Transient Response of the MOC and Climate to Potential Melting of the Greenland Ice Sheet in the 21st Century

Aixue Hu[1] Gerald A. Meehl[1], Weiqing Han[2], and Jianjun Yin[3]

[1]*Climate and Global Dynamics Division, National Center for Atmospheric Research, Boulder, CO 80305*

[2]*Department of Atmospheric and Oceanic Sciences, University of Colorado, Boulder, CO 80309*

[3]*Center for Ocean-Atmospheric Prediction Studies (COAPS), Florida State University, Tallahassee, FL 32306*

Abstract:

The potential effects of Greenland Ice Sheet (GrIS) melting on the Atlantic meridional overturning circulation (MOC) and global climate in the 21st century are assessed using the Community Climate System Model version 3 with prescribed rates of GrIS melting. Only when GrIS melting flux is strong enough to be able to produce net freshwater gain in upper subpolar North Atlantic does the MOC weaken further in the 21st century. Otherwise this additional melting flux does not alter the MOC much relative to the simulation without this added flux. The weakened MOC doesn't make the late 21st century global climate cooler than the late 20th century, but does reduce the magnitude of the warming in the northern high latitudes by a few degrees. Moreover, the additional dynamic sea level rise due to this weakened MOC could potentially aggravate the sea level problem near the northeast North America coast.

Introduction:

Recent observational studies show that the Greenland Ice Sheet (GrIS) possibly has become unstable since the 1990s with an estimated net mass loss up to more than 200 Gt/yr in the early 21st century [Lemke et al., 2007]. Model simulations show that if atmospheric greenhouse gas concentrations are stabilized at about 1000 ppm, the GrIS could melt away in 3000 years [Ridley et al., 2005; Alley et al., 2005]. However, most of the melting would happen in the first millennium with a possible rise of the global sea level by more than 3 meters. Since Greenland is close to ocean deep convection sites associated with the oceanic meridional overturning circulation (MOC), the discharge of the melting ice sheet water could potentially stabilize the upper ocean at these sites and lead to significantly weakened deep convection there. Presently, some modelling studies indicate a dramatic weakening of the MOC in a future warmer climate after the melting is taken into account [Fichefet et al., 2003; Swingedouw et al., 2007] and some do not [e.g. Huybrechts et al, 2002; Ridley et al., 2005; Jungclaus et al., 2006; Gerdes et al., 2006; Vizcaino et al., 2008]. Thus, it is still uncertain how much the GrIS melting would affect not only the MOC, but also the future global climate.

Paleo-records show many abrupt climate change events during last glacial period [Dansgaard et al., 1993; Ditlevsen et al., 2005; Heinrich, 1988; Hemming, 2004] and the causes of these abrupt events are most likely due to the dramatic changes of the MOC associated with a pulse of melt-water flux from the ice sheets of North America [e.g., Clark et al., 2002; Rahmstorf, 2002]. Previous studies without the melting flux from the GrIS indicate that the increase of greenhouse gas levels in the atmosphere could slow down the MOC [e.g., Manabe and Stouffer, 1994; Hu et al., 2004a; Gregory et al., 2005;

Schmittner et al., 2005], but would not collapse the MOC in the 21st century [Meehl et al., 2007]. On the other hand, paleo-records and modelling studies show that most of the GrIS was melted during the last inter-glacial period when the climate was a few degrees warmer than present day [Overpeck et al., 2006; Otto-Bliesner et al., 2006].

Given the observed accelerated mass loss of the GrIS in the past decade and the divergent modelling results, it is important to further clarify what potential role, if any, the addition of the GrIS melting flux into the climate system would play in the future changes of the MOC and global climate. Here, we evaluate the potential impact of the GrIS melting on the MOC and global climate by applying a series of idealized GrIS melting scenarios in the 21st century by using a state-of-art coupled climate model – the National Center for Atmospheric Research's (NCAR) Community Climate System Model version 3 (CCSM3) at T42 horizontal resolution for the atmosphere and land components, and 1 degree for the ocean and sea ice components [Collins et al., 2006]. There are two major differences of our approach in comparison to previous studies: 1. The melting flux of the GrIS is only added to the model during summer season; 2. Three reasonably chosen melting scenarios based on observations are applied to the coupled model under the same IPCC A1B scenario forcing. This approach provides the potential to assess the upper and lower bounds of the effect of the GrIS melting on MOC and global climate.

Experiments:

Four CCSM3 experiments are analyzed. All of the experiments are forced by the IPCC SRES A1B scenario and run from year 2000 to 2099. The first experiment is the standard A1B simulation without GrIS melting (A1Bexp). In the sensitivity experiments, it is assumed that the mass loss only occurs in the southern half of the GrIS in the 21st century. Thus, the melt-water from the GrIS, which is represented as a virtual salt flux in CCSM3, is added uniformly in the surrounding seas of the southern half of Greenland (58^o - 72^{o}N and 12^o - 60^{o}W, inserted on plot in Fig. 1a) with an initial annual mean melting rate of 0.01 Sv (Sv ≡ $10^6 m^3 s^{-1}$, or 1mm/yr global sea level equivalent) which is close to the observed current rate of the GrIS mass loss [Lemke et al, 2007]. The added freshwater flux is not compensated elsewhere. This initial rate of melting increases by 1% and 3% per year compound until 2099 in the second and third experiments (hereafter 1%exp and 3%exp, respectively), and by 7% till 2050 and then kept constant for the fourth experiment (7%exp). This 7% increase per year is close to the estimated mass loss of the GrIS in the last decade or so [Rignot and Kanagaratnam, 2006]. The maximum annual mean rate of the melt flux by 2100 is 0.03, 0.19 and 0.29 Sv, and the cumulative global sea level rise by 2100 due to this GrIS melting is 15, 54, and 165 cm for 1%exp, 3%exp and 7%exp, respectively (Figure 1a, b). The melt flux is only added in the summer months from May to October. All of these experiments are branched from a single realization of the 20th century simulation.

Results:

Time evolution of a 13-year lowpass filtered MOC index defined as the maximum strength of the annual mean Atlantic meridional streamfuction below 500-meter depth for all experiments is shown in Fig. 1c. The mean MOC in the 20th century is 19.4 Sv agreeing with the observed estimation [e.g., Ganachaud and Wunsch, 2000]. In the last 20-years of the 21st century, the mean MOC weakens by 24%, 26%, 33%, and 48% in the A1Bexp, 1% exp, 3%exp, and 7%exp, respectively, relative to the 20th century. As the MOC weakens, the Atlantic northward meridional heat transport (MHT) also reduces in almost all latitudes with a similar rate of reduction as the MOC (Fig 1d, e). In comparison with A1Bexp, the additional GrIS melting flux does not further weaken the MOC and MHT in the 1%exp, but does in the 3%exp and 7%exp, indicating that a freshwater forcing up to 0.03 Sv from the GrIS would not appreciably affect the MOC in the 21st century.

A surface freshwater budget analysis in the North Atlantic Ocean between 40 and 80^{o}N shows that the surface freshwater input from precipitation, evaporation, river runoff and net ice melting in this region decreases over the course of the 21st century in the A1Bexp relative to that in the late 20th century (Figure 1f), associated mainly to a reduction of the Arctic sea ice export from 0.2 Sv in the late 20th century to only 0.08 Sv in the late 21st century. On the other hand, the oceanic freshwater convergence in this region for the upper 1000-meters of the ocean is enhanced primarily due to the increased freshwater export from the Arctic associated with the loss of Arctic sea ice cover. Overall, the freshwater budget for the upper 1000 meters of the North Atlantic between 40 and 80^{o}N still indicates a freshwater deficit over the course of the 21st century in the A1Bexp and 1%exp (Figure 1g, positive values). This suggests that the less vigorous MOC in CCSM3 A1Bexp and 1%exp is mainly caused by the greenhouse gas induced warming, agreeing with Mikolajewicz and Voss [2000] and Gregory et al. [2005]. The freshwater deficit (salt gain) increases the upper ocean density, and thus opposes the MOC weakening [Hu et al., 2004a, b]. By including the additional freshwater flux from Greenland, the surface freshwater flux is greatly increased in the 3%exp and 7%exp simulations. The freshwater budget for these two simulations shows that the further slowdown of the MOC occurs only when the upper ocean of the subpolar North Atlantic starts to gain freshwater as shown in Figure 1g. As a result, the MOC further slows down by 9 to 24% in these two cases, in addition to the weakening induced by the increase of greenhouse gases.

The effect of this weakened MOC on the global annual mean surface temperature is shown in Fig. 1h. The global climate warms up by 2.43^{o}C in the last 20-years of the 21st century relative to the last 20-years of the 20th century. The addition of the GrIS melt-water does not significantly affect the global warming in the first half of the 21st century in all cases. At the end of the 21st century, a few tenths of a degree less warming shows up only in the 7%exp. This result indicates that the possible further weakening of the MOC due to melting of the GrIS would not reverse the global warming trend induced by the increase of the atmospheric greenhouse gas levels in the 21st century, agreeing with Swingedouw et al. [2006].

Regionally, the most significant warming occurs in the Arctic region (Figure 2a, b) where the warming in the central Arctic is more than 8^{o}C by the end of the 21st century in comparison to the end of the 20th century (Figure 2b). Less warming in the eastern part of the subpolar North Atlantic is associated with the weakening of the MOC due to the increase of atmospheric greenhouse gases. By adding the effect of the GrIS melting, there is less warming in the Labrador Sea and its surrounding region (between 50 and 65^{o}N in the Atlantic) in all ice sheet melting cases and all time slices relative to the A1Bexp (Figure 2). The area of this weakened warming expands and becomes even larger as the rate of the GrIS melting increases. In contrast to the few tenths of degree less global warming, the reduced warming in the 7%exp covers most parts of the

Northern Hemisphere with a magnitude of more than 3^{o}C in the polar and sub-polar regions by the end of the 21st century relative to the A1Bexp. In the Southern Hemisphere, the greenhouse gas induced warming is slightly intensified in the 7%exp by the further weakening of the MOC, which transports less heat from the other oceans into the North Atlantic. This pattern of the regional surface temperature anomaly is consistent with earlier studies with added freshwater forcing in the subpolar North Atlantic but without the greenhouse gas effect [e.g., Stouffer et al., 2006].

The sea ice extent in both Hemispheres, defined as the sum of the area where sea ice concentration is 15% or greater, does not change much in the experiments with GrIS melting throughout the 21st century in comparison with the A1Bexp, except in the Northern Hemisphere in the 7%exp. In the 7%exp, the annual mean sea ice extent starts to stabilize in the mid of 21st century (Figure 1i). The reduction of the annual mean sea ice coverage is about 50% smaller in the 7%exp than in the A1Bexp. In late 21st century, the summer minimum sea ice extent drops from about 8 million square kilometres in the late 20th century to only slightly higher than 1 million square kilometres in A1Bexp, 1%exp and 3%exp, but more than 4 million square kilometres in the 7%exp. This more extensive sea ice cover in 7%exp compared to the other future climate experiments is associated with largely reduced warming in the northern polar-subpolar region induced by the strong GrIS melt water flux, and the associated MOC weakening.

The changes of the MOC strength not only affect the MHT, but also the dynamic sea level (DSL) – the part of sea level controlled by oceanic currents. Presently, DSL is lower in the North Atlantic relative to the North Pacific [Rio and Hernandez, 2004], caused by the dense deep water formation in the former as part of the MOC. A weaker rate of deep water formation would raise the DSL in the North Atlantic faster than the other regions of the globe [Levermann et al., 2005]. In A1Bexp, DSL rises by about 10 to 30 cm in the coastal regions of the subpolar North Atlantic and in the Arctic, and lowers by 10 to 20 cm in the mid-ocean area of the above region by the end of 21st century (Fig. 3a), agreeing with Yin et al.[2009]. This DSL change strengthens the inflow to the Nordic Seas via the Iceland-Scotland channel as well as the southward outflow via the Denmark Strait, leading to a 14% increase of the MHT at 65^{o}N in the A1Bexp relative to that in late 20th century. This agrees with Hu et al. [2004a] who shows that the increase in atmospheric greenhouse gas levels would weaken the deep convection in the Labrador Sea, but strengthen it in the Nordic Seas resulting in an increased MHT into the Nordic Seas, contributing to the melting of Arctic sea ice. On the other hand, the contrast of the DSL changes on the east and west side of the Atlantic shows that the DSL in general rises more at the east than west, indicating an overall reduced northward flow in all cases in Figure 3, consistent with a weakened MOC.

The DSL anomaly does not change much in the 1%exp compared to that in the A1Bexp (Figure 3b), but changes considerably in the 3%exp and 7%exp due to the further MOC slowdown (Figure 3c,d). This induces a further increase of the DSL on the northeast North American coast by 10 to 30 cm. In addition to the sea level rise due to thermal expansion and GrIS melting (eustatic), the total rise of the DSL along northeast North American coast could be up to 50 cm associated to the MOC slowdown if the 7%exp case is realized by the end of the 21st century in comparison to the late 20th century. Note that in CCSM3, the sea level change due to thermal expansion and eustatic effect is not included since the total water volume of the ocean is not allowed to change. Although the further MOC slowdown due to the GrIS melting could not make the global mean climate cooler than the late 20th century, the additional rise of the DSL associated with this further MOC slowdown would pose significant impacts on the coastal regions of northeast North America.

Conclusion and Discussion:

In this study, we used a state-of-art coupled climate model (CCSM3) to assess the effect of the GrIS melt-water flux on the Atlantic MOC and the global climate. The applied Greenland melting flux in our experiments varies from 0.01 Sv at the beginning of the 21st century to 0.03 Sv and up to 0.29 Sv by the end of the 21st century. None of these experiments shows a collapse of the MOC. Freshwater budget analysis indicates that MOC weakening in the A1B and 1%exp is caused by surface warming induced by the increased greenhouse gas concentrations in the atmosphere. Changes of salinity, however, work against the MOC weakening because of the upper ocean freshwater deficit. In contrast, there is a net freshwater gain in the upper subpolar North Atlantic in the 3%exp and 7%exp simulations. It is this net freshwater gain that contributes to the further slowdown of the MOC. Thus we conclude that a moderate GrIS melting flux will not alter the MOC much in the 21st century, agreeing with some previous studies as mentioned in the introduction. If this GrIS melting flux is large enough, however, the MOC could slow down more.

Although the potential further slowdown of the MOC due to GrIS melting would not actually cool the 21st century global climate in comparison to the late 20th century in the model, it could cause the global climate to be a few tenths of a degree less warm if the

Greenland melting flux is strong enough, agreeing with Swingedouw et al. [2006]. Regionally this less warming could be as much as 3^{o}C than it otherwise would be in the Arctic region. Associated with this reduced warming, the Arctic sea ice cover could potentially increase significantly in comparison to the standard A1B simulation of future climate. In addition, the rise of the DSL related to the further slowdown of the MOC, added to the sea level rise caused by greenhouse gas induced warming, which could aggravate rising sea level in the northeast coast of North America.

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Correspondence and requests for materials should be addressed to Aixue Hu (ahu@ucar.edu).

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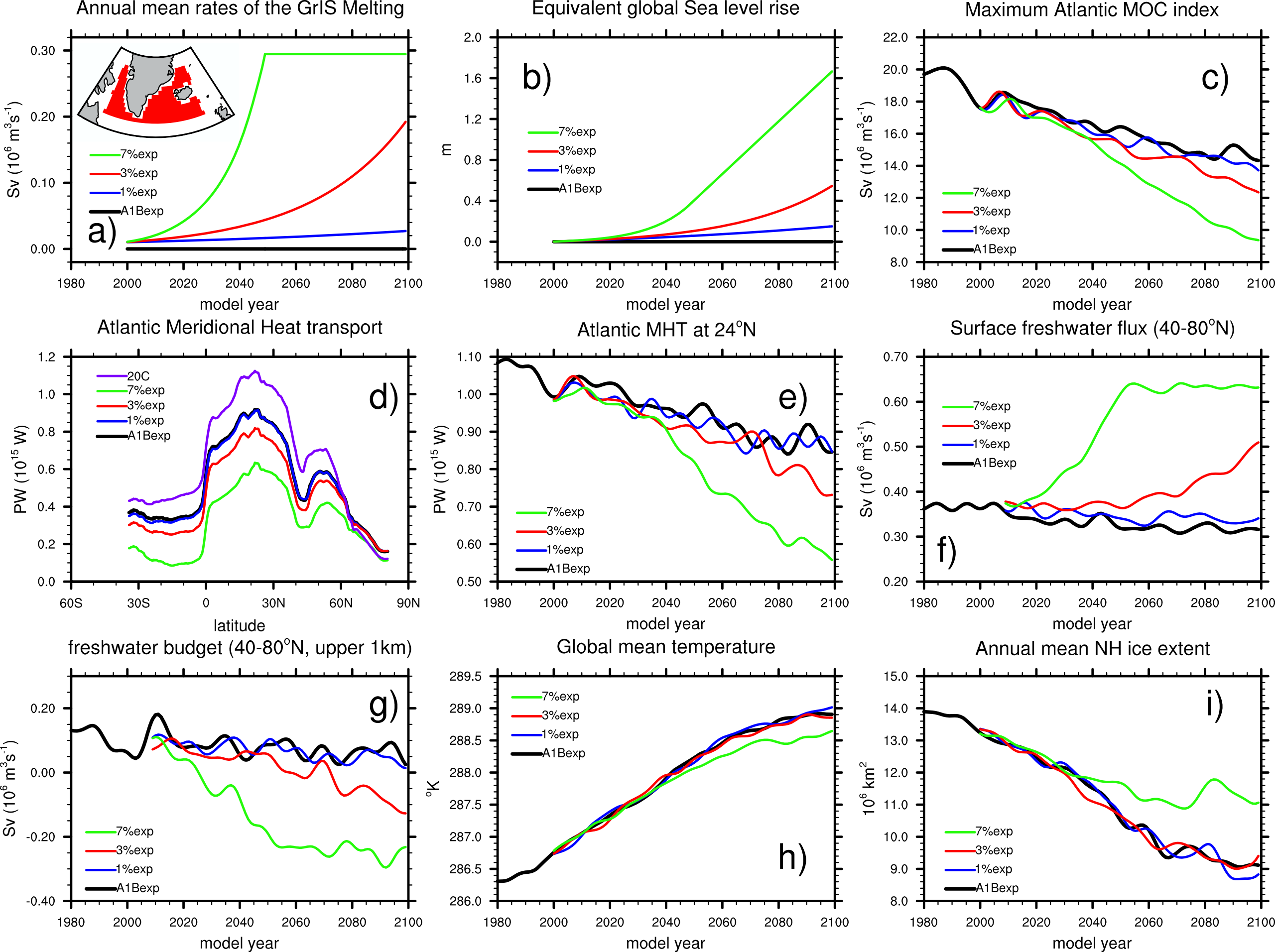
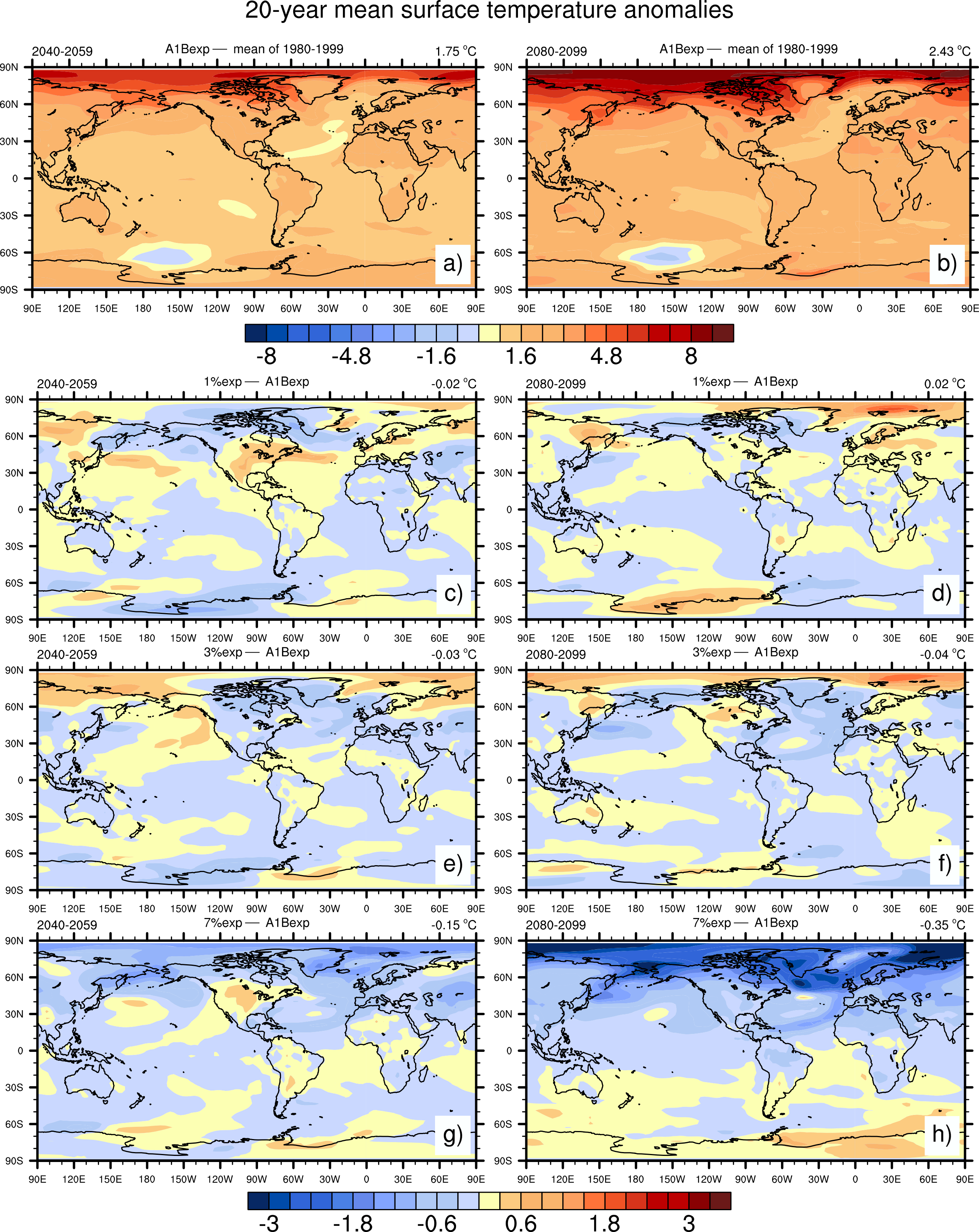
Figure legends:

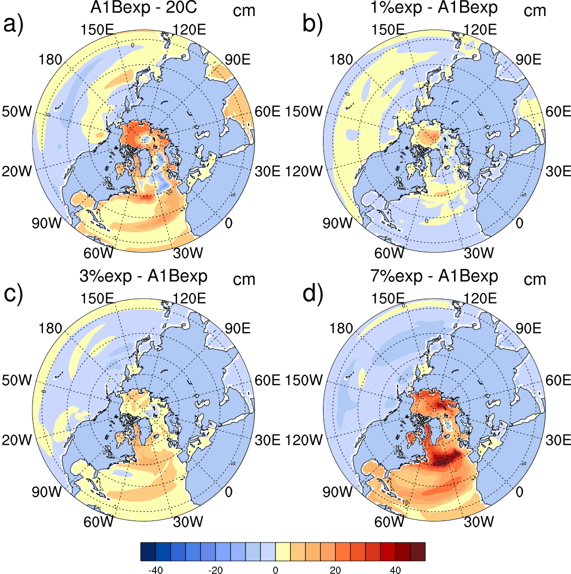
Figure 1. Time evolving annual mean rate of the GrIS melting in the experiments (a) and the equivalent cumulative global sea level rise due to this melt flux (b). Changes of the MOC index (c), the annual mean Atlantic meridional heat transport averaged over 2080-2099 for 21st century runs and averaged over 1980-1999 for the 20th century run (d) and changes of this transport at 24^{o}N (e) , changes of the surface freshwater input (f; positive represents oceanic freshwater gain) and the freshwater budget for the upper 1000 meter of the North Atlantic (40-80^{o}N, g, positive represents oceanic freshwater loss), the global annual mean surface temperature (h), and the annual mean Northern Hemisphere sea ice extent (i). The inserted figure in panel a shows the region where the Greenland melting is uniformly added into the model.

Figure 2. 20-year mean surface temperature anomalies: a) and b) are the temperature anomalies in the A1Bexp relative to the late 20th century (1980-1999); c) – h) are the temperature anomalies of the sensitivity simulations relative to the standard A1Bexp in the mid-21st century (left panels: a, c, e, g) and by the end of the 21st century (right panels: b, d, f, h).

Figure 3. Dynamic sea level anomalies: a) the mean of 2080-2099 in A1Bexp minus the mean of the last 20 year of the 20th century; b), c), and d) are the mean of 2080-2099 in 1%exp, 3%exp, and 7%exp minus the mean of 2080-2099 in A1Bexp, respectively.


Annual mean rates of the GrIS Melting
Sv (10^6 m^3s^{-1})
7%exp
3%exp
1%exp
A1Bexp
a)
model year
Equivalent global Sea level rise
m
b)
7%exp
3%exp
1%exp
A1Bexp
model year
Maximum Atlantic MOC index
Sv (10^6 m^3s^{-1})
c)
7%exp
3%exp
1%exp
A1Bexp
model year
Atlantic Meridional Heat transport
PW (10^{15} W)
20C
7%exp
3%exp
1%exp
A1Bexp
d)
latitude
Atlantic MHT at 24°N
PW (10^{15} W)
e)
7%exp
3%exp
1%exp
A1Bexp
model year
Surface freshwater flux (40-80°N)
Sv (10^6 m^3s^{-1})
7%exp
3%exp
1%exp
A1Bexp
f)
model year
freshwater budget (40-80°N, upper 1km)
Sv (10^6 m^3s^{-1})
g)
7%exp
3%exp
1%exp
A1Bexp
model year
Global mean temperature
°K
7%exp
3%exp
1%exp
A1Bexp
h)
model year
Annual mean NH ice extent
10^6 km^2
i)
7%exp
3%exp
1%exp
A1Bexp
model year


20-year mean surface temperature anomalies
2040-2059
A1Bexp — mean of 1980-1999
1.75 °C
2080-2099
A1Bexp — mean of 1980-1999
2.43 °C
a)
b)
-8
-4.8
-1.6
1.6
4.8
8
2040-2059
1%exp — A1Bexp
-0.02 °C
2080-2099
1%exp — A1Bexp
0.02 °C
c)
d)
2040-2059
3%exp — A1Bexp
-0.03 °C
2080-2099
3%exp — A1Bexp
-0.04 °C
e)
f)
2040-2059
7%exp — A1Bexp
-0.15 °C
2080-2099
7%exp — A1Bexp
-0.35 °C
g)
h)
90N
60N
30N
0
30S
60S
90S
90E
120E
150E
180
150W
120W
90W
60W
30W
0
30E
60E
90E
-3
-1.8
-0.6
0.6
1.8
3



a)
A1Bexp - 20C
cm
b)
1%exp - A1Bexp
cm
c)
3%exp - A1Bexp
cm
d)
7%exp - A1Bexp
cm
150E
120E
180
90E
150W
60E
120W
30E
90W
0
60W
30W
-40
-20
0
20
40