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Amplitude and phase of glacial cycles from a conceptual model

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Abstract

The astronomical theory of climate, in which the orbital variations of the Earth are taken to drive the climate changes, explains many features of the paleoclimatic records. Nevertheless, the precise link between insolation variations and climatic changes during the Quaternary remains mysterious in several aspects. In particular, the largest sea level changes of the past million years occurred when insolation variations were minimal, like during stage 11, and vice versa like during stage 7. Moreover, recent data from terminations II and III show surprising phase lead and lag between insolation and sea level variations. To explain these paradoxical amplitude and phase modulations, we suggest here that deglaciations started when a combination of insolation *and* ice volume was large enough. To illustrate this new idea, we present a simple conceptual model that simulates the sea level curve of the past million years with very realistic amplitude modulations, and with good phase modulations.

Keywords: glacial cycles; climate model; Milankovitch theory; insolation forcing

1. Introduction

Although we find astronomical frequencies in almost all paleoclimatic records [1,2], it is clear that the climatic system does not respond linearly to insolation variations [3]. The first well-known paradox of the astronomical theory of climate is the '100 kyr problem': the largest variations over the past million years occurred approximately every 100 kyr, but the amplitude of the insolation signal at this frequency is not significant. Although this problem remains puzzling in many respects, multiple equilibria and thresholds in the climate system seem to be key notions to explain this paradoxical frequency. In particular, the ice volume critical size is a good candidate to trigger the threshold [3,4]. Indeed, terminations occurred only after considerable build-up of ice volume; beyond this point, the next northern latitude summer insolation maximum, even a relatively weak one, will cause a deglaciation [5]. This simple idea allowed Paillard [4] to construct a conceptual model that successfully simulates the 100 kyr terminations.

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Another intriguing paradox is the relation between the amplitudes of the insolation extrema and the corresponding ice volume extrema. There is no simple relation between these two extrema. For example, transition V (from stage 12 to stage 11), which was probably the largest one over the past million years [6–8], occurred when insolation variations were very weak. This is known as the 'stage 11 problem'. Similarly, transition III (from stage 8 to stage 7) was rather small, whereas insolation variations during this time period were important [9]. The very small ice volume during stage 11 could be explained by its exceptional duration, two precessional cycles against only one for the other interglacial [4]. But what is the explanation for MIS 12.2, a stage with a weak minimum of insolation but probably the largest ice volume of the past 600 kyr [8]? (see Fig. 1) The same question also exists for MIS 16.2 and 2.2. A related paradox is the '400 kyr problem'. The amplitude of summer high latitude insolation variations is maximum every 400 kyr, due to the dominance of this periodicity in the eccentricity modulation of the precessional forcing. The 400 kyr problem is often presented as the absence of such a frequency in paleoclimatic records [10]. For the last 400 kyr, it is even the contrary: an amplitude modulation in the sea level curve does exist, but is opposite to the 400 kyr cycle of insolation. Sea level transitions were maximal when insolation variations were minimal, and vice versa (see Fig. 1). However, this inverse relationship is not so clear for the rest of the record all along the last million years.

Moreover, the phase relationship between a termination and the corresponding insolation extremum may not be constant through time. Termination II has been in advance with respect to the insolation maximum [11,12], whereas new U–Th datings seem to show the contrary for termination III [13].

To solve these amplitude and phase paradoxes, we suggest here that ice volume and insolation together play a role in the triggering of deglaciations. We suppose that the climatic system has two main states of variation: **g** (glaciation) and

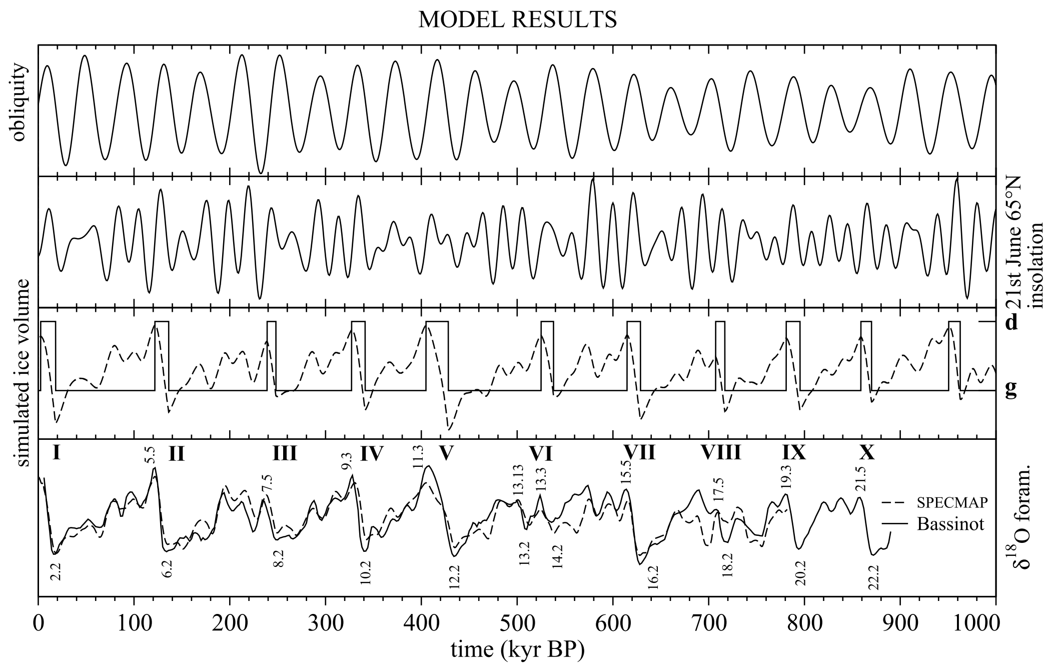

Fig. 1. Model results. From top to bottom: obliquity, June solstice insolation 65°N [16], modeled ice volume and model state (dashed line), foraminifera $\delta^{18}O$ from Bassinot et al. [20] or from SPECMAP [21] (dashed line), that can be interpreted as a proxy for global ice volume. Bold Roman numerals are terminations, and light decimal numbers are stages.

d (deglaciation), and that the **g**-to-**d** transition occurs when a combination of insolation and ice volume is large. More precisely, a deglaciation can occur when insolation forcing is moderate if ice volume is very large, or reciprocally when ice volume is moderate if insolation forcing is very large. We propose here a conceptual model based on this simple idea. It is driven by changes in the June Solstice insolation at 65°N and by obliquity. This simple model not only reproduces sea level transitions at the correct time, but also sea level extrema with the right amplitude. In addition, despite high latitude northern insolation being the only external forcing, we obtain significant phase variations between climatic transitions and insolation, in agreement with chronologies for terminations II and III. This proves that, in contrast to some previously published ideas [14,15], an astronomical theory of glacial cycles can easily accommodate for such phase variations. Furthermore, it proposes a conceptual explanation of how phase and amplitude variations are linked together.

2. Model description

We suppose here that the climatic system has two different states of evolution: the 'glaciation' state **g** and the 'deglaciation' state **d**. The evolution of these states is simply described by two linear equations:

$$\text{during state } \mathbf{g}: \quad \frac{dv}{dt} = -\frac{I_{tr}}{\tau_I} - \frac{O}{\tau_O} + \frac{1}{\tau_g} \quad (a)$$

$$\text{during state } \mathbf{d}: \quad \frac{dv}{dt} = -\frac{I_{tr}}{\tau_I} - \frac{O}{\tau_O} + \frac{(v_d - v)}{\tau_d} \quad (b) \qquad (1)$$

where v is the normalized ice volume. τ_d, τ_g, τ_I and τ_O are time constants. I_{tr} and O are the astronomical forcing. O is obliquity [16] normalized to unity variance and zero mean. I_{tr} is calculated from I, the June solstice insolation at 65°N [16], normalized to unity variance and zero mean, using a truncation function:

$$f(x) = (x + \sqrt{4a^2 + x^2}) - 2a \quad \text{if } x < 0$$
$$f(x) = x \quad \text{if } x > 0$$

(where a is a constant) and then normalized to unity variance and zero mean. This empirical adjustment accounts for the lower sensitivity of the ice volume with respect to summer northern hemisphere insolation when the latter is not large [4]. Eq. 1a,b is a simple model of the ice volume variation. The first term could represent mainly the ice melting during boreal summer, and the second one the accumulation of snow. Indeed, in a simple latitudinal model of moisture [17], accumulation of snow at poles is related to equator to pole annual insolation gradient [18], and therefore linearly related to obliquity [9]. Compared to previous conceptual models of ice volume variations, our model thus explicitly incorporate a term related to obliquity, an important parameter because it represents the mean annual insolation of polar regions. While the two first terms are related to the external orbital forcing, the third term represents a trend of slow glaciation for the **g** state, or of rapid relaxation to a deglaced state for the **d** state. This simple model implicitly represents the internal feedbacks of the climate system (like the carbon cycle, the oceanic and atmospheric circulations) by the occurrence of the two states **d** and **g** described before.

We still need to define when the model jumps from one state to the other. A recent study with a coupled GCM confirms that the decrease in summer northern insolation is probably the trigger towards a glaciation [19]. Snowfall increases over high northern plateaus might be also important to trigger towards a glaciation. Thus our model undergoes a **d**-to-**g** transition when a combination of I and O falls below a threshold I_0 (see Eq. 2a). The main new feature of this model concerns the triggering of deglaciations. Raymo [5] noticed that terminations occurred only after considerable build-up of ice sheet, and that beyond this point, the next northern latitude summer insolation maximum, even a relatively weak one, will cause deglaciation. The same idea appeared in the Paillard's model [4], with a first threshold on the ice volume and a second one on the insolation. But this formulation constrained the ice volume maxima to a constant amplitude, whereas they seem significantly variable in the paleoclimatic records. Therefore in our model we chose to express a condition on insolation and ice volume together for the **g**-to-**d** transition. The simplest possible criteria is to define a threshold on a linear combination of insolation and ice volume (see Eq. 2b; κ_I, κ_O and v_0 are constants):

$$\begin{cases} \mathbf{d}{-}{>}\mathbf{g}: \quad I+\frac{\kappa_O}{\kappa_I}O<I_0 \quad (\text{and } \kappa_I I+\kappa_O O+v<v_0) \quad (a) \\ \mathbf{g}{-}{>}\mathbf{d}: \quad \kappa_I I+\kappa_O O+v>v_0 \quad (\text{and } I+\frac{\kappa_O}{\kappa_I}O>I_0) \quad (b) \end{cases} \tag{2}$$

This formulation allows for **g**-to-**d** transitions when insolation is moderate if ice volume is large, or reciprocally when ice volume is moderate if insolation is large.

3. Results and discussion

Choosing $\tau_I = 9$ kyr, $\tau_O = 30$ kyr, $\tau_g = 23$ kyr, $\tau_d = 12$ kyr, $a = 0.6$, $\kappa_O = 0.35$, $\kappa_I = 0.6$, $I_0 = 0$, $v_0 = 6.25$ and starting at 1000 kyr BP in a **g** state with a normalized ice volume $v = 3.75$, we find an ice volume in very good agreement with the reconstructed ice volume from Bassinot et al. [20] or from the SPECMAP stacked curve [21] (Fig. 1). In particular, the timing of each glacial–interglacial transition is correct. There is one ambiguity on termination VI (stage 14 to stage 13), which occurs between stage 14.2 and stage 13.3 in our model whereas it is not obvious in the sea level records if the main termination is 13.2–13.13 or 14.2–13.3. The amplitude of sea level extrema from our model is in good agreement with the paleoclimatic records, in particular for the last four climatic cycles. The relative magnitude of the sea level maxima is (both in model and in data): stage 11 > stage 5 > stage 9 > Holocene > stage 7. The relative order of the sea level minima is (both in model and in data): stage 12 < stage 2 < stage 6 < stage 10 < stage 8. For the oldest part (between stage 13 and stage 22), the fit is quite good, except for transitions 13–14 and 17–18, which seem too large compared to the records.

It is worth emphasizing that the very good agreement between model and records is achieved with only a very small number of tunable parameters (that we have chosen with a trial and error method). This model is also quite robust. Time constants τ_I, τ_O, τ_g and τ_d could be chosen respectively in the intervals [6;16] kyr, [18;∞] kyr, [18;26] kyr and [10;15] kyr with only one deglaciation misplaced over the last million years, and with no significant changes in the relative amplitude of glacial cycles. With the same criteria, we can choose a, I_0, κ_O, κ_I, and v_0 respectively in the intervals [0.3;∞], [−0.1;0.5], [0;0.8], [0.55;1], [5.5;6.4]. In particular, when the obliquity forcing is omitted (i.e. for $\tau_O = \infty$ and $\kappa_O = 0$, see Fig. 2), timings of deglaciations are well estimated, but the ice volume amplitudes are not so well reproduced. With no truncation of insolation (i.e. with $a = 0$), termination III only is misplaced (see Fig. 2). Changing the initial condition on v and on model state only changes model results during the first two or three hundreds of thousand years (see Fig. 2).

MIS 7 seems to be very sensitive with respect to the model's parameters. Indeed, the start of termination III occurs almost at the maximum of insolation in our model, and thus a small change of one parameter can shift this termination to the following insolation maxima. Transition 8–7 appears to be a '*last minute deglaciation*'.

Concerning the ice volume phasing with respect to insolation, we find that it may significantly change during the past million years (see Table 1). Phasing is defined here as the time lag between a **g**-to-**d** 'threshold crossing' and the following insolation (June solstice, 65°N) maxima. This time lag rises from −2.5 to −8.7 kyr using this definition. The classical definition based on the time difference between maximum of insolation and the mid-transition of ice volume is not necessarily the most appropriate. Indeed, deglaciations are strongly non-linear processes, as illustrated by the occurrence of a large melt water pulse (MPWla) event during the last deglaciation. Such a large and rapid change cannot be captured by our simple linear model. It can be argued that, to some extent, 'deglaciation' is more an 'event-like' process than a slow continuous one. It therefore makes sense to compute the lag using our model 'threshold crossing', instead of the ice volume mid-transition. This simple model thus suggests an early termination II with respect to insolation changes (earlier than we could expect from phasing at termination I), and a late one for ter-

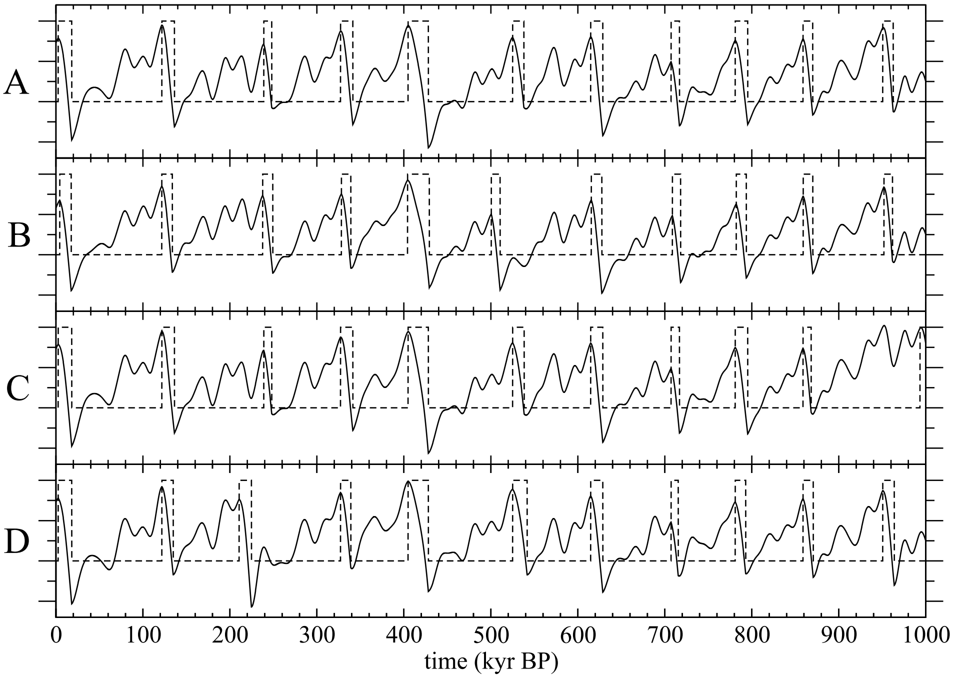

Fig. 2. Sensitivity of the model with respect to its parameters. (A) Model with optimal parameters (same as Fig. 1). (B) Without obliquity, i.e. with $\kappa_O = 0$ and $\tau_O = \infty$. (C) With a different initial condition: **d** state and $v = 0.2$ at 1000 kyr BP. (D) Without truncation of insolation (i.e. $a = \infty$).

mination III, whereas terminations I and IV are intermediate (see the last column of Table 1). Indeed, during MIS 8, the ice volume and the insolation maxima are not very large; therefore, the model crosses the **g**-to-**d** threshold only when the insolation is at its maximum. This conceptually explains why this deglaciation is late, which is indeed suggested by recent data [13]. In contrast, the ice volume and the insolation maximum are very large during stage 6, and this is why the

Table 1

To examine the phasing between insolation and ice volume variations, we show different time indicators against the 11 last terminations: minima of I (June solstice insolation at 65°N) [16], mid-transitions of I, maxima of I, timings of model's threshold, mid-transitions of v (modeled ice volume)

Term	Min. I	Trans. I	Max. I	Threshold	Trans. v	GSS97 age	Phasing
1	23.9	17.1	11.2	18.1	12.8		−6.9
2	139.7	132.8	127.5	136.2	130.2		−8.7
3	253.8	247.4	242.8	248.1	244.2	247.9	−5.3
4	346.1	339.8	334.3	341.3	335.9	339.3	−7
5	436.4	431.3	425.6	428.1	420.6	423.6	−2.5
6	516.6	510.8	531.5	538	533.8	534.5	−6.5
7	633.2	626.1	621	628.7	623.2	621.6	−7.7
8	723.9	718.5	713.5	716.8	713.1		−3.3
9	798.3	792.7	787.8	795.3	789.9		−7.5
10	875.4	869.8	864.8	870	865.8		−5.2
11	979.2	963.5	958.1	962.6	958.5		−4.5

GSS97 ages of terminations are from Raymo et al. [5] (mean age). The last column shows phasing calculated as the timing difference between max. of I and threshold.

termination occurs at an early time, when insolation is still increasing. This corresponds somewhat to observations of an early termination II, made in some records dated with the U–Th method [12,11]. It has been claimed that an early termination II contradicts the theory of the astronomical forcing on climate. Our interpretation is different, and we believe that variations in phasing only reflect a threshold mechanism in the system for deglaciations, and that this threshold depends on global ice volume. Changes in phasing are therefore directly linked to ice volume amplitudes, and the 'amplitude modulation' leads to a 'phase modulation'.

A conceptual model like this one must be viewed as a powerful means to highlight problems that must then be tackled from a more physical approach. Even if this simple model fits the proxy record in an impressive way, it remains to link its states and thresholds with a physical representation. Although the northern hemisphere summer insolation threshold for entering a glaciation seems a reasonable assumption [19], which closely corresponds to the traditional Milankovitch theory, the nature of the deglaciation threshold and state remains quite mysterious.

Previous studies [22,23] suggested that the isostatic bedrock depression following a glacial maximum could be the central phenomenon in the asymmetric 100 kyr cycle. However, these simple ice sheet models often contain unrealistic parameter values. Moreover, new evidences from Antarctic ice cores [24] show that during terminations, CO_2 and Antarctic temperature have already reached interglacial levels before any significant sea level rise. Thus, the central phenomenon of deglaciation cannot be the isostatic bedrock depression, even if this factor is certainly important during the rapid retreat of ice sheets. Moreover, CO_2 and Antarctica temperature seem to play an important role [25].

A possible hypothesis for the deglaciation threshold may be through the North American ice sheet. The idea of a considerable build-up of ice volume necessary for a complete deglaciation may correspond to a considerable build-up of the Laurentide ice sheet, the largest ice sheet that has disappeared since the last glacial maximum. The conjunction of a large Laurentide ice sheet and important northern hemisphere summer insolation may lead to large freshwater discharges in the North Atlantic ocean. These melt water pulses would lead to a warming of the southern ocean [26], through a change in the thermohaline circulation. This opposite phasing of both hemispheres, described as a bipolar sea-saw [27] has been observed in Greenland and Antarctica ice core records when synchronized with methane [28,26], but also in different types of oceanic models [29]. This warming, if in phase with the astronomical forcing, may lead to a significant out-gassing of CO_2, sufficient to induce a significant global planetary warming and thus a deglaciation [9]. This may be confirmed by the strong similarity between CO_2 and Antarctic temperature for the last four terminations [24]. For example during the last deglaciation, a drastic reduction of the thermohaline circulation occurred during Heinrich Event 1 (H1) around 17 kyr BP [30], which corresponds to a major increase of CO_2 concentration [31]. We have to note that the ice volume started to decrease several thousands of years before HI (between 21 and 18 kyr BP, depending on authors and dating methods [32]). This is not in contradiction with our explanation: the decrease of ice volume between $\sim$20 kyr and $\sim$17 kyr is due to the increase of northern latitudes summer insolation, whereas H1 marks the threshold crossing, the 'point of no return'. This entire hypothesis however remains to be demonstrated by physical evidences.

4. Conclusion

The response of the Earth climate to insolation forcing can be described by two different regimes, 'glaciation' and 'deglaciation', that can be switched according to threshold crossings. Our study indicates that the deglaciations start when a combination of insolation and ice volume is large enough. The ice volume is 'additive' with the insolation forcing in the triggering of deglaciations: either the ice volume or the insolation needs to be sufficiently large. This hypothesis allows us to precisely simulate the amplitude modulations of the sea level record over the past million years. This threshold model also allows for varying phase between insolation and ice volume. In particular, it could explain why termination II occurs earlier than expected and suggests that termination III may be an 'in extremis' deglaciation, which started rather late with respect to the other terminations, just before the maximum of insolation.

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