Climate response to the meltwater runoff from Greenland Ice Sheet: evolving sensitivity to discharging locations

Yonggang Liu[1], Robert Hallberg[2], Olga Sergienko[3], Bonnie L. Samuels[2], Matthew Harrison[2], and Michael Oppenheimer[4,5]

1. Department of Atmospheric and Oceanic Sciences, School of Physics, Peking Univeristy, Chengfu Road No. 209, Haidian, Beijing 100871, China
2. NOAA Geophysical Fluid Dynamics Laboratory, Ocean Group, 201 Forrestal Rd., Princeton, NJ 08540, USA
3. Program in Atmospheric and Oceanic Sciences, Princeton University, 201 Forrestal Road. Princeton, NJ 08540, USA
4. Woodrow Wilson School of Public and International Affairs, Princeton University, Princeton, NJ 08540, USA
5. Department of Geosciences, Princeton University, Princeton, NJ 08540, USA

Contact information of the corresponding author

Email: ygliu@pku.edu.cn

Phone: +86 10 62769125

Abstract Greenland Ice Sheet (GIS) might have lost a large amount of its volume during the last interglacial and may do so again in the future due to climate warming. In this study, we test whether the climate response to the glacial meltwater is sensitive to its discharging location. Two fully coupled atmosphere-ocean general circulation models, CM2G and CM2M, which have completely different ocean components are employed to do the test. In each experiment, a prescribed freshwater flux of 0.1 Sv is discharged from one of the four locations around Greenland - Petermann, 79 North, Jacobshavn and Helheim glaciers. The results from both models show that the AMOC weakens more when the freshwater is discharged from the northern GIS (Petermann and 79 North) than when it is discharged from the southern GIS (Jacobshavn and Helheim), by 15% (CM2G) and 31% (CM2M) during the first two (CM2G) or three (CM2M) hundred years. This is due to easier access of the freshwater from northern GIS to the deepwater formation site in the Nordic Seas. In the long term (>300 year), however, the AMOC change is nearly the same for freshwater discharged from any location of the GIS. The East Greenland current accelerates with time and eventually becomes significantly faster when the freshwater is discharged from the north than from the south. Therefore, freshwater from the north is transported efficiently towards the south first and then circulates back to the Nordic Seas, making its impact to the deepwater formation there similar to the freshwater discharged from the south. Such adjustment of the ocean circulation within the Nordic Seas should be observed in the model simulations with relatively high horizontal resolution and if the discharged freshwater is not directly injected at a region with strong formation of deepwater.

Key words: Climate change; Atlantic Meridional Ocean Circulation (AMOC); Greenland Ice Sheet; freshwater forcing; hosing experiment

1. Introduction

Injections of large bodies of freshwater into the northern oceans have a substantial impact on the North Atlantic deepwater formation. They also impact the northward heat transport by the Atlantic Ocean that, in its turn, affects the northern climate (Ganopolski and Rahmstorf 2001; Kageyama et al. 2010; Manabe and Stouffer 1995; Stouffer et al. 2006). The freshwater can come from either increased net precipitation and river runoff over the North Atlantic region that would occur in a warmer climate (Cheng et al. 2013), abnormal sea-ice transport to deepwater formation regions from the Arctic Ocean (e.g. the Great Salinity Anomaly event that occurred in the late 1960s (Dickson et al. 1988), and its influence on AMOC (e.g. Gelderloos et al. 2012), or from meltwater discharge and calving from ice sheets. The meltwater from ice sheets is expected to enter the ocean in a channelized form like river runoff. For example, during the deglaciation phase of the glacial cycles (the most recent one being that from ~21 thousand years ago (ka) to approximately ~8 ka), large fraction of meltwater from the Laurentide ice sheet sitting on the continent of North America probably entered the Atlantic Ocean primarily through Hudson Strait, Hudson river and St. Lawrence river (e.g. Licciardi et al. 1999), entered Gulf of Mexico through Mississippi river (e.g. Aharon 2003), and entered the Arctic Ocean through Mackenzie river (e.g. Tarasov and Peltier 2005). In the future, large amount of meltwater may be discharged from the Greenland Ice Sheet (GIS) and Antarctic Ice Sheet. The largest supply of this meltwater may enter the ocean through narrow “exit gates” for fast flowing ice streams or outlet glaciers (Fig. 1). These localized freshwater forcing may impact the ocean circulation and climate differently from a distributed forcing such as increased net precipitation.

In early modeling studies of climate response to freshwater forcing, the forcing was generally distributed over a large area of North Atlantic, e.g. from 50°N to 70°N, due to limitations of the ocean general circulation models in handling large local water fluxes (Stouffer et al. 2006). As models are improved in this aspect, more and more studies have appeared investigating the sensitivity of climate response to the location of freshwater forcing. Kleinen et al. (2009), for instance, has varied the freshwater forcing locations within the North Atlantic and Arctic Ocean. The forcing regions in their experiments, however, were still very large, spanning more than 50° in longitude and 10° in latitude, thus almost no difference in the response of AMOC between experiments emerged (see their Fig. 1). In a more systematic study, Roche et al. (2010) have demonstrated that the climate response was sensitive to the locations of freshwater discharge when the discharging regions were more localized. In their study, the base state upon which the freshwater forcing was applied was Last Glacial Maximum (LGM) but the physical mechanism they propose may be applicable to the present-day situation too. They tested 7 different locations within the North Atlantic and Arctic Ocean using a relatively coarse climate model (3°×3° horizontal resolution in the ocean), and found that the AMOC and Arctic sea ice were more sensitive to freshwater pulses that were applied near or upstream of the deepwater formation regions of the North Atlantic. Using a high-resolution (1/6°) ocean-only model, Condron and Winsor (2012) showed that AMOC was more significantly impacted if the freshwater was discharged into the Arctic Ocean from Mackenzie Valley than into the North Atlantic Ocean through the St. Lawrence Valley. In this study, however, the discharge of freshwater lasted for only one year and the whole simulation lasted for 25 years. It is, therefore, unclear how the climate response may differ if both the discharge and the simulations last much longer and the feedbacks from the atmosphere are included. Nonetheless, these latter two studies do suggest that the ocean currents should play an important role in modulating the impact of freshwater forcing on climate.

In this study, we focus on the climate impact of freshwater discharge from different sectors of GIS. We consider four locations, approximately coincident with the currently most rapidly discharging glaciers around GIS, specifically the Petermann and 79 North glaciers in the north, and the Jacobshavn and Helheim glaciers in the south (Fig. 1) (Joughin et al. 2010a; Joughin et al. 2010b). Additional simulations are done with the same amount of freshwater being discharged uniformly over the North Atlantic region from 50°N to 70°N (Fig. 1b, referred to as 'NA' experiment hereafter) as a comparison. We also test the sensitivity to the formulation of climate models by performing the identical experiments with two different fully coupled atmosphere-ocean general circulation models (AOGCM), CM2G and CM2M, both developed at Geophysical Fluid Dynamics Laboratory (GFDL) of National Ocean Atmosphere Administration (NOAA) and both participated in the Coupled Model Intercomparison Project Phase 5 (CMIP5). These two models have completely different ocean component and different characteristics of the deep convection sites in the North Atlantic.

The oceanic components of both models are run at the nominal 1° resolution for the ocean (see Section 2). This resolution is significantly higher than that used in Roche et al.

(2010) (3°×3°), as a result, the boundary currents are better resolved. Moreover, all the freshwater forcing regions are restricted to within one grid cell, rather than many cells as in Roche et al. (2010). The increased resolution and localized freshwater discharge may improve the fidelity of the results in terms of how freshwater is transported to the deepwater formation sites in the North Atlantic (Condron and Winsor 2011). The general conclusion reached by Roche et al. (2010) may not be applicable to the simulations done here.

The resolution used in this study is substantially lower than the 1/6° resolution used by Condron and Winsor (2011). As we use fully coupled climate models and run the simulations much longer, the higher resolution is prohibitively expensive. The coupling between the atmosphere and ocean is desired since it has been demonstrated by Stammer et al. (2011) that the climate response of a coupled model to GIS freshwater forcing may be significantly faster and larger than that of an ocean only model. The atmosphere can transfer the perturbative signals in the North Atlantic Ocean to other ocean basins within one year, rather than many tens of years for the ocean-only model. The change of surface latent heat flux over the northern North Atlantic in the coupled model provides a positive feedback to the initial freshwater perturbation by further reducing surface density of the northern North Atlantic Ocean, and causes much larger weakening of AMOC than that in the ocean-only model.

Many studies have examined the climate response to meltwater input from the GIS (Bakker et al. 2016; Gerdes et al. 2006; Hu et al. 2011; Hu et al. 2013; Jungclaus et al.

2006; Stammer 2008; Stammer et al. 2011; Swingedouw et al. 2007; Vizcaino et al. 2008). The models and experiments were set up very differently among these studies, but they all showed that the AMOC are weakened by the freshwater forcing. This weakening of AMOC and the associated reduction of northward heat transport are, however, not large enough to overcome the warming of the global climate, but do lessen the warming of the high northern latitude (e.g. Hu et al. 2011; Swingedouw et al. 2006). These studies are not focused on how AMOC changes differently if the freshwater is discharged from different sectors of GIS. In many of these studies, the meltwater has been discharged from the southern GIS (Hu et al. 2011; Hu et al. 2013; Stammer 2008; Stammer et al. 2011), and in others discharged uniformly from the whole ice sheet (Gerdes et al. 2006; Jungclaus et al. 2006; Swingedouw et al. 2013; 2015) or quasi-uniformly (Bakker et al. 2016).

A more recent study starts to test the sensitivity of the AMOC response to the discharging locations around the GIS. Using an AOGCM, the Bergen Climate Model, Yu et al. (2016) applied freshwater forcing of 0.1 Sv from the east and west edges of GIS, respectively, and compared with the experiment in which the same amount of freshwater is applied uniformly from the whole GIS, they found that the freshwater from the east GIS has a larger impact on AMOC and climate.

In the present study, we focus more on the difference in climate impacts produced by injection of meltwater from northern and southern GIS. Recent observations indicate that the surface melting and dynamical thinning of the northern GIS, especially the northeastern region, have started to accelerate during the past decades and may continue into the future due to amplified warming in that region (Khan et al. 2014; Mouginot et al. 2015). Two other mechanisms may also favor the loss of ice from the northern GIS. One is the enhanced basal melting in the northern GIS facilitated by the higher geothermal heat flux there than that under the southern GIS. The stronger basal melting promotes basal sliding, and as a consequence faster flow of ice streams. This likely has helped maintain the fast-flowing ice streams in the north for the present day (Rogozhina et al. 2016). The other is the enhanced surface melting over the northern GIS relative to the southern GIS due to a special pattern of atmospheric blocking (e.g. Hanna et al. 2016; Tedesco et al. 2016). The blocking has become stronger during the past few decades and may continue to the future (Liu et al. 2016). Numerous modeling studies also show that the northern GIS retreated substantially (Born and Nisancioglu 2012; Quiquet et al. 2013; Stone et al. 2013) during the last interglacial period (~115 ka), which was warmer than the present and may provide a clue of how GIS might change in the future. Therefore, it is not unreasonable to expect significantly increased freshwater discharge from the northern Greenland in the future.

2. Model and experimental design

The two climate system models used in this study are CM2G and CM2M, which have the same atmosphere, land, sea ice and iceberg model components. The atmosphere model (AM2) is a 3-D general circulation model solved using a finite-volume dynamic core. The horizontal resolution is 2° latitude × 2.5° longitude, and there are 24 layers in the vertical. The land model (LM3.0) includes a multi-layer snowpack above the soil;

continuous vertical representation of soil water; finite-velocity horizontal transport of river runoff; lakes, lake ice, and lake-ice snowpacks; vegetation and associated treatment of radiative transfer, evapotranspiration. The sea ice model has full ice dynamics, three-layer thermodynamics. The albedo of snow (on ice) and ice are 0.85 and 0.65, respectively. Snow on land and land ice cannot exceed 1 m thickness, excessive snow is transported by rivers to the coastal ocean cells, where they accumulate and are released as ice bergs. The iceberg model tracks their trajactory and mass (Martin and Adcroft 2010). The ice bergs release 0 °C freshwater on their path but do not affect surface albedo and momentum of the ocean-ice-atmosphere system.

The two models differ only in their ocean components; CM2G uses an isopycnal coordinate model (GOLD) (Hallberg and Adcroft 2009), while CM2M uses a z-coordinate model (MOM4p1) (Griffies 2007). Both ocean models were run at the same nominal 1° resolution, 360×210×63 (zonal×meridional×vertical) grid cells for CM2G and 360×200×50 for CM2M; both have enhanced meridional resolution up to 1/3° at the equator and both use a bipolar grid (one pole is located in Siberia, the other in North America) above ~65° N. Both models are free surface which allows the actual water flux rather than virtual salinity flux to be applied at the surface when simulating the freshwater forcing. Moreover, both models are able to handle the large amount of freshwater flux (0.1 Sv) applied to one grid box. More details about these components can be found in Dunne et al. (2012), where the model components of ESM2G and ESM2M are described. CM2G and CM2M are the same as ESM2G and ESM2M, respectively, except that the latter further include a biogeochemical component (Dunne et al. 2013).

The base state is preindustrial (year 1860) equilibrium state for both the CM2G and CM2M, for which the AMOC is approximately 24 Sv and 26 Sv, respectively (Fig. 2). Under present-day (year 1990) external forcing, the AMOC in both models decreases to approximately 19 Sv (not shown), consistent with the observed value at 26°N by the RAPID Array over years 2004-2009 although it has decreased to 15.5±1.9 Sv during years 2009-2014 (e.g. Frajka-Williams et al. 2016). Throughout this paper, AMOC is defined as the maximum value of the zonally integrated overturning stream function over the Atlantic north of 35°N.

Six simulations were carried out with each model: one control run and five experiments with injected extra freshwater. In all simulations, the freshwater forcing was applied at one location at a time; this was done to assess the climate impact of each location individually. In each experiment, a constant freshwater flux of 0.1 Sv (1Sv = 10^6 m^3/s) (equivalent to ~87 cm of global mean sea level rise per 100 years) was either applied locally within one grid cell at one of the four locations described above (Fig. 1) or spread over the North Atlantic between 50°N and 70°N in the 'NA' experiment for 300 years. The temperature of the freshwater was assumed to be at 0 °C. The flux magnitude of 0.1 Sv, same as that used in Yu et al. (2016) and Swingedouw et al. (2013; 2015), was chosen so as to produce a climate signal that was large enough to be readily distinguished from internal variability of the climate. For comparison, currently the entire GIS is losing mass at a rate ~0.01 Sv, and under an assumption of continuous acceleration of this rate similar to the past two decades, the mass loss rate will be ~0.073 Sv at the end of 21st century (Rignot et al. 2011; Yang et al. 2016). This is comparable to magnitudes of our applied perturbations at individual locations. Therefore, our study serves as a sensitivity study rather than a prediction. All simulations are carried out for 300 years except for three experiments ('79N', 'Petermann' and 'Helheim') with CM2M, which had to be continued for an additional 200 years in order to demonstrate more clearly the evolution with time of the climate impact of the freshwater forcing.

3. Results

Despite the completely different ocean modules within the two models, the general feature of the results obtained is very similar. The AMOC weakens in all freshwater forcing experiments. It reaches quasi-equilibrium after a relatively short transient period, ~50 years for CM2G and ~70 years for CM2M (Figs. 2a and 2c). The difference between the mean values of the AMOC for the freshwater forcing experiments and that of the respective control runs over this quasi-equilibrium period (from year 50 or 70 to year 300) are shown in Figs. 2b and 2d. In both models, the freshwater forcing applied at 79N has the largest impact on AMOC, which weakens by 5.6 Sv and 8.8 Sv for CM2G and CM2M, respectively. The freshwater forcing applied at Jacobshavn has the least impact on AMOC, which weakens by 4.6 Sv and 6.1 Sv for CM2G and CM2M, respectively. The AMOC reduction in the NA experiment is slightly smaller than that in the Jacobshavn experiment.

An apparent feature that appears in simulations with both models is that the AMOC weakens more when the freshwater is discharged from the northern GIS (79N and Petermann glaciers), than when it is discharged from the southern GIS (Jacobshavn and Helheim glaciers). The results of 79N and Petermann experiments may be averaged to represent the AMOC response to northern GIS forcing, and the results of Jacobshavn and Helheim experiments averaged to represent the response to the southern GIS forcing. Then AMOC weakens by 8.4 Sv (5.5 Sv) for the northern GIS forcing, but 6.4 Sv (4.8 Sv) for the southern GIS forcing, as obtained by CM2G (CM2M); the former is 31% (15%) larger than the latter. The difference is significant at 10% level in a two-tailed Student's t-test. The annual mean data points (251 points for CM2G and 231 points for CM2M simulations) were used in doing the significance test, with the effective sample sizes determined using the BART method of Thiebaux and Zwiers (1984). One standard deviation (σ) is calculated with the same effective sample size, and is indicated in Figs. 2b and 2d (vertical bars).

Both models also show that the weakening of AMOC due to freshwater forcing from east GIS (i.e. 79N and Helheim) is more than that due to forcing from west GIS (i.e. Petermann and Jacobshavn) (Fig. 2), consistent with the results obtained by Yu et al. (2015), although the simulations were carried out for much longer here than theirs (35 years). However, the contrast in the weakening of AMOC between the east and west GIS freshwater forcing is not always significant (at the 10% level), and is therefore not analyzed further.

The annual mean sea ice area in the northern hemisphere increases in all experiments (Fig. 3). This is unsurprising, as the northward ocean heat transport is reduced due to the AMOC weakening (Fig. 4), and the northern hemisphere is cooled significantly (Fig. 5), especially the mid to high-latitude of the Atlantic Ocean sector. Moreover, the surface of the Arctic Ocean is freshened compared to the base states (Fig. 6), which promotes sea ice formation. Maximum freshening is observed in the Petermann experiment in the simulation of both models, reaching ~1 PSU averaged over the whole Arctic basin. This freshening results in the increase of the freezing point by only 0.05 °C, therefore its effect on sea ice formation should be small. However, the surface freshening causes the shallow ocean to be stratified (Fig. 6), which may reduce vertical mixing and entraining of the warm subsurface water, resulting in thicker sea ice. The change of vertical mixing manifests itself in the change of vertical temperature profile; if mixing is reduced, the subsurface temperature increases. As shown in Fig. 7, all freshwater forcing experiments for CM2M exhibit subsurface warming, but only the Petermann experiment for CM2G exhibits subsurface warming. The magnitude of warming is generally consistent with the change of salinity stratification in different experiments shown in Fig. 6. Again, the sea ice area increases significantly (at 10% level) more (30% for CM2G and 37% for CM2M) when the freshwater is discharged from the northern GIS than when it is discharged from the southern GIS.

The annual mean sea ice thickness change averaged over year 100 to 200 is presented in Fig. 8 for the experiments ‘Petermann’ and ‘Jacobshavn’. The ice thickness increases substantially over the Arctic and North Atlantic region in both simulations, but the increase is larger when the freshwater is discharged from the northern location than from the southern location. Averaged over the Arctic basin, the mean thickness increases by 0.51 m (0.31 m) and 0.33 m (0.21 m) for the 'Petermann' experiment and 'Jacobshavn' experiment, respectively, for model CM2M (CM2G). Despite the overall thickening of sea ice, it may thin slightly in the vicinity of the freshwater discharge locations (blue regions in Fig. 8). This is due to the increased sea-ice export from those regions (see arrows in Fig. 8) — a consequence of high local sea surface height and increased divergence flow due to the addition of water mass locally. The sea ice exported from these regions results in thicker ice (by approximately 0.4 m and 0.2 m for CM2M and CM2G, respectively) in the eastern part of the Arctic Ocean in the Petermann experiment. In the other two GIS experiments, i.e. '79N' and 'Helheim', sea ice thickness increases more uniformly within the Arctic Ocean (not shown).

Comparison of the mean climates above show that the northern GIS experiments have larger climate impact (expressed as AMOC weakening, sea-ice expansion and surface temperature) than the southern GIS experiments (Figs. 2 and 3), but such comparison masks the temporal evolution of the climates. Taking AMOC as an example, they reach minima somewhere between year 50 – 200 depending on the model and experiments, and then recover slightly with time. The minima for the northern GIS experiments are all smaller than those for the southern GIS experiments. However, the AMOCs of the northern GIS experiments recover faster than those of the southern experiments, so at the end of simulations, AMOCs of all perturbed experiments actually become similar. To demonstrate this quantitatively, linear trends of AMOC are calculated for the time period from the lowest point of AMOC to the end of simulation for all experiments, and shown in Fig. 9. All trends are upward and are statistically significant at the 5% level except those for the control runs (black curves in Fig 9). To see the trends clearly, the CM2M experiments had been extended for another 200 years. The trends are indeed larger for the northern GIS experiments than the southern experiments. The sea ice area in the northern GIS experiments also converges towards those in the southern GIS experiments (not shown). Therefore, in the long term (> ~300 years), the climate impact due to freshwater discharge from any location of the GIS tends to become alike.

4 Mechanism Analyses

4.1 Different climate sensitivity to freshwater forcing locations in the first 200 years

In this section we focus on the period during which the difference in AMOC and sea ice response is approximately the largest, i.e., simulation year 101 – 200, and all the variables are averaged over this period. We focus on understanding the difference between the experiments with northern and southern freshwater forcing, because it is more significant than the difference between experiments with eastern and western freshwater forcing as described above.

We focus on the variability of the maximum mixed-layer depth in the North Atlantic changes because it is directly related to the deepwater formation and AMOC change (e.g. Thomas et al. 2015). In the absence of the freshwater forcing, deepwater forms in the Labrador Sea, Nordic Seas and the sea south of Iceland (referred to as Irminger Sea hereafter for the sake of convenience) as indicated by the deep convection/mixed-layer locations in Figs. 10a and 10c). The results of the freshwater experiments show that the deep convection (as well as deepwater formation) at the Labrador Sea shuts off quickly (within 20 years) in all 5 freshwater forcing experiments (Figs. 10b and 10e), similar to the previous results obtained in many other models (Stouffer et al. 2006). The change of mixed-layer depth in the Irminger Sea is relatively small in all experiments. Therefore, we focus on the deepwater formation site in the Nordic Seas when analyzing the cause of the differential AMOC responses in different experiments.

The time-series of mean mixed-layer depth in the North Atlantic averaged over the region north of 65°N but east of Greenland (for CM2M only) are shown in Fig. 11b. They have trends similar to those of AMOC in Fig. 9, and their variations lead AMOC by 1-3 years depending on the experiments (not shown). For the CM2G simulations, whose ocean component uses isopycnal coordinates in the vertical, the mean diapycnal velocity over the deepwater formation region is a better indicator of convective activity than the mixed-layer depth. Figure 11a shows time series of mean diapycnal velocity over the North Atlantic for this model. Their trends are significant (5% level) and consistent with those of the AMOC, but the variability of the mean diapycnal velocity for the 79N experiment is large near the end of simulation (year 250-300). The mean changes of either diapycnal velocity (CM2G; Fig 11b) or MLD (CM2M; Fig. 11d) over years 101-200 are also consistent with those of AMOC (Figs. 2b and 2d).

The precondition for the occurrence of deep convection is that the density stratification is weak. In the Nordic Seas, the density stratification is primarily determined by the salinity stratification (not shown). The depth profile of salinity anomaly at the deepwater formation site in the Nordic Seas (averaged horizontally over the region enclosed by the boxes shown in Fig. 10a) show stronger stratification in the freshwater forcing experiments compared to that in the control run (Fig. 12). The results are not sensitive to the size of the box. In addition, the upper ocean of the Nordic Seas is significantly (at 5% level) fresher when freshwater is discharged from the Petermann and 79N glaciers compared to that when the freshwater discharge is from the Jacobshavn and Helheim glaciers (Fig. 12), consistent with the respective AMOC changes.

Freshwater discharged from the northern locations is carried by the East Greenland Current towards the deep convection site in the Nordic Seas directly, increasing the stratification of the upper ocean there and weakening deepwater formation. The freshwater discharged from the southern locations of GIS, however, is carried by the Greenland currents into the Labrador Sea, where it is transported southward by the Labrador Current to Newfoundland. Here it merges into the North Atlantic Current and recirculates towards those two deepwater formation sites (see Fig. 10 for the mean currents). During this circulation, the freshwater experiences substantial mixing with ambient seawater in both horizontal and vertical directions, and therefore has less influence on the deepwater formation when it finally reaches the deepwater formation regions.

The simulations done by Roche et al. (2010), not considering GIS in particular, showed that climate response is most sensitive to the freshwater discharged from regions that are geographically near and dynamically upstream to the deepwater formation sites. It is consistent with our simulation results of the initial two hundred years. The mechanism described above is consistent with their summary. However, the summary holds only for a relatively short time period (~200 years), and a not-so-straightforward explanation is required to explain the evolution of AMOC and other associated climate impact in longer term, as will be presented in the next section.

4.2 Evolution of climate impact of the northern GIS experiments

In this section we analyze why the climate impact of the freshwater forcing from the northern GIS weakens with time after reaching a maximum. For both models, the AMOC in the 'Peterman' and '79N' experiments recovers, after reaching the minimum between simulation year 100 - 200, to a level similar to that in the 'Helheim' and 'Jacobshavn' experiments at the end of simulations (Figs. 2a and 5). Fig. 13 demonstrates that this recovery of AMOC is reflected in the weakening of stratification at the deepwater formation sites. Despite the different structure of the salinity profile between the two models, the salinity stratification (in the upper 2000 m of the water column) for both two northern GIS experiments weakens significantly (to 5% level) near the end of the simulations.

The weakening of the salinity stratification is due to both the freshening of the deeper part of the ocean (from ~200 m to the bottom) and the salinization of the upper part (<200 m). The freshening of the deep ocean is due to the deep vertical mixing in these regions. Since the surface water is freshened, the deep water should be freshened with time too. The salinization of the upper ocean, however, is not straightforward to understand. The change of net precipitation can be ruled out as a cause; its change does not have a clear and consistent trend among experiments and models (not shown). Then we investigate whether the change of circulation is able to salinize the shallow ocean at the deepwater formation sites, with a focus on the Nordic Seas.

Figure 14 shows the circulation patterns of the upper 100 m of the ocean around the Nordic Seas at the beginning (first 50 years) and end (last 100 years) of the 'Petermann' experiment obtained by model CM2M. It can be seen that at the beginning of the experiment, the freshwater coming from the northern GIS has a significant velocity component (as indicated by the yellow arrow in Fig. 14a) towards the deepwater formation site (see Fig. 10d for its location), while this component is greatly diminished at the end of the simulation (Fig. 14b). The change in circulation is manifested as a cyclonic circulation in the Norwegian Sea as shown in Fig. 14c. Compared to when the freshwater forcing is just applied, the East Greenland Current is enhanced after a few hundred years; it transports freshwater from the north towards the south more quickly. As a result, the area of high salinity water expands towards the end of the simulation (compare Figs. 14b with 14a). The change of circulation is similar in the '79N' experiment and in the model CM2G (Fig. 15). The evolution of net freshwater flux entering the deepwater formation region in the Nordic Seas (Fig. 16) confirms the hypothesized scheme above; it is initially larger for the northern GIS experiments than for the southern GIS experiments, but gradually decreases to the same level. This is especially clear for the CM2M (Fig. 16b), but less so for the CM2G (Fig. 16a), probably because the freshwater flux for the control run is still drifting and oscillating with a relatively large amplitude (black curve in Fig. 16a).

The change of circulation pattern is due to the adjustment of sea surface height (SSH) (Fig. 17). Rather than leaked out into the subtropical gyre like in the southern GIS experiments as also seen in Swingedouw et al. (2013; 2015), large amount of freshwater is stored in the Arctic Ocean at the initial stage of the northern GIS experiments (e.g. Fig. 6). Moreover, the East Greenland Current always traps lots of freshwater discharged from the northern GIS (see Fig. 18 below). This will increase the SSH over the Arctic Ocean and along the East Greenland Current and change the geostrophic ocean current in the way shown in Fig. 15. Such change of SSH pattern appears very soon, ~10 to 50 years depending on the experiments (not shown). This change then reinforces itself by increasing the salinity at the cyclonic center in Fig. 10 and decreasing the SSH there. After a few hundred years (~200 years in CM2G and 300 years in CM2M), a persistent SSH anomaly pattern is formed in the Nordic Seas (Fig. 17) that prevents freshwater from being advected directly to the deepwater formation site. This way the freshwater is mixed with seawater in the south before advecting back to the northern Nordic Seas where deep convection occurs. Therefore, the magnitude of the impact on the deepwater formation becomes very similar to the southern GIS experiments.

A series of snapshots of the salinity on a vertical section along 70°N (thick black curve in Fig. 17a) are shown in Fig. 18 to further demonstrate how salinity field and circulation adjust with time. During the initial stage of the freshwater forcing (Figs. 18b and 18f), horizontal spread of freshwater towards the east is apparent. This process actually reduces the slope of the isohalines, therefore does not trigger strong adjustment of the velocity field described above according the thermal wind relationship. The freshening reaches maximum after the freshwater forcing continues for 100 to 200 years (Figs. 18c and 18g), during which downward mixing of freshwater below the East Greenland Current (at the west boundary of each panel) is seen. This is especially clear for the CM2M model (see the deepening of the isohalines in Fig. 18g), and is equivalent to accumulation of freshwater under the East Greenland Current and rise of SSH as mentioned above. This deepening of isohaline increases the slope of isohalines and the speed of East Greenland Current itself, triggering the strong adjustment of the circulation pattern which weakens the freshening of the region to the east. The salinity in the region between 10°W and 10°E then increases (also see Fig. 14b and Fig. 15 for a top-down view).

Figure 19 shows the effect of the adjustment in the northern GIS experiment compared to the southern GIS experiment, taking 'Petermann' and 'Helheim' as the respective example. Compared to the 'Helheim' experiment, more freshwater is gradually accumulated in the Nordic Seas in the 'Petermann' experiment in the first two hundred years (Fig. 19b). Then the East Greenland current in 'Petermann' experiment becomes fresher and faster with time, so that most of the freshwater is transported out of the Nordic Seas. This freshwater comes back to the Nordic Seas in a similar way as that discharged at the Helheim glacier. Near the end of the simulation, the surface salinity in the Nordic Sea is similar between the two experiments (Fig. 19c).

4.3 The 'NA' experiment

The weakening of AMOC in the 'NA' experiment is close to that of the southern GIS experiments for both models (Fig. 2) during the whole simulation. Figure 20 shows the difference in the maximum mixed-layer depth between the 'NA' experiment and the 'Helheim' experiment. The mixed-layer depth is significantly shallower in the Nordic Seas in the 'NA' experiment than in the 'Helheim' experiment, but the opposite is true in the Irminger Sea region; the influence of the two regions on deepwater formation compensate with each other so that the AMOC change is similar between the two experiments. In the 'NA' experiment, the freshwater is not only forced directly on the deepwater formation site in the Nordic Seas, but also on the whole Norwegian-Atlantic Current that is feeding high salinity seawater into the Nordic Seas (see Figs. 1b and 14a), while the freshwater in the 'Helheim' experiment has to travel a relatively long way before reaching there. Therefore, the freshwater in the 'NA' experiment has a larger impact on the deepwater formation in the Nordic Seas than that in the 'Helheim' experiment. The amount of freshwater that reaches the Irminger Sea should be larger in the 'Helheim' experiment than that in the 'NA' experiment, since not only part of the freshwater in the latter ends up in the Nordic Seas directly and destroyed by strong deep mixing there, but also the freshwater elsewhere is applied over a broad region, thus subject to more mixing before reaching the Irminger Sea.

5. Conclusion and Discussion

By using two coupled climate models, the GFDL CM2G and CM2M with distinctly different ocean components, we investigate how the climate impact of meltwater from GIS glaciers depends on the discharge locations. The results from the two models are consistent; the climate impact is different on different time scales. In relatively short timescales (200 – 400 years depending on the model), the climate impact is larger when freshwater is discharged from the northern GIS than from the southern GIS, but at longer timescales, the climate impact of freshwater from the northern GIS weakens and becomes similar to that from the southern GIS.

When averaged over simulation year 50 – 300 for the CM2G and 70 – 300 for the CM2M, the reduction of AMOCs for the southern GIS experiments are 4.8 Sv and 6.4 Sv for CM2G and CM2M, respectively, the reduction for the northern GIS experiments are 5.5 Sv and 8.4 Sv for CM2G and CM2M, respectively. The larger reduction of AMOC in the northern GIS experiments results in larger reduction of ocean heat transport by the Atlantic Ocean, more significant Arctic sea ice expansion, and cooler surface air temperature in the northern hemisphere.

The larger climate impact in the first few hundred years of the northern GIS experiments is due to two reasons: 1) the deepwater formation site at the Labrador Sea is shut off quickly in all experiments and the change of mixed-layer depth in the region south of Iceland is modest, therefore the sensitivity of AMOC to freshwater forcing is determined primarily by the density stratification in Nordic Seas. 2) The freshwater discharged from the northern GIS is advected to this deepwater formation site more directly than that from the southern GIS.

The subsequent weakening of the climate impact of the northern GIS experiments is due to the adjustment of upper ocean circulation in the Nordic Seas. Freshwater is accumulated along the East Greenland current and in the Arctic Ocean in the first few hundred years, which increases the SSH there. The change of SSH strengthens the cyclonic circulation within the Nordic Seas, and enhances the East Greenland current. The enhanced East Greenland current then advects freshwater discharged from the northern GIS away from the deepwater formation, and the salinity of the upper ocean near the deepwater formation sites in the Nordic Seas increases. The deepwater formation and AMOC thus recovers slightly and the climate impact weakens.

The adjustment of ocean circulation in a relatively long timescale in the northern GIS experiments acts as a negative feedback of the ocean system to the freshwater forcing. The strength of the feedback may be larger or smaller in other models. In order for this feedback to exist, the ocean model resolution needs to be high enough, and the freshwater should be injected into a relatively small region. Otherwise, the boundary currents would be too wide and/or the injected freshwater would be diffused to the deepwater formation region easily, the adjustment of ocean circulation would not be important. Such feedback was not observed by Roche et al. (2010) in their relatively coarse model. The feedback may be important for the last deglaciation during which freshwater discharge lasted for many thousands of years (e.g. Liu et al. 2009). However, further investigation is required to understand its importance, as the climate state during the last deglaciation was very different from that during the pre-industrial.

The response of AMOC to 0.1 Sv of freshwater forcing is modest in all experiments performed here. This may be due to a bias in freshwater export by AMOC, which exists in all CMIP5 models (Liu et al. 2017). For the pre-industrial control runs of both CM2G and CM2M, there is a net freshwater convergence into the Atlantic by the AMOC, as opposed to divergence in the observations. For models that such bias is removed, the response of AMOC may be much larger (e.g. Liu et al. 2017), and the difference in AMOC response between the northern and southern GIS experiments may also become larger during the first few hundred years.

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Figure Captions

Figure 1 Locations at which freshwater is input into the ocean. a) Observed ice speed of the Greenland ice sheet (Joughin et al. 2010a; Joughin et al. 2010b), unit is m/yr. b) the sea surface temperature at the equilibrium state of the 'NA' experiment for CM2G. The locations where the freshwater is injected in our experiments are marked with black shapes in a) and red dots in b). The region enclosed by the black curve in b) is where the water is uniformly added to the ocean in the 'NA' experiment.

Figure 2 Variation of AMOC. a) and c) are the time series of AMOC for both the control runs and the freshwater forced runs obtained by CM2G and CM2M, respectively. A 10-year running average has been applied to the time series. b) and d) are the change of mean AMOC (filled circles) w.r.t. the control runs for the last 250 years (CM2G) and 230 years (CM2M) of simulations, respectively. The vertical bars show the $\pm 1\sigma$ error of the mean.

Figure 3 Similar to Fig. 2 but here the change of annual mean sea ice area in the northern hemisphere is shown. Note that, different from Fig. 2, a) and c) show the time series of the change of sea ice area w.r.t. the control runs, rather than the absolute values of sea ice area.

Figure 4 Meridional heat transport of the Atlantic Ocean for a) CM2G and b) CM2M, averaged over year 101-200.

Figure 5 Change of annual-mean surface air temperature relative to the control run, averaged over the year 101-200. Only the values that are significant at the 10% level are plotted. Only the results from the 'Petermann' experiment (upper panels) and 'Jacobshavn' experiment (lower panels) are shown. The results from CM2G are on the left, and CM2M on the right

Figure 6 Depth profile of salinity anomaly averaged over the whole Arctic basin for a) CM2G and b) CM2M. The salinity anomaly is calculated relative to the control run, averaged over year 101-200 of the simulations.

Figure 7 Similar to Figure 6 except here shows the depth profile of potential temperature anomaly of the whole Arctic basin for a) CM2G and b) CM2M. The results are averaged over year 101-200.

Figure 8 The pattern of sea-ice thickness change (color shading) and velocity change (arrows). The changes are relative to the control run, averaged over the year 101-200, and only the values that are significant at the 10% level are plotted. The thick black curve shows the sea-ice edges (defined by annual mean sea-ice thickness of 1 cm) in the control run. The green curves show the mean September sea-ice edges for both the control run (solid) and perturbed runs (dashed). Only the results from the 'Petermann' experiment (upper panels) and 'Jacobshavn' experiment (lower panels) are shown. The results from CM2G are on the left, and CM2M on the right.

Figure 9 Variation of AMOC for the control experiment and the perturbation experiments ‘Petermann’, ‘Helheim’ and ‘79N’ obtained by a) CM2G and b) CM2M. A 10-year running average has been applied to the time series. The CM2M experiments here are extended for another 200 years compared to those shown in Figs. 2 and 3. Moreover, the linear trend lines for periods from the lowest points to the end of runs are shown for all experiments, and the slopes are indicated with numbers (unit is 10^{-3} Sv/yr). All trends are significant to the level of 5% except the control run of CM2M (the black line in b)). The ‘NA’ and ‘Jacob’ experiments are not shown for the sake of cleanness.

Figure 10 The surface ocean currents (arrows) and maximum March mixed-layer depth (color shading) averaged over year 101-200. The panels on the left are for CM2G and right for CM2M. The upper panels (a and c) are from the control runs; the lower panels (b and d) are for the ‘Helheim’ experiment. The deepwater formation sites are approximate to regions where the mixed-layer depth is deep (> a few hundred meters). The region included in the black box is used to approximate the deep-water formation site in the Nordic Seas, for which the mean salinity anomaly profile is calculated and shown in Figs. 12 and 13.

Figure 11 Time series of a) diapycnal velocity at density interface 1034.0 kg m^{-3} from CM2G and c) mixed-layer depth from CM2M, averaged over the North Atlantic region where the mixed-layer depth is larger than 400 m. b) and d) are the changes of diapyncal velocity and mixed-layer depth of the freshwater forcing experiments relative to those in the control experiments, respectively, averaged over year 101-200. The deep blue, pink and light blue curves in a) have been shifted downwards by 0.4, 0.8 and 1.2 units from their original positions, respectively, for the sake of clearness. A 10-year running average is applied to all the time series. All linear trends shown are significant to the level of 5%.

Figure 12 Depth profile of salinity anomaly at the deepwater formation site in Nordic Seas (indicated by the box in Figure 10a). The salinity anomaly is calculated relative to the control run, averaged over year 101-200 of the simulations. The results for both a) CM2G and b) CM2M are shown.

Figure 13 Similar to Fig. 12 but here the values near the end of simulations are shown. The values are averaged over year 251-300 for the CM2G model (a and b), and over year 401-500 for the CM2M model (c and d), longer time period is used for the latter because it has oscillations of larger amplitude and longer period. Note that only 3 of the 5 experiments are carried out to year 500.

Figure 14 Mean salinity and velocity of the upper 100 m of the 'Petermann' experiment at the (a) beginning and (b) end of the simulation, where a constant value of 34 PSU has been subtracted from the salinity field (filled color). The difference in velocity between (b) and (a) is shown in (c), but the salinity is kept the same as that in (b). The values are averaged over the time period indicated at the upper right corner of each panel. The results are from the CM2M.

Figure 15 Similar to Fig. 14c except that here the '79N' experiment is shown for both CM2G (a) and CM2M (b).

Figure 16 The time series of net freshwater flux entering the deepwater formation region in the Nordic Seas through the upper 100 m for both a) CM2G and b) CM2M. All the northern GIS experiments show an initial increase in freshwater flux and then a gradual decline. The quality of the results of CM2G may suffer slightly from that the control run (black curve) has not reached perfect equilibrium state.

Figure 17 Sea surface height changes (filled colors) at the end relative to the beginning of the 'Petermann' experiments for both CM2G (a) and CM2M (b). The 'end' result is taken as the average of the last 50 years and 100 years of simulations for CM2G and CM2M, respectively, while the 'beginning' is the first 20 years and 50 years for the two models, respectively. Superimposed are the velocity changes (arrows) averaged over the upper 100 m.

Figure 18 Salinity on the vertical section along the latitude 70°N as indicated by the thick black curve in Fig. 17a. The results are from the '79N' experiment, and a constant value of 34 PSU has been subtracted from the field. The upper panels (a-d) are for model CM2G and the lower ones (e-h) for model CM2M. The salinity sections before the freshwater forcing is applied (left), at the beginning (middle left), largest climate impact (middle right) and end of the simulations (right) are shown. The field is averaged over multiple years to remove the influence of strong multidecadal oscillations, and the time period over which the field is averaged is indicated on the top right of each panel.

Figure 19 The difference in salinity and velocity between 'Petermann' and 'Helheim' experiment at different stages obtained by model CM2M. The fields are averaged over the time periods indicated on the upper-right corner of each panel. The salinity and velocity are averaged over the upper 100 m of the ocean. The white area in the ocean is where the difference in salinity is not significant to the 10% level.

Figure 20 Difference in March mixed-layer depth between 'NA' and 'Helheim' experiments. Values are averaged over year 101-200. Only where the difference is significant to the 10% level is shown.