学位論文（要約）

Impact of glacial ice sheets on the Atlantic meridional overturning circulation and climate
(氷期氷床が大西洋子午面循環と気候に与える影響に関する研究)

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Doctoral Dissertation

Impact of glacial ice sheets on the Atlantic meridional overturning circulation and climate

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Abstract

During glacial periods, the North American and Eurasian ice sheets grew across the continents (hereafter referred to as glacial ice sheets). These glacial ice sheets are suggested to have an impact on the Atlantic meridional overturning circulation (AMOC), which played a role in affecting the mean climate and inducing frequent climate changes (Dansgaard-Oeschger cycles; DO cycles) during glacial periods. However, the impact of glacial ice sheets on the AMOC is not fully understood. In particular, (i) processes by which the glacial ice sheets modify the AMOC (surface winds, surface cooling and atmospheric freshwater flux), (ii) the impact of glacial ice sheets on the DO cycles, for example, the duration of stadials, and (iii) the reason why the impact of glacial ice sheets on the AMOC depends on background climate (CO2) should be clarified. The objective of this study is to improve the understandings of the impact of glacial ice sheets on the AMOC and glacial climate using a general circulation model with different configurations.

In Chapter 2, the effect of wind change due to glacial ice sheets on the AMOC and the crucial region where the wind modifies the AMOC is explored. For this purpose, numerical simulations with an atmospheric general circulation model (AGCM) and an ocean general circulation model (OGCM) are conducted. First, from AGCM experiments, the effect of glacial ice sheets on the surface wind is evaluated. Second, from OGCM experiments, the influence of the wind stress change on the AMOC is evaluated by applying wind stress anomalies regionally or at different magnitudes as a boundary condition. These experiments demonstrate that glacial ice sheets intensify the AMOC through an increase in the wind stress at the North Atlantic mid-latitudes, which is induced by the North American ice sheet. This intensification of the AMOC is caused by the increased oceanic horizontal and vertical transport of salt. Changes in winds by the Eurasian ice sheet also appear to have an impact on the AMOC, though its effect is smaller compared to the North American ice sheet.

In Chapter 3, I investigate the impact of glacial ice sheets on the duration of stadials during mid-glacial. For this purpose, freshwater hosing experiments are conducted with an atmosphere-ocean coupled general circulation model (AOGCM) under several ice sheet configurations. I find that the expansion of glacial ice sheets during mid-glacial shortens the duration of stadials (recovery time of the AMOC). In order to explore the reason why glacial ice sheets modify the recovery time, partial decoupled experiments are conducted. In these experiments, atmospheric conditions that force the ocean model are replaced to different forcing. Thus the role of changes in surface winds, surface cooling and atmospheric freshwater flux by the glacial ice sheets on the recovery time can be estimated. These experiments show that differences in the surface wind is important in causing the shorter recovery under larger ice sheets, while differences in the surface cooling has an opposite effect. The impact of atmospheric freshwater flux appeared to be small. The wind shortens the recovery time through mainly increasing the surface salinity at the deepwater formation region, while surface cooling increased the recovery time through increasing the sea ice at the deepwater formation region. Thus these results show that differences in the surface wind induced by the glacial ice sheets play an important role in causing frequent DO cycles during mid-glacial through shortening the duration of stadials.

In Chapter 4, I explore the role of sea ice in modifying the impact of glacial ice sheets on the AMOC. For this purpose, results of AOGCM simulations are analyzed and numerical experiments are conducted using an AGCM and an AOGCM. It is shown that the expansion of sea ice over the northern North Atlantic weakens the surface wind. This is associated with a suppression of oceanic sensible heat flux due to the sea ice, which drastically increased the static stability of the air column. As a result, the surface wind cannot become strong even under the existence of glacial ice sheets and hence the wind effect of the ice sheet weakens. This modifies the relative strength of the wind effect and cooling effect of the glacial ice sheet. As a result, the cooling effect becomes stronger and thus the glacial ice sheet weakens the AMOC.

Together all, this study clarifies that the impact of glacial ice sheets on the AMOC is controlled by the relative strength of the surface winds and surface cooling. The relative strength of these effects depends on sea ice. Under warm climate and less sea ice condition, the glacial ice sheets can intensify the surface winds and the AMOC. However, under cold climate, sea ice expands and suppresses the atmosphere-ocean heat exchange. As a result, the surface winds cannot become vigorous even under the existence of the glacial ice sheets. Thus, the relative strength of wind and cooling effects changes. Since the strength of each effect depends on models, differences in the relative strength of the wind and cooling effects among models can cause a large difference in the AMOC among them. In addition, this study also suggests an important role of the glacial ice sheet on the frequent DO cycles during glacial periods. According to ice core data, the duration of the stadial is shortest during the mid-glacial period compared to the full glacial and earlier glacial periods. Over the early glacial to mid-glacial periods, the expansion of glacial ice sheet can reduce the duration of the stadials through strengthening the surface winds since the climate is relatively warm and there is less sea ice over the North Atlantic. Over the mid-glacial to full glacial periods, the North Atlantic is widely covered by sea ice due to the cold climate. As a result, the expansion of glacial ice sheet during this period can increase the duration of stadials through the cooling effect. Thus, this study suggests that the glacial ice sheet plays an important role in causing frequent DO cycles during mid-glacial period through modifying the duration of stadials.

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Contents

Chapter 1

General Introduction

Over the Quaternary, configurations of ice sheets have changed dramatically in response to insolation changes and associated nonlinear feedbacks (Fig. 1.1, e.g. Lisiecki and Raymo 2005, Abe-Ouchi et al. 2013). During Quaternary interglacial times, the ice sheets laid over the Greenland and Antarctica. These ice sheets exerted a large impact on the interglacial (modern) atmospheric and oceanic circulation (e.g. Petersen et al. 2004, Hakuba et al. 2012, Schmittner et al. 2011, Gong et al. 2015). During Quaternary glacial times, the North American (Laurentide/Cordilleran, e.g. Stokes et al. 2012) and Eurasian ice sheets grew individually across the continents (hereafter referred to as glacial ice sheets). Figure 1.2 shows a reconstruction of the glacial ice sheets over the last glacial period with an ice sheet model (Abe-Ouchi et al. 2013). The North American ice sheet first initiated at the northeastern North America. This is the region where a cold air flows from the Arctic, thus the temperature is lower compared to other regions in the North America (e.g. Brayshaw et al. 2009). After its initiation, the ice sheet expanded both westward and southward and finally combined with the Rockie at the Last Glacial Maximum (LGM, approximately 21,00 years ago). With respect to the Eurasian ice sheet, it first initiated near Kara Sea and expanded both meridionally and zonally. At the LGM, the western edge of the Eurasian ice sheet reached the United Kingdom (e.g. Abe-Ouchi et al. 2007). Since these glacial ice sheets are extensive and have a high altitude (approximately 3000-meter for the North American ice sheet and approximately 2000-meter for the Eurasian ice sheet), it is very likely that these glacial ice sheets exerted a large impact on the glacial climate, e.g. atmospheric and ocean circulations. Thus, in order to understand the climate change and climate system over the Quaternary, it is essential to understand how changes in glacial ice sheets modify the climate. In this dissertation, I particularly investigate how changes in glacial ice sheets modify the Atlantic meridional overturning circulation (AMOC) and climate.

In this General Introduction, I first describe differences between glacial and interglacial climate periods. Then I explain the importance of the AMOC on understanding the glacial climate. After that, I introduce previous atmospheric studies, which is important for the AMOC because changes in atmospheric forcing have a large impact on the AMOC. Subsequently, I review previous oceanic studies, which explored the dynamics of glacial AMOC and review recent coupled modeling studies. Lastly, I propose the objective of this study.

Differences between glacial and interglacial periods

Due to the expansion of glacial ice sheets and the lowering of greenhouse gases, especially the CO2, the glacial climate differed substantially from the interglacial (modern) climate mainly in two points (Fig. 1.1, e.g. CLIMAP 1981, Dansgaard et al. 1993, MARGO 2009). The first point is the surface air temperature. The climate of the LGM (e.g. Braconnot et al. 2012) has been studied as the representative of the glacial climate since it is the closest and largest glacial time over the last glacial period (e.g. CLIMAP 1981, MARGO 2009). During the LGM, the global average land temperature was lowered by 6.1 ($\pm$1.5) ˚C (Bartlein et al. 2011) and the global average sea surface temperature was lowered by 2.1 ($\pm$1.8) ˚C (Waelbroceck et al. 2009) compared to Preindustrial. This cooling was larger especially over the mid-high latitudes. The land temperature over mid-high latitudes was 7.5 ($\pm$1.6)˚C lower and the sea surface temperature was 3.2 ($\pm$2.7)˚C lower (Bartlein et al. 2009, Waelbroeck et al. 2009). In the tropics, land and sea surface temperature was lowered by 3.1 ($\pm$1.2)˚C and 1.5 ($\pm$1.2) ˚C, respectively.

The second point is the stability of the climate (e.g. McManus et al. 1999, Kawamura et al. 2017). Many people were surprised to find large and frequent fluctuations of δ^{18}O in the Greenland ice core during the glacial period. These climate changes are known as the Dansgaard-Oeschger cycles (DO cycles, Fig. 1.1, e.g. Dansgaard et al. 1993). Several studies first doubted whether these records represent true climate changes. However, similar signals were reported from different ice cores and proxies, thus people believed the existence of DO cycles as true climate signals (e.g. Johnsen et al. 1992). The DO cycles are characterized by an abrupt warming of 5˚C~16˚C within several decades (20~100 years;

Kindler et al. 2014, Orsi et al. 2014). After this abrupt warming, the climate first gradually cools and then returns abruptly to the cold period. The warm period is called as interstadials and the cold period is called as stadials. One DO cycle consists from these two climate modes. DO cycles are also famous for its global scale (e.g. Rahmstrof 2002, Clement and Peterson 2008). During the stadials, not only the Greenland, but also the entire North Atlantic cooled (e.g. Bond et al. 1993, Eliot et al. 2002). On the contrary, the temperature over the Antarctica increased (e.g. EPICA Community members 2006). In the tropics, proxies suggested a southward shift of Intertropical Convergence Zone (ITCZ) in the Atlantic. Schmit et al. (2006) showed with coupled Mg/Ca and $\delta^{18}O$ measurements in the western subtropical Atlantic that the salinity increased during stadials, while it decreased during interstadials. On the other hand, in the southern tropics, Wang et al. (2004) showed from speleothem in Brazilian cave that this region was wet during stadials. In addition, it has been widely shown from speleothem record from Timta and Hulu cave that the strength of summer Indian and east Asian monsoon reduced during stadials (e.g. Sinha et al. 2005, Wang et al. 2001).

As above, the climate of glacial period differed substantially from that of interglacial and modern era. However, the cause of the difference of climate between glacial and interglacial periods is not still fully understood. In particular, the following points remained unclear.

1. Why do DO cycles occur during glacial period, but not during interglacial period? What is the role of CO2 and glacial ice sheets?
2. What is the trigger of the abrupt climate shift?
3. Why are the transitions of DO cycle so abrupt?

In this dissertation, I mainly focus on the first question, though I plan to explore the other questions in the future.

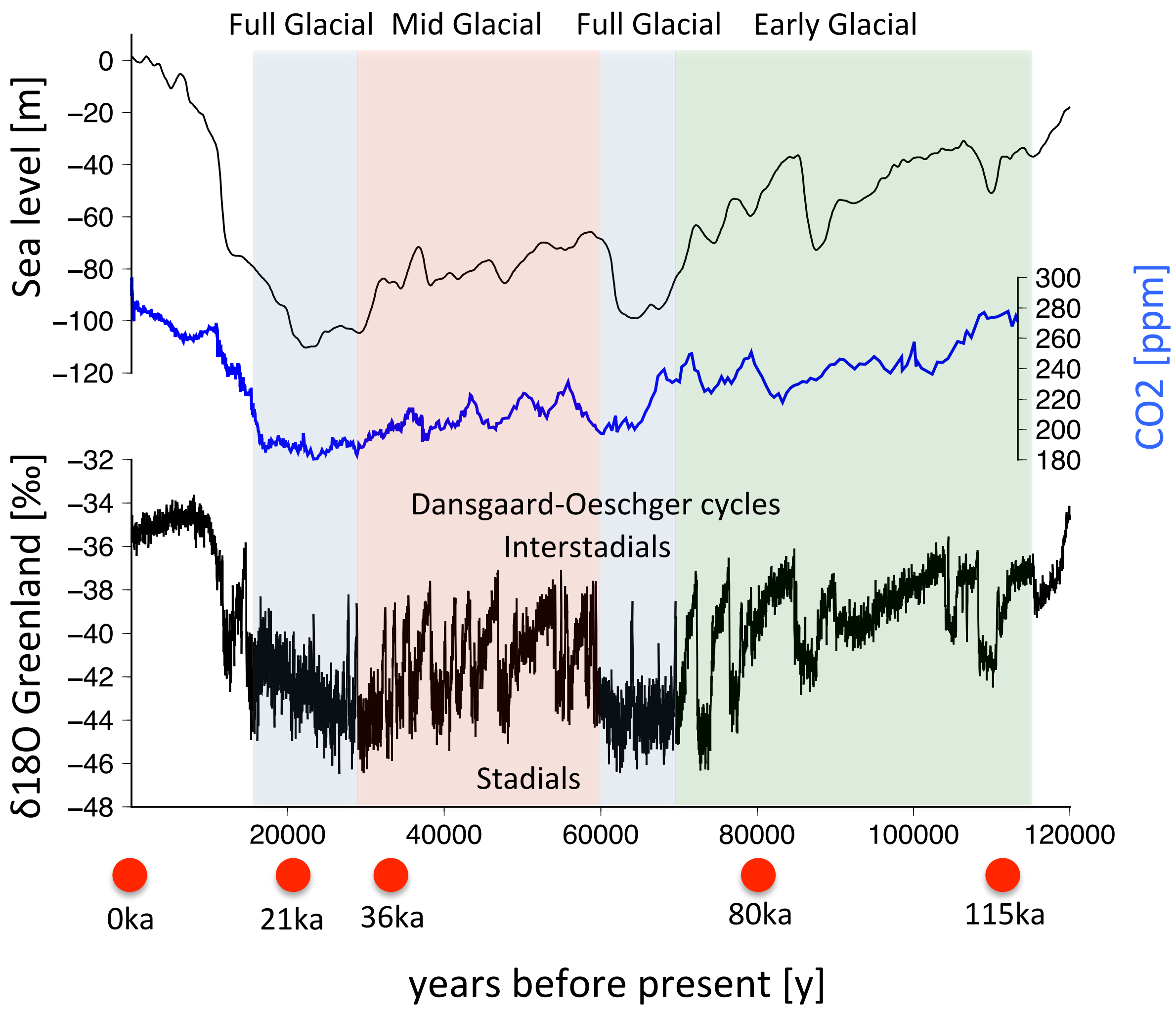

Fig. 1.1 Changes in ice volume (sea level, Grant et al. 2012), CO_2 (Beleiter et al. 2015) and Greenland ice core $\delta^{18}O$ (a proxy of temperature, Rasumussen et al. 2012) over the last glacial-interglacial cycle. Color shade separates the period within one glacial period. Blue: Full glacial, red: mid glacial and green early glacial. During full glacial, the glacial ice sheets are most extensive and CO_2 is low. During mid glacial, the glacial ice sheet is relatively smaller than the full glacial and the CO_2 is relatively higher than full glacial. During early glacial, the glacial ice sheet is smallest and the CO_2 is highest among other glacial periods. Red dots below the horizontal axis show the period whose reconstructions of glacial ice sheet are shown in Fig. 1.2.

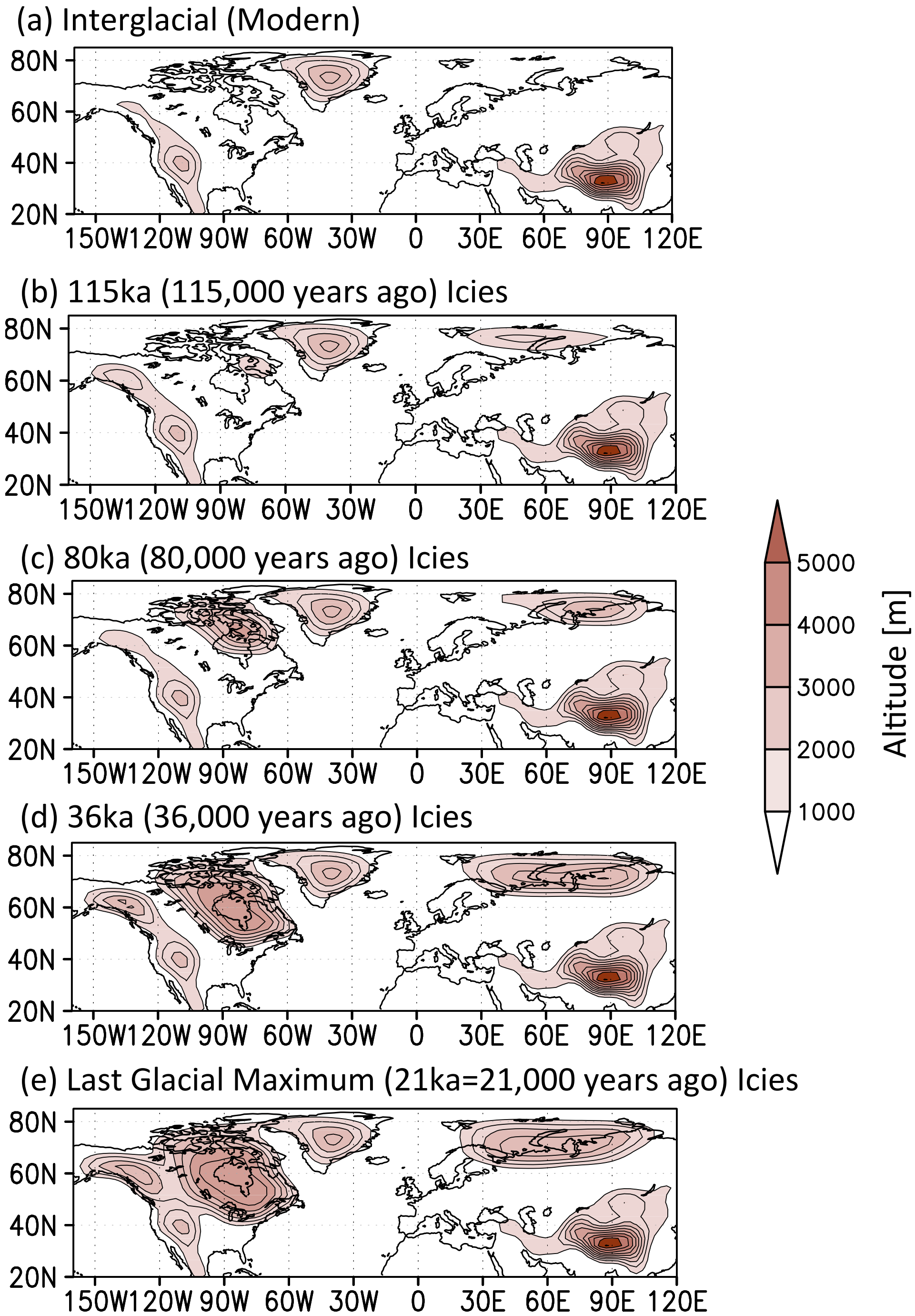

Fig. 1.2 Northern Hemisphere topography of (a) Interglacial (modern), (b) 115,000 years ago, (c) 80,000 years ago, (d) 36,000 years ago and (e) Last Glacial Maximum (21ka, 21,000 years ago). During the glacial period, huge ice sheets existed over North America and northern Europe. The ice sheet configuration of (b) – (e) is reconstructed with an ice sheet model Icies (Abe-Ouchi et al. 2007, 2013).

Importance of the AMOC on the climate

Changes in the AMOC are often considered to be important for the glacial climate because of its great impact on the global climate and deep ocean properties (e.g. Kulbroat et al. 2007). The AMOC is a large-scale ocean circulation in the Atlantic basin. It has a northward flow near the surface, which transports warm and salty water from the tropics and Southern Hemisphere to the northern North Atlantic. After reaching the northern North Atlantic, this northward flow releases the heat to the overlying cold atmosphere. As a result, the surface water loses buoyancy due to an intense cooling and eventually sinks and returns south at the depth of 2000-3000 meters, which is called the North Atlantic Deep Water (NADW). Below the NADW, a water mass originating from the Antarctica and Southern Ocean (Antarctica bottom water, AABW) occupies the bottom sea. In the modern climate, the NADW forms at the Greenland Sea, Labrador Sea and Irminger Sea (Killworth, 1983, Marshall and Schoot 1999, Pickart et al. 2003). The circulation closes as the NADW returns to the surface by diffusive processes and strong wind-driven upwelling in the Southern Ocean, due to the Drake Passage effect (e.g. Kulbroat et al. 2007). In total, since the AMOC transports warm water near the surface and cold water at depth, the AMOC plays an important role on the climate through meridional heat transport.

Proxy studies have suggested that the AMOC experienced large changes during glacial periods. Using $\delta^{13}C$ in benthic foraminifera in the Atlantic Ocean, Duplessy et al. (1988) showed that the AMOC was shallower during the LGM compared to that of modern period. Using $^{231}Pa/^{230}Th$, which is a proxy of the intensity of the past AMOC, from sediments recovered in Bermuda Rise (33˚ 41'N, 57˚37'W), McManus et al. (2004) showed that the AMOC was weaker during the LGM compared to modern period. Bohm et al. (2015) reconstructed changes in the AMOC over the last glacial period with combination of radiogenic neodymium (Nd) isotopes and particulate $^{231}Pa/^{230}Th$ from sediments recovered in Bermuda Rise (33˚ 41'N, 57˚37'W). They showed that the AMOC was weak during the LGM, while it remained vigorous most of the last glacial period. Also, using neodymium isotopes, which are considered to reflect the changes in source region of the deepwater formation, Piotrowki et al. (2005) showed that the origin of the deepwater in the South Atlantic switched between north (North Atlantic) and south (Antarctic) in concert with some of DO cycles. Bohm et al. (2015) also reported a significant weakening of the AMOC during Heinrich events (stadials). Kissel et al. (2008) showed that the strength of the NADW was perfectly concurrent with Greenland temperature from magnetic properties at various sites in the North Atlantic: the AMOC was vigorous during interstadial, while it was weak during stadial. To summarize, reconstruction studies showed that the AMOC was shallower and weaker during the LGM compared to modern interglacial. Before the LGM, reconstruction studies showed that the AMOC was vigorous and had a large variability, whose change was linked to changes in Greenland temperature during DO cycles.

From the modeling perspective, Ganopolski et al. (1998) explored the impact of changes in the AMOC on the LGM climate with an atmosphere-ocean-sea ice coupled climate model. By comparing results of the climate model and an atmospheric model coupled to a slab ocean, which assumes no changes in the oceanic heat transport, they showed that the weakening of the AMOC in the LGM played an important role in decreasing the temperature and in inducing southward expansion of sea ice over the North Atlantic. Modeling studies have also shown that the drastic weakening of the AMOC could reproduce climate changes that captures characteristics of DO cycles, e.g. cooling over the North Atlantic, warming over the Antarctic, southward shift of the ITCZ and a weakening of the Monsoon (e.g. Manabe and Stouffer 1995, Zhang and Delworth 2005, Kageyama et al. 2009). For example, assuming a large discharge from glacial ice sheets to the northern North Atlantic, which is considered as the trigger of DO cycles, Manabe and Stouffer (1995) conducted freshwater hosing experiments under modern climate. They showed that the freshwater hosing could drastically reduce the AMOC through decreasing the surface salinity and thus shutting down the deepwater formation. This drastic reduction of the AMOC induced global climate changes that resembled DO cycles. Menviel et al. (2014) and Zhang et al. (2014) further showed that the weakening of the AMOC due to freshwater hosing induced global climate changes under glacial background conditions. Therefore, modeling studies supported that changes in the AMOC played a crucial role in the glacial climate.

Theoretical and modeling studies have also shown that the AMOC has a very strong nonlinearity (e.g. Stommel 1961, Oka et al. 2012). This is important when considering nonlinear climate changes such as DO cycles. The nonlinearity of the AMOC is associated to a positive feedback of salt transport by the AMOC itself and a shutdown of deepwater formation regions. When the AMOC is in a vigorous mode, the northward flow by the AMOC itself transports saline subtropical water to the deepwater formation region (e.g. Oka et al. 2001, Timmerman and Goose 2004). This increases the salinity and density at the deepwater formation region, and thus stabilizes the strong AMOC. On the other hand, when the AMOC weakens, the weakening of the AMOC reduces the northward salt transport, which further weakens the AMOC. Once the weakening of the AMOC passes a threshold, which is associated to a shutdown of deepwater formation (e.g. Ganoposlki and Rahmstrof 2001, Kageyama et al. 2009, Oka et al. 2012), the AMOC drastically weakens and shifts into a weak mode. Thus, the strength of the AMOC can change nonlinearly to a subtle forcing when the AMOC is close to this threshold (e.g. Oka et al. 2012).

Based on the evidence from proxies and modeling studies, the AMOC is considered to play an important role in the glacial climate (very low surface air temperature at the northern North Atlantic during the LGM and DO cycles). Thus, a better understanding of glacial AMOC is critical for the understanding of glacial climate. Below, I first review previous studies those investigated glacial atmospheric conditions with atmospheric models. This is important since changes in atmospheric condition have a large impact on the AMOC.

Changes in the atmospheric circulation during glacial periods; role of glacial ice sheets

Previous studies explored the impact of glacial boundary conditions on the atmosphere with atmospheric general circulation models (AGCM, e.g. Manabe and Broccoli 1985, Kageyama et al. 1999, Hofer et al. 2012). In these studies, they conducted simulations of the LGM, because reconstructions of glacial ice sheet topography, sea surface conditions and other boundary conditions were available (e.g. CLIMAP 1981, Peltier 1994, Peltier et al. 2004, Abe-Ouchi et al. 2015). These studies showed that the existence of glacial ice sheets induced a large cooling over the Northern Hemisphere, especially over the glacial ice sheets and North Atlantic (e.g. Manabe and Broccoli 1985). The cooling over the ice sheets was induced by its high albedo and surface cooling due to the ice sheet, while cooling over the North Atlantic was induced by surface winds from the North American ice sheet. Manabe and Broccoli (1985) further showed that this cooling was amplified by the formation of sea ice.

With respect to the wind field, atmospheric modeling studies showed that the glacial ice sheets exerted a large impact on the large-scale atmospheric circulation. Manabe and Broccoli (1985) showed that the westerly intensified over the North Atlantic, while it split into two branches over the North America due to the existence of the North American ice sheet. The intensification and split of the westerly was associated to an anti-cyclonic circulation anomaly over the North American ice sheet and a cyclonic circulation anomaly downstream of the North American Ice sheet (Broccoli and Manabe 1987). Pausata et al. (2011) showed that the glacial ice sheets intensified the surface wind over the North Atlantic and thus showed that the wind anomaly has a barotropic structure (which means that the sense of the anomaly does not change with height). Merz et al. (2015) further revealed that the glacial ice sheets modified the maintenance mechanism of the westerly over the North Atlantic; it was maintained by convergences of zonal momentum by the stationary wave itself. This is different from the westerly in the modern climate, which is maintained by convergences of zonal momentum by transient waves (e.g. Hoskins et al. 1983, Brayshaw et al. 2008, 2009). The expansion of glacial ice sheets also modified the location of the westerly in the North Atlantic. For example, Laine et al. (2009) and Beghin et al. (2015) showed that the expansion of glacial ice sheets shifted the westerly to the south in some models in

Paleoclimate Model Intercomparison Project 2 and 3 (Braconott et al. 2007, 2012). However, they also showed that the meridional shift of the westerly depended on the model.

Precipitation field was also modified by the glacial ice sheets. Hofer et al. (2012) showed that the glacial ice sheets reduced the precipitation over the northern North Atlantic. They associated the decrease in precipitation with a reduction in humidity due to the lower temperature and the southward shift of the westerly over the northern North Atlantic.

Cook and Held (1988) explained the atmospheric circulation anomaly induced by the glacial ice sheets based on the linear theory (e.g. Hoskins and Karoly 1981). According to the linear theory, when the westerly flow hits a mountain, an anti-cyclonic circulation anomaly appears above the mountain and a cyclonic circulation anomaly appears downstream of the mountain. This is associated with vertical velocity anomalies, which are induced when the westerly flows over the mountain. In the upstream of the mountain, upward vertical velocity anomaly is induced as the westerly flows over the mountain. This upward motion shrinks the thickness of the air column. As a result, anti-cyclonic circulation anomaly is generated in order to conserve a quantity called potential vorticity. The potential vorticity in a homogeneous incompressible fluid can be expressed as follows:

$$potential\ vorticity = \frac{\eta}{h}.$$

Here, η indicates absolute vorticity and h indicates the thickness of air column. Under adiabatic and frictionless condition, the potential vorticity is conserved. Thus, when h decreases due to upward motion near the surface, so does η. Since a decrease of η corresponds to a generation of anti-cyclonic circulation anomaly, an anti-cyclonic circulation anomaly appears over the mountain. In the downstream of the mountain, a cyclonic circulation anomaly is induced in a similar manner. Since the North American ice sheet works as an orographic forcing on the westerly, the westerly that flows over the mountain can induce an anti-cyclonic circulation anomaly upstream of the ice sheet and a cyclonic circulation anomaly downstream of the ice sheet. Cook and Held (1988) confirmed this through a simulation with a linear model, in which they applied an orographic forcing under glacial background conditions (zonal mean temperature and zonal mean zonal wind simulated in an AGCM (Broccoli and Manabe 1987)). Roe and Linden (2001a, b) also supported their idea.

Liakka (2012) further showed that the surface cooling over the ice sheet played an important role in modifying the wind field; it intensified both the anti-cyclonic circulation anomaly over the ice sheet and the cyclonic circulation anomaly downstream (e.g. Ringler and Cook 1999). They explained the reason as follows. In their model, surface cooling induced a high-pressure anomaly over the ice sheet. This itself amplified the high-pressure anomaly induced by the topography of the ice sheet. In addition, the thermally forced high-pressure anomaly increased the westerly over the ice sheet. As a result, the vertical velocity increased and thus the cyclonic circulation anomaly increased. Thus, these studies showed that the glacial ice sheets enhanced the atmospheric circulation over the North Atlantic through the topography and cooling effects.

Above atmospheric studies demonstrated that glacial atmospheric conditions differed substantially from those of interglacial; stronger surface cooling, less precipitation and stronger winds over the North Atlantic. Therefore, it is very likely that changes in atmospheric conditions during glacial period affected the AMOC through changes in the surface cooling, winds and atmospheric freshwater flux (evaporation – precipitation – river runoff, hereafter E-P-Runoff).

What causes changes in the AMOC during glacial period?

Previous studies have first explored impacts of glacial atmospheric conditions on the AMOC mainly with an earth system model with intermediate complexity (EMIC) due to lack of computational resources. In EMICs, a very simple atmospheric model is coupled to a relatively sophisticated oceanic and sea ice component. In most EMICs, they use an energy balance model in their atmospheric component (e.g. Ganopolski et al. 1998). In the energy balance model, the wind is often fixed to modern observations and the advection of heat and water vapor is treated as diffusion. Thus, changes in atmospheric condition during glacial period are treated very crudely. Using an EMIC, Ganopolski et al. (1998) conducted an experiment of LGM. In their LGM simulation, the NADW formation region shifted south, which is associated with the southward expansion of sea ice. As a result, they obtained a shallower AMOC in their LGM simulation compared to that in their modern climate simulation, which is consistent with proxies (e.g. Bohm et al. 2015). They showed that the stronger surface cooling was important for the shallower AMOC. The importance of surface cooling on the AMOC was also reported in other EMIC studies (e.g. Ganopolski and Rahmstorf 2001, Loving and Vallis 2005, Colin de Veriere and the Raa 2010, Arzel et al. 2010, 2012). On the other hand, Schmittner et al. (2002) showed that changes in atmospheric freshwater flux (E-P-Runoff) was important in modifying the glacial AMOC. In this study, they showed that a decrease in atmospheric freshwater flux (E-P-Runoff) over the Atlantic, thus more freshwater input in the Atlanitc, weakened the AMOC in their LGM simulation. In addition, Schmittner et al. (2003) showed that changes in freshwater flux induced by sea ice formation over the Southern Ocean played an important role in glacial deep ocean circulation. However, since these studies used climate models with a simple atmospheric component, the effect of realistic changes in atmospheric conditions on the AMOC remained elusive.

Over the last decade, simulations of the LGM were conducted with AOGCMs in the context of PMIP (Braconnot et al. 2007, Kitoh et al. 2001, Shin et al. 2003). Thanks to this project, the impact of glacial atmospheric conditions and boundary conditions (glacial ice sheets and CO2) on the AMOC could be investigated in a realistic framework. However, another problem emerged. The response of the AMOC to LGM boundary conditions diverged among models. In PMIP2 (Otto-Bliesner et al. 2007, Weber et al. 2007), half of the models simulated a weaker and shallower AMOC in their LGM simulation compared to Preindustrial climate simulation, while other half showed a stronger and deeper AMOC in their simulation of the LGM. The simulation of the LGM was updated in PMIP3 (Bracconot et al. 2012). This time, all of the models in PMIP3 simulated a stronger AMOC in their simulations of the LGM compared to those of the Preindustrial. Thus, most AOGCMs failed to reproduce a weak and shallow LGM AMOC suggested by reconstructions (e.g. McManus et al. 2004, Bohm et al. 2015).

In order to understand the cause of the changes in the AMOC obtained in glacial climate simulations (LGM) with AOGCMs, two methods have been applied. The first method was a sensitivity experiments with a decoupled OGCM (Oka et al. 2012, Sun et al. 2016). In this method, one could apply atmospheric conditions simulated in an AOGCM as a boundary condition to the OGCM. Thus it helped to identify the important processes controlling the AMOC, which was complex in the AOGCMs due to the coupling process. By applying the surface condition computed in the Preindustrial and LGM simulations simulated in an AOGCM as a boundary condition to the OGCM, Oka et al. (2012) found that the stronger global cooling and glacial surface wind played an important role in modifying the glacial AMOC. They showed that the stronger surface cooling weakened the AMOC, while glacial winds intensified the AMOC. They suggested that the cooling weakened the AMOC through increasing the sea ice over the northern North Atlantic, which insulated the heat exchange between the atmosphere and ocean and suppressed the NADW formation. For the glacial winds, they suggested that changes in salt transport by wind-driven gyre circulation and wind-driven sea ice transport were important. From these, they suggested that stronger surface cooling or weaker winds in LGM simulations in AOGCMs might be important in simulating a weak LGM AMOC. However, since Oka et al. (2012) used the Preindustrial and LGM atmospheric conditions computed in the AOGCM, the important climate boundary conditions (e.g. CO2, glacial ice sheets), which modified the AMOC through these processes, remained elusive.

The second method was to assess the individual impact of glacial boundary conditions (e.g. CO2 and glacial ice sheets) on the AMOC through AOGCM sensitivity experiments. Stouffer and Manabe (2003) showed that the lowering of CO2 reduced the AMOC and decreased the depth of the AMOC. Liu et al. (2005) and Zhu et al. (2015) also reported similar results. Thus, it is widely accepted that the reduction of CO2 weakens the

AMOC. Two processes were shown to play a role. The first process was associated with an increase in sea ice formation over the Southern Ocean (e.g. Shin et al. 2003). This strengthened the formation of AABW through an increase in brine rejection. As a result, more saline and cold water occupied the bottom sea and caused a weakening and shallowing of the AMOC (e.g. Jansen 2016, Jansen and Nadeau 2016, Sun et al. 2016). The second process was related to the sea ice expansion in the North Atlantic. The expansion of sea ice insulated the heat exchange between the atmosphere and ocean (e.g. Oka et al. 2012, Zhu et al. 2015). As a result, the surface water could not become dense enough to sink, thus the AMOC weakened.

Studies with AOGCMs showed that the expansion of the glacial ice sheets exerted a large impact on the AMOC. For example, Kim (2004) showed that the glacial ice sheets reduced the AMOC. On the other hand, Eisenman et al. (2009), Smith and Gregory (2012), Brady et al. (2013), Zhang et al. (2014), Gong et al. (2015), Klockman et al. (2016) and Brown and Galbraith (2016) showed that the expansion of the glacial ice sheets intensified the AMOC. Therefore, the impact of glacial ice sheets on the AMOC depended on models.

Furthermore, Kawamura et al. (2017) and Abe-Ouchi et al. (personal communication) recently showed that the effect of glacial ice sheets on the AMOC depended on the background climate (CO2 level). They showed that the glacial ice sheet (both the LGM and mid glacial ice sheets) intensified the AMOC under warm climate (high CO2 level), while the expansion of the glacial ice sheet weakened the AMOC under cold LGM condition (low CO2 level). As a result of the reduction of the AMOC due to glacial ice sheets, they obtained a weak and shallow AMOC in their LGM simulations, which is consistent with proxies (e.g. Bohm et al. 2015). Therefore, a better understanding of the impact of glacial ice sheets on the AMOC is essential for interpreting the glacial climate as well as for interpreting the model-data inconsistency in the LGM AMOC.

In order to interpret the reason why the impact of glacial ice sheets depended on models and background climate (CO2 level), it is essential to understand the mechanism by which glacial ice sheets modified the AMOC. Previous studies suggested that the glacial ice sheets intensified the AMOC through affecting surface cooling (Smith and Gregory 2012), precipitation (Eisenman et al. 2009), and surface winds (Zhang et al. 2014, Gong et al. 2015). For example, Smith and Gregory (2012) suggested that stronger surface cooling over the northern North Atlantic by the glacial ice sheets enhanced the AMOC through increasing the atmosphere-ocean heat exchange. On the other hand, Eisenman et al. (2009) suggested that the decrease in precipitation over the northern North Atlantic due to glacial ice sheets intensified the AMOC through increasing the surface salinity over the deepwater formation region. Furthermore, Zhang et al. (2014) suggested that the strengthening of the surface wind and meridional shift of the westerly increased the surface salinity and decreased the sea ice over the deepwater formation region and intensified the AMOC. However, due to the complicated nature arising from the coupled process in the AOGCM, the impact of each processes on the AMOC remained elusive. As a result, it still remained unclear that why some model showed a weakening of the AMOC due to glacial ice sheets.

Regarding the background climate dependence of glacial ice sheet effect on the AMOC, Abe-Ouchi et al. (personal communication) suggested an important role of sea ice over the northern North Atlantic. Since modifications in the sea ice drastically affect the atmosphere-ocean heat exchange (e.g. Gildor and Tziperman 2003, Li et al. 2005, 2010) and thus trigger a large change in the AMOC (Oka et al. 2012), they suggested that difference in the sea ice under different CO2 levels might cause changes in the effect of the glacial ice sheet on the AMOC. However, the mechanism by which sea ice affected the effect of glacial ice sheets on the AMOC remained elusive.

Impact of glacial boundary conditions on the DO cycles

Recently, several studies started to explore how changes in glacial boundary conditions (e.g. CO2 level and glacial ice sheets) modified the stability of the climate and AMOC (DO cycles). Using the ice core data from Antarctic, especially that of Dome Fuji, Kawamura et al. (2017) first explored the relation between the likelihood of occurrence of DO cycles and the

Antarctic temperature over the last 720,000 years. By comparing the duration of an individual DO cycle (sum of interstadials and stadials) and the Antarctic temperature reconstruction, they showed that the DO cycles was most frequent during mid-glacial period, when the climate was significantly colder than interglacial, but not as full-glacial period. In the interglacial and full glacial periods, the recurrence time of DO cycles become very long and the climate was more stable. Buizert and Schmittner (2015) further analyzed the relationship between the duration of interstadials and stadials with the Antarctic temperature reconstruction from the Antarctic ice cores and Greenland ice cores over the last glacial period. They showed that the duration of the interstadials was highly correlated with the Antarctic temperature; the duration of interstadials decreased as the Antarctic temperature decreased. This can be shown in Fig. 1.3. This figure compares typical DO cycles of three periods; early glacial, mid-glacial and full glacial (see Fig.1.1 for the definition). As shown in Buizert and Schmittner (2015), it is clear that the duration of the interstadial decreases as the climate cools (from Fig. 1.3c to 1.3.a, although it is not shown explicitly, the Antarctic temperature decreases from Fig. 1.3c to 1.3.a). On the other hand, with respect to stadials, their duration is shortest during mid-glacial (Buizert and Schmittner 2015). Thus, the DO cycles were most frequent during mid-glacial because of a decrease in the duration of the interstadials during the glacial period and very short stadials.

To explore the reason why DO cycles occurred most frequently during mid-glacial period, Kawamura et al. (2017) conducted several numerical experiments with an AOGCM. First they conducted three experiments; Preindustrial, mid-glacial and full glacial (Fig. 1.1). In these experiments, they obtained a vigorous AMOC in Preindustrial and mid-glacial and obtained a weak AMOC in full glacial (LGM), which was consistent with modern observations (Talley et al. 2003) and paleo reconstructions (e.g. McManus et al. 2004, Bohm et al. 2015). Then, they assessed the sensitivity of each climate to a perturbation. For this purpose, they conducted freshwater hosing experiments under these three climates and explored in which climate does the AMOC weaken the most. As a result, they found that the mid-glacial period was most sensitive to the freshwater hosing. In the Preindustrial climate experiment, the same amount of freshwater did not trigger a drastic weakening of the AMOC as in mid-glacial. In the full glacial experiment, the AMOC was already in a weak mode before the hosing, which is consistent with reconstructions (e.g. McManus et al. 2004). Thus the freshwater input could not cause a drastic weakening of the AMOC. From this, they first showed that the increase in the sensitivity of the AMOC to a perturbation during mid-glacial is the key to have a frequent DO cycles. To assess the role of glacial ice sheets and CO2 in causing sensitive climate during mid-glacial period, they conducted additional sensitivity experiment, which they applied either the lowering of CO2 or glacial ice sheets. They found that the lowering of CO2 was important for the sensitivity. From these, they showed that the global cooling induced by the lowering of CO2 was important for the frequent DO cycles during mid-glacial through increasing the sensitivity of the AMOC to perturbations (e.g. freshwater hosing).

In order to understand the reason why DO cycles occurred frequently during mid-glacial, it is also important to understand the reason why the duration of stadial shortened during mid-glacial. According to the ice core data (e.g. Buizert and Schmittner 2015), there is no clear relationship between the duration of stadials and Antarctic temperature or CO2 (e.g. shortest stadials during mid-glacial, when the CO2 is relatively lower than early glacial). This suggested that another climate forcing play a role in controlling the duration of stadials. Based on previous studies, which suggested that the glacial ice sheets have a large impact on the AMOC (e.g. Kim et al. 2004, Zhang et al. 2014), changes in the ice sheet may play a role. However, there is no study that explored the impact of changes in the glacial ice sheet on the duration of stadials.

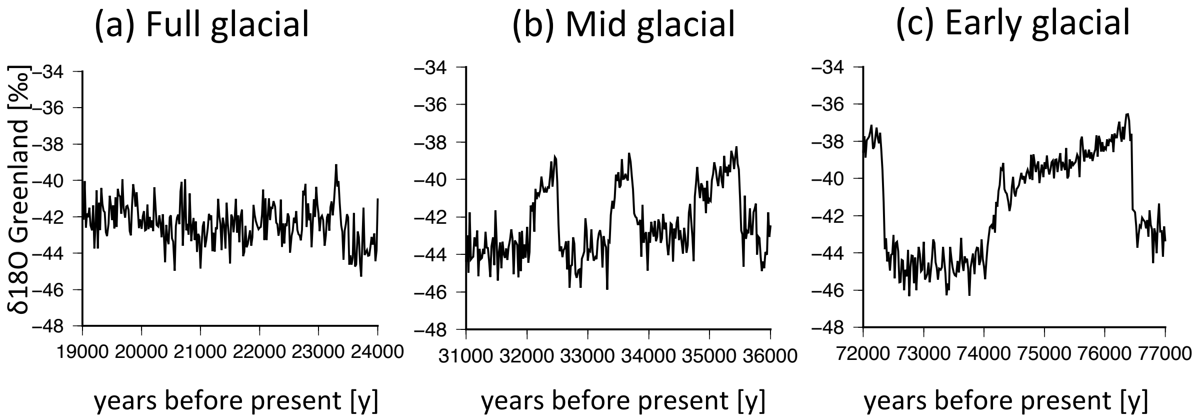

Fig. 1.3 Characteristics of typical DO cycles during (a) full glacial, (b) mid glacial and (c) early glacial. Data of Rasumussen et al. (2012) is used.

Summary

It has been shown that changes in the AMOC played an important role in the glacial climate, for example, the DO cycles (e.g. Kawamura et al. 2017) and changes in the temperature field during the LGM (Ganopolski et al. 1998). Previous studies have reported that the changes in CO2 and glacial ice sheets exerted a large impact on the AMOC and climate. However, while the effect of lowering of CO2 on the AMOC and climate was qualitatively consistent among models, the impact of glacial ice sheets on the AMOC and climate was not straightforward; the effect depended on background climate (e.g. CO2 level) and models qualitatively. The reason for this remained unclear. In addition, recent ice core and modeling studies suggested that the expansion of glacial ice sheets might played a role in modifying the duration of stadials over the last glacial period. This is important to understand why DO cycles occurred frequently during mid-glacial. Thus, a better understanding of the impact of glacial ice sheets on the AMOC and climate is essential for the understanding of glacial climate.

The reason why the impact of glacial ice sheets on the AMOC remained unclear is associated to the coupling process in the AOGCMs. In the AOGCMs, when I modify the ice sheet configurations, changes in winds, surface cooling and atmospheric freshwater flux (E-P-Runoff) are all passed to the oceanic component, hence the individual role on the AMOC becomes unclear. In addition, changes in the oceanic condition can further affect the winds, surface cooling and precipitation. This can modify the ocean circulation again. Thus, because of the strong feedbacks in the AOGCM arising from the coupling process, it is difficult to track the relation of cause and results.

There are two ways to overcome this complexity of AOGCMs and interpret the cause-results relation. The first method is decoupled experiments (e.g. Oka et al. 2012). In this method, the atmospheric model and oceanic model are simulated separately. Thus, we can apply changes in atmospheric condition as a boundary condition and assess the impact of individual process on the AMOC. Another advantage of this method is that, it does not require a lot of computational resources, thus several sensitivity experiments can be conducted. However, this method also has a disadvantage. The decoupling induces unrealistic atmosphere and ocean heat exchange, which greatly affects the stability of the AMOC (e.g. Schmittner et al. 2002). Therefore, it is difficult to assess the effect of changes in heat exchange by glacial ice sheets on the AMOC (cooling effect) with decouple experiments. To overcome this disadvantage, I conduct partial decouple experiments with an AOGCM as a second method. In this method, the atmospheric forcing that drives the oceanic component is switched one by one to a different forcing (e.g. Shcmittner et al. 2002, Krebs and Timmerman 2007). The method itself is similar to the decoupled experiments, though, unlike the decoupled experiments, the partial decoupled experiments are conducted with an AOGCM. Thus, the exchange of heat between the atmosphere and ocean is calculated realistically. By comparing the results, with and without the replacement of atmospheric forcing, I can evaluate the impact of atmospheric forcing (e.g. cooling effect) on the AMOC. However, since I use the AOGCM, it requires a lot of computational resources. Thus I cannot conduct a lot of long-duration experiments with this method.

Through these two methods, I can interpret the result of the complex AOGCMs. To my knowledge, no studies have conducted partial decouple experiments with AOGCMs in paleoclimate AOGCM studies at least for the glacial climate.

Objective of this study

The objective of this study is to explore the impact of glacial ice sheets on the AMOC and climate to improve our understandings of glacial climate and the DO cycles. By means of decoupled and partial decouple experiments, I clarify unclear points in previous studies as follows.

- How do glacial ice sheets affect the AMOC? What is the impact of changes in winds, surface cooling and atmospheric freshwater flux (E-P-Runoff) on the AMOC?
- Why does the impact of glacial ice sheets depend on background climate (CO2 level)?

- Do glacial ice sheets modify the duration of stadials? If so, do they play a role in reducing the duration of stadials during mid-glacial period?

In Chapter 2, I evaluate the impact of changes in the surface winds induced by the glacial ice sheets on the AMOC under LGM conditions. I further assess the important regions of the wind changes on the AMOC and the relative importance of the North American ice sheet and Eurasian ice sheets on the AMOC. For this purpose, I conduct decoupled experiments with an AGCM and an OGCM, decoupled from an AOGCM.

In Chapter 3, I investigate the impact of glacial ice sheets on the duration of stadials during mid-glacial. For this purpose, freshwater hosing experiments are conducted with an AOGCM under several ice sheet configurations. In addition, in order to explore the mechanism by which the glacial ice sheets modify the duration of stadials, partial decouple experiments are conducted. From these experiments, I clarify the impact of changes in surface winds, surface cooling and atmospheric freshwater flux on the AMOC and the recovery time of the AMOC.

In Chapter 4, I investigate the reason why the impact of glacial ice sheets on the AMOC depends on the background climate (CO2 level). Particular emphasis is placed on the role of sea ice, which was suggested by Abe-Ouchi et al. (personal communication). For this purpose, I analyze the data of Kawamura et al. (2017) and Abe-Ouchi et al. (personal communication) and further perform numerical experiments with the AGCM. Partial decouple experiments are also conducted with MIROC4m AOGCM to assess the impact of changes in the winds due to sea ice expansion on the AMOC.

The general conclusion of this study is presented in Chapter 5.

Note that the content of Chapter 2 has recently been accepted to *Climate Dynamics* (23, June, 2017, Sherriff-Tadano et al. 2017).

Chapter 2

Influence of glacial ice sheets on the Atlantic meridional overturning circulation through surface wind change

Abstract

Coupled modeling studies have recently shown that the existence of the glacial ice sheets intensifies the Atlantic meridional overturning circulation (AMOC). Consequently, most models show a strong AMOC in their simulations of the Last Glacial Maximum (LGM), which is however, inconsistent with reconstructions that indicate both a weaker and stronger AMOC during the LGM. Therefore, a detailed investigation of the mechanism behind this intensification of the AMOC is important for a better understanding of the glacial climate and the LGM AMOC. Here, various numerical simulations are conducted to focus on the effect of wind changes due to glacial ice sheets on the AMOC and the crucial region where the wind modifies the AMOC. First, from atmospheric general circulation model experiments, the effect of glacial ice sheets on the surface wind is evaluated. Second, from ocean general circulation model experiments, the influence of the wind stress change on the AMOC is evaluated by applying wind stress anomalies regionally or at different magnitudes as a boundary condition. These experiments demonstrate that the North American ice sheets intensify the AMOC through an increase in the wind stress at the North Atlantic mid-latitudes. This intensification of the AMOC is caused by the increased oceanic horizontal and vertical transports of salt, while the change in sea ice transport has an opposite, though minor, effect. Experiments also show that the Eurasian ice sheet intensifies the AMOC by directly affecting the deep-water formation in the Norwegian Sea, but the effect is relatively small compared to the North American ice sheet.

2.1 Introduction

Ice sheets play a crucial role in the climate system. The existence of the ice sheets not only affects the global sea level, but also the surface air temperature, atmospheric circulation and ocean circulation. Over the Quaternary, the configuration of the ice sheets has changed dramatically in response to insolation changes and associated nonlinear feedbacks (e.g. Lisiecki and Raymo 2005, Abe-Ouchi et al. 2013). During Quaternary glacial times, the North American (Laurentide/Cordilleran, e.g. Stokes et al. 2012) and Eurasian ice sheets grew individually across the continents (hereafter referred to as glacial ice sheets). The glacial ice sheets exerted a substantial impact on the atmospheric circulation and glacial climate (e.g. Manabe and Broccoli 1985, Broccoli and Manabe 1987, Chiang et al. 2003, Chiang and Bitz 2005, Abe-Ouchi et al. 2007, Yanase and Abe-Ouchi 2010, Singarayer and Valdes 2010, Chavaillaz et al. 2012). Especially in the North Atlantic, the Icelandic Low and Azores High intensified and the wind fields at high latitudes were also modified due to the topography of the North American ice sheet as well as the albedo and cooling over the ice sheet (Cook and Held 1988, Ringler and Cook 1997, 1999, Pausata et al. 2011, Liakka et al. 2011, Liakka 2012).

Recently, in the context of the Paleoclimate Modeling Intercomparison Project (PMIP, Braconnot et al. 2007, 2012), simulations of the Last Glacial Maximum (LGM) have been conducted with atmosphere-ocean coupled general circulation models (AOGCMs). These simulations made it possible to explore the impact of the glacial boundary conditions on the Atlantic Meridional Overturning Circulation (AMOC). While reconstructions suggested both a weaker and stronger AMOC during the LGM (McManus et al. 2004; Lynch-Stieglitz et al. 2007, Gerardi et al. 2009, Lippold et al. 2012, Böhm et al. 2015), most of the AOGCMs (except CCSM3 and HadCM3 in PMIP2) in the PMIP showed a strong AMOC in their LGM simulations (Weber et al. 2007, Otto-Blisner et al. 2007, Muglia and Schmittner 2015). Moreover, recent sensitivity studies using AOGCMs showed that the existence of the glacial ice sheets intensified the AMOC (Eisenman et al. 2009; Smith and

Gregory 2012, Brady et al. 2013, Zhang et al. 2014, Zhu et al. 2014, Gong et al. 2015, Brown and Galbraith 2016 and Kawamura et al. 2017). Therefore, in order to improve our understanding of the glacial climate and the LGM AMOC, it is essential to understand how glacial ice sheets affected the AMOC.

To investigate why the AMOC during the LGM was stronger than the Preindustrial as simulated in the Model for Interdisciplinary Research on Climate 4m (MIROC4m; Hasumi and Emori 2004) AOGCM in the PMIP2 (Weber et al. 2007), Oka et al. (2012, hereafter OHA12) conducted sensitivity experiments with an ocean general circulation model (OGCM) decoupled from MIROC4m. The decoupling of the OGCM from the AOGCM helped identify the important processes controlling the AMOC, which was more complex in the AOGCMs due to the coupling process. By applying the surface condition computed in the MIROC4m Preindustrial and LGM simulations as a boundary condition to the OGCM, they found that the LGM wind stress was crucial for the intensification of the AMOC. They suggested that the changes in salt transport were important, which affected the sea surface density and deep-water formation (e.g. Oka et al. 2001, Born et al. 2010). While the study of OHA12 was based on the glacial wind simulated by MIROC4m, the role of the glacial wind was further confirmed by Muglia and Schmittner (2015), who also showed that the glacial wind over the Northern Hemisphere (Atlantic) intensified the AMOC in all of the PMIP3 and PMIP2 models.

Changes in the salt transport could be induced by wind-driven ocean processes, wind-driven sea ice processes or both. For example, wind anomalies at low, mid and high latitudes could modify the gyre circulation as well as Ekman upwelling, downwelling and convective mixing. This has a substantial impact on the horizontal and vertical salt transport, which affects the AMOC through modifying the stratification of the ocean (Oka et al. 2001, Timmerman and Goosse 2004, Montoya and Levermann 2008, Saenko 2009a, b, Montoya et al. 2011, Zhong et al. 2011). In addition, wind anomalies at mid-high latitudes could modify the wind-driven sea ice transport (Born et al. 2010, Zhong et al. 2011, Zhu et al. 2014, Zhang et al. 2014). This modifies the sea ice coverage and regions of sea ice melt, which affects the AMOC through modifying the atmosphere-ocean heat exchange and surface salinity.

To summarize, previous studies have shown that the glacial wind over the North Atlantic intensified the AMOC in the PMIP LGM simulations. The glacial ice sheets were considered to play an important role and changes in the salt transport were shown as key to the enhancement of the AMOC. However, several points remained elusive in these studies. First, since both OHA12 and Muglia and Schmittner (2015) used only the Preindustrial and LGM wind and did not distinguish between the impact of glacial ice sheets and that of other factors (CO_2, sea surface conditions), the impact of glacial ice sheet wind on the AMOC was somewhat vague. In addition, the role of individual ice sheets (the North American and Eurasian ice sheets) remained unclear. This is very important, especially when we want to apply our knowledge to other glacial periods as the North American and Eurasian ice sheets did not wax and wane in concert during the glacial periods (Abe-Ouchi et al. 2013, Hughes et al. 2016). Second, it still remained unclear as to which region of the wind field in the North Atlantic was important for the intensification of the AMOC as the wind field was largely modified over the entire North Atlantic. This is also related to the uncertainty in the important mechanism, as changes in the wind in the North Atlantic could modify the salt transport and hence the AMOC by affecting the wind-driven oceanic circulation or by affecting the wind-driven sea ice transport.

In this chapter, I first clarify the important factors for the intensification of the AMOC through modification of the surface wind (e.g. the North American ice sheet, the Eurasian ice sheet or other effects). Then I investigate in which regions the wind anomalies are important, and the mechanism by which they modify the AMOC. For this purpose, I conduct a series of sensitivity experiments using an atmospheric general circulation model (AGCM) and an OGCM decoupled from an AOGCM. In this method, the atmosphere and the ocean do not interact with one another so that I can apply surface conditions as boundary conditions to the OGCM, which enables us to clearly evaluate the effect of the wind change due to the glacial ice sheets on the ocean circulation and on the AMOC. Moreover, by taking advantage of the decoupled nature, I can conduct several sensitivity experiments and decompose the glacial ice sheet effect into the North American effect and the Eurasian effect. In addition, I can investigate the regions in which the wind change is crucial and the mechanism by which it influences the AMOC. In this way I can systematically investigate the wind change effect due to the glacial ice sheets on the AMOC.

This chapter is organized as follows. In section 2.2, I describe the model and the experimental setup. In section 2.3, I investigate the effect of glacial ice sheets on the atmospheric circulation by running AGCM experiments. In section 2.4, I investigate the effect of the glacial ice sheets on the AMOC through surface wind change and the regions of crucial importance. In section 2.5, I discuss our results and lastly, I present our conclusions in section 2.6.

2.2 Methodology

2.2.1 Model and Experimental setup

I use an atmospheric component and an oceanic component of the coupled model MIROC4m, which participated in the PMIP2 (Braconnot et al. 2007). The performance of MIROC4m has been assessed extensively for both modern climate (e.g. Nozawa et al. 2005, O'ishi and Abe-Ouchi 2009, Ohgaito and Abe-Ouchi 2009, Yanase and Abe-Ouchi 2010, Watanabe et al. 2010) and paleoclimate, the latter of which is compared to other models and reconstructions in the context of PMIP (e.g. Kageyama et al. 2006, Masson-Delmotte et al. 2006, Yanase and Abe-Ouchi 2007, Murakami et al. 2009, Rojas et al. 2009, Otto-Bliesner et al. 2009, Lü et al. 2010). In particular, it is known that MIROC4m simulates a stronger AMOC in its LGM simulation compared to the Preindustrial climate simulation (e.g. Weber et al. 2007, Otto-Bliesner et al. 2007; note that the model integration has been extended to more than 6000 years after these studies, though the result does not change). Even though I use a model that simulates a stronger AMOC in the LGM, our main goal is to understand how glacial ice sheets affect the AMOC by modifying the surface wind field and thus results of MIROC4m are utilized for this purpose.

The AGCM that I use is identical to the uncoupled atmospheric component of MIROC4m (Numaguchi et al. 1997). It is coupled to a land surface model and has a horizontal resolution of T42, which corresponds to an approximate grid resolution of 2.8˚, and 20 layers in the vertical. Unlike the coupled model, the sea surface temperature and sea ice are fixed as boundary conditions. The OGCM used in this study is also identical to the uncoupled oceanic component of MIROC4m (Hasumi 2006). It is coupled to a sea ice model, and has a resolution of ~1.4˚ in longitude and 0.56˚~1.4˚ in latitude (latitudinal resolution is finer near the equator) and 43 unevenly spaced layers in the vertical. Unlike the coupled model, atmospheric conditions are fixed as boundary conditions (further information can be found below). See Hasumi and Emori (2004) and Chan et al. (2011) for further information on the AGCM and the OGCM of MIROC4m. The AGCM reproduces the surface westerlies over the North Atlantic reasonably well, while it underestimates the southwest-northeast orientation and the strength (Fig. 2.1). The OGCM reproduces the Preindustrial AMOC and the North Atlantic Deep Water (NADW) formation in the Greenland-Iceland-Norwegian Sea reasonably well (Oka et al. 2008, OHA12).

To run the AGCM, several boundary conditions are required: ice sheet configuration, greenhouse gases, orbital parameters, sea surface temperature and sea ice. I apply either Preindustrial or LGM boundary conditions, the latter of which follows the protocol of the LGM simulation in PMIP2 (Braconnot et al. 2007). This protocol specifies glacial ice sheets (ICE-5G, Peltier 2004, Fig. 2.2), greenhouse gas concentration (CO_2 = 185 ppm, CH_4 = 350 ppb and NO_2 = 200 ppb; Fluckiger et al. 1999; Dallenbach et al. 2000; Monnin et al. 2001) and orbital parameters. Sea surface temperature and sea ice are taken from the climatology of the MIROC4m Preindustrial or LGM simulations. See Table 2.1 for a detailed design of the experiments. I use the ICE-5G LGM ice sheets in order to be consistent with OHA12. However, some studies have show that differences between the ice sheet reconstructions of the PMIP3 and ICE-5G have a significant impact on the climate over the North Atlantic (e.g. Abe-Ouchi et al. 2015, Fig. 1.2b). The impact of differences in the ice sheet reconstructions on the surface wind will be presented in the discussion. Integration time is 32 years and I use the last 23 years for analysis and for OGCM boundary conditions.

Three types of surface fluxes, heat, momentum and freshwater (E-P-Runoff), are required to run the OGCM. The heat flux is diagnosed from the surface temperature, humidity, short wave radiation and long wave radiation, all of which are taken from the coupled model Preindustrial and LGM simulations as in OHA12. The freshwater flux (E-P-Runoff) is calculated using the evaporation, precipitation and runoff and these are taken directly from the coupled model. Note that the freshwater flux (E-P-Runoff) used here does not include the freshwater flux form the sea ice. This is calculated explicitly in the OGCM. For the momentum flux, surface wind stresses from AGCM experiments are applied. The model is integrated for more than 1000 years from modern climatology of the observed temperature and salinity (Steele et al. 2001). I analyze the last 100 years after verifying that the strength of the AMOC (Fig. 2.3) and deep ocean temperature (below 2000m) has reached a quasi-equilibrium state (not shown, the decreasing trend of the deep ocean temperature is less than 0.005˚C in the last 200 years for all the experiments). In order to focus on the impact of the wind forcing on the ocean circulation, I fix the land-sea mask to modern in all the OGCM experiments following OHA12 (although changes in bathymetry have a large impact on the AMOC (Hu et al. 2015), its evaluation is not our main purpose). Note that the importance of the wind forcing does not depend on the land-sea mask since OHA12 and Muglia and Schmittner (2015) showed that changes between the LGM and Preindustrial wind field enhance the AMOC under modern and glacial land-sea masks, respectively. However, since I use the modern land-sea mask for OGCM experiments, this will cause a mismatch between the land-sea masks of the AGCM and OGCM (e.g. in the AGCM, the Barents Sea is treated as ice sheet while it is treated as ocean in the OGCM). Thus, when I directly apply wind stress anomalies computed in the AGCM (e.g. differences between LGM and Preindustrial simulations) to the OGCM, unrealistic wind anomalies will appear in these regions. To avoid this unrealistic wind forcing, I do not apply wind stress anomalies over the regions where there is a mismatch of land-sea mask between AGCM and OGCM. However, note that the main results do not depend on whether I apply the unrealistic wind or not. In fact, when I conduct an additional experiment to test whether this unrealistic wind forcing itself has an impact on the AMOC, results show that the impact is very small. This supports our claim that the results presented in this study are not affected by the unrealistic wind forcing.

2.2.2 Design of experiments

Using the AGCM, I first conduct three experiments; A-PI, A-NOICE and A-FULLICE. A-PI is a Preindustrial climate simulation; in A-NOICE I lower the greenhouse gas concentrations and impose LGM orbital forcing. Importantly, I do not change either the surface albedo or topography; and in A-FULLICE (LGM simulation), I further impose LGM surface albedo and topography (Table 2.1). From these experiments, I define the glacial ice sheet effect as the difference between A-FULLICE and A-NOICE. The residual effect (changes in sea ice, sea surface temperature, greenhouse gases, and insolation) is defined as the difference between A-NOICE and A-PI. As a next step, I decompose the effect of the glacial ice sheets into albedo, topography, North American and Eurasian ice sheets by conducting three sensitivity experiments (A-FLATICE, A-NAFULL, A-EUFULL in Table 2.1). In A-FLATICE, I impose the albedo of the glacial ice sheet, but do not change the topography. In A-NAFULL, I impose the albedo and topography of the ice sheet over North America. In A-EUFULL, I impose the albedo and topography of the ice sheet over Europe and Asia. Using these experiments, I define the albedo effect as the difference between A-FLATICE and A-NOICE, the topography effect as the difference between A-FULLICE and A-FLATICE, the North American ice sheet effect as the difference between A-NAFULL and A-NOICE and the

Eurasian ice sheet effect as the difference between A-EUFULL and A-NOICE (Table 2.1).

For the OGCM, I first conduct two experiments (O-PI and O-LGM); O-PI is a Preindustrial climate simulation, in which I apply the wind field from A-PI and the thermal and freshwater (E-P-runoff) conditions from the Preindustrial AOGCM simulation; O-LGM is a LGM simulation in which I apply the wind field from A-FULLICE and the thermal and freshwater (E-P-runoff) conditions from the LGM AOGCM simulation (Table 2.2). As a next step, in order to explore the effect of glacial ice sheet wind on the glacial AMOC, I apply the wind stresses from the AGCM sensitivity experiments to the OGCM under glacial thermal condition. In O-HEAT I apply the wind stress from A-PI, in O-NOICE I apply wind stress from A-NOICE, in O-FULLICE I apply wind stress from A-FULLICE, in O-NAFULL I apply wind stress from A-NAFULL and in O-EUFULL I apply wind stress from A-EUFULL. From these experiments, I define the glacial ice sheet wind effect as the difference between O-FULLICE and O-NOICE, the residual wind effect as the difference between O-NOICE and O-HEAT, the North American ice sheet wind effect as the difference between O-NAFULL and O-NOICE and the Eurasian ice sheet wind effect as the difference between O-EUFULL and O-NOICE (Table 2.2). In these experiments, I apply Preindustrial freshwater fluxes (E-P-Runoff) in order to carry out this investigation in a simple manner. The fact that the changes in the freshwater flux (E-P-Runoff) has only a small impact on the AMOC (OHA12 and subsection 2.2.3) supports the rationale for investigating the impact of the glacial ice sheet wind on the AMOC under Preindustrial freshwater flux (E-P-Runoff).

To clarify the important regions where the surface wind anomaly forced by the glacial ice sheet (difference of A-FULLICE and A-NOICE) intensifies the glacial AMOC in the North Atlantic, which was unclear in Muglia and Schmittner (2015), I subdivide the North Atlantic into three latitudinal bands, 0˚~30˚N, 30˚N~65˚N and 65˚N~90˚N, and apply the wind stress anomaly to each of these regions separately in the OGCM. I apply the wind anomaly over 0˚~30˚N in O-LOW, I apply the wind anomaly over 30˚N~65˚N in O-MIDDLE and I apply the wind anomaly over 65˚N~90˚N in O-HIGH. I have set the boundary at 65˚N because it is the lowest latitude where sea ice exists throughout the year in O-NOICE.

Table 2.1 (a) Experiments conducted with the AGCM. For LGM greenhouse gases (GHG) I set CO_2 levels to 185 ppm, CH_4 to 350 ppb and NO_2 to 200 ppb (Fluckiger et al. 1999; Dallenbach et al. 2000; Monnin et al. 2001). For the glacial ice sheet, I used the LGM configuration of ICE-5G (Peltier 2004). The sea surface temperature (SST) and sea ice distribution required to run the AGCM is taken from the climatology of the MIROC4m LGM experiment. For North America (NA) and Eurasia (EU), I only apply the North American and Eurasian glacial ice sheets, respectively. (b) Relationship between the effect of the individual boundary condition and sensitivity experiments. I define the albedo effect as A-FLATICE – A-NOICE, the topography effect as A-FULLICE – A-FLATICE, the North American ice sheet effect as A-NAFULL – A-NOICE and the Eurasian ice sheet effect as A-EUFULL – A-NOICE. Many previous studies decompose the albedo and topography effect in the same manner (e.g.: Abe-Ouchi et al. 2007, Kageyama and Valdes 2000, Ogura and Abe-Ouchi 2001).

(a)

Exp name	SST and Sea ice	Ice sheet albedo	Ice sheet topography	GHG and insolation
A-PI	MIROC4m Preindustrial	Preindustrial	Preindustrial	MIROC4m Preindustrial
A-NOICE	MIROC4m LGM	Preindustrial	Preindustrial	PMIP2 LGM
A-FULLICE	MIROC4m LGM	ICE5G LGM	ICE5G LGM	PMIP2 LGM
A-FLATICE	MIROC4m LGM	ICE5G LGM	Preindustrial	PMIP2 LGM
A-NAFULL	MIROC4m LGM	Laurentide only	Laurentide only	PMIP2 LGM
A-EUFULL	MIROC4m LGM	Eurasian only	Eurasian only	PMIP2 LGM

(b)

Boundary condition	Difference between two experiments
Glacial ice sheets	A-FULLICE minus A-NOICE
Topography	A-FULLICE minus A-FLATICE
Albedo	A-FLATICE minus A-NOICE
North American	A-NAFULL minus A-NOICE
Eurasian	A-EUFULL minus A-NOICE
Residual effect	A-NOICE minus A-PI

Table 2.2 (a) Experiments conducted with the OGCM. Wind stress from the AGCM is used. For the thermal and freshwater (E-P-Runoff) conditions, I use either the glacial condition from the MIROC LGM simulation or the Preindustrial condition from the MIROC Preindustrial simulation. Value inside brackets for O-PI refers to the strength of the modern AMOC from Talley et al. (2003). (b) Relationship between the boundary condition effect and sensitivity experiments.

(a)

Exp name	wind stress	thermal condition	freshwater flux	AMOC max
O-PI	A-PI	MIROC4m Preindustrial	MIROC4m Preindustrial	16.3 [18±5]
O-HEAT	A-PI	MIROC4m LGM	MIROC4m Preindustrial	4.7
O-NOICE	A-NOICE	MIROC4m LGM	MIROC4m Preindustrial	8.4
O-NAFULL	A-NAFULL	MIROC4m LGM	MIROC4m Preindustrial	22.9
O-EUFULL	A-EUFULL	MIROC4m LGM	MIROC4m Preindustrial	21.5
O-FULLICE	A-FULLICE	MIROC4m LGM	MIROC4m Preindustrial	25.4
O-LGM	A-FULLICE	MIROC4m LGM	MIROC4m LGM	26.2

(b)

Boundary condition	Difference between two experiments
Glacial ice sheets	O-FULLICE minus O-NOICE
North American ice sheet	O-NAFULL minus O-NOICE
Eurasian ice sheet	O-EUFULL minus O-NOICE
Residual effect	O-NOICE minus O-HEAT

Table 2.3 The maximum value of the AMOC obtained from additional OGCM sensitivity experiments, in which I applied changes to the wind field regionally, is shown. Regional changes in the wind stress between A-FULLICE and A-NOICE are superimposed to A-NOICE wind field and then applied to the OGCM. Therefore these results should be compared with O-NOICE (8.4 Sv).

Exp name	Region where A-FULLICE wind is applied	AMOC max (Sv)
O-HIGH	(0°-360°, 65°N-90°N)	17.3
O-MIDDLE	(80°W-0°, 30°N-65°N)	22.5
O-LOW	(80°W-0°, 0°N-30°N)	8.6

2.2.3 Reproducibility of the results of AOGCM and OHA12 by decoupled simulations

Here, I assess whether our decoupled simulations can reproduce the results of MIROC4m and OHA12, because the present study is based on these two sets of results. Figure 2.1 shows the changes between the LGM and Preindustrial surface wind stresses over the North Atlantic for both the AOGCM and the AGCM. The AGCM reproduces the changes in the AOGCM surface wind field reasonably well, while it overestimates the strengthening of the westerly wind stress at North Atlantic mid-latitudes.

Figure 2.4 compares the AMOC and the NADW formation regions of the AOGCM and our decoupled OGCM. Decoupled OGCM simulations (O-PI and O-LGM) reproduce both the changes in the strength and the spatial structure of the AOGCM AMOC reasonably well in that the AMOC is stronger and deeper in the LGM simulation compared to the Preindustrial simulation. The AMOC in O-PI is slightly weaker than that in the AOGCM simulations whereas the AMOC in O-LGM is slightly stronger than that in the coupled experiment (Fig. 2.4). Note that in O-LGM, the meridional streamfunction around 60˚N is strong compared to that in the AOGCM LGM simulation. This is possibly related to the slight difference in the region of NADW formation, which lies slightly farther north in O-LGM compared to that of the AOGCM LGM simulation (Fig. 2.4d, h). However, since O-LGM at least reproduces the maximum strength of the AMOC and also the deepening of the NADW cell compared to O-PI, OGCM simulations successfully reproduce the overall structure of the AMOC in AOGCM.

The wind from the AGCM also reproduces the results of OHA12 in that the Preindustrial wind (A-PI) produces a weak AMOC (O-HEAT) and the LGM wind (A-FULLICE) produces a strong AMOC (O-FULLICE) under glacial thermal and Preindustrial freshwater (E-P-Runoff) conditions (Fig. 2.3 and Table 2.2). Note that OHA12 used the winds from MIROC4m. They suggested that the weakening of the AMOC in O-HEAT is caused by the southward expansion of sea ice in the North Atlantic due to the cooling, which insulates heat exchange between the atmosphere and ocean and weakens the deep-water formation. The same results are obtained under glacial thermal and freshwater conditions (not shown). This is also consistent with OHA12, who showed that changes in the freshwater flux (E-P-Runoff) between the Preindustrial and glacial climates have a relatively small impact on the strength of the AMOC compared to changes in the thermal and momentum conditions.

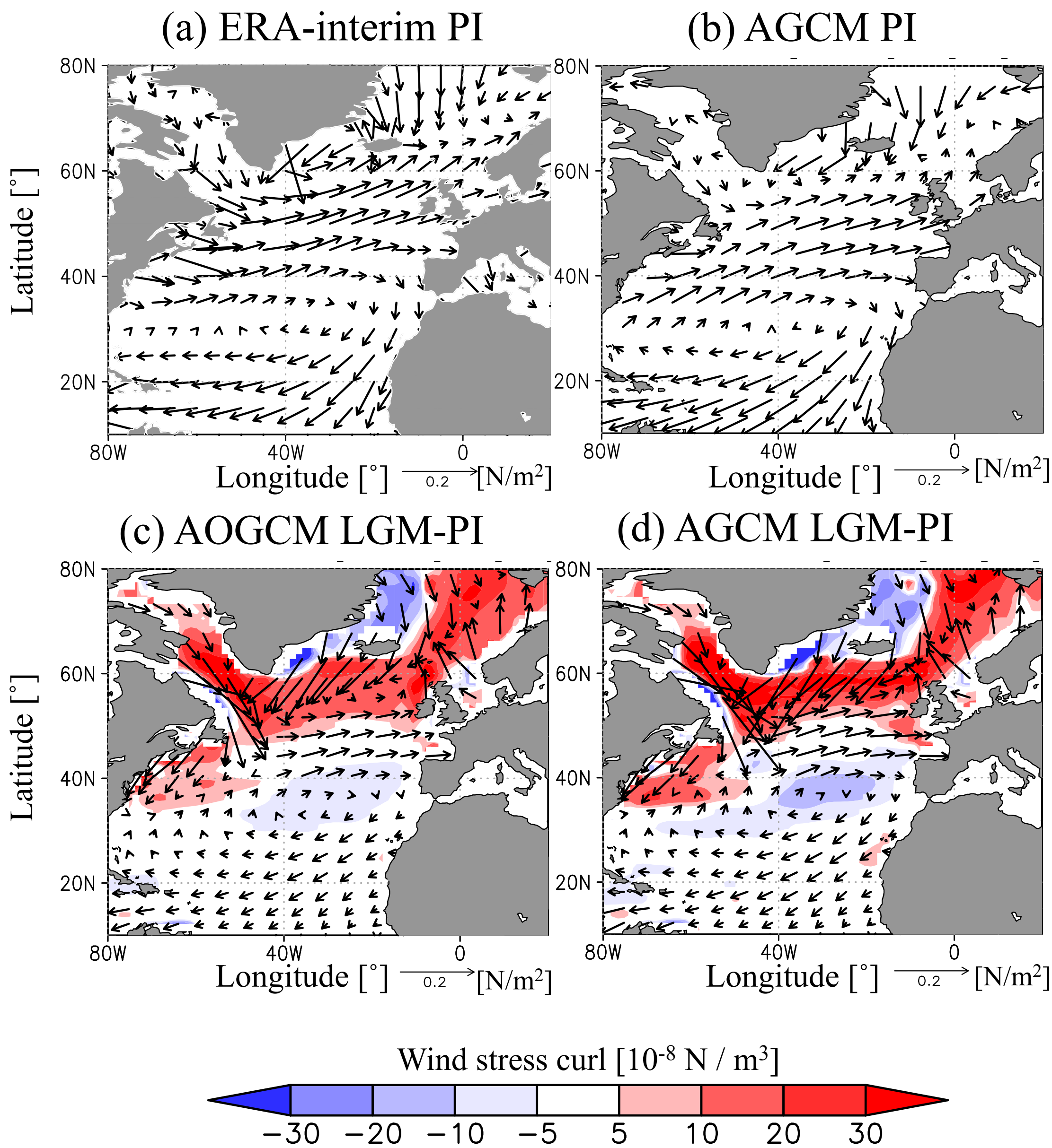

Fig. 2.1 Validation of surface wind stress from AGCM. Top figures show annual climatology of surface wind stress (N/m^2; averaged over 1981-2010) from ERA-interim (Dee et al. 2011; a) and the AGCM Preindustrial climate simulation (A-PI, b). In the bottom figures (c, d) changes in wind stress (arrow, N/m^2) and wind stress curl (color, 10^{-8} N/m^3) in the North Atlantic between LGM and Preindustrial in the AOGCM and AGCM are shown.

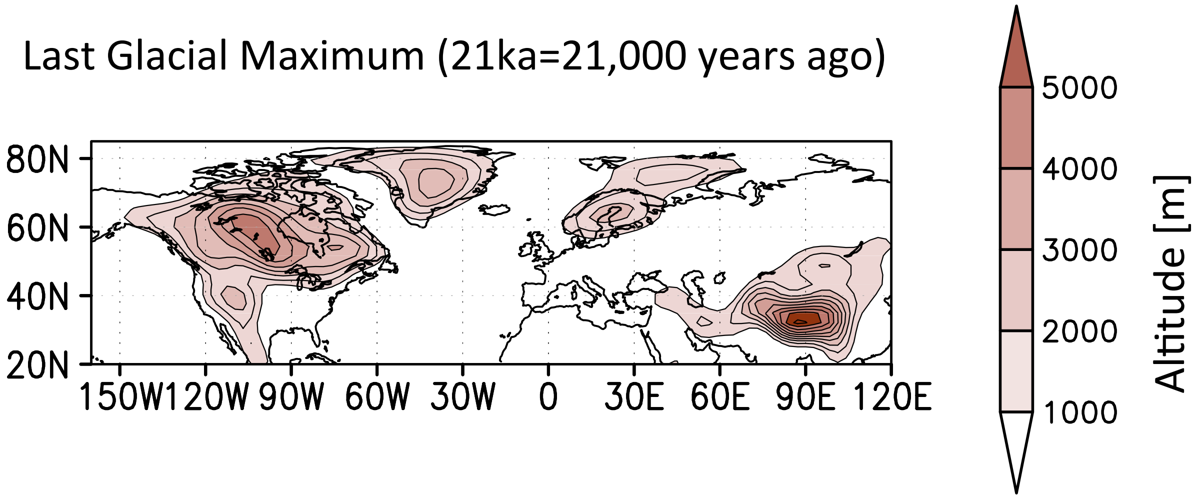

Fig. 2.2 Northern Hemisphere topography of Last Glacial Maximum used in the simulations (21ka, Peltier 2004).

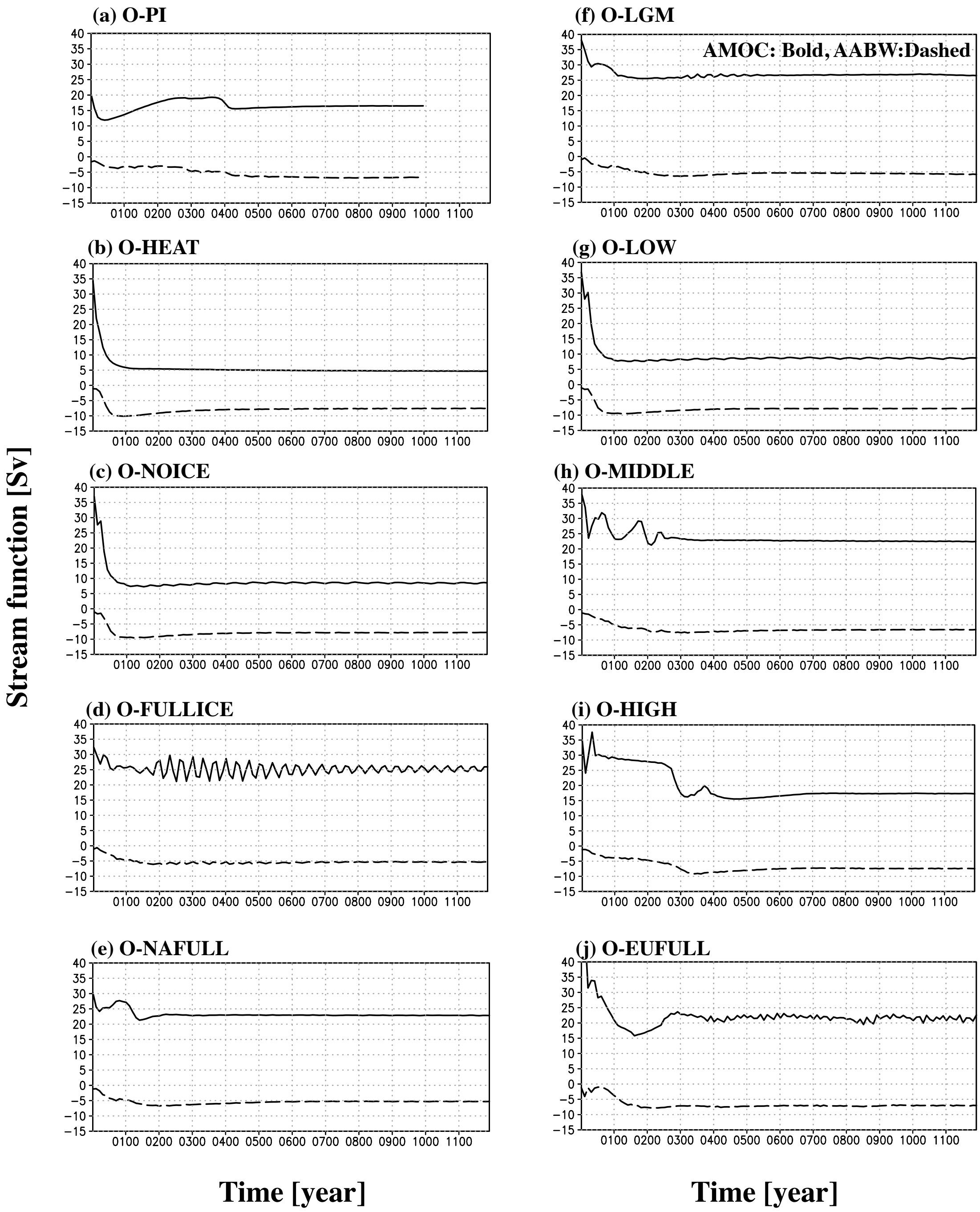

Fig. 2.3 Evolution of the strength of the AMOC (bold) and AABW (dashed) in uncoupled OGCM experiments. (a) O-PI, (b) O-HEAT, (c) O-NOICE, (d) O-FULLICE, (e) O-NAFULL, (f) O-LGM, (g) O-LOW, (h) O-MIDDLE, (i) O-HIGH and (j) O-EUFULL.

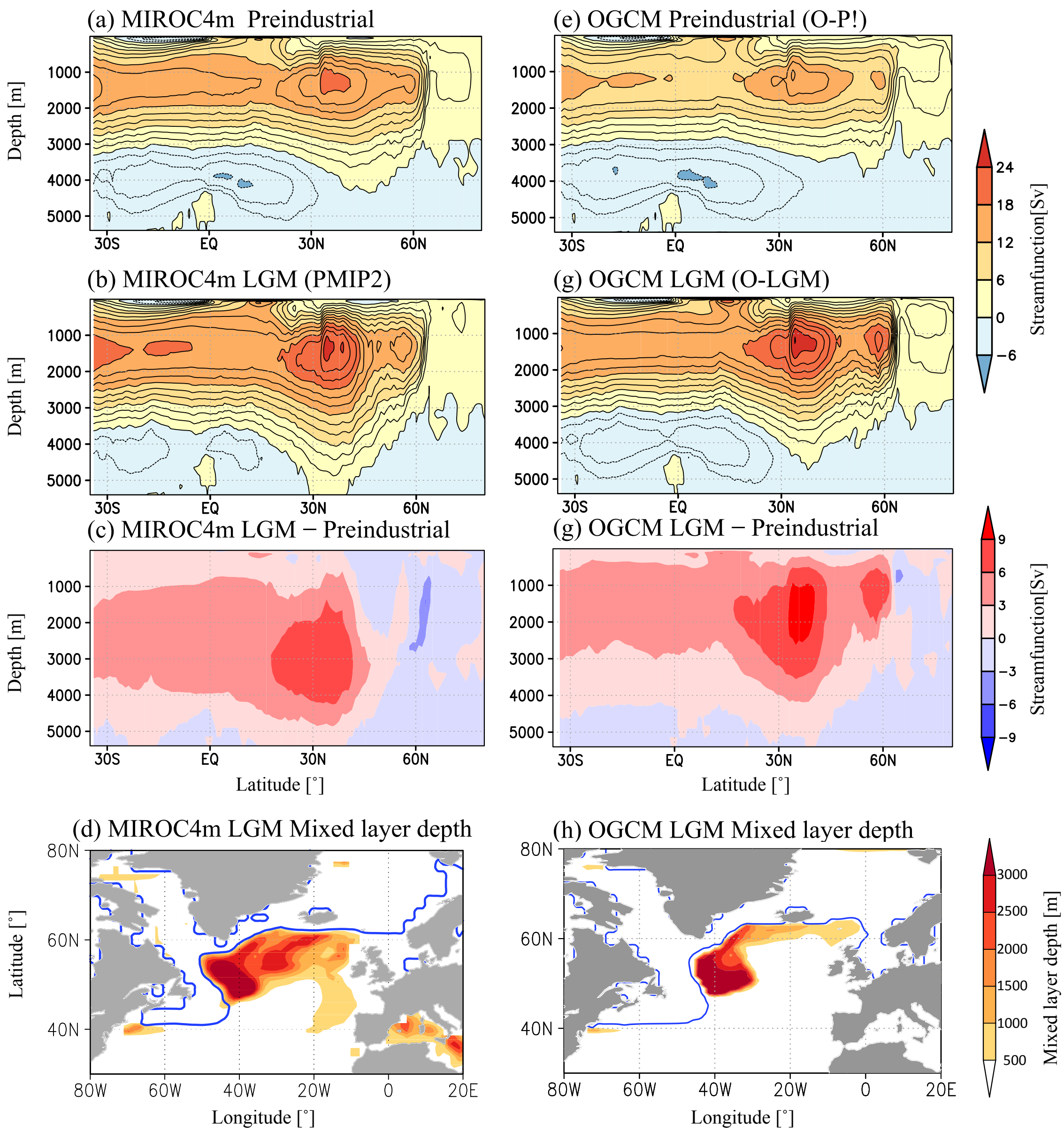

Fig. 2.4 Comparison of the results from MIROC4m (left) and OGCM (right). In the OGCM simulations, heat and freshwater fluxes are calculated using the variables taken from the corresponding MIROC4m simulations, while the wind stress is taken from corresponding AGCM simulations. The meridional streamfunction in the North Atlantic from the Preindustrial simulation (a, e), the LGM simulation (b, f) and the difference between the LGM and Preindustrial simulations (c, g) are shown. In (d, h) the February mixed layer depth from the LGM simulations are shown, when the deep-water formation becomes vigorous.

2.3 Results from the AGCM

In this section, I first decompose the difference between the LGM and Preindustrial winds computed in MIROC4m into glacial ice sheet effect and residual effects. I then investigate the general features of the atmospheric circulation change induced by the glacial ice sheets and further decompose the glacial ice sheet effect.

Figure 2.5 shows the annual mean surface wind field from the AGCM experiments A-PI, A-NOICE and A-FULLICE. By examining the anomaly field, it is obvious that the difference between A-FULLICE and A-PI is dominated by the glacial ice sheet effect, which is deduced from the difference between A-FULLICE and A-NOICE. Changes in sea ice, sea surface temperature, greenhouse gases and insolation (residual effect) have some impact on the surface wind, while the impact is relatively small compared to those induced by the glacial ice sheets. However I do not explore them further since they have only a relatively small impact on the AMOC (section 2.4.1).

Figure 2.6 shows the general structure of the change in the atmospheric circulation due to the glacial ice sheets. The December-January-February (DJF) average is shown because the change in the wind field reaches its maximum and deep-water formation intensifies before reaching a maximum at the end of this period. Glacial ice sheets induce a cyclonic circulation anomaly downstream of the North American ice sheet and an anti-cyclonic circulation anomaly over the North American ice sheet throughout the troposphere (e.g. Pausata et al. 2011); these anomalies shift westward with height (Löfverström et al. 2015). This is mainly induced by the topography of the North American ice sheet (Fig. 2.7, e.g. Cook and Held 1988, see also Hoskins and Karoly 1981, Held 1983, Trenberth and Chen 1988, Cook and Held 1992, Ringler and Cook 1997, Liakka et al. 2011 for the atmospheric response to a large orographic forcing). However, other factors, such as surface cooling over the ice sheet, also play an important role (Cook and Held 1988, Ringler and Cook 1999, Liakka 2012, Löfverström et al. 2014). Associated with the stationary wave field, the westerlies also intensify throughout the troposphere in the North Atlantic (Li and Battisti 2008, Merz et al. 2015).

The influence of the Eurasian ice sheet is small compared to that of the North American ice sheet, although it has a substantial effect over the Norwegian Sea, where a strong southwesterly wind anomaly is induced. This is associated with a high-pressure anomaly over the Eurasian ice sheet. Comparison of A-EUFULL and A-FLATICE shows that this high-pressure anomaly is induced by the topography of the Eurasian ice sheet. However, unlike the North American ice sheet, the anomaly is confined near the surface, which is perhaps associated with the small size of the ice sheet. This suggests the importance of analyzing both the changes in the wind field at the surface and in the free troposphere, since it is difficult to estimate changes in the former by only looking at the latter (Figure 2.7).

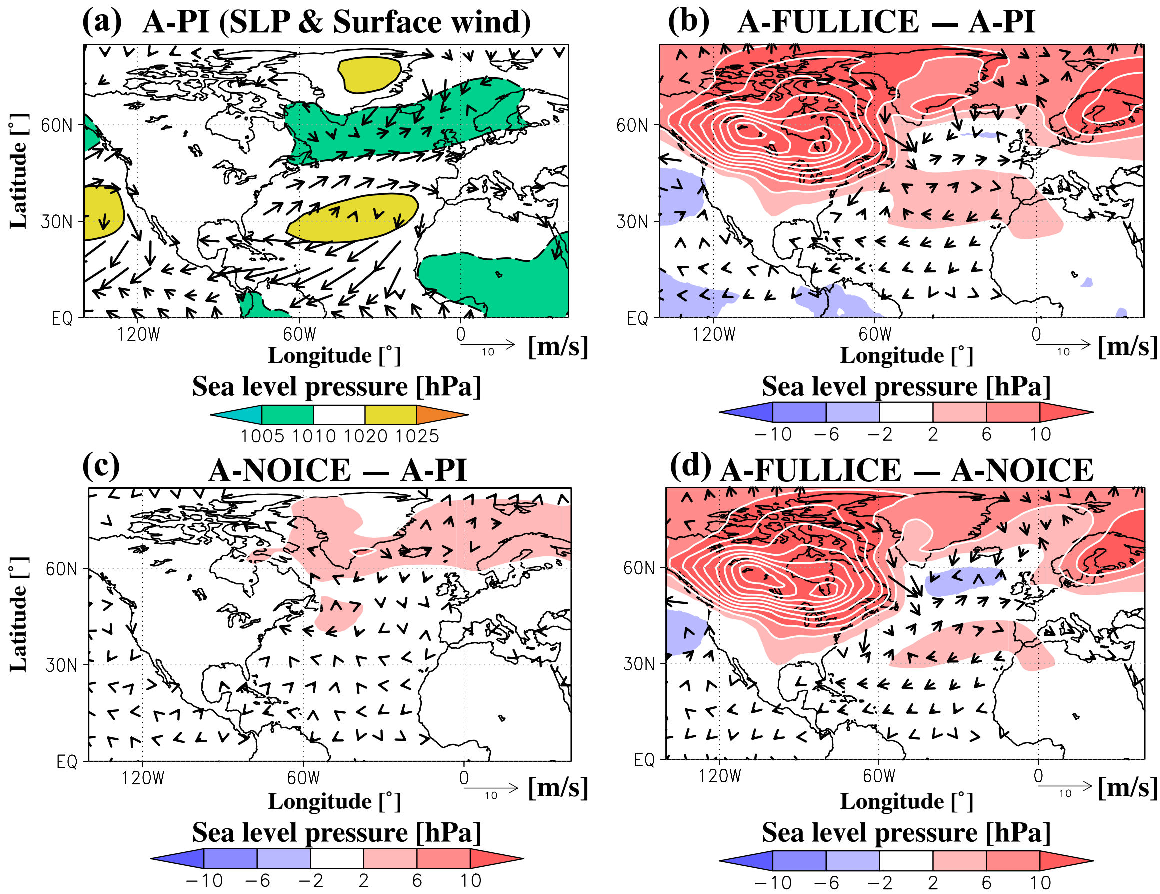

Fig. 2.5 Surface wind (arrow, m/s) and sea level pressure (SLP, color, hPa) from AGCM experiments. The annual average climatology of (a) A-PI and the differences between A-FULLICE and A-PI (b), A-NOICE and A-PI (c, residual effect) and A-FULLICE and A-NOICE (d, ice sheet effect) are shown.

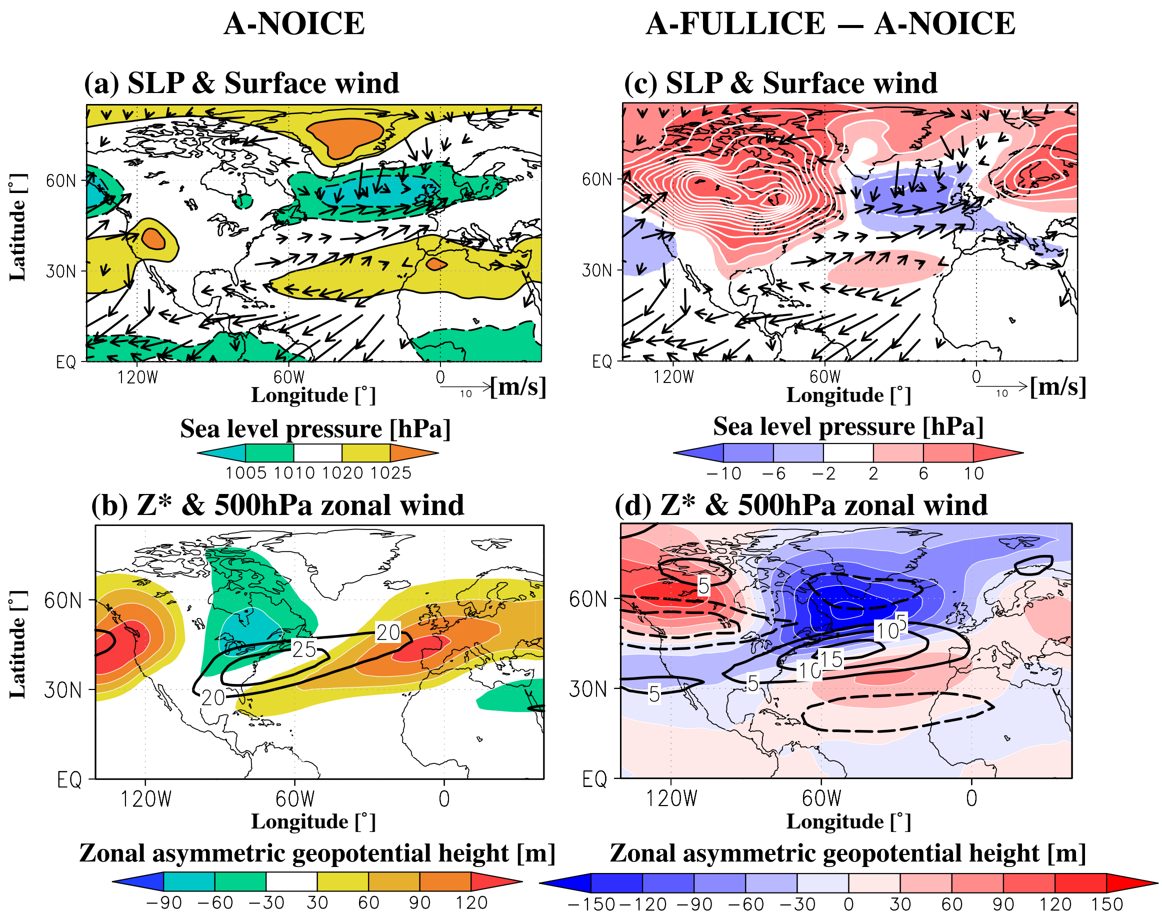

Fig. 2.6 General structure of glacial ice sheet induced circulation change throughout the troposphere. DJF average is shown. Climatology from A-NOICE is shown on the left and glacial ice sheet effect is shown on the right. Top figures: 10 m height wind (arrow, m/s) and sea level pressure (SLP: contour and color, hPa). Bottom figures: 500 hPa eddy geopotential height (color: anomaly from the zonal mean, m) and 500 hPa zonal wind (contour, m/s). For the SLP anomaly field, regions where the anomaly is outside the range -2 hPa to 2 hPa are shaded. Contour interval is 4 hPa, starting from -2 hPa and 2 hPa.

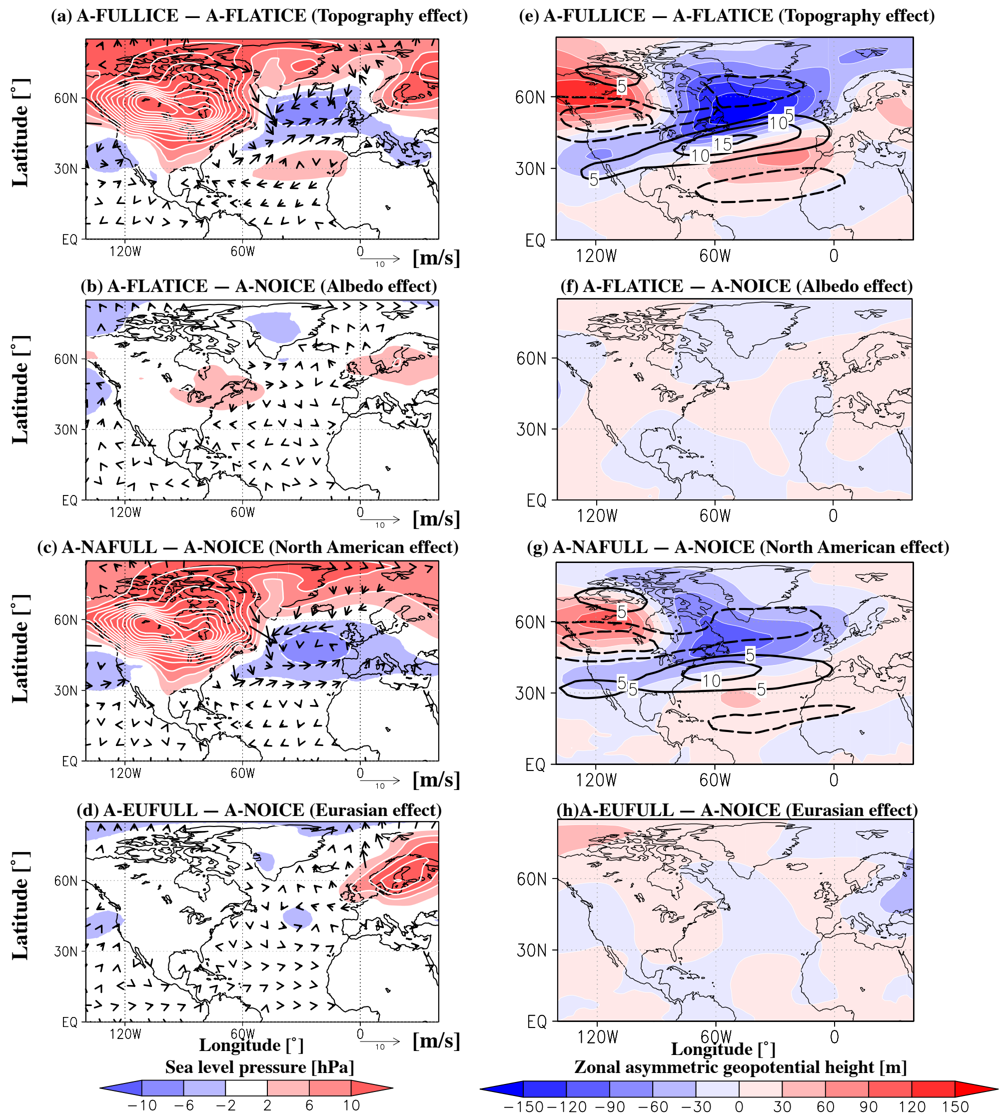

Fig. 2.7 Decomposition of the glacial ice sheet effect into topography, albedo, North American and Eurasian effect. DJF average is shown. Left figures: 10 m height wind (arrow, m/s) and sea level pressure (SLP: contour and color, hPa). Right figures: 500 hPa eddy geopotential height (color: anomaly from the zonal mean, m) and 500 hPa zonal wind (contour, m/s). For the SLP anomaly field, regions where the anomaly lies outside the range -2 to 2 hPa are shaded. Contour interval is 4 hPa, starting from -2 hPa and 2 hPa. Note that a different contour interval is used for the albedo and Eurasian effects in the right figures since their effect were especially small.

2.4 Results from the OGCM

In this section, I first identify the important factor for the intensification of the AMOC through modification of the surface wind stress by using the wind field from the AGCM experiments in the previous section (Table 2.2). Then I investigate how the wind change intensifies the AMOC.

2.4.1 Influence on the AMOC

Using the wind stress from the AGCM sensitivity experiments, I run the OGCM under the glacial thermal and Preindustrial freshwater (E-P-Runoff) conditions (O-FULLICE, O-NOICE and O-HEAT; in O-FULLICE, winds from A-FULLICE are applied, in O-NOICE, winds from A-NOICE are applied and in O-HEAT, winds from A-PI are applied, see Table 2.2). As a result, I find that the glacial ice sheet effect, which is deduced from the difference between O-FULLICE and O-NOICE, plays an important role in intensifying the AMOC (an increase of 17.0 Sv compared to O-NOICE). By applying the wind stress from A-NAFULL and A-EUFULL to the OGCM (O-NAFULL and O-EUFULL, respectively), the effect of the wind from individual glacial ice sheets is further investigated. Results reveal that both of the ice sheets play an important role in intensifying the AMOC, while the impact of the North American ice sheet is larger (compared to O-NOICE, there is an increase of 14.5 Sv in O-NAFULL and an increase of 13.1 Sv in O-EUFULL, respectively, see Table 2.2). On the other hand, the modification of the wind from A-PI to A-NOICE has only a relatively small impact on the AMOC (an increase of 3.7 Sv in O-NOICE compared to O-HEAT). These experiments demonstrate that the glacial ice sheets, especially the North American ice sheet, play an important role in modifying the AMOC through surface wind.

In order to assess the importance of the separate localities within the North Atlantic, which was unclear in the previous studies (OHA12, Muglia and Schmittner 2015), I apply regional changes (differences between A-FULLICE and A-NOICE) to the wind field in the low latitudes (O-LOW), mid-latitudes (O-MIDDLE) and high latitudes (O-HIGH) of the North Atlantic. As a result, I find that the AMOC drastically intensifies only when the wind field is modified at mid and high latitudes (compared to O-NOICE, I see an increase of 8.9 Sv in O-HIGH, an increase of 14.1 Sv in O-MIDDLE and an increase of 0.2 Sv in O-LOW, see Table 2.3). From these experiments, I identify the North Atlantic mid-high latitudes as a key region, with the mid-latitudes being of primary importance. Note that I also conduct an experiment, which I apply regional changes to the winds in the Southern Hemisphere, though the impact on the AMOC is small (an increase of 0.4 Sv compared to O-NOICE).

The drastic change in the AMOC is accompanied by a shift in the region of NADW formation (see Fig. 2.8 and Kageyama et al. 2009). In the weak AMOC cases, the production of the NADW is confined to the center of the subpolar gyre (O-NOICE) or even ceases (O-HEAT). On the other hand, in the strong AMOC cases, NADW formation also takes place outside or at the rim of the subpolar gyre (O-FULLICE, O-NAFULL, O-EUFULL, O-HIGH and O-MIDDLE). Note that the AMOC can become strong, even if the NADW formation does not occur north of the Greenland-Iceland-Scotland ridge (Fig. 2.8). This is somewhat different from OHA12, who showed that, under Preindustrial wind conditions, the increase in global cooling triggers a large weakening of the AMOC by preventing NADW formation in the Greenland-Iceland-Norwegian Seas. This suggests that stronger wind over the North Atlantic can generate a vigorous AMOC with NADW formation south of the Greenland-Iceland-Scotland ridge. Note also that, although the depth of the NADW cell generally increases as the AMOC strengthens (Fig. 2.9), the depth can differ substantially with a similar strength (Fig. 2.9f, g, e.g. O-NAFULL and O-EUFULL). This may have some implications on our understanding of past changes in the deep ocean, for example a vigorous, though shallow AMOC (e.g. Lippold et al 2012). However, detailed analyses on these topics are not covered here as I am focussing on how the glacial ice sheets intensify the AMOC. These will be reported elsewhere.

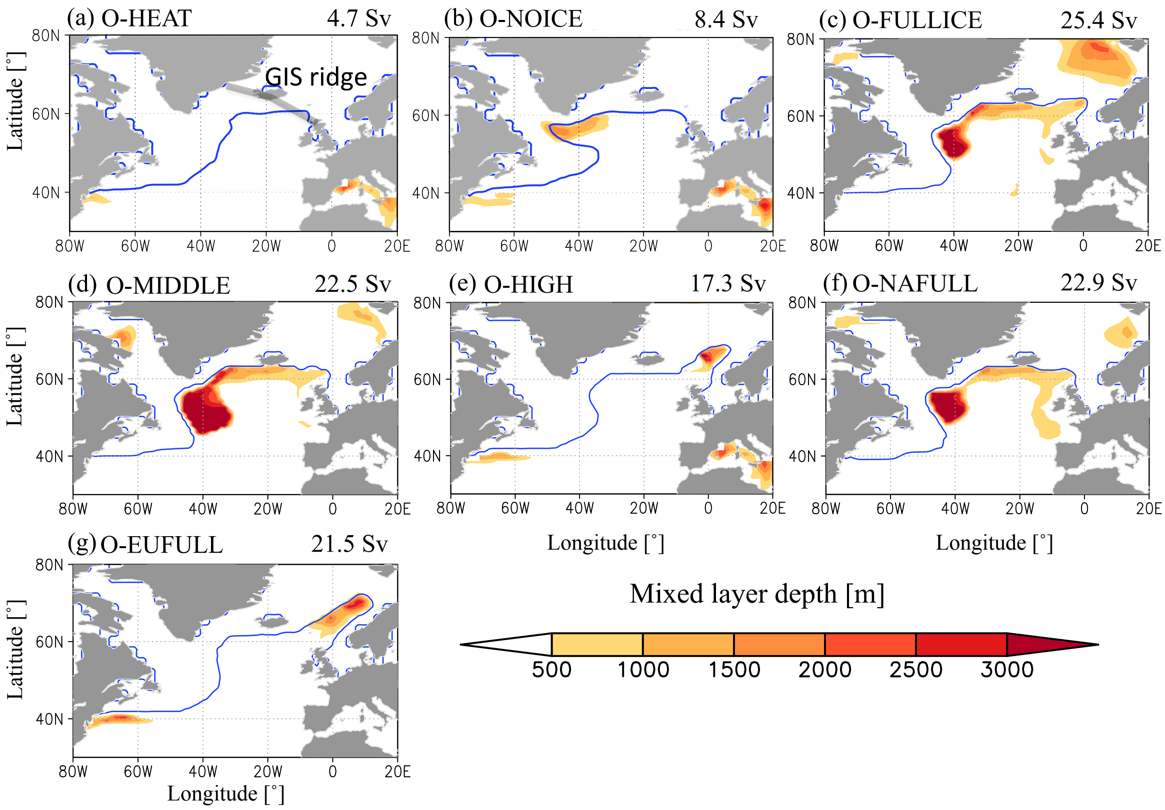

Fig. 2.8 Region of NADW formation (color: February mixed layer depth, m) and February 50% sea ice concentration (bounded by blue line) for (a) O-HEAT, (b) O-NOICE, (c) O-FULLICE, (d) O-MIDDLE, (e) O-HIGH, (f) O-NAFULL and (g) O-EUFULL. Value shown in the upper right corner in each panel indicates the maximum strength of the AMOC. In (a) the location of the Greenland-Iceland-Scotland Ridge is indicated in grey line.

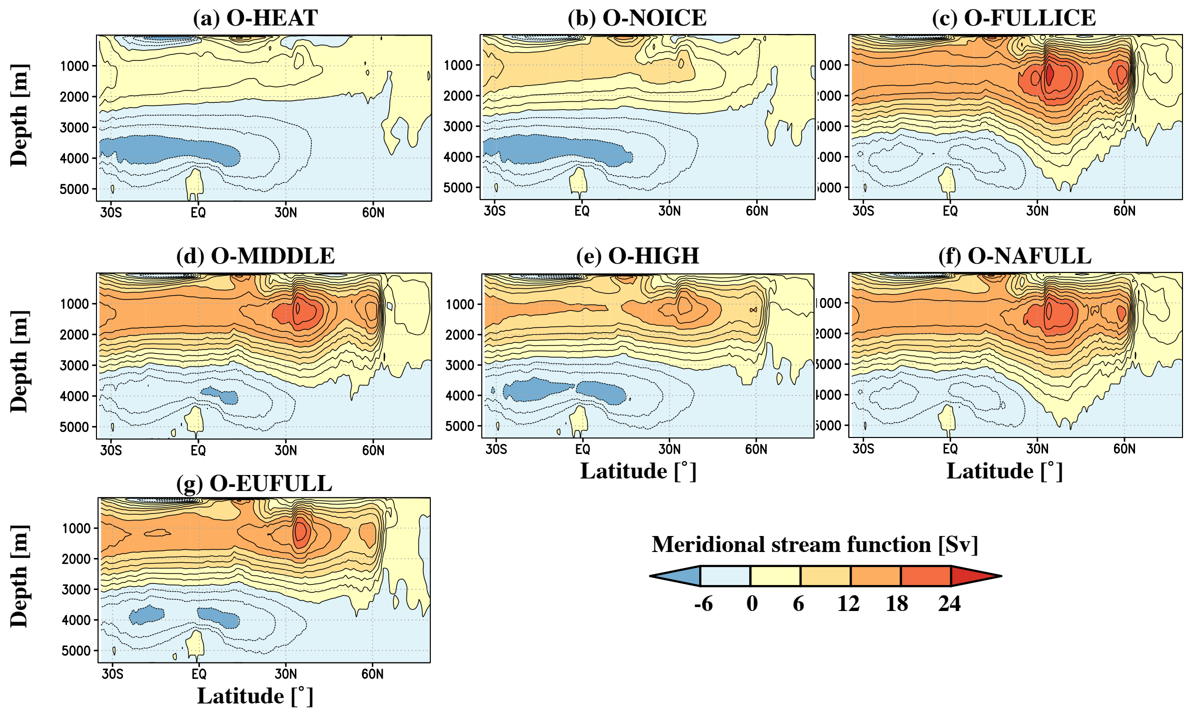

Fig. 2.9 The meridional streamfunction in the Atlantic for (a) O-HEAT, (b) O-NOICE, (c) O-FULLICE, (d) O-MIDDLE, (e) O-HIGH, (f) O-NAFULL and (g) O-EUFULL.

2.4.2 Wind stress change in the North Atlantic and its effect on the AMOC

The mechanism by which the North Atlantic mid-latitude and high latitude winds intensify the AMOC is investigated in this section. For this purpose, the annual mean wind stress and wind stress curl is shown as it is directly related to the wind-driven ocean circulation and sea ice transport in the North Atlantic (Fig. 2.10). In A-NOICE, the wind stress curl is generally positive north of 50˚N and negative south of 50˚N, which is associated with the Icelandic Low and Azores High. These mainly drive the subpolar and subtropical gyres, respectively. From the anomaly field (Fig. 2.10b), it is clear that the glacial ice sheet, especially the topography effect of the North American ice sheet (Fig. 2.7), intensifies the background wind stress curl by strengthening the Icelandic Low. The high-pressure anomaly induced over the North American continent also contributes to the increase in the positive wind stress curl by enhancing the northwesterly wind near the Labrador Sea. Quantitatively speaking, the zonal mean positive wind stress curl over the Atlantic basin doubles in magnitude north of 45˚N (Fig. 2.10c). South of 40˚N, the negative wind stress curl increases by a factor of 1.5 compared to A-NOICE. At high latitudes, the glacial ice sheet enhances the northwesterly wind at Baffin Bay and the northerly wind along the east coast of Greenland to the Irminger Sea, and it induces southwesterly wind in the Norwegian Sea. The enhanced cyclonic circulation in the Norwegian Sea results in an increase in the positive wind stress curl of that region. Overall, the glacial ice sheets intensify A-NOICE wind stress curl over the North Atlantic.

In order to explore the direct impact of the wind forcing on the ocean and its mechanism in detail, I conduct additional experiments with the OGCM. In these experiments, I vary the wind stress linearly from A-NOICE to A-FULLICE globally and calculate the quasi-equilibrium AMOC for each experiment. These sets of experiments are named O-LINEAR. These experiments enable us to clarify the direct impact of wind anomalies on the ocean circulation. This is because the differences in the ocean circulation between

O-FULLICE and O-NOICE are largely affected by the drastic changes in the AMOC and the NADW formation region (e.g. Figs. 12, 13 and 15, see also Oka et al. 2001, Montoya and Levermann 2008, Saenko 2009a, Montoya et al. 2011) as well as the wind anomalies. Thus, it is difficult to extract the direct impact of the wind anomalies on the oceanic circulation only by exploring the difference between O-FULLICE and O-NOICE. The wind stress applied in each experiment is as follows:

$$wind\ stress = (1 - \alpha) \times \text{A\_NOICE} + \alpha \times \text{A\_FULLICE}$$

Here, α denotes a wind stress factor and ranges from 0 to 1. When $\alpha = 1$ ($\alpha = 0$), the wind stress is identical to A-FULLICE (A-NOICE). Therefore, as the factor increases, the wind stress varies from A-NOICE to A-FULLICE. I use the last year of O-NOICE as initial conditions for these sensitivity experiments. The results of O-LINEAR experiments are summarized in Figure 2.11. The AMOC intensifies abruptly at $\alpha = 0.27$, after a gradual increase. This abrupt intensification is associated with the initiation of NADW formation outside and at the rim of the subpolar gyre (not shown). After the transition, the AMOC responds linearly to a gradual increase in the wind forcing. Thus, the initiation of the new NADW formation region amplifies the strengthening of the glacial AMOC due to the glacial ice sheet wind. Therefore, by analyzing the difference in the oceanic properties of O-NOICE ($\alpha = 0$) and $\alpha = 0.25$ experiment (prior to the abrupt intensification in the AMOC), I can investigate the direct impact of changes in the wind on the oceanic circulation and also the cause of the shift of the NADW formation region.

Differences in the zonally averaged potential density, salinity and potential temperature in the Atlantic basin of O-NOICE and $\alpha = 0.25$ experiment (Fig. 2.12) are analyzed because the shift in the region of the NADW formation is related to the changes in the surface water properties of the northern North Atlantic (e.g. Montoya and Levermann 2008). From Fig. 2.12, it is clear that the sea surface density increases at mid-high latitudes and thus the stratification weakens before the large abrupt increase in the AMOC. This density increase is primarily attributed to an increase in surface salinity, while the increase in temperature has only a small impact on the surface density. Therefore, the large abrupt increase in the AMOC can be triggered as a result of the increase in salinity at the surface of the northern North Atlantic. This increase in salinity can be attributed to either wind-driven oceanic circulation (Oka et al. 2001, Timmerman and Goose 2004, Montoya and Levermann 2008, Montoya et al. 2011, Zhang et al. 2014) or/and wind-driven sea ice transport (Born et al. 2010, Zhong et al. 2011, Zhu et al. 2014, Zhang et al. 2014).

The increase in oceanic salt transport appears important for the large increase in surface salinity (Figs. 2.13 and 2.14). Fig. 2.13 shows the surface salinity and barotropic streamfunction, which defines the subpolar and subtropical gyres, and Fig. 2.14 shows vertical-latitudinal cross-section of isopycnal surfaces zonally averaged over the subpolar gyre (45˚W to 25˚W). Fig. 2.14 illustrates the exposure of the subsurface warm, saline water to the sea surface due to Ekman upwelling driven by the positive wind stress curl in this region (e.g. Marshall and Schott 1999, Kuhlbrodt et al. 2007, Montoya et al. 2011). In O-NOICE, saline water is transported horizontally/vertically from the subtropics/subsurface by the wind-driven gyre/Ekman upwelling. This maintains the salinity maximum at high latitudes, as there is no minimum in the freshwater flux in this region (not shown). Note that the subpolar gyre does not extend into the Norwegian Sea because the production of NADW is absent in this region as a result of the sea ice expansion due to surface cooling (OHA12). As the surface wind strengthens due to the glacial ice sheet, both the subpolar and subtropical gyres are enhanced (Fig. 2.13b). This strengthening of the gyres enhances horizontal transport of salt (Oka et al. 2001, Timmerman and Goose 2004, Montoya and Levermann 2008), thus contributing to the increase in surface salinity of the northern North Atlantic. The strengthening of the surface wind also enhances the vertical salt transport. This is demonstrated in Fig. 2.14b - the glacial ice sheet wind enhances the upwelling of the subsurface saline water to the surface, which, in turn, triggers the convective mixing. This further mixes the saline subsurface water with fresher surface water and increases the surface salinity (Zhong et al. 2011). Thus, the strengthening of both horizontal and vertical oceanic salt transport plays an important role.

On the other hand, changes in the sea ice transport appear to play an opposite role. Figure 2.15 shows the sea ice volume transport and sea ice formation rate superimposed on the annual mean 10% sea ice concentration. In O-NOICE, sea ice forms near the continental shelf, is transported mainly by the wind field (Steele et al. 1997, Zhu et al. 2014) and melts near the region of NADW formation and at mid-latitudes (e.g. Bitz et al. 2005). In the experiment of $\alpha = 0.25$ (Fig. 2.15b), the glacial ice sheet wind enhances the sea ice transport into the North Atlantic (Fig. 2.15c), which is induced by intensified northwesterly wind in the Labrador Sea and northerly wind in the Irminger Sea (e.g. Fig. 2.10b). In particular, the zonally averaged southward sea ice volume transport increases by 0.06 $m^2\ s^{-1}$ in the Labrador Sea and 0.014 $m^2\ s^{-1}$ in the Denmark Strait compared to O-NOICE. As a result, the area-averaged sea ice melt (negative sea ice growth rate across 70˚W-5˚W, 40˚N-65˚N) increases from 0.92 cm/day to 1.23 cm/day. This will reduce the surface salinity and weaken the AMOC (Saenko et al. 2002, Born et al. 2010). However, the fact that the glacial ice sheet wind intensifies the AMOC suggests that the oceanic salt transport processes play the more important role in intensifying the AMOC, which dominates the opposite effect of the sea ice process. (Note that the increase in freshwater transport by sea ice will be partly compensated by a decrease in freshwater transport by liquid water (Saenko et al. 2002), which can be expected from a large increase in salinity at the Baffin Bay. However changes in sea ice have larger impact on the deep-water formation as they directly affect the deep-water formation region (Born et al. 2010).)

Changes in the wind over the Greenland-Iceland-Norwegian Sea also modify the AMOC (O-HIGH, O-EUFULL). In this region, positive wind stress curl is enhanced due to the glacial ice sheet topography (Figs. 2.7 and 2.10) and the Eurasian ice sheet also plays an important role (Fig. 2.16). The increase in the positive wind stress curl intensifies both the surface cyclonic ocean circulation and the Ekman upwelling (Fig. 2.17). This enhances horizontal and vertical transport of salt to the surface and thus increases the surface salinity

(Fig. 2.13). In addition, the southwesterly wind stress anomaly exports the sea ice out of the Norwegian Sea and leads to a reduction in sea ice (Fig. 2.15c). Thus, once the wind forcing becomes sufficiently large, deep-water formation can be initiated in the Norwegian Sea owing to higher surface salinity and a reduced amount of sea ice (Fig. 2.8c, e, g). As a result, the AMOC exhibits an abrupt intensification (e.g. Fig. 2.11). Thus, these results show that local changes in surface winds over the Norwegian Sea can trigger drastic changes in the AMOC by directly affecting the deep-water formation in this region.

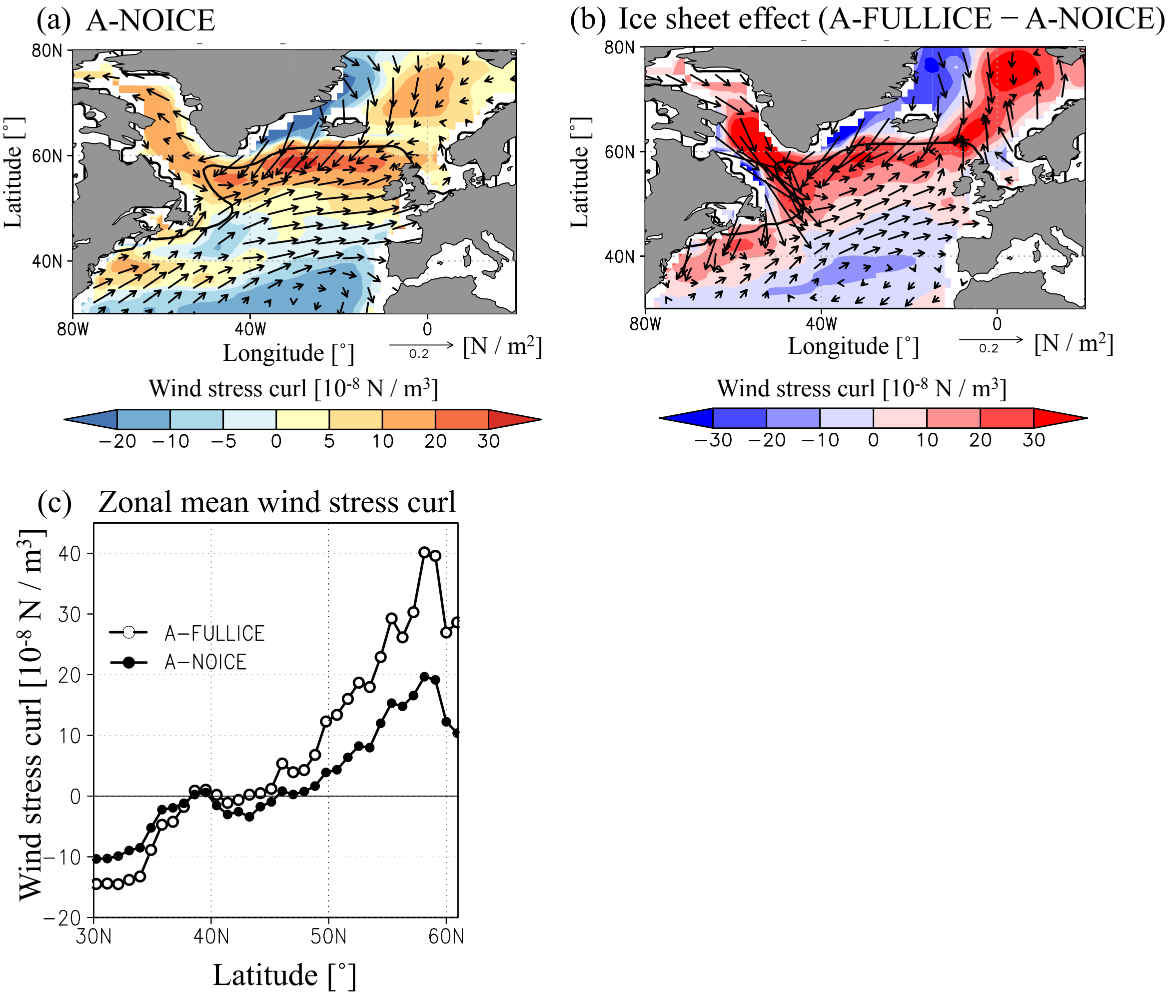

Fig. 2.10 Annual mean surface wind stress (arrow, N/m^2) and the wind stress curl (color, 10^{-8} N / m^3) for (a) A-NOICE and (b) the glacial ice sheet effect (A-FULLICE - A-NOICE). (c) Wind stress curl zonally averaged over the Atlantic basin for A-NOICE (black circle) and A-FULLICE (white circle). In (a) and (b), the black line indicates annual mean 50% sea ice concentration in O-NOICE.

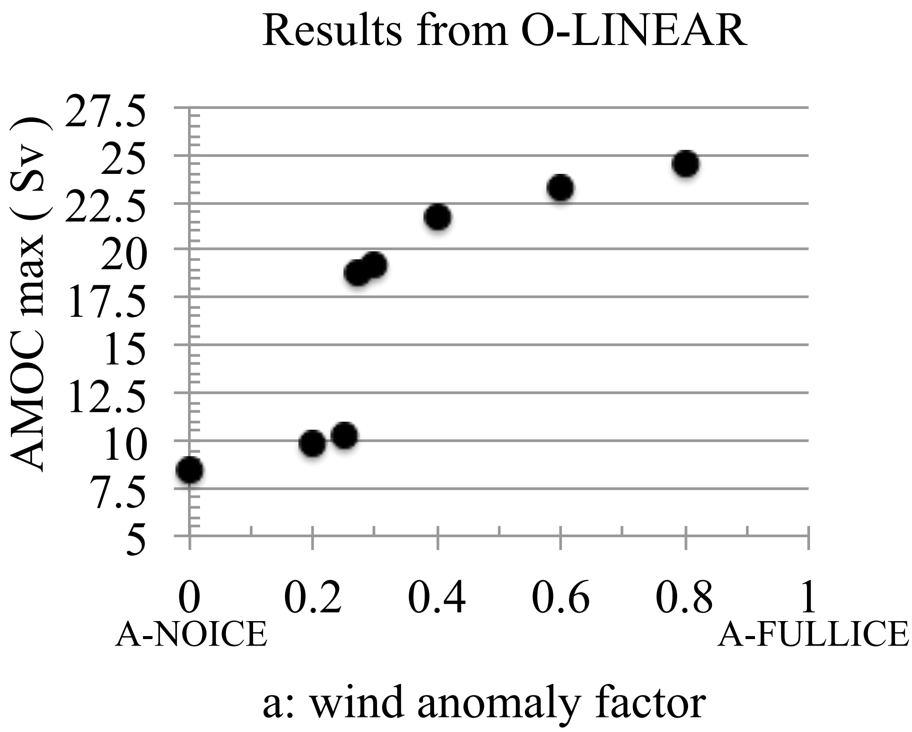

Fig. 2.11 Results from linear sensitivity experiments (O-LINEAR). Scatter plot of AMOC strength and wind stress anomaly factor. Experiments are initialized from O-NOICE. An abrupt increase in the AMOC occurs at a=0.27.

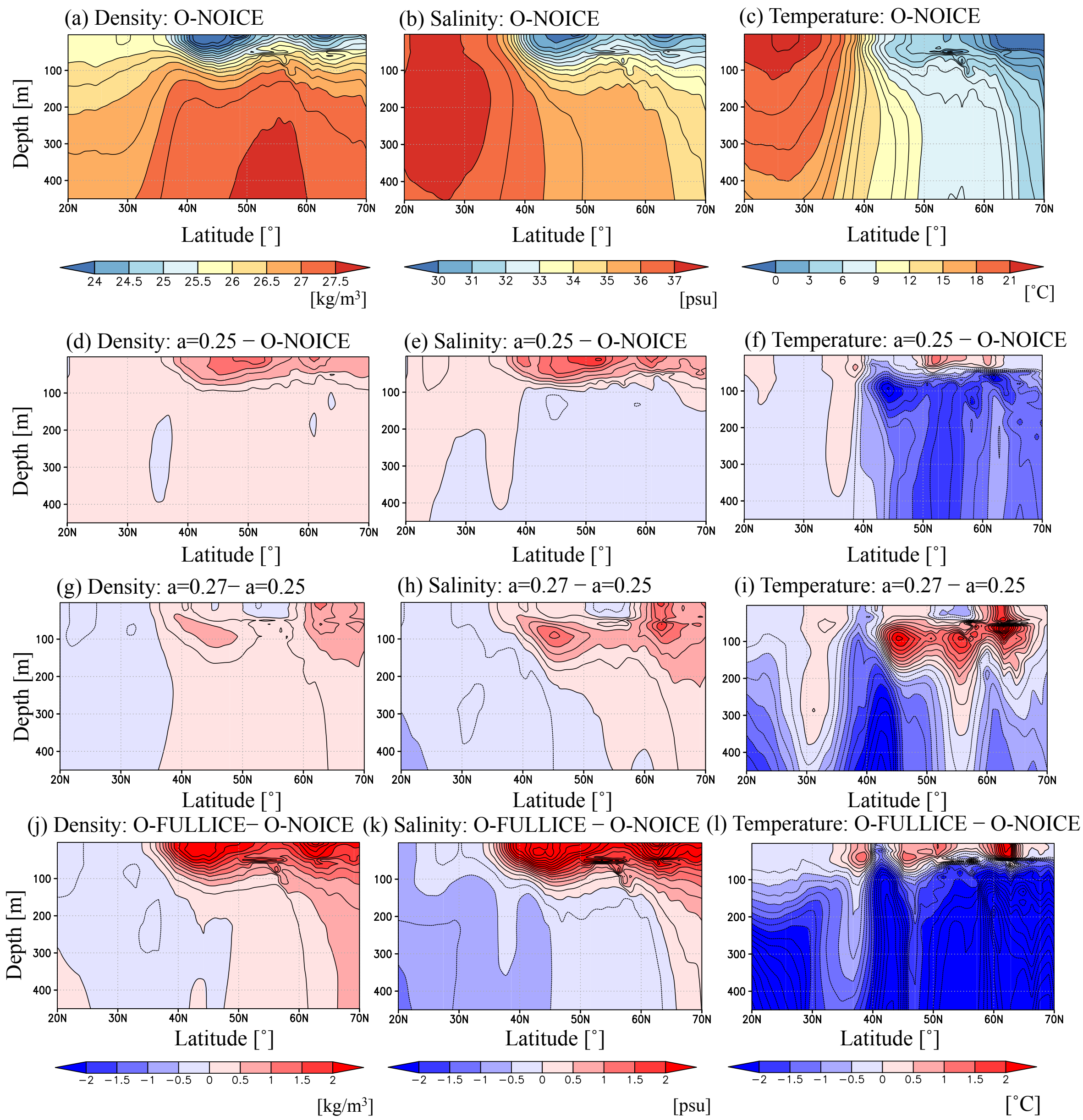

Fig. 2.12 (a) Potential density, (b) salinity and (c) potential temperature, zonally averaged across the Atlantic basin in O-NOICE (a=0). (d, e, f) the differences between their values in a=0.25 and O-NOICE. (g, h, i) the differences between their values in a=0.27 and a=0.25. (j, k, l) the differences between their values in O-FULLICE and O-NOICE.

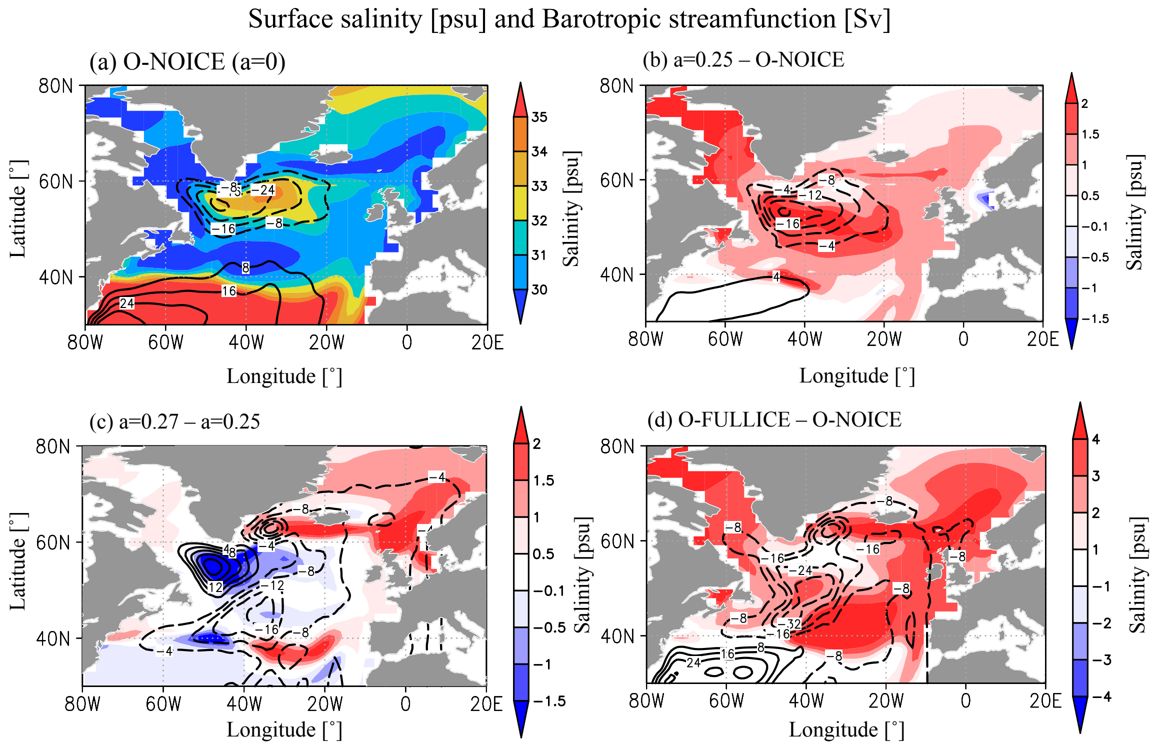

Fig. 2.13 Surface salinity (color, psu) and baratropic streamfunction (contour; m^2/s) for (a) O-NOICE (a=0) and (b) the differences between their values in a=0.25 and O-NOICE. (c) Differences between a=0.27 and a=0.25. (d) Differences between O-FULLICE and O-NOICE. Note that the color scale of (d) is different from that in (b) and (c).

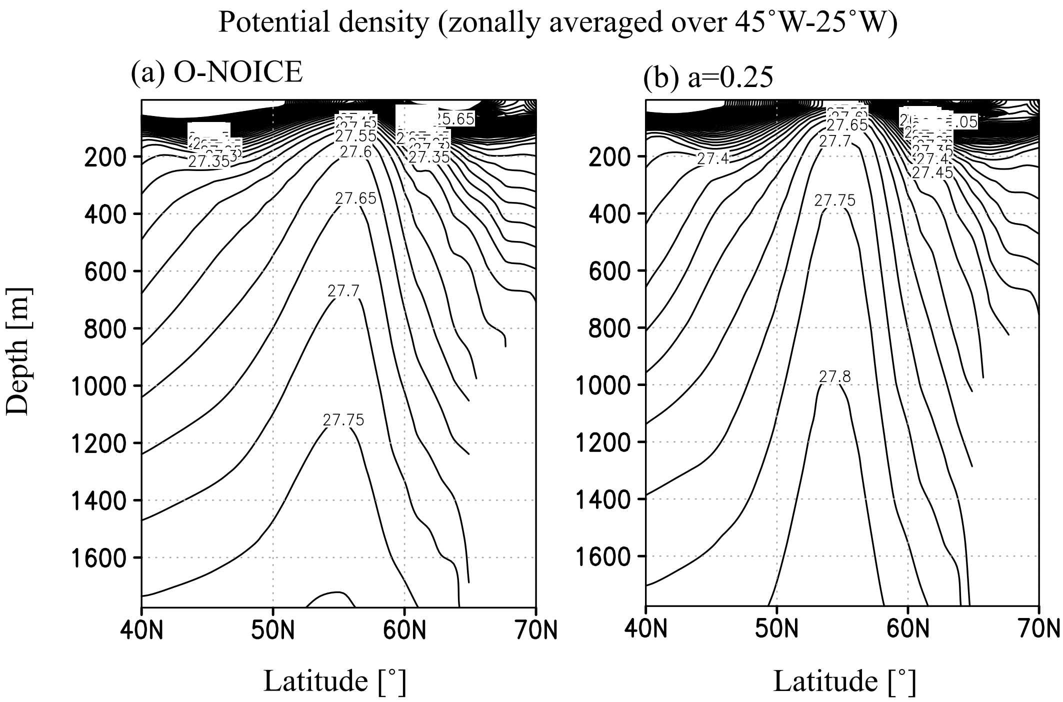

Fig. 2.14 Vertical-latitudinal cross-section of potential density (kg/m^3), zonally averaged over 45˚W-25˚W. (a) O-NOICE (a=0), (b) a=0.25. This figure illustrates the exposure of the subsurface warm and saline water to the sea surface (outcropping) by the positive wind stress curl in this region (e.g. Montoya et al. 2011).

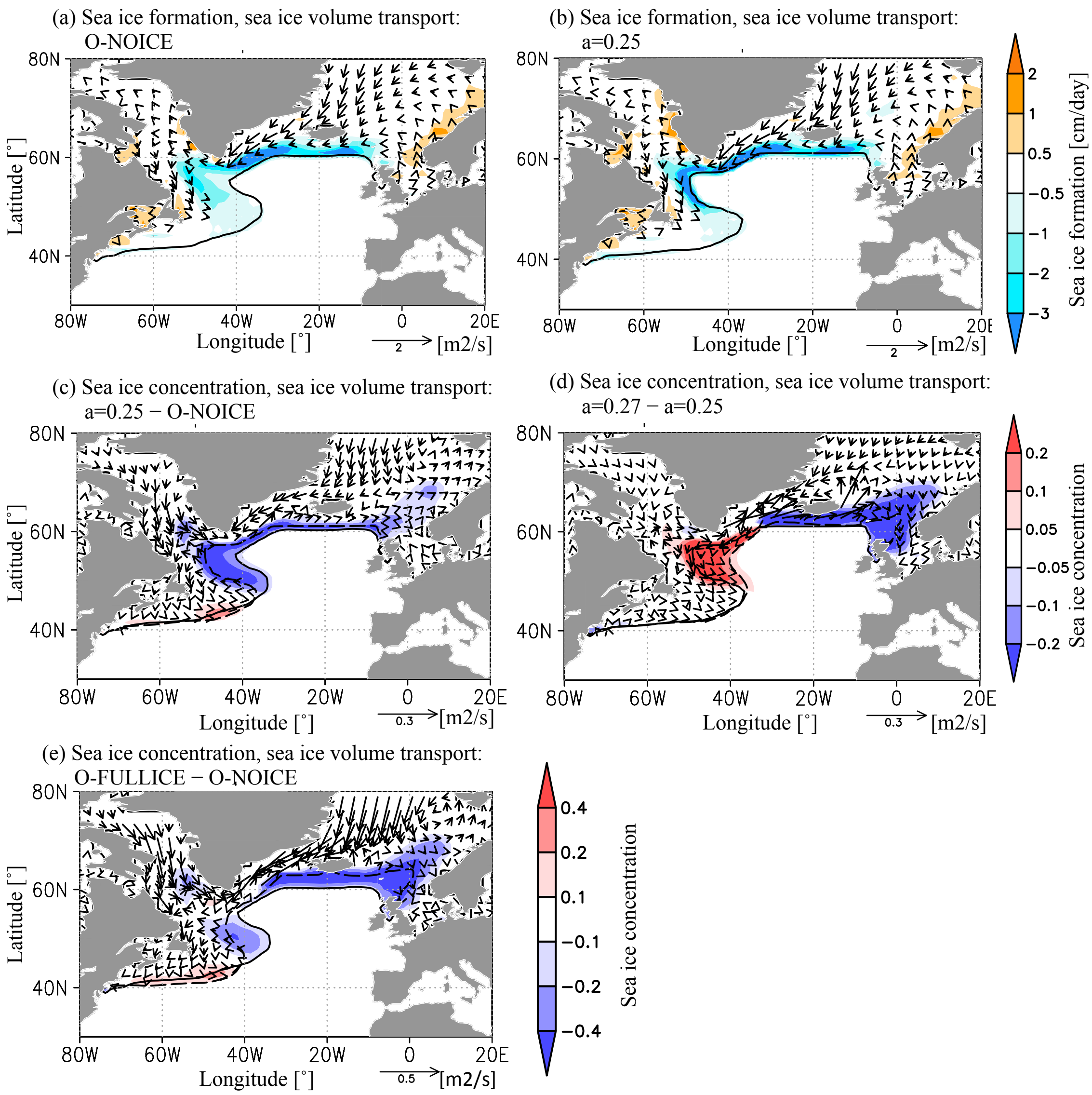

Fig. 2.15 Annual mean sea ice volume transport (arrow: m^2/s), sea ice formation/melt (color: cm/day) and 10 % sea ice concentration (contour) for (a) O-NOICE, (b) a=0.25. In (c), differences in the sea ice volume transport (arrow) and sea ice concentration (color) between a=0.25 and O-NOICE is shown. Bold and dashed contours indicate the 10 % sea ice concentration for O-NOICE and a=0.25 respectively. In (d), differences between a=0.27 and a=0.25 is shown, and in (e) differences between O-FULLICE and O-NOICE is shown. Bold and dashed contours indicate the 10 % sea ice concentration for a=0.25 and a=0.27 in (d) and for O-NOICE and O-FULLICE in (e) respectively. Note that the color and arrow scales of (e) is different from that in (c) and (d).

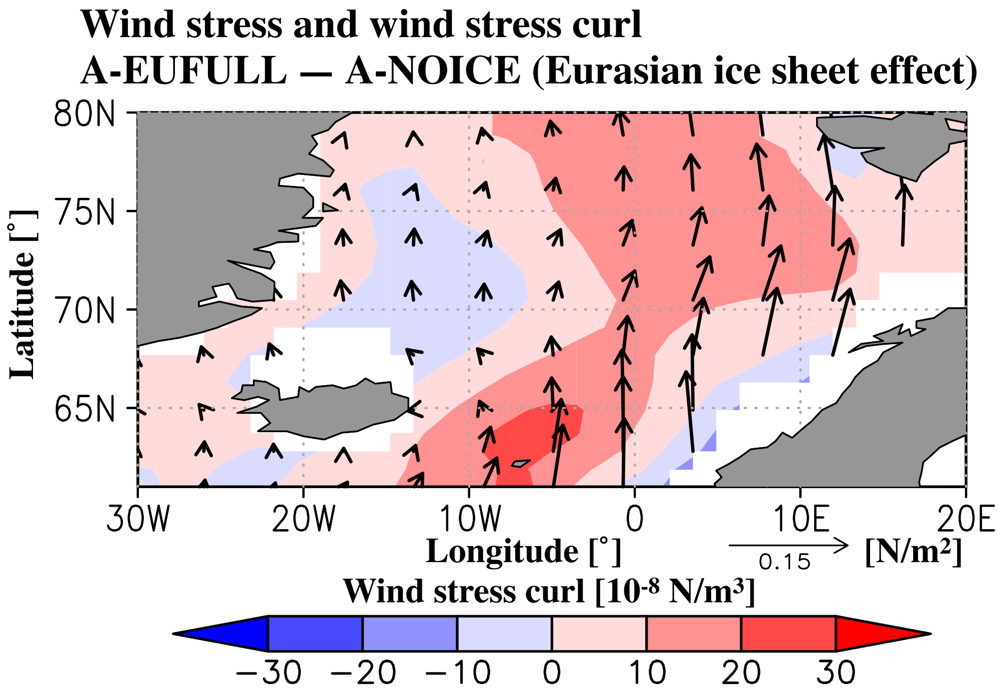

Fig. 2.16 Annual mean surface wind stress (arrow, N / m^2) and the wind stress curl (color, 10^{-8} N / m^3) over the Norwegian Sea for the Eurasian ice sheet effect (A-EUFULL - A-NOICE).

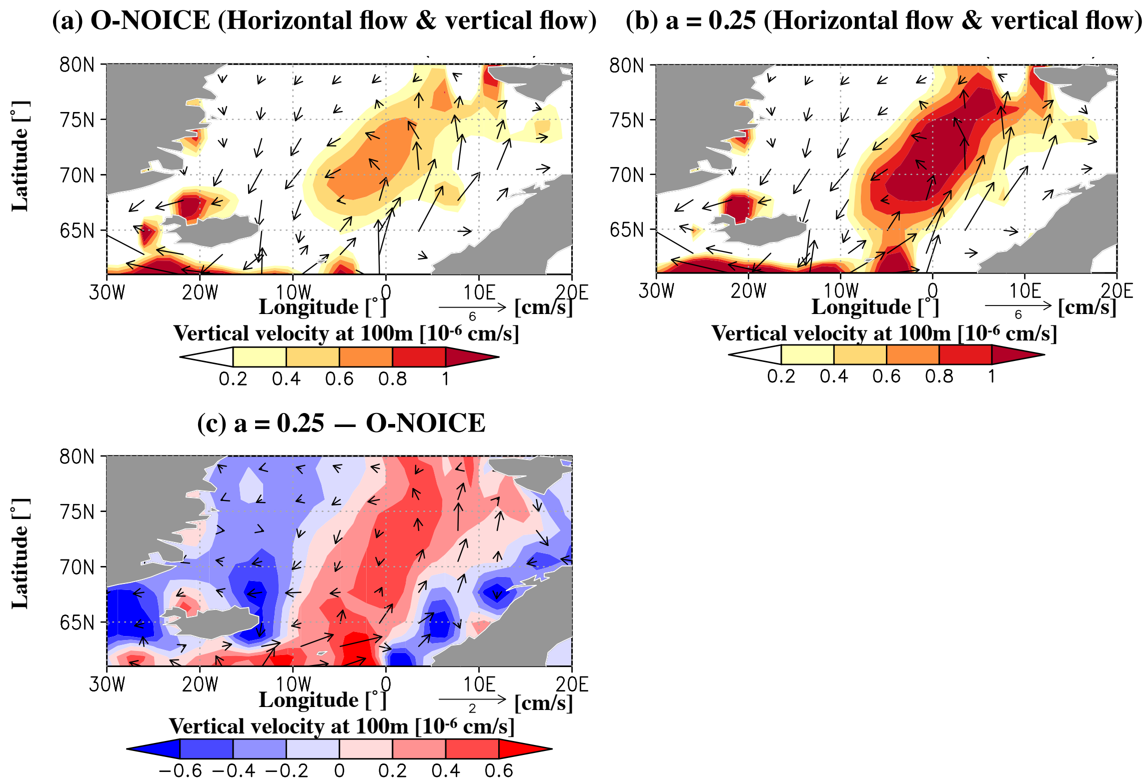

Fig. 2.17 Change in oceanic circulation before abrupt AMOC change. Annual mean horizontal velocity averaged between the surface and 100m depth (arrow: cm/s) and vertical velocity at 100m depth (color: 10^{-6} cm/s) for (a) O-NOICE, (b) a=0.25. In (c), difference between a=0.25 and O-NOICE is shown.

2.5 Discussions

From a series of sensitivity experiments using the AGCM and the OGCM, I find that the glacial ice sheets cause the AMOC to strengthen by intensifying the atmospheric circulation at the North Atlantic mid- to high latitudes, with the mid-latitudes being of primary importance. For the mid-latitude process, our decoupled simulations confirm the importance of increase in wind stress, which has been suggested in previous coupled modelling studies (e.g. Zhang et al. 2014, Gong et al. 2015). Therefore, the mid-latitude process can be considered as a robust and common feature among models. From sensitivity experiments using the AGCM and the OGCM, this intensification of the wind is attributed to the topography effect of the glacial ice sheets, especially that of the North American ice sheet, which is consistent with previous atmospheric studies (e.g. Cook and Held 1988, Li and Battisti 2008, Ullman et al. 2014, Löfverström et al. 2014, Merz et al. 2015). Our decoupled simulations also show that the strengthening of the atmospheric cyclonic circulation over the Greenland-Iceland-Norwegian Seas can have a large impact on the AMOC. This is related to the initiation of the deep-water formation in the Norwegian Sea (Fig. 2.8). To our knowledge, the role of high latitude winds related to the ice sheets has not been pointed out in previous coupled modeling studies. AGCM and OGCM sensitivity experiments show that the Eurasian ice sheet plays a role by inducing the southwesterly wind stress anomaly over the Norwegian Sea. However, there are very few studies that explore changes in the wind field over the Norwegian Sea due to the Eurasian ice sheet. Thus, it is difficult to assess the robustness of the high-latitude process. I suggest that model comparison of the wind field in this region is important.

Results from O-FULLICE, O-NAFULL and O-EUFULL show that the responses of the AMOC to wind forcing are highly nonlinear: if they are linear, the sum of the wind effects of the North American and Eurasian ice sheets should be similar to that of the full glacial ice sheet. The reason for the nonlinearity of the AMOC can be explained using Fig. 2.11. This figure shows that changes in the wind forcing can cause a drastic change in the AMOC by triggering an initiation of a new NADW formation region (e.g. Fig. 2.8). Since all of O-FULLICE, O-NAFULL and O-EUFULL have passed this transition, the AMOC is drastically intensified in these experiments. After passing this threshold, the AMOC responds linearly to a gradual increase in the wind forcing. Thus, the strength of the AMOC in the experiments where the threshold has been passed (e.g. O-NAFULL and O-EUFULL) become similar to one another, compared to those where the threshold has not been passed (e.g. O-NOICE). However, note that the general structure of the AMOC in O-FULLICE is mainly controlled by the North American ice sheet, as the strength of the AMOC, the spatial structure of the NADW cell (Fig. 2.9) and the NADW formation region of O-FULLICE (Fig. 2.8) is much closer to those of O-NAFULL. This is reasonable considering the larger wind changes due to the North American ice sheet.

In the present study, changes in the wind-driven sea ice transport worked as an effect in weakening the AMOC. This is associated with the increase of sea ice transport into the northern North Atlantic due to the enhanced northerly wind over the Labrador Sea and the Denmark Strait. Since the enhancement of the northerly wind in these regions is a common feature among models (Fig.S1 in Muglia and Schmittner 2015), this effect can be regarded as robust. However, the meridional shift of the westerlies can also greatly affect the wind-driven sea ice process, as suggested by Zhang et al. (2014). In Zhang et al. (2014), the westerlies split into northern and southern branches when they applied glacial ice sheets, resulting in weaker westerlies at their original position. They suggested that this results in a reduction of eastward sea ice transport into the NADW formation region and therefore an intensification of the AMOC. In our case, because the westerlies did not show any meridional shift but only an increase in speed, this effect may be absent. If the westerlies had shifted south (north), this may have led to a reduction (an increase) in eastward sea ice transport to the region of NADW formation, and thus modify the strength (or perhaps the sign) of the sea ice process. Since the meridional shift of the westerlies in response to glacial ice sheets depends on the model (Laîné et al. 2009, Rivière et al. 2010, Hofer et al. 2012, Merz et al. 2015, Beghin et al.

2015) and the configuration of the ice sheets (Kageyama et al. 1999, Li and Battisiti 2008, Laîné et al. 2009, Beghin et al. 2015), the effect of the sea ice process related to the meridional shift of the westerlies should also depend largely on individual simulations**.** More studies are required to understand the precise role of the sea ice process related to the meridional shift of the westerlies.

The experiments in the present study have been carried out with modern land-sea mask in order to focus on the impact of wind forcing on the AMOC. However, changes in the land-sea mask (e.g. closure of the Bering Strait and shrinking of the Greenland-Icelandic-Norwegian Sea) also have an impact on the freshwater transport (sea ice and liquid water) and the AMOC. In fact, previous studies have shown that the closure of the Bering Strait restricts the Bering throughflow from the Pacific and reduces the transport of sea ice from the Arctic to the Atlantic (e.g. Hasumi 2002, Hu et al. 2015), while the shrinking of the Greenland-Icelandic-Norwegian Sea restricts the exchange of sea ice and seawater between the Arctic and the Greenland-Icelandic-Norwegian Seas. Thus, these changes in the land-sea mask may have an impact on the high-latitude process, which is partly related to the sea ice export out of the Norwegian Sea. Since the expansion of the ice sheet restricts the northward sea ice transport out of the Norwegian Sea by narrowing the pathway into the Arctic, the efficiency of the high latitude wind may change. For the mid-latitude process, it is reasonable to assume that the wind effect is less affected by changes in the land-sea mask because this process is localized in the North Atlantic, away from the large changes in the land-sea mask. This assumption is supported by the fact that the importance of the wind forcing does not depend on the land-sea mask (OHA12, Muglia and Schmittner 2015).

The glacial ice sheet used in the present study (ICE-5G) has elevations higher than those of other reconstructions (e.g. ICE-4G, Peltier 1994; PMIP3, Abe-Ouchi et al. 2015). Some studies have shown that the difference in the glacial ice sheet distribution could alter the strength and configuration of the AMOC and sea ice distribution (Justino et al. 2006, Vettoretti and Peltier 2013, Ullman et al. 2014). Hence I conducted additional experiments, in which I applied the PMIP3 ice sheet to assess the impact of changes in the ice sheet on the wind stress in the North Atlantic (Fig. 2.18). Results show that the cyclonic surface wind weakens at mid-latitudes, while the northwesterly wind strengthens near the Labrador Sea with the PMIP3 ice sheet. This causes a decrease in the zonal mean wind stress curl over the mid-latitudes, though it does not change around 60˚N as the increase in positive wind stress curl near the Labrador Sea counteracts a decrease in wind stress curl induced by a weaker cyclonic wind field. The weaker cyclonic wind can be attributed to the lower ice sheet configuration (e.g. Zhang et al. 2014), though the mechanism of a stronger northwesterly near the Labrador Sea is not straightforward and that local differences in the ice sheet configuration may be important. However, detailed investigation on the cause of the northwesterly wind stress anomaly is beyond the scope of this study. Overall, with respect to mid-latitude wind stress curl, the wind effect on the AMOC is weaker under the PMIP3 ice sheet forcing. Although, reconstructions show both a weaker and stronger AMOC during the LGM, the fact that all of the PMIP3 models show a strong AMOC suggests that more investigations are needed for a future study, for example the processes around the Southern Ocean (e.g. Roche et al. 2012, Kobayashi et al. 2015).

Previous studies have suggested the importance of a Southern Ocean process for simulating a weaker AMOC in the LGM simulation, which is related to the production of sea ice and bottom water formation in this region (Stouffer and Manabe 2002, Schmittner 2003, Shin et al. 2003, Liu et al. 2005). Our results also suggest that a weaker wind over the North Atlantic mid-high latitudes or the increase in the sea ice transport into the North Atlantic can contribute towards the weakening of the AMOC. Since reconstructions of ice sheet height have uncertainties, it may be useful to conduct several ensemble simulations and assess whether the coupled models can reproduce the weak AMOC in their LGM simulations within the uncertainty of the ice sheet reconstructions.

I clarified the role of wind changes induced by the glacial ice sheets on the AMOC in this chapter, but the role of surface cooling and atmospheric freshwater flux (E-P-Runoff) on the AMOC remains elusive. The impact of these atmospheric conditions on the AMOC will be explored in the next chapter.

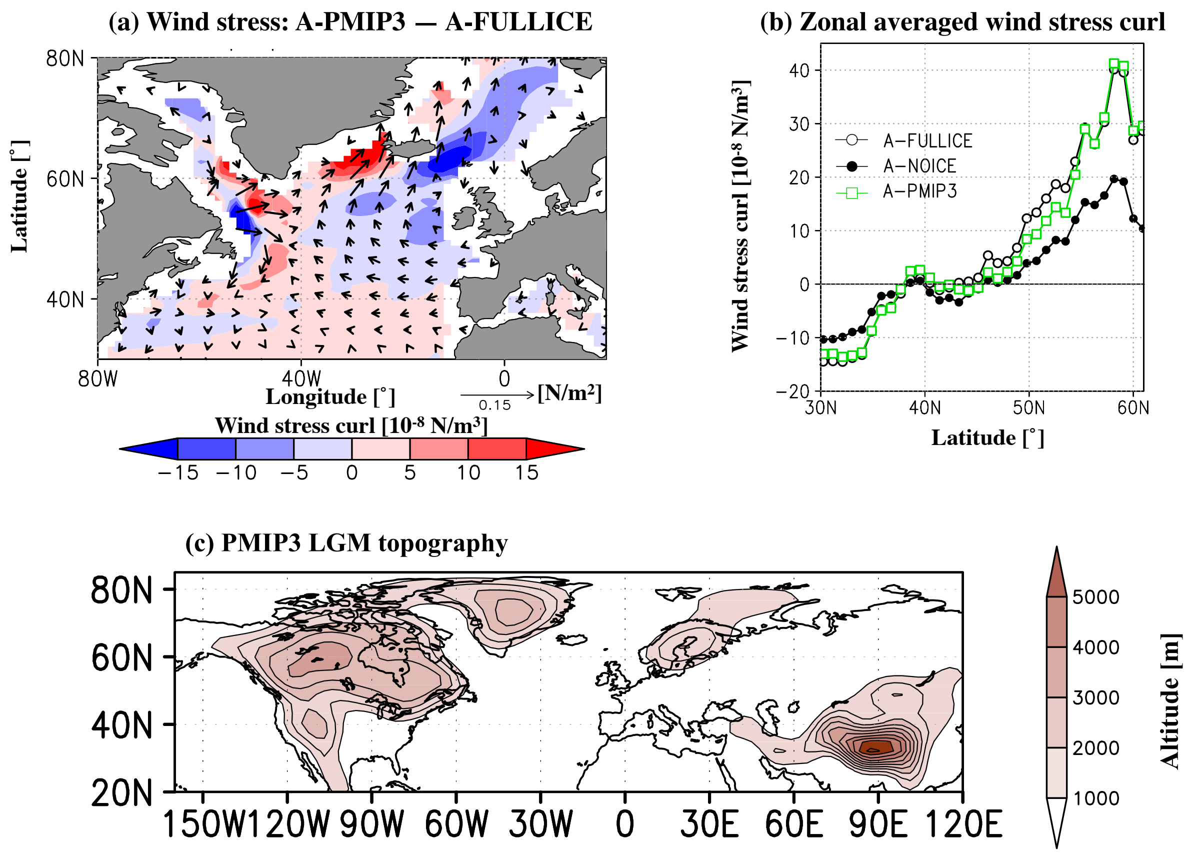

Fig. 2.18 (a) Differences in surface wind stress (arrow; N m^{-2}) and wind stress curl (color; 10^{-8} N m^{-3}) between A-PMIP3, to which I applied the PMIP3 ice sheet (Abe-Ouchi et al. 2015; green contour), and A-FULLICE. (b) Zonal averaged wind stress curl over the Atlantic basin (unit: 10^{-8} N m^{-3}). The results from A-NOICE and A-FULLICE are also shown. (c) Northern Hemisphere topography of the LGM used in PMIP3 (Abe-Ouchi et al. 2015).

2.6 Conclusions

In this chapter, I conduct a series of sensitivity experiments using the AGCM and the OGCM decoupled from the AOGCM and investigate the effect of wind change due to glacial ice sheets to the AMOC and the region where wind change is important in modifying the AMOC. Our main findings are summarized below.

- I show from decoupled experiments (A-FULLICE, A-NOICE, A-PI, O-FULLICE, O-NOICE and O-HEAT) that the surface wind change induced by the glacial ice sheets plays a crucial role in intensifying the AMOC and maintaining a strong AMOC.
- From sensitivity experiments in which I modify the wind field regionally or to which I apply wind change due to individual ice sheets (O-HIGH, O-MIDDLE, O-LOW, O-NAFULL and O-EUFULL), I find the wind change at North Atlantic mid-high latitudes to be crucial. In this region, surface wind stress and wind stress curl are enhanced due to the glacial ice sheet topography effect, especially that of the North American ice sheet. This is associated with an intensified Icelandic Low and a surface high pressure anomaly over northern North America, which are found throughout the troposphere. I also find that the Eurasian ice sheet can have an impact on the AMOC by inducing a southwesterly wind stress anomaly over the Norwegian Sea, which directly affects the deep-water formation in this region.
- The intensified wind enhances the AMOC by strengthening the wind-driven ocean circulation and, thus, the salt transport. Changes in the wind also increase the sea ice transport into the North Atlantic, which has as a weakening effect on the AMOC by increasing the sea ice melt at the region of NADW formation near the Labrador Sea. However, this effect is minor compared to those of the salt processes in our simulation (O-LINEAR). I suggest that a weaker wind over the mid-high latitudes and/or stronger sea ice transport into the North Atlantic may help couple models to produce a weak AMOC. Further investigation on this topic will be presented in Chapter 4.

In this chapter I focused on the impact of glacial ice sheet during the LGM. In the next chapter, the impact of glacial ice sheet during the mid-glacial will be explored.

第 3 章は 5 年以内に論文として Journal of Climate に出版する予定のため、非公開。出版後、公開する予定。

第 4 章は 5 年以内に論文として Geophysical Research Letters に出版する予定のため、非公開。出版後、公開する予定。

Chapter 5

General Conclusion

In this study, I explored the impact of glacial ice sheets on the AMOC and climate to improve our understandings of glacial climate and the DO cycles. I particularly investigated;

- How do glacial ice sheets modify the AMOC? What is the impact of changes in winds, surface cooling and atmospheric freshwater flux (E-P-Runoff) on the AMOC?
- Why does the impact of glacial ice sheets depend on background climate (CO2 level)?
- What kind of impact do glacial ice sheets have on the duration of stadials? Do they play a role in reducing the duration of stadials during mid-glacial period?

These were unclear in previous studies due to complicated nature of atmosphere-ocean coupled process.

In Chapter 2, I explored the impact of wind changes induced by the LGM ice sheet on the AMOC. In particular, I explored the important regions of wind changes in modifying the AMOC and the relative importance of the North American ice sheet and Eurasian ice sheet in modifying the AMOC through winds. This was important to understand why most climate models simulate a vigorous and deep AMOC in their simulations of the LGM, which contradicts the proxies. This was also important to understand how changes in glacial ice sheet modify the AMOC.

For this purpose, I conducted decoupled AGCM and OGCM experiments. In this method, I could specify changes in the winds as boundary conditions, hence I could extract the impact of glacial ice sheet winds on the AMOC, which was difficult in the AOGCM. I found that the topography of the North American ice sheets played the major role in intensifying the AMOC (Fig. 5.1). The topography of the North American ice sheet intensified the cyclonic circulation throughout the troposphere over the northern North Atlantic. This wind anomaly intensified the transport of salt by wind-driven ocean circulation. As a result, the surface salinity increased at the deepwater formation region and the AMOC intensified. The effect of surface winds was amplified by shifts in the deepwater formation region. On the other hand, I found that changes in sea ice transport induced by the wind anomaly tried to reduce the AMOC through increasing sea ice melt over the deepwater formation region. However, this effect only played a secondary role in the simulation. I also found that the wind-driven sea ice effect on the AMOC depend largely on the models or ice sheet configurations.

In Chapter 3, I explored the impact of glacial ice sheets on the duration of stadials. For this purpose, I conducted freshwater hosing experiments with MIROC4m AOGCM and compared the recovery time of the AMOC between two ice sheet configurations (mid-glacial and early glacial). I further evaluated the impacts of changes in winds, surface cooling and atmospheric freshwater flux (E-P-Runoff) induced by the glacial ice sheets on the recovery time of the AMOC. This remained unclear in previous studies and in Chapter 2. For this purpose, I performed partial decouple experiments with an AOGCM for the first time at least in the paleoclimate modeling community. In this method, the atmospheric forcing that drives the ocean model in the coupled model is switched to a different forcing. By comparing the results, with and without the replacement, one can evaluate the impact of atmospheric forcing (e.g. surface cooling) on the AMOC.

I found that the expansion of glacial ice sheets from the early glacial to mid glacial decreased the recovery time of the AMOC. This result showed that the changes in glacial ice sheet played an important role in inducing short stadials during mid-glacial. Partial decouple experiments revealed that differences in the winds were important in causing the shorter recovery. On the other hand, the cooling effect of the glacial ice sheet appeared to reduce the AMOC and increased the recovery time of the AMOC. Thus I clarified for the first time that the wind and the cooling effects on the AMOC were completely different. Analysis showed that the cooling effect reduced the AMOC through increasing the sea ice in the deepwater formation regions (Fig. 5.2). This strengthened the suppression of atmosphere-ocean heat exchange and hence suppressed buoyancy loss of the ocean. Partial decouple experiment also showed that changes in atmospheric freshwater flux (E-P-Runoff) decreased the recovery time. This was related to a southward shift of the mid-latitude rain belt due to the larger ice sheet (Fig. 5.3). However, the effect of the freshwater flux was smaller compared to the wind and surface cooling effects.

In chapter 4, I explored the reason why the impact of glacial ice sheets on the AMOC depended on background climate (e.g. CO2 level). In particular, I focused on the role of sea ice, which was suggested by previous studies (Kawamura et al. 2017, Abe-Ouchi et al. personal communication). For this purpose, I first investigated the cause of the weakening of the wind induced by changes in sea ice and the impact of wind changes on the AMOC. I then applied the results to interpret the background climate dependence of glacial ice sheet effect on the AMOC. Analysis of MIROC4m and simulations with AGCM showed that the expansion of sea ice over the northern North Atlantic caused a weakening of the surface wind. The weakening of the surface wind was mainly induced by the suppression of sensible heat flux, which increased the static stability of the air column near the surface and induced the anti-cyclonic circulation anomaly near the surface. The suppression of precipitation (latent heat flux) had a small impact on the winds on its own, although it was shown that precipitation had a significant impact on the amplitude of the wind anomaly.

Partial decouple experiments clearly showed that the weakening of surface winds due to sea ice expansion played a role in maintaining a weak AMOC. Thus, it is shown that the feedback by the wind works as a positive feedback and hence plays a role in inducing contrasting AMOC and climate. This result showed that the interaction of sea ice and surface wind is the key for modifying the impact of glacial ice sheets on the AMOC. Under warm climate and less sea ice condition, the glacial ice sheets could intensify the surface winds and the AMOC. However, under cold climate, the sea ice expanded and suppressed the atmosphere-ocean heat exchange. As a result, the surface winds could not become vigorous even under the existence of the glacial ice sheets. Thus, the strength of the wind effect by the glacial ice sheets reduced. As a result, the glacial ice sheet weakened the AMOC through the cooling effect. Therefore, the interaction of sea ice and surface wind played an important role for the background climate (CO2) dependence of the ice sheet effect on the AMOC.

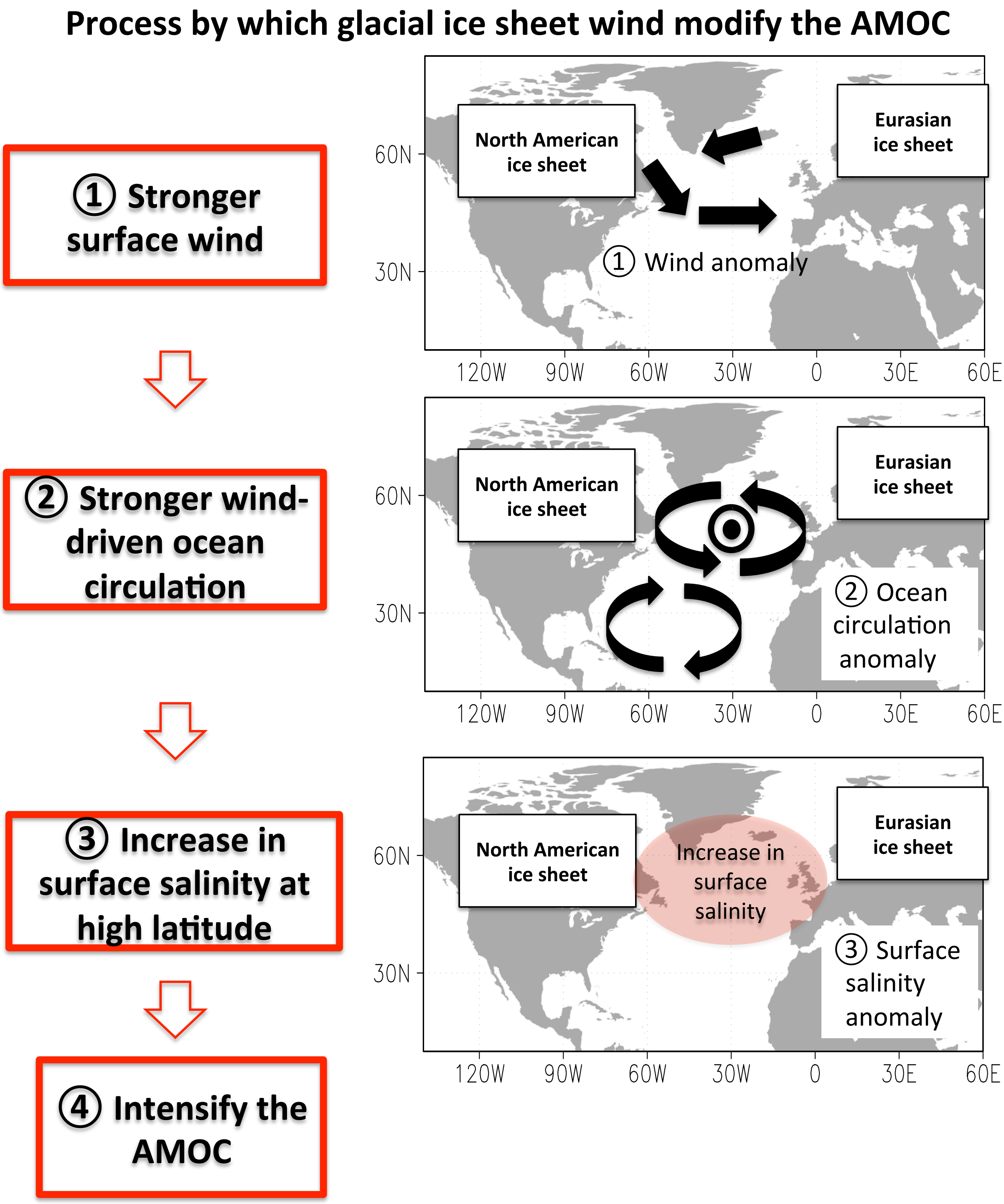

Fig 5.1 Schematic figure of processes by which glacial ice sheets modify the AMOC through surface winds.

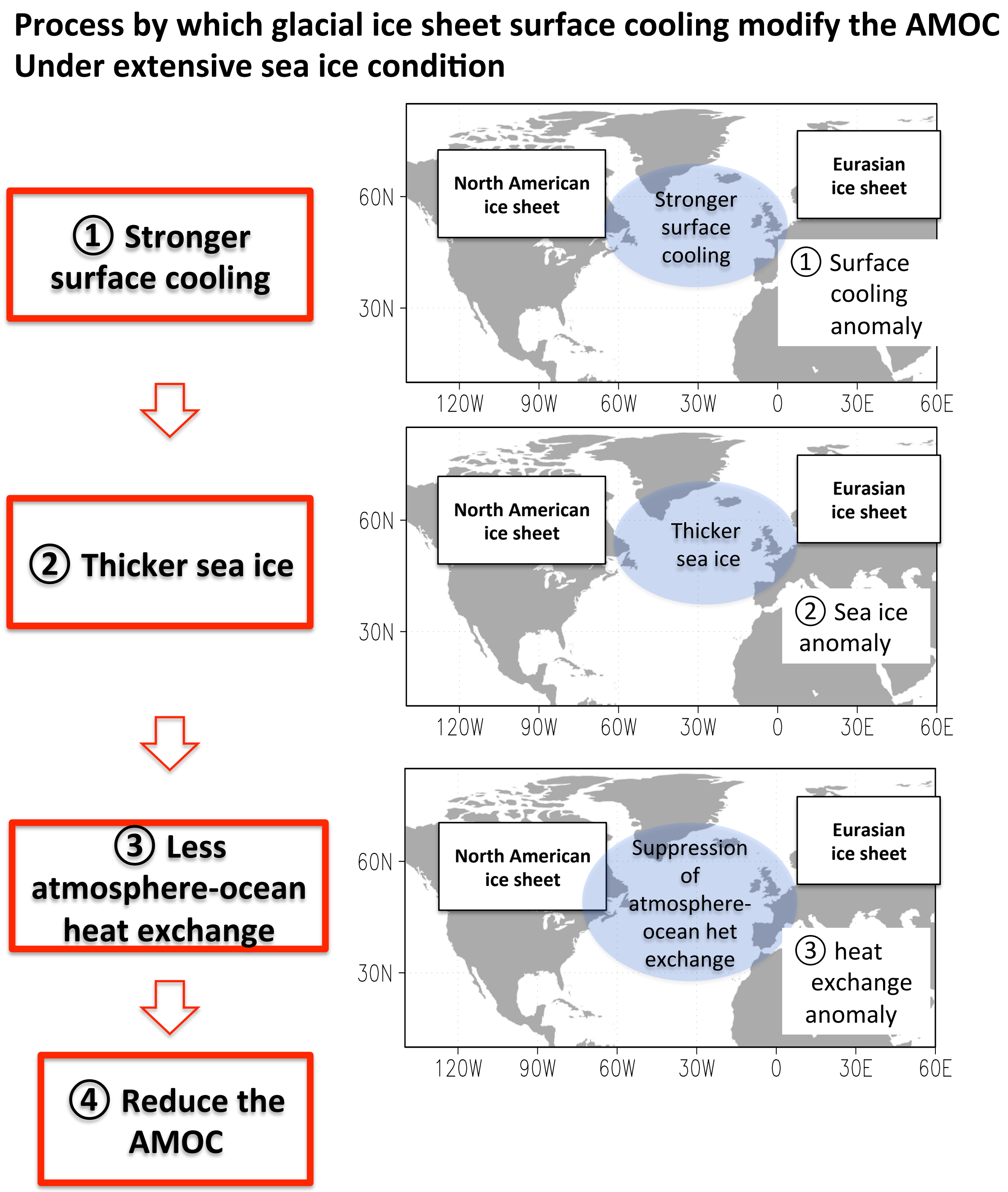

Fig 5.2 Schematic figure of processes by which glacial ice sheets modify the AMOC through surface cooling under extensive sea ice condition.

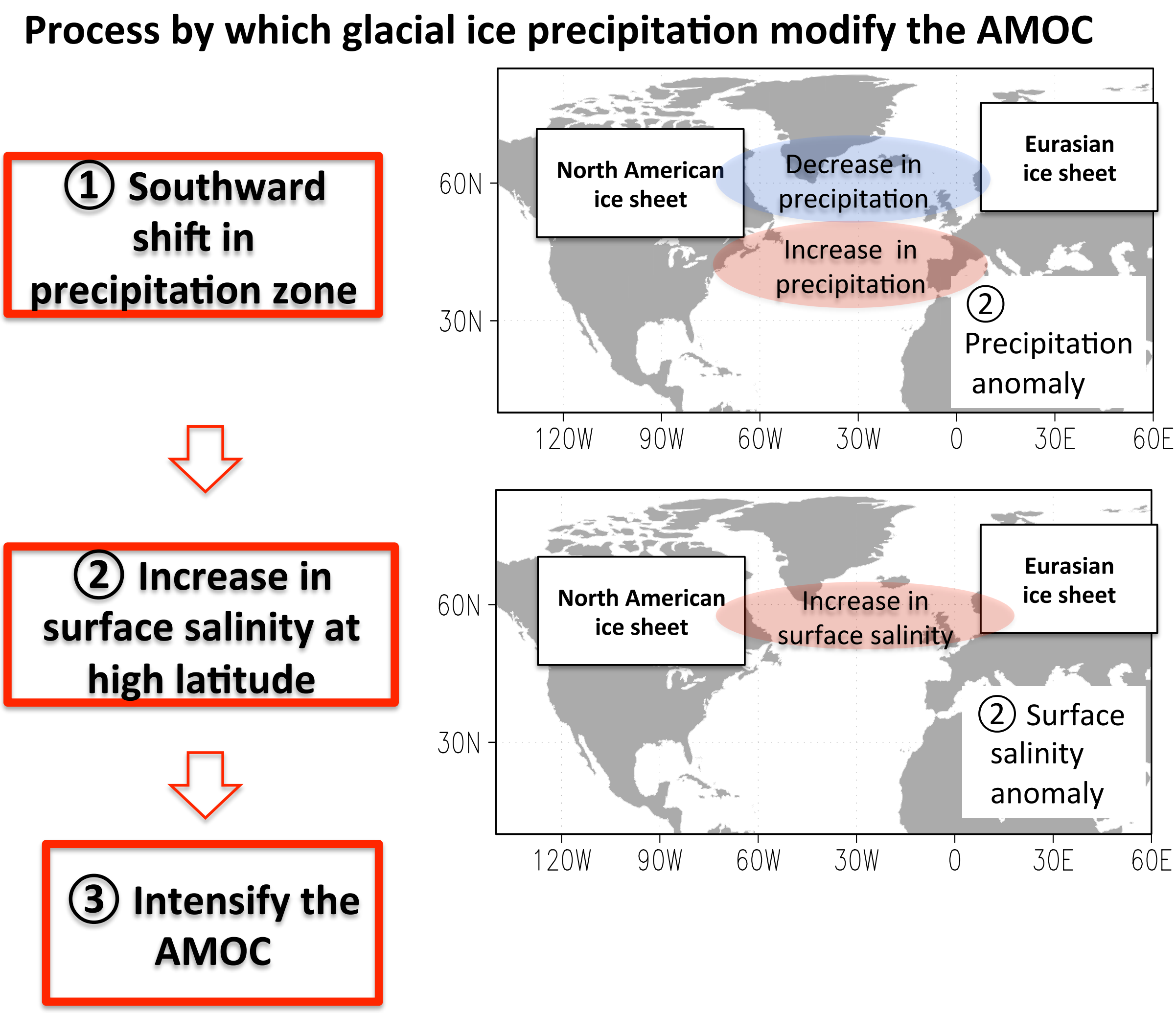

Fig 5.3 Schematic figure of processes by which glacial ice sheets modify the AMOC through atmospheric freshwater flux (E-P-Runoff).

Together all, this study can be summarized in Figure 5.4. I show that the strengthening of the winds by the glacial ice sheets increases the surface salinity over the deepwater formation region through enhancing the wind-driven ocean circulation. The increase in surface salinity enhances the AMOC. Changes in the wind can also strengthen the AMOC through affecting the wind-driven sea ice transport, which modifies the atmosphere-ocean heat exchange and sea ice melt. However, this effect depends on models and ice sheet configurations. On the other hand, I also show that strengthening of the surface cooling by the glacial ice sheets increases the sea ice over the deepwater formation. This directly weakens the AMOC through suppressing the atmosphere-ocean heat exchange. Thus, this study clarify that the impact of glacial ice sheets on the AMOC and climate is controlled by the relative strength of the surface winds and surface cooling.

The relative strength of the wind and cooling effects is found to depend on the background climate (CO2, insolation). For this, interaction of sea ice and surface wind plays the key role (Fig. 5.4). The expansion of sea ice has an effect to weaken the surface wind through suppressing the heat flux from the ocean to the atmosphere. Thus, as the climate cools and sea ice expands due to a decrease in CO2, the strength of the wind effect weakens. This causes a switch between the relative strength of the wind and cooling effects. For this reason, the impact of the glacial ice sheets on the AMOC depends on the background climate (e.g. CO2, insolation). Also, since the strength of each effect depends on the model, differences in the relative strength of the wind and cooling effects among models can cause large model discrepancy.

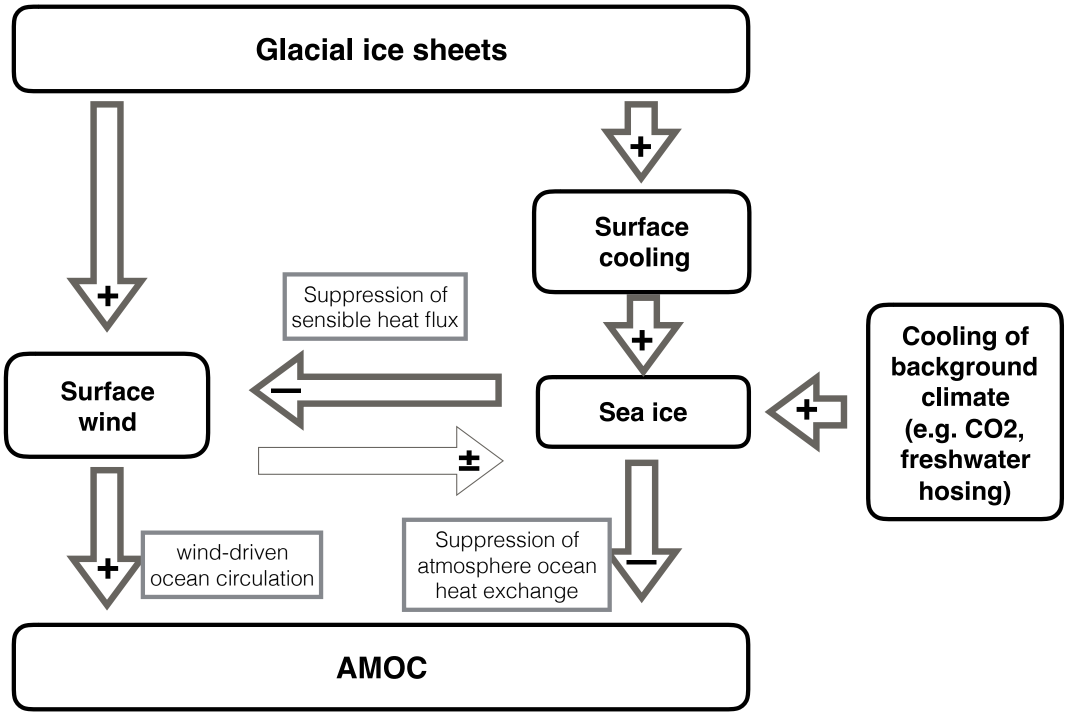

Fig 5.4 Schematic of processes by which glacial ice sheets modify the AMOC. The sign on the arrows are linked to the impact of the forcing on the target.

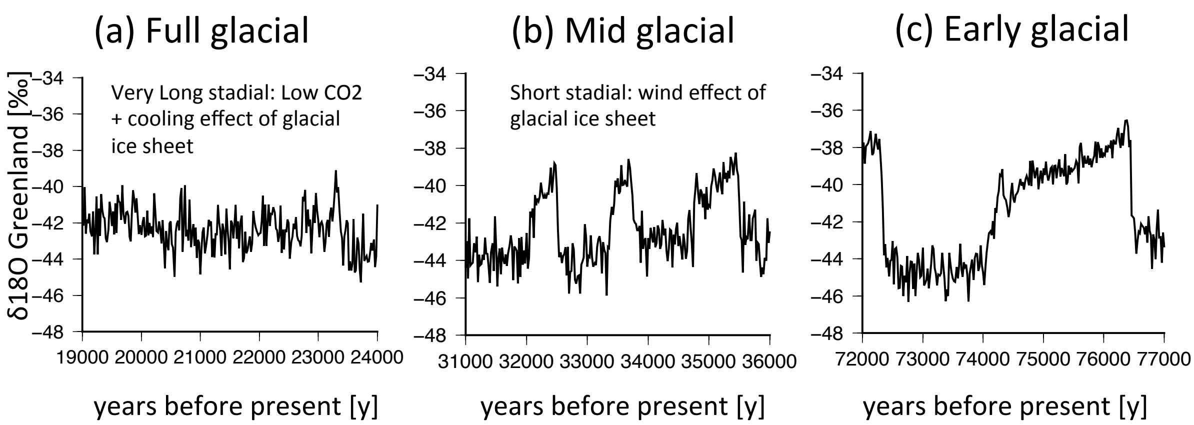

Fig. 5.5 Characteristics of typical DO cycles during (a) full glacial, (b) mid glacial and (c) early glacial. Impact of glacial ice sheets on the duration of stadials suggested from this study is described. Data of Rasumussen et al. (2012) is used.

Our results are applicable for the interpretation of proxy data, especially the changes in duration of stadial during glacial period (Fig. 5.5). Buizert and Schmittner (2015) showed that the duration of the stadial is shortest during the mid-glacial period compared to the full glacial and earlier glacial periods. Since the lowering of temperature over the Southern Ocean or lowering of CO2 weakens the AMOC (Buizert and Schmittner 2015, Kawamura et al. 2017), the short stadial during the mid-glacial cannot be explained only by the Antarctic temperature or CO2 (Buizert and Schmittner 2015). Our study suggests that the glacial ice sheet play an important role. Over the early glacial to mid-glacial period, since the climate is relatively warm, it is reasonable to assume that the amount of sea ice is small. Under this warm condition, the expansion of the glacial ice sheet can reduce the duration of the stadials through strengthening the surface winds. Thus, our result suggests that the expansion of the glacial ice sheet plays an important role in inducing frequent DO cycles during mid-glacial period. Over mid-glacial to full glacial period, the duration of the stadial increases drastically (Buizert and Shcmittner 2015). Although the contribution from the lowering of CO2 itself may cause this very long stadial, our results suggest that the increase in sea ice by the glacial ice sheets (cooling effect) can contribute to the increase the duration of stadials.

The most significant points in this study can be summarized as follows.

1. I clarified the impact of changes in winds, surface cooling and atmospheric freshwater flux (E-P-Runoff) by the glacial ice sheets on the AMOC. This was possible through decoupled AGCM, OGCM and partial decouple AOGCM experiments. Our results suggested that the differences in the relative strength of the each process among models cause differences in the response of AMOC to glacial ice sheets.
2. I suggested that the background dependence of glacial ice sheet effect, which arises from the wind and cooling effect, played a role in controlling the duration of stadials over the glacial period.
3. I performed partial decouple experiments with the AOGCM for the first time at least in the paleoclimate modeling community. Partial decouple experiments are a very useful

method to understand the results of AOGCM, which is very complicated. In fact, this method can be applied not only to investigate the role of glacial ice sheets on the ocean circulation, but also to investigate the role of other orographic forcing on the ocean circulation, such as the Himalaya, Rookie, Greenland and so on. Moreover, this method can be applied for investigations of the interaction of atmosphere and ocean, which is critical for decadal-to millennial time scale climate variability. Because of the increase in computational resources as well as development of efficient techniques to run the AOGCMs, a lot of studies have started to use the AOGCMs instead of EMICs or OGCMs in the paleoclimate modeling community. This study will serve as a reference for interpreting the results of complex AOGCMs.

Future studies

There are mainly four points that needs to be investigated in the future. The first point is a better understanding of model discrepancies. In this study, I clarified that the impact of glacial ice sheets on the AMOC is determined by the combination the wind effect, cooling effect and sea ice. Since the strength of each effect depends on models, the total impact of glacial ice sheets on the AMOC diverge among models. Thus it is important to understand the cause of differences in the wind, surface cooling and sea ice among models. Two methods can be used for the investigations. The first method is to perform a detailed comparison among several models. For example, a detailed comparison of the strength atmosphere-ocean heat exchange over the northern North Atlantic may give useful information, because it plays an important in modifying the surface wind, deepwater formation and sea ice. In the second method, parameter studies with an AGCM should give useful information. For example, it is interesting to explore how the wind and cooling effects are affected by changes in the extent and height of the ice sheets. By modifying the height and extent of the ice sheet in the AGCM, we can get a general view of how changes in the shape of the ice sheets modify the wind and surface cooling. This will help to interpret differences in the atmospheric circulation between models. Also, since the shape of the glacial ice sheet altered drastically with respect to the extent and height over the glacial periods (e.g. Abe-Ouchi et al. 2013), this will give useful information when comparing a glacial period to a different glacial period.

The second point of future studies is to better understand the role of glacial ice sheet on DO cycles that is excited without external forcing (internal oscillation within the atmosphere-sea ice-ocean system, e.g. Peltier and Vettoretti 2014, Brown and Galbraith 2016). Although the mechanism of the recovery of the AMOC from a weak mode to a vigorous mode seems similar to that in the freshwater hosing experiment (e.g. Brown and Galbraith 2016, Vettoretti and Peltier 2016), it is important to assess the role of glacial ice sheets in modifying the duration of the recovery time. In addition, Timmerman et al. (2003) and Brown and Galbraith (2016) showed that this internal oscillation appears in a particular parameter space. It is important to assess whether the existence of glacial ice sheets plays a role in inducing the internal oscillation or not. Also it is important to explore the impact of glacial ice sheets and CO_2 on the characteristics of the internal oscillations (frequency, amplitude and abruptness). Very long model integrations with several ice sheet configurations will be a first step to explore this topic.

The third point of future studies is to explore the full evolution of DO cycles. In this study, I fix the ice sheet configuration and CO_2 during the experiment, though these have changed during stadials (e.g. Grant et al. 2012, Buizert and Schmittner 2015). Thus the role of temporal changes in ice sheet height and CO_2 concentration during stadials remains elusive. It would be interesting to perform experiments, in which I modify the height of the ice sheets and CO_2 concentrations during stadials, following the evidence of sea level change and CO_2 change from proxies. Also, it would be very interesting to perform a fully coupled climate model simulation (AOGCM coupled with an ice sheet model and a carbon cycle model).

The main focus of this study was to clarify the impact of glacial ice sheets on the climate. However, for a complete understanding of the glacial climate, it is important to understand the role of changes in insolation as well. In fact, a recent study showed that changes in obliquity might play an important role on the AMOC (Turney et al. 2015). Thus, it is important to assess the relative importance of glacial ice sheets, CO2 and insolation on the AMOC. This is the forth point of the future study.

Although, not explored in detail in this study, there are several future studies that should be conducted. The first study is related to the abruptness of the recovery of the AMOC. Reproducing an abrupt recovery of the AMOC, which resembles the ice core data, was difficult and has been one of the main targets in paleoclimate modelers (e.g. Clement and Peterson 2008). Recent studies succeeded in simulating an abrupt recovery of the AMOC within the range of the ice core data (less than 100 years, Zhang et al. 2014), though it is still difficult in other models to reproduce (Zhang et al. 2014). In addition, although some models show a consistent abrupt recovery with proxy, the reason for this is not fully understood (Zhang et al. 2014). Thus, detailed analysis on the abrupt recovery of the AMOC is important.

The second study is associated to the impact of DO cycles on the glacial-interglacial cycle. During DO cycles, the surface air temperature and snowfall over the ice sheet changes drastically and can affect the mass balance of the ice sheets. Since the glacial ice sheets modify the duration of stadials, changes in the duration of stadial can feedback on the mass balance of the ice sheet. This can further affect the frequency or amplitude of the glacial-interglacial cycle. It would be interesting to explore the impact of DO cycles on the glacial-interglacial cycle.

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