Vegetation dynamics amplifies precessional forcing

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Received 2 March 2006; revised 30 March 2006; accepted 3 April 2006; published 12 May 2006.

[1] The astronomical theory of climate variations predicts that the climatic precession which changes the seasonal distance between Earth and Sun does not affect the annual mean irradiation at any given latitude. However, previous modeling studies suggest that during interglacials, the interaction between atmosphere, vegetation and ocean can transform the seasonal forcing by precession into an annual mean global signal. Here, we show that this result can be generalized. A distinct precessional signal emerges in a climate system model over many precessional cycles. While neither the atmosphere-ocean nor the atmosphere-vegetation model are able to produce a large amplitude of global temperature in the precessional band, only the mutual amplification of biogeophysical feedback and sea ice-albedo feedback allows a strong amplification of the precessional signal. **Citation:** Claussen, M., J. Fohlmeister, A. Ganopolski, and V. Brovkin (2006), Vegetation dynamics amplifies precessional forcing, *Geophys. Res. Lett.*, *33*, L09709, doi:10.1029/2006GL026111.

1. Introduction

[2] There exists ample evidence from a variety of climate archives that periodic changes in the Earth orbit and the tilt of the Earth axis affect the climate system. However, it is also evident that orbital forcing does not explain past climate variations in a linear manner. Instead, orbital forcing can be viewed a trigger of changes in the climate system [*Paillard*, 2001] which are amplified by a number of feedbacks in the climate system [e.g., *Berger*, 2001; *Calov et al.*, 2005]. Here, we address the problem of transformation of the precessional signal in orbital forcing into an annual mean climate signal.

[3] It is known that the climatic precession, which reveals a double-peak spectrum at periods of some 23,000 and 19,000 years, changes the insolation with season. On average over a year, however, the mean insolation for any given latitude does not depend on the climatic precession [*Paillard*, 2001]. Nonetheless, some climate archives such as oxygen isotopes in ice cores [*Petit et al.*, 1999] which are assumed to record the annual mean temperature [*Delaygue et al.*, 2000] reveal a significant precessional signal. It has been discussed earlier that variations in ice sheets could amplify the precessional signal [e.g., *Köppen and Wegener*, 1924]. But even in periods of little or no changes in ice sheets such as the present interglacial, the precessional signal appears in paleoclimate records.

[4] There are several modeling studies which have addressed this problem. For example, *Lohmann and Lorenz* [2006] found in their model that boreal summer temperatures follow mainly local orbital forcing, whereas the boreal winter climate is strongly modified by varying modes of atmospheric circulations. On annual and global average, the temperature is only marginally affected by precession. The simulations were undertaken with a coupled atmosphere – ocean - sea ice model which does not include changes in ice sheets nor dynamic vegetation nor changes in atmospheric CO_2 concentrations. Other model studies [e.g., *Ganopolski et al.*, 1998; *Crucifix et al.*, 2002; *Wohlfahrt et al.*, 2004] emphasize the non-linear interaction between atmosphere, ocean, and vegetation dynamics. These studies indicate that a northward shift of boreal forest caused by mid-Holocene summer warming leads to a decrease in the surface albedo during the snowy season (because snow-covered tall vegetation appears to be darker than snow-covered flat vegetation or bare ground) and, hence, further warming [e.g., *Otterman et al.*, 1984]. The sea ice-albedo feedback works in a similar way, as a reduction in sea-ice area leads to a darker and warmer region and to a stronger heat flux from the ocean. If vegetation dynamics and sea-ice dynamics operate in a model, then the biogeophysical, or taiga-tundra, feedback tends to amplify the effect of the sea ice-albedo feedback, and *vice versa*. This so-called synergy between feedbacks could cause a winter-time warming in the mid-Holocene in some boreal regions as reported by *Cheddadi et al.* [1997]. In the model of *Ganopolski et al.* [1998] the synergy is stronger than the pure contribution of vegetation while in the transient simulations by *Crucifix et al.* [2002], the opposite is seen. *Crucifix and Loutre* [2002] yield similar model results for the last interglacial, the Eemian, which persisted approximately from 126 ky BP to 115 ky BP (1000 years before present). They highlight the synergy between snow and vegetation feedbacks as being crucial for the glacial inception at the end of the Eemian.

[5] In a recent paper, *Jahn et al.* [2005] explored the role of vegetation in a model simulation for the last glacial maximum in which low atmospheric CO_2 concentrations and larger ice sheets on the Northern Hemisphere yield a strong global mean cooling in the atmosphere-ocean model. If interactive vegetation dynamics is included, then this cooling is amplified by vegetation dynamics, mainly by the taiga-tundra feedback. Therefore, we suggest that previous results obtained for interglacial and glacial climate could be generalized, and we hypothesize that vegetation dynamics amplifies the precession signal throughout the late Quaternary.

2. Setup of Model Experiments

[6] To test this hypothesis, a number of simulations have been conducted by using the Earth system Model of

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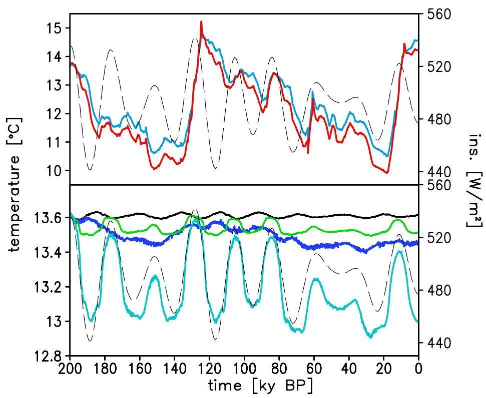

Figure 1. Simulated global annual mean near-surface air temperatures for the last 200 ky BP (ky = 1000 years before present). (top) The red line in depicts results of the model run AOV-IC in which changes in insolation and reconstructed changes in inland ice and atmospheric CO_2 concentration are prescribed to force an atmosphere-ocean-vegetation model. The blue line refers to results from a simulation AO-IC which is the same as AOV-IC with the exception that vegetation cover is kept constant at values simulated for 200 ky BP. (bottom) Results of the atmosphere-only model (black line), the atmosphere-vegetation model (green line), the atmosphere-ocean model (dark blue line), and the atmosphere-ocean-vegetation model (light blue line) in which inland ice, atmospheric CO_2 concentration are kept constant with time (see text). The dashed line shows the monthly mean insolation in June at 60°N.

Intermediate Complexity CLIMBER-2 (see Appendix A). The numerical simulations span the last 200,000 years, i.e., some nine precessional cycles, in which orbitally induced variations in insolation according to *Berger and Loutre* [1991] were prescribed as external forcing. Furthermore, changes in atmospheric CO_2 concentration were prescribed using data by *Petit et al.* [1999]. Finally, a scenario of inland ice variations has been constructed where the total ice volume was supposed to follow the sea-level changes according to *Waelbroeck et al.*, [2002]. The area covered by ice sheets was assumed to follow ice volume being scaled by a power of 2/3. This simple assumption implies a fast change in ice area when the ice volume is small and a slower change with increasing ice volume - in line with modeling results by *Calov et al.* [2005]. At the last and the pen-ultimate glacial maximum, the inland ice was assumed to cover the areas indicated in reconstructions by *Peltier* [1994]. Given the coarse geographical resolution of CLIMBER-2, this simple scenario of past inland ice variations can be viewed as sufficiently realistic.

[7] Three sets of simulations were performed. Firstly, an equilibrium simulation was done in which the coupled atmosphere-ocean-vegetation system is in equilibrium with orbital forcing, constructed inland ice masses and atmospheric CO_2 concentration representative for 200 ky BP (referred to as 200k boundary conditions in the following). Secondly, four transient simulations were performed which span the last 200 ky: an atmosphere-only simulation, labeled A, a simulation with interactive atmosphere and vegetation dynamics, AV, a simulation with interactive atmosphere and ocean dynamics, AO, and a simulation with interactive atmosphere, ocean, and vegetation dynamics, AOV. In these simulations, 200k boundary conditions as well as vegetation cover, sea-surface temperatures and sea-ice cover from the 200k equilibrium simulation were taken in various combinations for simulations A, AV, AO. Thirdly, in two transient simulations, AOV-IC and AO-IC, spanning the last 200 ky, inland ice and atmospheric CO_2 concentrations are varied according to the scenario specified above. In AO-IC, vegetation cover is kept constant at 200k boundary conditions.

3. Results

[8] Figure 1 depicts the global, annual mean, near-surface air temperature obtained from simulation AOV-IC and AO-IC (Figure 1, top). The simulations start at a global mean temperature which is smaller than the (simulated) present-day, pre-industrial temperature. This reflects the assumption that some 200 ky ago, the inland ice masses were slightly larger (by some 5 10^6 km^3 in terms of ice volume) and the atmospheric CO_2 concentration was lower (by some 35 ppmv) than present-day, pre-industrial value. A spectral analysis of AOV-IC (Figure 2) reveals all main orbital periodicities which are also present in the prescribed variations of inland ice and atmospheric CO_2 concentration. In Figure 1 (bottom), the temperature changes resulting from simulations A (black line), AV (green line), AO (dark blue line), and AOV (light blue line) with 200k boundary conditions are depicted together with the 60°N mean June insolation (dashed line).

[9] Obviously, most of the glacial - interglacial global temperature changes can be attributed to (prescribed) variations in ice sheets and atmospheric CO_2 concentration (compare simulations AOV-IC, AO-IC and AOV). When keeping vegetation cover constant, then temperature maxima at relatively warm periods (at 200 ky BP, around 125 ky BP, 102 ky BP, and 85 ky BP) are nearly the same in AOV-IC and AO-IC. Glacial temperatures are, however, smaller in AOV-IC then in AO-IC mainly because the taiga-

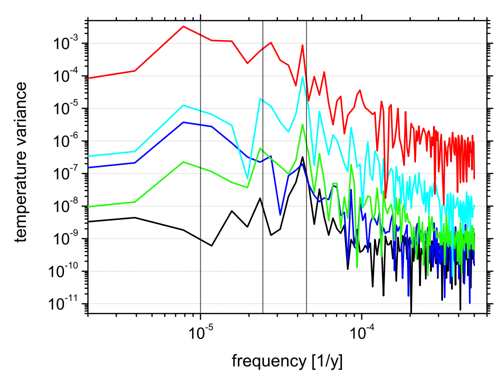

Figure 2. Spectra of simulated global annual mean temperature variance for the last 200 ky BP. The meaning of colors is the same as in Figure 1 (the spectrum of simulation AO-IC is not shown). The vertical dark lines indicate the periods of variations in the Earth's orbit eccentricity (at some 100 ky), tilt of the Earth's axis (at some 41 ky) and climatic precession (around some 22 ky). The spectra are computed using a Fast Fourier Transform.

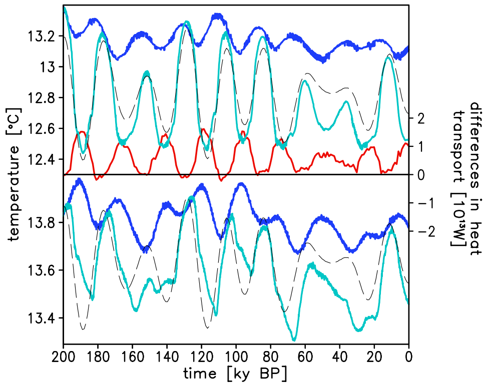

Figure 3. Annual mean temperature on average (top) over the northern hemisphere and (bottom) over the southern hemisphere as a result of simulation AO (boundary conditions, including vegetation cover, are kept constant at values representative for 200 ky BP), dark blue line, and of simulation AOV, light blue line. Also shown is the difference between simulations AOV and AO (red line) in vertically integrated, northward oceanic meridional heat transport at the equator of the Atlantic. As in Figure 1, the dashed lines in both figures show the monthly mean insolation in June at 60°N (here without scale).

tundra feedback amplifies the cooling - in line with earlier simulations by *Jahn et al.* [2005]. Further simulations - not to be discussed here - with prescribed or interactively simulated inland ice extent, atmospheric CO_2 concentration, sea-surface temperatures, sea-ice extent, and vegetation cover in various combinations yield similar results: whenever vegetation dynamics are included, then global glacial climate is cooler than in simulations with vegetation cover prescribed at interglacial values.

[10] To analyze the amplification of the precessional cycle by vegetation dynamics more closely, we focus on simulations A, AV, and AOV. In the atmosphere-only model (simulation A) in which the temperature of roughly 2/3 of the Earth surface is kept constant, a small temperature response with an amplitude of less than 0.05K is found which can be traced back to variations in snow cover over the ice-free part of the continents. If vegetation dynamics are included, then the precessional forcing is clearly visible in the temperature variations with an amplitude of some 0.15 K which then are in phase with June insolation at 60°N. If ocean and sea-ice dynamics are considered (simulation AO) and, hence, sea-surface temperatures are allowed to vary while vegetation cover was kept constant, then the resulting global mean temperature varies much stronger with the eccentricity and much weaker with the climatic precession. The power of temperature variance turns out to be an order of magnitude larger at some 100 ky than around some 22 ky, while the opposite is valid for simulation AV (see Figure 2). Simulation AOV reveals larger temperature amplitudes of some 0.6 K being clearly in phase with the course of 60°N mean June insolation (see Figure 1).

[11] To analyze the differences between simulations AO and AOV, annual mean temperatures on average of the northern and southern hemisphere, respectively, are depicted in Figure 3. In AO, the temperature over the northern hemisphere mainly follows the insolation at high northern latitudes during boreal summer. Over the southern hemisphere, it follows the insolation at high southern latitudes during local (austral) summer which are in antiphase with those at high northern latitudes. On average over both hemispheres, the influence of precessional forcing on temperature appears to nearly compensate such that the precessional signal is barely visible on global scale. In contrast to AO, simulation AOV yields a by-and-large synchronous signal on both hemispheres. Further analysis reveals that changes in surface albedo are much stronger in AOV than in AO. When boreal summer insolation decreases (increases), then the surface albedo increases (decreases) stronger in AOV than in AO. At the same time, the strength of the meridional overturning circulation in the Atlantic decreases (increases) stronger in AOV than in AO. The difference between AOV and AO in oceanic meridional heat transport at the equator of the Atlantic follows accordingly (see Figure 3). This difference is positive (nearly zero or negative) when the temperature over the southern hemisphere attains a relative maximum (minimum) in simulation AO, which explains a stronger synchronization of hemispheric temperatures in simulation AOV than in AO. Hence, we see the same processes and changes in processes at work as already discussed by *Ganopolski et al.* [1998] for mid-Holocene climate.

[12] Variations in zonally averaged tree fraction and grass fraction reveal a clear precessional signal in all latitudes, except for the inner tropics near the equator for which the model shows only small changes in tree cover. Hence the emergence of a precessional cycle cannot be attributed solely to boreal or to subtropical vegetation dynamics. However, there is an interesting difference between simulations AOV and AV. In AOV, the variations in tree fraction and grass fraction at high northern latitudes are larger than those in AV (Figure 4). Furthermore, the variation in fractional sea-ice coverage in the Arctic is stronger in AOV than in AO (not shown here). Hence, it is the synergy between the taiga-tundra feedback and the sea ice-albedo feedback, described in the introduction, which causes the difference in surface albedo and, finally, in global mean temperature between AOV and AO.

4. Summary and Concluding Remarks

[13] Previous studies have shown that the interaction of atmospheric processes, vegetation dynamics and sea ice dynamics, including the so-called taiga-tundra feedback

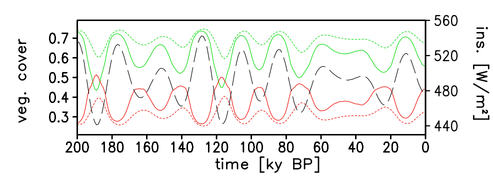

Figure 4. Simulated fractional coverage of land areas by trees (green lines) and by grass (red lines) for the last 200,000 years. Shown are zonal averages over the latitudinal belt between 50°N and 60°N for simulation AOV (full lines) and for simulation AV (dotted lines). Both simulations are forced with 200k boundary conditions. As in Figure 1, the dashed line shows the monthly mean insolation in June at 60°N.

and the sea ice-albedo feedback, can transform the precessional forcing of insolation into an annual mean temperature signal in the precessional band. Here we demonstrate that this result, which was obtained for snap shots of interglacial and glacial climate [*Ganopolski et al.*, 1998; *Jahn et al.*, 2005] or short transient interglacial simulations [*Crucifix et al.*, 2002; *Crucifix and Loutre*, 2002] can be generalized. In a simulation spanning the last nine precessional cycles, the precessional signal in the global mean temperature emerges mainly because of vegetation dynamics. Hence our results suggest that in a climate state without ice sheets or even sea ice, a precessional cycle should be visible in annual mean temperature. Interestingly, neither the atmosphere-ocean nor the atmosphere-vegetation model are able to produce a large amplitude of global temperature changes in the precessional band. Obviously the interaction, or synergy, between biogeophysical feedback and sea-ice albedo feedback allows a strong amplification of the precessional signal.

[14] Finally, a word of caution is warranted: our simulations include only biogeophysical feedbacks. In nature, the precessional forcing might be additionally amplified by the dust feedback and by biogeochemical feedbacks due to variations in terrestrial methane and carbon which in this study are implicitly, and only to some extent, taken into account in the prescribed changes in atmospheric CO_2 concentration.

Appendix A

[15] CLIMBER-2 encompasses a 2.5-dimensional statistical dynamical model of the atmosphere, a multibasin, zonally averaged ocean model, including sea ice dynamics, and a dynamical model of terrestrial vegetation which are coupled by fluxes of energy, water, and momentum [*Petoukhov et al.*, 2000]. The atmosphere model has a coarse resolution of 10° in latitude and about 51° in longitude. The transport equations for temperature and humidity are solved on 10 vertical levels utilizing universal vertical profiles for temperature and humidity. The shortwave and long-wave radiation fluxes are calculated for 16 vertical layers accounting for evolving stratus and cumulus cloud coverage and average aerosol and ozone concentrations. The ocean model has zonally averaged basins for the Atlantic, the Indian Ocean, and the Pacific Ocean which are connected by the Antarctic circumpolar current. The meridional resolution is 2.5°, and the vertical is resolved by 20 layers. Land grid cells have fractions of glacial cover and ice-free surface. The latter is separated into forest, grass and bare soil fractions which interactively evolve under changes in climate. CLIMBER-2 has been validated against present-day climate and tested against comprehensive general circulation models and has been used successfully for a variety of paleoclimate studies (see the list of references in the Table of EMICs [*Claussen*, 2005]).

[16] **Acknowledgment.** The authors appreciate the constructive comments of anonymous reviewers.

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