Effect of ice sheet interactions in anthropogenic climate change simulations

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[1] We investigate the effect of ice sheets on climate change under elevated atmospheric CO_2 concentrations with an atmosphere ocean general circulation model (AOGCM) coupled to a thermomechanical ice sheet model and a vegetation model. The effect of increased meltwater fluxes from ice sheets turned out to be negligible in the phase of initial weakening of the North Atlantic meridional overturning circulation (AMOC), and more important during the recovery in subsequent centuries. Lower surface height of the Greenland ice sheet (GRIS) leads locally to a warming, especially in winter, and remotely to a cooling over northern Eurasia due to modified atmospheric circulation. With quadrupling of the atmospheric CO_2 concentration the entire GRIS is exposed to surface melt in summer. On formerly ice-covered grid points climate locally warms strongly via increased albedos, with positive feedbacks due to boreal forest expansion.
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1. Introduction

[2] The AMOC with its associated large northward heat transport is one of the key factors determining the climate of Europe. The future development of the strength of this overturning cell is one of the main uncertainties in predicting anthropogenic climate change. Most models participating in the IPCC AR4 predict a weakening of this overturning circulation associated with global warming [*Schmittner et al.*, 2005; *Meehl et al.*, 2007]. None of these models takes the effect of ice sheets into account. From stand-alone ice sheet models a reduction in size of the GRIS is predicted [e.g., *Huybrechts et al.*, 2004]. Numerous studies have shown that a substantial meltwater input into the North Atlantic will reduce the AMOC or even lead to its collapse [e.g., *Maier-Reimer and Mikolajewicz*, 1989; *Stouffer et al.*, 2006], whereas meltwater input from the Antarctic ice sheet (AAIS) can strengthen the AMOC [*Mikolajewicz*, 1998]. A reliable estimate of the future development of the AMOC requires the interactive coupling between an AOGCM and an ice sheet model.

[3] So far only few studies exist where AOGCMs have been coupled to dynamic ice sheet models. *Fichefet et al.* [2003] presented a 21st century simulation with an AOGCM coupled to a dynamic model of the GRIS without modifying the topography of the atmosphere model. The additional meltwater from the GRIS led to a collapse of the already rather weak AMOC. *Ridley et al.* [2005] found a slight reduction of the AMOC with a model with a more complete coupling. *Winguth et al.* [2005], *Vizcaíno* [2006] and *Mikolajewicz et al.* [2007] did not find a substantial effect of ice sheet melting on the AMOC, as the changes in atmospheric moisture transport were dominating. All these studies were suffering from the fact that the climate of the AOGCMs was not realistic enough to simulate reasonable ice sheets. Thus, for the forcing of the ice sheet model only anomalies superimposed onto a climatology were used (a technique in the following termed 'flux correction'). In all these models a simple parameterisation to estimate the surface melting from seasonal-mean atmospheric near-surface temperatures (degree-day method) has been used. Here we investigate the effect of meltwater input from the ice sheets with an ice sheet model coupled interactively without artificial flux correction to an AOGCM. The surface mass balance has been calculated using an energy balance scheme. This reduces uncertainties due to artefacts in the coupling compared to previous model studies. The effect of ice sheet interactions on anthropogenic climate change is investigated here by comparing simulations with and without interactive dynamic ice sheet model.

2. Model and Experimental Setup

[4] The AOGCM applied in this study is a coarse resolution version of ECHAM5/MPIOM [*Jungclaus et al.*, 2006a]. The spectral atmospheric model ECHAM5 was run in a T31 resolution (approx. 3.75°) with 19 levels. The ocean model MPIOM was used in a resolution of roughly 3° near the equator and with poles located on central Greenland and Antarctica with 40 levels. The dynamical vegetation model LPJ [*Sitch et al.*, 2003] yields potential vegetation. Anthropogenic effects on plants other than climate change and CO_2 concentration are not included.

[5] The 3-dimensional thermomechanical ice sheet model SICOPOLIS [*Greve*, 1997] is defined on two polar stereographic grids with a resolution of 80 km. The model is subject to the shallow ice approximation and does not include ice shelves, ice streams and outlet glaciers. In contrast to the standard version of SICOPOLIS, the surface mass balance is calculated with a new energy balance scheme, using 6h data of precipitation, air and dew point temperature, wind speed, and downward shortwave and longwave radiation with corrections for the deviations

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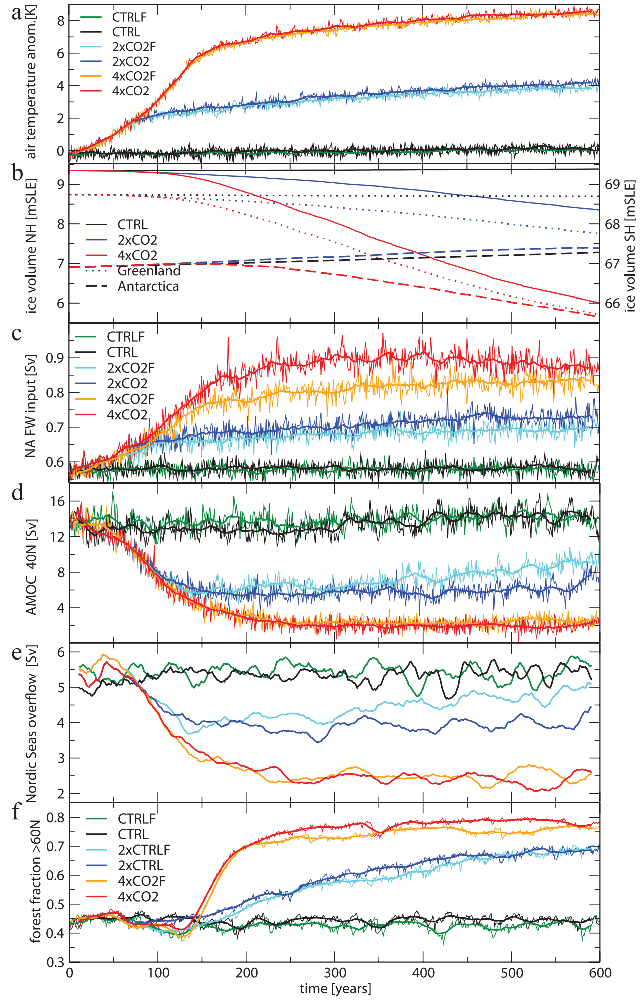

Figure 1. Timeseries of (a) global mean near surface air temperature anomaly (relative to climate of CTRL) (K), (b) volume (mSLE) of the entire Northern Hemisphere ice sheets (solid, left scale), the GRIS (dotted, left scale), and the AAIS (dashed, right scale), (c) net freshwater input into the Arctic and North Atlantic (north of 40°N) (Sv), (d) strength of the meridional overturning circulation at 40°N and 1040 m depth (Sv), (e) sum of the overflow through Denmark Strait (below 340 m) and the Faroe-Shetland Channel (below 520 m) (Sv), and (f) fraction of land north of 60°N covered with forest. In Figures 1a, 1c, 1d, and 1f annual mean values (thin lines) and running means over 20 years (thick lines) are displayed. Figure 1e shows only 20-year running mean data, Figure 1b shows annual mean data.

between the actual height and the surface height of the coarser atmosphere model. The ice sheets modify the glacier mask, atmospheric topography (except for the gravity wave drag), sea level and freshwater fluxes. Details of the model coupling and the spinup are given by *Vizcaíno* [2006]. Some information about the climate of the model and the coupling of the vegetation model can be found in the auxiliary material.[1] After longer separate spinups of the model components, the coupled model was integrated with and without ice-sheet model for 300 years. From these states all experiments were started. In the greenhouse simulations, the prescribed atmospheric pCO_2 was increased with a rate of 1%/year and stabilized at 560 (2xCO2 and 2xCO2F) and 1120 ppm (4xCO2 and 4xCO2F), respectively. In the control runs CTRL and CTRLF the pCO_2 was kept at 280 ppm. The suffix F indicates fixed ice sheets. Each experiment was integrated for 600 years.

3. Results

[6] In the last century of 4xCO2, the global mean near surface air temperature is approximately 8 K warmer, for 2xCO2 this is 4 K (Figure 1a). The interactive ice sheets have hardly any effect on the simulated global mean surface air temperature. At the end of the greenhouse experiments the model still shows a substantial warming (about 0.2 to 0.3 K/century). The simulated warming is strongest over the Arctic and the adjacent regions (Figures 2a and 2b). In 2xCO2 the Arctic Ocean becomes essentially ice-free in summer, in 4xCO2 the entire year. The simulated warming in the northern North Atlantic is rather small. In 2xCO2 even a cooling is simulated southwest of Iceland. The warming over southern Greenland is relatively low, in accordance with the temperature signal over the adjacent ocean.

[7] The northward expansion of the boreal forests towards the coast of the Arctic ocean is the most significant vegetation change (Figure 1f) in the greenhouse simulations. Even on parts of western Greenland boreal forest occurs. The increased forest cover enhances high-latitude warming by darker surface albedos and by a stronger masking of snowcover by trees in spring and early summer.

[8] At the end of the greenhouse simulations, the extent of the ice sheets on the northern hemisphere is reduced by 12% (2xCO2) and 33% (4xCO2), the ice volume (Figure 1b) has shrunk by 1.03 and 3.38 mSLE (m sea level equivalent). Approximately 90% of this stem from the GRIS. The mass loss of the GRIS is strongest on the northern half and on the northeast coast, where many grid points become ice-free (Figures 3a and 3b). In southern Greenland the reduction in ice thickness is rather small, which is consistent with the relatively small warming in this area. In 2xCO2 the pattern is similar, but weaker. In several high elevation areas of central and southeastern Greenland the thickness is increasing due to increased snowfall. Whereas in CTRL summer melt is restricted to low-elevation areas, in 4xCO2 the entire ice sheet shows surface melting in summer.

[9] The ablation rate of the GRIS is more than doubled in 4xCO2 between the years 200 and 400 compared to CTRL. After year 400 it decreases slowly from 0.17 to 0.13 Sv due to the reduced area of the ice sheet. The warming associated with the reduction in height enhances the melt and further reduces the fraction of the precipitation falling as snow. In 2xCO2 the ablation rate increases continuously. During the first century the mean net meltwater input rate from the ice sheets into the North Atlantic and the Arctic ocean does not exceed 0.02 Sv in both greenhouse simulations. In 4xCO2 a peak of almost 0.1 Sv is reached between years 200 and 400, before it decreases below 0.06 Sv. In 2xCO2 the meltwater input rate lies between 0.025 and 0.03 Sv. Our

[1] Auxiliary materials are available in the HTML. doi:10.1029/2007GL031173.

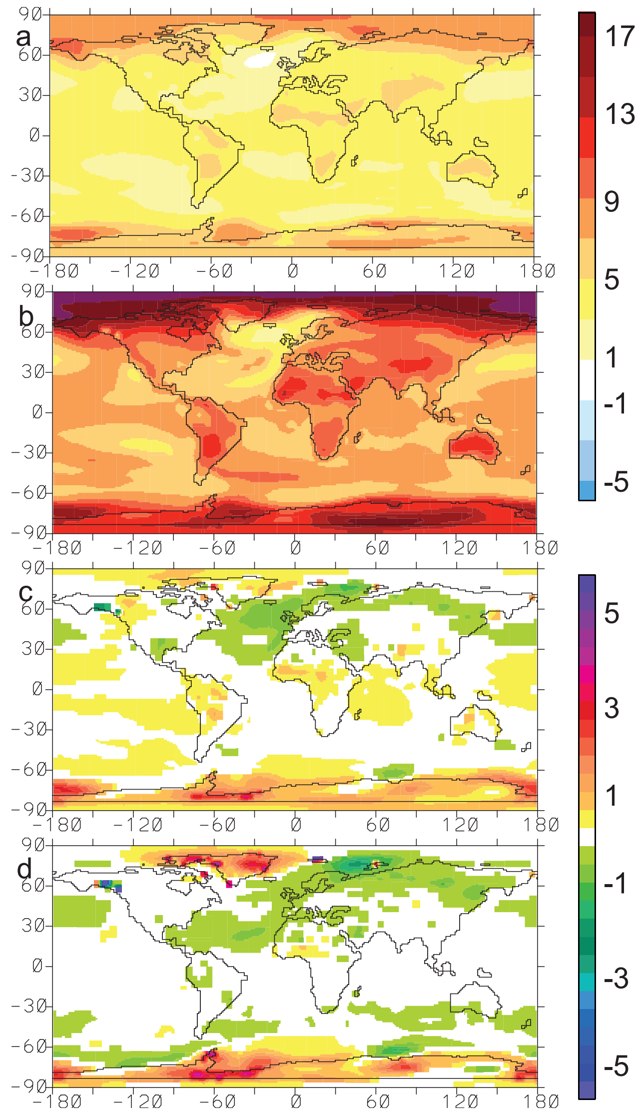

Figure 2. Anomaly of near surface air temperature (averaged over years 501 to 600) (K) relative to CTRL for (a) 2xCO2 and (b) 4xCO2. Effect of the interactive ice sheets on climate change for the (c) 2xCO2 and (d) 4xCO2 experiments calculated as the difference between the signals in the experiments with and without interactive ice sheets. In Figures 2c and 2d signals are shown only, where the statistical significance of a two-sided t-test exceeds 95%.

model yields rather similar meltwater input rates as the Hadley Centre model [*Ridley et al.*, 2005].

[10] At the end of 4xCO2 the accumulation on the AAIS is increased by almost 50%, the ablation is enhanced to ≈0.2 Sv. The net meltwater input rate is ≈0.05 Sv. Melting is still restricted to coastal areas. At the end of the simulation, the total mass loss of the AAIS is 1.63 mSLE (Figure 1b). About 60% of this stem from the Antarctic Peninsula. In the interior ice thickness is growing. At the coast several points become ice-free (Figure 3d). In 2xCO2 the increase of accumulation in the interior exceeds the relatively small increase of ablation (Figure 3c), leading to a growth of the AAIS of 0.11 mSLE. It should be stressed that our model does not adequately represent ice shelves which could be a key element in the future development of the AAIS.

[11] The strength of the AMOC weakens as the climate warms (Figure 1d). After 100 years all greenhouse simulations have a weakened AMOC (8 Sv at 40°N and 1040 m depth compared to 14 Sv in CTRL). In 4xCO2 the AMOC decreases to 2 to 3 Sv in the year 250. The strength of the overflow over the sills between the Nordic Seas and the North Atlantic weakens from 5.5 Sv to 4 Sv in 2xCO2 within 150 years (Figure 1e). In 4xCO2 the overflow is reduced even further to 2.5 Sv after 250 years. At the end of 4xCO2, a shallow overturning cell remains, with shallow, gradually deepening convection in the Greenland Sea. Below 2000 m the deep Atlantic is stagnating. In 2xCO2 the AMOC remains at ≈6 Sv for several centuries with a slight recovery during the last 50 years. Below 2500 m the deep Atlantic is stagnating. Convection in the north occurs in the Greenland Sea and in the Labrador Sea with a gradual deepening with time. The weak AMOC is associated with a strongly reduced northward heat transport of the Atlantic, which explains the relatively small warming of the northern North Atlantic.

[12] Comparison of 4xCO2 and 4xCO2F reveals that for this scenario the effect of ice sheet melting on the AMOC is not very important. The initial reduction of the AMOC is independent of the melting of the GRIS. The dominant factors are surface warming and the changes in atmospheric moisture transports. The net freshwater input into the North Atlantic and Arctic (precipitation minus evaporation over the ocean, river runoff and fluxes from the ice sheet) increases in 4xCO2 by more than 0.3 Sv around year 400, before it slowly decreases (Figure 1c). In year 150, when the AMOC is already strongly weakened (Figure 1d), the total freshwater flux anomaly is 0.2 Sv with 0.05 Sv meltwater from the GRIS. When the melting reaches maximal strength, the AMOC is already very weak/collapsed. The effect of additional meltwater input on an already collapsed AMOC is rather small. Towards the end of the experiments, however, the AMOC in 4xCO2F is slightly stronger.

[13] In 2xCO2 and 2xCO2F the AMOC shows an almost identical behaviour until year 150. Whereas it slowly recovers in 2xCO2F, the AMOC in 2xCO2 stays at the low level until year 550. The relative small meltwater input of 0.02 to 0.03 Sv is sufficient to reduce the AMOC's ability to recover substantially. Whereas in 2xCO2F the strength of the overflow over the sills (Figure 1e) increases continuously from 4 Sv to 5 Sv between years 200 and 600, in 2xCO2 it remains at 4 Sv for another 400 years. In 2xCO2F, the convection in both the Greenland Sea and at the mouth of the Labrador Sea is always deeper than in 2xCO2. The surface salinity averaged over the subpolar North Atlantic (between 45°N and the sills) is more than 0.3 permille lower in 2xCO2 due to the meltwater input from the GRIS, which explains the shallower convection depths and the weaker overturning.

[14] Other comparable long-term simulations also found only a moderate/small effect of Greenland melting on the AMOC: *Ridley et al.* [2005], *Winguth et al.* [2005], and *Mikolajewicz et al.* [2007] in simulations with interactive ice sheet models, *Jungclaus et al.* [2006b] in experiments with prescribed meltwater input. In contrast, *Swingedouw et al.* [2006] report a rather strong effect of the meltwater input

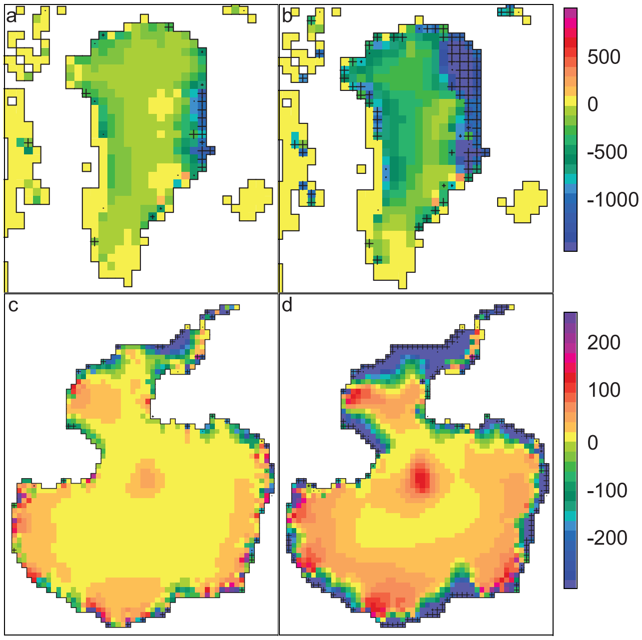

Figure 3. Difference in simulated ice thickness (m) for Greenland in (a) 2xCO2 and (b) 4xCO2 and Antarctica for (c) 2xCO2 and (d) 4xCO2. Displayed are averages over the years 501 to 600 relative to CTRL. Crosses indicate a change in the glacier mask.

of 0.2 Sv in experiments with an AOGCM coupled to a snowpack model. Similar simulations with stand-alone AOGCMs have given quite diverse results. *Manabe and Stouffer* [1994], similar to this study, reported a complete collapse of the AMOC in a 4xCO2 simulation, whereas *Voss and Mikolajewicz* [2001] and *Wood et al.* [2003] obtained a strong reduction with a subsequent slow recovery. Shorter simulations comparing the response of several AOGCMs indicate a relatively high sensitivity of our model's AMOC to an increase of greenhouse gas concentrations [*Gregory et al.*, 2005].

[15] The influence of interactive ice sheets on near-surface air temperature at the end of 4xCO2 is strongest on Greenland and on the Antarctic Peninsula, where the changes in surface topography and glacier mask are largest. The lower surface elevation leads to a substantial warming over most of Greenland (Figure 2d) with a maximum over the northeast and the western half of the Nordic Seas. The warming signal is strongest in winter and is to first order consistent with the lowering of the ice sheet. In summer the temperature signal due to the lower ice sheet surface is rather small as in both 4xCO2 simulations almost the entire surface of the GRIS is at the melting point, which limits the surface warming. At the margins of the ice sheets and on some locations of ice caps (e.g. Spitsbergen) the retreat of the ice strongly modifies the surface albedo with strongest warming (locally more than 10 K) in summer. Over northern Eurasia cooler temperatures are caused by changes in the stationary wave field and the path of the stormtracks due to the reduced size of the GRIS. The simulated circulation and temperature changes due to a reduction of the GRIS are consistent with experiments in which the effect of a removal of the GRIS on atmospheric circulation has been investigated [e.g., *Ridley et al.*, 2005; *Vizcaíno*, 2006]. In the 2xCO2 experiments a very similar, but weaker effect of topographic changes on the near-surface temperature is simulated. A difference to the 4xCO2 experiments is the marked cooling over the northeastern Atlantic (see Figure 2c) due to the stronger AMOC in 2xCO2F compared to 2xCO2.

4. Conclusions

[16] The state-of-the-art technique to estimate the future development of the ice sheets (and thus sea-level) are simulations with stand-alone ice sheet models with forcing from AOGCM simulations. The difference in topography between the fixed topography of the atmosphere and the time-varying surface of the ice-sheets is in general accounted for by assuming a constant height correction [e.g., *Huybrechts et al.*, 2004]. In our simulations this would lead to an overestimation of the melting on the central GRIS as the increase in summer temperature is limited due to the presence of ice at the surface. On the other hand, the changes in ice mask strongly enhance the simulated warming at the margins of the ice sheet. Future studies would benefit from a fractional glacier mask in the atmospheric model.

[17] During the rapid initial warming phase the meltwater input from the GRIS is small compared to the freshwater input into the North Atlantic and Arctic due to enhanced poleward atmospheric moisture transports. The effect of meltwater from the ice sheets becomes more influential during later phases of the experiment, where it contributes to prolong the state with very weak and shallow NADW cells. Like most other studies, we found only a moderate effect of ice sheets on the AMOC. The coupling between ice sheet and climate models has pronounced effects on the estimates of the future development of the ice sheets, but only small effects on large-scale climate, with possibly bigger effects if the model is close to a bifurcation point of the AMOC.

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