The Roles of CO2 and Orbital Forcing in Driving Southern Hemispheric Temperature Variations during the Last 21 000 Yr*

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Abstract (Summary)

Transient climate model simulations covering the last 21 000 yr reveal that orbitally driven insolation changes in the Southern Hemisphere, combined with a rise in atmospheric pCO^sub 2^, were sufficient to jump-start the deglacial warming around Antarctica without direct Northern Hemispheric triggers. Analyses of sensitivity experiments forced with only one external forcing component (greenhouse gases, ice-sheet forcing, or orbital forcing) demonstrate that austral spring insolation changes triggered an early retreat of Southern Ocean sea ice starting around 19-18 ka BP. The associated sea ice-albedo feedback and the subsequent increase of atmospheric CO2 concentrations helped to further accelerate the deglacial warming in the Southern Hemisphere. Implications for the interpretation of Southern Hemispheric paleoproxy records are discussed. [PUBLICATION ABSTRACT]

Full Text (6886 words)

[Headnote]

ABSTRACT

Transient climate model simulations covering the last 21 000 yr reveal that orbitally driven insolation changes in the Southern Hemisphere, combined with a rise in atmospheric pCO^sub 2^, were sufficient to jump-start the deglacial warming around Antarctica without direct Northern Hemispheric triggers. Analyses of sensitivity experiments forced with only one external forcing component (greenhouse gases, ice-sheet forcing, or orbital forcing) demonstrate that austral spring insolation changes triggered an early retreat of Southern Ocean sea ice starting around 19-18 ka BP. The associated sea ice-albedo feedback and the subsequent increase of atmospheric CO2 concentrations helped to further accelerate the deglacial warming in the Southern Hemisphere. Implications for the interpretation of Southern Hemispheric paleoproxy records are discussed.

(ProQuest: ... denotes formula omitted.)

1. Introduction

Ratios of heavy-to-light oxygen and hydrogen isotopes (δ^sup 18^O, SD) stored in the ice matrix of Antarctic ice cores reflect the local temperature evolution over the last eight glacial cycles (Augustin et al. 2004; Jouzel et al. 2007). With the exception of oxygen isotopes recorded in the Taylor Dome ice core (Steig et al. 1999), temperature proxies from different Antarctic sites, such as the European Project for Ice Coring in Antarctica (EPICA) Dome C (Augustin et al. 2004), Dronning Maud Land (Barbante et al. 2006), Vostok (Petit et al. 1999), Dome Fuji (Watanabe et al. 2003; Kawamura and et al. 2007), and Siple Dome (Brook et al. 2005), as well as sea surface temperature (SST) proxies from the Southern Ocean (Pahnke et al. 2003; Stott et al. 2007; Lamy et al. 2007), consistently show that the major rise in local temperatures during the last glacial termination began between 20 and 17.5 ka BP and ended around 9-10 ka BP. Within the same time period, atmospheric CO2 concentrations measured in air bubbles trapped in ice cores rose from glacial to interglacial concentrations by about 80 ppmv (Monnin et al. 2001). Furthermore, sea ice proxies, derived from sea salt fluxes in Antarctic ice cores (Wolff et al. 2006; Fischer et al. 2007) (see Fig. 1) and marine sediment cores (Shemesh et al. 2002; Schneider-Mor et al. 2005), indicate that Antarctic sea ice started to retreat around 20-17 ka ago. It is still unclear what processes controlled the timing of the last glacial termination in the Southern Hemisphere and the corresponding CO2 increase.

Previous ice-core analyses demonstrated that the glacialinterglacial swings recorded in isotope data from Antarctica are orbitally paced (Petit et al. 1999). In Augustin et al. (2004) and Kawamura and et al. (2007), it was suggested that Northern Hemispheric summer insolation leads Antarctic temperature, as recorded at EPICA and Dome Fuji by a few millennia. Whereas the statistical coherency between Northern Hemispheric summer forcing and Antarctic temperatures is well established (Jouzel et al. 2007), a viable physical mechanism that transports the Northern Hemispheric climate signal to Antarctica has not been proposed yet.

An alternative pacemaking mechanism involving local spring insolation changes in the Southern Hemisphere was recently suggested in Stott et al. (2007). As demonstrated in Fig. 1, the local austral spring insolation changes at 63°S, computed by using a fixed season length, correlate well with sea-salt-based reconstructions of sea ice around Antarctica (Wolff et al. 2006; Fischer et al. 2007). Issues, however, arise from the different possibilities of defining seasons (Joussaume and Braconnot 1997; Timm et al. 2008). The differences between a fixed-length season and a fixed-angle season are shown for the austral spring insolation at 63°S in Fig. 2. While there are significant differences in the amplitude and timing of the austral spring insolation curves, they both support the notion that spring insolation changes of up to 15-30 W m^{-2} could have contributed to the Southern Hemispheric deglacial temperature rise between 20 and 10 ka BP. Starting from 18 ka BP, the austral spring insolation shows a very significant increase that coincided with major reorganizations of climate and marked the onset of the last glacial termination in the Southern Hemisphere.

Another possibility for regional orbital pacemaking was recently proposed in Huybers and Denton (2008). These authors argue that the length of the austral summer season may provide an important climate forcing for Southern Hemispheric climate. In fact, changes in the length of the boreal summer season affect the boreal spring insolation anomalies using a fixed-season-length framework Stott et al. (2007). Hence, there appears to be a significant overlap between the proposed orbital forcing mechanisms of Antarctic change in Stott et al. (2007) and Huybers and Denton (2008).

On precessional time scales, the time series of fixedseason austral spring insolation at 63°S (Fig. 1), boreal summer insolation at 63°N, and summer duration at 77°N are all highly coherent with the EPICA SD-based temperature reconstruction (Jouzel et al. 2007) (not shown), attaining values for the squared coherence of up to 0.6. Distinguishing which of these different forcing scenarios is responsible for the onset of the deglacial temperature rise in Antarctica and the Southern Ocean based solely on statistical inferences is virtually impossible.

To further quantify the effects of different external forcings on the climate evolution of the Southern Hemi- sphere, it is essential to understand the seasonal sensi- tivities of ice-core records. Isotope-based temperature reconstructions (Petit et al. 1999; Jouzel et al. 2007) implicitly assume that the ice-core isotope variations capture annual mean signals. Numerous modeling re- sults have provided independent support for a rela- tionship between isotope variations and annual mean temperatures (Jouzel et al. 2003, 2007). If this assump- tion is correct, the oxygen and hydrogen isotope varia- tions would be sensitive to CO_2 and the annually aver- aged local insolation changes associated with the ob- liquity cycle, as well as other annual mean climate signals that may originate from a nonlinear rectification of the precessional signal (Huybers and Wunsch 2003). In fact, the obliquity cycle is one of the major spectral components of isotope variations in Antarctic ice cores (Kawamura and et al. 2007) and is particularly strong in the deuterium excess records from Vostok (Vimeux et al. 2001) and Dome Fuji (Watanabe et al. 2003). However, as shown in Jouzel et al. (2007, their figure S7), the annual mean orbital forcing due to obliquity is leading the obliquity signal in δD by about 5-6 ka. Why changes of annual mean insolation of about 5 W m^{-2} would result in an Antarctic warming 5 ka later is still unresolved. In fact, this result could either imply that isotopic variations of hydrogen and oxygen in Antarctica are influenced by remote (advective) effects, as described (Vimeux et al. 1999), or that more complicated seasonal relationships between orbital forcing, climate, and isotopes have to be invoked.

Kawamura et al. (2007) and Bender (2002) discuss how local summer insolation over Antarctica controls the physical properties and metamorphism of near-surface snow, which in turn determines the magnitude of the O_2/N_2 fractionation used for orbital tuning of the Dome Fuji ice core. While the effect of local summer insolation on the O_2/N_2 fractionation is clearly visible, this forcing does not seem to affect the $\delta^{18}O_{ice}$ temperature signal very significantly. It is important to note that $\delta^{18}O_{ice}$ and δD are most sensitive to the temperature above the atmospheric inversion layer, where the precipitation forms. Unlike the locally generated O_2/N_2 variations, $\delta^{18}O_{ice}$ and δD represent a convolution of regional, seasonally integrated, and also remote hydrological and thermal processes. A possible explanation for the discrepancy between O_2/N_2 and $\delta^{18}O_{ice}$ might be that surface temperature variations could be somewhat decoupled from the temperatures above the inversion layer. Such a decoupling would occur if the isotope variations represented a signal that is affected by lateral advection of heat and moisture and remote effects. Another way to explain this discrepancy is that $\delta^{18}O_{ice}$ in Dome Fuji and δD in EPICA capture more complicated seasonally weighted or annual mean temperature signals, rather than austral summer signals, as for the O_2/N_2 ratio.

The aim of this paper is to study the first-order local forcing mechanisms of glacial-interglacial temperature variations in Antarctica. By focusing on a relatively large area in our analyses (50°-90°S), more regional features will average out. To understand the climate evolution captured in ice cores, it is important to elucidate the local energy balance as well as lateral heat advection effects. The paper is organized as follows: After a brief description of the climate model used in this study and its preindustrial performance in the Southern Hemisphere (section 2), the main results from transient externally forced climate simulations covering the last 21 000 yr will be presented (section 3). This section also studies the response of sea ice variations to external forcings using simple energy balance arguments. Section 4 summarizes our main results and discusses several open unresolved questions, such as the role of lateral advection of heat and remote forcing of Antarctic climate change.

2. Model description

The diagnostics of our modeling study is based on a series of transient climate change simulations covering the last 21 000 yr; these simulations were conducted with the earth system model of intermediate complexity ECBilt-Clio (Goosse et al. 2002). The atmospheric component of the coupled model is version 2 of ECBilt, a spectral T21, three -level quasigeostrophic model that is extended by diagnostic estimates of the neglected ageostrophic terms. The model contains a full hydrological cycle that is closed over land by a bucket model for soil moisture. Synoptic variability associated with weather patterns is explicitly computed. Diabatic heating due to radiative fluxes, the release of latent heat, and the exchange of sensible heat with the surface are parameterized, and cloudiness is prescribed. The sea iceocean component CLIO (Goosse et al. 1999) consists of a free-surface primitive equation model with 3° × 3° resolution coupled to a thermodynamic-dynamic sea ice model.

We use a slightly modified version of ECBilt-Clio (Timm and Timmermann 2007), which has a higher climate sensitivity than the standard version with respect to CO2 forcing. The intent is to mimic the general climate response to CO2 forcing in accordance with the majority of the Intergovernmental Panel on Climate Change's (IPCC) Fourth Assessment Report (AR4) models. The simulated global mean temperature response of our model version to a CO2 doubting amounts to about 2°-3°C (Timm and Timmermann 2007). As shown in Fig. 3, the preindustrial control simulation is capable of simulating Southern Hemispheric climate quite realistically. The simulated and "observed" temperature, sea ice, and precipitation climatology from the preindustrial control run and the 40-yr European Centre for Medium-Range Weather Forecasts Reanalysis (ERA-40; Kållberg et al. 2004) agree well with each other, except for the amplitude of the sea ice area that is somewhat overestimated in ECBilt-Clio. More information on the preindustrial-present-day performance of ECBilt-Clio (LOVECLIM) can be found in Justino et al. (2005), Goosse et al. (2001), and Petoukhov et al. (2005).

To quantify the global climate response to timevarying glacial-interglacial boundary conditions, a series of transient model simulations was conducted using ECBilt-Clio. The simulations started from an equilibrated Last Glacial Maximum (LGM) simulation (Timm and Timmermann 2007).

In the ALL simulation (Timm and Timmermann 2007), the ICE-4G ice-sheet orography (Peltier 1994) and the corresponding land-ice albedo variations were prescribed globally in 1000-yr-long chunks. The greenhouse gas (GHG) forcing is prescribed in 1000-yr-long chunks during the transient model simulation, which includes changes in CO2, CH^sub 4^, and N^sub 2^O concentrations (see Fig. 4). Concentration values were estimated from the Taylor Dome ice core in Antarctica (Indermühle et al. 1999; Smith et al. 1999). The high-resolution record of the Holocene was spliced with the lower sampled record of the LGM/transition period. The data are aligned to the Greenland Ice Sheet Project Two (GISP2) time scale by using methane and oxygen isotope synchronization. Both CH^sub 4^ and N^sub 2^O were measured in samples from the GISP2 ice core (Brook et al. 1996; Sowers et al. 2003). The most relevant greenhouse gas changes are associated with CO2, which corresponds to an estimated radiative forcing of about -2 W m^sup -2^ during the LGM (compared to -0.22 and -0.25 W m^sup -2^ for CH^sub 4^ and N^sub 2^O, respectively). As shown in Fig. 4, CO2 forcing starts to rise at 17 ka BP, which is basically consistent with Monnin et al. (2001). The uncertainties involved with the CO2 age scale (Monnin et al. 2001) are less than the 1000-yr chunk length for the CO2 forcing in our model. Recent assessments of firn densification models (Loulergue et al. 2007) have helped to further reduce the uncertainties in gas-ice ice-age differences. Potential future adjustments of the timing of the observed deglacial CO2 rise might have some influence on the simulated timing of the onset of the glacial termination, but are not likely to affect our main conclusions.

Furthermore, the vegetation (i.e., forest fraction) mask in ECBilt-Clio was modified in land-ice-covered regions, assuming vegetation-free surfaces. The orbitally driven time evolution of solar insolation is computed from Berger (1978). More technical details on this simulation and its initial state can be found in Timm and Timmermann (2007). Millennial-scale forcing mechanisms associated with freshwater perturbations during Heinrich event I, meltwater pulse Ia, and the Younger Dryas were not explicitly included in the model simulations. Therefore, the model did not simulate such features as the Antarctic Cold Reversal, the Younger Dryas, and the Bølling-Allerød. However, it should be noted that some relatively weak millennial-scale forcing is present in the ice-sheet topography and the radiative forcing associated with the methane concentrations.

To further separate the effects of the individual forcings on the climate evolution in the Northern and Southern Hemispheres, a series of transient sensitivity experiments was conducted and analyzed. Experiment ORBonly incorporated the time-varying orbital forcing, while all other boundary conditions were fixed at LGM values. In experiment GHGonly, only the observed temporal evolution of the atmospheric greenhouse gas concentrations was prescribed. Orographic forcing and orbital forcing were fixed to LGM values. Experiment TOPOonly is forced only by the time-varying orographic and ice-sheet albedo effects in both the Northern and Southern Hemispheres. All experiments conducted here start from an equilibrated LGM state that is very similar to the LGM state described in Timmermann et al. (2004), Timm and Timmermann (2007), Roche et al. (2007), and Menviel et al. (2008).

To save computing time and reduce internal model drifts, all experiments use an acceleration factor of 10. For more details on the acceleration technique and its effects on the Southern Hemisphere, see Timm and Timmermann (2007) and Lunt et al. (2006).

3. Results

The ALL simulation covers the climate evolution of the last 21 000 yr and can be directly compared to existing paleoproxy data. The simulated surface temperature history over Antarctica during the glacialinterglacial transition shows pronounced seasonality1 (Fig. 5). Whereas the summer to winter seasonal temperature variations do not match the ice-core data, the simulated austral spring and, to a lesser extent, the annual mean temperature evolution track the ice-core temperature reconstruction (Jouzel et al. 2007) reasonably well. In austral spring the model is capable of simulating the onset of deglacial warming at around 18-17 ka BP, as well as the stable Holocene. However, it should be noted here that due to the use of the acceleration technique, these runs are not yet in full equilibrium with the forcing and might exhibit spurious Holocene trends. For a more direct comparison between model results and ice-core records, we computed the simulated precipitation-weighted temperature (Figs. 5c and 5d). The resulting time series correlates well with the simulated annual mean temperature, thereby supporting previous ice-core interpretations. However, the modelderived pseudoproxy does not match the ice-core temperature evolution better than the simulated annual mean. Again, this might be partly due to the fact that the Southern Ocean response in our accelerated simulations lags the forcing by several centuries (Timm and Timmermann 2007).

Because millennial-scale forcing associated with Heinrich event I was not included, the bipolar seesaw response to an Atlantic Meridional Overturning Circulation (AMOC) weakening at around 17 ka cannot be reproduced. Figure 5 suggests that the deglacial temperature rise in Antarctica and the surrounding oceans can be generated without the bipolar seesaw response to Northern Hemispheric millennial-scale forcing. However, as will be shown below, the atmospheric CO2 increase around 17 ka BP plays a key role in the deglacial temperature rise in Antarctica and the Southern Hemisphere. Under preindustrial conditions, Menviel et al. (2008) simulated a 20-ppmv increase of atmospheric CO2 in response to a complete shutdown of the AMOC. The mechanism invokes vegetation changes in the tropics and marine CO2 buffering mainly due to increased solubility. It might explain the atmospheric CO2 increase (Monnin et al. 2001) around 17 ka that coincided with a reconstructed weakening of the AMOC (McManus et al. 2004). Hence, a significant fraction of the deglacial warming shown in Fig. 5a might indirectly be associated with millennial-scale climate change in the Northern Hemisphere and, in particular, with Heinrich event I.

Figure 5b shows the simulated seasonal temperature evolution near Greenland in comparison with the North GRIP (NGRIP) δ^sup 18^O ice core data (NGRIP and members 2004). Except for the millennial-scale features associated with Heinrich event I, Bølling-Allerød, and the Younger Dryas, simulated summer temperatures reproduce the δ^sup 18^O signal reasonably well. Even the early Northern Hemispheric warming between 21 and 16 ka BP described in Alley et al. (2002) is simulated well in response to local orbital forcing as can be demonstrated by the ORBonly summer temperature evolution (not shown). A model-derived precipitation-weighted temperature over Greenland does not track the NGRLP δ^sup 18^O record very well (Fig. 5d) but reproduces the simulated annual mean temperature evolution.

The time series in Fig. 5 can be further decomposed into the contributions from the different local and remote forcings. The greenhouse gas effects on Southern Hemispheric seasonal mean temperatures are obtained from the GHGonly sensitivity experiment. Figure 6a shows that the first significant increase in the CO2 concentration occurring around 17 ka BP leads to a rapid wanning of about 1°C about 1-2 ka later than in the ORBonly experiment. While austral fall, winter, and spring temperatures show a very similar evolution throughout the GHGonly simulation, the temperature changes for the summer season are significantly smaller. Overall, the CO2 forcing accounts for about 80% of the glacial-interglacial annual mean warming in ALL between 20 and 10 ka BP. It should be noted here that a fraction of the Holocene trends observed in GHGonly and TOPOonly can be attributed to the acceleration technique, rather than to the external forcing itself.

Albedo changes in the Southern Hemisphere represent primarily changes in the sea ice area. The GHGonly experiment simulates a substantial reduction of the sea ice area in response to greenhouse gas forcing, with only small seasonal variations. This demonstrates that the sensitivity of Southern Ocean sea ice to greenhouse gas forcing is only marginally dependent on the seasonally varying mean sea ice area. Similar conclusions can be drawn from the TOPOonly experiment (Fig. 6d). While some observational evidence points to a local thickening of the Antarctic ice-sheet during the Holocene (Parrenin et al. 2007), the prescribed spatially averaged thinning of ICE -4G of 500 m leads to a net warming of about 3°C, in rough accordance with mean lapse rate estimates. As shown in Justino et al. (2006) the difference between the latest ICE-5G orographic reconstruction (Peltier 2004) and the previous ICE-4G reconstruction (Peltier 1994) can account for Southern Hemispheric temperature anomalies in ECBilt-Clio that amount to less than 1°C for the region 90°-60°S. For both topographic reconstructions, the thinning of the Antarctic ice sheet during the early Holocene plays an important role for glacial-interglacial sea ice retreat (Fig. 6d) as well as for annual mean and seasonal polar warming in the Southern Hemisphere. It should however be noted that there still exist significant uncertainties for reconstructions of LGM ice-sheet topography for central Antarctica.

Only orbital forcing leads to significant differences in the seasonal response. Its effects on the temperature and sea ice evolution are shown in Figs. 6e and 6f, respectively. The strongest response in ORBonly occurs in austral spring with a maximum temperature anomaly of about 47deg;C occurring around 9 ka BP. During the Holocene a spring cooling trend can be observed in the Southern Hemisphere that partly offsets the warming trends induced by topographic and greenhouse gas changes (Figs. 6a and 6c) and stabilizes the Holocene temperature evolution. The sea ice evolution (Fig. 6f) does not directly follow the simulated temperature evolution. In particular, the strong summer sea ice response is not directly matched by a similarly strong summer temperature change. Annual mean, austral spring, and summer and winter sea ice start to retreat around 18 ka, about 1 ka earlier than in the GHGonly simulation.

The orbitally forced glacial-interglacial sea ice changes in ORBonly are small compared to the GHGonly response. However, they still play an essential role in generating the early deglacial warming that is evident in many marine paleoproxy records from the Southern Hemisphere (Pahnke et al. 2003; Stott et al. 2007; Lamy et al. 2007).

Under present-day conditions, there is a typical phase lag of about 6-8 weeks between sea ice area and areaaveraged surface temperature (see Fig. 3). Due to the heat capacity of sea ice, it integrates the heat fluxes, thereby generating a slightly delayed response. Because we focus on large-scale averages (zonal average from 90° to 60°S for temperature and 80° to 50°S for sea ice) sea ice transport changes can be neglected. This delayed response between sea ice and air temperature on orbital time scales is further illustrated in Fig. 7. Temperature anomalies lead the sea ice anomalies by about 50-60 days, but eventually the resulting changes in sea ice albedo will also affect the temperature evolution, thereby providing a positive feedback.

To further disentangle how orbital forcing influences the near-surface temperatures, we decompose the absorbed shortwave radiation Q^sub abs^ into ..., where i, t represent the season and the year, respectively. The incoming shortwave radiation, taken here at a reference latitude of 63°S, is denoted as Q^sub inc^. Bared quantities represent a long-term climatology averaged over 21 000 yr, whereas primed quantities characterize the orbitally and seasonally varying anomaly part. For reasons of simplification we assume an idealized annual mean and time-averaged cloud-cover weighted cloud albedo of α^sub cloud^_ = 0.45. The surface albedo is decomposed into a long-term climatology, α(i), and a seasonally and orbitally varying contribution, α(i, t)'. As shown in Fig. 8a, insolation variations in spring and summer oppose one another over the delaciation and Holocene. Fall, winter, and annual mean solar forcings are small and shall be neglected in our discussion. The simulated albedo changes illustrated in Fig. 6f lead to large seasonal anomalies in absorbed shortwave radiation. Summer and spring are the dominant terms because they constitute the largest climatological incoming shortwave radiation changes (using our fixed season-length definition).

The decreased sea ice extent in austral spring and during summer between 18 and 6 ka BP is accompanied by warming of the ocean mixed layer and an increase in surface air temperatures. Moreover, the spring and summer sea ice retreat between 18 and 6 ka BP is itself partly generated by the temperature changes during the preceding 2 months. The sum of the shortwave contributions shown in Fig. 8c closely matches the temperature evolution in Fig. 6e, providing further diagnostic support for the causalities identified here. These feedbacks are illustrated schematically in Fig. 9.

The simulated spring insolation changes of net shortwave radiation in ORBonly provide a significant contribution to the annual mean net shortwave radiation changes simulated by the ALL experiment. Their amplitude attains values of up to 13 W m^sup -2^ (Fig. 8), which is comparable to the magnitude of the glacial-interglacial differences of annual mean shortwave radiation (see Fig. 10) in ALL. As can be seen from Fig. 10 (note the different scales on the ordinates), the shortwave radiation changes outweigh the changes in longwave radiation on an annual mean and also seasonal (not shown) basis. Due to the retreat of sea ice in ORBonly and TOPOonly, more longwave radiation is emitted from the surface. The resulting regionally averaged longwave flux amounts to about ~0.5 W m^sup -2^. In GHGonly the effect of sea ice retreat is counteracted by the overall increase of backscattered longwave radiation due to the increasing CO2 concentrations. The result is a glacial and interglacial net longwave heat flux change into the ocean of less than 0.5 W m^sup -^2, which is more than an order of magnitude less than the shortwaveinduced warming in GHGonly. Figure 10 also reveals from the perspective of the net shortwave fluxes that a stable

Holocene temperature evolution requires a reduction of annual mean shortwave radiation, which is largely due to the reduction of springtime insolation and the corresponding sea ice albedo response (Fig. 8).

4. Summary and discussion

The seasonally stratified analyses of the ALL simulation and the sensitivity experiments have revealed that orbital forcing affects the sea ice extent around Antarctica. The resulting changes in the absorption of shortwave radiation lead to temperature changes that feed back onto the sea ice extent. This process works to amplify, in particular, the austral spring temperature changes due to orbital forcing alone. Spring solar inso- lation forcing increases substantially between 20 and 10 ka BP, leading to a reduction in sea ice area, and in- creased absorption of shortwave radiation. This warms the surface and leads to further sea ice reduction. The spring cooling trend from 10 ka to the present partially offsets the effects of ice-sheet thinning (10-7 ka BP) and increasing greenhouse gas concentrations. In the model simulations the result is a relatively stable Holocene temperature evolution over and around Antarctica. Although shorter and more regional-scale features of Antarctic climate variability cannot be simulated with our model setup, the low-frequency externally forced behavior is simulated in accordance with ice core data. Furthermore, recent paleoreconstructions of sea ice area changes (Shemesh et al. 2002; Bianchi and Gersonde 2004; Wolff et al. 2006) around Antarctica are consistent with the idea that spring insolation forcing is an important pacemaker for glacial-interglacial Southern Hemispheric climate change, as is further illustrated in the supplementary figure S5 in Stott et al. (2007).

Still an unresolved question is whether sea ice variations around Antarctica can significantly affect temperatures in central Antarctica. To quantify the effects of sea ice and temperature changes in the Southern Ocean on central Antarctic temperatures via lateral heat advection, we conducted an additional partially coupled sensitivity experiment with ECBilt-Clio using preindustrial boundary conditions. In a region between 60°S and 50°S, the atmospheric model is forced by a zonally homogenous 2°C temperature anomaly that is added onto the preindustrial SST climatology. Elsewhere, the model is fully coupled. Mimicking a 5°-10° meridional shift in the annual mean sea ice margin, the resulting 2-m temperature anomaly pattern (Fig. 11) clearly demonstrates the importance of lateral heat advection from the Southern Ocean to central Antarctica. In this relatively coarse-resolution model, a Southern Ocean warming of about 2°C results in temperature changes of 1°C in Antarctica. However, it should also be noted that local radiative forcings play a key role in generating glacial-interglacial climate variations in central Antarctica. To further assess and quantify the relative roles of remote versus local forcing in this region, higherresolution modeling studies (Noone and Simmonds 2004) are needed. In the analysis presented above, the spatial averaging over a relatively large area (90°S-60°S) averages out some sea ice advection and lateral advection effects.

Oxygen and hydrogen isotope ratios measured in ice cores in Antarctica represent a proxy for the precipitationweighted condensation temperature. Understanding isotope variations in terms of a precipitation-weighted temperature requires an accurate knowledge of the seasonal cycle of precipitation and its changes through time. Even for present-day conditions, there exists a troubling uncertainty in observing or diagnosing an accurate snowfall climatology (Cullather et al. 1998) in central Antarctica. On orbital time scales the seasonal cycles of both temperature and precipitation can be subject to change. In fact our model simulates very significant long-term changes of the precipitation climatology around Antarctica (not shown). On the other hand, it has been argued (e.g., Krinner and Werner 2003; Jouzel et al. 2003, 2007) that the effect of seasonality changes of precipitation on precipitation-weighted temperatures is likely to be weak in most parts of central Antarctica. While our model-based inferences are far from conclusive, they suggest that simulated precipitation-weighted temperature tracks the simulated annual mean temperature reasonably well, supporting the annual-mean interpretation of isotopic records from Antarctica. Nevertheless, on regional scales a better understanding of the seasonality issue could provide further insight into the forcing mechanisms of deglacial climate change.

Several caveats and uncertainties of our modeling approach need to be mentioned. The CO_2 sensitivity of our climate model greatly determines the amplitude of the temperature signal in GHGonly and ALL. A weaker sensitivity would influence the magnitude of the glacial-interglacial temperature change as well as the time evolution during the Holocene. Furthermore, significant uncertainties exist for the orographic forcing in Antarctica. Different orographic reconstructions (ICE4G, ICE-5G) might induce significant regional errors and uncertainties that can average up to 1°C or more in zonally averaged Southern Hemispheric temperatures. As has already been mentioned, the acceleration technique, while on the one hand reducing internal model drift, can, on the other hand, generate spurious trends that result from incomplete equilibration with the external forcing (Lunt et al. 2006; Timm and Timmermann 2007).

Within our modeling framework that uses prescribed greenhouse gas forcing, two important questions remain unanswered: What brought about the increase in atmospheric CO_2 concentrations at nearly the same time as the Antarctic temperature rise (Fig. 1) and why has Antarctica experienced glacial-interglacial transitions on a 100-120-ka time scale (Augustin et al. 2004), rather than on precessional time scales? We suggest that the mechanism underlying the link between Antarctic temperature and atmospheric CO_2 rise at glacial terminations involves increased austral spring-to-summer insolation, which causes sea ice retreat in the Southern Ocean during the melting season (Stephens and Keeling 2000) in accordance with available paleoreconstructions of sea ice extent (Shemesh et al. 2002; Bianchi and Gersonde 2004) and increased oceanic ventilation in an expanded Southern Ocean (Spero and Lea 2002; Stott et al. 2007).

In addition, the Northern Hemispheric warming (Fig. 5b) between 21 and 18 ka BP may have destabilized the Laurentide ice sheet, resulting in a major freshwater discharge into the North Atlantic (Heinrich event I) and a subsequent weakening of the Atlantic Meridional Overturning Circulation (McManus et al. 2004). A weakened overturning circulation may further increase the atmospheric CO2 due to oceanic (Marchai et al. 1998) or terrestrial processes (Obata 2007; Menviel et al. 2008). While these Northern and Southern Hemispheric mechanisms of CO2 rise could in principle occur independently from each other, there is some evidence that they have synchronized during glacial terminations. Synchronization of prolonged positive trends between Northern Hemispheric boreal summer insolation and Southern Hemispheric austral summer insolation occurs every 100-120 ka and coincides with the onset of the major glacial terminations (Schulz and Zeebe 2006).

This study discussed the effects of orbital forcing, greenhouse gas changes, orographic forcing, and sea ice-albedo feedbacks on the Southern Hemispheric climate evolution over the past 21 000 yr.

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[Footnote]
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[Footnote]
1 The seasonal model output uses a fixed-length season of 90 days and a year of 360 days. Effects of orbitally varying season length on the temperature evolution in the ECBilt-Clio simulation have been discussed elsewhere (Timm et al. 2008). For consistency reasons the seasonal output will be compared with a fixed-season orbital forcing.

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