Theories and Modeling of Glacial–Interglacial Cycles and Glacial Inception

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Contents

1 Introduction

Today we live in an interglacial period that started about 11,000 years ago. Interglacials occurred before, alternating with longer glacial periods. Before the so-called mid-Pleistocene revolution (MPR) at around 0.9 million years before present (900 kyr BP), glacial periods occurred every 41 kyr and were characterized by smaller ice volumes as after the MPR (Imbrie et al., 1993). After the mid-Brunhes event (MBE) around 430 kyr BP, the 100 kyr cycle dominated the glacial-interglacial cycles, the amplitude of glacial and interglacial states became larger and the ice volume during glacials increased (e.g., Berger and Wefer, 2003). Between the MPR and the MBE, glacial maxima were slightly less cold than after the MBE and interglacials were less warm than the last four interglacials, but lasted longer (see figure 1e).

Numerous researchers have attempted to simulate these glacial-interglacial cycles in order to understand what drives them and which feedbacks alter their properties. Milankovitch (1930) was the first to suggest that glacial-interglacial cycles are orbitally forced. He recognized that the seasonal and latitudinal distribution of energy received by the sun is modulated by oscillations of the earth's orbital parameters, particulary by the climatic precession (19 kyr and 23 kyr cycles), changes in the eccentricity of the earth's orbit (400 kyr and 100 kyr cycles) and the obliquity cycle (41 kyr) (see figure 1). The eccentricity of the earths orbit is thereby the only parameter that changes the globally and annually averaged solar radiation received by the earth, while the precession of the equinoxes and the obliquity change the seasonal and latitudinal variation of the insolation. However, the insolation change caused by the 100 kyr cycle is rather small and the reason for the dominance of this cycle over the 41 kyr cycle (which has a larger amplitude) during the last 420 kyr is still unknown. At present, it is believed that the orbital forcing is the main driver for the onset and termination of glaciations; however, glacial inceptions and terminations are probably altered by other forcings and feedbacks within the climate system.

In 1972, Kukla et al. (1972) summarized the results from a meeting with the title "When will the present interglacial end?". One of the results presented was that the present interglacial, when compared with the previous interglacials, should be in its final phase and its end should be expected soon, "possibly within the few next centuries (...) if man does not intervene." (Kukla et al., 1972). They reached this conclusion based on paleo-analogues because the last three interglacials lasted 10 kyr to 11 kyr years, while the present one already has lasted 11 kyr. Another prediction was that the Holocene interglacial would end abruptly with a jump into another state of the climate system, similar to how the last interglacial ended (Kukla and Matthews, 1972).

As deeper ice cores were drilled and temperature-proxy records that reached further and further back in time became available (e.g., EPICA community members, 2004) scientists noticed that the last three interglacials were different from the fourth interglacial before today's, the so called Marine Isotope Stage 11 (MIS-11) that occurred about 400 kyr BP. MIS-11 lasted longer than the following three interglacials (Oppo et al., 1998) and occurred during a time when

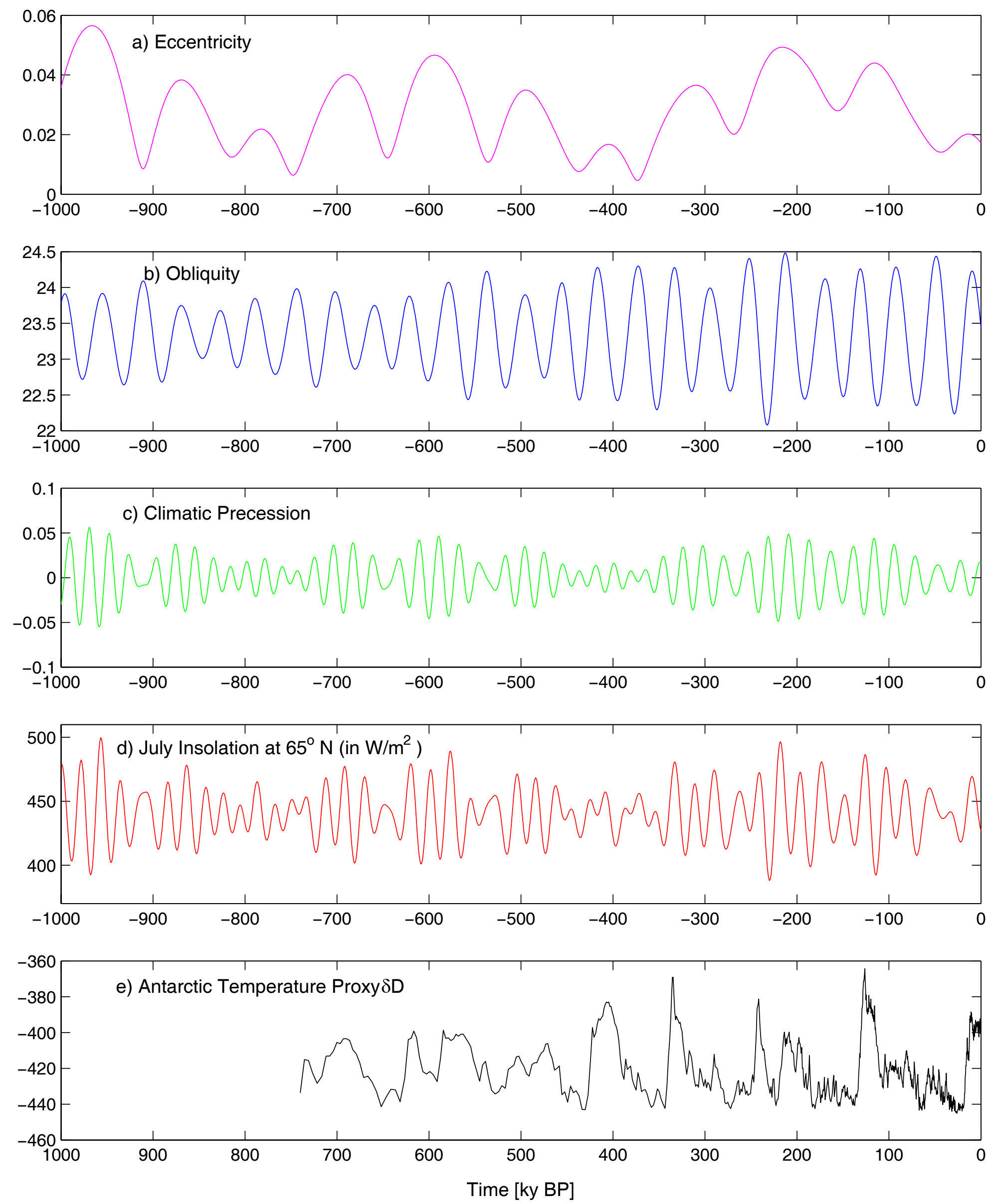

Figure 1: Shown are changes in the orbital parameters eccentricity (a), obliquity [in °] (b) and climatic precession (c). In (d) the resulting insolation in July at 65°N is shown [in W/m^2]. A proxy for Antarctic temperature, the deuterium/hydrogen ration in ice (δD), is plotted in (e) [in ‰], with colder temperatures corresponding to lower values of δD. The data for plots (a-d) are from Berger (1992), published in Berger and Loutre (1991). For graph (e) the data is from Jouzel et al. (2004), published in EPICA community members (2004).

the orbital eccentricity was at a minimum, similar to today (can be seen in figure 1a and 1e). Therefore, MIS-11 might be a better analogue for the present interglacial than the last three interglacials, as mentioned, for example, by Broecker (1998) and Claussen et al. (2005). Thus the Holocene does not need to end anytime soon but could last much longer than the last three interglacials.

Nevertheless, there is still no agreement about when and how the present interglacial will end, since the processes behind the dynamics of glacial-interglacial cycles remain unknown. Some researchers propose that the next glacial should already have started 5000 to 6000 years ago while others do not predict the next glacial before 50,000 or even 100,000 years after present (AP). Therefore the dynamics of glacial-interglacial cycles are still a controversial and therefore very interesting topic, and I will use section 2 of this report to summarize some of the most important theories, concepts and predictions about glacial-interglacial cycles and glacial inceptions and terminations. Section 3 will then give an overview of the results obtained with the McGill Paleoclimate Model (MPM) concerning the last and the next glacial inception and compare them with some of the results presented in section 3. Some ideas about future research topics in this area and the model improvements that should precede them will also be presented at the end of this section. Section 4 will be used to present conclusions and a summary.

2 Glacial–Interglacial Cycles, Inceptions and Terminations

2.1 Glacial–Interglacial Cycles

Glacial-interglacial cycles have been studied with many model types, including simple conceptual models, dynamic ice sheet models coupled to energy balance models (EBMs), and Earth System Models of Intermediate Complexity (EMICs). General Circulation Models (GCMs) cannot be used so far to simulate entire glacial-interglacial cycles due to their long computational time; this makes it impossible to run them over 100 kyr, which is the typical timescale of these cycles. Each of these model types has its advantages and disadvantages and all are useful tools to learn more about the physical principles behind glacial-interglacial cycles. Nevertheless it is important to note which results are obtained with which model type in order to obtain an idea of the limitations of the results presented.

Gallée et al. (1992) and Berger et al. (1998) showed that, provided the prescribed constant CO_2 concentration was below 220 ppm, the LLN-2D northern hemisphere climate model (a model of intermediate complexity) could reproduce the observed glacial-interglacial cycles when forced by changes in insolation. On the other hand, Loutre and Berger (2000a) found that under fixed orbital forcing the same model failed to reproduce the glacial-interglacial cycles when atmospheric CO_2 was varying according to Vostok reconstructions. In particular, they found that an astronomical forcing fixed at values typical for glacial times caused a climate cold enough for ice sheets to grow; however, CO_2 changes were then not sufficient to melt them again, leading to constant glacial conditions. Fixed interglacial astronomical forcing, on the other hand, created a climate that was too warm for ice sheets to form, leading to a constant interglacial. Therefore they concluded that the variations of the CO_2 concentration as reconstructed from Vostok cores were unable to drive the climate system into glacial-interglacial cycles when CO_2 was used as sole forcing. Their results confirmed the fundamental role of the orbital forcing for the glacial-interglacial cycles, showing that CO_2 forcing can alter the climate response but that it is not sufficient to drive the climate into glacial-interglacial cycles. However, in a modeling study of MIS-11, Berger and Loutre (2001) showed that during periods when the amplitude of insolation changes are too small to drive glacial-interglacial cycles, changes in CO_2 concentration are more important than during other times.

Paillard (1998) performed experiments with a simple conceptual model that was able to successfully simulate each glacial-interglacial cycle during the Pleistocene. This simple climate system model has three different equilibrium states: an interglacial mode, a mild glacial mode and a full glacial mode. The transition between the interglacial and the mild glacial occurred when insolation values crossed a threshold, while the system switched into the full glacial when the ice volume exceeded a threshold. Glacial terminations occurred when the insolation increased enough to cross a second threshold. He also found that MIS-11 was a particularly robust result, insensitive to changes in model parameters. In a simulation over the last 2 million years, Paillard (1998) was also able to correctly simulate the onset of the prominent 100 kyr cycle around 0.8 to 1 million years ago. That this simple three-stage model was able to successfully simulate the glacial-interglacial cycles during the Pleistocene led Paillard (1998) to the conclusion that the climate might be a three-stage system that is forced by insolation changes.

2.2 Glacial Inceptions

Glacial inceptions have also been investigated with climate models, using both time-slice experiments and transient simulations. Experiments were thereby mainly focused on the simulation of the most recent and the next glacial inception. All available model classes have been used in these simulations: GCMs, mainly for equilibrium experiments or short transient runs, and EMICs and conceptual models, for long transient as well as equilibrium experiments. There are problems, however, associated with equilibrium experiments of glacial inceptions: they happen when the climate is not in equilibrium but by definition in a very transient state (Kubatzki et al., 2005). Kubatzki et al. (2005) pointed out that equilibrium experiments might therefore simulate a permanent snow cover, but in transient runs the inception might occur much later or never.

2.2.1 The Last Glacial Inception

Pollard and Thompson (1997) used a high resolution, two-dimensional dynamic ice-sheet model to investigate the initiation of ice sheets during the last glacial inception. They forced the ice sheet model with the climate at 116 kyr BP, as simulated in equilibrium experiments with the GENESIS atmospheric GCM (AGCM), coupled to a slab ocean. In order to use the AGCM output to force their ice sheet model, they developed an elevation correction to downscale the AGCM results from a horizonal resolution of $3.75° \times 3.75°$ to a finer horizontal resolution of $0.5° \times 0.5°$ that is needed for dynamic ice sheet models. In addition, they introduced a meltwater refreezing correction into their model. Pollard and Thompson (1997) found large ice sheets over Alaska and western Canada that developed rapidly, and slower growing, smaller ice sheets over Baffin Island and the Canadian Archipelago. They argued that the simulation of rather small ice sheets over eastern Canada might have been caused by a warm bias in the GENESIS model and that the simulated large ice sheets over western North America could have been the result of an "inaccurate response of the GCM to the prescribed orbital changes" (Pollard and Thompson, 1997). In general they proposed that with the downscaling method they introduced, it would be possible to asynchronously couple GCMs to ice sheet models in the future. This would make it possible to account for the significant effect that the ice sheets have on the climate, an effect which has been neglected in their study since they only forced the ice sheet model with the AGCM results without feeding the simulated ice sheets back into the AGCM.

Kohdri et al. (2001) investigated the amplification of the orbital forcing by ocean feedbacks at 115 kyr BP in equilibrium simulations with an atmosphere-ocean GCM (AOGCM). Their experiment was the first that utilized a coupled atmosphere-ocean GCM to investigate the last glaciation; before only AGCMs with slab oceans or prescribed sea-surface temperatures (SSTs) were used for glacial inception studies. Kohdri et al. (2001) found that the North Atlantic deep water (NADW) was shallower and that the Atlantic meridional overturning circulation weakened due to the orbital forcing at 115 kyr BP, compared to the orbitally forced circulation we see today. This amplified the initial cooling of the high latitude ocean and the warming of the tropical ocean that was caused by the insolation forcing alone, leading to an even stronger pole-to-equator temperature gradient, and consequently an increased northward moisture transport. The colder temperatures in the northern latitudes together with the increased moisture transport to the polar regions were both favorable for the development of permanent snow cover in the Canadian Archipelago, and later in the simulation also over Norway and northeastern Eurasia. They therefore concluded that their results confirm "the considerable effects that modifications of ocean dynamics can have on the climate of Northern Hemisphere middle and high latitudes" (Kohdri et al., 2001).

Yoshimori et al. (2002) used an AGCM to investigate the role of SST and sea ice cover as well as the impact of vegetation changes on the dynamics of the last glacial inception. They prescribed the SST, sea ice cover and vegetation distribution obtained from a coupled climate system model for 116 ky BP as boundary conditions for the equilibrium simulations with the AGCM. A permanent snow cover developed in their simulation with present day SST and sea ice cover over the Canadian Archipelago. When they prescribed the colder sea surface temperatures and larger sea ice cover, as simulated for 116 ky BP by the coupled model, the area covered by permanent snow cover increased significantly. They therefore concluded that the formation of permanent snow cover (which they use as an indicator for possible ice sheet growth) in northern high latitudes of North America and Scandinavia was favored by colder than present-day SSTs and larger ice sheet cover. When they prescribed a vegetation distribution that was in equilibrium with the simulation for 116 ky BP with 116 ky BP SST and sea ice cover (in which tundra expands significantly in northern high latitudes), they found that the permanent snow cover increased over northern Canada and started to occur over Scandinavia. They therefore concluded that changes in vegetation should be taken into account, as noted earlier by Gallimore and Kutzbach (1996) and de Noblet et al. (1996). They closed with the statement that it would be necessary to couple an ice sheet model interactively to the AGCM in order to further investigate the glacial inception. In particular it would be important to investigate whether the permanent snow cover simulated in their study would turn into ice sheets, and whether the simulated ice sheet growth rate and ice volume would be in agreement with paleo-data.

Meissner et al. (2003) performed equilibrium experiments with the UVic Earth System Model, an earth system model of intermediate complexity which includes with an ocean GCM. They investigated the effect of the new land surface scheme and vegetation module on the simulation of the last glacial inception. To accomplish this task, they compared the results of equilibrium runs for 116 ky BP with and without the land surface scheme and the vegetation module with a control run for present day climate. They found a southward shift in the treeline in northern latitudes and a decrease in the amount of carbon stored in vegetation, due to the replacement of trees with grass. These vegetation changes caused an increase of perennial snow cover, which they used as an indicator for possible ice sheet growth since their model at that time was missing an ice sheet model. The vegetation changes also led to an increase in the strength of the thermohaline circulation (THC) by 3.8 Sv as compared to the present day simulation (and by 1.9 Sv as compared to the inception experiment without the land surface scheme and vegetation module). Furthermore, they found that in the tropics 88% of the broadleaf trees were replaced by shrubs in reaction to the reduced atmospheric CO_2 and the associated global cooling. Meissner et al. (2003) concluded that their results are in line with other studies (e.g., Yoshimori et al., 2002) that already showed the importance of vegetation feedbacks on glacial inception. They noted, however, that the importance of the vegetation feedbacks seems to be highly model dependent, as Brovkin et al. (2003) were able to show for the example of the two EMICs MoBiDic and CLIMBER-2: CLIMBER-2 showed a lower sensitivity to boreal deforestation than MoBiDic did.

Kageyama et al. (2004) investigated the last glacial inception and especially the importance of different ice sheet feedbacks during that time. They used CLIMBER-2 (an EMIC), coupled to the northern hemisphere (NH) ice sheet model GREMLINS, to perform transient experiments from 126 to 106 ky BP. In their simulation, two ice sheets appeared over the northwestern Rocky Mountains and the Canadian Archipelago at 121 ky BP. They reached their full areal extent at 113 ky BP and a thickness of 4000 m at the end of the run (106 ky BP), which is equivalent to a 17 m sea-level drop. The simulated total ice volume at 110 ky BP is similar to the ice volume found by Wang and Mysak (2002, discussed in section 3.1), but in contrast to Wang and Mysak (2002), Kageyama et al. (2004) found no ice sheet over Eurasia. Kageyama et al. (2004) showed that in their model the climatic difference between North America and Eurasia was caused by the initially warmer climate over Eurasia. This led to the presence of vegetation further from the taiga-tundra threshold, so that the orbitally induced cooling did not lead to a strong taiga-tundra feedback over Eurasia, which however occurred over North America. They also found that if vegetation was fixed at its initial interglacial state at the beginning of the run, no glacial inception occurred. Another result they found was that the summer (i.e., June, July and August) temperature is the limiting factor for the initiation of glaciation, as stated in the Milankovitch theory. Once the glaciation had started in their model, the ice extent feedback (i.e., the ice-albedo feedback) was the accelerating feedback in their model, while altitude and freshwater feedbacks were found to be unimportant. Overall their simulated ice volume is too small compared to paleo-data, a result which they blamed that on the rather small ice sheet growth in their model.

Vetoretti and Peltier (2004) performed transient sensitivity studies with the Canadian Climate Center Atmospheric GCM, coupled to a mixed layer ocean model, to investigate the role of orbital parameters and CO_2 forcing as drivers of glacial inceptions. They found in their transient experiments that the obliquity dominates the occurrence of the glacial inception in their model, and that the eccentricity-precession forcing is about as large as the effect of the CO_2 forcing at high latitudes. The eccentricity-precession forcing and the CO_2 forcing are both about half as large as the obliquity forcing. Therefore, eccentricity and CO_2 forcing combined also caused a glacial inception. They concluded that a glacial inception can be caused either by a strong obliquity forcing or by a eccentricity-precession forcing combined with a CO_2 forcing.

Kubatzki (2005, in preparation) also investigated the role of different orbital parameters on the last glacial inception. In transient experiments from 128 ky BP to 100 kyr BP with CLIMBER-2.3 (an EMIC), she found that changes in the perihelion (part of the precession signal) are necessary to initiate the growth of a small ice sheet in North America. However, to cause a full glacial inception, perihelion and obliquity changes had to be combined. Adding eccentricity changes on top of these two changes led to an additional increase in ice volume by 25%. However, perihelion and eccentricity forcing together were not sufficient to pass the threshold for glacial inceptions, leading only to an increase in ice volume by 10%. Furthermore, obliquity and eccentricity changes together and also individually were not sufficient to trigger any ice sheet growth.

Calov et al. (2005a) also used transient simulations with CLIMBER-2.3 to investigate the last glacial inception. They found that the rapid expansion of inland ice starts at 117 ky BP in their model and showed that the glacial inception represents a bifurcation transition in the climate system, caused by a strong snow-albedo feedback. In their model the transition into the last glacial is therefore a strong non-linear process; it occurs when summer insolation at northern high latitudes drops below a threshold value. Moreover they found that the positive snow-albedo feedback is the primary driver of the rapid climate transition towards the glacial. The ice-sheet-elevation feedback played only a secondary role in their model, in agreement with the results of Kageyama et al. (2004) discussed above. The snow-albedo feedback could only trigger the glaciation in their model if the snow cover and the ice dynamics were simulated with sufficiently high resolution. A resolution of 300 km in the horizontal, as used in most GCM's, did not lead to a glacial inception in their model, as it did not allow this feedback to work effectively enough.

In their companion paper, Calov et al. (2005b) included increased dust during glacial times. The dust was only allowed to influence the snow albedo (it acts as a negative feedback); its radiative effect in the atmosphere will be investigated in a future study. They found that with the increased dust, their simulated ice sheet cover is in general in better agreement with paleo-data. One example is northern Alaska, which had an ice cover in the simulations in Calov et al. (2005a), but was ice free (as paleo-data suggests) when increased dust was included. The glaciation in north-eastern Eurasia turned out to be highly sensitive to changes in the dust concentration. This was caused by the fact that the dust concentration in snow for a given dust deposition rate is inversely proportional to the snowfall. Since the snowfall is relatively low in north-eastern Eurasia, this region is very sensitive to changes in the dust deposition rate (Calov et al., 2005b)

In addition to the influence of dust, the effects of vegetation and the ocean circulation were also investigated. They found that including these features amplified the growth of ice sheets. They also confirmed the result of Loutre and

Berger (2000a), namely that changes in atmospheric CO_2 alone are not able to initiate a glaciation, but that CO_2 has nevertheless a strong amplifying effect on the ice sheet growth and the onset of the glaciation.

2.2.2 Next Glacial Inception

Saltzman et al. (1993) found in experiments with a dynamical system model that the climate system could be displaced from its oscillating mode with interglacials and glacials into a stable regime with lower ice masses if an anthropogenic increase in atmospheric CO_2 to values of over 350 ppm would be maintained for a long time. This would mean that the current period, in which we have seen repeated changes between glacials and interglacials, would end and we would enter a stable, non-oscillating climate state with constant interglacial conditions (Saltzman et al., 1993).

Ledley (1995) investigated the importance of summer solstice radiation and summer caloric half-year solar radiation (solar radiation averaged over the summer half of the year) in producing glacial-interglacial cycles. She found that the summer solstice solar radiation is more important as it is representative of the energy available for snow melt. Another conclusion was that it is unlikely that an ice age will begin in the next 70 kyr, as the summer solar radiation at 75°N will not decrease by more than $14\,\mathrm{Wm}^{-2}$ while most of the past major increases in ice volume involved decreases of 20 to $30\,\mathrm{Wm}^{-2}$; the radiation minima at 10,000 and 50,000 years in the future therefore will not cause a glaciation.

Broecker (1998) agreed with the prediction made by Kukla and Matthews (1972) that the Holocene will end abruptly. He argued that this is plausible because the climate has not responded significantly to the decrease in summer insolation that occurred during the last 10kyr; therefore, there will be no slow shift into a glacial. Broecker (1998) also mentioned that the present interglacial seems to be in a stable state; however, as soon as a threshold is crossed, the climate system will jump into another stable state, which would be the end of the Holocene. Broecker (1998) pointed out that the MIS-11 is different from the last three interglacials and that therefore a simple paleo-analogue does not give a definite answer to the question when the Holocene will end. Therefore, he focused on the mechanisms which ended previous interglacials. For the end of the Eemian he found that the end of the interglacial was preceded by a growth of ice caps and an associated sea level drop. Hence he argued that if the present interglacial ends similarly, its end is at least several thousand years away as the sea level has not yet started to decrease. He also pointed to the reorganization of the ocean circulation as an important mechanism in the glacial-interglacial cycles. Hence he argued that anthropogenic warming could trigger a change in the ocean circulation that would lead to a cooling. On the other hand, he pointed out that this could only happen after a significant anthropogenic warming occurred, so that in the end the climate might be just the same as today.

Loutre and Berger (2000b) investigated the future climate and its sensitivity to different CO_2 scenarios in transient experiments with the LLN 2-D NH model, a climate model of intermediate complexity. The found that the climate is likely to experience a long lasting interglacial ($\sim$50 kyr) and that a small glacial maximum will occur at 60 kyr AP with a larger one at 100 kyr AP. They also found that the CO_2 scenarios they used had an effect on the timing and the amplitude of the glaciation. In particular, they found that during times when insolation variations are small, as was the case during MIS-11 and today, the simulated ice volume depends strongly on CO_2 concentration. In their sensitivity studies they found that the ice volume simulated is influenced significantly by the present-day state of the Greenland ice sheet. For an anthropogenic scenario where the CO_2 concentrations reached 750 ppm over the next 200 years and decayed afterwards, the Greenland ice sheet melted almost completely between 10 and 14 kyr AP and did not reach the volume simulated under the natural CO_2 scenario until 50 kyr AP. Therefore they concluded that mankind could perturb the climate for up to 50 kyr into the future. They admitted, however, that the Greenland ice sheet melts rather easily in their model when compared to other models.

Ruddiman (2003) proposed that early anthropogenic warming suppressed a glaciation 5000 to 6000 years ago. He argued that the anthropogenic era started already 8000–5000 years ago, and not 150 to 200 years before present as generally believed. Ruddiman (2003) based this theory on the fact that the CO_2 and CH_4 levels decreased during the last three interglacials, and so, by paleo-analogue, they should have decreased during the Holocene as well. However, the CO_2 began to increase 8000 years ago and the CH_4 started to increase 5000 years ago. Ruddiman stated that these increases were caused by human activities, specifically by the start of forest clearing in Europe 8000 years ago and the beginning of rice irrigation 5000 years ago. He calculated that these early land cover changes caused a CH_4 anomaly of 250 ppb and a CO_2 anomaly of 40 ppm. These numbers are anomalies relative to the typical interglacial values Ruddiman inferred from the previous three interglacials, but not the anomalies that occurred relative to the Holocene CO_2 and CH_4 values previous to 5000 and 8000 years BP; the latter anomalies are much smaller as the ones quoted by Ruddiman (2003). He calculated the warming caused by these early gas emissions to be 0.8 °C globally and 2 °C in high latitudes. Furthermore, he argued, based on the results of two climate models, that this warming was large enough to stop a natural glaciation of northeastern Canada 5000 to 6000 years ago.

Claussen et al. (2005), in a reply to Ruddiman's (2003) hypothesis, argued that Ruddiman did not take into account biogeophysical effects of land cover changes, which, especially in snow covered regions, could compensate or even overcompensate the global warming due to emission of CO_2 and CH_4. As a consequence, the warming due to the early land cover changes could probably not have stopped a naturally occurring glaciation in northern high latitudes of Canada, as suggested by Ruddiman (2003). They also pointed out that during the first 10 kyr years of MIS-11, CO_2 concentrations changed little or even showed an increase, so that the increase of 20 ppm seen in the Holocene CO_2 level is not unique in climate history. They also questioned Ruddiman's estimate of the CO_2 emission due to past land use, as even today the uncertainty in the estimate of CO_2 emissions associated with land use changes is very high. For the past, where not even the exact extent of deforestation is known, the uncertainty of the CO_2 change would be even higher. To test how their model, CLIMBER-2, reacts to the Ruddiman-proposed natural development (i.e., decrease) of CO_2 during the last 10 kyr, Claussen et al. (2005) ran their model with a 40 ppm reduction of CO_2 and a 250 ppb reduction of CH_4 values in the atmosphere. They found a cooling in the high northern latitudes of 1–1.5 °C for the pre-industrial climate, but not the 2 °C Ruddiman calculated. They also tried to simulate a Holocene glaciation as suggested by Ruddiman. For that purpose they ran the model from 10,000 years BP to 3000 years into the future, prescribing a CO_2 decrease from 265 ppm 10,000 years ago to 240 ppm today to 230 ppm at 3000 years into the future. In this experiment they found some increase in ice cover, but no glacial inception was simulated, in contrast to the end of the last interglacial, which they were able to simulate successfully. They explained the difference between the end of the last interglacial and the present interglacial by the much smaller variations in insolation during the Holocene compared to the Eemian. Therefore, they concluded that the threshold for a glacial inception in the Holocene has not been crossed yet. This stayed true even when it was assumed that the Holocene CO_2 evolution was similar to the CO_2 evolution observed during the previous three interglacials. Hence, they suspect that the threshold will probably not even be crossed during the next 50,000 years, as the insolation minima until then are all going to be much smaller that the insolation minima that ended the Eemian. They further concluded that paleo-analogues are probably not the right tool to predict climate changes, but that an earth system approach is more likely to help us understand climate dynamics. Claussen at al. (2005) also stated that they believe that the climate system would respond to insolation changes in a strongly nonlinear manner, which is in agreement with others (e.g., Broecker, 1998; Paillard, 1998, 2001).

2.3 Glacial Terminations

Yoshimori et al. (2001) investigated the sensitivity of the last glacial termination to orbital and CO_2 forcing. They used a predecessor of the UVic Model that included atmosphere, ocean and sea ice components, asynchronously coupled to a dynamic ice sheet model. In their experiments they used orbital parameters for 21 kyr and 11 kyr BP as well as CO_2 concentrations of 200 and 280 ppm to investigate the individual contribution of both orbital and CO_2 forcing. In addition they performed a control run for the present-day climate to evaluate the performance of their model. They found that both orbital and CO_2 forcing was necessary to obtain a full deglaciation. However, orbital forcing was found to be more important during deglaciation, due to its maximal effect on the temperature in summer while the CO_2 forcing had its maximum impact during winter. Since it is known that summer temperatures are more important for ice sheet development or degeneration than winter temperatures (Milankovitch 1930), it is consistent with Milankovitch theory that the orbital forcing is the main driver of deglaciations. Nevertheless CO_2 was found to cause a strong feedback that helped to end the last glacial. How the increase in CO_2 that preceded the last deglaciation by about 2 kyr was caused, however, remains an open question Yoshimori et al. (2001) did not study (for some hypotheses about the cause of the CO_2 variations see section 2.4.2)

Petit et al. (1999) suggested, based on Vostok data, that during the last four glacial terminations that were captured in the Vostok core (which reaches back 420 kyr), the same sequence of events led to glacial terminations: First orbital forcing, which was then amplified by the increases of greenhouse gases in the atmosphere. The decrease in ice sheets caused by them was then further enhanced by the ice-albedo feedback until the end of the deglaciation was reached. In respect to the phase relationship between CO_2 increase and warming in Antarctica, Petit et al. (1999) arrived at the conclusion that the uncertainty in the dating of ice-age relative to the gas-age is with about 1000 years too large to determine the sign of the relationship between CO_2 and temperature, which was estimated to be 600±400 years by Fischer at al. (1999). Petit at al. (1999) further concluded that, based on the Vostok record, Antarctic temperature and atmospheric CO_2 lead global ice volume and Greenland temperature during terminations.

2.4 Changes in Greenhouse Gas Concentrations

2.4.1 Reconstructed Changes in Greenhouse Gas Concentrations

Greenhouse gases are thought to constitute an important feedback to the orbitally forced glacial-interglacials cycles (e.g., Raynaud et al., 1993; Lorius et al., 1990). Reconstructions of atmospheric concentration of CO_2 and CH_4 from ice core records like the Vostok core revealed that their atmospheric concentration is smaller during glacials than during interglacials (Barnola et al., 1987; Petit et al., 1999). Petit et al. (1999) found in the Vostok greenhouse gas record that

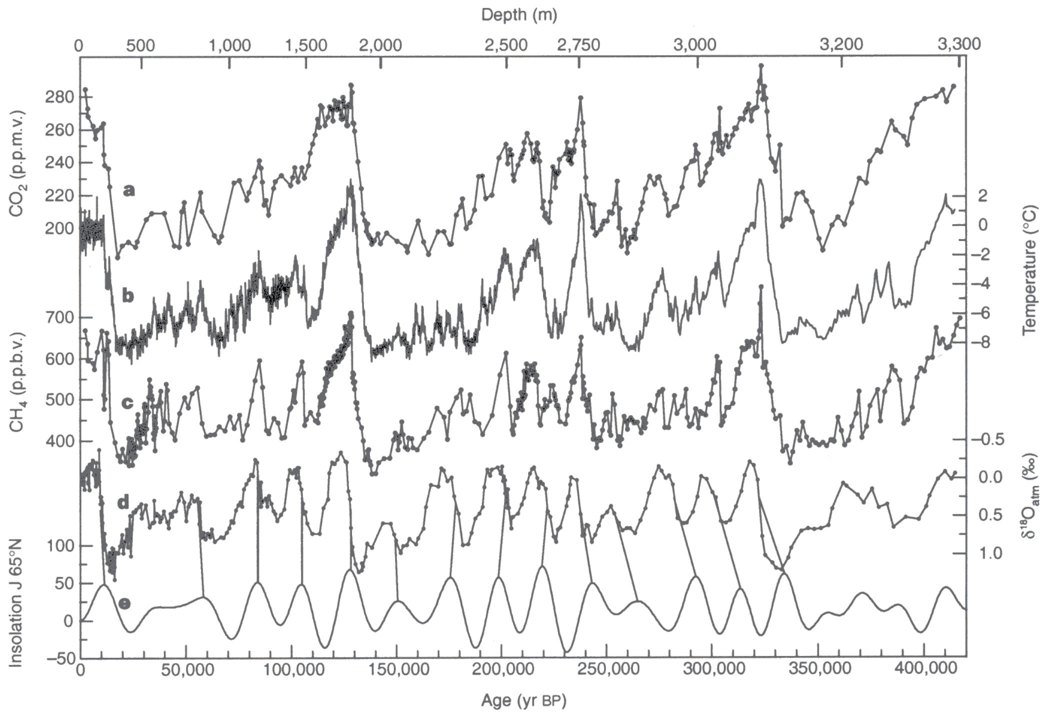

Figure 2: Time series from the Vostok ice core of (a) atmospheric CO_2, (b) a temperature proxy showing changes in Antarctic air temperature, (c) atmospheric CH_4, (d) $\delta^{18}O_{atm}$ (a proxy for global ice volume and the hydrological cycle) and (e) mid-June insolation at 65°N [in W m^{-2}]. Figure taken from Petit et al. (1999).

CO_2 varies between 180 ppm during glacials and 280–300 ppm during interglacials, while CH_4 varies between 320–350 ppb and 650–770 ppb (see figure 2).

During glacial terminations, increases in CH_4 vary between the four cycles covered by the Vostok record. For CO_2 the differences between individual terminations are much smaller as for CH_4. CH_4 thereby shows a two-step increase, with a slow increase first, followed by a rapid jump to interglacial values. Petit et al. (1999) also found that the Antarctic temperature (determined from proxies), CH_4 and CO_2 increased in phase during terminations, at least as far as it was possible to determine their timing, due to uncertainties in the gas-age/ice-age differences. These uncertainties are on the order of 1000 years, which is the same order as proposed lags between temperature and CO_2 of 600±400 years (Fischer et al., 1999). More recent results from Dome C showed, however, that at termination V (around 430 kyr BP) CH_4 lagged the Antarctic temperature and CO_2 increase by 4–5 kyr, and also that the jump in the CH_4 increase happened when CO_2 had almost reached its maximum (EPICA community members, 2004). The EPICA community members therefore concluded "that the pattern of climate before MIS 11 was different to that which has followed for the past four glacial cycles" (EPICA community members, 2004).

The NH warming and global ice volume were found to lag behind the southern hemisphere (SH) warming by several thousand years (Petit et al., 1999), but the magnitude of the lag varies between 4 and 9 kyr for different deglaciations (can be seen in figure 2). Alley et al. (2002), however, argued against a lead of SH warming over NH warming. They analyzed the timing of extrema in records from Greenland and Antarctica instead of the relative phasing of climatic variables. In their analysis they found that extrema of "northern high latitude insolation and northern temperature were nearly synchronous, southern temperature and CO_2 were nearly synchronous with each other and lagged northern insolation by approximately 2 ka or more,...". This is contradictory to the results of Petit et al. (1999) and others (e.g., Fischer et al., 1999; Monnin et al., 2001), so that better records are needed in order to determine if there was a southern or northern lead.

It is apparent from figure 2 that the increase to interglacial CO_2 levels is much faster than the decrease to glacial CO_2 levels, as was pointed out by Petit et al. (1999). During glacial inceptions, Petit et al. (1999) found an in-phase decrease of Antarctic temperature and the CH_4 in the atmosphere, while the CO_2 stayed at high interglacial levels for several more thousand years before it also decreased (see figure 2).

Overall, a correlation of r^2=0.71 and r^2=0.73 between the CO_2 and CH_4 records, respectively, and the Antarctic temperature was found (e.g., Petit et al., 1999; Barnola et al., 1987; Chappellaz et al., 1990; Raynaud et al., 1993). This high correlation suggests that CH_4 and CO_2 changes have amplified the orbital forcing of glacial-interglacial cycles, probably along with other feedbacks (Raynaud et al., 1993; Lorius et al., 1990).

2.4.2 Causes of Glacial-Interglacial Greenhouse CO_2 Variations

As described in the previous section, the shape of the CO_2 variations over the glacial-interglacial cycles is reasonably well know from ice core data. Furthermore, their high correlation with the proxy for Antarctic temperature suggests that CO_2 has contributed significantly to the glacial-interglacial climate change (Petit et al., 1999). The cause of the observed changes in CO_2 is still unknown, even though it has been an active area of research for the past 20 years. There are, however, many hypotheses that try to explain part or all of the observed changes.

Petit et al. (1999), for example suggested that the Southern Ocean around Antarctica plays a key role in the long term CO_2 changes, via changes of the sea-ice extend and the intensity of the deep ocean circulation. Stephens and

Keeling (2002) argued that an extensive (99%) sea-ice cover of the Southern Ocean south of 55°S during glacial times would trap 65 ppm of CO_2 in the ocean, which are about 80% of the glacial-interglacial CO_2 change of 80–100 ppm. The sea-ice cover would thereby decrease the atmospheric CO_2 level by limiting the outgassing from the ocean. Stephens and Keeling (2002) found in their study that the sea-ice only significantly reduces the outgassing of CO_2 when the sea-ice cover south of the Antarctic Polar Front is larger than 95% during winter. For smaller sea-ice coverage, they found that a negative feedback counteracts the decrease of atmospheric CO_2 due to limited outgassing, since the partial pressure gradient increases when atmospheric CO_2 decreases and oceanic CO_2 increases. This leads to an intensified CO_2 flux over the open water areas, so that only for a very large sea-ice coverage of 95% or more the decrease in atmospheric CO_2 due to the decrease in outgassing is large enough, compared to the increase due to the stronger CO_2 flux from the ocean, to cause a significant reduction of atmospheric CO_2. Maqueda and Rahmstorf (2002) found that during the last glacial maximum (LGM), their coupled upper-ocean–sea-ice model only simulated a maximum sea ice coverage of 92%. This corresponds to a CO_2 decrease of only 35 ppm. They therefore concluded that the increase of sea-ice in the Southern Ocean could explain only 15–50% of the CO_2 decrease during glacials, and not 80% as suggested by Stephens and Keeling (2002). Nevertheless, this means that the trapping of CO_2 in the ocean by sea-ice could be one of the important mechanisms which caused the glacial-interglacial CO_2 variations (Maqueda and Rahmstorf, 2002).

Changes in the biosphere were found to work in the opposite direction of the observed glacial-interglacial CO_2 variations, since the terrestrial carbon storage was smaller during glacials than during interglacials (Crowley, 1995). It was estimated by Bird et al. (1994) that during the last glacial the terrestrial carbon storage was reduced by 300–700 Pg C (10^{15} g carbon), which means that atmospheric CO_2 would have been about 45 ppm higher. However, this increase in inorganic carbon would have caused an increase in dissolution of deep-sea sedimentary calcite, so that only an effective CO_2 increase of 15 ppm would have occurred (Sigman and Boyle, 2000). Hence, the biosphere was a source for CO_2 during glacials, and not a sink.

The colder ocean temperatures during glacials increased the solubility of CO_2 in the ocean, while the higher salinity of the ocean at the same time decreased the CO_2 solubility (Weiss 1974). Overall it was estimated that the temperature decrease caused a decrease of atmospheric CO_2 by 30 ppm while the increase in salinity caused an increase in atmospheric CO_2 by 10 ppm (Broecker and Peng, 1998; Sigman and Boyle, 2000). This means that the temperature and salinity changes in the ocean together caused a decrease in atmospheric CO_2 concentration by 20 ppm during glacials. This effect can therefore also not explain the total interglacial-glacial CO_2 variations.

Another important process that could explain the CO_2 decrease during glacials is the increase in the intensity of the biological carbon pump of the ocean during glacials, as first suggested by Broecker (1982). The intensity of the biological carbon pump can be increased by different processes in different regions. In the low-latitude oceans the intensity of the biological carbon pump can increase when nutrient availability is increased, since the amount of CO_2 that can be extracted from the atmosphere by biological production in the low latitude oceans is limited by the supply of the nutrients nitrate and phosphate. An increase in these nutrients by about 50% compared to present day values could cause a 80 ppm decrease in atmospheric CO_2 (Sigman et al., 1998). Studies indicate that the denitrification of the water column was reduced in low latitudes during the last glacial (Ganeshram et al., 1995; Altabet et al., 1995), and it was also suggested that N_2 fixation rates were higher during the last glacial due to the larger supply of dust containing iron (Falkowski 1997). Together these two processes would have increased the nitrate concentration in the ocean during the last glacial. However, traditionally it is believed that phosphate is the limiting nutrient on glacial-interglacial timescales (Broecker and Peng, 1982). If the increase in nitrate concentration in the low latitude surface ocean is really driving the glacial-interglacial CO_2 variations, as suggested by Falkowski (1997) and others, the marine biota must be able to deviate from the so-called Redfield ratio[1], so that the biota can use the excess nitrate instead of the missing phosphate (Sigman and Boyle, 2000)

In the polar oceans the productivity of the biological pump can be increased by increased nutrient utilization (Knox and McElroy, 1984). Today CO_2 is transferred from the ocean to the atmosphere in the regions of upwelling in the polar

[1]The Redfield ratio describes the ratio in which phosphate, nitrate and inorganic carbon are being incorporated into biomass during biological production (Redfield et al., 1963).

Southern Ocean, due to incomplete nutrient utilization so that CO_2 can escape into the atmosphere. If the available nutrients would be used more complete, less CO_2 would be transferred to the atmosphere, as it would be fixed by increased biological productivity (Sigman and Boyle, 2000). Two possible causes have been proposed that could have increased nutrient utilization during the last glacial: an increase in biological production (due to the increase of air-blown dust containing iron, which is rare in the polar regions and limits biological productivity there (Martin, 1990) or a decrease in the upwelling of deep waters at the surface, so that less CO_2 would be transported to the surface (Francois et al., 1997). Isotope data in the Antarctic region suggests that the nitrate utilization rate was indeed twice as large as today (Francois et al., 1997), which would be enough to lower the atmospheric CO_2 concentration by the full glacial-interglacial value (Sigman et al., 1999). Francois et al. (1997) suggested that the increased stratification (and therefore decreased upwelling) in the polar Southern Ocean during the last glacial were the real drivers behind the interglacial-glacial cycles in CO_2, since proxy data show that export production in the oceans around Antarctica was lower during the last glacial. This implies that nitrate utilization at the surface only increased because the nitrate supply from the deep ocean was decreased due to stronger stratification (Sigman and Boyle, 2000).

Which process or which combination of processes really drive the glacial-interglacial CO_2 variations remains an open question that has to be answered in the next years. From modeling studies it appears that the CO_2 variations have an important amplifying effect on the glacial-interglacial cycles (e.g., Yoshimori et al., 2001; Weaver et al.,1998), so that an understanding of their cause is crucial in order to understand the influence of the anthropogenic CO_2 emissions on our climate in the future.

3 Glacial-Inceptions in the MPM

The McGill Paleoclimate Model (MPM)[2] is an EMIC, which has been used to successfully simulate different climate states, and also the last and the next glacial inceptions. The results of these experiments are presented on the following pages and some differences and agreements with the results presented in section 2 are also mentioned.

3.1 Modeling of the Last Glacial Inception

Wang and Mysak (2002) simulated the last glacial inception with the MPM. They performed seven transient simulations with different setups from 122–110 ky BP to investigate the effect of a parametrization for the freezing of rain/refreezing of meltwater, the elevation effect of orography and the use of an interactive ocean component versus a fixed SST. All experiments were forced with varying orbital forcing and CO_2 concentrations from Vostok reconstructions. They could show that Milakovitch forcing alone was sufficient to initiate the formation of ice sheets over Eurasia and North America around 120 kyr BP. However, to simulate the rapid ice sheet growth in the 10 kyr afterwards they needed to include the

[2]Detailed descriptions can be found in Wang and Mysak (2002) and the references therein.

parametrization of the freezing of rain/refreezing of meltwater, the elevation effect and the active ocean model. Without both the elevation effect and the refreezing parametrization their simulated ice volume was only 8% of the ice volume found in the full simulation, while neglecting either one led to an ice volume that was one third of the volume found in the full simulation. In simulations that showed rapid ice sheet growth, their THC was intensified, due to the cooling in the high latitudes and the decreased freshwater flux into the ocean. The stronger THC led to an increased land-sea thermal contrast at northern high latitudes, causing increased northward moisture transport, which is favorable for ice sheet growth and, therefore, worked as a positive feedback. The ice sheets simulated in the experiments of Wang and Mysak (2002) reached about two thirds of the ice volume inferred from paleo-data. Their ice sheets were located over Alaska, the northern Laurentide, Scandinavia and northern Siberia (see figure 3). Wang and Mysak (2002) concluded that the elevation feedback as well as the freezing/refreezing parametrization and the positive feedback with the THC are important for the correct simulation of the rapid ice sheet growth after the initial ice sheets started to appear. That they found the elevation feedback to be important stands in contrast to results of Kageyama et al. (2004) and Calov et al. (2005b), who both found the elevation effect to only have a small influence. In their simulations, the ice-albedo feedback was the dominant feedback.

After the results of Wang and Mysak (2002) were published, the MPM was improved significantly. Wang, Z. et al. (2004) introduced a new solar energy disposition scheme and Wang, Y. et al. (2005b) added an interactive vegetation

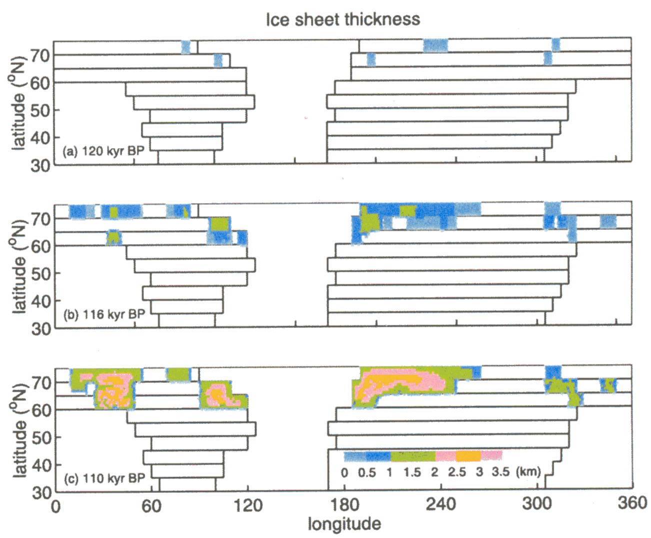

Figure 3: Ice sheet thickness and distribution over North America and Eurasia for the fully coupled run of Wang and Mysak (2002) at (a) 120 kyr BP, (b) 116 kyr BP and (c) 110 kyr BP. Graph taken from Wang and Mysak (2002).

component to the MPM that includes one of the biogeophysical climate vegetation feedbacks (i.e., the taiga-tundra feedback). Wang, Z. et al. (2005) then used this so called "green" MPM to repeat the simulation of the last glacial inception performed by Wang and Mysak (2002). They ran the model from 122 to 80 kyr BP, again forced by varying solar insolation and atmospheric CO_2 concentrations derived from Vostok ice core data. The glacial inception was simulated around 19 kyr BP, as compared to 20 kyr BP in Wang and Mysak (2002). The total ice volume simulated was 10.3×10^6 km^3 (see figure 5), which is less as found by Wang and Mysak (2002) (13×10^6 km^3). This was mainly due to a smaller Eurasian ice sheet in the work of Wang, Z. et al. (2005) (see figure 4), which was caused by the new interactive vegetation component that simulated forest cover

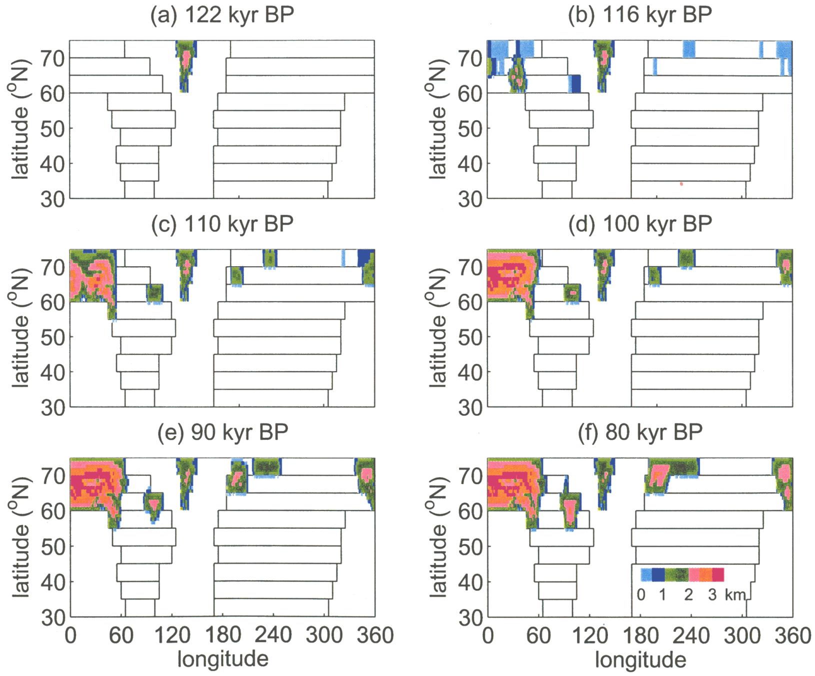

Figure 4: Ice sheet thickness and distribution over North America and Eurasia for the fully coupled run (CON) of Wang et al. (2005) at (a) 122 kyr BP, (b) 116 kyr BP, (c) 110 kyr BP, (d) 100 kyr BP, (e) 90 kyr BP and (f) 80 kyr BP. Graph taken from Wang, Z. et al. (2005).

over northern Europe. The forest cover warmed this area over the taiga-tundra effect, so that ice buildup was suppressed or limited. In contrast to Kageyama et al. (2004) (discussed earlier), who found no ice cover over Eurasia, Wang, Z. et al. (2005) found ice sheets over Eurasia; however they are much smaller than those in Wang and Mysak (2002), which is in better agreement with paleo-data. When they fixed the vegetation at interglacial conditions they found that Eurasia was ice free and that the North American ice volume was only 64% for the ice volume simulated with interactive vegetation. They therefore concluded that the interactive vegetation-albedo effect in high northern latitudes was crucial to simulate the Eurasian ice sheets and that it had a strong amplifying effect on the North American ice growth. This is in overall agreement with Kageyama et al. (2004), who also argued that the vegetation played a crucial role during the glacial inception. However, Kageyama et al. (2004) found that the effect of a forest cover in Europe prevented ice sheet growth over Eurasia, instead of leading to it as found by Wang, Z. et al. (2005). The use of a fixed interglacial vegetation cover led to no ice sheet growth anywhere in the study of Kageyama et al. (2004). In Wang, Z. et al. (2005), the fixed interglacial vegetation prevents ice sheet development over Eurasia, and limits ice volume increase over North America. This leads to the conclusion that the climate is closer to the glaciation threshold in Wang, Z. et al. (2005) than in Kageyama et al. (2004), as without the additional cooling caused by the vegetation, no glaciation takes place in the latter while Wang, Z. et al. (2005) see a glaciation over North America even without the vegetation feedback. In contrast to both of these studies, Calov et al. (2005b) (discussed earlier) found that the vegetation component was not a primary mechanism for the inception.

The total ice volume at 110 kyr BP simulated by Wang, Z. et al. (2005) was even smaller than that found by Wang and Mysak (2002), and hence was also too small compared to paleo-data. However, after 95 kyr BP the simulated ice volume is comparable to estimates from paleo-data. One aspect mentioned by Cochelin (2004) (who performed part of the experiments leading to the paper of Wang, Z. et al. (2005) and analyzed them in her MSc thesis) that could contribute to the too small ice volume before 95 kyr BP is that the model only reached

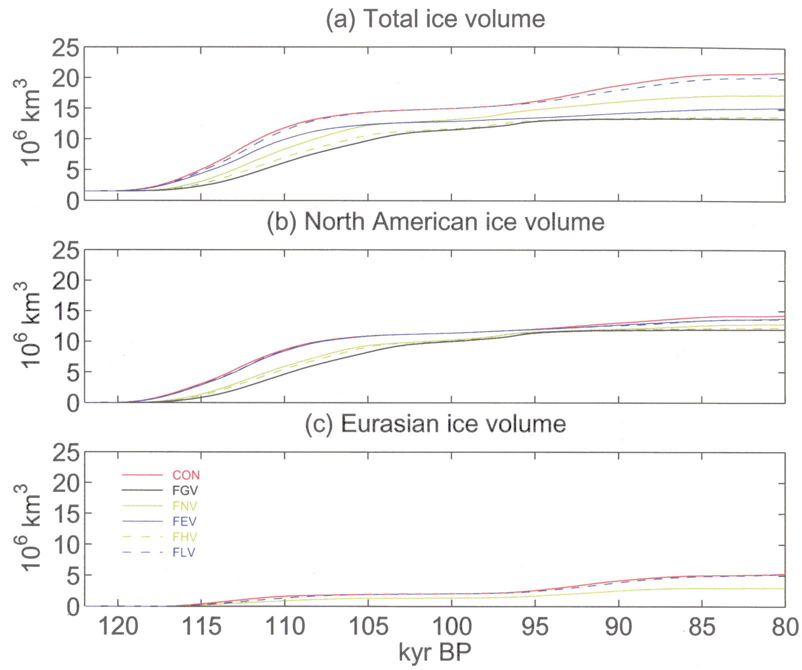

Figure 5: (a) Total ice volume simulated by the "green" MPM between 122 and 80 kyr BP. (b) Ice volume over North America. (c) Ice volume over Eurasia. In experiment "CON", vegetation is interactive, in "FGV" global vegetation is fixed, in "FNV" vegetation is fixed over Eurasia, in "FHV" Vegetation is fixed over 60-75°N, and in "FLV" vegetation is fixed over 30-60°N. Figure taken from Wang, Z. et al. (2005).

up to 75 °N, so possible ice sheets north of 75 °N (i.e., on Ellesmere Island and northern Greenland) were not included in the simulated ice volume. The ice sheets that formed in the "green" MPM were located over the Laurentide, Cordilleran, Scandinavia and Siberian regions and over Alaska (see figure 4). The ice sheets over Alaska are problematic since many scientists believe that Alaska was ice free during the last glacial; however, paleo-data can not rule out that ice sheets existed over Alaska (e.g., Muhs et al., 2003). Cochelin (2004) argued that the ice sheet over Alaska might be a result of a cold bias in the MPM over Alaska due to the downscaling scheme and also due to the spatial averaging used in the MPM, which can not resolve the deep valleys and fjords of Alaska. Overall Cochelin (2004) concluded that the MPM could successfully simulate the last glacial inception, even though the simulated ice volume was too small

3.2 Modeling of the Next Glacial Inception

Cochelin et al. (2005) ran the MPM into the future using different CO_2 scenarios. They investigated the climatic response of the MPM to a doubling of atmospheric CO_2 concentrations, followed by a constant plateau phase, as well as the climate resulting from different constant long term CO_2 concentrations. In the "doubling of CO_2" experiment, the CO_2 level increased linearly from 280 to 560 ppm in 70 years and then stabilized at 560 ppm. They found a warming of 1.8 °C during the first 70 years and an final warming of 3.1 °C after the CO_2 had stabilized. The THC decreased substantially during the first 100 years of the simulation. Its strength increased afterwards again but stayed 2.6 Sv below the initial value.

In the 100-kyr long future simulations, two different sets of scenarios were used: Various constant CO_2 concentrations were used (240, 259, 260, 270, 280, 290 and 300 ppm) as well as scenarios with a rapid increase and then a slow decrease to stable CO_2 levels during the first 1000 years, followed by constant atmospheric CO_2 concentrations for the remaining 99 kyr. In the case of constant atmospheric CO_2 levels, they found two thresholds in the atmospheric CO_2 level for the next glacial inception. For CO_2 levels of less than 270 ppm an instantaneous glacial inception occurred within the next 1000 years. For CO_2 concentrations between 280 and 290 ppm, the glacial inception in the MPM was simulated to start at 50 kyr AP, while for CO_2 levels of 300 ppm or more no glacial inception occurred within the next 100 kyr. The also found that the Laurentide ice sheet appeared later and that the total ice volume was smaller the higher the CO_2 concentration was, which is in agreement with results of Loutre and Berger (2000b) (presented earlier). In the simulations with an anthropogenic warming scenario in the first 1000 years and a stabilization at 280, 290 or 300 ppm afterwards, Cochelin et al. (2005) found that for CO_2 levels of 290 ppm or less a glacial inception at 50 kyr occurred, while for 300 ppm or more no inception within the next 100 kyr was simulated. This means that the anthropogenic warming scenario at the beginning did not influence the limit for which no glacial inception occurs. The warming scenario only influenced the climate at the beginning of the simulation, but after about 10 kyr the simulated climate was the same as in the simulations without the warming episode. Cochelin et al. (2005) therefore concluded that the climate system will not be perturbed for longer than 10 kyr at the most by an anthropogenic warming within the next thousand years, in contrast to Loutre and Berger (2000b) (presented earlier), who found that mankind could perturb the climate for up to 50 kyr due to a melting of the Greenland ice sheet in their model. Cochelin et al. (2005) further concluded that the timing of the next glacial inception depends on whether the CO_2 concentration in the future is below or above the two CO_2 thresholds they found. What determines these thresholds remains an open question and more research is necessary to investigate this topic;

it is clear, however, that the start of the glacial inception is governed by a non-linear process. The biogeophysical vegetation feedback that was included in the model did not act as a driver of the glacial inception; it did however reinforce the cooling of the climate and the buildup of ice sheets once they had started. The same was found to be true for the THC.

The simulation of an imminent glaciation for CO_2 values of 270 ppm or less by Cochelin et al. (2005) differs from the results of Claussen et al. (2005) (presented earlier), who did not find an imminent inception in the next 3000 years for CO_2 values of less than 270 ppm (240 ppm for today to 230 ppm at 3000 years AP).

3.3 Future Modeling Approach

3.3.1 Model Improvements

To improve the simulation of glacial inceptions in the MPM, some aspects of the model could be improved.

1. The atmospheric CO_2 is used as an external forcing in the simulations so far, but in reality, in so far as natural CO_2 changes are concerned, it is part of the climate system, influenced by changes in other components of the climate system like the ocean and the biosphere. In order to simulate and not prescribe the atmospheric CO_2 concentration, the release and uptake of carbon by vegetation and ocean has to be simulated. To be able to do that, the carbon cycle in the MPM still has to be improved. A. Antico is working on including the ocean carbon cycle while Wang, Y. et al. (2005b) included

the carbon release and uptake by the terrestrial vegetation. When this work is finished, it will be possible to simulate the natural variations of atmospheric CO_2 concentrations within the MPM. This will make it possible to run idealized (i.e., non-anthropogenic) simulations for the future to simulate the "natural" development of the climate as well as paleo-simulations, without artificially prescribing CO_2 concentrations.

To take into account the anthropogenic CO_2 release into the atmosphere, different scenarios of anthropogenic CO_2 emissions will need to be prescribed. In difference to the present model, these scenarios would only prescribe the emissions caused by humans, not the part that stays in the atmosphere, since the MPM would be able to simulate the CO_2 uptake by the ocean and biosphere and the amount of CO_2 resting in the atmosphere interactively when the carbon-cycle is included in the MPM. This is very important, as carbon uptake and release by the terrestrial biosphere and the ocean are likely to change during a changed climate, so that calculations of the amount of CO_2 resting in the atmosphere based on present-day climate might be under or over predicting the CO_2 uptake and release.

For paleo-simulations the interactive simulation of atmospheric CO_2 levels is also very valuable, as it will allow the atmospheric CO_2 content to respond to changes in the other components of the climate system. This will enable us to better understand why the CO_2 values were the way they are recorded in ice cores, which will finally lead to a better understanding of the climate system itself.

2. The ice sheet model used in the MPM is the 2-D dynamic ice sheet model of Marshall and Clarke (1997). In the MPM the thermodynamic component is missing; the ice sheet model is therefore only an isothermal ice sheet model. In order to simulate the full glacial cycle which includes D/O events and HE's and also to better simulate glacial inceptions, the thermodynamic component has to be included. This will be another step towards a fully integrated climate system model, which will in the end hopefully be able to simulate the past climate development correctly, forced only by insolation changes. This will then enable us to have increased confidence in simulations of the future climate.

3. The downscaling scheme used to downscale the sectionally averaged surface air temperature between 30 °N and 75 °N on a $5° \times 5°$ grid is problematic, as already discussed by Cochelin (2004). For the downscaling, anomalies simulated by UK Universities Global Atmospheric Modeling Program (UGAMP) AGCM for present day climate were used. This is problematic for simulations of climates different from the present-day climate, as it assumes that the anomalies are the same as today, which is probably not the case. In addition, Cochelin (2004) showed that the UGAMP AGCM has a warm bias over the center of Canada and a cold bias over southern Alaska, as compared to present-day observations from ECMWF. Therefore it might be advisable to think about ways to improve the downscaling scheme.

4. Wang, Z. (2005) already improved the MPM by adding the Antarctic and

Arctic regions into the model and thus making the MPM a true global climate model. Antarctica thereby is ice and snow covered and has a prescribed elevation corresponding to its present-day level. The Arctic Ocean is included as a mixed-layer ocean. No freshwater storage is considered in the Arctic Ocean, so the net freshwater budget of the Arctic Ocean (precipitation, evaporation, runoff and salt rejection/freshwater release due to sea-ice formation/melting) in added to the northernmost box of the North Atlantic. Wang, Z. (2005) also improved the atmospheric part of the MPM by introducing active winds instead of prescribed winds. In addition, flux adjustments are eliminated in this new version of the MPM, which is now called the MPM-2. The MPM-2 has been used in a study of the THC (Wang, Z. 2005); however, the influence of the improvements in the MPM-2 on the simulation of the last and next glacial inception still has to be determined.

3.3.2 Future Plans

Once the thermodynamic part of the ice sheet model is coupled to the MPM, simulations over the last glacial could be performed to investigate how and if the MPM can simulate the dynamics of D/O events and HE's, which is interesting since the physical processes behind them are still unclear and there are many hypothesis that can be tested.

After the carbon cycle is closed in the MPM and the thermodynamic part of the ice sheet model is coupled as well, the simulation of the last glacial inception should be repeated to test the new model. The model should also be run over the full glacial-interglacial cycle to investigate the climate dynamics and to find out how good the model reproduces paleo-data. If the results are satisfactory, the MPM can then be used to again simulate the next glacial inception. this time without prescribing the CO_2 concentration, since the carbon cycle module will be able to calculate the atmospheric CO_2. To simulate anthropogenic influences, scenarios of predicted CO_2 emissions need to be prescribed, so that the carbon cycle module can calculate the corresponding atmospheric CO_2 concentration.

Before all this is possible, the effect of the already implemented improvements of the "green" MPM (inclusion of precipitation-vegetation feedback and the extension from 75° to 90°) on the simulation of the last glaciation could be tested. It would be interesting to see if they would lead to a larger ice volume, which would bring the model results in better agreement with paleo-data.

4 Summary

Much progress has been made in the understanding of glacial-interglacial cycles in the past 30 years, since Kukla et al. (1972) predicted that the present interglacial would end soon. Nevertheless, the dynamics which govern the glacial-interglacial cycles are still not fully understood. The results from Paillard (1998) suggest that the glacial-interglacial cycles are indeed driven by insolation, as predicted by Milankovitch (1930), since Paillards (1998) simple, orbitally forced model could successfully simulate the observed glacial-interglacial cycles of the past 2 million years, including the change from the dominant 41 kyr cycle to the 100 kyr cycle after the MPR around 900 kyr BP. However, which of the orbital parameters is the primary driver of glacial-interglacial cycles remains debated. Vetoretti and Peltier (2004) found that glacial inceptions can be caused either by a strong obliquity forcing or by a combination of eccentricity-precession forcing and low CO_2 values, which is in line with results from Berger and Loutre (2001) who found that CO_2 is important during times like the MIS-11, when the insolation variations are too small to drive glacial-interglacial cycles. However, Kubatzki (2005) found that a simultaneous changes in the perihelion and obliquity forcing is necessary to initiate a full glaciation, while obliquity forcing alone or in combination with eccentricity forcing is not able to cause a glaciation.

The cause of the observed variations in greenhouse gases in the atmosphere over glacial-interglacial cycles remains unknown, even though changes in the productivity of the oceanic biological pump in the Southern Ocean or the low latitude ocean, as well as changes in the sea-ice coverage of the polar Southern Ocean seem to offer promising explanations (see Sigman and Boyle (2000) for a review). The results of modeling studies (e.g., Loutre and Berger, 2000a; Kubatzki, 2005; Vetoretti and Peltier, 2004; Calov et al., 2005b) suggest that the observed changes in greenhouse gases are not the main drivers of glacial-interglacial cycles, but that these cycles exist even without greenhouse gas variations. Nevertheless, greenhouse gas variations seem to strongly amplify the effect of orbital forcing on climate (e.g., Calov et al., 2005b, Loutre and Berger, 2000a).

The importance of vegetation feedbacks on glacial inceptions was found to be important (e.g., Meisser et al., 2003; Yoshimori et al, 2002; Kageyama et al., 2003; Wang, Z. et al., 2005), even so their effect appears to be highly model dependent (Brovkin et al., 2003).

After a glaciation has started, the ice-albedo feedback was found to be the most important mechanism that led to rapid expansion of ice sheets in some studies (Calov et al., 2005a; Kageyama et al., 2004), while other studies found the elevation effect to be the most important feedback (Wang and Mysak, 2002).

Overall, the results from modeling studies suggest that glacial inceptions are threshold processes (e.g. Calov et al., 2005a; Cochelin et al., 2005; Paillard, 1998), which explains some differences between results obtained with different models, since threshold-processes are strong non-linear processes, so that small changes that exceed the threshold can trigger large changes. Different thresholds within the models as well as different climate sensitivities of models to changes in various variables can therefore lead to the simulation of different climate states, even for similar forcing.

Concluding, it is obvious that much works remains to be done until we will be able to fully understand the dynamics of glacial-interglacial cycles. In order to predict our future climate it is crucial to gain this understanding, since until we understand which role the greenhouse gases in the atmosphere played during past glacial inceptions and terminations, we will not be able to successfully predict the long term effect that our anthropogenic greenhouse gas emissions will have on the future climate. Therefore better models, which should include all processes that are important on the time-scale of glacial-interglacial cycles as well as on the shorter timescale of terminations and inceptions, are necessary.

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Glossary

AGCM .... Atmospheric GCM

AOGCM .... Atmosphere-Ocean GCM

AP .... After Present

BP .... Before Present

CH_4 .... Methane

CO_2 .... Carbon dioxide

EBM .... Energy Balance Model

EMIC .... Earth System Models of Intermediate Complexity

GCM .... General Circulation Model

kyr .... thousands of years

LGM .... Last Glacial Maximum

MBE .... Mid-Brunhes Event

MIS-II .... Marine Isotope Stage 11

MPM .... McGill Paleoclimate Model

MPR .... Mid-Pleistocene Revolution

NADW .... North Atlantic Deep Water

NH .... Northern Hemisphere

Pg C .... 10^{15} g Carbon

ppb .... Parts per billion

ppm .... Parts per million

SH .... Southern Hemisphere

SST .... Sea Surface Temperature

THC .... Thermohaline circulation