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A movable trigger: Fossil fuel CO_2 and the onset of the next glaciation

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[1] The initiation of northern hemisphere ice sheets in the last 800 kyr appears to be closely controlled by minima in summer insolation forcing at 65°N. Beginning from an initial typical interglacial pCO$_2$ of 280 ppm, the CLIMBER-2 model initiates an ice sheet in the Northern Hemisphere when insolation drops 0.7 σ (standard deviation) or 15 W/m^2 below the mean. This same value is required to explain the history of climate using an orbitally driven conceptual model based on insolation and ice volume thresholds (Paillard, 1998). When the initial baseline pCO$_2$ is raised in CLIMBER-2, a deeper minimum in summertime insolation is required to nucleate an ice sheet. Carbon cycle models indicate that ~25% of CO_2 from fossil fuel combustion will remain in the atmosphere for thousands of years, and ~7% will remain beyond one hundred thousand years (Archer, 2005). We predict that a carbon release from fossil fuels or methane hydrate deposits of 5000 Gton C could prevent glaciation for the next 500,000 years, until after not one but two 400 kyr cycle eccentricity minima. The duration and intensity of the projected interglacial period are longer than have been seen in the last 2.6 million years.

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1. Orbital Variations and the Onset of Glaciation

[2] Pleistocene climate variability is dominated by the glacial/interglacial cycles documented in ice core and ocean sedimentary records. The timing of the glacial cycles is tightly correlated with variations in the shape of the Earth's orbit. The history of ice volume seems to follow the various beats of solar insolation at 65°N in June or July [*Ruddiman*, 2003b]. A northern hemisphere sweet spot for orbital driving of climate is consistent with the presence of land in northern high latitudes that can support ice sheets, and with sensitivity of ocean circulation in the North Atlantic to atmospheric forcing, affecting ocean heat transport to high latitudes and deep ocean temperature and hence pCO$_2$. The standard deviation of insolation at this time and place is about 20 W m^{-2}, which is large compared with the global mean effect of doubling CO_2, about 4 W m^{-2}. Three components of orbital variation affect the magnitude and distribution of solar insolation: precession at periods of 19 and 23 kyr, obliquity at about 40 kyr, and eccentricity

at 400 and 100 kyr [*Berger*, 1978]. At 65°N, obliquity accounts for about a third of the insolation variability, and precession two thirds. The direct effect of eccentricity is small, but eccentricity regulates the intensity of the precessional variability. The Earth's orbit will be nearly circular in the coming 50 kyr, so that precession effects will be smaller than usual.

[3] The recent history of ice ages is recorded in the $\delta^{18}O$ of $CaCO_3$ precipitated in the deep ocean. $\delta^{18}O$ reflects both ice volume and deep sea temperature, but both components of the signal move in the same direction through glacial cycles, and thus $\delta^{18}O$ is taken as a standard climate state indicator. Through the past 800 kyr, $\delta^{18}O$ seems to respond linearly to obliquity and precessional orbital forcing [*Imbrie and Imbrie*, 1980]. The difficulty with understanding the effect of orbits on climate is that $\delta^{18}O$ variations are stronger in the 100 kyr frequency band, than is the orbital forcing which drives it. This is a recent phenomenon geologically; before ~800 kyr ago the 100 kyr and the 19 and 23 kyr components of ice sheet variability were much weaker [*Imbrie et al.*, 1993].

[4] Simple climate models are able to reproduce this behavior given some nonlinearity in ice sheet response to orbital forcing, presumably by rectifying the precession cycle (the "envelope" of which is determined by eccentricity). The idea is analogous to the operation of a crystal AM radio, which uses a diode to nonlinearly select only the positive parts of a radio signal of varying amplitude, then smoothing it, resulting in variability at a slower (audible) frequency. The *Calder* [1974] model introduces the nonlinearity of differing rate constants for ice accumulation and melting to generate a 100 kyr component to ice sheet variability. The model predicts the timings of climate transitions quite well but deglaciates too easily, and its spectral maximum is weak in the 100 kyr frequency. The model of *Imbrie and Imbrie* [1980] uses a similar ice-growth kinetics nonlinearity to generate output that is stronger at 100 kyr, but generates variability at the other eccentricity frequency of 400 kyr that is not observed in the $\delta^{18}O$ record [*Paillard*, 2001]. *Paillard* [1998] proposed a model in which the nonlinearity arises from insolation and ice volume thresholds governing transitions between distinct climate states. In this model, a transition from interglacial to glacial climate state is driven by a dip of northern hemisphere summer insolation below a threshold value, denoted i_0, about 15 W m^{-2} or 0.75 σ below the mean. After an ice sheet is nucleated, it is not allowed to collapse until, in one formulation, some period of time has passed (the range of 25–60 kyr seemed to work), or in another, until ice volume reaches some critical size. At this point, the model is eligible to deglaciate, triggered by insolation exceeding another threshold value. The apparent 100 kyr cycle arises from 5 ± 1 precession cycles [*Raymo*, 1997]. The model thus formulated is able to generate sufficient power at 100 kyr without spillover into the 400 kyr frequency band [*Paillard*, 2001]. The aim of this paper is to use the intermediate-complexity CLIMBER-2 model, described below, to diagnose how the parameters of the Paillard model should change in response to anthropogenic CO_2 forcing.

[5] The aspect of glacial climate cycles that concerns us is the onset of glaciation. Nucleation of an ice sheet from an interglacial climate is perhaps the easiest part of the glacial cycle to understand and forecast. Glacial onset begins from the interglacial climate state, for which cloudiness, humidity, heat transports, etc. are known (in contrast to the glacial climate state). Also, the nucleation of an ice sheet is easier to understand than its subsequent growth or collapse. Models require some amplifying feedback, from sea ice [*Khodri et al.*, 2001] or the terrestrial biosphere [*DeNoblet et al.*, 1996], to nucleate on the basis of insolation forcing, but insolation is always the primary driver. Paillard's choice of a simple insolation trigger for ice sheet nucleation from an interglacial climate therefore seems a sensible one, by process of elimination; what else is there?

[6] The coincidence of insolation extrema with $\delta^{18}O$ variations in the climate record is very clear, although in fairness it must be said that the chronology of the climate record is tuned to orbital forcing. *Raymo* [1997] showed a correlation between insolation maxima and glacial terminations. Insolation trigger-minimum events result in increasing $\delta^{18}O$, consistent with increasing ice volume or decreasing ocean temperature. Ambiguity in the $\delta^{18}O$ climate proxy makes it difficult to confidently pick out ice sheet nucleation events from an interglacial climate state, as opposed to growth events of already-existing ice sheets. However, all known glacial initiation events are correlated to insolation trigger-minimum events.

[7] The coupled climate-ice sheet model CLIMBER-2 also shows this insolation trigger for glacial nucleation, at very close to the insolation value required by Paillard. The CLIMBER-2 model

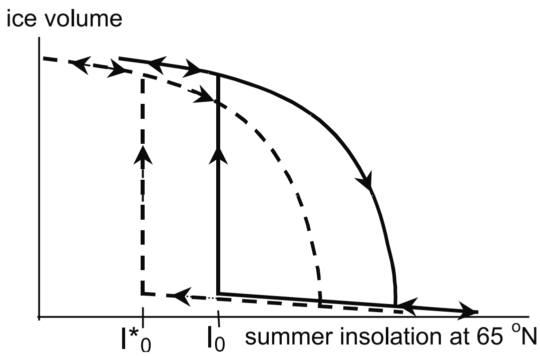

Figure 1. Schematic stability diagrams for two different atmospheric pCO_2 values, showing dependence of the Northern Hemisphere total ice volume on summer solar insolation in the CLIMBER-2 model. The dashed line corresponds to higher CO_2 concentration than the solid line. i_0 and i_0* are the corresponding threshold values of the summer insolation. Stability diagrams are obtained by very slowly changing of eccentricity resulting in the gradual decrease of summer solar insolation at 65°N. (Based on Calov and Ganopolski, manuscript in preparation, 2005).

consists of a coarse resolution atmosphere-ocean-vegetation component [*Brovkin et al.*, 2002; *Petoukhov et al.*, 2000] coupled with the high-resolution 3-D thermomechanical ice sheet model SICOPOLIS [*Greve*, 1997]. The climate and ice sheet components are coupled bidirectionally using a physically based energy and mass balance interface described in detail by *Calov et al.* [2005]. The coupled climate-ice sheet version of CLIMBER-2 has been used already for simulation of Heinrich events [*Calov et al.*, 2002] and the last glacial inception [*Calov et al.*, 2005]. Using this model, a systematic stability analysis of climate-cryosphere system in the phase space of orbital forcing and CO_2 was performed. In these model experiments (R. Calov and A. Ganopolski, Multistability and hysteresis in the climate-cryosphere system, manuscript in preparation, 2005; hereinafter referred to as Calov and Ganopolski, manuscript in preparation, 2005), the existence of different equilibrium states as a function of summer solar insolation at 65°N was traced by gradual changes of summer solar insolation for several CO_2 concentrations. Different values of summer solar insolation were obtained by gradual variations of the Earth's orbital eccentricity within its observed range (0–0.07). In these experiments obliquity was set to its present value and the precessional parameter was set to either "cold" orbit (northern hemisphere summer occurs during aphelion) or "warm" orbit (northern hemisphere summer occurs during perihelion). These experiments demonstrate pronounced hysteresis behavior of the climate-cryosphere system and existence of a range of summer solar insolations where two different equilibrium states, glacial and interglacial, exist. When summer solar insolation drops below some threshold value, the interglacial climate state becomes unstable and the climate-cryosphere system experiences a bifurcation transition (Figure 1) to a state with extensive glaciation over North America and Eurasia. This threshold value for summer solar insolation depends on CO_2 concentration.

[8] Under preindustrial CO_2 concentration (280 μatm) ice appears when summertime insolation (averaged between June 21 and July 20) drops below 455 W/m^2, or 0.7 σ below the mean, and grows to full glacial size at very close to this insolation value. The rate of ice growth depends on the value of insolation below this threshold, and it could be that a slightly more negative insolation minimum may be required to nucleate a stable ice sheet within a timescale of the orbital variation. The insolation trigger value may depend somewhat on the duration of the insolation minimum [*Vettoretti and Peltier*, 2004]. Once the ice sheet is established in CLIMBER-2, it persists until the model is subjected to higher insolation than the glaciation trigger, supporting qualitatively Paillard's assumption forbidding immediate deglaciation once the ice sheet is born.

[9] The nucleation threshold in CLIMBER-2 depends strongly on pCO_2 (Figures 1 and 2), such that a higher pCO_2 requires a deeper minimum in insolation to trigger glaciation. In addition to the standard 280 μatm base case, we ran values of 200, 400, and 560 μatm. At 400 μatm the trigger decreases to −1.5 σ below the mean, and at 560 μatm the model will not glaciate at all within

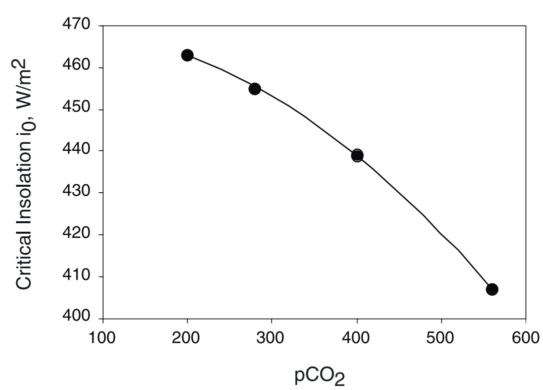

Figure 2. Critical insolation value as a function of pCO_2. Circles indicate model experiments; the smooth curve was used to interpolate.

a reasonable range of orbital eccentricity (i.e., the trigger insolation is 407 W/m^2, more than 3σ below the mean). An increase in global radiative forcing from CO_2 of 1 W/m^2 decreases the model-predicted value of i_0 by 5–20 W/m^2. It makes sense that 1 W/m^2 of CO_2 forcing is more effective than 1 W/m^2 of orbital forcing, because CO_2 forcing is global in distribution and steady in time, whereas a negative orbital insolation anomaly in the northern hemisphere summer is always accompanied by a positive anomaly in another place or season. Orbital cooling could therefore be mitigated somewhat by heat transport or storage. Watt for watt, CO_2 forcing becomes more effective than orbital forcing as pCO_2 increases. The insolation trigger varies linearly with pCO_2 itself, although this is probably just a coincidence, since the climate effect of CO_2 is logarithmic rather than linear.

[10] In summary, it seems clear that the onset of glaciation in the past must have been fairly tightly controlled by the intensity of solar heating in the high northern latitude summer. Ice core data show that pCO_2 during an interglacial does not begin to decline until after the ice sheet has started growing. Insolation minima are tightly correlated with ice sheet nucleation and growth, and the CLIMBER-2 climate model nucleates ice on the basis of an insolation trigger that is quantitatively similar to the trigger value estimated from the climate record. It makes intuitive sense that an increase in baseline pCO_2 will require a deeper minimum insolation event in order to nucleate an ice sheet. Forecasting the magnitude of this effect is more difficult than estimating the trigger insolation at typical interglacial pCO_2 levels, because a change in pCO_2 alters the climate from that of the present-day, and because we are deviating from the conditions of the paleoclimate record which served as ground truth. However, the climate sensitivity of the CLIMBER-2 model is similar to that of full mechanistic climate models [*Ganopolski et al.*, 2001], and the CLIMBER-predicted climate of the last glacial maximum is similar to paleoclimatic observations [*Ganopolski et al.*, 1998]. The sensitivity of ice sheets to CO_2 has been predicted by other models [*Berger et al.*, 1999] and empirical studies [*Paillard*, 2003].

2. Long-Term Impact of Fossil Fuel Carbon Release

[11] Past history and future forcing for atmospheric pCO_2 are shown in Figure 3a. The CO_2 time history in Figure 3a was derived from trapped air bubbles in the Vostok ice core in Antarctica [*Petit et al.*, 1999]. The future pCO_2 trajectories are anthropogenic CO_2 forcing based on model projections of 300, 1000, and 5000 Gton C releases (blue, orange, and red, respectively) [*Archer*, 2005], neglecting natural carbon cycle variability. The equilibrium partitioning of a slug of new CO_2 between the atmosphere and the $CaCO_3$-buffered oceans is such that, in the absence of natural CO_2 forcing such as glacial inception, approximately 7% of the CO_2 remains in the atmosphere 100 kyr after the perturbation, ultimately to be neutralized by the silicate weathering cycle [*Berner and Caldeira*, 1997]. This silicate thermostat has a time constant of approximately 400 kyr [*Sundquist*, 1991].

[12] The source of the carbon could be combustion of fossil fuels (5000 Gton C available [*Rogner*, 1997; *Sundquist*, 1985]) or release by meltdown [*Archer and Buffett*, 2005] of methane clathrate deposits (5000 [*Buffett and Archer*, 2004] to 10,000 [*Kvenvolden*, 1993] Gton C). The anthropogenic perturbation has the potential to be larger than the CO_2 cycles of the past that accompanied and to some extent drove the dramatic glacial/interglacial climate cycles.

3. Results

[13] We have reconstructed the second of the two *Paillard* [1998] models, in which a simple ice volume kinetic model determines the timing of deglaciation by crossing a threshold ice volume quantity. In practice, the exact timings of the glacial/interglacial transitions are very sensitive to the parameters of the ice volume growth parameters, and our model trajectories are not exactly the same as Paillard's, but they are close. We make the modification of allowing the anthrogenic pCO_2 forcing (Figure 3a) to affect the critical glaciation trigger insolation value i_0 in the future, using the relationship derived from CLIMBER-2 results in Figure 2. The insolation time series is shown in Figure 3b, superimposed on the time course of i_0. Earth's orbit is entering a period of low eccentricity and thus low insolation variability in the northern hemisphere summer [*Loutre and Berger*, 2000]. The next predicted natural glaciation should occur the next time that northern hemisphere insolation drops below the natural threshold i_0, a constant −0.75 σ below the mean. The insolation minimum in the next few millennia comes very close to Paillard's choice of i_0, but the model, such as it

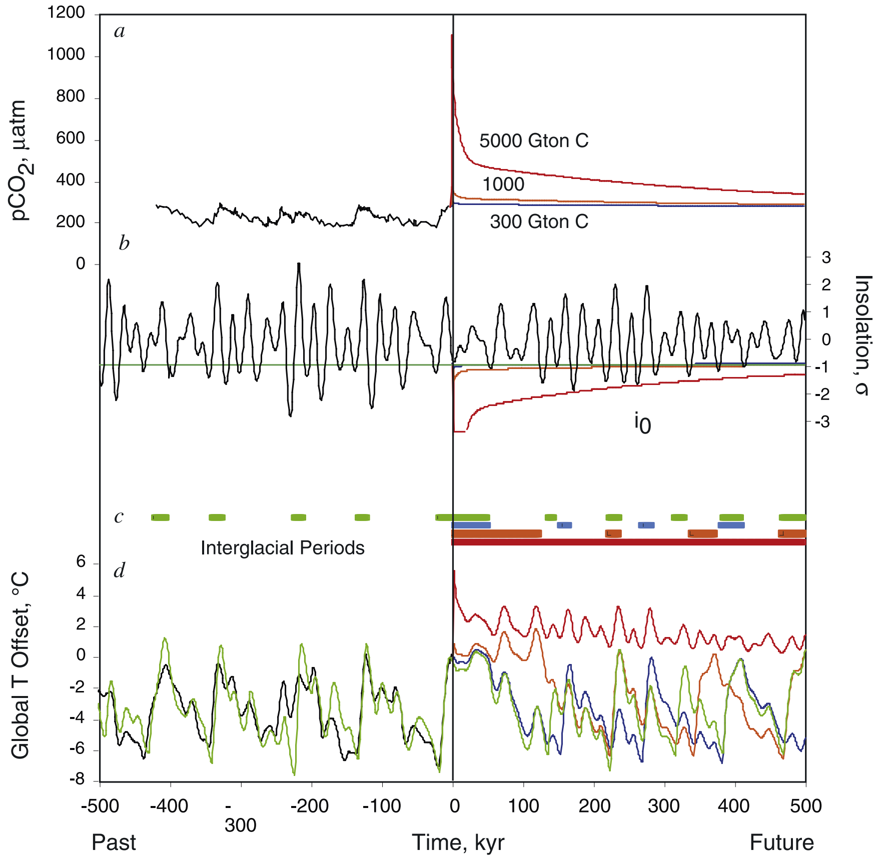

Figure 3. Effect of fossil fuel CO_2 on the future evolution of climate. Green represents natural evolution, blue represents the results of anthropogenic release of 300 Gton C, orange is 1000 Gton C, and red is 5000 Gton C. (a) Past and future pCO_2 of the atmosphere. Past history is from the Vostok ice core [*Petit et al.*, 1999], and future anthropogenic perturbations are from a carbon cycle model [*Archer*, 2005]. (b) June insolation at 65°N latitude, normalized and expressed in σ units. 1 σ equals about 20 W m^{-2}. Green, blue, orange, and red lines are values of the critical insolation i_0 that triggers glacial inception. The i_0 values are capped at -3 σ to avoid extrapolating beyond model results in Figure 3; in practice, this affects only the 5000 Gton C scenario for about 15 kyr. (c) Interglacial periods of the model. (d) Global mean temperature estimates.

is, misses i_0 this time and glaciates in 50 kyr, a result consistent with some forecasts [*Berger and Loutre*, 2002], and in conflict with others [*Imbrie and Imbrie*, 1980; *Ruddiman*, 2003a]. Paillard's i_0 was a somewhat arbitrary choice within a wide range of acceptable values, however, and a slight change in i_0 could easily tip the simulation into the onset of glaciation now rather than in 50 kyr. It appears that the natural evolution of the next few thousand years is a close call whether to glaciate or not, an issue of subtle differences in models rather than a fundamental difference between them.

[14] Assuming for the moment that natural CO_2 forcing in the future will be small, the model predicts that the available fossil fuel carbon reserves have the capacity to impact the evolution of climate hundreds of thousands of years into the future. An anthropogenic release of 300 Gton C (as we have already done) has a relatively small impact on future climate evolution, postponing the next glacial termination 140 kyr from now by one precession cycle. Release of 1000 Gton C (blue lines, Figure 3c) is enough to decisively prevent glaciation in the next few thousand years, and given the long atmospheric lifetime of CO_2, to prevent glaciation until 130 kyr from now. If the anthropogenic carbon release is 5000 Gton or more (red lines), the critical trigger insolation value exceeds 2 σ of the long-term mean for the next 100 kyr. This is a time of low insolation variability because of the Earth's nearly circular orbit. The anthrogenic CO_2 forcing begins to decay toward natural conditions just as eccentricity (and hence insolation variability) reaches its next minimum 400 kyr from now. The model predicts the end of

the glacial cycles, with stability of the interglacial for at least the next half million years (Figure 3c).

[15] Figure 3d compares past and future climates using the metric of global mean temperature. Past temperatures are derived by scaling $\delta^{18}O$ variations to global mean temperature variations of 6°C (the difference between LGM and preanthropogenic). Although ice volume and temperature changes are not simultaneous in the past, the analysis gives a rough idea of the scale and patterns of past temperature cycles. Changes in future ice volume are converted to temperature with the same scaling. We presume that the temperature change associated with a future glaciation would be amplified by a natural CO_2 drawdown, as it has done in the past. The scaling factor between ice volume and temperature, 6°C between full glacial and interglacial, implicitly includes this natural CO_2 amplification effect. To the glaciation forcing we add anthropogenic CO_2 temperature forcing, using a climate sensitivity ΔT_{2x} of 3°C for doubling CO_2. For the 5000 Gton C case, the global mean temperature perturbation, averaged over the next 500 kyr, is 4.7°C.

[16] How could natural CO_2 variation modify this forecast? The natural CO_2 drawdown associated with a descent into glacial conditions will not be an issue if the transition from interglacial to glacial climate does not occur. There may however be a positive feedback release of CO_2 from the terrestrial biosphere [*Cox et al.*, 2000] or the oceans [*Archer et al.*, 2004], analogous to the poorly understood amplifying role of CO_2 during deglaciation. The timing of the CO_2 rise during deglaciations [*Broecker and Henderson*, 1998] would be consistent with a positive feedback mechanism by which the temperature of the ocean for example affects atmospheric pCO_2 through changes in the CO_2 solubility. Models of the ocean chemistry do not show anything like the sensitivity required to drive the observed 80–100 ppm pCO_2 change, however, so other factors must be at work [*Archer et al.*, 2000]. The future pCO_2 trajectories presented here include the expected feedback from deep sea warming, but any further positive feedbacks would increase the long-term climate impact beyond what we present here.

4. Summary

[17] The combination of relatively weak orbital forcing and the long atmospheric lifetime of anthropogenic carbon could generate a longer interglacial period than has been seen in the last 2.6 million years. This will have consequences for the major ice sheets in Antarctica and Greenland [*Huybrechts and De Wolde*, 1999], and for the methane clathrate reservoir in the ocean [*Archer and Buffett*, 2005].

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References

Archer, D. (2005), The fate of fossil fuel CO_2 in geologic time, *J. Geophys. Res.*, doi:10.1029/2004JC002625, in press.

Archer, D., and B. Buffett (2005), Time-dependent response of the global ocean clathrate reservoir to climatic and anthropogenic forcing, *Geochem. Geophys. Geosyst.*, *6*, Q03002, doi:10.1029/2004GC000854.

Archer, D. E., A. Winguth, D. Lea, and N. Mahowald (2000), What caused the glacial/interglacial atmospheric pCO_2 cycles?, *Rev. Geophys.*, *38*, 159–189.

Archer, D., P. Martin, B. Buffett, V. Brovkin, S. Rahmstorf, and A. Ganopolski (2004), The importance of ocean temperature to global biogeochemistry, *Earth Planet. Sci. Lett.*, *222*, 333–348.

Berger, A. L. (1978), Long-term variations of daily insolation and quaternary climatic changes, *J. Atmos. Sci.*, *35*, 2362–2368.

Berger, A., and M. F. Loutre (2002), An exceptionally long interglacial ahead?, *Science*, *297*, 1287–1288.

Berger, A., X. S. Li, and M. F. Loutre (1999), Modelling northern hemisphere ice volume over the last 3 Ma, *Quat. Sci. Rev.*, *18*, 1–11.

Berner, R. A., and K. Caldeira (1997), The need for mass balance and feedback in the geochemical carbon cycle, *Geology*, *25*, 955–956.

Broecker, W. S., and G. M. Henderson (1998), The sequence of events surrounding Termination II and their implications for the cause of glacial-interglacial CO_2 changes, *Paleoceanography*, *13*(4), 352–364.

Brovkin, V., J. Bendtsen, M. Claussen, A. Ganopolski, C. Kubatzki, V. Petoukhov, and A. Andreev (2002), Carbon cycle, vegetation, and climate dynamics in the Holocene: Experiments with the CLIMBER-2 model, *Global Biogeochem. Cycles*, *16*(4), 1139, doi:10.1029/2001GB001662.

Buffett, B., and D. E. Archer (2004), Global inventory of methane clathrate: Sensitivity to changes in environmental conditions, *Earth Planet. Sci. Lett.*, *227*, 185–199.

Calder, N. (1974), Arithmetic of ice ages, *Nature*, *394*, 216–218.

Calov, R., A. Ganopolski, V. Petoukhov, M. Claussen, and R. Greve (2002), Large-scale instabilities of the Laurentide ice sheet simulated in a fully coupled climate-system model, *Geophys. Res. Lett.*, *29*(24), 2216, doi:10.1029/2002GL016078.

Calov, R., A. Ganopolski, M. Claussen, V. Petoukhov, and R. Greve (2005), Transient simulation of the last glacial inception. Part I: Glacial inception as a bifurcation in the climate system, *Clim. Dyn.*, doi:10.1007/s00382-005-0007-6.

Cox, P. M., R. A. Betts, C. D. Jones, S. A. Spall, and I. J. Totterdell (2000), Acceleration of global warming due to carbon-cycle feedbacks in a coupled climate model, *Nature*, *408*, 184–187.

DeNoblet, N., I. C. Prentice, S. Joussaume, D. Texier, A. Botta, and A. Haxeltine (1996), Possible role of atmospheric biosphere interactions in triggering the last glaciation, *Geophys. Res. Lett.*, *23*, 3193–3194.

Ganopolski, A., S. Rahmstorf, V. Petoukhov, and M. Claussen (1998), Simulation of modern and glacial climates with a coupled global model of intermediate complexity, *Nature*, *371*, 323–326.

Ganopolski, A., V. Petoukhov, S. Rahmstorf, V. Brovkin, M. Claussen, A. Eliseev, and C. Kubatzki (2001), CLIMBER-2: A climate model of intermediate complexity. Part II: Model sensitivity, *Clim. Dyn.*, *17*, 735–751.

Greve, R. (1997), Application of a polythermal three-dimensional ice sheet model to the Greenland ice sheet: Response to steady-state and transient scenarios, *J. Clim.*, *10*, 901–918.

Huybrechts, P., and J. De Wolde (1999), The dynamic response of the Greenland and Antarctic ice sheets to multiple-century climatic warming, *J. Clim.*, *12*, 2169–2188.

Imbrie, J., and J. Z. Imbrie (1980), Modeling the climatic response to orbital variations, *Science*, *207*, 943–955.

Imbrie, J., et al. (1993), On the structure and origin of major glaciation cycles: 2. The 100,000-year cycle, *Paleoceanography*, *8*, 699–736.

Khodri, M., Y. Leclainche, G. Ramstein, P. Braconnot, O. Marti, and E. Cortijo (2001), Simulating the amplification of orbital forcing by ocean feedbacks in the last glaciation, *Nature*, *410*, 570–574.

Kvenvolden, K. A. (1993), Gas hydrates: Geological perspective and global change, *Rev. Geophys.*, *31*(2), 173–187.

Loutre, M. F., and A. Berger (2000), Future climatic changes: Are we entering an exceptionally long interglacial?, *Clim. Change*, *46*, 61–90.

Paillard, D. (1998), The timing of Pleistocene glaciations from a simple multiple-state climate model, *Nature*, *391*, 378–381.

Paillard, D. (2001), Glacial cycles: Toward a new paradigm, *Rev. Geophys.*, *39*, 325–346.

Paillard, D. (2003), Variations climatiques au cours du Quaternaire: L'apport des modèles "simples," thesis, Univ. de Paris, Paris.

Petit, J. R., et al. (1999), Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica, *Nature*, *399*, 429–436.

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

Raymo, M. E. (1997), The timing of major climate terminations, *Paleoceanography*, *12*, 577–585.

Rogner, H.-H. (1997), An assessment of world hydrocarbon resources, *Annu. Rev. Energy Environ.*, *22*, 217–262.

Ruddiman, W. F. (2003a), The Anthropocene greenhouse era began thousands of years ago, *Clim. Change*, *61*, 261–293.

Ruddiman, W. F. (2003b), Orbital insolation, ice volume, and greenhouse gases, *Quat. Sci. Rev.*, *22*, 1597–1629.

Sundquist, E. T. (1985), Geological perspectives on carbon dioxide and the carbon cycle, in *The Carbon Cycle and Atmospheric CO_2: Natural Variations Archean to Present*, *Geophys. Monogr. Ser.*, vol. 32, edited by E. T. Sundquist and W. S. Broecker, pp. 105–173, AGU, Washington, D. C.

Sundquist, E. T. (1991), Steady- and non-steady-state carbonate-silicate controls on atmospheric CO_2, *Quat. Sci. Rev.*, *10*, 283–296.

Vettoretti, G., and W. R. Peltier (2004), Sensitivity of glacial inception to orbital and greenhouse gas climate forcing, *Quat. Sci. Rev.*, *23*, 499–519.