38. Mechanisms Leading to the Last Glacial Inception over North America: Results From the CLIMBER-GREMLINS Atmosphere–Ocean–Vegetation Northern Hemisphere Ice-Sheet Model

Masa Kageyama[1], Sylvie Charbit[1], Catherine Ritz[2], Myriam Khodri[3] and Gilles Ramstein[1]

[1]Laboratoire des Sciences du Climat et de l'Environnement/IPSL (UMRI572 CEA-CNRS-UVSQ), CE Saclay, L'Orme des Merisiers, Bâtiment 701, 91191 Gif sur Yvette Cedex France
[2]Laboratoire de Glaciologie et Géophysique de l'Environnement, Saint-Martin-d'Hères, France
[3]Geophysical Fluid Dynamics Laboratory, U.S. Department of Commerce, National Oceanic and Atmospheric Administration, Princeton, New Jersey, USA, Now at: UR PALEOTROPIQUE (IRD) and LSCE, 32 Avenue Henri Varagual, 93143 Bondy Cedex France

ABSTRACT

The CLIMBER-GREMLINS model is an atmosphere–ocean–vegetation–northern hemisphere ice-sheet model able to simulate ice-sheet growth in response to the transient forcings (insolation and CO_2 changes) of the period 126–106 kyr BP. In the present version of the model, this growth mainly occurs over North America and reaches an equivalent of 17 m in sea-level drop. To quantify the role of the vegetation, ocean and ice-sheet feedbacks in this glaciation of North America, we have conducted sensitivity experiments in which the feedback of each of these components is sequentially switched off. These experiments show that, in this model. (1) glacial inception does not occur when vegetation is fixed to its interglacial state (experiment testing the response of the atmosphere–ocean–boreal land–ice system), (2) glacial inception occurs faster than in the standard experiment when the ocean surface characteristics (surface temperature and sea ice extent) are prescribed to their interglacial seasonal cycle (experiment testing the sensitivity of the atmosphere–vegetation–boreal land–ice system), (3) the ice-sheet albedo and altitude feedbacks are not crucial for starting the glaciation, but the albedo feedback doubles the ice volume growth rate, (4) the potential effect of a reduction in the thermohaline circulation is tested via an additional experiment in which we have forced it to its 'off' mode. The results show that this mode favours the fastest land ice growth of all our experiments.

38.1 INTRODUCTION

The last glacial inception occurred at a time (ca. 115 000 years ago) when the northern high-latitude summer insolation was close to a minimum (Berger, 1978; Berger *et al.*, this volume). The Milankovitch theory (Berger, 1988) relates weak summer insolation values (typically under a certain threshold) to cold summers favouring perennial snow at high latitudes and therefore, eventually, ice-sheet growth. The study of palaeoclimatic records has suggested that this relationship between summer insolation and northern hemisphere ice-sheet growth follows many pathways. For instance, on the one hand, the oceanic temperatures in the Greenland–Iceland–Norwegian seas significantly decrease at around 120 kyr BP, much before the minimum in summer insolation (Cortijo *et al.*, 1999). On the other hand, vegetation changes, consisting of the southward retreat of the boreal tree line, help amplify the snow albedo feedback which is somewhat inhibited in a forest environment (Sánchez Goñi *et al.*, 2005). Numerical experiments have confirmed that the feedbacks from the vegetation and the ocean, in addition to the ice-sheet albedo feedback, play an important role in amplifying the initial insolation signal and creating the climatic conditions ideal for ice-sheet growth (see Vettoretti and Peltier (2004) for a review). However, at present, there are only few models in which the feedbacks from all three of these components can be compared within a single framework. Here we present the results from one of these models: CLIMBER-GREMLINS. This model simulates significant ice-sheet growth over North America in response to the insolation and CO_2 changes occurring between 126 and 106 kyr BP (Kageyama *et al.*, 2004). In the present paper, we present the results from three sensitivity experiments designed to evaluate the feedbacks from ice sheet, vegetation and ocean. The fourth sensitivity experiment was inspired by the evidence for abrupt climate change as early as 115 kyr BP (see for instance NGRIP Project Members, 2004). One of the possible explanations for these abrupt climate changes over Greenland is the sudden switch of the thermohaline circulation from an active mode (with the formation of deep water in the North Atlantic and large heat and mass transport to the North in the Atlantic Ocean) to a 'sluggish' or 'off' mode in which these nprocesses are nearly stopped. We have not attempted to mimic each of these events (see Kubatzki *et al.*, in this volume, for another approach in the simulation of these events). Rather, we have forced the thermohaline circulation (THC) in its off mode (in the sense of Ganopolski and Rahmstorf 2001) to examine the maximum effect of an extremely slow THC on northern hemisphere ice-sheet growth.

38.2 MODEL DESCRIPTION

The CLIMBER-GREMLINS model has been developed by coupling the atmosphere–ocean–vegetation model CLIMBER2.3 (Petoukhov *et al.*, 2000) to the northern hemisphere ice-sheet model GREMLINS (GREnoble Model for Land Ice in the northern hemisphere: Ritz *et al.*, 1997). Here we provide a brief summary of the model characteristics. The atmosphere in CLIMBER2.3 is represented on a coarse grid (seven sectors in longitude and 18 10°-wide latitudinal bands) by a statistical-dynamical model which includes a description of the atmospheric hydrological cycle (moisture transport by the mean circulation and transport by mid-latitude eddies which is parameterised). The ocean is described by three latitude–depth basins for the Atlantic, Indian and Pacific oceans. The vegetation is described by two vegetation types: grass and trees, and provides a simple representation of the modulation of the snow albedo by vegetation.

The ice-sheet model GREMLINS is a three-dimensional thermomechanical model for grounded ice, which is forced by temperature and mass balance fields at its surface. In the present version of the model, the ablation is computed from the mean annual and summer surface temperatures via a positive-degree-day method. Therefore, the forcing fields for GREMLINS are the annual and summer temperatures and annual snowfall. GREMLINS distinguishes three surface types: ice sheets, ice-free land and ocean. Grid points are attributed to the ocean surface type if they are not covered by ice sheets and if the altitude of the grid point is below sea level, which is given to the model as an input.

In all the experiments presented here, the climate and ice-sheet models exchange information every 200 years. A downscaling procedure (described in Kageyama *et al.*, 2004; Charbit *et al.*, 2005) has been developed in order to transfer the results from the CLIMBER coarse atmosphere to the fine grid (45 km × 45 km) of GREMLINS. This procedure is based on the calculation, in the atmosphere component of CLIMBER, of the surface radiative balance on 15 different levels, as if they were the grid box level. These levels are regularly spaced every 300 m between 0 and 4200 m. Thus, within each CLIMBER grid box, we obtain the surface temperatures for 15 different altitudes. The total precipitation vertical profiles are computed accounting for the free atmosphere temperature vertical profiles computed in CLIMBER, and snowfall is obtained on the 15 vertical levels as a fraction of the total precipitation which is itself a function of the temperature at these 15 different levels. Furthermore, these calculations are performed for each CLIMBER surface type (ocean, sea ice, ice sheets, forests, grassland and desert) and averaged over the three surface types of GREMLINS (ocean, ice-free land and ice sheets). The mean annual and summer temperatures, as well as the annual accumulation, are then computed on the GREMLINS grid by trilinear interpolation and serve as forcing for GREMLINS. In turn, GREMLINS computes the fraction of ocean, land ice and ice-free land as well as the altitude over the continents, which are given back to CLIMBER as new boundary conditions. Thus, the altitude of the ice sheet is taken into account in two different ways in the model. The effect of the altitude on the atmospheric dynamics is computed using the average altitude of each grid box in CLIMBER. The dependence of temperature and snowfall on altitude is taken into account via our downscaling procedure and the calculation of these fields on the 15 vertical levels defined above. In our sensitivity experiments, when we refer to the altitude feedback of the ice sheet on the climate, we have only isolated the first effect (effect of the altitude on the atmospheric dynamics), not the second one.

In the present version, this model has some shortcomings which are currently being corrected. As far as the glacial inception is concerned, it does not simulate a fast enough ice-sheet growth, probably partly because no inception occurs, even after 115 kyr BP, over Fennoscandia. However, ice growth over North America is substantial, which provides the opportunity to study the role of the feedbacks from the different components of the climate system during the glacial inception.

38.3 EXPERIMENTAL DESIGN

In the standard experiment (STD), the model is simply forced by the evolving insolation from 126 to 106 kyr BP (Berger, 1978; Berger *et al.*, this volume, Fig. 5) and the CO_2 concentration measured in the Vostok ice core (Petit *et al.*, 1999), which fluctuates between 260 and 280 ppm from 126 to 112 kyr BP and decreases from 260 to 230 ppm from 112 to 106 kyr BP. Furthermore, the northern ice-sheet model is forced by the time-dependent sea-level values from Bassinot *et al.* (1994). For simplicity of the experimental design, the initial ice-sheet and bedrock altitudes are set to their present values: the only ice sheet in the domain is Greenland. This is a simplification in the sense that the Greenland ice sheet is known to have been smaller than present during the Eemian. However, its extent and volume are not very well constrained (see for instance Cuffey and Marshall, 2000). In the experiments presented here, the ice-sheet model adjusts to the Eemian external forcings within the first thousand years of the experiment (Kageyama *et al.*, 2004). The climatic (atmosphere–ocean–vegetation) initial state is in equilibrium with this ice-sheet distribution and 126 kyr BP insolation and CO_2. In particular, the THC is its active mode, with a maximum overturning at around 20 Sv.

From this STD experiment, four sensitivity experiments have been conducted, which are summarised in Table 38.1. In the first one (125VEG), the vegetation is fixed to its 125 kyr value, consisting of extensive forests over northern North America. The

Table 38.1 Summary of the numerical experiments used in the present study

Experiment	Atmosphere	Ocean and sea ice	Vegetation	NH ice sheets
STD	Interactive	Interactive	Interactive	Interactive
125VEG	Interactive	Interactive	Fixed to 125 kyr BP value	Interactive
122ICE	Interactive	Interactive	Interactive	Fixed to 122 kyr BP value. Ice sheet model then used as a diagnostic
125OCN	Interactive	Seasonal cycle of ocean and sea ice characteristics fixed to 125 kyr BP values	Interactive	Interactive
THCoff	Interactive	Interactive, but with thermohaline circulation in off mode	Interactive	Interactive

The atmosphere, ocean, sea ice and vegetation models are the components of the CLIMBER2.3 model. The northern hemisphere ice-sheet model is the GREMLINS (GREnoble Model for Land Ice in the Northern hemisphere) model.

atmosphere, ocean, sea ice and ice sheets are interactive. In the second one (122ICE), the ice-sheet feedback is inhibited from 122 kyr BP onwards: from this date, the ice-sheet extent and thickness and land–sea distribution computed by GREMLINS are not transferred to the CLIMBER model, which is forced, until the end of the run, by the 122 kyr BP values (i.e. minimal ice sheets) of these variables. The extent and volume of the Greenland ice sheet at this stage of the simulation are in broad agreement with reconstructions (see for instance Tarasov and Peltier, 2003). On the other hand, GREMLINS is still forced by the climatic output from CLIMBER in order to evaluate the strength of the ice-sheet feedback. All three of these experiments have been analysed by Kageyama *et al.* (2004).

To investigate the role of the ocean, we have performed a third and fourth sensitivity run. The third sensitivity experiment (125OCN) mimics the approach we followed to study the role of the vegetation: the ocean and sea ice models are disconnected from the earth system model from 125 kyr BP onwards, and from this date the atmosphere–vegetation–ice-sheet model is forced by the 125 kyr BP sea-surface temperature and sea ice cover seasonal cycle. This experimental design has two main consequences: (1) The ocean and sea ice cannot react anymore to changes in incoming insolation and do not play any role in the system equilibration to the evolving incoming insolation and changing CO_2; (2) From 125 kyr BP onwards, the northern SSTs in summer are warmer and sea ice extent smaller than if they had been computed, with a maximum difference at 115 kyr BP. The difference in winter SSTs is not as large since winter insolation at high latitudes is very small anyway.

The fourth sensitivity experiment (THCoff) lets the ocean and sea ice adjust to the evolving incoming insolation but with the constraint of the THC being in its ‘off’ mode. This state has been obtained by imposing a permanent 0.06 Sv freshwater flux to the north Atlantic ocean in the fashion of Ganopolski and Rahmstorf (2001)’s northern North Atlantic scenario (i.e. between 50 and 70 °N). In the present case, we obtain an ‘off’ mode in which the THC is completely shut down, rather than a slow mode which can be obtained by perturbing a glacial state. In the palaeorecords, there is some evidence for THC instabilities as early as 110 kyr BP (Lehman *et al.*, 2002), but these are short-lived compared to the duration of our experiments. However, they could have a important impact on ice-sheet growth by modifying the northern high-latitude climate and therefore the possibilities of ice sheet growth. The THCoff experiment is a simple experiment which is not intended to be realistic but is designed to test the sensitivity of ice-sheet growth to the state of the THC in a drastic (and exaggerated) scenario.

38.4 COMPARISON OF THE SIMULATED ICE-SHEET EVOLUTIONS

The evolutions of the volume (upper panels) and area (lower panels) of land ice over North America in each experiment is shown in Fig. 38.1 as a function of time (left panels) and of June–July–August average insolation (right panels). In the STD experiment, inception occurs at 121 kyr BP, and the glaciated area (Fig. 38.1, lower left panel, black curve) increases steeply to reach 3.5 million km^2 in 3.5 kyr. From 117.5 kyr BP, the ice-sheet area increases slowly and levels at the value of 4.5 million km^2 in the last two to three thousand years of the simulation. In terms of land–ice area, the inception is therefore fast and concomitant with the large decrease in insolation, which is at its minimum at ~114.4 kyr BP. In terms of volume, the inception is more gradual but also reaches an apparent equilibrium during the last thousands of years of the simulation (Fig. 38.1, upper left panel, black curve). In fact, the ice-sheet extent and volume increase as long as the summer insolation decreases (Fig. 38.1, right panels, black curves), but do not decrease when the June-July-August insolation increases, from 114.4 kyr BP onwards. Rather, their growth rate drastically decreases and reaches zero by the last thousand years of the simulation.

The importance of the ice-sheet feedback suggested by this result is further shown

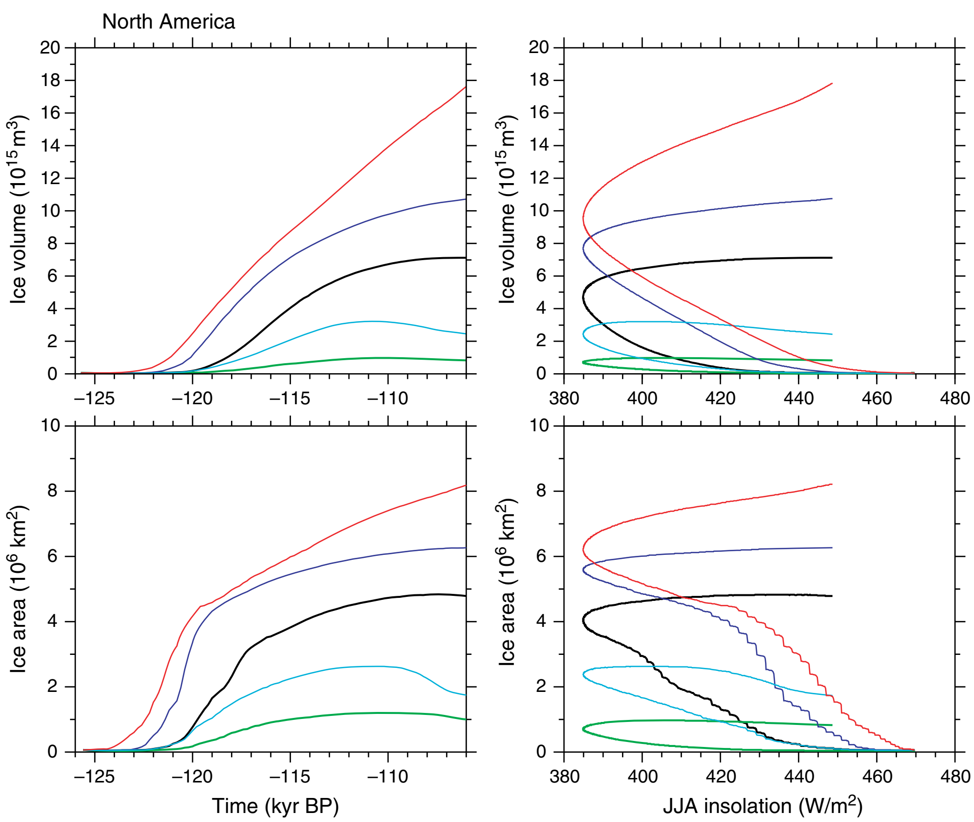

Fig. 38.1 (top left) Evolution of the ice-sheet volume covering North America in all the simulations: STD (black), 125VEG (green), 125OCN (blue), 122ICE (cyan), THCoff (red); (bottom left) same for the ice-sheet area; (top right) ice-sheet volume as a function of June–July–August insolation (in W/m^2). The simulations start at the r.h.s. of the graph with ice volume equal to zero. The minimum insolation is reached at 114 400 yr BP; (bottom right) same graph for the ice-sheet area.

by the 122ICE sensitivity experiment (Fig. 38.1, cyan curve). In this experiment, the atmosphere–ocean–vegetation model does not 'know' about the inception over North America. As a result, the glaciated area and volume, computed off-line, increase smoothly from 121 to 110 kyr BP and then react to the increase in summer insolation. Despite the increase of summer snow cover from the beginning of the run, with a summer snow fraction of 20 to 25% at 75 °N, to 115 kyr BP, with a summer snow fraction of 55% in the same latitude band, the summer albedo increase is much smaller than in the STD run, in which summer snow fraction equals 100% at 75 °N at 115 kyr BP. The snow albedo feedback alone is therefore not strong enough, in our model, to initiate a significant ice sheet, in terms of both volume and area of land–ice. In the 122ICE experiment, the ice-sheet extent, with a maximum of 3 million km^2, is always smaller than the corresponding STD value. The ice-sheet feedback (and not only the snow albedo feedback) is therefore important in determining the ice-sheet growth rate, maximum ice-sheet area and volume, and as a consequence, the delay of the ice-sheet response to the evolving external forcings.

The influence of vegetation in simulating a glacial inception with our model is crucial (experiment 125VEG, Fig. 38.1, green curves). When vegetation is prescribed to its interglacial value, the glaciated area only reaches 1 million km^2, and the ice-sheet volume is negligible compared to the results of the other simulations. To disentangle the role of the vegetation from that of the ice sheets, we have conducted experiments equivalent to 125VEG and STD, but without any coupling to the ice-sheet model. In the latter experiment, the summer surface air temperature over northern North America is up to 2.4°C cooler than in the experiment with fixed vegetation, with a surface albedo up to 0.12 larger and a summer snow fraction up to 0.18 larger. Once coupled to the ice-sheet model, this temperature difference translates into a positive snow mass balance and therefore inception over northern North America for the experiment with interactive vegetation. Without the transition from forest to grass over northern North America, the summer albedo is not high enough and the summer temperature does not decrease enough to allow for a positive snow mass balance over these regions and inception does not occur. The ice-sheet feedback further amplifies the difference between the experiments with interactive vegetation and with fixed vegetation.

Although designed in a very different fashion, both experiments testing the role of the ocean result in a much larger glaciated area than in STD. In the case of 125OCN, the inception occurs 1.5 kyr earlier than in STD, and the continental ice area grows very fast until 120 kyr BP, reaching 4.5 million km^2 at that time. It then increases smoothly to reach more than 6 million km^2 by the end of the run. The ice area and volume in this simulation are therefore larger than in STD for the whole simulated period. This is unexpected: prescribing warm summer SSTs and less extensive sea ice in the mid- and high northern latitudes could have resulted in warmer temperatures over land as well. This is not what is observed in this experiment. As summer insolation decreases, summer temperature over land and over Arctic sea ice decreases more than in the STD experiment. This is also the case when we compare two equivalent experiments in which the ice-sheet model is not coupled to CLIMBER2.3, showing that this over-reaction of the land and sea ice surfaces to insolation changes when the surface ocean is prescribed compared to the interactive ocean case occurs primarily in the atmosphere–ocean–vegetation model. In the prescribed SST and sea ice area experiment, the atmosphere appears to compensate for the fact that the SSTs cannot adjust to the insolation changes, by being colder than in the STD experiment in summer and warmer in winter. A direct result of these colder land temperatures is decreased ablation and therefore faster ice-sheet growth. Also, more precipitation falls as snow, increasing the ice-sheet growth rate. Lastly, in summer, the equator-to-pole temperature gradient is reinforced in 125OCN, which results in slightly amplified heat and moisture transport to high latitudes. However, the corresponding total precipitation changes are rather small when compared to the annual snow fall increase. Therefore, it is the local temperature decrease over land and sea ice, in 125OCN, compared to STD that is primarily responsible for faster ice-sheet growth, and not accumulation changes due to increased moisture fluxes.

In the experiment in which we constrain the THC in its 'off'-mode, the ice-sheet growth is even faster than in the previous experiment: glaciation starts earlier, at 124 kyr BP, the fast growth phase ends at 121 kyr BP with an area of 5 million km^2, followed by a slower growth phase until the end of the simulation. The growth rate during this second phase is constant and much larger than in any other simulation. The final area is more than 8 million km^2. In this run, the SSTs are free to adjust to the evolving external forcings, with the constraint of the THC being in its off mode. The resulting climate over both land and ocean at high latitude is very cold, both in summer and in winter. The largest climate differences compared to STD are in the mid-latitudes (not shown), which corresponds to the reaction to the THC documented in Ganopolski and Rahmstorf (2001). This suggests that as soon as the THC slows down, the ice-sheet growth can be much faster because of a decrease in ablation, and an increased accumulation related to larger precipitation, of which a larger fraction falls as snow because of the lower temperatures. The increase in precipitation could itself be related to an increase in mid-latitude northward moisture fluxes, simulated both in summer and in winter in the 40 to 60°N latitude band. Thus, periods of slow THC, which have been documented for the last inception (Lehman *et al.*, 2002), will therefore be included in more realistic experiments in the future.

The final distribution of glaciated areas in STD is shown on the top graph of Fig. 38.2. The southern tip of the Greenland ice sheet has melted as a result of the interglacial high summer insolation values and has not recovered by the end of the simulation. Inception has occurred over Alaska, the Canadian Rockies, the western Canadian plains and the Canadian Archipelago. This distribution is not accurate but could be a result of our rather crude horizontal interpolation of the CLIMBER results onto the GREMLINS grid. Here, the final distribution of the ice sheets in the different runs is given for the purpose of comparing the experiments. In all of them, inception occurs over the Canadian Rockies, because of the altitude effect on temperatures and relatively high precipitation there, which is related to our bilinear interpolation of precipitation fields onto the GREMLINS grid. For an inception to occur over the Canadian archipelago, both the ice-sheet and vegetation feedbacks are needed, as shown by the 125VEG and 122ICE experiments. The inception occurs separately over this area and on the western Canadian plains, where inception starts afterwards.

In the case of 125OCN and THCoff runs, the final North American ice sheet extends from Hudson Bay to Alaska. The southern tip of Greenland is ice-covered, and so is eastern Siberia. In our model, as long as Hudson Bay is below sea level in the ice-sheet model, the computed mass balance over this area is based on CLIMBER's results over the corresponding ocean (see section 2), which induce relatively high values of ablation over this area. Moreover, the ice-sheet model does not include any representation of ice shelves which could help getting a glacial inception over this region. The glaciation over Alaska and Siberia could be related to each other, in the sense that when an ice sheet develops over Alaska, the surrounding region gets colder, which results in an ice sheet growing over Siberia. Specific experiments exploring the climate–cryosphere dynamics over this region are currently being analysed.

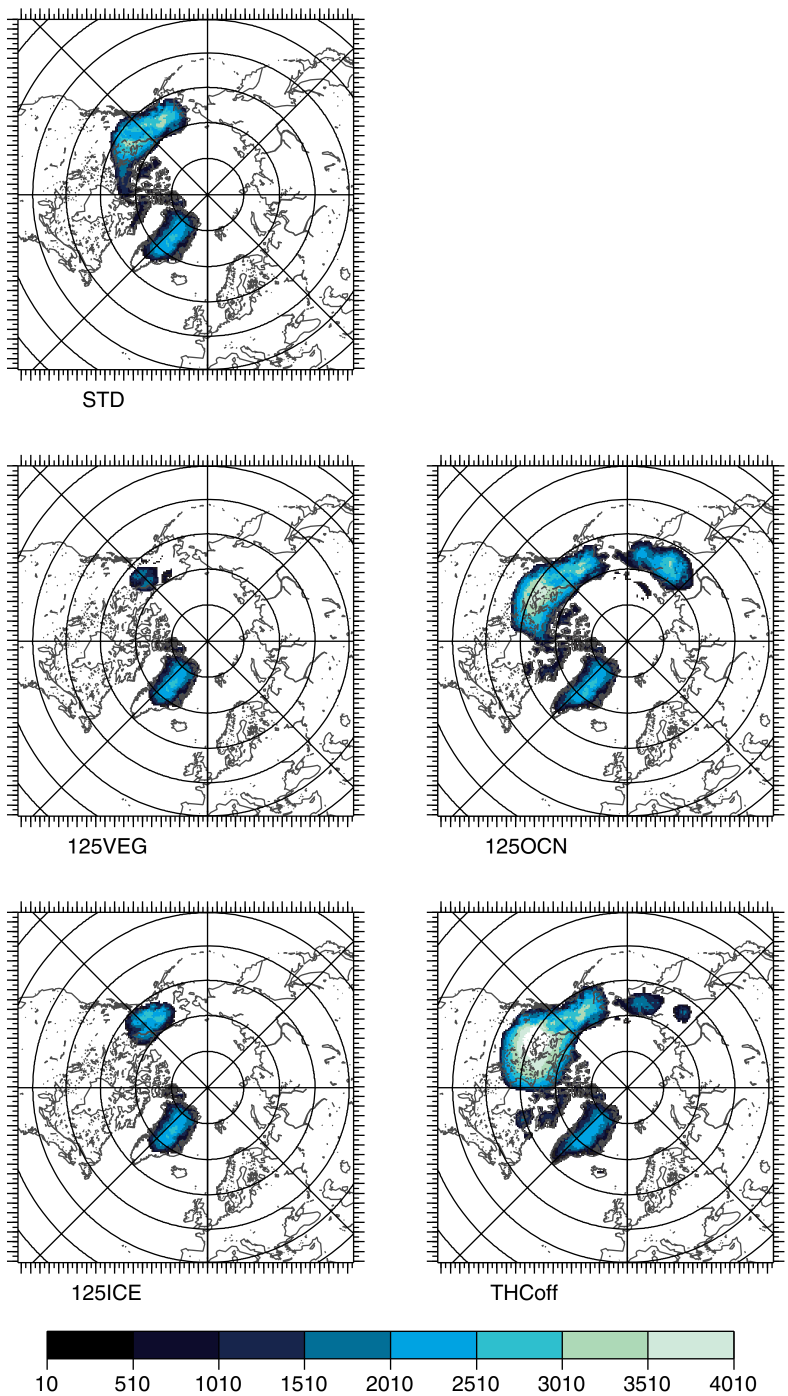

Fig. 38.2 Ice-sheet thickness (in m) at the end of the STD and sensitivity experiments.

38.5 DISCUSSION AND PERSPECTIVES

The CLIMBER-GREMLINS model, in its present version, is able to simulate significant land ice over North America as a result of the variations in incoming insolation and CO_2. Even if the location and size of the ice sheets are far from being perfect, such a model, which includes the atmosphere, the ocean, the vegetation and the northern hemisphere ice sheets, allows for a comparison, within a single framework, of the role of each of these components in the last glacial inception. Our sensitivity experiments show that both vegetation and ice-sheet feedbacks are essential to simulate significant ice-sheet growth. The role of the ocean is more ambiguous. The experiment in which SSTs and sea ice are prescribed to their 125 kyr BP seasonal cycle shows that glaciation can occur even when summer SSTs are warm. This, however, could be an artefact of the experimental design, which forces the model to adjust to the decreasing incoming insolation by decreasing the temperature over land, which is logical since ocean temperatures cannot be adjusted in this experiment. Whether the same behaviour occurs in other models, and in particular in general circulation models, remains to be investigated. The second experiment testing the role of the ocean shows that if SSTs are allowed to adjust, but with an 'off'-mode THC, ice-sheet growth is even faster. This experiment, in which the THC has been switched off during the whole duration of the run, is unrealistic, but there is evidence of some periods of abrupt THC variations within the inception that would be worth being investigated.

Current and future work on our model includes developing a new downscaling procedure to account for precipitation redistribution as a function of topography and land–sea mask. This appears to be essential to correct the locations of glacial inception and could be crucial to simulate a glaciation over Fennoscandia. In the present version, our model results are different from those obtained at PIK (Calov *et al.*, 2005a, 2005b). A comparison between our ice-sheet models and coupling procedure would be very informative about the dominant processes leading to glacial inception.

ACKNOWLEDGEMENTS

We thank the PIK Climber team for providing us their CLIMBER2.3 model and the organisers of the 2004 DEKLIM-EEM workshop for stimulating discussions. We are also grateful to Marie-France Loutre and Michel Crucifix, whose reviews greatly helped in clarifying the initial manuscript.

REFERENCES

Bassinot, F. C., L. D. Labeyrie, E. Vincent, X. Quidelleur, N. J. Shackleton and Y. Lancelot, 1994. The astronomical theory of climate and the age of the Brunhes-Matuyama magnetic reversal. Earth and Planetary Science Letters 126, 91–108.

Berger, A., 1978. Long-term variations of daily insolation and quaternary climatic changes. Journal of the Atmospheric Sciences 35, 2362–2367.

Berger, A., 1988. Milankovitch theory and climate. Reviews of Geophysics 26, 624–657.

Calov, R., A. Ganopolski, M. Claussen, V. Petoukhov and R. Greve, 2005a. Transient simulation of the last glacial inception. Part I: glacial inception as a bifurcation in the climate system. Climate Dynamics 24, 545–561.

Calov, R., A. Ganopolski, V. Petoukhov M. Claussen, V. Brovkin and C. Kubatzki, 2005b. Transient simulation of the last glacial inception. Part II: sensitivity and feedback analysis. Climate Dynamics 24, 563–576.

Charbit, S., M. Kageyama, D. Roche, C. Ritz and G. Ramstein, 2005. Investigating the mechanisms leading to the deglaciation of past continental Northern Hemisphere ice sheets with the CLIMBER-GREMLINS coupled model. Global and Planetary Change 48, 253–273.

Cortijo, E., S. Lehman, L. Keigwin, M. Chapman, D. Paillard and L. Labeyrie, 1999. Changes in meridional temperature and salinity gradients in the North Atlantic ocean (30°–72°N) during the last interglacial period. Paleoceanography 14, 23–33.

Cuffey, K. M. and S. J. Marshall, 2000. Substantial contribution to sea level rise during the last interglacial from the Greenland ice sheet. Nature 403, 591–594.

Ganopolski, A. and S. Rahmstorf, 2001. Rapid changes of a glacial climate simulated in a coupled climate model. Nature 409, 153–158.

Kageyama, M., S. Charbit, C. Ritz, M. Khodri and G. Ramstein, 2004. Quantifying ice sheet feedbacks during the last glacial inception. Geophysical Research Letters 31, doi:10.1029/2004GL021339.

Lehman, S. J., J. P. Sachs, A. M. Crotwell, L. D. Keigwin and E. A. Boyle, 2002. Relation of subtropical Atlantic temperature, high-latitude ice rafting, deep water formation, and European climate 130,000–60,000 years ago, Quaternary Science Reviews 21, 1917–1924.

North Greenland Ice Core Project Members, 2004. High–resolution record of Northern Hemisphere climate extending into the last interglacial period. Nature 431, 147–151.

Petit, J.-R., J. Jouzel, D. Raynaud, N. I. Barkov, J.-M. Barnola, I. Basile, M. Benders, J. Chappellaz, M. Davies, G. Delaygue, M. Delmotte, V. M. Kotlyakov, M. Legrand, V. Y. Lipenkov, C. Lorius, L. Pépin, C. Ritz, E. Saltzman and M. Stievenard, 1999. Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. Nature 399, 429–436.

Petoukhov, V., A. Ganopolski, V. Brovkin, M. Claussen, A. Eliseev, C. Kubatzki and S. Rahmstorf, CLIMBER-2: a climate system model of intermediate complexity. Part I: model description and performance for present climate, 2000. Climate Dynamics 16, 1–17.

Ritz, C., A. Fabre and A. Letréguilly, 1997. Sensitivity of a Greenland ice sheet model to ice flow and ablation parameters: consequences for the evolution through the last climatic cycle. Climate Dynamics 13, 11–24.

Sánchez-Goñi, M. F., M.-F. Loutre, M. Crucifix, O. Peyron, L. Santos, J. Duprat, B. Malaizé, J.-L. Turon and J.-P. Peypouquet, 2005. Increasing vegetation and climate gradient in Western Europe over the Last Glacial Inception (122–110 ka): data-model comparison. Earth and Planetary Science Letters 231, 111–130.

Tarasov, L. and W. R. Peltier, 2003. Greenland glacial history, borehole constraints and Eemian extent. Journal of Geophysical Research 108(B3), 2143, doi:10.1029/2001JB001731.

Vettoretti, G. and W. R. Peltier, 2004. Sensitivity of glacial inception to orbital and greenhouse gas climate forcing. Quaternary Science Reviews 23, 499–529.