantic (30°S-30°N)	yes	1000

This series of experiments will help us to tease apart the oceanic and atmospheric feedbacks involved in the response of the AMOC to a freshwater-induced weakening.

3.3 Fully coupled climate response to a shut-down of the AMOC

The anomalous North Atlantic freshwater input during the simulation years 50 to 200 influences the strength of the AMOC and hence climate conditions worldwide. To study the global impact of the Heinrich event scenario on the climate system as well as its atmospheric and oceanic responses, we focus on the fully coupled *HE* experiment.

3.3.1 Collapse of the AMOC in the Atlantic

In experiment *HE* the freshwater flux perturbation generates sea surface salinity (SSS) anomalies which exceed –8 psu and spread over the entire Atlantic ocean (Figure 3.3). Advected northward by the North Atlantic current it causes an increased stratification

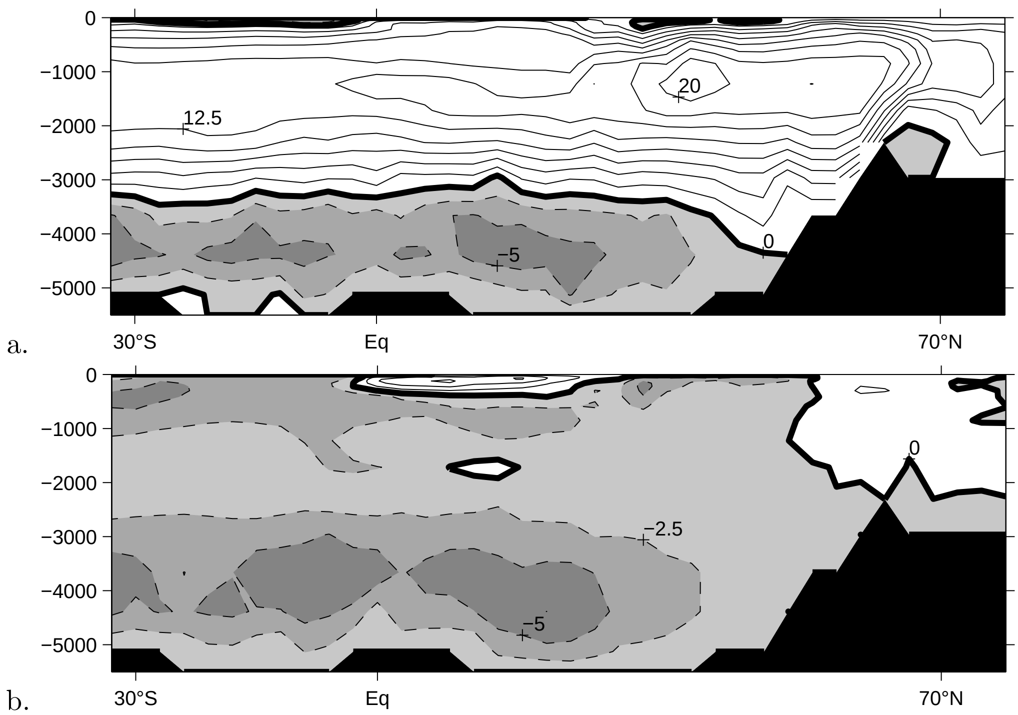

Figure 3.2: Meridional streamfunction [Sv=10^6 m^3/s] in the Atlantic for a. preHE (time average over years 1-50) and b. the collapsed AMOC state after freshwater perturbation (years 201-250). Shaded areas indicate negative stream function values, contour interval is 2.5 Sv

in the sub-polar Atlantic. High latitude deep convection is thus suppressed, causing a radical weakening of the AMOC (Figure 3.2). As a result of the halted circulation the deep southward transport of North Atlantic Deep Water (NADW) vanishes almost completely. The meridional overturning cell reduces to a shallow tropical circulation cell, mainly driven by Ekman transport. The dramatic reorganization of the entire ocean circulation is also documented in (Figure 3.3), which shows a substantial weakening of the North Brazil current and hence of the cross-equatorial flow in the tropical Atlantic and a strengthening of the Brazil current. Figure 3.2 also reveals that in our model simulation a shut-down of the AMOC leaves the Antarctic Bottom Water (AABW) overturning cell almost unchanged.

As a result of the AMOC shut-down the meridional heat transport in the North Atlantic reduces from 1 PW to 0.2 PW (Figure 3.5). In the southern hemisphere it even changes

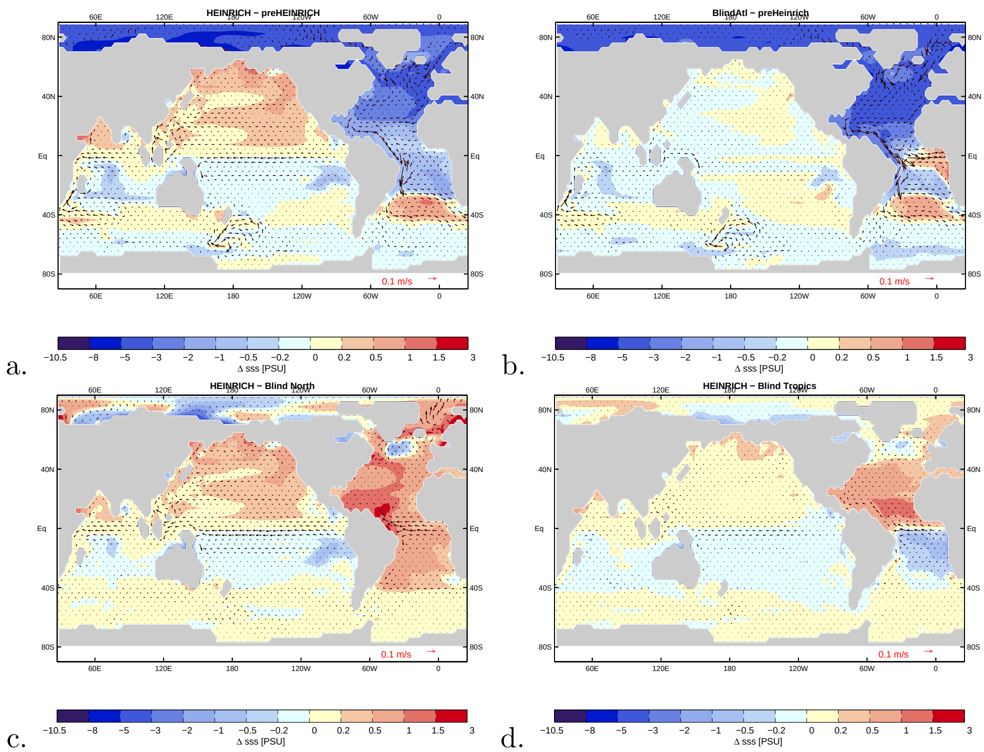

Figure 3.3: Difference of sea surface salinity [psu] and sea surface currents [0.1 m/s] between the time averaged fields of the perturbation experiments (HE (a), BlindAtlHE (b)) (average over years 251-300,) and the time average of the first 20 years of the unperturbed glacial simulation. c. Difference between the time-averaged SSS and sea surface velocity fields of HE and BlindNAtlHE; d. same as c. but for HE and Blind-TrAtlHE. Panels a., b., c., d., represent the full response, the oceanic response without atmospheric teleconnections, the response to the North Atlantic cooling only and the effect of the tropical Atlantic air-sea interactions, respectively.

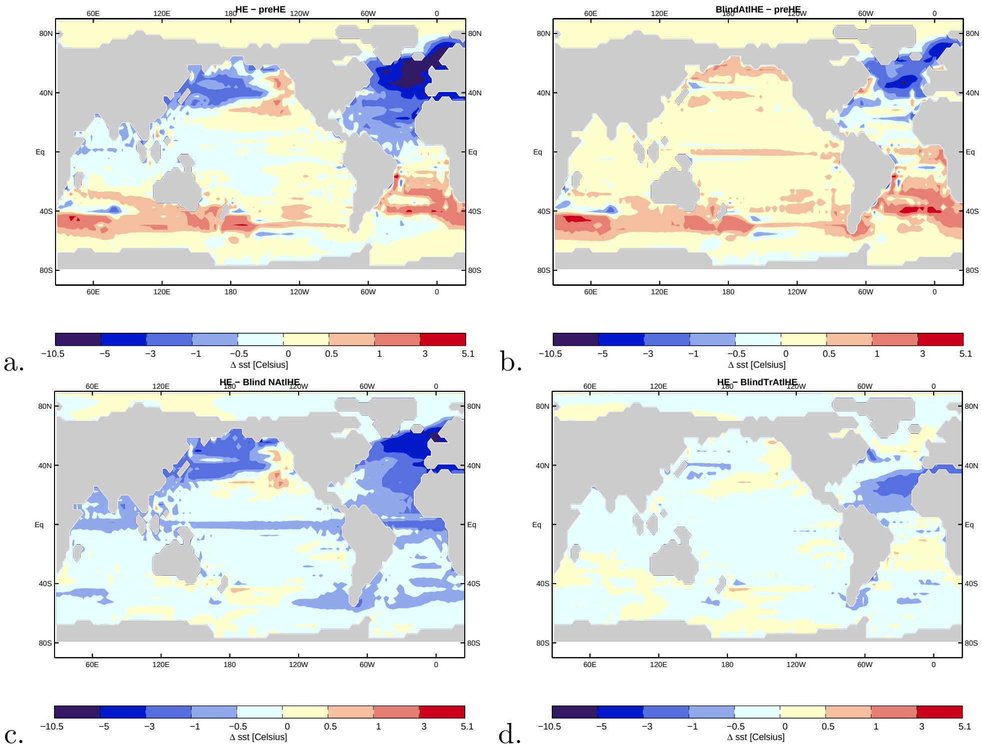

Figure 3.4: As in Figure 3.3, but for the simulated SST [K].

direction from 0.4 PW heat import from the southern ocean in the North Atlantic to –0.3 PW export. As a result of the reduced North Atlantic heat import, North Atlantic SST decrease by up to 10°C while the southern hemisphere exhibits a warming of up to 3°C in some regions of the Antarctic Circumpolar Current (Figure 3.4). In the tropical Atlantic an SST-dipole with an amplitude of 1°C is generated with a zonal SST front located at 5°N. Both the magnitude of the North Atlantic cooling and the anti-correlation between northern and southern hemisphere are characteristics of glacial meltwater events, which are well documented in various paleoclimatic archives (Broecker, 1994; Bond et al., 1993; Blunier and Brook, 2001) and climate model simulations (Knutti et al., 2004; Timmermann et al., 2005a; Stouffer et al., 2006).

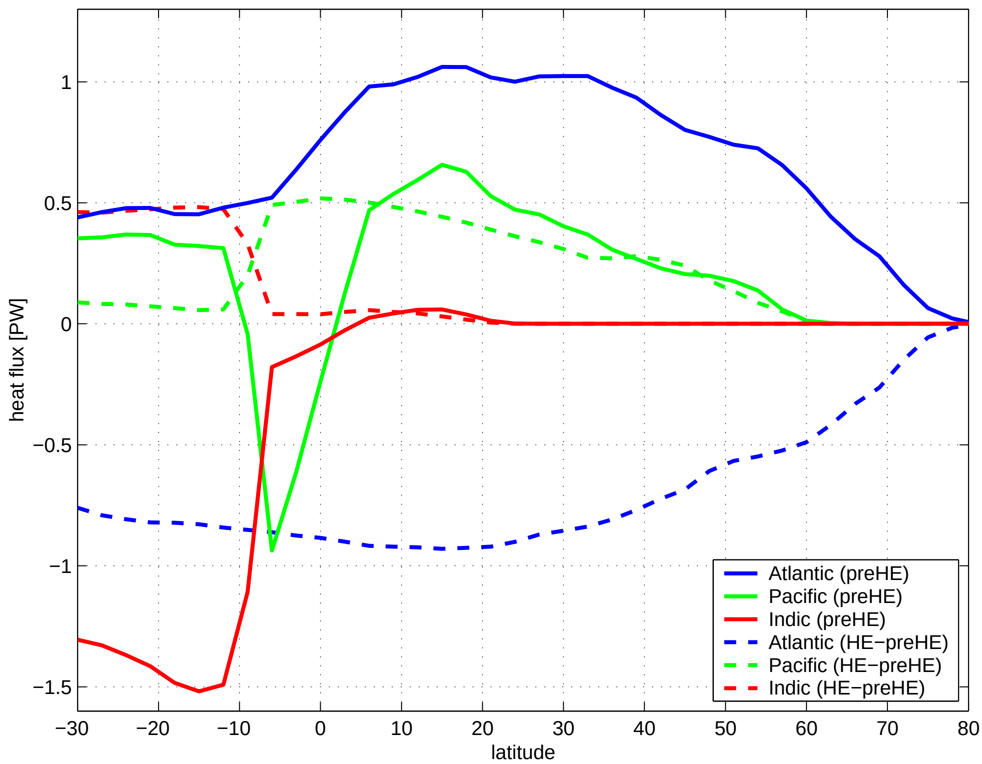

Figure 3.5: zonally and depth integrated mean oceanic heat flux [PW] of the *preHE* experiment (years 1-50) for the Atlantic (solid blue), Pacific (solid green) and Indic Ocean (solid red), and difference of the *HE* experiment after freshwater perturbation (years 201-250) and the *preHE* experiment (years 1-50) for the Atlantic (dashed blue), Pacific (dashed green) and Indic Ocean (dashed red)

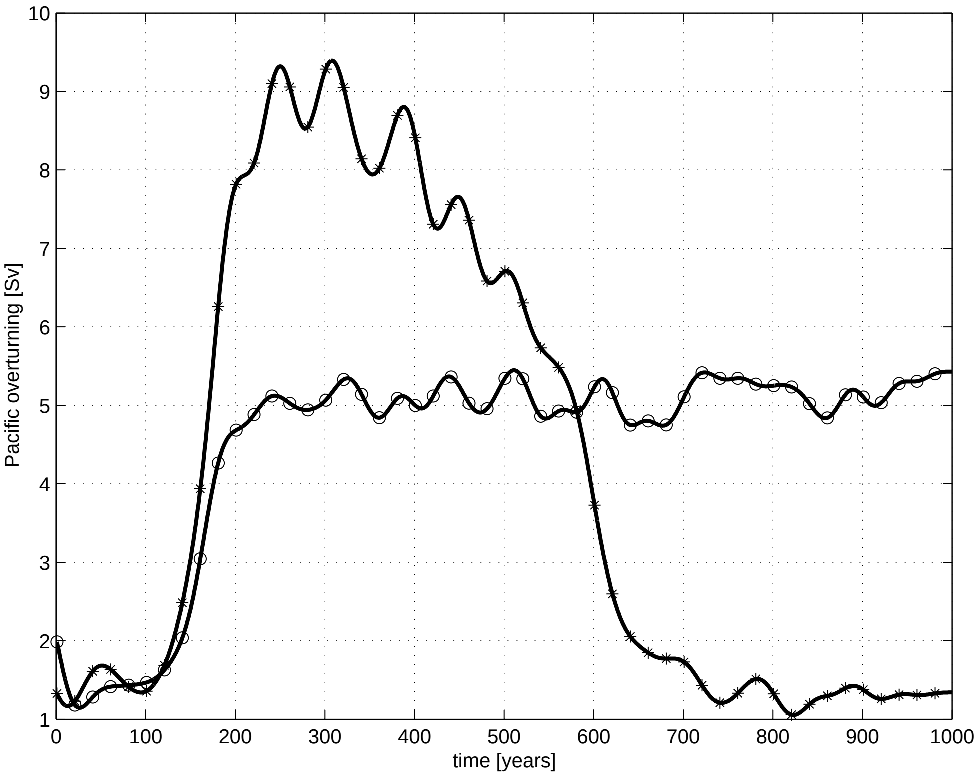

Figure 3.6: Maximum of the meridional streamfunction [Sv] in the North Pacific below 1000 m for experiments HE (as indicated by the asterisk) and BlindAtlHE (circles). The time series is filtered with a 50-year low-pass filter, reducing higher frequency variability.

3.3.2 Impacts of an AMOC shutdown on the global ocean circulation

Changes of the thermohaline circulation in the Atlantic trigger large-scale changes of the ocean circulation in the Indian and Pacific oceans via wave adjustment (Huang et al., 2000; Cessi et al., 2004) and advective processes (Goodman, 2001). Our model simulation *HE* shows that a shut-down of the AMOC leads to a weakening of the Indonesian throughflow by approximately 6 Sv (not shown) and to an increase of North Pacific deep water formation by up to 9 Sv (Figure 3.6). Furthermore, the northward transport of heat is increased in the southern Indian Ocean (by 0.5 PW) and the northern North Pacific (by ~0.2 PW).

It should be noted here that simulated SSTs are only partly controled by the oceanic heat transport convergence, but also by atmospheric heat flux anomalies associated with local and large-scale changes of the atmospheric circulation. In the North Indic and the western and central North Pacific negative SST-anomalies develop in spite of decreased oceanic heat exports from the respective basins (Figure 3.4).

The increased sea surface salinities (Figure 3.3) in the North Pacific and North Indic can be partly explained in terms of anomalous surface freshwater flux anomalies partly by the anomalies in the large-scale oceanic salinity transports. TIMMERMANN ET AL. (2005a) e.g. document that positive SSS anomalies in the Pacific warm pool occuring shortly after a shutdown of the AMOC are related to fast atmospheric teleconnections and meridional shifts of the ITCZ as well as to inter basin seiching of the thermocline. The relative contributions of atmospheric and oceanic teleconnections to the global climate response of an AMOC shut-down will be further quantified below by an analysis of carefully designed blind atmosphere experiments (*BlindAtlHE* and *BlindNAtlHE*).

3.3.3 North Atlantic cooling

As a result of the AMOC shut-down North Atlantic SSTs decrease substantially and the Arctic sea ice margin shifts southward. This results in heat flux anomalies of up to $--250$ W/m^2 in the Norwegian Sea (see Figure 3.7, a). This response helps to set-up surface buoyancy conditions which are favorable for the resumption of the AMOC (RAHMSTORF AND WILLEBRAND, 1995). The mean atmospheric circulation spreads the North Atlantic cooling by anomalous transport of sensible heat, leading to wide-spread cooling of the northern Hemisphere (Figure 3.7). Both, the atmospheric boundary layer and the free atmosphere respond to the cooling by increased sea-level pressure and by intensifying the mean westerlies, respectively (not shown). The increased meridional temperature gradient in the Northern Hemisphere spins up a thermal wind which helps to amplify the mean westerlies. Changes of the mean atmospheric baroclinicity also lead to substantial changes of the midlatitude transient eddy activity (not shown). Furthermore, an increase of the northeasterly trade winds is simulated in response to the AMOC shutdown (Figure 3.8, d). The implications of circulation change on the recovery processes of the AMOC will be discussed further below.

Lower air temperatures (Figure 3.7, a) reduce the saturation vapor pressure over the North Atlantic. This reduces evaporation rates (Figure 3.9, a) inspite of increasing wind speeds (Figure 3.8, a). In the sub-polar Atlantic evaporation rates are further reduced due to an increase of the sea ice covered area. Relative humidity decreases in areas where anomalously dry and cold air is advected into regions of smaller negative air-

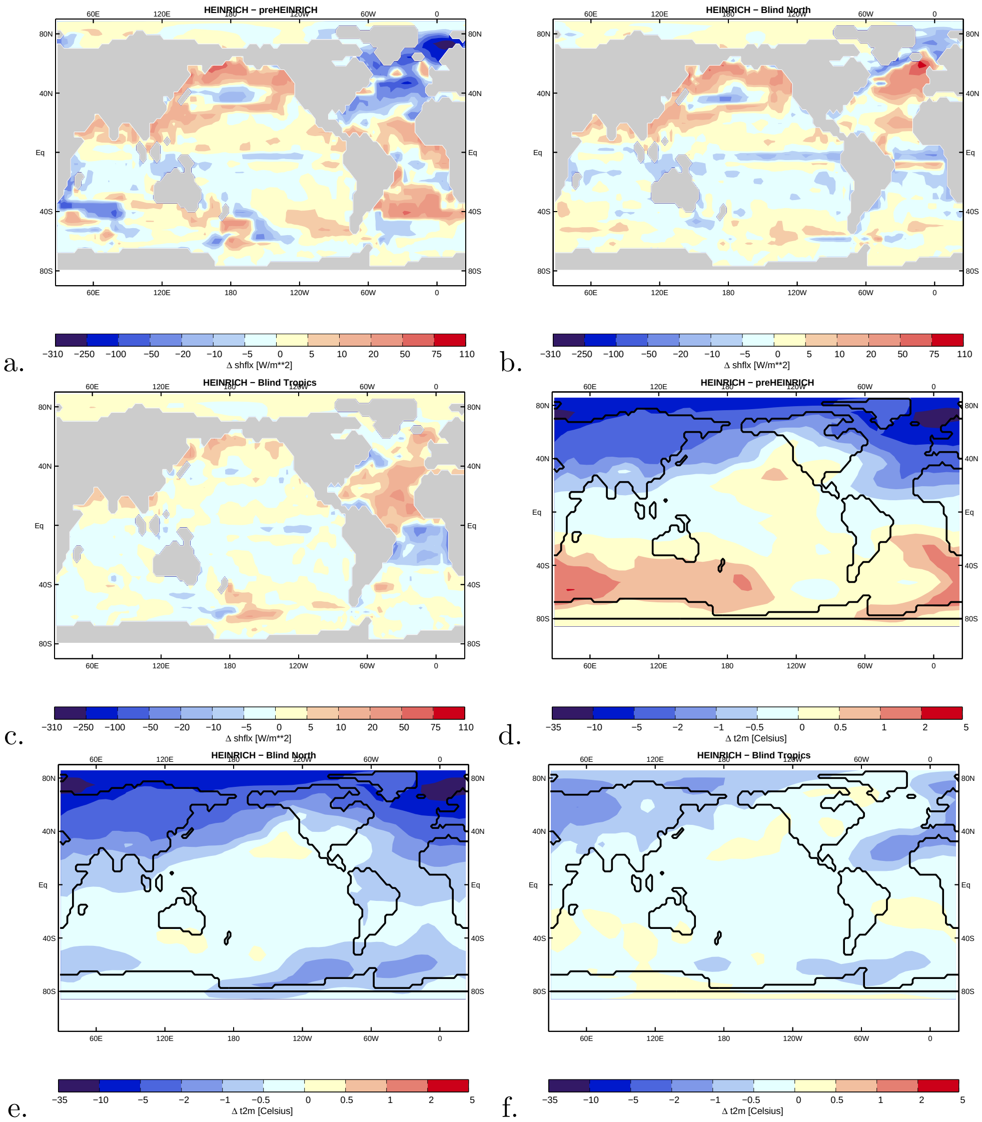

Figure 3.7: As in Figure 3.3, but for the simulated ocean heat flux anomaly [W/m^2] obtained from the difference of 50-year-long averages of the a: HE and preHE ; b: HE and BlindNAtlHE; c: HE and BlindTrAtlHE experiments. Positive values represent a warming tendency of the ocean., d., e., f., as in a., b., c., but for 2m air-temperature [K]

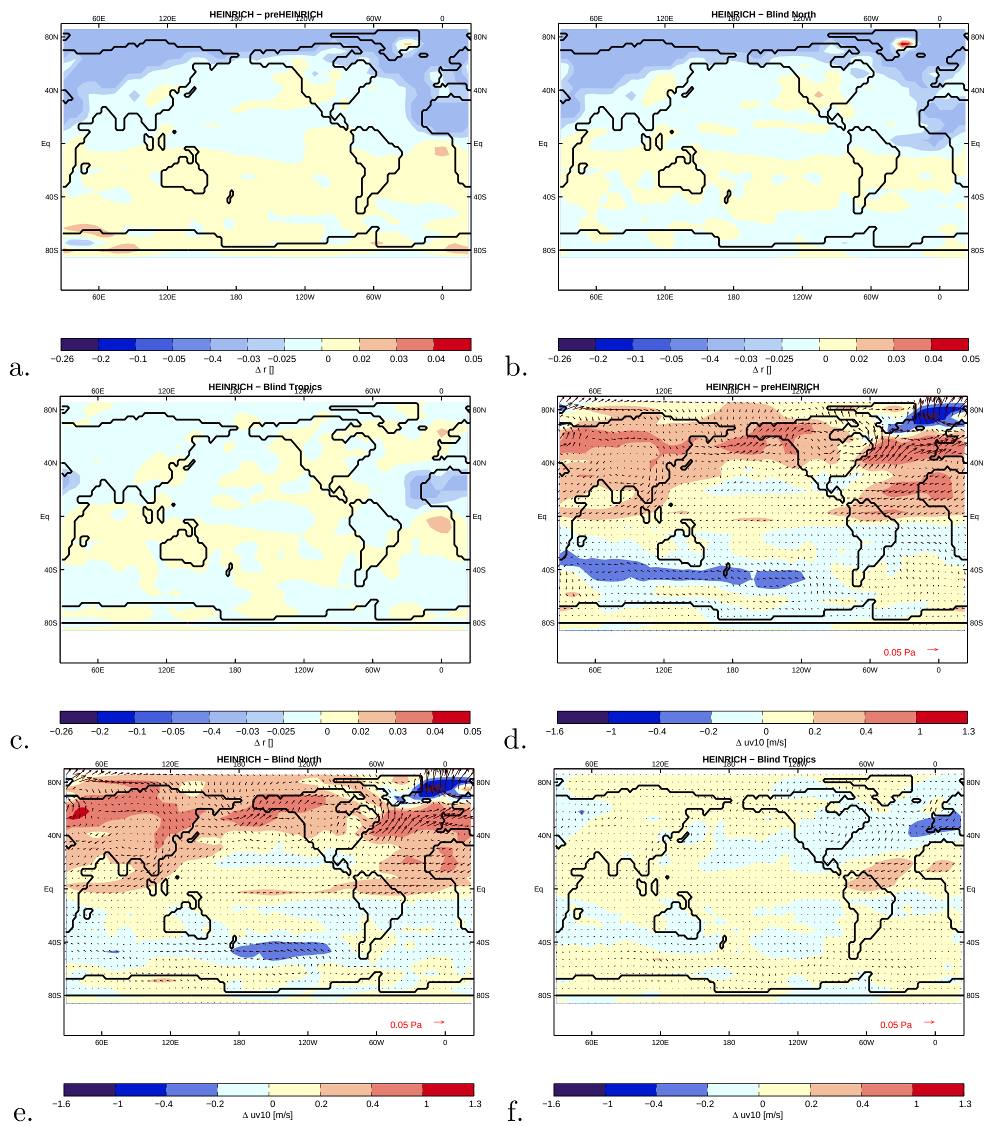

Figure 3.8: a.,b., c. as in Figure 3.7 a.,b.,c. but for the atmospheric relative humidity; d.,e.,f., as im a.,b.,c, but for windstress vectors at sea surface and 10 m winds.

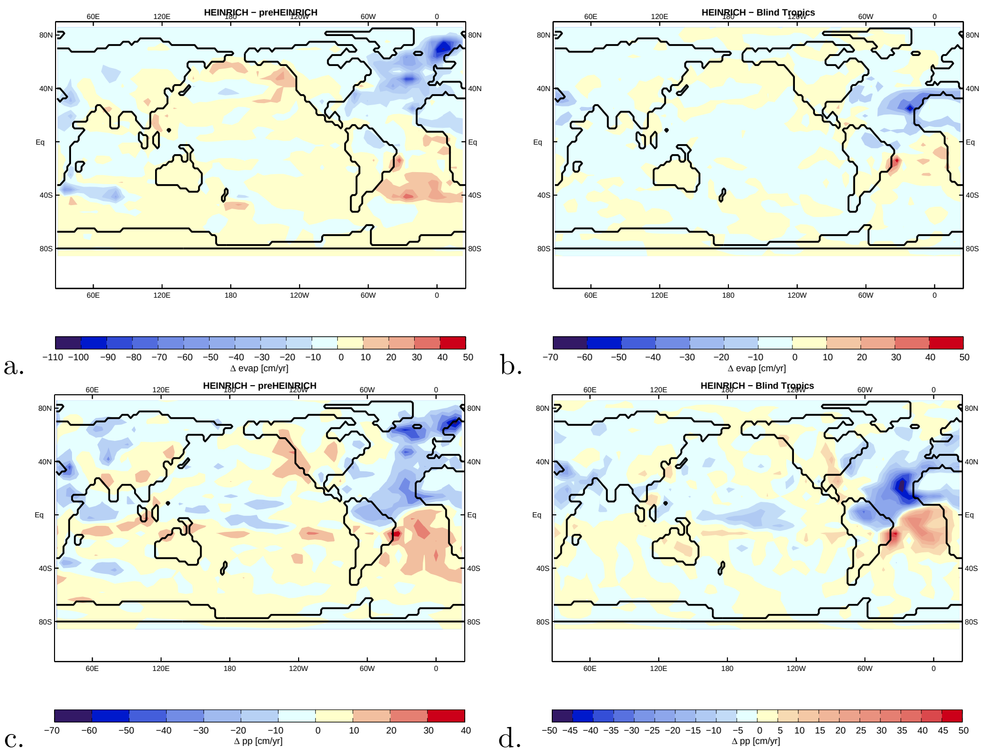

Figure 3.9: a., b., same as Figure 3.3 a., d., but for the evaporation [cm/year]. Positive values represent increased evaporation from the ocean to the atmosphere; c.,d., same as a. and b., but for precipitation [cm/years]

temperature anomalies (Figure 3.8). This effect can be clearly seen for northern Africa and the eastern North Atlantic between 10°N and 40°N. Our model simulations suggest that the Eurasian continent and North Africa experienced extreme arid conditions during Heinrich events, in agreement with paleo-reconstructions (e.g. Bartov et al. (2003), Geraga et al. (2005))

3.3.4 Atmospheric response in the tropical Atlantic

Previous modeling studies (Dong and Sutton, 2002; Vellinga and Wu, 2004; Timmermann et al., 2005a; Zhang and Delworth, 2005; Dahl et al., 2005; Yin et al., 2006; Broccoli et al., 2006)) as well as many paleo-records (Stott et al., 2003; Turney et al., 2004; Peterson et al., 2000) suggest a relation between North Atlantic cooling and southward shifts of the ITCZ both in the tropical north Atlantic

(Wang et al., 2004; Peterson et al., 2000) and the tropical western Pacific (Wang et al., 2001; Timmermann et al., 2005a). This finding is corroborated by the precipitation anomalies in simulation *HE*. Figure 3.9 c shows an equatorial dipole pattern in precipitation anomalies with maximum anomalies reaching up to 40cm/yr in some areas. Furthermore, the strengthening/weakening of northeasterly/southeasterly trade winds in the North Atlantic in particular (Figure 3.8) is reminiscent of a southward shift of the ITCZ.

The southward shift of the ITCZ results from a southward displacement of the Hadley cell which balances the increased temperature gradient between high and low latitudes (Broccoli et al., 2006). The latitudinal zero-crossing of the annually averaged meridional atmospheric heat transport is a measure for the position of the zonal mean ITCZ. As shown in Figure 3.10 the zero-crossing shifts southward by 3-6° during the AMOC collapse. A recent model intercomparison study by Stouffer et al. (2006) reveals that our simulated changes in the meridional structure near the equator are smaller than those simulated by freshwater perturbation experiments conducted with state-of-the-art Coupled General Circulation Models (CGCMs). This may be partly due to the low atmospheric resolution employed in our experiments, partly due to the representation of ageostrophic dynamics in the atmosphere model.

Even though the tropical response is somewhat underestimated in our experiments, the mean atmospheric freshwater transport across the equator (Figure 3.11) is reduced by up -0.2 Sv during the AMOC shut-down phase. In certain regions this anomaly can even compensate the effect of the northern North Atlantic freshwater perturbation of maximal 1.4 Sv. As a result of the increased export of freshwater from the North Atlantic to the Southern Hemisphere and the reduced cross-equatorial surface currents (Figure 3.3, a), positive salinity anomalies develop in the northern tropical Atlantic which exceed pre-Heinrich levels already in year 300 (nolt shown). Advected poleward, the salinity anomalies play a crucial role in the recovery of the AMOC by destabilizing the stratification in the regions of deep water formation (Krebs and Timmermann, 2006a).

The ITCZ shift leads a meridional displacement of the Hadley circulation and an associated increase of the cross-equatorial heat transport (Figure 3.5). The heat flux convergence has the tendency to weaken the negative SST anomaly in the North Atlantic. Furthermore, increased heat fluxes south of 20°N (Figure 3.7, originating partly from the reduced evaporation (Figure 3.9) in this area also lead to a damping of the negative SST anomalies.

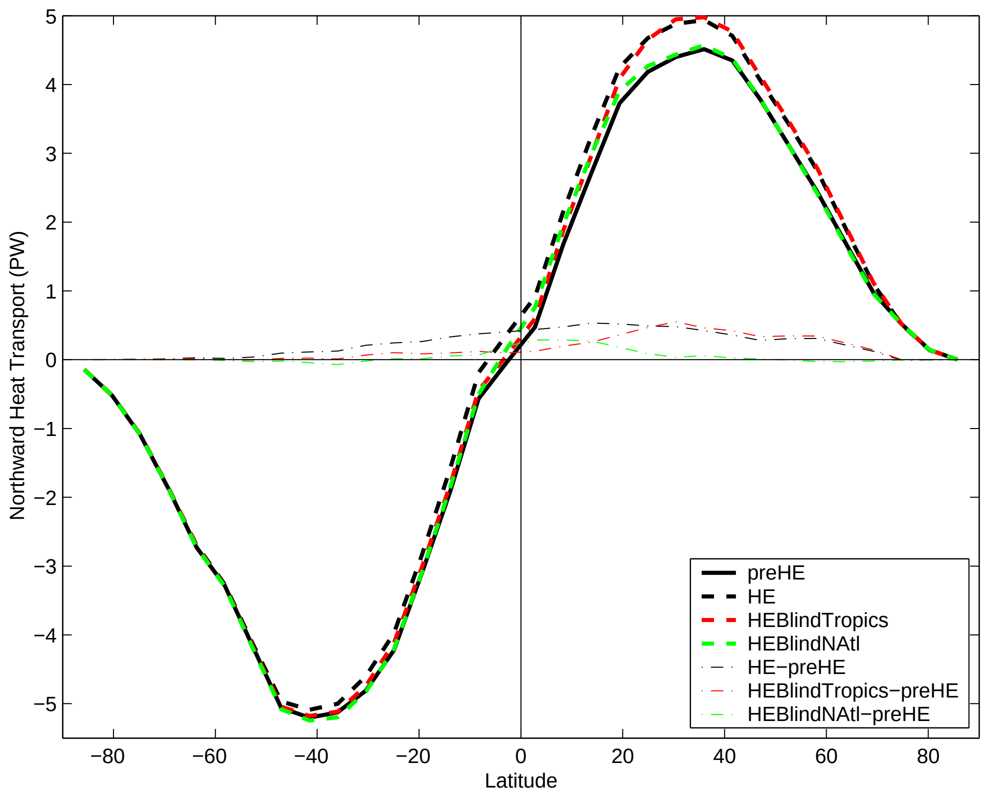

Figure 3.10: Annually averaged zonally intergrated total northward energy flux by the atmosphere [PW=10^{15}W] for *preHE* (black solid), *HE* (black dashed), *BlindTrAtlHE* (red dashed), *BlindNAtlHE* (green dashed), *HE* - *preHE* (black dash-dotted), *BlindTrAtlHE-HE* (red dash-dotted), *BlindNAtlHE -HE* (green dash-dotted).

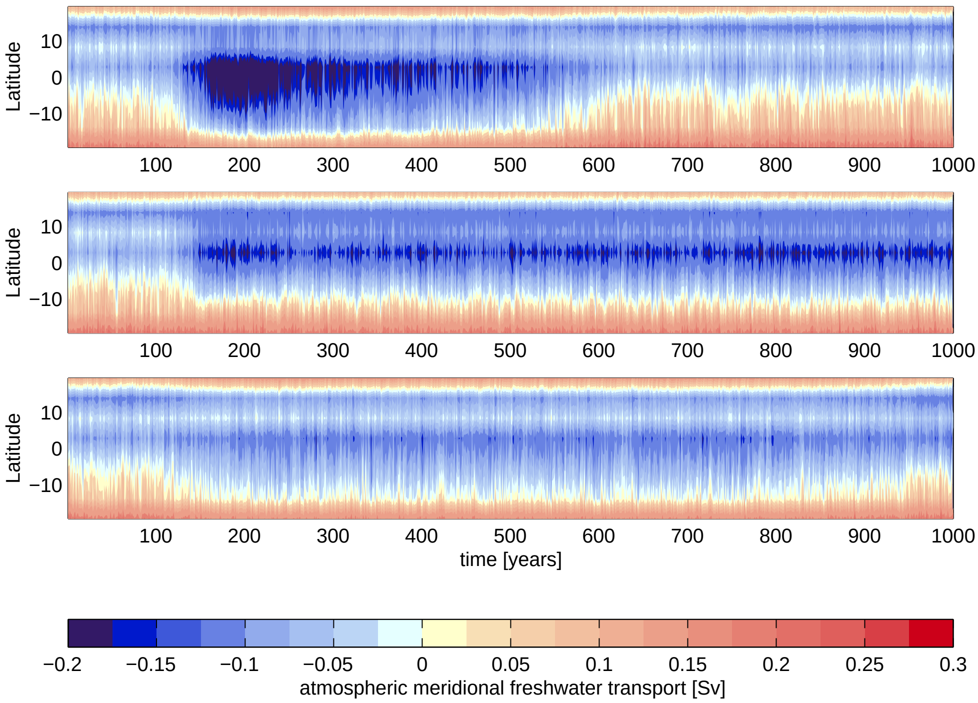

Figure 3.11: Atmospheric meridional freshwater transport (computed by zonal integration of surface freshwater fluxes): **upper**: experiment *HE* **middle**: experiment *BlindNAtlHE* **lower**: experiment *BlindTrAtlHE*

3.4 Air-sea interactions and their role in the recovery of the AMOC

To quantify the local and remote influences of SST anomalies in certain areas on the atmosphere and back to the ocean we decouple the atmosphere from the underlying SST anomalies induced by the AMOC shutdown by using the blind atmosphere technique.

3.4.1 Global climate response to Atlantic SST anomalies during the AMOC shut-down phase

In experiment *BlindAtlHE* the atmosphere over the Atlantic north of 30°S is decoupled from the SST anomalies by the technique described in section 3.2. Decoupling the atmosphere from the SST anomalies primarily eliminates the temperature feedback (Rahmstorf and Willebrand, 1995) which provides a negative feedback for the AMOC. In *BlindAtlHE* the atmosphere warms the ocean and North Atlantic heat fluxes change sign. This further enhances stratification in the regions of deep water formation and thus delays the recovery of the AMOC by several thousand years (Figure 3.12). During the 5000-year-long *BlindAtlHE* simulation the AMOC does not recover. Ultimately a recovery of the AMOC might be triggered by the diffusive warming of the deep ocean (Winton and Sarachik, 1993; Timmermann and Goosse, 2003). This impressively illustrates that the choice of boundary conditions (blind atmosphere is quite similar to mixed boundary conditions) may fundamentally alter the stability characteristics of the AMOC.

The prime virtue of this experiment, however, is the elimination of all important atmospheric teleconnections. The atmosphere only responds to SST changes outside the Atlantic area. We will particularly study the *BlindAtlHE - preHE* and the differences for the years 250-300, which primarily reflect the global climate response without Atlantic atmospheric teleconnections. In the off-equatorial Atlantic surface current anomalies computed from *BlindAtlHE - preHE* largely resemble the results from experiment *HE*, while near-equatorial anomalies are remarkably different (Figure 3.3, b). The weakening of the South Equatorial current in experiment *BlindAtlHE* can be explained in terms of the missing intensification of the trades. Furthermore, the meridional overturning circulation in the North Pacific is substantially weakened as compared to experiment *HE* (Figure 3.6). This can be explained by the absence of negative surface freshwater fluxes and by the absence of negative SST anomalies (Figure 3.4, b). This indicates that in our experiments the Pacific-Atlantic see-saw (Saenko et al., 2004) is driven

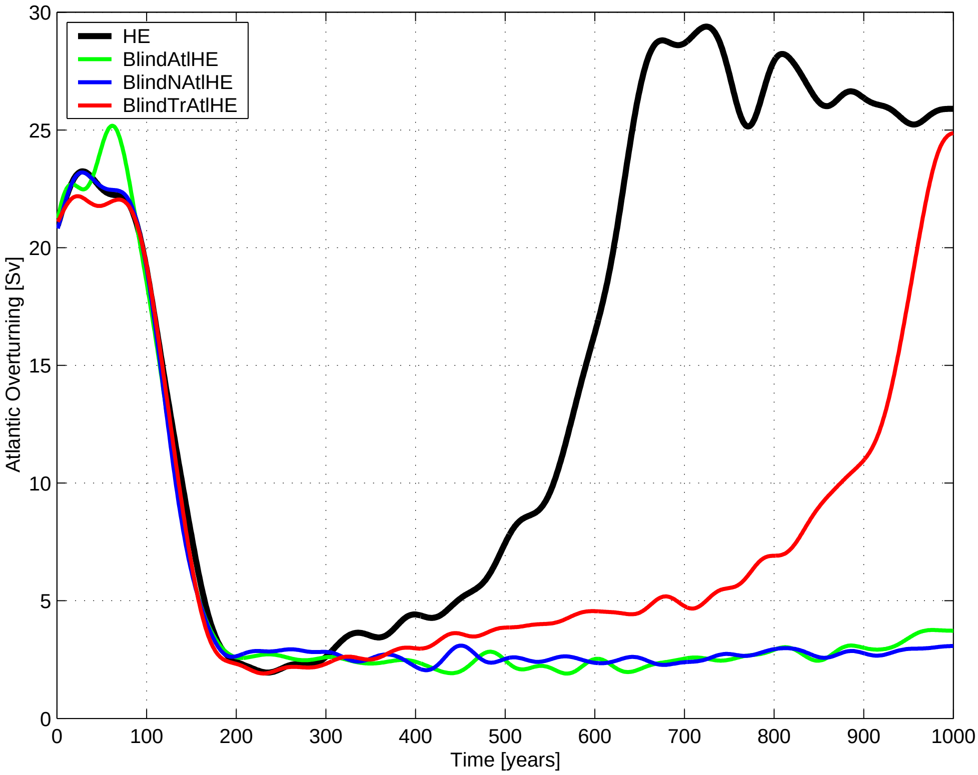

Figure 3.12: Time series of NADW export at 20°S [Sv] for *HE* (black), *BlindTrAtlHE* (red), *BlindNAtlHE* (blue), and *BlindAtlHE* (green).

by atmospheric as well as oceanic teleconnections in almost equal parts. In contrast, the Indonesian throughflow reduces by 6 Sv both in experiment *HE* and experiment *BlindAtlHE* (not shown). This is somewhat surprising because the throughflow can be largely obtained from the wind-stress integral along the boundaries (Godfrey, 1989) and the JEBAR term. The fact that the Indonesian throughflow hardly changes, in spite of substantial differences in the wind-stress fields in *BlindAtlHE* and *HE* suggests that the density changes during the AMOC shutdown in combination with the JEBAR effect play an important role in controlling the inter-basin exchange between Indian ocean and Pacific.

Timmermann et al. (2005b) propose that in the ECBilt-Clio model North Atlantic density changes are partly communicated via viscous boundary/ Kelvin waves to the other ocean basins and yield a global adjustment of sea level and thermocline which eventually drives the pan-oceanic circulation changes.

In experiment *BlindAtlHE* and in response to the shutdown of the AMOC, upper ocean temperatures increase in the South Atlantic, the Southern Ocean and the eastern Pacific Figure 3.4). This supports the idea that global oceanic teleconnections play an important role in the generation of temperature anomalies outside the Atlantic basin. Within the North Atlantic we observe that temperature anomalies between 0°N and 40°N are much weaker in *BlindAtlHE* as compared to experiment *HE*. This clearly illustrates that during the Heinrich event the atmospheric circulation response helps to spread the cooling from the northern North Atlantic equatorward.

3.4.2 Global climate response to North Atlantic SST anomalies during the AMOC shut-down phase

In the experiment *BlindNAtlHE* the atmosphere over the Atlantic is decoupled from the ocean north of 30°N by using the same strategy as described above. Air-sea coupling in the tropical Atlantic is fully captured. Due to the lack of the negative temperature feedback (Rahmstorf and Willebrand, 1995) we again do not observe any AMOC recovery during the 1000 years of integration of this experiment. In the following we will quantify the effect of the atmospheric response to the cooling in the northern North Atlantic by computing the time averaged difference fields of air temperature, atmosphere to ocean freshwater flux, sea surface salinities and SST between the standard *HE* experiment and the experiment *BlindNAtlHE*.

The strong cooling in the northern North Atlantic leads to an almost global cooling (Figure 3.7,c,e). Hence in *HE*, strong oceanic heat flux convergence has to overcom-

pensate this cooling in the Southern Hemisphere to generate the net warming of the bi-polar seesaw.
Our analysis of the differences of *HE* and *BlindNAtlHE* also reveals that the northern North Atlantic cooling generates an intensification of the trade winds in the tropical North Atlantic (Figure 3.8, e), a southward shift of the ITCZ, a decrease of the relative humidity in the North Atlantic (Figure 3.8, b), the generation of positive salinity anomalies in the entire Atlantic (Figure 3.3, c) and an intensification of the South equatorial Current in the tropical North Atlantic. Figures 3.9 also reveal that North Atlantic SST anomalies can only partly explain the simulated freshwater flux and wind anomaly patterns in *HE*. Tropical SST changes are in fact contributing significantly to the tropical response as will be shown below. Moreover, the reduction of the atmospheric freshwater export from the Northern to the Southern Hemisphere in *HE* (as depicted in Figure 3.11) can only be partly explained in terms of the Atmospheric response to the northern North Atlantic SST anomalies induced by the AMOC shut-down (Figure 3.11, middle panel).
These results support the notion that extratropical SST anomalies can significantly influence tropical climate during Heinrich events (Broccoli et al., 2006) and will ultimately play an important role in the recovery of the AMOC as shown in Figure 3.12 (difference between red and green lines)

3.4.3 Global climate response to tropical Atlantic SST anomalies during the AMOC shut-down phase

In this section we quantify the impact of the tropical Atlantic SST anomalies (which can be partly triggered by the atmospheric response to cooling in the northern North Atlantic, as illustrated in Figure 3.4, c) on the atmospheric circulation and its impact back on the ocean by analyzing the *BlindTrAtlHE* experiment. In *BlindTrAtlHE* the atmosphere is decoupled from the Atlantic SST between 30°N and 20°S.
Unlike the North Atlantic response the atmosphere does not develop strong teleconnections to the other ocean basins in response to tropical Atlantic SST anomalies. The *HE-BlindTrAtlHE* decompositions of surface heat flux (Figure 3.7, c), air temperature, and near-surface winds (Figure 3.8) exhibit no pronounced signals in the Pacific or Indian Ocean. Given the fact that our tropical-air sea interactions are somewhat underestimated, the teleconnections from the tropical Atlantic to the other oceans may be also underestimated, as compared to state-of-the-art CGCMs.
As shown in Figure 3.7, f tropical SST anomalies have an effect on extratropical climate

in the Atlantic and lead to a reduction of air temperatures of up to $-3°$C. Furthermore, tropical Atlantic SST anomalies are responsible for the reduction of relative humidity in the eastern tropical Atlantic (Figure 3.8, c). This can be explained in terms of mean advection of drier air from the anomalously cold subtropics to the tropical Atlantic. This in turn reduces precipitation in the northern tropical Atlantic (Figure 3.9 d) and allows cross equatorial southward atmospheric moisture transport (Figure 3.11, lower panel) in connection with the southward shift of the ITCZ. Accordingly, north equatorial surface salinities rise in response to these local air-sea interactions (Figure 3.3 d). Being transported poleward with the mean wind-driven circulation they reduce the stratification and enhance the vertical mixing in the northern North Atlantic.
The relevance of the atmospheric response to tropical Atlantic SST for the AMOC becomes evident by comparing the recovery timescales of the AMOC in experiments *HE*, *BlindNAtlHE* and *BlindTrAtlHE*. In comparison to the standard experiment *HE* the recovery of the AMOC in *BlindTrAtlHE* is delayed by almost 300 years (Figure 3.12, in spite of very similar surface buoyancy fluxes in the regions of deep water formation. This highlights the importance of tropical air-sea interactions for the AMOC recovery. On the other hand the AMOC in *BlindTrAtlHE* recovers in contrast to the *BlindNAtlHE*. This implies that the coupling in the North Atlantic (included in *BlindTrAtlHE*, not included in *BlindAtlHe*) is another crucial element for the recovery of the AMOC.

3.5 Summary and discussion

Using the earth system model of intermediate complexity ECBilt-Clio, this study explored the role of the atmosphere for the recovery of the AMOC after a freshwater-induced shutdown. After a spin-up under glacial boundary conditions a freshwater perturbation, mimicking a glacial Heinrich event, was applied to the northern North Atlantic. The simulated response (amplitude of North Atlantic cooling and southern hemispheric warming, southward ITCZ shift and recovery within centuries) is generally consistent with paleoclimatic reconstructions of Heinrich events (Hemming, 2004; Dahl et al., 2005).
In this study the atmospheric response to an AMOC collapse was assessed by suppression of air-sea coupling in certain key areas. Our analysis focused on the

- **Large-scale oceanic response** which is responsible for significant oceanic anomalies in the Atlantic, temperature anomalies in the southern Ocean, the

Indian and Pacific Ocean.

Due to the shut-down of the AMOC a net export of heat from the North Atlantic to the South Atlantic takes place. In turn, North Atlantic surface waters cool by more than 5°C. The density changes in the North Atlantic are transmitted globally by baroclinic Kelvin waves. Ultimately through shedding of Rossby waves and the advection of different water masses, the Indonesian throughflow reduces by 6 Sv, North Pacific deep water formation increases by 5 Sv and the global thermocline readjusts (Timmermann et al., 2005a). These effects were clearly identified by suppressing the atmospheric teleconnections from the Atlantic to the other oceans (experiment *BlindAtlHE*). This experiment also revealed that SST changes in the tropical and subtropical North Atlantic strongly depend on the atmospheric response, whereas those in the southern ocean can be attributed mostly to oceanic circulation changes.

- **The North Atlantic cooling** does not only trigger global teleconnections but it is also responsible for an intensification of the trade winds and a southward shift of the ITCZ in the entire tropics. The overall cooling induced by the cooling in the North Atlantic leads to a substantial reduction of the relative humidity. Extratropical SST anomalies lead to an enhancement of the Hadley Cell and can induce a southward shift of the ITCZ and hence reduced freshwater fluxes in the northern tropical Atlantic. Both effects together lead to the generation of positive salinity anomalies in the entire North Atlantic (Figure 3.3). The associated increase of the surface buoyancy fluxes in the northern North Atlantic is sufficient to trigger a resumption of the AMOC after about 800 years (Figure 3.12).

  While the AMOC decreases, North Pacific deep overturning increases in the fully coupled experiments. In the case of a suppressed atmospheric SST response in the North Atlantic the North Pacific deep overturning is still present, but reduced by about 50%. This suggests that the Atlantic-Pacific seesaw (Saenko et al., 2004) can be explained both in terms of changes in the atmospheric forcing and pan-oceanic connections.

- **Atmospheric feedbacks in the tropical Atlantic** considerably accelerate the recovery of the AMOC. The AMOC shutdown leads to substantial changes of the Atlantic heat transport and to the development of an anomalous zonal SST dipole centered at 5°N of ±1°C. This dipole causes an atmospheric reorganizations between 30°S and 30°N and provides an important negative feedback for the

AMOC as documented by the precipitation and salinity difference fields between the fully coupled and the partially coupled *BlindTrAtlHE* experiments.

It has to be noted here that the experiments *BlindNAtlHE* and *BlindTrAtlHE* are not entirely independent from each other. North Atlantic SST anomalies trigger an atmospheric response which will eventually generate SST anomalies in the tropical North Atlantic.

Yang and Liu (2005) recently adopted a similar decoupling strategy to demonstrate that two thirds of the tropical Atlantic SST anomalies during an AMOC weakening can be explained by the atmospheric response to extratropical North Atlantic SST anomalies. This atmospheric bridge is further maintained by local air-sea interactions in the tropical Atlantic. In our model simulations we find that the tropical response is almost entirely suppressed in experiment *BlindTrAtlHE* but only partly in experiment *BlindNAtlHE*, thereby corroborating the results of Yang and Liu (2005).
As illustrated by the *BlindNAtlHe* experiment, the local temperature feedback in the northern North Atlantic plays a key role in the recovery of the AMOC. Another crucial negative feedback for the AMOC shut-down is provided by the accumulation of positive salinity anomalies in the North Atlantic induced by the southward displacement of the ITCZ in combination with reduced cross-equatorial surface flows. This feedback has already been identified and analyzed by Vellinga and Wu (2004) in the context of multidecadal variability and in Yin et al. (2006) in the context of the AMOC recovery. This negative feedback is quite important in the sense that it offsets the positive density-advection feedback (Stommel, 1961). It is the interplay of these positive and negative feedbacks which determines the recovery time of the AMOC after a major shut-down. Many two-dimensional models of intermediate complexity do not capture tropical air-sea interactions and may hence overestimate the recovery time of a collapsed AMOC. Moreover, the recent compilation of state-of-the-art CGCM waterhosing experiments (Stouffer et al., 2006) reveals that the tropical temperature and ITCZ responses simulated by ECBilt-Clio are somewhat underestimated. This may imply that the negative feedback provided by the tropical air-sea coupling in the Atlantic may be even more efficient in reality than in our coupled model.
More research needs to be done to constrain the relative roles of the density advection feedback, the temperature feedback, the ITCZ feedback and the role of heat diffusion in the ocean for the AMOC stability.

Chapter 4

The relative effects of vertical diffusion and tropical air-sea coupling on the recovery of the AMOC

Abstract Vertical diffusion has previously been considered as a key factor controlling the strength and stability of the Atlantic Meridional Overturning Circulation (AMOC). Simple models without mixing energy constraints predict an intensification of the AMOC with increasing vertical diffusivities following a power-law with an exponent of $\sim 2/3$. Overall, the strong sensivity of the AMOC stability characteristics to vertical diffusion has been confirmed using ocean general circulation models. Recent coupled general circulation model studies have revealed that tropical air-sea coupling may be another important controlling factor for the strength and stability of the AMOC.
Here an attempt is made to quantify the relative roles of tropical air-sea coupling and vertical diffusion on the stability characteristics of the AMOC using a coupled dynamical atmosphere-ocean-sea ice model of intermediate complexity. Our focus is on the recovery processes of the overturning circulation after a major freshwater-induced collapse. In model simulation without air-sea coupling in the tropical Atlantic, the recovery of the AMOC is strongly controlled by vertical diffusion, whereas this sensitivity is absent in a fully coupled simulation. Our analysis suggests that meridional advection of anomously saline tropical waters is a more efficient negative feedback for the off-state of the AMOC than density homogenisation due to vertical exchange processes.

4.1 Introduction

Providing a poleward heat transport of about 1 PW (Petawatt = 10^{15} W) (Wunsch, 2005a), the Atlantic Meridional Overturning Circulation (AMOC) is a key element in reducing the pole-to-equator heat contrast, thereby regulating climate in high latitudes. Modelling studies (Stouffer et al., 2006) and paleoclimatic evidence (McManus et al., 2004) suggest that the associated transport of heat and mass can be considerably weakened in response to extratropical freshwater perturbations. During the last glacial period occasional instabilities of the northern hemispheric ice sheets led to surges of icebergs into the northern North Atlantic and hence a freshening of the subpolar waters. These events – referred to as Heinrich events (Broecker, 1994) – were linked to large-scale reorganizations of the atmospheric and oceanic circulations, encompassing remote areas such as Antarctica (Blunier and Brook, 2001; Stocker and Johnsen, 2003) or the Indian (Ivanochko et al., 2005) and the Pacific ocean (Stott et al., 2003).
Understanding the response of the AMOC to past extratropical density perturbations may also help to assess the sensitivity of the AMOC to future changes of the high latitude buoyancy. Some modeling studies suggest that anthropogenic warming and associated changes of the hydrological cycle in high latiudes may lead to a substantial weakening of the AMOC (Dixon et al. (1999), IPCC (2001), Gregory et al. (2005)). However, recent multi-model ensemble intercomparisons (Schmittner et al., 2005), have revealed that the simulated AMOC response to increasing greenhouse gas concentrations is still quite uncertain.
Understanding this uncertainty requires a dynamical understanding of the processes controlling the transient and stationary response of the AMOC to North Atlantic density perturbations.
According to Stommel (1961) the sinking of North Atlantic Deep water can be – at least partly – balanced by uniform upwelling, which can be provided by vertical diffusion. Simple scaling arguments (Bryan, 1987) suggest that the strength of the meridional overturning circulation in the North Atlantic Φ obeys the scaling law $\Phi \sim \Delta\rho^{1/3}\ \kappa_v^{2/3}$, where $\Delta\rho$ and κ_v represent the meridional density gradient and the vertical diffusivity, respectively. However, the derivation of this scaling law presented in Bryan (1987) assumes that the typical meridional velocity is similar to the zonal velocity. This assumption is probably not well justified and has caused widespread misunderstandings. Several numerical studies have basically confirmed that enhanced vertical diffusion (representing subgrid scale vertical mixing) results in a stronger present-day meridional

overturning circulation (Bryan, 1987; Wright and Stocker, 1992; Marotzke, 1997; Zhang et al., 1999). Its effect on the stability of the AMOC, however, has been shown to be quite model-dependent. While enhanced vertical diffusion was found to destabilize the off-state of the AMOC in coupled climate models with two-dimensional ocean components (Ganopolski et al. (1998), Schmittner and Weaver (2001)), three-dimensional ocean general circulation models (Prange et al., 2003) can even exhibit an opposite sensitivity. As argued by Prange et al. (2003) the horizontal gyre circulation, which is not represented in zonally integrated models might play an important role in offsetting the vertical diffusive effects on the off-state.

Another important process which may weaken the sensitivity of the AMOC to vertical mixing is air-sea coupling (Weber, 1998). Our paper investigates the influence of vertical diffusion on the stability of the off-state in a fully coupled atmosphere-ocean-sea ice model of intermediate complexity and a partially coupled version of this model. Mimicing glacial Heinrich events, we analyse the recovery of the AMOC after a freshwater-induced shutdown for different values of the vertical diffusion and for different types of coupling.

4.2 Simulation of a freshwater induced AMOC collapse

We use the three-dimensional atmosphere-sea ice-ocean model ECBilt-Clio. The atmospheric component is version 2 of ECBilt (Opseegh et al. (1998)), a spectral T21, three-level quasi-geostrophic model extended by estimates of the neglected ageostrophic terms in order to close the equations at the equator. Mainly due to the low atmospheric and oceanic resolution ECBilt does not reproduce interannual tropical variability associated with El Nino/Southern Oscilation (ENSO). The sea ice-ocean component Clio (Goosse et al., 1999; Goosse and Fichefet, 1999; Campin and Goosse, 1999) consists of a free-surface primitive equation ocean model with 3°x3° resolution coupled to a thermodynamic-dynamic sea ice model. The ocean model incorporates a mixed layer model following the scheme of Mellor and Yamada (e.g. Mellor and Yamada (1982), Kantha and Clayson (1994)), which is applied only to the fully turbulent regions such as the surface mixed layer. In regions below, a minimum (background) diffusivity κ_v is chosen, which follows a vertical profile similar to that used by Bryan and Lewis (1979). Above 1000 m κ_v takes values of 10^{-5} m^2/s, below κ_v increases up to a maximum value of 2×10^{-4}m^2/s towards the bottom. To avoid a singularity at the North Pole the oceanic component makes use of two subgrids: The first one is

Exp. name	**coupling**	κ_v **scaling factor**
A	fully coupled	1
B	uncoupled trop. Atl.	1
C	fully coupled	10^{-1}
D	uncoupled trop. Atl.	10^{-1}

Table 4.1: List of Experiments

based on the normal longitude and latitude coordinates and covers the whole ocean except for the North Atlantic ocean and the Arctic ocean. These are covered by the second spherical subgrid, which is rotated and has its poles at the equator in the Pacific (111° W) and the Indian Ocean (69° E). Instead of the standard pre-industrial set-up we mimic climate conditions of the last glacial maximum (LGM), 21,000 years ago. This allows for a more realistic assessment of the effects of Heinrich events on global climate. Glacial climate conditions are simulated by forcing the coupled model with the LGM ice sheet topography ICE-4G (Peltier, 1994), a LGM vegetation index (Crowley and Baum, 1997), reduced atmospheric CO_2 concentrations (200 ppm) and an albedo mask which captures the presence of glacial ice-sheets and changes in vegetation. Our experiments analysed here start from a 5000 year coupled spin-up run forced with glacial boundary conditions.

Similar LGM set-ups for the ECBilt-Clio model were used in Justino et al. (2005) and Timmermann et al. (2004, 2005a,b).

Our experiments simulate glacial meltwater events by delivering a pulse of anomalous freshwater (Figure 4.1) to the northern North Atlantic. The freshwater forcing is evenly distributed between 45°N and 60°N, attains a maximum of 1.3Sv and has a duration of 200 years. This corresponds to an integrated global sea level rise of about 8 m. In this study we focus on four different experiments as listed in table 1: Experiment A is performed under standard glacial settings while in experiments B the atmosphere and ocean are decoupled in the tropcial Atlantic between 30°N and 30°S. In experiments C and D the vertical background diffusion is reduced by one order of magnitude at all depth levels. Additionally experiment D also uses full air-sea coupling only outside of the tropical Atlantic ocean.

In Experiment B,D a decoupling technique is used which is very similar to the partial coupling method proposed by Wu et al. (2003). In the decoupled tropical Atlantic region the atmospheric component responds to climatological (glacial) SST and albedo forcing only. In response to this local climatological forcing, the atmosphere generates freshwater- and heat and momentum fluxes. Outside of the tropical Atlantic belt, full air-sea coupling is applied, meaning that the atmosphere can respond to the locally generated SST anomalies due to the AMOC shutdown.

4.3 Collapse and subsequent recovery of the AMOC

As can be seen in Figure 1, the freshwater-induced shutdown-phase of the AMOC is very similar for experiments A-D. The resulting negative salinity anomalies are advected poleward by the North Atlantic Current. This leads to the establishment of a subpolar/polar halocline, eventually suppressing deep convection in the northern North Atlantic. After about 150 years of simulation North Atlantic Deep Water (NADW) formation halts and Atlantic meridional overturning rates reduce from 24 Sv to 7 Sv (Figure 4.1). The associated reduction of the meridional heat transport results in a decrease of North Atlantic sea surface temperatures (SST) by up to 10°C and a large-scale cooling of the entire northern hemisphere (not shown). A subsequent resumption of NADW formation and recovery of the AMOC occurs in all four experiments, but on very different time-scales. Without tropical atmospheric response (experiment B) the recovery of the AMOC is delayed by 300 years as compared to the standard experiment A, thereby suggesting that tropical air-sea coupling is a very important negative feedback for the off-state. Comparing experiment A and C it becomes apparent that a reduction of the vertical diffusivity by a factor of ten does not influence the recovery of the AMOC in the fully coupled simulation. However, vertical diffusion is a very important factor in determining the AMOC recovery timescale in the experiments which neglect air-sea coupling in the tropical Atlantic (experiments B, D). In fact a change of the vertical diffusivity by a factor of 10 delays the complete recovery of the AMOC by about 500 years (Figure 4.1).

For the standard setting (experiment A), it was shown (Krebs and Timmermann, 2006a,b) that the resumption of NADW formation is induced by the wind-driven advection of anomalously saline waters into the subpolar North Atlantic. Figure 4.2 illustrates that after year 300 a positive salinity anomaly is generated in the northern tropical Atlantic due to a negative surface freshwater forcing (Figure 4.3). This freshwater flux anomaly originates from a decrease in precipitation caused by a southward shift of the intertropical convergence zone (ITCZ). This ITCZ shift is a typical feature of Heinrich events, as documented both, by paleo-proxy data (Hemming, 2004) and climate model

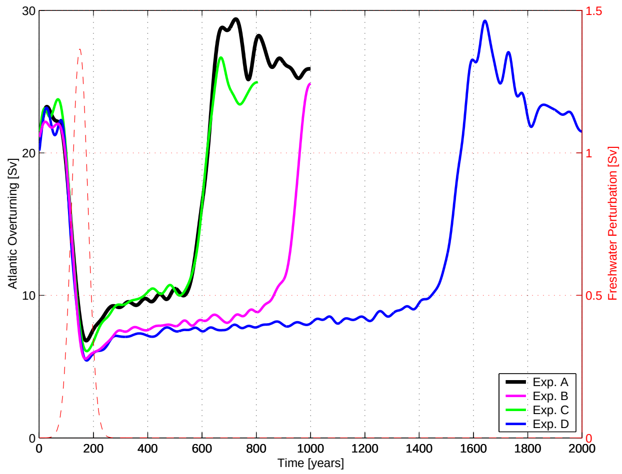

Figure 4.1: Freshwater induced collapse of the Atlantic meridional overturning circulation: time series of freshwater perturbation (right axis, dashed red) and annual mean of maximum Atlantic meridional transport for experiment A (black), experiment B (magenta), experiment C (green) and experiment D (blue) (all left axis, higher frequency variability is removed by a 50-year low-pass filter)

simulations (Zhang and Delworth, 2005; Timmermann et al., 2006). It can be attributed to a trade-wind intensification which can be decomposed into two parts: a linear atmospheric response to the tropical SST gradient (Gill, 1980) and an intensification of the Hadley circulation due to cooling in the extratropical North Atlantic (Broccoli et al., 2006; Krebs and Timmermann, 2006b). The important role of anomalous poleward salinity advection is further substantiated by the delayed AMOC recovery in the uncoupled experiment B (Figure 4.1). In experiment B no negative tropical freshwater flux anomaly is generated and tropical salinities increase at a much slower rate than in experiment A. The slow recovery of the AMOC in experiment B is initiated by the poleward advection of relatively weak positive salinity anomalies originating from the southern hemisphere (Figure 4.2).

A more detailed look into the salinity perturbations (Figure 4.3) will give us a better understanding of the different sensitivities of experiments A-D. In response to the freshwater perturbation the zonally averaged upper ocean salinity decreases by more than 4 psu (Figure 4.3). Comparing the zonally averaged salinity difference between experiment A and B, we find that tropical air-sea coupling in the Atlantic increases upper ocean salinities by more than 0.8 psu(Figure 4.3, lower left). Furthermore up-

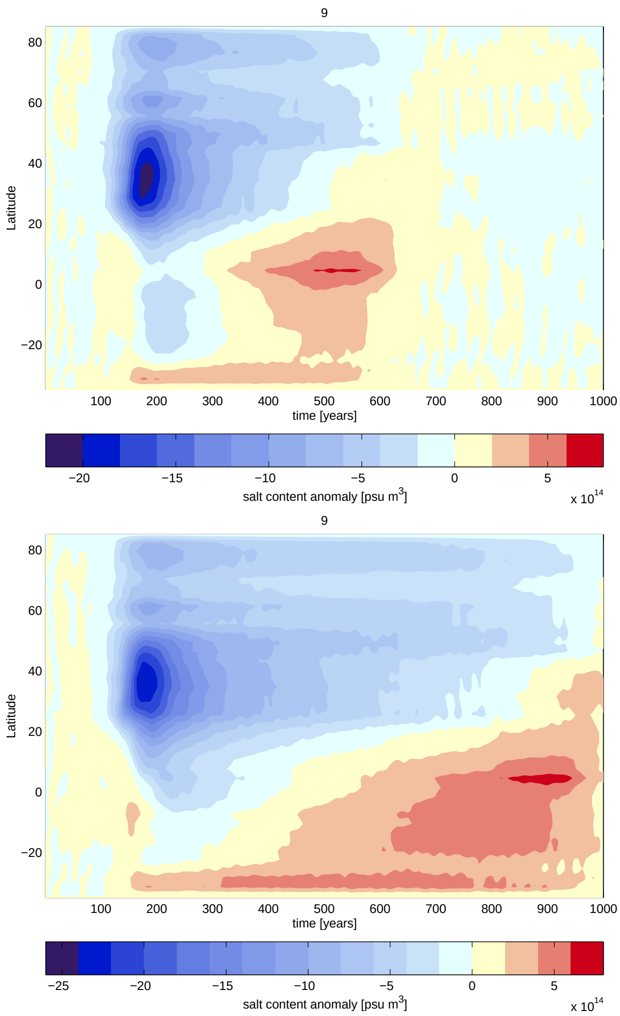

Figure 4.2: Hovmoeller diagram of zonally and vertically integrated salinity anomaly $[psum^3]$ over the top 250m, relative to the time mean of years 0-50, with full atmospheric response (exp. A) and without atmospheric response in the tropical Atlantic (exp. B) in the upper and lower panel respectively

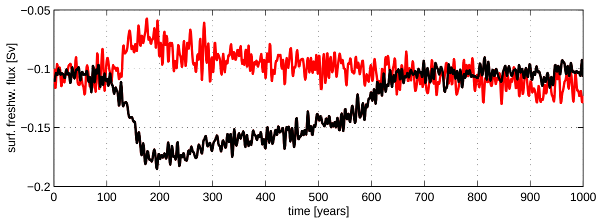

Figure 4.3: Total Atlantic air-sea freshwater flux into the ocean between equator and 30° N in Sv (5 year low-pass filtered) for fully coupled experiment A (black) and tropically uncoupled experiment B (red)

per ocean salinities can increase due to diffusive and advective vertical exchange with deeper, saltier water masses. The effect of increased vertical diffusion on the recovery of the AMOC can be assessed by computing the salinity difference between experiments A and C and experiments B and D (Figure 4.3, upper right, lower right panels).
In the fully coupled experiment A, where the surface freshwater lens is effectively removed by the positive salinity anomalies generated by tropical air-sea coupling, vertical diffusion is of minor importance. During model years 250-300 tropical air-sea coupling and enhanced vertical diffusion are responsible for a salinity increase of 0.8 and 0.2 psu, respectively (Figure 4.3, lower left, upper right). Clearly, tropical air-sea coupling provides a more efficient negative feedback for the off-state of the AMOC than an increase of the vertical diffusivity by a factor of 10. In experiment B the tropical feedback is missing and hence the recovery time is much longer than in experiment A. During the off-state of the AMOC, higher diffusion in experiment B compared to experiment D results in a salinity increase of more than 0.5 psu. Consequently the AMOC in experiment B recovers 600 years earlier than in experiment D. In experiments without a tropical ITCZ feedback, the sensitivity of the recovery time-scale to vertical diffusion can thus be attributed to the role of vertical diffusion in removing the surface freshwater lens.

4.4 Discussion

Our experiments suggest that the ITCZ response to the meridional temperature gradient destabilizes the freshwater induced AMOC off-state. In contrast, vertical diffusion does not have a pronounced impact on the recovery of the AMOC in our fully coupled simulations. However, when tropical air-sea coupling is suppressed, vertical diffusion becomes an important factor in restoring upper ocean salinities and in this case en-

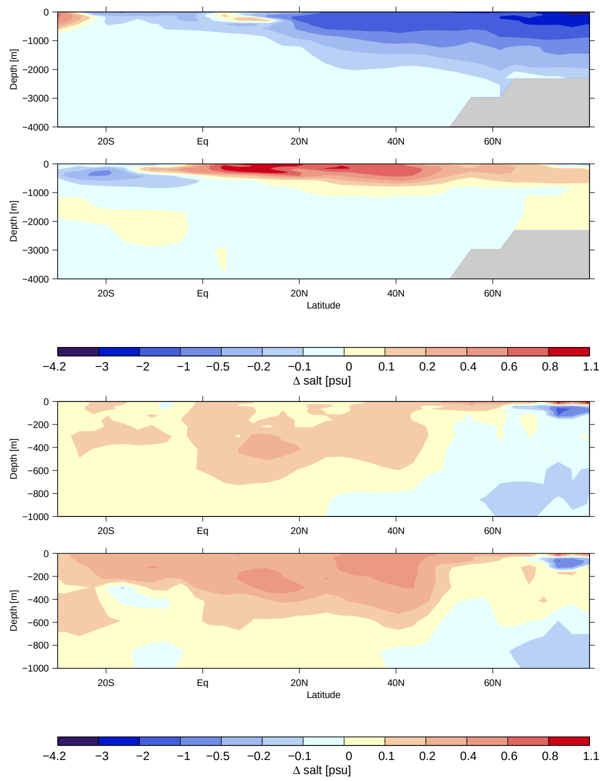

Figure 4.4: Mean zonally averaged Atlantic salinity differences [psu]: upper left: effect of the AMOC collapse: Exp. A (year 250-300 average) - Exp. A (year 0-50 average); lower left: effect of tropical atmospheric response: Exp. A - Exp. B (year 250-300 average); upper right: effect of vertical diffusion with tropical coupling: Exp. A - Exp. C (year 250-300 average); lower right: effect of vertical diffusion without tropical coupling: Exp. B - Exp. D (year 250-300 average)

hanced vertical diffusion accelerates the recovery process considerably. Thus, the strong sensitivity of the AMOC on vertical diffusion, described in Prange et al. (2003); Ganopolski et al. (1998); Schmittner and Weaver (2001), might result from the fact that tropical air-sea coupling in these models was not fully resolved.
Our results on the role of tropical air-sea interactions for the AMOC are also confirmed by recent modeling studies (Vellinga and Wu, 2004; Yin et al., 2006; Krebs and Timmermann, 2006b). A recent intercomparison of state-of-the-art CGCM waterhosing experiments (Stouffer et al., 2006) reveals that the tropical response simulated by the ECBilt-Clio model is somewhat underestimated. This may imply that the negative feedback provided by the tropical air-sea coupling in the Atlantic may be even more efficient in reality than in our coupled model.

Chapter 5

Synthesis

5.1 Summary

This study was designed to provide a better understanding of climate dynamics during a freshwater induced shut-down of the AMOC. The main emphasis was placed on the processes which may have controlled the rapid recovery of North Atlantic climate after glacial Heinrich events. Different proxy data indicate that glacial Heinrich events represent a collapse of the AMOC, caused by massive ice surges into the northern North Atlantic (Oppo and Lehmann, 1995; Vidal et al., 1997; Johnsen et al., 1995). The subsequent abrupt warming in the North Atlantic sector suggests a fast return to an active AMOC regime. While the AMOC shut-downs are commonly attributed to the ice surge events and the associated high latitude freshwater perturbations, the mechanisms which led to the resumption of the AMOC after such events are still relatively poorly understood. The abruptness of the observed warmings is reminiscent of the rapid transitions between multiple equilibria found in highly idealized models of the thermohaline circulation (Stommel, 1961; Marotzke et al., 1988; Stocker et al., 1992). However, the large scale ocean circulation relys on complex interactions with the atmosphere and local small-scale processes, which are subject to large uncertainties. Consequently, simulations with comprehensive climate models yield highly model dependent results with respect to the reversibility of a collapsed AMOC state and the recovery time-scale (Stouffer et al., 2006).

In this study the collapse and subsequent recovery of the AMOC has explicitely been analysed as a coupled air-sea phenomenon using a coupled climate model of intermediate complexity. The investigation of the AMOC recovery after glacial Heinrich events particularly focused on the following three major topics:

Processes responsible for the fast recovery of the climate system after glacial meltwater events The relatively fast recovery of the AMOC can be associated with two important processes.

- Shorty after the AMOC collapse, thermal processes establish weaker stratification in the northern North Atlantic which is particularly vulnerable to perturbations. During the weak overturning stage, the transport of warm upper ocean water into the sinking regions is strongly reduced. Relatively warm water at intermediate depths of about 800-1000m mixes with the cold water above. The associated upper ocean heat-flux reduces the sea-ice thickness and coverage. Overall, in the absence of meridional transports, mixing and heat diffusion homogenize the vertical temperature gradients in the water column and thereby pre-condition the system for buoyancy instabilities.
- During the collapsed AMOC state, the mean gyre circulation whichid primarily driven by the surface wind-stress transports positive surface salinity anomalies into the Nordic seas and the convection regions. This leads to a destabilisation of the water column and a resumption of deep ocean convection. Eventually, the reorganised vertical density profiles lead to a readjustment of the entire ocean circulation via Kelvin and Rossby wave adjustment (Kawase, 1987; Huang et al., 2000)

The role of atmospheric feedbacks for the evolution and climatic impact of Heinrich events By analyzing fully and partially coupled freshwater perturbation experiments under glacial conditions it is shown that modification of North Atlantic heat transport leads to cooling north of the thermal equator. The associated strengthening of the northeasterly trade winds and advection of cold air leads to a southward shift of the Intertropical Convergence Zone (ITCZ). Changes in accumulated precipitation lead to generation of a positive salinity anomaly in the northern tropical Atlantic and a negative anomaly in the southern tropical Atlantic. During the shut-down phase of the AMOC, cross-equatorial oceanic surface flow is halted, preventing a dilution of the positive salinity anomaly in the North Atlantic. Advected northward by the wind driven ocean circulation the positive salinity anomaly increases the upper ocean density in the regions of deep water formation, thereby accelerating the recovery of the AMOC considerably. In partially coupled experiments which neglect tropical air-sea coupling the AMOC recovery times are almost twice as long as in the fully coupled case. The impact of a shut-down of the AMOC on the Indian and Pacific ocean can be decomposed into atmospheric and oceanic contributions. Temperature anomalies in the northern hemisphere are largely controlled by atmospheric circulation anomalies, whereas those in the southern hemisphere are strongly determined by oceanic changes. An intensification of the Pacific meridional overturning cell in the northern North Pacific during the AMOC shut-down can be explained in terms of wind-driven ocean circulation changes acting in concert with global ocean adjustment processes.

The relative roles of tropical air-sea coupling and vertical diffusion on the stability characteristics of the AMOC Previous model studies suggest that vertical diffusion has a strong impact on the stability of the AMOC. This was systematically studied using quasi-equilibrated hysteresis experiments which, owing to computational costs, utilized simplified atmospheric models. However, little is known about the importance of vertical diffusion in transient climate simulations with considerable atmospheric feedbacks at work. The present work studied the relative roles of tropical air-sea coupling and vertical diffusion on recovery processes of the overturning circulation after a major freshwater-induced collapse of the AMOC. When atmospheric response is suppressed in the tropical Atlantic, the recovery of the AMOC is strongly controlled by vertical diffusion as it represents an important factor for the removal of upper ocean freshwater anomalies. In contrast, this sensitivity is absent in a fully coupled simulation where the tropical precipitation feedback is primarily responsible for the reestablishment of high upper ocean salinities in the North Atlantic. This suggests that meridional advection of anomalously saline tropical waters is a more efficient negative feedback during the off-state of the AMOC than density homogenisation due to vertical exchange processes.

The "conveyor belt" metaphor as a simple model for large scale ocean circulation suggests a coherently moving, buoyancy-driven oceanic circulation with an "Achilles heel" (Broecker, 1997) in the region of North Atlantic Deep Water formation. However, the present work demonstrates the entanglement of the oceanic conveyor with atmospheric feedbacks and demonstrates the necessity to understand the AMOC as a coupled air-sea phenomenon. The tropical Atlantic was identified as a region of important atmospheric feedbacks whose realistic representation appears to be essential for a successful simulation of both past and future climates.

5.2 Discussion and outlook

The climate model utilized in the present study (ECBilt-Clio) allows to analyze climate variations on centennial time-scale owing to relatively low computational costs compared to state-of-the-art general circulation models. However, its reduced complexity implicates some limitations, particularly arising from the relative coarse resolution and the simplified (though dynamic) atmospheric component. Small-scale ocean processes such as vertical mixing and deep water formation as well as the complexity of equatorial atmospheric dynamics can not be represented in full detail. Furthermore the climate sensitivity to glacial boundary conditions seems to be somewhat underestimated. Nevertheless, the simulated response to a glacial North Atlantic meltwater event (amplitude of North Atlantic cooling and southern hemispheric warming, southward ITCZ shift and recovery within centuries) is generally consistent with paleoclimatic reconstructions of Heinrich events (Hemming, 2004; Dahl et al., 2005). The influence of tropical atmospheric response on the AMOC stability has been confirmed by recent studies, which consistently find tropical air-sea interactions to represent a strong negative feedback for the AMOC (Yin et al., 2006; Vellinga and Wu, 2004). Moreover, the recent compilation of water-hosing experiments (Stouffer et al., 2006) suggests that the tropical temperature and ITCZ responses simulated by ECBilt-Clio are underestimated. This may imply that the negative feedback provided by the tropical air-sea coupling in the Atlantic may be even more efficient in reality than in the coupled model used in this study.

The assessment of spatially integrated density fluxes into the convection regions has allowed to identify the responsible process for the fast recovery of meridional overturning in the simulated Heinrich scenario. In addition the "blind atmosphere" technique has proven to be a powerful tool to distinguish and assess atmospheric and oceanic feedbacks and teleconnections. Both approaches could also provide valuable insight in the stability of the present-day climate and serve as benchmarks for model intercomparison projects such as Covey et al. (2003) and Stouffer et al. (2006).

The freshwater pertubation scenario in the present study was designed to represent an idealized, "typical" Heinrich event under constant LGM boundary conditions. However, amplitude, duration and location of glacial meltwater pulses are subject to large uncertainties (e.g. Hemming, 2004) and Heinrich events H1 - H6 are likely to have differed considerably from each other. For more realistic simulations it is also necessary to account for overlaying climate cycles and variations in glacial boundary conditions. The H1 Heinrich event coincides with southern hemispheric warming during the deglacia-

tion which possibly influenced the chronology of this particular event (Knorr, 2003). Futhermore, Heinrich events seem to be embedded in intermillenial climate cycles (so-called Bond cycles (Dansgaard et al., 1993; Bond et al., 1993). They appear at the end of a series of near millennial Dansgaard-Oeschger events (Dansgaard et al., 1993). These so-called Bond-cycles are characterised by a succession of progressively cooler relatively warm periods (interstadials). The interplay between Heinrich events and D-O cycles is not well understood, in terms of triggers and responses. Ice surges might not only be associated with the internal instability of continental ice sheets (MacAyeal, 1993) but also be triggered by changes in sea level and oceanic heat transport (Lekens et al., 1005; Flückiger et al., 2006).
Hence coupled climate models which incorporate the effect of sea-level changes on the continental ice sheets may help to investigate the mechanisms controlling glacial melt-water events (e.g. Philippon et al., 2006).
Comparisons of paleoceanographic data with ensemble simulations of glacial Heinrich scenarios may additionally help to constrain boundary conditions and background climate. However, the transfer function required to gain physical climate variables (such as temperature) from proxy data (such as oxygen isotopes) usually depends on various other variables, which are particularly uncertain for periods of dramatic climate change. On this account the opposite approach of directly simulating paleoceanographic proxies (e.g. Roche et al., 2004) - though still in its infancy - may yield valuable insight in the near future.

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Erklärung

Meine Abhandlung zur Promotion ist – abgesehen von der Beratung durch meine akademischen Lehrer – nach Inhalt und Form meine eigene Arbeit. Sie hat weder ganz noch teilweise hier oder an anderer Stelle im Rahmen eines Prüfungsverfahrens vorgelegen.
Kapitel 2 bis 4 basieren auf Manuskripten, die bei wissenschaftlichen Fachzeitschriften eingereicht wurden.

Kapitel 2:

Krebs, U. and A. Timmermann (2006), Fast advective recovery of the Atlantic meridional overturning circulation after a Heinrich event, *Paleoceanography*, accepted for publication.

Kapitel 3:

Krebs, U. and A. Timmermann (2006), Tropical air-sea interactions accelerate the recovery of the Atlantic Meridional Overturning Circulation after a major shutdown, *Journal of Climate*, submitted.

Kapitel 4:

Krebs, U. and A. Timmermann (2006), The relative effects of vertical diffusion and tropical air-sea coupling on the recovery of the AMOC, *Geophysical research Letters*, submitted.

Southampton, 30. Oktober 2006

(Uta Krebs)