Constraining atmospheric CO$_2$ content during the Middle Miocene Antarctic glaciation using an ice sheet-climate model

P. M. Langebroek$^1$, A. Paul$^{1,2}$, and M. Schulz$^{1,2}$

$^1$Faculty of Geosciences, University of Bremen, Bremen, Germany
$^2$MARUM – Center for Marine Environmental Sciences, Univ. of Bremen, Bremen, Germany

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Correspondence to: P. M. Langebroek (petra@palmod.uni-bremen.de)

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Abstract

Foraminiferal oxygen isotopes from deep-sea sediment cores suggest that a rapid expansion of the Antarctic ice sheet took place in the Middle Miocene around 13.9 million years ago (Ma). The origin for this transition is still not understood satisfactorily. Among the proposed causes are a drop in the partial pressure of atmospheric carbon dioxide ($p$CO$_2$) in combination with orbital forcing. An additional complication is the large uncertainty in the magnitude and age of the reconstructed $p$CO$_2$ values and the low temporal resolution of the available record in the Middle Miocene. We used an ice sheet-climate model with an energy and mass balance module to assess variations in ice-sheet volume induced by $p$CO$_2$ and insolation forcing and to better constrain atmospheric CO$_2$ in the Middle Miocene. The ice-sheet sensitivity to atmospheric CO$_2$ was tested in several scenarios using constant $p$CO$_2$ forcing or a regular decrease in $p$CO$_2$. Small, ephemeral ice sheets existed under relatively high atmospheric CO$_2$ conditions (between 400–450 ppm), whereas more stable, large ice sheets occurred when $p$CO$_2$ is less than 400 ppm. Transitions between the states were largely CO$_2$-induced, but were enhanced by extremes in insolation. In order to explain the Antarctic glaciation in the Middle Miocene as documented by the oxygen isotope records from sediment cores, $p$CO$_2$ must have decreased by approximately 150 ppm in about 30 ka, crossing the threshold $p$CO$_2$ of 400 ppm around 13.9 Ma. Forcing the ice sheet-climate model with cyclic $p$CO$_2$ variations at a period of 100 ka and amplitudes of approximately 40 ppm generated late Pleistocene glacial-interglacial like ice-volume variations, where the ice volume lagged $p$CO$_2$ by 11–16 ka.

1 Introduction

In the Middle Miocene, around 13.9 million years ago (Ma), a large shift towards heavier benthic oxygen isotope values ($\delta^{18}$O) is found in deep-sea sediment records (Zachos et al., 2001; Holbourn et al., 2005). This increase coincided with a global sea-level
drop (Miller et al., 1998, 2005) and is interpreted as an expansion of the Antarctic ice cap and a global transition into a colder climate (e.g., Zachos et al., 2001; Shevenell et al., 2004). Several causes for the transition are proposed, ranging from the effect of ocean circulation and gateways (e.g., Flower and Kennett, 1995), enhanced chemical weathering and burial of organic matter (e.g., Raymo, 1994), to orbital forcing in combination with variations of partial pressure in the atmosphere ($p\text{CO}_2$) (Holbourn et al., 2005, 2007). A remarkable difference of this glaciation with respect to previous large-scale events (e.g. the Eocene-Oligocene transition; Coxall et al., 2005; Pollard and DeConto, 2005) is the relatively small decline in atmospheric CO$_2$, varying only in the order of 100–200 ppm (cf., Pearson and Palmer, 2000; Royer et al., 2001; Pagani et al., 2005; Kürschner et al., 2008; Zachos et al., 2008). This raises questions about the origin of the Middle Miocene transition. Was orbital forcing by itself sufficient to cause a large-scale continental glaciation? Or was some additional reduction of atmospheric CO$_2$ needed? And if so, how large and how quick was this $p\text{CO}_2$ drop? Considering the large uncertainties in the most recent $p\text{CO}_2$ reconstructions for the Middle Miocene (Pearson and Palmer, 2000; Royer et al., 2001; Pagani et al., 2005; Kürschner et al., 2008; Zachos et al., 2008), this study aims to give constrains on for timing, duration and speed of the $p\text{CO}_2$ transition using a modelling approach. We examine these questions using a geometrically simplified, but physically comprehensive ice sheet-climate model, which is forced by insolation (derived from the orbital parameters) and atmospheric CO$_2$ only.

2 Methods and experimental set-up

2.1 Ice sheet-climate model

We used a coupled ice sheet-climate model. The climate component consists of three large-scale boxes covering the entire southern hemisphere: a low (0–30° S), middle (30–60° S) and high (60–90° S) latitude box (Fig. 1). Forcing consists of seasonal or-
bital forcing following the work of Laskar et al. (2004) in combination with prescribed atmospheric CO$_2$ levels. In the large-scale boxes of the climate model, energy is conserved and is redistributed by meridional energy transport, taking into account the latent heat fluxes due to evaporation and snow accumulation. The physical processes within the high latitude box are resolved in 0.5° latitude bands. In these boxes energy balances the for atmosphere and surface are resolved separately, but computed simultaneously. Additionally, the mass balance for the ice-sheet component is modeled. Daily computation is necessary, because the orbital cycle as well as processes of snow accumulation and melting have a strong seasonal imprint (Pollard, 1983).

The atmospheric and surface energy balances include parameterizations for short- and longwave radiation, latent heat of evaporation and snowfall, sensible heat exchange, heat flux into the surface and energy used by melting of ice and snow (Pollard, 1982, 1983; Jentsch, 1987, 1991; Wang and Mysak, 2000). Total accumulation and its latitudinal distribution is tuned to the present-day (total) Antarctic accumulation and depends on surface temperature, distance from the coast, surface height and daily surface temperature (Oerlemans, 2002, 2004). It therefore includes important processes such as the elevation-desert effect (Pollard, 1983). The ice-sheet model is symmetric around the axis of the South Pole. Within the ice sheet velocities and temperatures are computed with a vertical resolution of 12 layers. The altitude and ice thickness of every latitude grid cell are derived by solving the continuity equation using basal melting, local bedrock isostasy and a surface mass balance (Sima, 2005; Sima et al., 2006). The initial ice-free bedrock topography is reconstructed using the BEDMAP project database (Lythe et al., 2000) for bedrock elevation and ice thickness, considering local isostasy. For the axially symmetric ice-sheet model, the high spatial resolution of the dataset is reduced, averaging the topography into the 0.5°-wide latitude bands. The initial bedrock used by the model is a simplified version of the zonally-averaged topography that includes a bulge close to the continental shelf and a flatter hinterland. Although no separate ocean component is included in the model, the energy and mass balances within the Antarctic box include the albedo of (seasonally varying) sea-ice.
This is parameterized depending on the near-surface temperature of the appropriate grid cells. The surface albedo of the Antarctic continent depends on the ice and snow content of the corresponding grid cell and combines albedos of land (0.3), ice (0.35) and snow (0.75). A more detailed description of the climate forcing can be found in the appendix.

2.2 Climate sensitivity of the model

The equilibrium climate sensitivity as estimated from 19 atmospheric general circulation models ranges from 2.1 to 4.4°C, with an average of 3.2°C (Randall et al., 2007). These models do not include a dynamic ice-sheet component, but do account for changes in snow cover and albedo. The large spread in sensitivity is introduced by differences in feedback parameterizations. To tune the climate sensitivity of our model, we first ran the coupled ice sheet-climate model for the last 100 ka with constant pre-industrial $p$CO$_2$ of 280 ppm and varying orbital parameters. The modeled present-day ice sheet is in equilibrium with the radiative forcing and has a volume of $25.1 \times 10^{15}$ m$^3$, similar to its estimated present-day size (e.g., Huybrechts et al., 2000; Oerlemans, 2002; Huybrechts, 2004). The mean hemispheric surface temperature is 14.8°C. The model is tuned such that a doubling of $p$CO$_2$ gives a reasonable temperature increase. We deliberately enhanced the sensitivity to changes in $p$CO$_2$ in order to account for the missing water vapor feedback (see Appendix). In the tuned model, a doubling of $p$CO$_2$ while maintaining fixed ice-sheet height and (seasonal) insolation distribution resulted in a hemispheric mean temperature increase of 2.8°C. This value falls well within the range of the values reported by the IPCC report (2007). The largest increase is found in atmospheric and surface temperatures in the Antarctic, high latitude box, with values up to 11.6°C. The large polar amplification is due to the included ice-albedo feedback.
2.3 Insolation and $p$CO$_2$ forcing

Based on Earth’s orbital elements as computed by Laskar et al. (2004) we compute daily insolation (Berger, 1978a,b) at the top of the atmosphere (for every latitude box) and at the surface for the high resolution Antarctic cells (60–90°S). We will use two different averages for comparison to ice-volume variations: annual mean and caloric summer (half-year of highest values) insolation. Since the atmospheric CO$_2$ level in the Middle Miocene is not very well constrained, the model is forced by prescribed scenarios of constant $p$CO$_2$, constant decrease in $p$CO$_2$ and a $p$CO$_2$ forcing including a 100-ka cycle (eccentricity).

2.4 Experimental set-up

First, hysteresis experiments are carried out, in order to find $p$CO$_2$-threshold values at which the Antarctic continent (de)glaciates. In addition, different levels of constant atmospheric CO$_2$ are applied for model runs of 1 Ma, from 14.2 to 13.2 Ma (preceded by a 100 k spin-up time), in order to investigate the effect of insolation fluctuations on ice-sheet volume under different constant $p$CO$_2$ conditions. Between 200 and 450 ppm, every 10 ppm is used for constant model forcing, with an increased resolution of 5 ppm between 390 and 410 ppm. Second, sensitivity experiments involving a reduction in $p$CO$_2$ are carried out, focusing on ice-sheet response to the level, speed and timing of $p$CO$_2$ decrease. Finally, $p$CO$_2$ forcing with a frequency of 100 ka is used to look into the mechanism causing eccentricity cycles within the sedimentary records. In all experiments specific daily insolation for appropriate latitudes is applied.
3 Results

3.1 Hysteresis experiments

To assess the sensitivity of the simulated ice volume to $p\text{CO}_2$ changes and in order to find the critical range of $p\text{CO}_2$ at which the Antarctic continent (de)glaciates, two types of hysteresis experiments were performed. The first included orbital variations, while the second is only forced by atmospheric $\text{CO}_2$ (Fig. 2). In both experiments the (rapid) transition into a large ice-sheet occurred around 400 ppm, preceded by a semi-stable small ice sheet. In case of omitting orbital variations, deglaciation occurred at approximately 550 ppm, accounting for a hysteresis window of $\sim$150 ppm. Orbital forcing acts as noise, therefore the ice sheet melts under much lower $p\text{CO}_2$ when it is included ($\sim$425 ppm). The modeled rate of $p\text{CO}_2$ change was slow at 50 ppm/Ma, comparable to the 280 ppm/5 Ma used by Pollard and DeConto (2005).

3.2 Constant $p\text{CO}_2$ experiments

The critical $p\text{CO}_2$ for glaciation is approximately 400 ppm. Below this threshold the entire Antarctic continent is glaciated, with mean ice volumes between 23 and $25\times10^{15}$ m$^3$ (Fig. 3). Between $\sim$405 and $\sim$430 ppm (almost) continuous small ice sheets existed for the modeled period. Higher $p\text{CO}_2$ levels resulted in small, ephemeral ice sheets. Under constant $p\text{CO}_2$ forcing and insolation derived from orbital parameters of the Middle Miocene (between 14.2 and 13.2 Ma) either large or small ice sheets occurred (Figs. 3 and 4). Only constant $p\text{CO}_2$ values close to the threshold of 400 ppm caused a transition between these two states. Using constant 400 ppm forcing, Antarctic glaciated at 13.43 Ma. All small ice sheets showed large variations in ice volume, up to $\sim2.5\times10^{15}$ m$^3$ for constant $p\text{CO}_2$ of 410 ppm. Volume of large ice sheets varied less under constant $p\text{CO}_2$ conditions, with a maximum variance of $\sim1.1\times10^{15}$ m$^3$ for $p\text{CO}_2$ values close to the threshold and nearly no variance at lower $p\text{CO}_2$ levels. For further comparison two runs of constant $p\text{CO}_2$ close to the glaciation threshold value and with
maximum ice-volume variance are used to represent the large ice sheet (390 ppm) and the small ice sheet (410 ppm).

The correlation between ice-volume variations and annual and summer mean insolation was computed for every 5° of southern latitude (Fig. 5). Both ice-sheet variations correlated better to high (around 70° S), than to low latitudes. In the case of the large ice-sheet, highest correlation coefficients were reached when ice volume lagged insolation by approximately 2 ka. Maximum correlation coefficient values were 0.29 and 0.75 for annual and summer mean, respectively. The small ice-sheet matched insolation averages best for a lag of 5 to 6 ka. Maxima for annual and summer mean insolation were 0.49 and 0.60, respectively.

3.3 Sensitivity experiments

In the first sensitivity experiment atmospheric CO₂ decreased linearly at a rate of 50 ppm/ka (Fig. 6 – red curves). For every experiment, the timing of the drop was simultaneous, only the extent, and therefore the initial and final levels of ρCO₂ were different. The resulting ice volume transitions occurred around the same moment in time, whereby the largest difference in ρCO₂ forced the most rapid ice-sheet transition (Fig. 6 – blue curves). This experiment focusing on initial and final levels of ρCO₂ is repeated for different slopes of the ρCO₂ transition, with identical results (not shown).

The second test focused on the effect of the pace at which the atmospheric CO₂ is decreasing on the ice-sheet transition (Fig. 6). Experiments forced by a slow decrease in ρCO₂ resulted in a variable duration of ice-sheet transition, between 20-30 ka. In runs with a fast ρCO₂ drop, the transition length was independent of the speed of ρCO₂ drawdown. This relation also holds for different timing of the ρCO₂-transition (not shown). In the last sets of sensitivity experiments the forcing was applied at different moments in time (Fig. 7). Antarctica would have glaciated during the appropriate time interval (13.84–13.88 Ma) if a fast ρCO₂ transition occurred around 13.9 Ma or due to a slower ρCO₂ drawdown between 14.03 and 13.83 Ma.
4 Discussion

4.1 Constant $p$CO$_2$ experiments

In our coupled ice sheet-climate model, which is tuned to present-day conditions and a climate sensitivity of 2.8°C, relatively stable, large ice sheets occurred under $p$CO$_2$ below ~400 ppm (Fig. 3). Very stable ice sheets under relatively low $p$CO$_2$ levels (below 235 ppm) showed extremely small ice-volume variations. Correlations between continental ice volume and benthic oxygen isotopes vary between 1‰ (Zachos et al., 2001) and 2.2‰ (Pekar et al., 2002) for 100 m sea-level change. Taking the present ocean area ($3.6 \times 10^6$ km$^2$) and the densities of water and ice (1000 kg/m$^3$ and 910 kg/m$^3$, respectively), an apparent sea level (ASL) drop of 100 m is equivalent to the build-up of ice with a volume of approximately $33 \times 10^{15}$ m$^3$. Using the maximum oxygen-isotope sea-level calibrations ($\sim 2.2\% \approx 33 \times 10^{15}$ m$^3$), the standard deviation in the oxygen-isotope ratio derived from ice volume would be approximately 0.002‰ (Table 2) and resulting global sea-level fluctuations would be impossible to detect in the geological record. Larger variances are found in the modeled ice-volume record for a constant $p$CO$_2$ of 390 ppm (for a large ice sheet) and 410 ppm (for the small one) (Fig. 3). Ice-volume fluctuations in the small (big) ice sheet accounted for a maximum of 77% (41%) of the mean of the standard deviations in the benthic oxygen-isotope records (standard deviations are computed from the original oxygen-isotope data and afterwards averaged for comparison to the modeled ice-volume fluctuations; Tables 1 and 2). The modeled result that small ice sheets were more easily perturbed than large ice sheets can partly be traced back to the oxygen-isotope record (Table 1). An F-test showed that the variances of the original oxygen-isotope data in the restricted time domains were significantly different (significance level of 95%). This difference might be explained by the fact that small ice sheets are more easily perturbed by changes in the forcing. Additionally, ablation, which plays a major role in ice-volume variations, occurred on two sides of the small ice sheet, whereas the large ice sheet only had an ablation zone at the outer rim. Considering the very crude calibration used to compute these values, not
taking care of any effects caused by changes in (deep sea) temperature, salinity, local runoff, oxygen-isotope ratio of the ice and so on, the correlation between the standard deviations of data- and model derived oxygen-isotope ratios is surprisingly high. Further more, these experiments were performed under constant $p\text{CO}_2$-levels. In the final part of this paper an experiment forced by large $p\text{CO}_2$ fluctuations will be discussed.

The synchronous minima of eccentricity and obliquity at $\sim 13.84$ Ma (Abels et al., 2005) result in a relatively constant and average-to-low insolation, which has been proposed as being partly responsible for the large-scale glaciation in the Middle Miocene (Holbourn et al., 2005). In our ice sheet-climate model, the occurrence of such a special configuration or the natural variability in insolation by itself is not sufficient for an Antarctic glaciation. Only constant atmospheric $\text{CO}_2$ levels at or very close to the threshold of 400 ppm induced a transition from small to large ice volume (Fig. 3), in a period that does not precisely corresponding to the transition depicted from oxygen isotope records. Therefore, in order to appropriately glaciate the Antarctic continent, some decrease in $p\text{CO}_2$ must have occurred.

4.2 Sensitivity experiments

In order to constrain the atmospheric $\text{CO}_2$ transition in the Middle Miocene, experiments with a constant $p\text{CO}_2$ decrease were performed. Three important factors defining the transition (the amount of $p\text{CO}_2$ drawdown, the slope and the timing of the event) are highly unknown, and were therefore tested in the following sensitivity experiments. First the ice-sheet model is forced by a decrease in atmospheric $\text{CO}_2$ with identical slopes and timing (Fig. 6 – red curves). The resulting glaciations occurred around the same time, with some difference in ice-sheet growth efficiency. The larger the difference between initial and final $p\text{CO}_2$, the faster the transition was. This can be explained by the different variability of the ice sheet at different $p\text{CO}_2$ levels. The larger standard deviations of $1.059 \times 10^{15}$ m$^3$ and $2.542 \times 10^{15}$ m$^3$ (Table 1) in the 390– and 410 ppm-runs, respectively, enhance the possibility for insolation variations to act against a rapid ice-volume transition. Because of the fact that sedimentary records
do not indicate such a rapid glaciation (e.g., Holbourn et al., 2005) and do show large variance in time, the relatively small difference in $pCO_2$ (410 to 390 ppm) is used for the remaining experiments.

The second sensitivity test dealt with the slope of the atmospheric CO$_2$ drawdown (Fig. 6 – blue curves). Six experiments with slopes varying between 20 ppm/2 ka to 20 ppm/200 ka were overlapping in time and had equivalent initial and final $pCO_2$ levels. For most runs, the ice-volume transition took place at the same moment. Exception was the slowest forcing (20 ppm/200 ka), which glaciates much later. The duration of the ice-sheet transition was defined as the period in which ice volume is larger than the maximum volume of the small ice sheet and smaller than the minimum size of the large ice sheet. Additional experiments show that slow forcing does not have a strong correlation to the time necessary for glaciation (not shown). It is also evident that the rate of $pCO_2$ drop has no effect on the length of the ice-sheet transition. The duration merely depends on the timing of the $pCO_2$ drop. This quite constrained timing (see next paragraph) limits the glaciation event to a length of approximately 30 ka. Compared to the duration of glaciation depicted by available data (30–40 ka according to $\delta^{18}O$ records by e.g. Holbourn et al., 2005) this is on the fast side, indicating that the ice-sheet model may respond more rapidly than the real Antarctic ice sheet. Most probably the difference in $pCO_2$ before and after the transition is relatively small, in the order of 100–200 ppm cf. Pearson and Palmer (2000); Royer et al. (2001); Pagani et al. (2005); Kürschner et al. (2008); Zachos et al. (2008), which make the slowest $pCO_2$ transitional experiments not very likely.

Nevertheless, the third set of sensitivity experiments investigated the timing of the $pCO_2$ decrease, for the (too) slow pace of 20 ppm/200 ka and for the faster and more realistic speed of 20 ppm/4 ka (Fig. 7). The best fit to data (glaciation time shown by vertical green lines) occurred either with a fast drawdown around 13.9 Ma, or a slow drawdown starting centered around 13.925–13.95 Ma (bold lines). Shifting this by only 2 ka resulted into a much earlier ($\sim$13.91 Ma) or later ($\sim$13.78 Ma) glaciation. This experiment was based on a 20 ppm difference in $pCO_2$, but extending this range would
give similar results (see first sensitivity test).

The comparison of the three sensitivity experiments to glaciation- and $pCO_2$-estimates from sedimentary records indicates that the $pCO_2$ drop should cross the threshold at approximately 400 ppm and should have occurred just before the ice-sheet transition (around 13.9 Ma) with a slope of approximately 20 ppm/4 ka (corresponding to ~150 ppm/30 ka). Because of the fact that the exact values are model dependent, these numbers should only be taken as guidelines, but they do put some constraints on the amount, pace and extent of the $pCO_2$-decrease.

Our results also contradict the hypothesis that Antarctic glaciation partly originated due to synchronous minima in eccentricity and obliquity around 13.84 Ma (e.g., Hollbourn et al., 2005; Abels et al., 2005). The ice sheet-climate model is forced by daily insolation, which were averaged over the whole year (annual mean) and over the caloric summer (half year of largest daily insolation) to promote comparison to the resulting ice-volume variations. The two upper panels in Fig. 3 show the annual and summer mean insolation over the Middle Miocene period. The latitude of 70° S is chosen, because the large as well as the small ice-sheet volume variations correlated best to this latitude, although with a different lag time (see Result section). The combined minima in eccentricity and obliquity at ~13.84 Ma resulted in an average to high summer and annual mean insolation at 70° S. From the sensitivity experiments it can be seen that Antarctic glaciation is favored by maxima in ice-volume variations of the small ice sheet. These maxima are correlated to minima in insolation. For an ice-sheet extension in the Middle Miocene to occur, similar as illustrated by benthic oxygen-isotope records, the minima in high-latitude insolation at approximately 13.88 Ma is the most suitable candidate. So, instead of pointing to 13.84 Ma as an important insolation moment for the Antarctic glaciation, 13.88 Ma is more appropriate.

4.3  100-ka cycles

Without a carbon cycle implemented in the ice sheet-climate model, it is of course difficult to discuss possible interactions between the carbon cycle, ice volume and global
climate. On the other hand the low computation time allowed us to easily test the effect of various $p\text{CO}_2$ scenarios on ice volume. Here we focus on $p\text{CO}_2$-cycles with a frequency of 100 ka, similar to glacial-interglacial cycles of the Quaternary. It has been proposed that these cycles are caused by variability in the carbon cycle (e.g., Shackleton, 2000; Pälike et al., 2006) in contrast to studies showing that internal ice-sheet dynamics could result in the appropriate periodicity (Pollard, 1982, 1983). Following the previous ideas, the model was forced by atmospheric $\text{CO}_2$ changes that included a 100-ka cycle. Glacial-interglacial-like ice-volume cycles are found when the mean $p\text{CO}_2$ is around 420 ppm and the amplitude is about 40 ppm (Fig. 8). This is comparable to the ~50 ppm amplitude in $p\text{CO}_2$ found in ice-core records (e.g., Petit et al., 1999). When maxima in $p\text{CO}_2$ are close to maxima in eccentricity, the 100-ka cycle is most apparent. Similar to ice-volume cycles in the late Quaternary recorded in marine sediment records (e.g., Lisiecki and Raymo, 2005) the modelled ice volume showed slow ice build-up and rapid terminations. This asynchronous, threshold behaviour originates from internal model feedbacks, as the eccentricity rhythm within the input forcing was purely sinusoidal. Ice volume lagged $p\text{CO}_2$ with approximately 11–16 ka, which is in range with the 14 ka deducted by Shackleton (2000).

5 Conclusions

Despite the relatively simple geometry of our ice sheet-climate model, the realistically tuned climate sensitivity and hysteresis experiments indicate that the mechanism described in the following conclusions can be considered robust. However, exact numbers are model dependent and should only be taken as a guideline.

1. It is very unlikely that a constant $p\text{CO}_2$ forcing induced the large-scale Antarctic glaciation in the Middle Miocene. Constant levels produced either a large (below ~400 ppm threshold) or a small (above ~400 ppm) ice sheet. Large ice sheets covered the whole Antarctic continent and had a smaller ice-volume variation than...
expected from sedimentary deep-sea records (~41%). The variance in the small ice sheet explained the fluctuation in oxygen-isotope ratios in these records for ~78%. Residual variation in the isotope records can originate from fluctuations in $pCO_2$ or other changes in climatic conditions. Ice-volume variations correlated best to insolation at relatively high latitudes (~70° S). In case of the small ice sheet, summer mean insolation was the main forcing, which lead ice volume by ~2 ka. Ice volume lagged insolation by 5-6 ka in the large ice sheet and correlated only to some extent better to summer than annual mean values.

2. The extent of the $pCO_2$ drawdown was not important for timing or duration of the glaciation event, as long as it crosses the 400 ppm threshold. Moderate or quick $pCO_2$ reductions resulted in comparable and realistic ice-sheet extension. The ice-sheet response was fast, which limited the $pCO_2$ drawdown to happen around 13.9 Ma. A relatively slow drop in $pCO_2$ caused a delayed glaciation and had to occur at 13.925–13.950 Ma. The best guess for the Middle Miocene $pCO_2$ decline was a scenario crossing the threshold of 400 ppm around 13.9 Ma with a speed of approximately 150 ppm/30 ka.

3. Forcing the ice sheet-climate model with 100-ka $pCO_2$ cycles of 40 ppm amplitude resulted in late Pleistocene ice-age-like behavior, with slow ice-volume build-up, and rapid terminations. Ice-volume variations lagged $pCO_2$ cycles by 11–16 ka, similar to what had been found by Shackleton (2000).

Appendix A

Model description

The ice sheet-climate model is controlled by energy and mass balances. Orbital elements are derived following the work of Laskar et al. (2004). They drive the seasonal
solar radiation at the top of the atmosphere and define, together with the $pCO_2$, the amount of energy entering the entire climate system.

A1 Energy and temperature balances

The model consists of three large-scale boxes covering the entire southern hemisphere: a low (0–30° S), middle (30–60° S) and high (60–90° S) latitude box (Fig. 1). Within the climate system energy is conserved and changes in time are described by (Pollard, 1983; Hartmann, 1994):

$$\frac{\partial E_{ao}}{\partial t} = R_{TOA} + \Delta F_{ao} + LH$$  \hspace{1cm} (A1)

where $E_{ao}$ is the total energy in the system, $t$ the time of year, $R_{TOA}$ the net incoming solar radiation on the top of the atmosphere, $\Delta F_{ao}$ the divergence of the meridional energy transport in the ocean as well as in the atmosphere, and $LH$ the latent heat added to the atmosphere after condensation and freezing of water vapor. The net incoming radiation is the sum of the incoming short-wave radiation ($SW_p$) and the outgoing short- and long-wave radiation ($LW_p$):

$$R_{TOA} = SW_p^\downarrow - SW_p^\uparrow - LW_p^\uparrow.$$  \hspace{1cm} (A1)

Radiation fluxes in the two lower latitude boxes (0–30° S and 30–60° S) are parameterized as:

$$SW_p^\downarrow - SW_p^\uparrow = Q(1 - \alpha_p)$$

$$LW_p^\uparrow = \varepsilon_p \sigma T_a^4 + f_{CO_2}$$

where $Q$ is the solar insolation at the top of the atmosphere, $\alpha_p$ the planetary albedo, $\varepsilon_p$ the planetary emissivity and $\sigma$ the Stefan-Boltzmann constant. $T_a$ is interpreted as
the near-surface air temperature and $f_{CO_2}$ as the effect of the atmospheric CO$_2$ content (cf., Myhre et al., 1998):

$$f_{CO_2} = -\frac{\ln(\frac{CO_2}{280})}{\ln(2)} \approx 2.8 - 0.7\ln(CO_2).$$

Therefore, a doubling of $pCO_2$ from 280 ppm (pre-industrial conditions) to 560 ppm accounts for a reduction of 4 W/m$^2$ in the outgoing longwave radiation of the two lower latitude boxes.

The physical processes in the high latitude box are deciphered in much higher resolution and complexity. For every 0.5° latitude energy and mass balances for the atmosphere and for the surfaces are simultaneously solved. Atmospheric temperature ($T_a$) is described by:

$$C_a \frac{dT_a}{dt} = R_a + LW + SH + LH_{eva} + LH_{snow}$$

and surface temperature ($T_s$) by:

$$C_s \frac{dT_s}{dt} = R_s - LW - SH - LH_{eva} - F_s - F_m$$

where $C_{a,s}$ is the heat capacity for the atmosphere and surface, respectively.

The incoming energy at the top of the atmosphere and at the surface is represented as (Jentsch, 1987; Wang and Mysak, 2000):

$$R_a = SW_a^{\downarrow} - SW_a^{\uparrow} - LW_a^{\uparrow}$$

$$= Q(1 - \alpha_a)(1 - \tau)(1 + \tau\alpha_s) - (\varepsilon_2\sigma T_a^4 + (1 - \varepsilon_1)\sigma T_s^4)$$

$$R_s = SW_s^{\downarrow} - SW_s^{\uparrow}$$

$$= \tau Q(1 - \alpha_a)(1 - \alpha_s)$$
where $\tau$ is the atmospheric transmissivity of solar radiation, $\alpha_{a,s}$ the atmospheric and surface albedos, $\varepsilon_2$ an emissivity constant and $\varepsilon_1$ a term describing the greenhouse effect (see below).

The longwave and sensible heat fluxes between the atmosphere and surface are parameterized as:

$$LW = \sigma T_s^4 - \varepsilon_1 \sigma T_a^4$$

$$SH = \lambda (T_s - T_a)$$

where $\lambda$ is a heat exchange coefficient which in principle depends on wind speed, atmospheric density and heat capacity, but is taken to be constant. The heat flux into the subsurface soil or upper ice layer ($F_s$) is given by:

$$F_s = \frac{2k_1}{\Delta z_1} (T_s - T_a)$$

where $k_1$ is the thermal conductance of snow and $\Delta z_1$ the depth range of conduction.

The latent heat of evaporation ($LH_{eva}$) is parameterized as (Hartmann, 1994):

$$LH_{eva} = \rho_{air} L_v C_{DE} U [q_s^{*} (1 - RH) + \frac{RH c_p}{B_e L_v} (T_s - T_a)]$$

where $\rho_{air}$ is the air density, $L_v$ is the latent heat of vaporization, $C_{DE}$ an exchange coefficient, $U$ the wind speed, $q_s^{*}$ the sea surface humidity, $B_e$ the equilibrium Bowen ratio, $c_p$ the specific heat of dry air and $RH$ the relative humidity.

The latent heat associated with snowfall ($LH_{snow}$) depends on the accumulation of snow:

$$LH_{snow} = L_s A$$
where $L_s$ is the latent heat of sublimation and $A$ the accumulation. The snow is considered to be evaporated in the low latitude box, accounting for the $LH$-term in the energy equation (Eq. ??). The total accumulation and its latitudinal distribution is tuned to the present-day total Antarctic accumulation and depends on the distance to the South Pole ($r$), the surface height ($h_{sfc}$) and the daily surface temperature ($T_s$) (Oerlemans, 2002, 2004). It therefore includes processes such as the elevation-desert effect (Pollard, 1983):

$$A = (c_a + c_b r) e^{-h_{sfc}(r)} e^{\kappa T_s}$$

where $c_{a,b}$ are (tuning) constants, $c_d$ is a characteristic length scale and $\kappa$ a constant describing the precipitation dependence on temperature. Only when the local temperature is below $2^\circ C$, snow is accumulated (Oerlemans, 2001).

The amount of energy available for melting ($F_m$) depends on the incoming energy and the thickness and heat capacity of the top layer (Fraedrich et al., 2005). The affected layer is 20 cm deep ($d_{top}$) and consists of snow ($d_{snow}$), soil ($d_{soil}$) or a mixture of both. The heat capacity ($C_s$) used for computation of the surface temperature is therefore computed by:

$$C_s = \frac{C_{snow} C_{soil} d_{top}}{C_{snow} d_{soil} + C_{soil} d_{snow}}$$

The atmospheric and surface temperature equations are simultaneously solved. Daily computation is necessary, because the orbital cycle as well as processes of snow accumulation and melting have a strong seasonal imprint (Pollard, 1983). The meridional heat transport ($\Delta F_{eo}$) accounts for the coupling between the boxes, and is proportional to the temperature gradient based on the diffusion approximation (Sellers, 1970; North, 1975). The atmospheric temperatures, and also the surface temperatures, are further extrapolated towards their altitudes ($h_{sfc}$) according to the prescribed lapse rate, $\Gamma_{\text{lapse}}$:

$$T_a = T_a + \Gamma_{\text{lapse}} h_{sfc}(r).$$
A2 Mass balance

The mass balance is solved cumulatively on a daily basis. The specific mass balance, the total amount of accumulation or ablation (per latitude) within one year, possibly reduced by (surface or bottom) melting, evaporation or calving, is annually added to or subtracted from the snow/ice-sheet.

The ice sheet is allowed to grow into the surrounding ocean as long as it is hydrostatically floating. When the total weight of the ice column exceeds the floating criteria, calving occurs (Pollard, 1982) and the total mass balance \( G \) will be set to a negative value \( c_{\text{bal}} \):

\[
G = c_{\text{bal}} \text{ if } \rho_{\text{air}} h_{\text{ice}} < \rho_{w} (h_{\text{sfc}} - h_{\text{ice}})
\]

where \( \rho_{\text{ice}} \) and \( \rho_{w} \) are the densities of ice and water, respectively, \( h_{\text{ice}} \) is the ice thickness, and \( h_{\text{sfc}} \), the elevation of the surface with respect to the current sea level, which is taken as a constant reference level. This crude calving parameterization also accounts for occurrence of proglacial lakes and/or marine incursions (Pollard, 1982).

Bottom melting \( S \) occurs when the temperature in the basal layer \( T_{\text{base}} \) exceeds the pressure melting point \( T_{\text{pmp}} \):

\[
S = \frac{C_{\text{ice}}}{L_{m}} (T_{\text{base}} - T_{\text{pmp}}) \frac{\Delta z_{\text{base}}}{\Delta t}
\]

where \( C_{\text{ice}} \) is the specific heat of ice and \( L_{m} \) the specific latent heat of fusion of ice and \( \Delta z_{\text{base}} \) the thickness of the basal layer.

A3 Albedo

A separate snow balance is computed to parameterize the surface albedo. The formulas for this cumulative balance resemble the previous surface mass and energy balance
equations, except for the fact that the snow depth cannot become negative. The daily derived surface albedo ($\alpha_s$) depends on the snow depth ($d_{\text{snow}}$), when the snow layer is thicker than 10 cm:

$$\alpha = \frac{\alpha_{\text{snow}} + \alpha_{\text{ice}}}{2} + \frac{\alpha_{\text{snow}} - \alpha_{\text{ice}}}{2} \tanh(A_{\text{snow}}(d_{\text{snow}} - B_{\text{snow}}))$$

where the slope ($A_{\text{snow}}$) and shift ($A_{\text{snow}}$) are constant and $\alpha_{\text{snow}}$ and $\alpha_{\text{ice}}$ are the albedos of snow and ice, respectively. When there is less or no ice/snow, the land, ocean (low and middle latitude boxes) or sea-ice albedos (high latitude box) are used. The latitudinal extent of sea-ice ($\text{lat}_{\text{si}}$) is given by (Jentsch, 1987):

$$\text{lat}_{\text{si}} = \sin^{-1}[\tanh(x_0(T_{pd}/T_a)^{x_1})] - C_{\text{si}}$$

where $x_0$ and $x_1$ are tuning constants, $T_{pd}$ a measure for the present-day value of sea-water temperature and $C_{\text{si}}$ a latitudinal shift.

The planetary ($\alpha_p$) and atmospheric ($\alpha_a$) albedos are parameterized as functions of latitude (Wang and Mysak, 2000):

$$\alpha_p = 0.6 - 0.4 \cos(\text{lat})$$

$$\alpha_a = 0.3 - 0.1 \sin(\text{lat})$$

A4 Greenhouse effect

The longwave radiation constant $\varepsilon_1$ accounts for the greenhouse effect due to $pCO_2$ and other greenhouse gases:

$$\varepsilon_1 = \varepsilon_{10} + \varepsilon_{11} \sqrt{e'}$$
where, $e'$ is the atmospheric vapor pressure, related to the saturation specific humidity ($q_{sat}$) and relative humidity (RH):

$$e' = 1.6 \times 10^3 \text{RH} q_{sat}$$

where:

$$q_{sat} = \frac{1.57 \times 10^{11}}{\rho_{air} R_{air} T_a} e^{-\frac{-5421}{T_a}}$$

with $R_{air}$ being the gas constant for dry air.

According to Staley and Jurica (1970) and Jentsch (1991), the CO$_2$-emission factor can be parameterized by:

$$\varepsilon_{CO_2}^{10} = 0.1 + 0.025 \ln(\text{CO}_2).$$ \hspace{1cm} (A2)

The other main greenhouse gas, water vapor (H$_2$O), also contributes about half to the (present-day) greenhouse effect. Because of the fact that we do not explicitly compute the hydrological cycle, this feedback cannot be parameterized separately. To still include the effect of water vapor, we increased the climate sensitivity to $p$CO$_2$ Eq. A2 is therefore expanded and retuned to:

$$\varepsilon_{10} = \varepsilon_{10}^{CO_2} + \varepsilon_{10}^{H_2O} = 0.1 + 0.025 \ln(\text{CO}_2).$$

A doubling of atmospheric CO$_2$ now results in a climate sensitivity of 2.8°C and modeled present-day ice-sheet size, accumulation and temperature distribution are similar to estimates (Huybrechts et al., 2000; Oerlemans, 2002).

A5 List of constant parameters

Table A5 gives an overview of the parameters used in the ice sheet-climate model.

Acknowledgements. This project was funded by the DFG (Deutsche Forschungsgemeinschaft) within the European Graduate College “Proxies in Earth History”.
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Jentsch, V.: Cloud-ice-vapor feedbacks in a global climate model, in: Irreversible Phenomena and Dynamical Systems Analysis in Geosciences, edited by Nicolis, C. and Nicolis, G.,
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Table 1. Standard deviation of benthic oxygen-isotope records (‰) of Holbourn et al. (2005).

<table>
<thead>
<tr>
<th>Time interval (Ma)</th>
<th>References and comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>13.2–13.8</td>
<td></td>
</tr>
<tr>
<td>13.9–14.5</td>
<td></td>
</tr>
<tr>
<td>0.160</td>
<td>0.207 Site 1146</td>
</tr>
<tr>
<td>0.127</td>
<td>0.159 Site 1237</td>
</tr>
<tr>
<td>0.144</td>
<td>0.183 Mean of above two standard deviations for comparison to modeled ice-volume fluctuations</td>
</tr>
<tr>
<td>(Fig. 4 left)</td>
<td>(Fig. 4 right)</td>
</tr>
</tbody>
</table>
Table 2. Standard deviation of modeled ice-volume variation and equivalent oxygen-isotope ratios using $33 \times 10^{15} \text{ m}^3 = 1\%$ (Zachos et al., 2001) and $33 \times 10^{15} \text{ m}^3 = 2.2\%$ (Pekar et al., 2002).

<table>
<thead>
<tr>
<th>Time interval (Ma)</th>
<th>$p\text{CO}_2$ (ppm)</th>
<th>Standard deviation (10$^{15}$ m$^3$)</th>
<th>ASL (m)</th>
<th>$\delta^{18}$O (%) (33$\times$10$^{15}$ m$^3$ = 1%)</th>
<th>$\delta^{18}$O (%) (33$\times$10$^{15}$ m$^3$ = 2.2%)</th>
<th>Percentage of data (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>13.2–13.8</td>
<td>220</td>
<td>0.041</td>
<td>0.104</td>
<td>0.001</td>
<td>0.002</td>
<td>1% of 0.144</td>
</tr>
<tr>
<td>13.2–13.8</td>
<td>390</td>
<td>1.059</td>
<td>2.677</td>
<td>0.027</td>
<td>0.059</td>
<td>41% of 0.144</td>
</tr>
<tr>
<td>13.9–14.5</td>
<td>410</td>
<td>2.542</td>
<td>6.426</td>
<td>0.064</td>
<td>0.141</td>
<td>77% of 0.183</td>
</tr>
</tbody>
</table>
### Table A1. List and description of constant parameters.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Value</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \varepsilon_p )</td>
<td>Planetary emissivity</td>
<td>0.61 (15° S) – 0.66 (45° S) – 0.69 (75° S) –</td>
<td>–</td>
</tr>
<tr>
<td>( \varepsilon_{11} )</td>
<td>Emissivity constant (Sellers, 1970)</td>
<td>0.05</td>
<td>–</td>
</tr>
<tr>
<td>( \varepsilon_2 )</td>
<td>Emissivity constant (Jentsch, 1987)</td>
<td>0.30</td>
<td>–</td>
</tr>
<tr>
<td>( \sigma )</td>
<td>Stefan-Boltzmann constant</td>
<td>( 5.67 \times 10^{-8} )</td>
<td>( \text{W m}^{-2} \text{K}^{-4} )</td>
</tr>
<tr>
<td>( \tau )</td>
<td>Atmospheric transmissivity (Wang and Mysak, 2000)</td>
<td>0.65</td>
<td>–</td>
</tr>
<tr>
<td>( \lambda )</td>
<td>Heat exchange coefficient</td>
<td>10.0</td>
<td>( \text{W m}^{-2} \text{K}^{-1} )</td>
</tr>
<tr>
<td>( k_1 )</td>
<td>Thermal conductance of snow</td>
<td>0.31</td>
<td>( \text{W m}^{-1} \text{K}^{-1} )</td>
</tr>
<tr>
<td>( \Delta z_1 )</td>
<td>Depth range of subsurface conduction</td>
<td>3.0</td>
<td>m</td>
</tr>
<tr>
<td>( \rho_{\text{air}} )</td>
<td>Density of air</td>
<td>1.2</td>
<td>( \text{kg m}^{-3} )</td>
</tr>
<tr>
<td>( \rho_{\text{ice}} )</td>
<td>Density of ice</td>
<td>910</td>
<td>( \text{kg m}^{-3} )</td>
</tr>
<tr>
<td>( \rho_w )</td>
<td>Density of water</td>
<td>1000</td>
<td>( \text{kg m}^{-3} )</td>
</tr>
<tr>
<td>( C_{\text{DE}} )</td>
<td>Exchange coefficient for latent heat</td>
<td>( 1.0 \times 10^{-3} )</td>
<td>–</td>
</tr>
<tr>
<td>( U )</td>
<td>Wind speed</td>
<td>5.0</td>
<td>( \text{m s}^{-1} )</td>
</tr>
<tr>
<td>( q_s^* )</td>
<td>Sea surface specific humidity</td>
<td>( 0.8 \times 10^{-3} )</td>
<td>( \text{kg kg}^{-1} )</td>
</tr>
<tr>
<td>RH</td>
<td>Relative humidity (Bintanja, 1999, p. 122)</td>
<td>0.75</td>
<td>–</td>
</tr>
<tr>
<td>( B_e )</td>
<td>Equilibrium Bowen ratio</td>
<td>2.0</td>
<td>–</td>
</tr>
<tr>
<td>( R_{\text{air}} )</td>
<td>Gas constant for dry air</td>
<td>287.04</td>
<td>( \text{J kg}^{-1} \text{K}^{-1} \text{yr}^{-1} )</td>
</tr>
<tr>
<td>( \Gamma_{\text{lapse}} )</td>
<td>Atmospheric temperature lapse rate</td>
<td>( -0.012 )</td>
<td>( \text{C m}^{-1} )</td>
</tr>
<tr>
<td>( L_v )</td>
<td>Latent heat of vaporization of ice</td>
<td>( 2.26 \times 10^6 )</td>
<td>( \text{J kg}^{-1} )</td>
</tr>
<tr>
<td>( L_m )</td>
<td>Latent heat of melting of ice</td>
<td>( 0.334 \times 10^6 )</td>
<td>( \text{J kg}^{-1} )</td>
</tr>
<tr>
<td>( C_p )</td>
<td>Specific heat capacity of dry air</td>
<td>1005</td>
<td>( \text{J kg}^{-1} \text{K}^{-1} )</td>
</tr>
<tr>
<td>( C_{\text{ice}} )</td>
<td>Specific heat capacity of ice</td>
<td>2009</td>
<td>( \text{J kg}^{-1} \text{K}^{-1} )</td>
</tr>
<tr>
<td>Symbol</td>
<td>Description</td>
<td>Value</td>
<td>Unit</td>
</tr>
<tr>
<td>---------</td>
<td>---------------------------------------</td>
<td>-------</td>
<td>-----------</td>
</tr>
<tr>
<td>$d_{\text{top}}$</td>
<td>Affected snow/soil layer</td>
<td>0.2</td>
<td>m</td>
</tr>
<tr>
<td>$c_d$</td>
<td>Precipitation constant</td>
<td>3000.0</td>
<td>m</td>
</tr>
<tr>
<td>$\kappa$</td>
<td>Precipitation dependence on temperature</td>
<td>0.0345</td>
<td>K$^{-1}$</td>
</tr>
<tr>
<td>$A_{\text{snow}}$</td>
<td>Tangent hyperbolicus constant</td>
<td>50</td>
<td>–</td>
</tr>
<tr>
<td>$B_{\text{snow}}$</td>
<td>Tangent hyperbolicus constant</td>
<td>0.05</td>
<td>–</td>
</tr>
<tr>
<td>$c_{\text{bal}}$</td>
<td>Calving constant</td>
<td>–2</td>
<td>m yr$^{-1}$</td>
</tr>
<tr>
<td>$\alpha_{\text{snow}}$</td>
<td>Albedo of snow</td>
<td>0.75</td>
<td>–</td>
</tr>
<tr>
<td>$\alpha_{\text{ice}}$</td>
<td>Albedo of ice</td>
<td>0.35</td>
<td>–</td>
</tr>
<tr>
<td>$\alpha_{\text{seaice}}$</td>
<td>Albedo of seaice</td>
<td>0.60</td>
<td>–</td>
</tr>
<tr>
<td>$\alpha_{\text{land}}$</td>
<td>Albedo of land</td>
<td>0.30</td>
<td>–</td>
</tr>
<tr>
<td>$\alpha_{w}$</td>
<td>Albedo of ocean water</td>
<td>0.10</td>
<td>–</td>
</tr>
<tr>
<td>$x_0$</td>
<td>Constant for sea-ice extent</td>
<td>2.1</td>
<td>–</td>
</tr>
<tr>
<td>$x_1$</td>
<td>Constant for sea-ice extent</td>
<td>0.6</td>
<td>–</td>
</tr>
<tr>
<td>$T_{pd}$</td>
<td>Temperature constant for sea-ice extent</td>
<td>–41</td>
<td>°C</td>
</tr>
<tr>
<td>$C_{si}$</td>
<td>Latitudinal shift constant for sea-ice extent</td>
<td>19</td>
<td>°</td>
</tr>
</tbody>
</table>
Fig. 1. Set-up of the model. Upper: Large-scale box model consisting of low (0–30°C S), middle (30–60°C S) and high (60–90°C S) latitude cells. Each compartment is forced by shortwave (SW) and longwave (LW) radiation at the top of the atmosphere and sensible heat transport by eddies (HT), as well as latent heat transport induced by evaporation and snowfall (LH). The two lower latitude boxes are described by one general temperature (T) and albedo (α). Lower: The high-latitude, Antarctic box is subdivided into smaller grid cells with a resolution of 0.5°C latitude. For each cell the energy and mass balances are solved for surface and atmospheric temperatures (Ts and Ta, respectively). Fluxes include incoming and outgoing shortwave radiation at the top of atmosphere (SWa) and at the land/ice/snow surface (SWc); reflected longwave radiation at the top of atmosphere (LWa); longwave (LW), sensible heat (SH) and latent heat of evaporation (LHeva) fluxes between the surface and atmosphere; latent heat of snowfall in atmosphere (LHsnow); heat flux into underlying bedrock (Fs). In all boxes ice flow velocities and ice height are computed, depending on the mass balance, local temperature (T), albedo (α) and isostasy.
Fig. 2. Hysteresis experiment. Starting from no-ice conditions and high $pCO_2$ (orange solid) or starting from full ice sheet and low $pCO_2$ (green dashed). Upper panel shows hysteresis including orbital forcing, lower panel without orbital variations. Rate of $pCO_2$ change is 250 ppm/5 Myr, comparable to 280 ppm/5 Myr used by Pollard and DeConto (2005).
Fig. 3. Constant $pCO_2$ experiments in the Middle Miocene (14.2–13.2 Ma). (a) Annual mean insolation at 70° S. (b) Summer mean insolation at 70° S. (c) Resulting ice-volume variations of four typical $pCO_2$ forcing (220 ppm (black), 390 ppm (blue), 400 ppm (green) and 410 ppm (red). (d) Mean ice volume (dot) and standard deviation (arrows) of large (black/blue) and small (red) ice sheets, defined by their $pCO_2$ level. The blue rectangle encompasses the Middle Miocene glaciation period as depicted by oxygen isotope records.
Fig. 4. Cross-section of large (left) and small (right) Antarctic ice sheets. Color scale corresponds to annual mean ice temperatures.
Fig. 5. Ice-sheet variations correlated to insolation at latitudes between 0 and 90° S. Insolation is shifted backwards per 1 ka on horizontal axis (ice volume lags insolation). Correlation coefficients are given for a small (a and c) and large (b and d) ice sheet and for annual (a and b) and summer mean (c and d) insolation. Best correlation is found for latitudes around 70° S and a shift of 5-6 ka (small ice sheet) or 2 ka (large ice sheet). Highest correlation coefficients for a small ice sheet are 0.49 and 0.60, for annual and summer mean insolation respectively. Maxima for a large ice sheet are 0.29 (annual) and 0.75 (summer mean insolation).
Fig. 6. $p$CO$_2$ sensitivity experiments – level of initial and final $p$CO$_2$ (red colors) and speed of $p$CO$_2$ decrease (blue colors). Colors in upper panel show $p$CO$_2$ forcing and correspond to ice-volume transition in lower panel. Blue box indicates approximate Antarctic glaciation as retrieved from sedimentary records. Glaciation is independent from initial and final $p$CO$_2$ levels (red to orange). On the contrary, the speed of the $p$CO$_2$ drawdown is important. Extremely slow drop in $p$CO$_2$ (dark blue) results in delayed ice-sheet extension, relatively slow decrease (light blue) causes appropriate timing with glaciation. A $p$CO$_2$ drop of 20 ppm/50 ka or faster (for example 20 ppm/4 ka in orange) give the same ice-sheet transition as 20 ppm/50 ka.
Fig. 7. $p\text{CO}_2$ sensitivity experiment – timing of $p\text{CO}_2$ decrease. Colors in upper panel show $p\text{CO}_2$ forcing and correspond to ice-volume transition in lower panels. Blue box indicates approximate Antarctic glaciation as found in sedimentary records. Green/orange curves result from fast $p\text{CO}_2$ transition (20 ppm/4 ka); purple/blue ones from a slow drop (20 ppm/200 ka). The center lines (black text in legend) indicate best fitting solutions (see Discussion section). Orange and dark blue curves correspond to their counterparts in Fig. 6.
Fig. 8. Forced 100-ka $p$CO$_2$ cycles. (a) Normalized eccentricity (grey). (b) and (c) $p$CO$_2$ scenarios (colors) with mean levels of 420 ppm and amplitude of 40 ppm. Maxima in $p$CO$_2$ are tuned to maxima in eccentricity (best fit in green). Resulting glacial-interglacial ice-volume cycles. Normalized power spectral densities, with a strong 100-ka periodicity. Spectral analyses were performed using function pwelch in MATLAB 7.3.0.