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A movable trigger: Fossil fuel CO\textsubscript{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\textsubscript{2} of 280 ppm, the CLIMBER-2 model initiates an ice sheet in the Northern Hemisphere when insolation drops 0.7 \( \sigma \) (standard deviation) or 15 W/m\textsuperscript{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\textsubscript{2} is raised in CLIMBER-2, a deeper minimum in summertime insolation is required to nucleate an ice sheet. Carbon cycle models indicate that \( \sim 25\% \) of CO\textsubscript{2} from fossil fuel combustion will remain in the atmosphere for thousands of years, and \( \sim 7\% \) 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\textsubscript{2}. The standard deviation of insolation at this time and place is about 20 W m\textsuperscript{-2}, which is large compared with the global mean effect of doubling CO\textsubscript{2}, about 4 W m\textsuperscript{-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.

[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).

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

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.

[8] 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...
a reasonable range of orbital eccentricity (i.e., the trigger insolation is 407 W/m², more than 3σ below the mean). An increase in global radiative forcing from CO₂ of 1 W/m² decreases the model-predicted value of i₀ by 5–20 W/m². It makes sense that 1 W/m² of CO₂ forcing is more effective than 1 W/m² of orbital forcing, because CO₂ 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, CO₂ forcing becomes more effective than orbital forcing as pCO₂ increases. The insolation trigger varies linearly with pCO₂ itself, although this is probably just a coincidence, since the climate effect of CO₂ 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 pCO₂ 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 pCO₂ 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 pCO₂ levels, because a change in pCO₂ 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 CO₂ 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 pCO₂ are shown in Figure 3a. The CO₂ time history in Figure 3a was derived from trapped air bubbles in the Vostok ice core in Antarctica [Petit et al., 1999]. The future pCO₂ trajectories are anthropogenic CO₂ 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 CO₂ between the atmosphere and the CaCO₃-buffered oceans is such that, in the absence of natural CO₂ forcing such as glacial inception, approximately 7% of the CO₂ 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 CO₂ 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 anthropogenic pCO₂ forcing (Figure 3a) to affect the critical glaciation trigger insolation value i₀ 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 i₀. 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 i₀, a constant −0.75 σ below the mean. The insolation minimum in the next few millennia comes very close to Paillard’s choice of i₀, but the model, such as it
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 $\sigma$ 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 anthropogenic 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

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**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 $p$CO$_2$ 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 $\sigma$ units. 1 $\sigma$ 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 $-3$ $\sigma$ 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.
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 δ18O 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 CO2 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 CO2 amplification effect. To the glaciation forcing we add anthropogenic CO2 temperature forcing, using a climate sensitivity ΔT2x of 3°C for doubling CO2. 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 CO2 variation modify this forecast? The natural CO2 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 CO2 from the terrestrial biosphere [Cox et al., 2000] or the oceans [Archer et al., 2004], analogous to the poorly understood amplifying role of CO2 during deglaciation. The timing of the CO2 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 pCO2 through changes in the CO2 solubility. Models of the ocean chemistry do not show anything like the sensitivity required to drive the observed 80–100 ppm pCO2 change, however, so other factors must be at work [Archer et al., 2000]. The future pCO2 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

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