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Effect of the potential melting of the Greenland Ice Sheet on the Meridional Overturning Circulation and global climate in the future

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ABSTRACT

Multiple recent observations indicate an accelerated mass loss of the Greenland Ice Sheet since the mid-1990s. This increased ice sheet mass loss might be an evidence of global warming and could be related to elevated atmospheric greenhouse gas concentrations. Here, we use the National Center for Atmospheric Research Community Climate System Model version 3 to assess the potential influence of a shrinking Greenland Ice Sheet on the Atlantic Meridional Overturning Circulation (MOC), the surface climate, and sea level in the next two centuries under the IPCC A1B scenario with prescribed rates of Greenland Ice Sheet melting. Results show that a low rate of Greenland melting will not significantly alter the MOC. However a moderate or high rate of Greenland melting does make the MOC weaken further. This further weakened MOC will not make the global climate in the next two centuries cooler than in the late 20th century, but will lessen the warming, especially in the northern high latitudes. Moreover, the sea level changes due to steric effect and ocean dynamics could potentially aggravate the sea level problem near the northeast North America coast and the islands in the western Pacific region.

1. Introduction

Multiple observational evidences suggest that the Greenland Ice Sheet is losing mass since the 1990s (Lemke et al., 2007; Pritchard et al., 2009; Velicogna, 2009). The estimated annual net mass loss is more than 200 Gt in the early 21st century. The rate of this mass loss seems to be accelerating since the mid-1990s, which might be related to the greenhouse gas induced global warming (Rignot and Kanagaratnam, 2006). A few model simulations indicate that if the atmospheric greenhouse gas concentrations are stabilized at about 1000 parts per million (ppm), the whole Greenland Ice Sheet could totally melt away in about 3000 years (Ridley et al., 2005; Alley et al., 2005). Moreover, the majority of the melting would occur in the first millennium with a potential global sea level rise by more than 3 m. This means a huge amount of freshwater will flood into the North Atlantic, which potentially could significantly affect the ocean circulation there.

The Meridional Overturning Circulation (MOC, or the thermohaline circulation, THC) is a global scale ocean circulation that transports warmer and saltier upper ocean water into the North Atlantic, where this water losses its heat, becomes dense, then sinks to depth, flows southward, and upwells elsewhere in the world ocean. Many available paleo-climate records indicate a frequent occurrence of abrupt climate change events in the last glacial period, e.g., the Dansgaard/Oeschger events or Heinrich events (Dansgaard et al., 1993; Ditlevsen et al., 2005; Heinrich, 1988; Hemming, 2004). Studies suggest these abrupt events could be caused by drastic changes of the MOC due to a pulse of ice sheet related melt-water flux from North America (e.g., Clark et al., 2002; Rahmstorf, 2002). For future climates, without the inclusion of the potential Greenland melting flux, the increase of atmospheric greenhouse gas levels could induce a weakening of the MOC (e.g., Manabe and Stouffer, 1994; Hu et al., 2004; Gregory et al., 2005; Schmittner et al., 2005), but not a MOC shutdown in the 21st century (Meehl et al., 2007). In addition, paleo-observations and modeling studies both indicate a much smaller Greenland Ice Sheet in the last inter-glacial period when the climate was a few degrees warmer than the present day (Overpeck et al., 2006; Otto-Bliesner et al., 2006), suggesting a large melt water influx from Greenland into the North Atlantic occurred in the past.

The potential freshwater discharge from Greenland Ice Sheet melting could increase the upper ocean stratification in the subpolar North Atlantic, and result in significantly weakened deep convection and a more muted MOC, which could modulate the future global climate. To date, modeling studies of the effect of the Greenland Ice Sheet melting on the MOC give mixed results—some suggest a dramatic weakening of the MOC in the future warmer climate (Fichefet et al., 2003; Swingedouw et al., 2007) and some show a negligible effect (e.g., Huybrechts et al.,

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2002; Ridley et al., 2005; Jungclaus et al., 2006; Gerdes et al., 2006; Vizcaino et al., 2008). Hu et al. (2009) used a fully coupled climate model with different rates of prescribed Greenland Ice Sheet melting to indicate a lower rate of the Greenland Ice Sheet melting would not change the MOC much relative to the simulation without the inclusion of the Greenland Ice Sheet melting, but a moderate to high rate of the Greenland Ice Sheet melting does weaken the MOC further. There are two significant differences of this research in comparison to previous studies of a similar type: (1) different rates of Greenland Ice Sheet melting are proposed based on observations; and (2) the melt water is only added into the ocean during the summer half of the year.

Based on the research of Hu et al. (2009), here we expand the scope by including more detailed analysis of the effects of Greenland Ice Sheet melting on MOC, surface climate, and sea level in the next two centuries. The rest of the paper is organized as follows: Section 2 outlines the model and experiment design; Section 3 describes the MOC changes; Section 4 relates the changes of the surface climate to the MOC changes; Section 5 details the potential change of the sea level due to thermal expansion of the sea water and the ocean current induced dynamic changes; and Section 6 gives the conclusions.

2. Model and experiments

2.1. Model

The state-of-art coupled climate model used in this study is the National Center for Atmospheric Research (NCAR) Community Climate System Model version 3 (CCSM3), which has been developed by NCAR scientists, in collaboration with the Department of Energy research laboratories and university scientists (Collins et al., 2006). The atmospheric component in CCSM3 is the Community Atmospheric Model version 3 (CAM3), using spectral dynamics at T42 resolution (grid points roughly every 280 km) and 26 hybrid levels vertically. The ocean model is a version of the Parallel Ocean Program (POP) developed at Los Alamos National Lab with 1° horizontal resolution and enhanced meridional resolution (1/2°) in the equatorial tropics and the North Atlantic with 40 vertical levels. The sea ice model is the Community Sea Ice Model version 5 (CSIM5) with elastic–viscous–plastic dynamics, a subgrid-scale thickness distribution, and energy conserving thermodynamics. The land model is the Community Land Model version 3 (CLM3).

2.2. Experiment design

Four experiments are carried out using this model. All of these experiments are branched from a single realization of the 20th century simulation and are forced by the IPCC SRES A1B scenario, which is a mid-range greenhouse gas emission scenario. The CO_2 concentrations change from 368.5 ppm at year 1999 to 688.5 ppm at year 2099, and then kept constant at year 2099 levels to year 2199. The first experiment is the standard A1B simulation without the inclusion of the Greenland Ice Sheet melting (A1Bexp). This serves as a reference for our sensitivity experiments.

In the sensitivity experiments, it is assumed that the melt water from the Greenland Ice Sheet only flows into the seas surrounding the southern half of Greenland (Fig. 1A). This melt-water from the

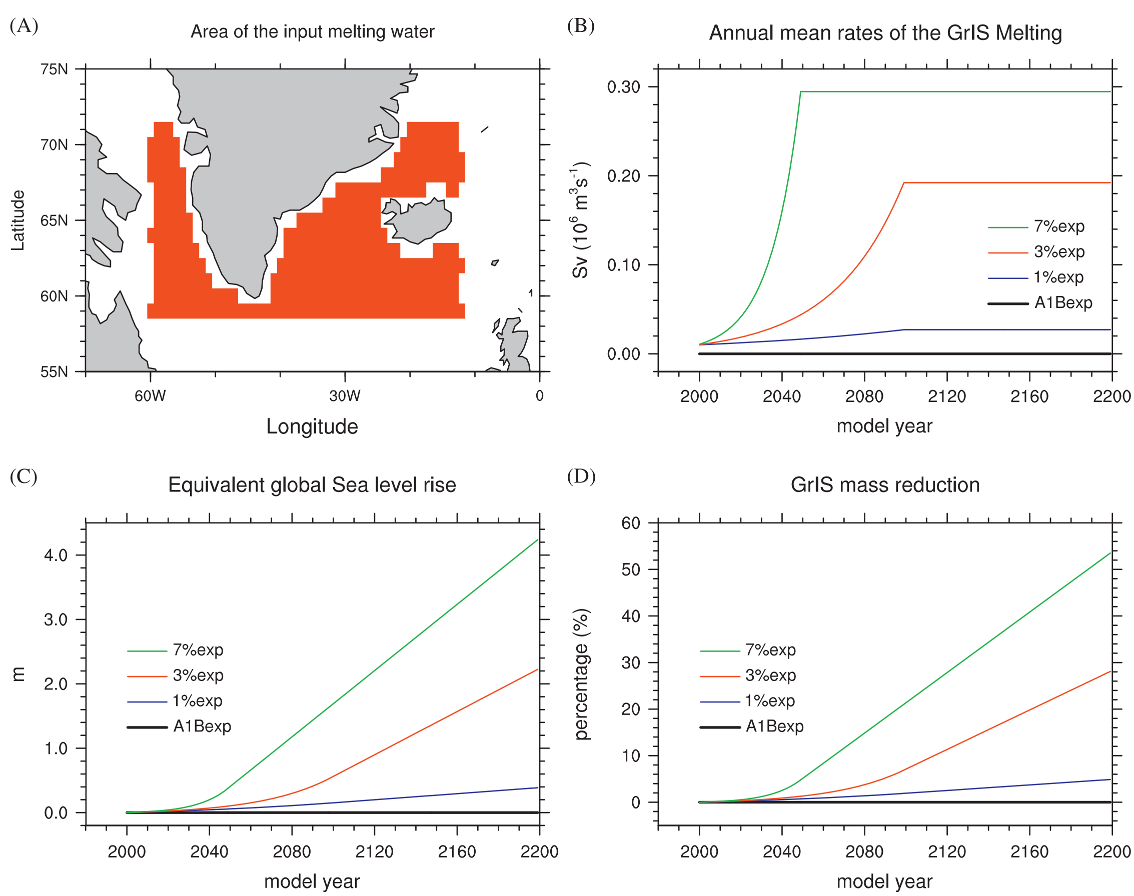

Fig. 1. Greenland Ice Sheet melt water and associated global mean sea level and Greenland Ice Sheet mass changes: (A) the region where the Greenland melt water is uniformly added into the model; (B) the assumed annual mean rate of the Greenland Ice Sheet melting; (C) the equivalent global mean sea level change due to the addition of Greenland melt water into the ocean; and (D) the percentage reduction of the Greenland Ice Sheet mass. The total mass is assumed to be 2.85 million km^3 Church et al., 2001.

Greenland Ice Sheet is added uniformly in these seas (58–72°N and 12–60°W, Fig. 1A), with an initial annual mean melting rate of 0.01 Sv (Sv ≡ 10^6 m^3 s^{-1} or 1 mm/yr global sea level equivalent), which is close to the observed current rate of the Greenland Ice Sheet mass loss (Lemke et al., 2007). This added freshwater flux is not compensated elsewhere. This initial rate of melting in our simulations increases by 1% and 3% per year compounded until 2099, then is kept at the year 2099 level until 2199 in the second and third experiments (hereafter 1%exp and 3%exp, respectively), and by 7% till 2050 and then kept constant to 2199 for the fourth experiment (7%exp). The maximum annual mean rates of the melt flux by 2100 is 0.03, 0.19, and 0.29 Sv in the 1%exp, 3%exp, and 7%exp, respectively. This 7% increase per year is close to the observationally estimated increase rate of the Greenland Ice Sheet mass loss in the last decade or so (Rignot and Kanagaratnam, 2006; Velicogna, 2009).

2.3. Sensitivity of the MOC to the timing of the freshwater forcing

Considering that the ice sheet melting mainly occurs in summer, this Greenland Ice Sheet melt flux is only added in the summer months from May to October in our simulations. To determine the effect of the freshwater input on the MOC in summer months vs. the whole period, we did two simulations with the same amount of total annual freshwater input. In one experiment, the freshwater is added year round with a rate of 0.01 Sv, and in another, the freshwater is only added in summer months from May to October with a rate of 0.02 Sv. Results show that the MOC weakens a bit more when the freshwater forcing is added year round (Fig. 2). When the freshwater is only added in summer months, part of this freshwater flux can be transported out of the regions where the deep convection occurs in winter, reducing its impact on the MOC. When the freshwater forcing is added year round, the freshwater forcing can strengthen the winter upper ocean stratification more directly, thus producing a stronger impact on the MOC.

2.4. Effects of the freshwater forcing on global mean sea level

The melt-water flux from the Greenland is represented as a virtual salt flux in CCSM3, thus the total volume of the ocean water does not change in our model simulations. Comparing the results from a freshwater volume flux version and a virtual salt flux version of the Geophysical Fluid Dynamics Laboratory's Climate Model version 2.1 (GFDL CM2.1) Yin et al. (2010) indicates that the usage of the virtual salt flux instead of the actual freshwater volume flux does not alter the model simulation much in many ways. Especially, the dynamic sea level change (i.e., the sea level change deviation from the global mean) is simulated by the virtual salt flux model with high accuracy. Since this freshwater volume flux does change the global sea level in our model, we calculated the equivalent cumulative global sea level rise due to this melt water flux. It shows that the cumulative global mean sea level rise due to the Greenland Ice Sheet melting is 15, 54, and 165 cm by year 2100, and 40, 220, and 420 cm by year 2200, for 1%exp, 3%exp, and 7%exp, respectively (Fig. 1B and C). The associated total mass reduction of the Greenland Ice Sheet is 2%, 7%, and 22% by the end of the 21st century, and 5%, 28%, and 54% by the end of the 22nd century based on the assumption that the total volume of the Greenland Ice Sheet is 2.85 million km^3 (Church et al., 2001). A further discussion about how this melt water would be distributed locally can be found in Section 5.4.

3. MOC, MHT, and freshwater budget analysis

3.1. Changes of the MOC and the associated MHT

Fig. 3 shows the time evolution of a 13-year lowpass filtered MOC index, the Atlantic meridional heat transport (MHT) at 24°N, and the 20-year mean MHT for the late 21st and 22nd centuries. The MOC index used here is defined as the maximum strength of the annual mean Atlantic meridional streamfuction below 500 m depth. The mean MOC in the 20th century is 19.4 Sv, agreeing with the observed estimates (e.g., Ganachaud and Wunsch, 2000). As the climate becomes warmer due to the increase of greenhouse gas concentrations, the mean MOC weakens to 14.7 and 13.8 Sv in the last 20 years of the 21st and 22nd centuries, respectively, in the A1Bexp. In the 1%exp, 3%exp, and 7%exp, the mean MOC weakens further and becomes 14.4 (13.4), 13.4 (9.7), and 10.1 (7) Sv, respectively, in the last 20 years of the 21st (22nd) century due to the effect of the melting water from the Greenland Ice Sheet. Our simulations indicate that a small melt-water flux (up to 0.03 Sv) from Greenland would not appreciably affect the MOC in the next two centuries; however a large melt-water flux can significantly affect the strength of the MOC.

The Atlantic meridional streamfuction shows that the North Atlantic deep water reaches a depth more than 3000 m in the late 20th century (Fig. 4A). The depth of this deep water shallows to about 2700 m by the end of the 22nd century in the A1Bexp and 1%exp when MOC weakens (Fig. 4B and C), but to 2500 m in the 7%exp when MOC weakens further (Fig. 4D). As the North Atlantic deep water shoals, the Antarctic bottom water fills its space. But the overall production of the Antarctic bottom water in the Atlantic sector is decreased. This less production of the Antarctic bottom water may be associated with the melting of the Southern Ocean sea ice, which increases the surface ocean stratification and prohibits the occurrence of the deep convection.

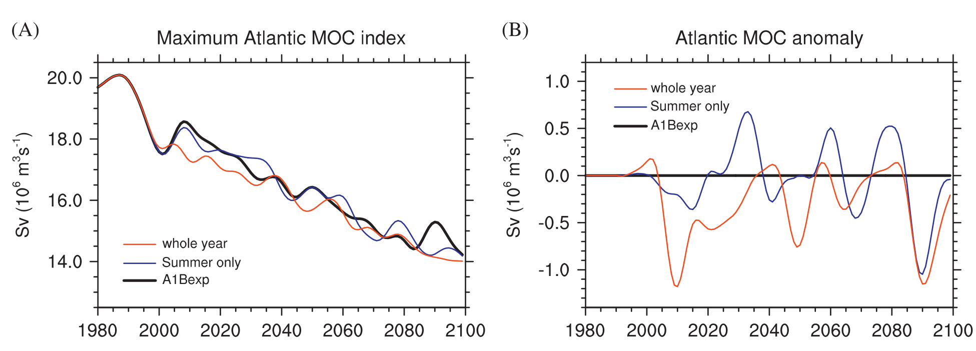

Fig. 2. MOC variations due to the addition of Greenland melt water in the summer 6 months (May–October) only vs. the whole year. (A) Time-evolving MOC index; and (B) the MOC index anomaly relative to the A1Bexp.

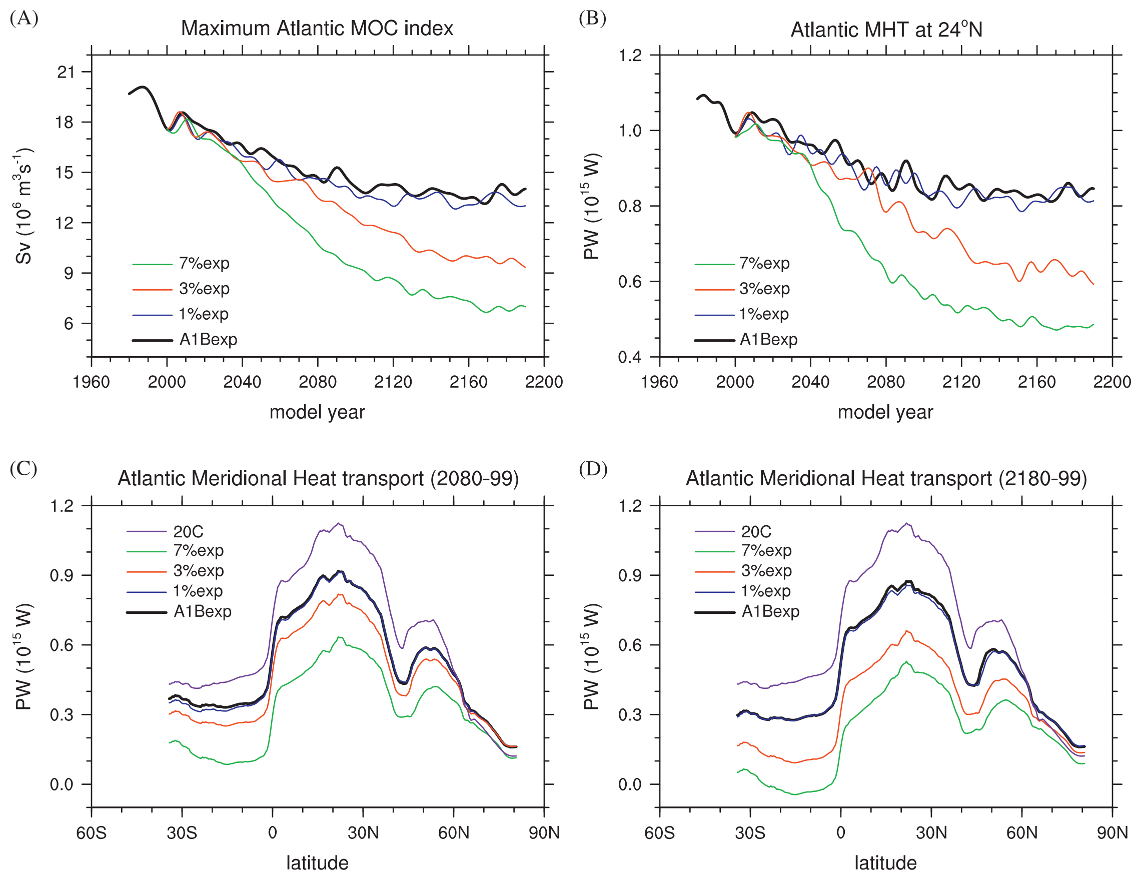

Fig. 3. Changes of the MOC and MHT in the Atlantic. (A) Time-evolving MOC index; (B) time-evolving Atlantic MHT at 24°N; (C) the annual mean Atlantic meridional heat transport averaged over 2080–2099 and 1980–1999 (20C); and (D) the annual mean Atlantic meridional heat transport averaged over 2180–2199 and 1980–1999 (20C).

As the MOC weakens, the Atlantic northward meridional heat transport (MHT) also reduces with a similar rate of reduction as the MOC (Fig. 3B). In comparison with the MHT in the late 20th century, the northward MHT reduces at almost all of the latitudes in our model simulations, except the latitudes north of 60°N (Fig. 3C and D). Only in the late 22nd century when the MOC is really weak in the 7%exp, the MHT north of the 60°N becomes less than that in the late 20th century. As indicated by Hu et al. (2004), when MOC weakens due to the elevated greenhouse gas levels, the deep convection weakens in the Labrador and Irminger Seas associated mostly with the warming, but strengthens in the Nordic Seas due to the lack of melting sea ice there; as a result, the northward MHT increases north of 60°N.

3.2. Freshwater budget analysis in the subpolar North Atlantic

To assess the reasons for these non-uniform effects of different Greenland melting rates on the MOC, a surface freshwater budget analysis in the North Atlantic Ocean between 40°N and 80°N is performed. In comparison to the late 20th century, the surface freshwater input, including precipitation, evaporation, and river runoff, reduces in the A1Bexp and 1%exp over the 21st and 22nd century, but increases significantly in the 3%exp and 7%exp due to the larger amount of the Greenland melt water input (Fig. 5A). Sea ice melting reduces similarly in this region in our simulations, except in the 7%exp, which shows signs of recovery since the late 21st century (Fig. 5B). This reduction of the sea ice melt water is resulted from the shrinkage of Arctic sea ice cover, leading to a significant reduction of sea ice export from the Arctic into the North Atlantic.

On the other hand, the upper 1000 m southward oceanic meridional freshwater transport from the Arctic into the North Atlantic increases due to the melting of the sea ice caused by the greenhouse gas induced warming in the A1Bexp and 1%exp (Fig. 5C). This southward freshwater transport increases initially in both 3%exp and 7%exp, and then starts to stabilize by the end of the 21st century in the 3%exp, but decreases after the mid-21st century in the 7%exp. At 40°N, the southward freshwater transport in the upper 1000 m Atlantic Ocean reduces slightly in all simulations (Fig. 5D). As a result, the ability of the ocean to diverge the freshwater gain in this region decreases in our simulations, except in the 7%exp, which shows an increased freshwater divergence due to the reduced freshwater export from the Arctic (Fig. 5E).

Overall, the freshwater budget for the upper 1000 m North Atlantic between 40°N and 80°N indicates a freshwater deficit (net divergence) over the course of the next two centuries in the A1Bexp and 1%exp (Fig. 5F, positive values). This freshwater deficit (salt gain) would increase the upper ocean density and destabilize the oceanic stratification in the North Atlantic if this deficit cannot be transported out, thus opposing the MOC weakening (Hu et al., 2004, 2009). Therefore, this suggests that the less vigorous MOC in the CCSM3 A1Bexp and 1%exp is not caused by the changes of the surface freshwater input and the oceanic freshwater divergence (salinity changes), but is mainly due to the greenhouse gas induced warming, agreeing with Mikolajewicz and Voss (2000) and Gregory et al. (2005).

However, in the 3%exp and 7%exp, the additional freshwater forcing from the Greenland Ice Sheet melting is strong enough to generate a freshwater gain (net convergence) in the upper 1000 m North Atlantic between 40°N and 80°N (Fig. 5F). Comparing

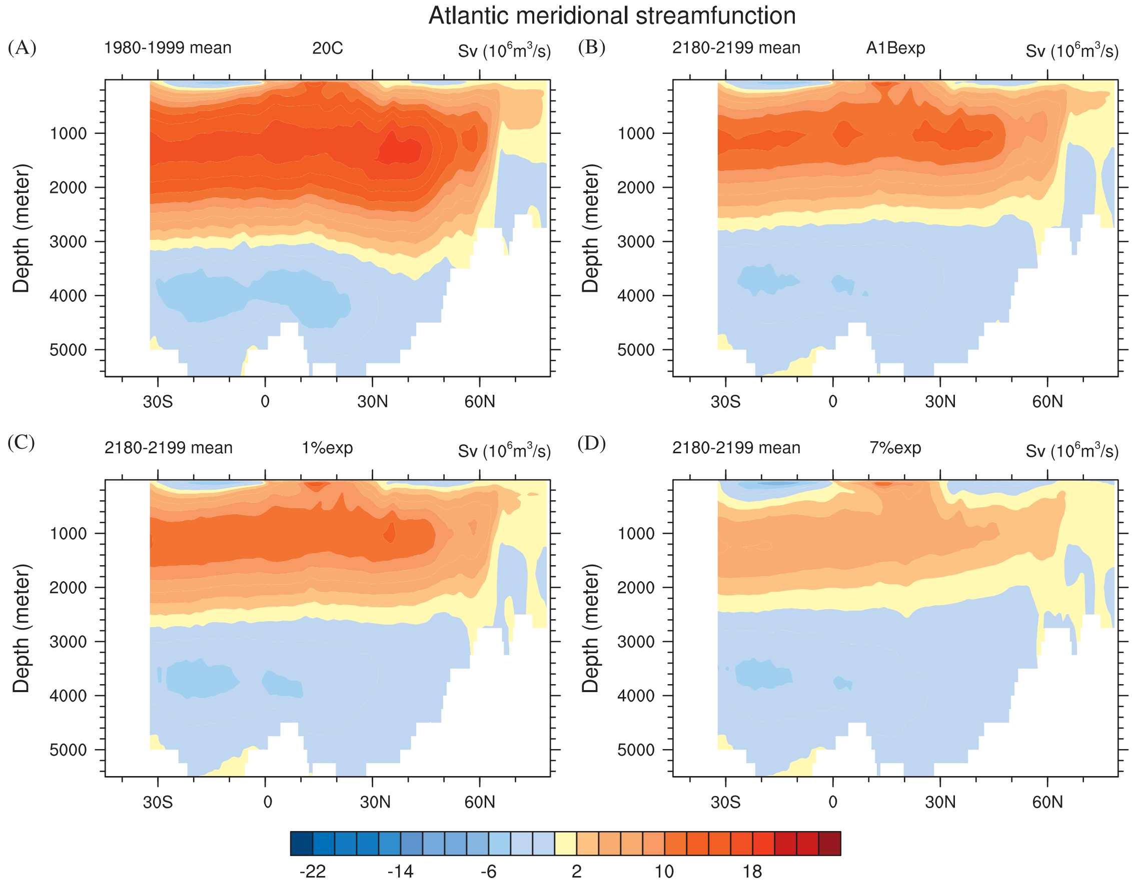

Fig. 4. 20-year mean Atlantic meridional streamfunction: (A) late 20th century; (B) late 22nd century in A1Bexp; (C) late 22nd century in 1%exp; and (D) late 22nd century in 7%exp.

with Fig. 3A, the separation of the MOC index in these two simulations from the A1Bexp occurs only after the upper 1000 m ocean freshwater budget changes from a net divergence (freshwater loss) to a net convergence (freshwater gain). Thus it is this freshwater gain (convergence) in the upper ocean that strengthens the upper ocean stratification, amplifies the weakening of the deep convection induced by the warming effect in the subpolar North Atlantic, and results in a more muted MOC.

3.3. *Contribution of the Pacific freshwater flux to MOC*

A slower MOC will reduce the transport of the fresher North Pacific water into the Arctic and North Atlantic via the Bering Strait (Hu and Meehl, 2005; Hu et al., 2007, 2008, 2010). The Bering Strait mass transport is 0.9 Sv in the late 20th century in our model, slightly higher than the observed estimation (Woodgate and Aagaard, 2005). As shown in Fig. 6 and summarized in Fig. 7, when the MOC slows down due to the increase of the greenhouse gases and the addition of Greenland Ice Sheet melt water, the Bering Strait mass transport reduces from about 0.9 Sv in the late 20th century to about 0.6 Sv in the 22nd century in the A1Bexp and 1%exp, and to about 0.2 Sv in the 3%exp, but to a negative value in the 7%exp. This negative value means that the flow through the Bering Strait reverses its direction, i.e., instead of the Pacific water flowing into the Arctic, the Arctic water flows into the Pacific when the MOC is very weak.

Corresponding to the reduced mass transport at the Bering Strait from the North Pacific into the Arctic, the freshwater transport does not change much in the A1Bexp and 1%exp. This is due to the increased precipitation in the North Pacific associated to the increased greenhouse gas levels. A small reduction of the mass transport would not change the freshwater transport much since the upper ocean water in the North Pacific is fresher than before. However, when the MOC slows down further in the 3%exp and 7%exp, the freshwater transport through Bering Strait also decreases. When the mass flow reverses its direction, the freshwater flux also changes its direction in the 7%exp. This will reduce the impact of the Greenland melt water on the North Atlantic stratification, and the MOC since part of this melt water effect is compensated by the reduction of the North Pacific freshwater inflow.

If this mass transport at the Bering Strait would not reduce, more freshwater would be available to be transported into the North Atlantic, leading to a more stratified upper ocean and a more muted MOC. Moreover, this reduced/reversed Bering Strait flow also causes a weakening of the trans-Arctic flow, resulting in a reduced water mass exchange between the Arctic and the North Atlantic, and leading to less fresher upper Arctic water being transported into the North Atlantic.

4. Effect of the weakened MOC on surface temperature and sea ice

The weakened MOC has a moderate effect on the global annual mean surface temperature (Fig. 8). The mean global climate warms up by 2.43 °C in the last 20 years of the 21st century relative to the last 20 years of the 20th century, and increases an additional 0.43 °C by the end of the 22nd century in the A1Bexp. Adding the Greenland Ice Sheet melt-water, the global mean temperature is not significantly affected in the first half of the 21st century in all cases. By the end of the 21st century, a few

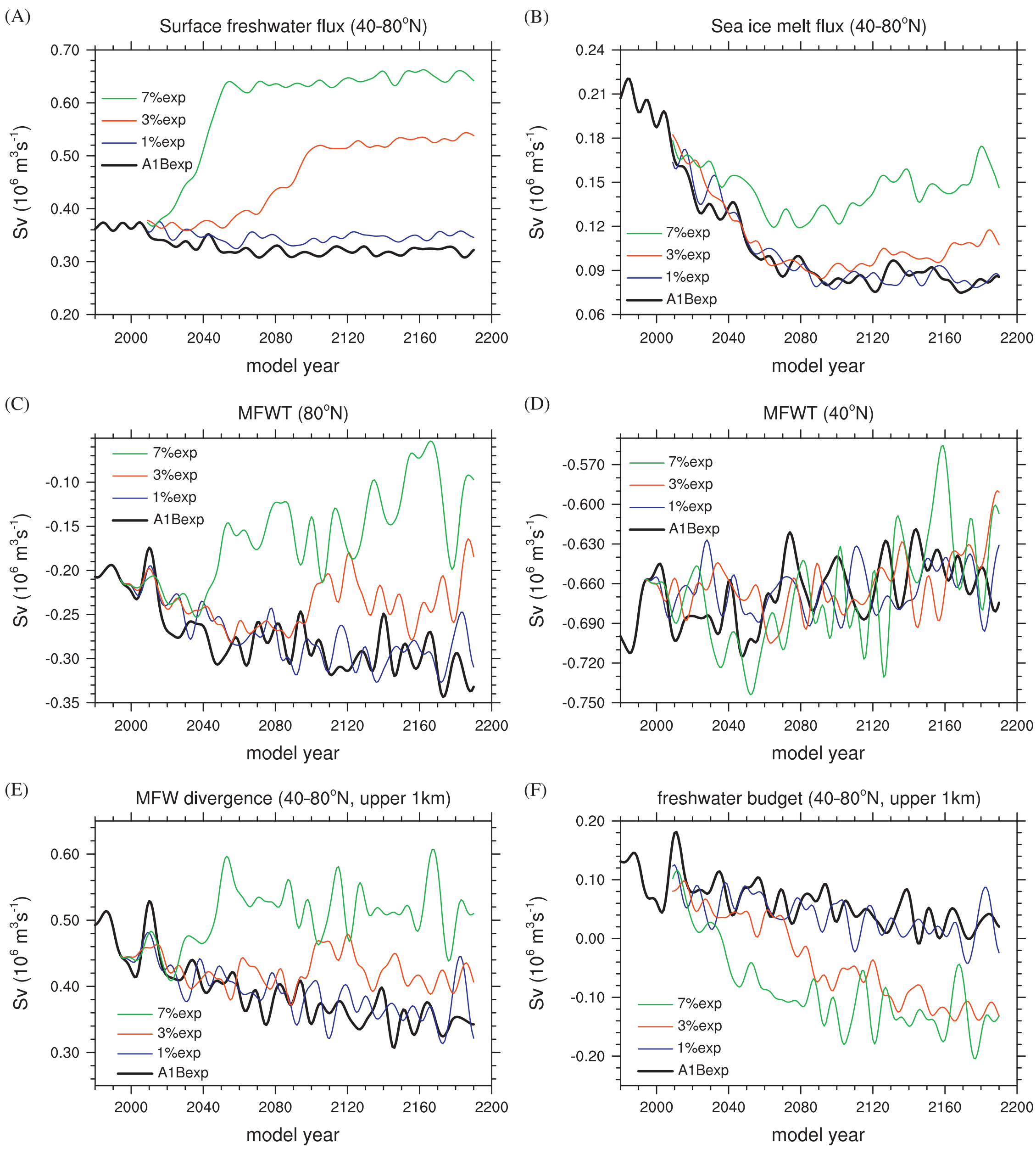

Fig. 5. Freshwater budget analysis. (A) Changes of the surface freshwater input (positive represents oceanic freshwater gain); (B) surface freshwater input due to melting of sea ice; (C) meridional freshwater transport at 80°N from the Arctic into the North Atlantic (negative indicates a southward meridional freshwater transport); (D) the same as in (C), but for 40°N in the Atlantic; (E) meridional freshwater divergence in the North Atlantic between 40°N and 80°N; and (F) the freshwater budget for the upper 1000 m of the North Atlantic (40–80°N, positive represents oceanic freshwater divergence).

tenths (0.35) of a degree less warming shows up only in the 7%exp. By the end of the 22nd century, this less warming is 0.25 °C in the 3%exp and 0.50 °C in the 7%exp. In the 1%exp, the evolution of the global mean temperature is almost the same as in the A1Bexp. All of these results indicate: (1) the MOC will possibly weaken further if the melting of the Greenland Ice Sheet intensifies; (2) the global mean climate will become warmer in the next two centuries caused by the increased atmospheric greenhouse gas level; and (3) this further weakened MOC will not be able to reverse the global warming trend in the next two centuries, agreeing with previous studies (e.g., Gregory et al., 2005; Swingedouw et al., 2006; Meehl et al., 2007).

The sea ice extent, defined as the sum of the area where sea ice concentration is 15% or greater, is decreased in all experiments (Fig. 9). The reduction of the summer sea ice extent in the Arctic is the greatest. In the Southern oceans, the reduction of the sea ice cover in both winter and summer is similar among the different simulations (Fig. 9C and D), indicating that the changes of the MOC do not have significant impacts on the Southern Ocean sea ice in the future.

The changes of sea ice cover in the Arctic are quite different from case to case. In winter, the maximum sea ice extent changes from about 18.5 million square kilometers in the late 20th century to about 14 million square kilometers in the A1Bexp and 1%exp in the 22nd century (Fig. 9B). Because of less warming due to the more muted MOC, the Arctic sea ice extent starts to stabilize around 2040 at about 17 million square kilometers in the 7%exp, and around 2100 at 15 million square kilometers in the 3%exp. In summer, the sea ice is nearly vanished in the A1Bexp, 1%exp, and 3%exp, but is stabilized at about half of the late 20th century value in the 7%exp (Fig. 9B). These results show that although the additional freshwater forcing from Greenland could

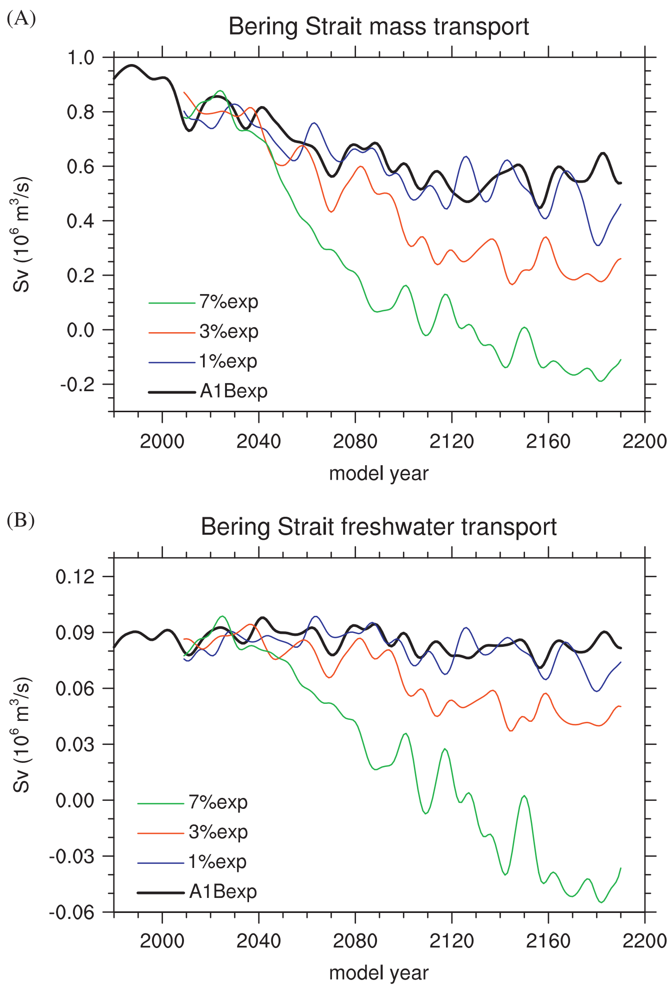

Fig. 6. Time-evolving Bering Strait mass (A) and freshwater transport (B). Positive indicates a northward transport.

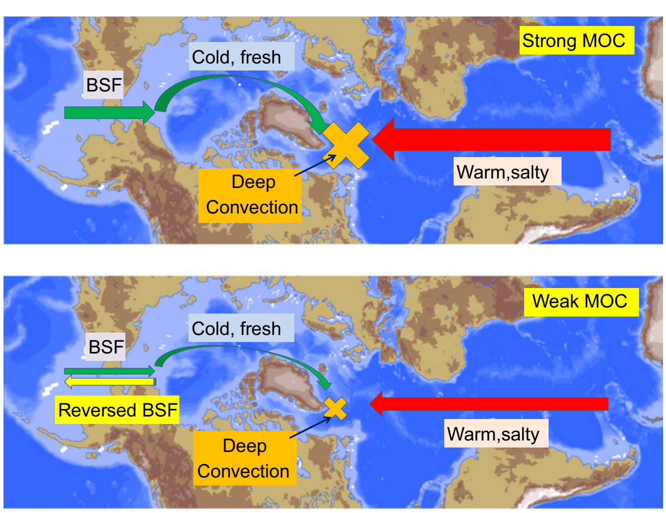

Fig. 7. Schematic diagram. Interactions between the MOC and the Bering Strait transport. When MOC is strong, more warm, salty upper ocean water flows into the North Atlantic convection sites, and the Bering Strait flow (BSF) is also stronger, which transports more fresher water into the North Atlantic. When MOC is weak, less warm, salty upper ocean water flows into the subpolar North Atlantic, and the Bering Strait flow also reduces, and even reverses direction.

weaken the MOC, the associated reduced warming in the Arctic region is still not strong enough to make the Arctic sea ice cover recover to late 20th century levels, especially in summer.

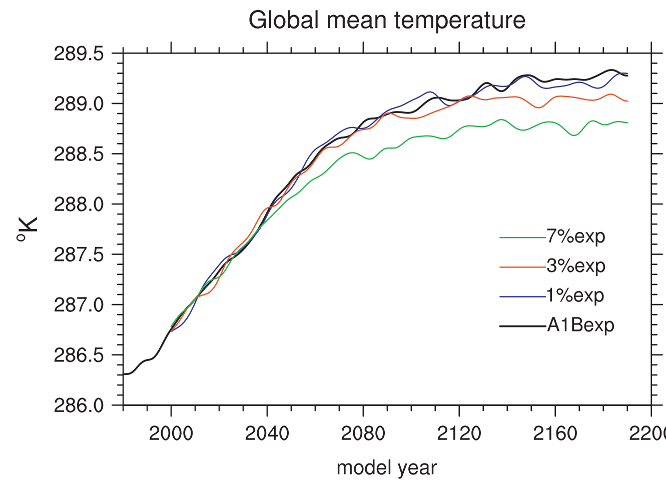

Fig. 8. The global annual mean surface temperature.

5. Changes of the sea level

5.1. Global mean steric sea level changes

The increase of greenhouse gas concentrations in the atmosphere and the melting of the Greenland Ice Sheet not only affect the surface climate, but also influence the global and regional sea level that could have a significant impact on the continental coastal region and the small islands in the middle of the ocean. A significant portion of the heat trapped in the earth system due to the elevated atmospheric greenhouse gas concentrations is absorbed by the ocean since water has a much larger heat capacity than air and land. As the ocean warms up, the total volume of the sea water increases due to the changes of the sea water density, which is called the steric sea level change. This steric sea level change includes the sea water density change due to the increase of the water temperature—the sea water's thermal expansion (thermosteric sea level rise), and due to the changes of the sea water salinity associated to the changes of the precipitation, evaporation, river runoff, and the growth/melt of land-based ice sheets or ice caps (halosteric sea level rise).

As shown in Fig. 10, the total steric sea level change in the A1Bexp is mainly due to thermosteric effect (the thermal expansion). The halosteric contribution (sea water salinity change) to the steric sea level rise is only a few tenths of a centimeter in the 21st and 22nd centuries (Fig. 10C), which is associated mostly to the melting of the sea ice and may also be associated to the salt redistribution in the ocean. The global mean steric sea level rise in A1Bexp due to thermal expansion of sea water is about 25 cm by the end of the 21st century, and 46 cm by the end of the 22nd century relative to that at year 2000 (Fig. 10B), comparable to the thermosteric sea level changes in the T85 version of the CCSM3 (Meehl et al., 2005, 2006). The thermosteric sea level change in the sensitivity experiments is similar to that in the A1Bexp during most of the 21st century. The halosteric contribution to total steric sea level change becomes more significant in the 3%exp (7%exp) since the late (mid-) 21st century, and reaches about 6 (11) cm by the end of the 22nd century (Fig. 10C).

Interestingly, the thermosteric sea level rise is about 2 cm higher in 7%exp in comparison to the A1Bexp by the end of the 21st century, and 6 cm higher by the end of the 22nd century (Fig. 10B), although the global mean surface temperature increase is smaller in the 7%exp than in the A1Bexp (Fig. 8). Fig. 11 shows the zonal mean temperature difference between the 7%exp and the A1Bexp in the Pacific basin and the Atlantic basin at the end of the 22nd century. It is clear from this figure that although the

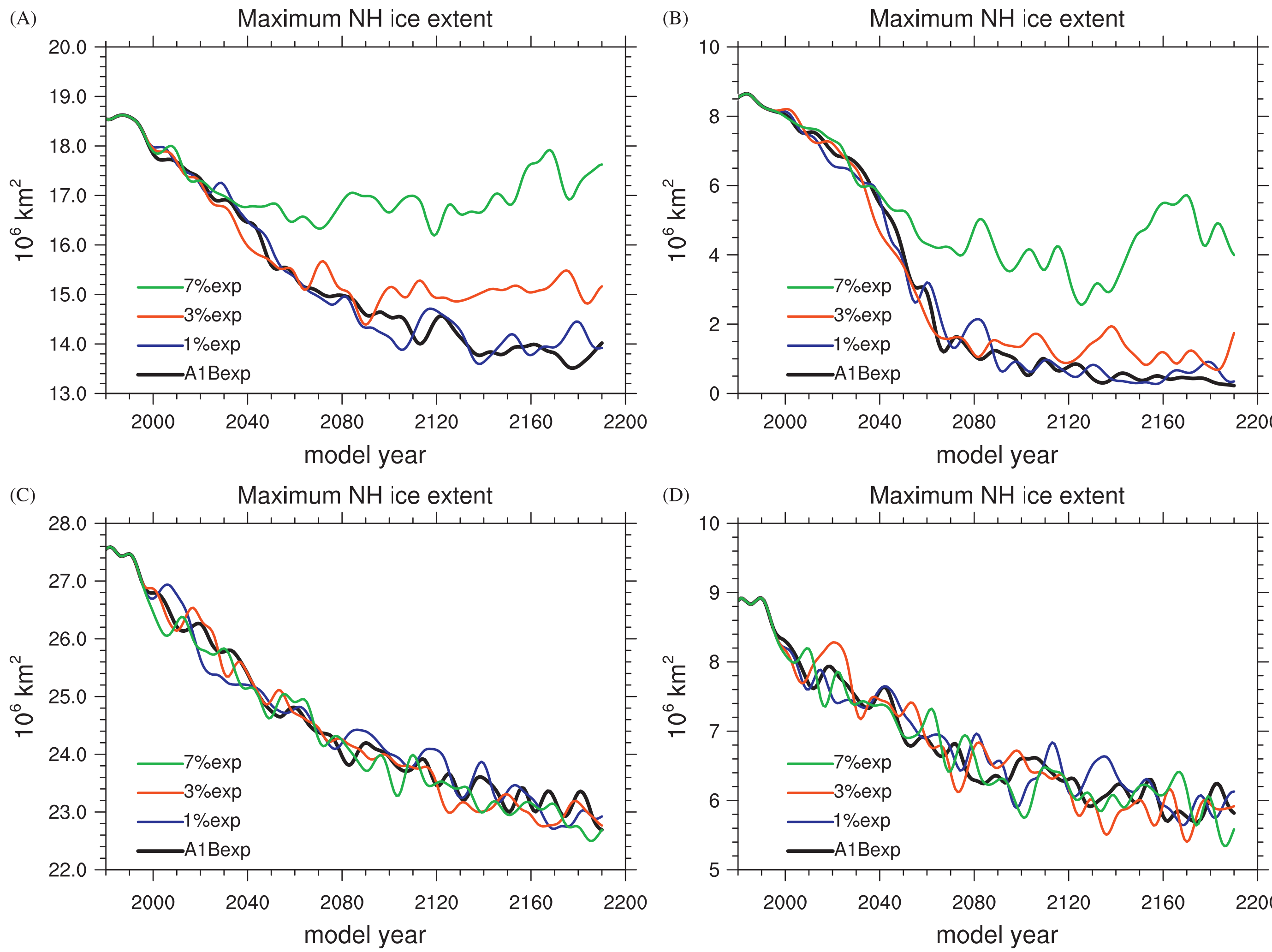

Fig. 9. Time-evolving winter maximum and summer minimum sea ice extent in the northern hemisphere (panels A and B, respectively) and in the southern hemisphere (panels C and D, respectively).

surface warming is less in the 7%exp than in the A1Bexp, the warming in the subsurface ocean is much larger in the 7%exp, which leads to the larger sea level rise due to thermal expansion in the Pacific basin. This larger subsurface warming in the Pacific is due to two reasons. First, as the MOC slows down, the freshwater transport from the North Pacific into the Arctic is reduced, and could even reverse its direction, resulting in a freshwater gain in the North Pacific region, a more strongly stratified upper ocean, more restricted convection, and reduced winter heat loss, leading to warming of the subsurface ocean. Second, the weakened MOC also pulls less heat from the Pacific into the Atlantic, resulting in a reduced southward meridional heat transport at 30°S in the South Pacific and an increased northward MHT from the tropics to the North Pacific.

In the Atlantic, relative to the A1Bexp, significant warming in the 7%exp appears in the region south of the 20°N to a depth more than 1500 m, caused also by the slowing MOC that transports less heat into subpolar North Atlantic. This warm anomaly corresponds to a more thermal expansion-induced sea level rise in these regions. The warm anomaly around 45°N is associated with the Gulf Stream. As the northward MHT reduces due to a slower MOC, more heat is trapped in the tropical Atlantic, resulting in a warmer Gulf Stream carrying more heat northward. This warmer Gulf Stream leads to more sea level rise in its path, such as along the east coast of the United States. Cold anomalies appear in most parts of the North Atlantic north of 20°N. The cold anomaly in the subpolar North Atlantic results directly from the further weakened MOC that transports less heat there. This cold anomaly is transported downward and southward with the lower limb of the MOC. In the subtropical North Atlantic, the cold anomaly is related to the transport of colder subpolar water into the subtropical ocean by the wind driven gyres. These cold anomalies produce less thermal-induced sea level rise in these regions in the 7%exp than in the A1Bexp. Overall, the subsurface water in the 7%exp warms more than in the A1Bexp, resulting in more thermal induced steric sea level rise.

5.2. *Regional dynamic sea level changes*

Regionally, the sea level is not only affected by the thermal expansion of sea water and the water salinity changes, but also by the ocean dynamics—the part of sea level change controlled by oceanic currents. The changes of the ocean currents, such as the MOC or wind driven gyres, can push the water to pile up in difference places, resulting in local sea level changes. But for a global average, the dynamic sea level does not change the total water volume or the global mean sea level. Because of the dense deep water formation in the North Atlantic as part of the MOC, the present dynamic sea level is lower in the North Atlantic

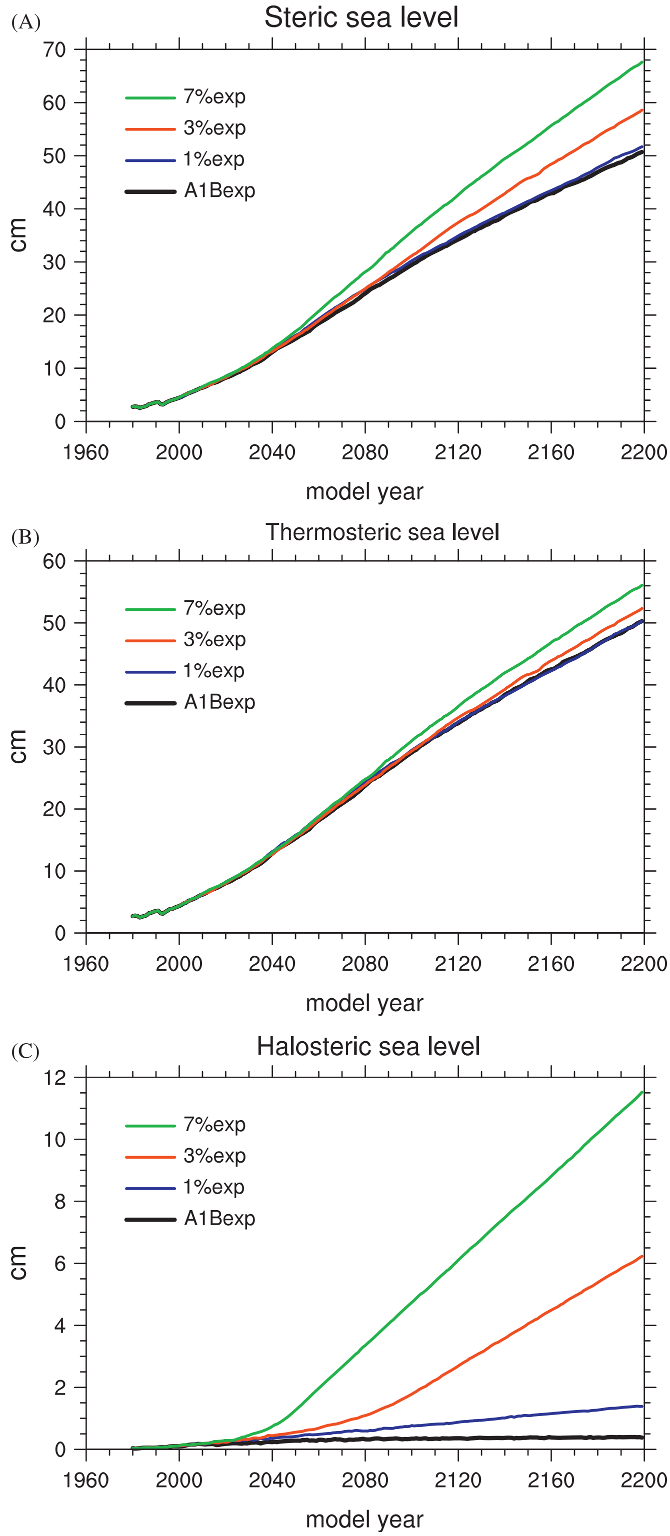

Fig. 10. Global mean steric sea level change (A), and the thermosteric (B) and halosteric (C) sea level changes.

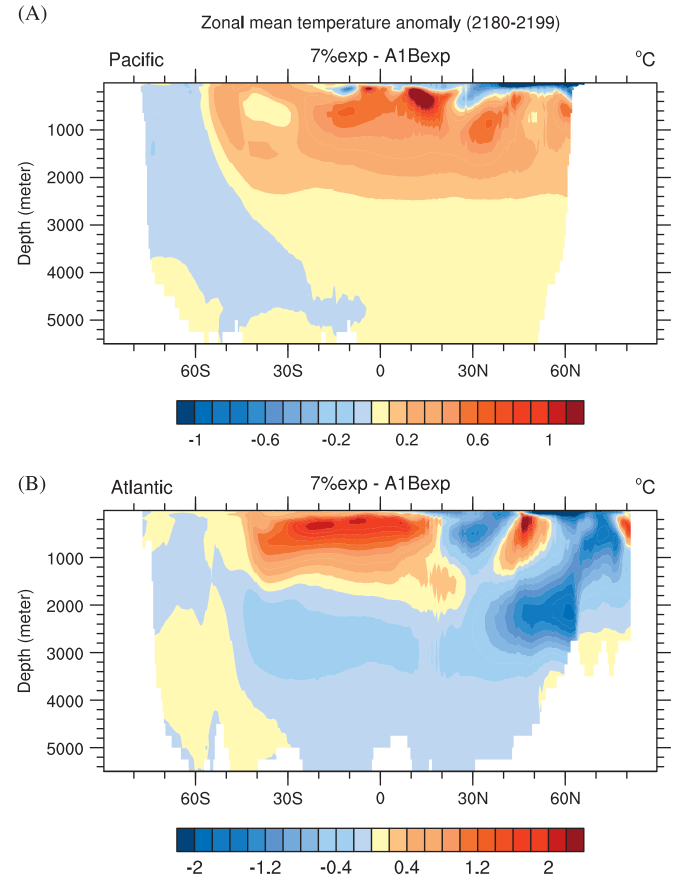

Fig. 11. Zonal mean oceanic temperature anomaly between the 7%exp and the A1Bexp at the end of the 22nd century in the Pacific (A) and Atlantic (B) basins.

relative to the North Pacific (Rio and Hernandez, 2004). In general, as the deep water formation weakens in the North Atlantic, the dynamic sea level rises faster in the North Atlantic than the other ocean regions of the globe (Levermann et al., 2005). Fig. 12 clearly shows that as the MOC slows down in the next two centuries, the dynamic sea level rises in the Atlantic and Arctic basins and lowers in the Pacific basin. As indicated by Hu et al. (2009), the dynamic sea level rises by about 10–30 cm in the coastal regions of the subpolar North Atlantic and in the Arctic, and lowers by 10–20 cm in the mid-ocean areas of those regions in the A1Bexp by the end of 21st and 22nd centuries (Fig. 12A and B), agreeing with Yin et al.(2009). There is also a general reduction of the dynamic sea level in the Southern Oceans and the eastern side of the Pacific, and a general rise of the dynamic sea level in the western parts of the Pacific and most of the Indian Ocean. The pattern of the dynamic sea level change in the North Atlantic strengthens the inflow to the Nordic Seas via the Iceland–Scotland channel as well as the southward outflow via the Denmark Strait, leading to an increase of the MHT at 65°N in the A1Bexp relative to that in the late 20th century (Fig. 3C and D), contributing to the melting of Arctic sea ice (Hu et al., 2004).

The patterns of the dynamic sea level changes are very similar in the 1%exp and the A1Bexp relative to the late 20th century (Fig. 12), since the changes of the MOC in these two simulations are almost the same. However, the dynamic sea level changes are considerably larger in the 3%exp and 7%exp associated with the further MOC slowdown (Fig. 12E–H). The dynamic sea level rise on the northeast North American coast is about 10–30 cm by the end of the 21st century, and about 40–50 cm by the end of the 22nd century in the 7%exp.

5.3. Regional dynamic and steric sea level changes

In reality, the sea level change due to steric and dynamic effects is tightly coupled together (Lowe and Gregory, 2006). Because of the current configuration of the CCSM ocean component, Boussinesq approximation is used. As a result, the model conserves the total water volume, which prevents us to include

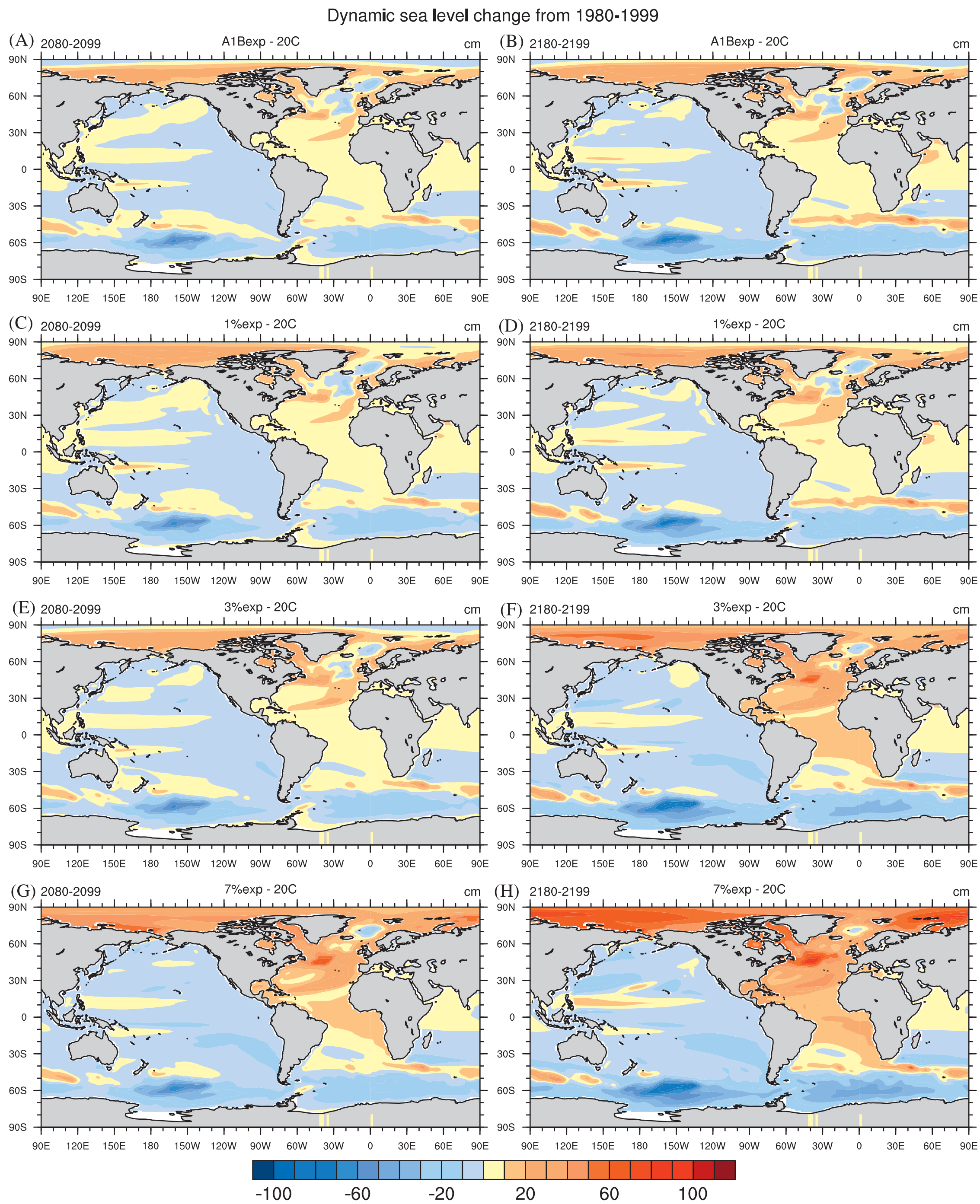

Fig. 12. Dynamic sea level anomalies in the late 21st (left panels: A, C, E, G) and 22nd (right panels: B, D, F, H) centuries relative to that in the late 20th century. Panels A and B are for the A1Bexp, C and D for the 1%exp, E and F for the 3%exp, and G and H for the 7%exp.

the steric effect on sea level while the model is running. An easy remedy to this is proposed by Greatbatch (1994), who suggest that the steric effect on sea level can be included by adding a globally uniformed, time-evolving steric sea level change to the model produced dynamic sea level change on a time scale longer than a month. Fig. 13 shows the combined global steric and dynamic sea level changes. In general, the sea level rises more in the Atlantic basin than in the Pacific basin in all experiments, resulting in mostly a less vigorous MOC in these simulations than that in the late 20th century. The sea level rise could be significantly more along the east coast of the North America if a high rate of Greenland melting is realized in the future, e.g., the 7%exp. This sea level rise due to steric and dynamic effect could reach above 1 m along the east coasts of North and South America, the west coast of Africa, and the coastal regions of the Arctic by the end of the 22nd century (Fig. 13H).

In comparison with the A1Bexp, the further slowdown of the MOC and the freshening of the ocean due to the melting of the Greenland Ice Sheet could cause an additional sea level rise of about 20–30 cm along the east coast regions of North America by the end of the 21st century, and 40–50 cm by the end of the 22nd century in the 7%exp (Fig. 14). Additional sea level rise of 20–40 cm can also be seen in the western coast of Africa, and the Philippine and Indonesian islands in the 7%exp, which could threaten the coastal infrastructure and ecosystems there.

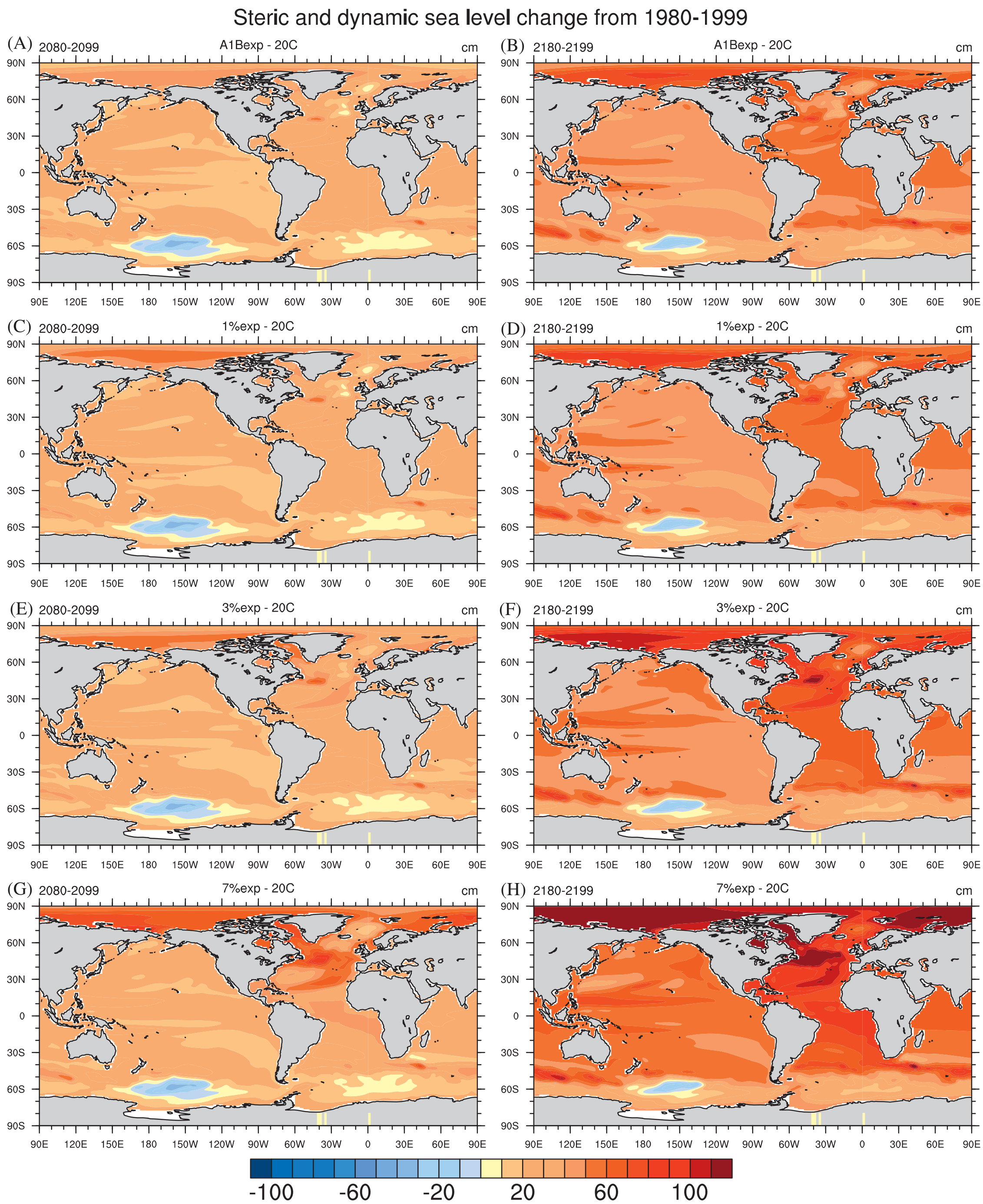

Fig. 13. Sea level anomalies due to steric and dynamic effects in the late 21st (left panels: A, C, E, G) and 22nd (right panels: B, D, F, H) centuries relative to that in the late 20th century. Panels A and B are for the A1Bexp, C and D for the 1%exp, E and F for the 3%exp, and G and H for the 7%exp.

5.4. Effect of the melt water volume on regional sea level

One additional factor that can significantly affect the sea level is the freshwater volume flux from the melting of the Greenland Ice Sheet. This effect is not included in our model simulation since our model formulation requires the model to conserve the total water volume in the ocean. The sea level change due to the addition/subtraction of water mass from the ocean, such as the growth or decay of the land-based ice, is called eustatic sea level change. When the melting water flows into the ocean, it will increase the total volume of the sea water even if there are no sea water temperature changes. Currently, our model cannot accurately distribute this melt water globally.

On the contrary to the more commonly stated that the melt water from land based-ice sheet will be distributed evenly in the ocean, as indicated by Mitrovica et al. (2009), due to the isostatic deformation of the solid earth and the gravitational and rotational changes caused by the land–sea water mass exchanges, a collapse of West Antarctic Ice Sheet would raise the sea level more in the far field and lower the sea level in the surrounding region of the West Antarctic Peninsula. A melting of the Greenland Ice Sheet may also cause more eustatic sea level rise in the far field, and less eustatic sea level rise (or even a eustatic sea level drop) near the Greenland (Milne et al., 2009). A more recent work of Kopp et al. (2010) indicates that when Greenland ice mass loss exceeds about 20 cm equivalent sea level changes, the regional total sea level

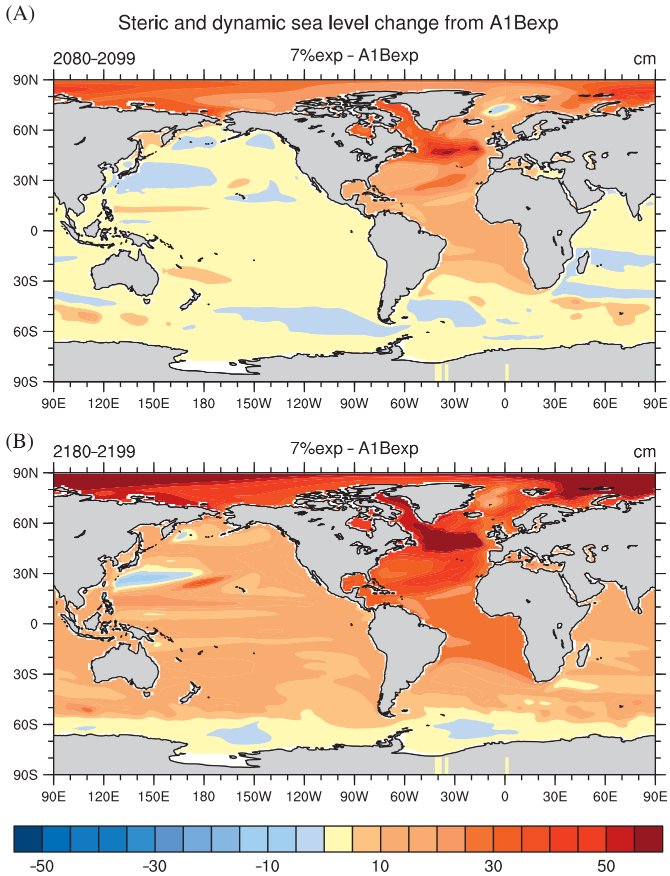

Fig. 14. Sea level difference between 7%exp and A1Bexp in the late 21st (A) and late 22nd (B) centuries due to both steric and dynamic effects.

changes will be dominated by eustatic sea level changes, especially in the regions close to the Greenland.

Moreover, a recent satellite observation (Velicogna, 2009) suggests that the rate of the Greenland and the Antarctic ice sheet mass loss increased significantly in the last 10 years and the amount of the mass loss in these two ice sheets is similar in recent years. Considering the eustatic effect from both ice sheets, there will be, to some extent, a compensation effect on the local sea level change in regions near Greenland and Antarctic. The melting of the Greenland Ice Sheet would rise the local sea level more in regions outside of the North Atlantic (Milne et al., 2009; Kopp et al., 2010) and the melting of the Antarctic ice sheet would rise sea level more in regions outside of the southern oceans (Mitrovica et al., 2009; Milne et al., 2009). Thus, we suspect that for the near future the steric and dynamic sea level change may be still the dominate factor controlling the local deviation of sea level rise in the Greenland and North American coast regions in all our model simulations. After 2050, the eustatic sea level change may become more dominant due to a cumulative large mass loss from the Greenland (and Antarctic) ice sheet(s). Therefore, the regional total sea level change – thermosteric, dynamic, and eustatic – may not be the same as discussed previously in this paper after the mid of 21st century, particularly in the Greenland and North America east coastal regions, but this is beyond the scope of this paper.

6. Conclusion and discussion

Using a state-of-art coupled climate model (CCSM3), we studied the effect of a potentially destabilized Greenland Ice Sheet on the Atlantic MOC, surface climate, and sea level in the next two centuries. Different rates of the potential melting of the Greenland Ice Sheet are tested. Results show that a low rate of Greenland Ice Sheet melting will not change the MOC and the associated climate much in comparison to the simulation without the inclusion of the Greenland melting in the next two centuries. This is because the MOC weakening in the future warmer world is mainly caused by warming of the surface ocean, which strengthens the upper ocean stratification and weakens deep convection. A freshwater budget analysis shows that the MOC would weaken further only if a net freshwater gain in the upper subpolar North Atlantic is maintained. In the A1Bexp and 1%exp, it appears a net freshwater deficit is maintained in the upper subpolar North Atlantic throughout the simulations, thus working against the slowdown of the MOC due to greenhouse gas induced surface warming. However, in the simulations with a moderate to high rate of Greenland melting, such as the 3%exp and 7%exp, a net freshwater gain is maintained in the upper subpolar North Atlantic, which leads to a further weakened MOC.

This further weakened MOC can lessen global mean warming by a few tenths of a degree in the moderate to high rates of the Greenland Ice Sheet melting simulations (3%exp and 7%exp) in comparison to the simulation without the Greenland melting. Regionally, the lessened warming mainly appears in the northern high latitudes, especially in the Arctic and the subpolar North Atlantic. This pattern of surface temperature change does not affect the Southern Ocean sea ice much, but does affect the Arctic sea ice. The Arctic sea ice extent shows a sign of stabilization in the 3%exp or 7%exp in the 22nd century or mid-21st century, respectively, but is still below the extent of late 20th century. Therefore, even if the MOC significantly weakens, e.g., in the 7%exp, the cooling effect associated with this more muted MOC is still not strong enough to overcome the greenhouse gas induced warming effect in the next two centuries, agreeing with Swingedouw et al.(2006). In other words, a drastic weakening of the MOC in the future will not cause a global cooling effect in the next few centuries in comparison to the climate in the late 20th century.

Local sea levels can be affected by sea water density changes (steric), the ocean dynamics associated to the change of the ocean circulation, and the addition/subtraction of the water to/from the ocean (eustatic). In this paper, we provided a more detailed discussion of the steric and dynamic sea level changes, and only briefly touched on the eustatic sea level change due to the addition of Greenland melt water into the ocean. Because of the elevated greenhouse gas levels in our simulations, steric sea level is increased everywhere, which primarily produced the thermal expansion of the sea water (thermosteric effect). The dynamic sea level shows increase in the Atlantic and Arctic, and decrease in the Pacific, which is associated to the weakened MOC. combining both global steric and dynamic effects, the sea level rise is more significant along the east coast of North America, and in the islands of the Philippines and Indonesia, which could have significant impacts on those regions. Considering the eustetic sea level change due to the addition of Greenland melt water into the ocean, the sea level near Greenland may not rise much (or may even drop a bit), but the sea level may rise more in the far field, and this needs further study.

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