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Simulating late Pliocene Northern Hemisphere climate with the LLN 2-D model

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Abstract. Deep-sea sediment records suggest Northern Hemisphere glaciation intensified at approximately 2.75 Ma. In this paper we simulate the fluctuations of the late Pliocene Northern Hemisphere ice-sheets volume using the LLN 2-D model, forced by the astronomically derived insolation and by scenarios of CO_2 concentrations. The model simulates the waxing and waning of the Northern Hemisphere ice volume in an acceptable agreement with geological reconstructions, and in particular supports the geological evidence that the development of significant Northern Hemisphere ice sheets started between 2.75 and 2.55 Ma, a time interval accompanied by quite high and stable eccentricity values, and increased obliquity amplitude. When the CO_2 concentration is lower than a threshold value, the Milankovitch forcing explains the suddenness and specific timing of this entrance into the Ice Age during the late Pliocene.

Introduction

The glaciation of Greenland and of the Arctic region seems to have started during late Miocene [*ODP Leg 151*, 1994]. However the first significant ice sheets in the Northern Hemisphere did not develop until about 2.75 Ma when the Eurasian Artic and Northeastern Asia entered into glaciation, about 100 ka before the entrance into glaciation of Alaska and about 200 ka before a significant glaciation occurred in North East America [*e.g.*, *Raymo et al.*, 1992; *Maslin et al.*, 1995; 1997].

Attempts have been made to explain the significant intensification of Northern Hemisphere glaciation between 2.75 and 2.55 Ma [*e.g.*, *Li*, 1997]. A recent suggestion is that changes in orbital forcing may have been an important mechanism controlling the gradual cooling and the subsequent rapid intensification of Northern Hemisphere glaciation. *Maslin et al.* [1995] proposed that they were forced in particular by the gradual increase in the amplitude of obliquity from 3.5 to 2.5 Ma and a sharp rise in the amplitude of precession and thus insolation between 2.8 and 2.55 Ma.

In this paper we will simulate the late Pliocene Northern Hemisphere ice-sheet volume with the LLN 2-D model, and compare the simulation with the geological records in order to understand better the climatic mechanisms controlling the entrance into glaciation during the late Pliocene in the Northern Hemisphere.

Simulating the late Pliocene Northern Hemisphere glaciation

The LLN 2-dimensional climate model links the atmosphere, the mixed layer of the ocean, the sea-ice, the continents, the ice sheets and their underlying lithosphere of the Northern Hemisphere. Its description and validation are given in *Gallée et al.* [1991, 1992] and *Berger and Loutre* [1996], for example.

To simulate the late Pliocene climate, insolation and atmospheric CO_2 variations must be known. For the insolation variation, the orbital elements calculated by *Berger and Loutre* [1991] and updated by *Loutre and Berger* [1993] will be used. For CO_2, there are some evidences that its atmospheric concentration was generally higher during the Pliocene than during the Pleistocene. For example, *Van Der Burgh et al.* [1993] suggested that the late Neogene CO_2 concentrations fluctuated between about 280 and 370 ppmv based on the evidence from fossil tree leaves. This is why a first scenario will test the response of the LLN model to a constant CO_2 concentration of 300 ppmv from 4 to 2 Ma BP. The simulated ice volume (not shown here) fluctuates with long interglacials and rather weak glacial peaks; the NH ice volume is always less than 20×10^6 km^3. The spectrum is rather noisy which is actually not the case for

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a constant 220 ppmv CO_2 experiment where the astronomical periods appear clearly [*Berger et al.*, 1997]. There is no long-term cooling, but a glacial is simulated at ~2.7 Ma, a possible indication of the important role played by insolation. It was therefore decided to follow the original proposal by *Saltzman and Verbitsky* [1993] where the CO_2 concentration is linearly decreasing from 330 ppmv to 290 ppmv between 4 and 2 Ma. The experiment starts at 3.94 Ma. This date was chosen chosen as it is a warm period in marine oxygen isotope data [*Tiedemann et al.*, 1994; *Shackleton et al.*, 1995], and there is no evidence of ice rafted debris in either the Pacific [*Maslin et al.*, 1995] or the Atlantic [*Raymo et al.*, 1992]. We have therefore assumed there were no major Northern Hemisphere ice sheets at that time. The solid line in Fig. 1c indicates the Northern Hemisphere ice-sheet volume from 4 Ma to 2 Ma simulated by the model with all the hypotheses described above. Interglacial conditions prevail, especially before 3.2 Ma, where only 4 weak short-lasting glacials appear. Between 3.2 and 2 Ma BP, 3 intervals with more extensive increase in ice volume are simulated: between 3.2 and 2.9 Ma, between 2.75 and 2.45 Ma, and between 2.35 and 2.1 Ma. The simulated variation of the ice volume at about 3 Ma has the smallest amplitude. The ice-sheet volume between 2.6 and 2.2 Ma reaches a maximum of $\sim 17 \times 10^6$ km^3, comparable in amplitude to the abortive glaciations of isotopic substage 5b or 5d during the last interglacial-glacial cycle. Linear regression indicates a slight trend of increase in the simulated ice volume from 3 Ma onwards.

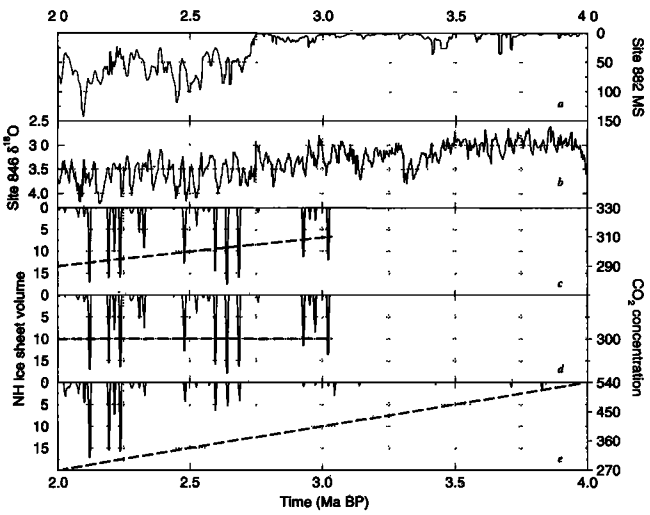

Figure 1. Comparison between geological data and the simulated Northern Hemisphere ice sheet volumes (10^6 km^3). (a) Magnetic susceptibility from Site 882, in c.g.s. unit [*Maslin et al.*, 1995]; (b) benthic δ^{18}O record from the equatorial Pacific at Site 846, in $^o/_{oo}$ [*Shackleton et al.*, 1995]; (c) simulation from 3.940 to 2 Ma (solid line) forced with insolation variations [*Loutre and Berger*, 1993] at the top of the atmosphere and a CO_2 concentration linearly decreasing from 350 ppmv at 5 Ma to 250 ppmv at present [*Saltzman and Verbitsky*, 1993] (bold dashed line); and (d) simulation from 3.940 to 2 Ma (solid line), with CO_2 linearly decreasing from 540 ppmv at 4 Ma to 270 ppmv at 2 Ma (bold dashed line)

The simulated ice volume is compared with the Site 882 (North Pacific) magnetic susceptibility records [*Maslin et al.*, 1995], with the Site 846 (equatorial Pacific) benthic δ^{18}O isotope record [*Shackleton et al.*, 1995] and some other data that are not included in Fig. 1. In the geological data, the 2.75 Ma glaciation recorded in Site 882 sediments can be correlated with the benthic δ^{18}O enrichment in Site 846 and the dust flux increase at Site 659 off west Africa [*Tiedemann et al.*, 1994], implying that the rapid cooling at 2.75 Ma was a global event. The simulation indicates that for about 150 ka before ~2.75 Ma there are virtually no ice sheets in the Northern Hemisphere; between 2.75 and 2.45 Ma, significant glaciation take place, four glacial maxima occurring at 2.69, 2.64, 2.60 and 2.48 Ma, the first 3 being the largest. These 4 simulated cooling events are coeval with the oxygen isotopic stages G4, G2, 104 and 98 in the marine sediment record, and with the Site 659 dust flux increases at the corresponding times. The ice volume increases before 2.9 Ma and around 2.2 Ma especially well correspond to the Site 659 high dust fluxes. The time intervals without significant Northern Hemisphere ice sheets correspond to the relatively low δ^{18}O, dust flux and magnetic susceptibility. There are discrepancies, however, if we compare further in detail. The δ^{18}O stages 96 and 100 are underestimated in the simulation. The simulated glaciation at 2.12 Ma, with an estimated ice volume of 17×10^6 km^3, has no counterpart in the records. The differences may be due to the very simplified boundary conditions and CO_2 scenario assumed in the simulations, because of inaccuracies in the data themselves and/or because of a lack of adequate mechanisms in the model.

An obvious question related to this simulation is that the long-term cooling trend seems too weak, and the simulated Northern Hemisphere ice volume before 3 Ma are too large considering that no significant ice sheets existed before ~2.7 Ma in either Eurasia or North America according to the geological records.

As the CO_2 scenario used in that simulation shows only a 40 ppmv difference from 4 Ma to 2 Ma, a larger CO_2 gradient scenario is constructed assuming a value of 270 ppmv at 2 Ma, about the late Pleistocene interglacial level, and a linear increase towards the past to reach 540 ppmv at 4 Ma. This CO_2 concentration is close to that deduced by *Cerling* [1992] from C3/C4 vegetation shift which argues persuasively for a CO_2 value of 500-600 ppmv at 7 Ma BP. We start the simulation at 3.94 Ma, assuming no ice sheet in the Northern Hemisphere. The model simulates very small ice sheets around 3.8 Ma, 3.7 Ma, 3.4 Ma, 3.1 Ma and 3.0 Ma (Fig. 1d), which seem to correspond to cooling events reflected in the Site 882 magnetic susceptibility and the Site 846 δ^{18}O. In this scenario, it is only after ~2.7

Ma that the simulated Northern Hemisphere ice volume starts to exceed 3×10^6 km^3. Moreover, the trend of ice volume increase is larger than in the former simulation. The simulated ice-sheet waxing and waning between 3 and 2 Ma can be well correlated to those simulated in the earlier experiments, except that the ice volumes of this simulation are usually smaller before 2.3 Ma and slightly larger after 2.1 Ma.

Finally, the model was forced by a 420 ppmv constant CO_2 from 4 to 3 Ma, and a CO_2 decreasing linearly to 270 ppmv from 3 Ma to 2 Ma. Significant Northern Hemisphere ice sheets appear also only after ~2.7 Ma, although it is difficult to get rid of the formation of an ice sheet at about 3.88 Ma due to the very large insolation forcing at that time.

Link between insolation and glacial - interglacial cycles during late Pliocene

For a glaciation to develop, the astronomical theory [*Milankovitch,* 1941; *Berger,* 1988] requires that the summer in northern high latitudes must be cold enough to prevent the winter snow from melting, in such a way as to allow a positive value in the annual budget of snow and ice, which generates a positive feedback and a cooling through a subsequent increase of the surface albedo.

Fig. 2 compares the Earth's orbital parameters and the resultant July insolation at 65° N for the late Pliocene [*Loutre and Berger,* 1993]. This traditional high latitude summer insolation is used here only as an illustration [*Milankovitch,* 1941]; actually the LLN model uses insolations for all days and latitudes and similar conclusions can be drawn from any insolation curve (except those close to the polar night [*Berger et al.* 1993a]). Between 3 and 2 Ma, the eccentricity-modulated amplitude of the precessional index is dominated by the ~400 ka period, with the largest amplitude centered around 3.0, 2.6, and 2.2 Ma. Since July 65°

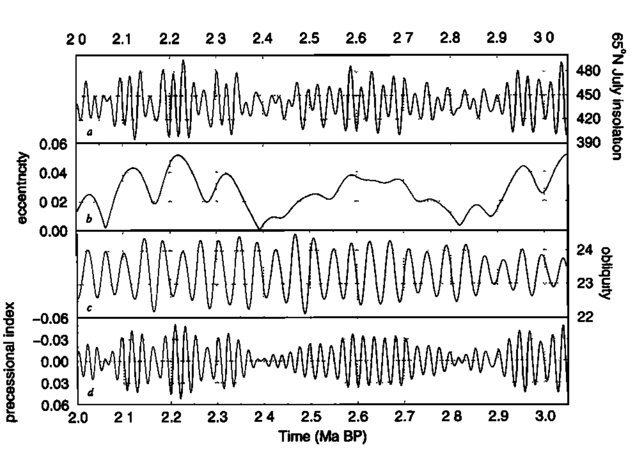

Figure 2. The Earth's orbital elements and the resultant insolation variation from 3.05 to 2 Ma. (a) The 65°N July insolation (W m^{-2}), and (b - d) the Earth's orbital parameters [*Loutre and Berger,* 1993]

N insolation is dominated by the precessional signal, its amplitude variation shows also a strong 400-ka period. Three time intervals with large amplitudes can be identified: roughly before 2.9 Ma, from 2.75 to 2.45 Ma, and from 2.35 to 2.1 Ma. These insolation stages correspond clearly to the three simulated and observed glaciations over the same time span. Moreover, as observed by *Maslin et al.* [1995; 1997], from 3.5 Ma until 2.5 Ma the amplitude of obliquity gradually increases, enhancing the insolation amplitude.

Results with a large CO_2 gradient (Fig. 1d) show that low summer insolations can lead to the development of late Pliocene NH ice sheets only when the CO_2 concentration allows winter snow to persist from year to year. This CO_2 threshold seems to be reached sometime between ~2.9 and ~2.7 Ma in our last sensitivity experiment. But the deepest summer insolation minima required for ice sheets to grow occur only within time intervals when the amplitude of the variations is large. This requires high eccentricity, which explains why the significant glaciations can not be initialized until 2.75 - 2.55 Ma.

High eccentricity leads to warm but also to cool summers. These last (corresponding to deep minima in the high latitude summer insolation) lead to sharp glacials, each lasting usually about 20 ka, as it can be seen in Fig. 1c and 1d. This is particularly true for the three insolation minima of ~400 Wm^{-2} at 2.691, 2.646 and 2.600 Ma which coincide with the 3 simulated distinct glacials centered at 2.686, 2.640 and 2.596 Ma in Fig. 1c, or at 2.689, 2.643 and 2.597 Ma in Fig. 1d. This is also the case for the insolation minima occurring at 2.333, 2.241, 2.219 and 2.126 Ma.

As a consequence, the relation between eccentricity and climate during the late Pliocene — glaciation under high eccentricity — is opposite to the relation which prevails during the late Pleistocene, when the interglacials are related to high eccentricitiy [*e.g., Imbrie et al.,* 1984]. This originates from the different responses of the climate system to the insolation forcing under different boundary conditions. In the warm late Pliocene, the Northern Hemisphere ice sheets could only develop during very cool summers; in the colder late Pleistocene, significant Northern Hemisphere ice sheets exist during most of the time and only high summer insolation in the polar regions can start to melt the ice sheets and lead to an interglacial.

Conclusions

The major characteristics of the late Pliocene Northern Hemisphere climate seems to be approximately well reconstructed by the LLN 2-D model, forced both with insolation and a linearly decreasing atmospheric CO_2 concentration. It is the conjunction of reaching a threshold value of CO_2 concentration (about 400 ppmv for the LLN 2-D model) and of an orbital forcing with high eccentricity and high amplitude obliquity that seems to explain best both the entrance into late Pliocene glaciation at 2.75 Ma and the start of significant glacial-interglacial cycles. Contrary to the late Pleistocene, the late Pliocene ice ages correspond to times when the insolation variations had a large amplitude, it means to times of high eccentricity.

The simplified structure of our model prevents us from discussing the origin of the CO_2 trend that we have adopted. This model is indeed far from being complete and needs to be improved (by including the deep ocean circulation, the interactive biogeochemical cycles and atmospheric chemistry, and improving the hydrological cycle and cloud physics, for example). Moreover, other forcings may play a role in shaping the climatic evolution, like the distribution of the continents, some particular feature of paleogeography (*e.g.*, the emergence of the Panama Isthmus), and atmospheric dust originating from volcanic activity, from land erosion, or from interplanetary media.

It remains nevertheless that, as it stands, the model shows that the CO_2 concentration threshold and the particular characteristics of the astronomical elements at that time may have played a decisive role. Although the possible interplay between falling CO_2 and insolation is not a new concept, our results shed light on the way this could have occurred, and it helps counter the likely misconception that some kind of extreme forcing occurred at 2.75 Ma, rather than just the gradual crossing of a cooling threshold.

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