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The Antarctic ice sheet and the triggering of deglaciations

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Abstract

A new physical mechanism involving the Antarctic ice-sheet extent is able to link climatic and CO_2 glacial–interglacial changes. It is furthermore able to explain many features of the glacial cycles, like the 100 kyr oscillations or the peculiarities of stage 11 (about 400 kyr BP). Indeed, from recent results, the glacial ocean bottom waters were possibly much more saline and consequently, may have an unsuspected large density. This glacial deep stratification could account for a significant part of the glacial–interglacial CO_2 difference. The formation of these waters around Antarctica involves brine rejection over the continental shelves and is directly linked to changes in sea ice formation and Antarctic ice-sheet extent. Using this new scenario, we formulate a conceptual model that reproduce the succession and the amplitude of glaciations over the last few million years. This new mechanism furthermore provides the clue to understanding the changes in the main climatic frequency from 23 to 41 ka about 3 Myr ago and from 41 to 100 ka about 1 Myr ago. The Antarctic ice-sheet influence on bottom water formation could provide the "missing piece" to complete the astronomical theory of Quaternary climates.

Keywords: paleoclimatology; glacial cycles; Milankovitch theory; carbon cycle; climate dynamics

The successive climates of the Quaternary are increasingly well documented, in particular, through marine sediment and ice core analysis. The ultimate cause of these glacial–interglacial variations is the changing parameters of the Earth's orbit [1], but the physical mechanisms involved are still mostly unknown. This problem can be better formalized in terms of a relaxation oscillation, as illustrated by multiple-state or threshold models [2–4]. A definition of distinct climatic states "glacial" and "interglacial" are quite natural from records of deep ocean temperature, CO_2, or Southern Ocean temperature [5–8], suggesting a link between the rapid disintegration of Northern Hemisphere ice sheets, the high atmospheric CO_2 levels during this time [3] and the higher deep-ocean temperatures [5,6,9]. Many authors have

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focussed on the glacial pCO_2 problem and tried to explain why glacial conditions led to such lower CO_2 levels. But when considering a dynamical climate, the key question is almost the opposite. Indeed, if we want to relate CO_2-induced deglaciations to glacial maxima, the problem becomes: how can a glacial maximum trigger a deglaciation through oceanic CO_2 release [3]?

Box models of the oceanic carbon cycle have provided many possible scenarios for low glacial atmospheric CO_2, most of them disproven by more realistic 2D or 3D models. Box models use small or no vertical mixing. Consequently, they can store large amounts of carbon in the deep ocean [10,11]. More realistic models explicitly represent vertical diffusion. They have difficulties to decrease significantly atmospheric CO_2 without assuming drastic changes, for instance in global fertilization or alkalinity. The problem is to keep carbon efficiently away from the ocean surface, if possible in its deepest levels. In today's deep ocean, vertical density gradients are very small. Deep waters are cold but rather fresh and can easily be mixed upwards, inasmuch as intermediate waters have a similar density. The recent suggestion [12] that glacial deep waters were cold and very salty has been experimentally verified by the analysis of pore waters in marine sediments [13]. The up-to-now longest, multithousand year experiment of a glacial-type climate with a coupled ocean-atmosphere model [14] also produces very cold and salty bottom waters. Although the local details of AABW formation are not represented in such a model, this result suggests that, on a larger scale, the Antarctic region becomes drier during glacials, sea ice further isolates the ocean and is transported from the continental margins to more northern latitudes. Consequently, the surface waters around Antarctica become saltier. This reversal of the surface ocean freshwater balance in the high-latitude Southern Ocean should lead to the formation of extremely dense bottom waters by brine rejection above continental margins. These very dense waters should fill the deep ocean up to about 2000 m [12], a strong vertical density difference between intermediate and bottom waters should appear, and deep waters could store considerable amounts of carbon. The impact on atmospheric CO_2 is illustrated on Fig. 1, using a simple geochemical box model [15]. First, the increase of deep stratification favors the oceanic storage of carbon, as illustrated by the decrease in vertical mixing m on Fig. 1. Second, and most importantly, the sensitivity of the carbon cycle is greatly enhanced, and any change in the physical or biological parameters of the carbon cycle will induce much larger atmospheric CO_2 changes, as shown on the figure when changing the large scale advection v. Besides, explaining about a 40 ppm lower glacial pCO_2 is sufficient for our purpose, inasmuch as additional known processes are also involved, which could almost account for the other 40 ppm, like surface ocean cooling, increase in dust transport, and Fe-fertilization in the Southern Ocean [16].

If two distinct "glacial" and "interglacial" states are defined as above by different deep-ocean stratification and, consequently, different pCO_2 levels, we need to understand how and why climate switches from one state to the other by looking at the mechanisms influencing bottom water formation in the South. In order to switch from "interglacial" to "glacial", the traditional Milankovitch theory is probably sufficient. Indeed, the build-up of continental ice in the Northern Hemisphere cools sufficiently the global climate to reverse the hydrological budget around Antarctica: climate becomes drier, sea ice better insulates ocean waters from precipitations, sea ice and icebergs melt further north, brine formation increases. Some insolation forcing in the Southern Hemisphere may also influence Antarctic brine formation. But the crucial switch is from "glacial" to "interglacial", and here, the Antarctic ice sheet has a central role. The Antarctic ice sheet grows slowly during glaciations under the influence of changing sea level and oceanic heat flux [17]. It reaches its maximum extent at the limits of the continental shelf a few thousand years after glacial maxima, as attested by geological data [18,19]. This affects bottom water formation. Indeed, brine rejection is almost permanent around Antarctica, newly formed sea ice being blown away from the coast by the catabatic winds. Because of the strong negative thermal feedback associated with brine rejection, this process is favoured above continental shelves. Indeed, when salt is rejected from forming sea ice, it sinks and is replaced by deeper, warmer waters which prevent further formation of sea ice. Brine rejection is more difficult when the water depth is large because it is necessary to cool the whole column down to almost the freezing point. Consequently, even if open ocean

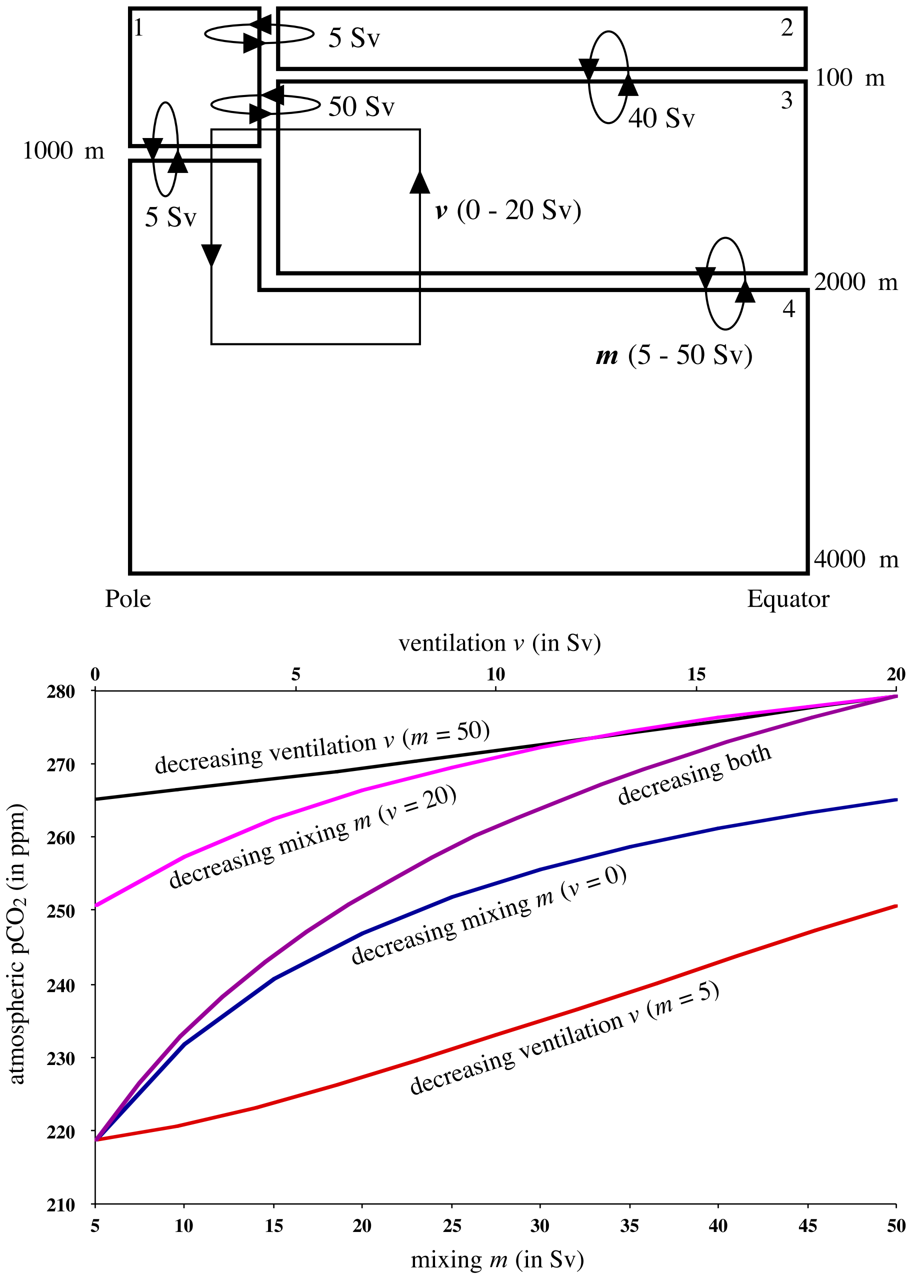

Fig. 1. Top: box model configuration (see [15] for details). Bottom: results obtained when varying the deep mixing *m* and the overturning *v*. Following glacial data [13], a ten-fold decrease in vertical mixing *m* is a reasonable assumption. Decreasing *m* alone can account for about 30 ppm, which is already quite large. But more importantly, the sensitivity to other changes is largely enhanced. For instance, a decrease in *v* is much more efficient in driving atmospheric CO_2 down with low deep-mixing, while decreasing *v* alone with high mixing accounts for less than 15 ppm. Decreasing both *m* and *v* could account for up to 60 ppm, which would explain a significant amount of the glacial interglacial CO_2 difference.

sea ice formation is possible in stratified areas, it cannot produce large amounts of salty bottom waters. The Antarctic ice sheet will invade completely the continental shelves at the glacial paroxysms. Salty bottom water formation being more difficult, the deep ocean stratification will cease after a few thousand years, and atmospheric CO_2 will increase significantly, leading to a global warming and a major reduction of the Northern Hemisphere ice sheets. The following Northern Hemisphere summer insolation maximum, even a modest one like during MIS 11, will be able to melt the remaining Northern Hemisphere ice sheets.

A mathematical expression of this conceptual model follows. We need three variables: global ice volume V forced both by Northern Hemisphere summer insolation and atmospheric pCO_2; extent of Antarctic ice sheet A forced by sea level changes (i.e., by V); and atmospheric CO_2 C linked primarily to deep-water state «glacial» or «interglacial». The oceanic switch is forced by the «salty bottom waters formation efficiency» parameter:

$$F = aV - bA - cI_{60} + d$$

which increases with V and decreases with A. a, b, c, and d are constant coefficients. Indeed, F should increase when global climate cools (through V) and decrease when continental shelf areas are reduced (through A). I_{60} is the daily insolation (60°S, 21st February) inasmuch as a reduced sea-ice extent during late Austral summer could affect the Southern Ocean heat budget, and consequently, warm the regional climate, thus affecting F. Coefficient c is typically very small. When F is negative, the ocean is in "interglacial" mode and reciprocally. The model equations are:

$$dV/dt = (V_R - V)/\tau_V$$

$$dA/dt = (V - A)/\tau_A$$

$$dC/dt = (C_R - C)/\tau_C$$

with $V_R = -xC - yI_{65} + z$ (i.e., ice volume is driven by insolation and CO_2); $C_R = \alpha I_{65} - \beta V + \gamma H(-F) + \delta$ (i.e., CO_2 is driven by some precessional forcing, here I_{65}, by global climate V and by deep ocean stratification); H is the Heaviside function (H=1 if F<0; H=0 otherwise); I_{65}=insolation 65°N, 21st June; τ_V, τ_A, and τ_C are time constants; x, y, z, α, β, γ, and δ are constants. Contribution of Antarctica to V is neglected. V is limited by a minimum zero value. This system is linear, except for the discontinuity represented by $H(-F)$, which reflects either a nonlinearity in the carbon cycle, or more probably, a nonlinearity of the interactions between deep stratification, bottom water formation, and thermohaline circulation. The parameter values (Table 1) are chosen in order to reproduce qualitatively the main features of the glacial–interglacial cycles, but no precise tuning was performed. For example, α=0 (no precessional forcing in the CO_2) or c=0 (no southern insolation forcing) do not change the timing of terminations but only the qualitative agreement with data; other precessional or obliquity forcings could very well be used here. The integration of this simple system is illustrated on Fig. 2 for the last 500 kyr. The last four cycles are reasonably well represented in comparison with the available data [7,20–22]. pCO_2 is leading ice volume changes by a few thousand years at each termination in agreement with data [7,8], and main transitions occur after the Antarctic ice sheet reaches a maximum. Although on this example the spectral power in the 100 ka band is probably too low, the success of this model in reproducing simultaneously the ice volume and the atmospheric CO_2 data is

Table 1
Parameter values used to produce Fig. 2

Parameter	Value	Range
τ_V	15 (kyr)	13.1–18.1
τ_C	5 (kyr)	3.1–15
τ_A	12 (kyr)	9.5–26
x	1.3	1.23–1.44
y	0.5	0.4–0.64
z	0.8	0.77–0.82
α	0.15	0–0.35
β	0.5	0.46–0.54
γ	0.5	0.37–0.6
δ	0.4	0.39–0.42
a	0.3	0.26–0.39
b	0.7	0.63–0.74
c	0.01	0–0.15
d	0.27	0.253–0.302

The range given in the table is defined as the interval within which one parameter can be changed without modifying the timing of the last five terminations. A narrow range often indicates that two parameters need to be adjusted simultaneously to keep results qualitatively unchanged over a much wider interval.

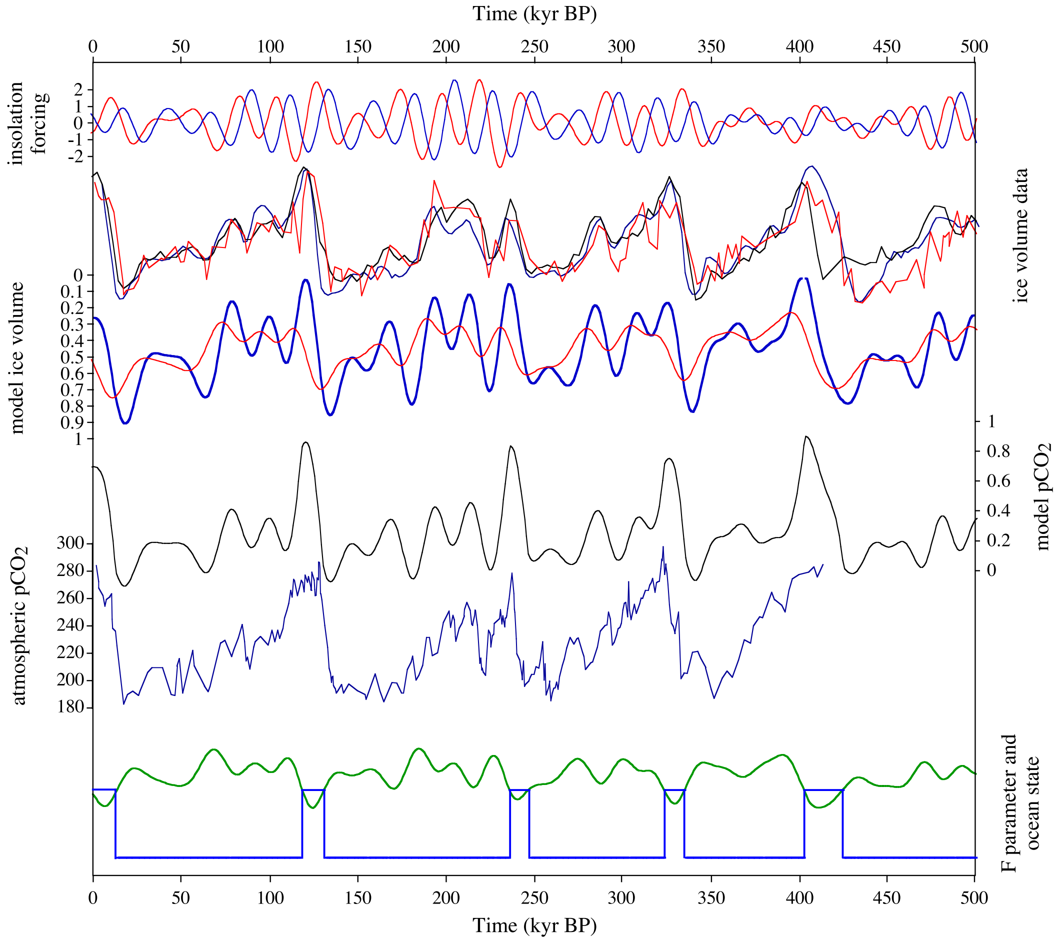

Fig. 2. Model results and data over the last 500 kyr. Top, the forcing: daily insolation at 65°N, 21st June (red) and at 60°S, 21st February (blue), following [24]. Below, different sea-level curves from the isotopic composition of planktonic foraminifera: blue, from [20]; red, from [21]; black, from [22]. Below, the simulated ice volume (blue) and the Antarctica area (red). Below, the simulated CO_2 levels (black) and the Vostok CO_2 data from [7] (blue). Bottom, the critical F parameter (green) and the state of the ocean (violet): glacial or interglacial.

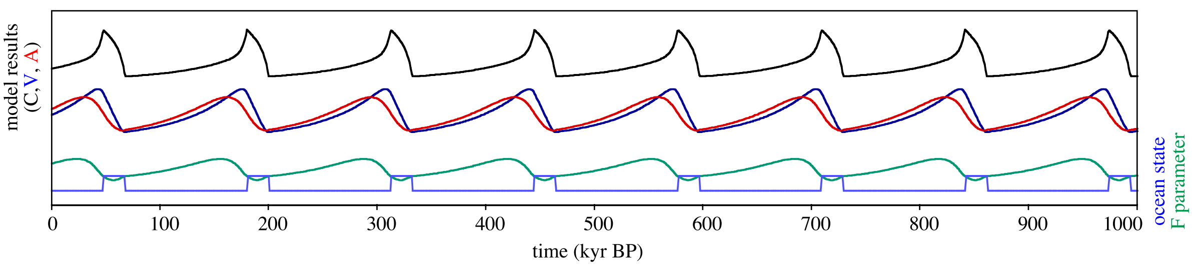

Fig. 3. Model results in the absence of forcing. For the parameter values listed in Table 1, the model generates a 132-kyr oscillation. By changing, for instance, the value of d, periodicities from a few tens of kyr up to a few hundreds of kyr are possible. Top, the simulated CO_2 levels (black line). Below, the simulated ice volume (blue line) and Antarctica area (red line). Bottom, the critical F parameter (green) and the state of the ocean (blue): glacial or interglacial.

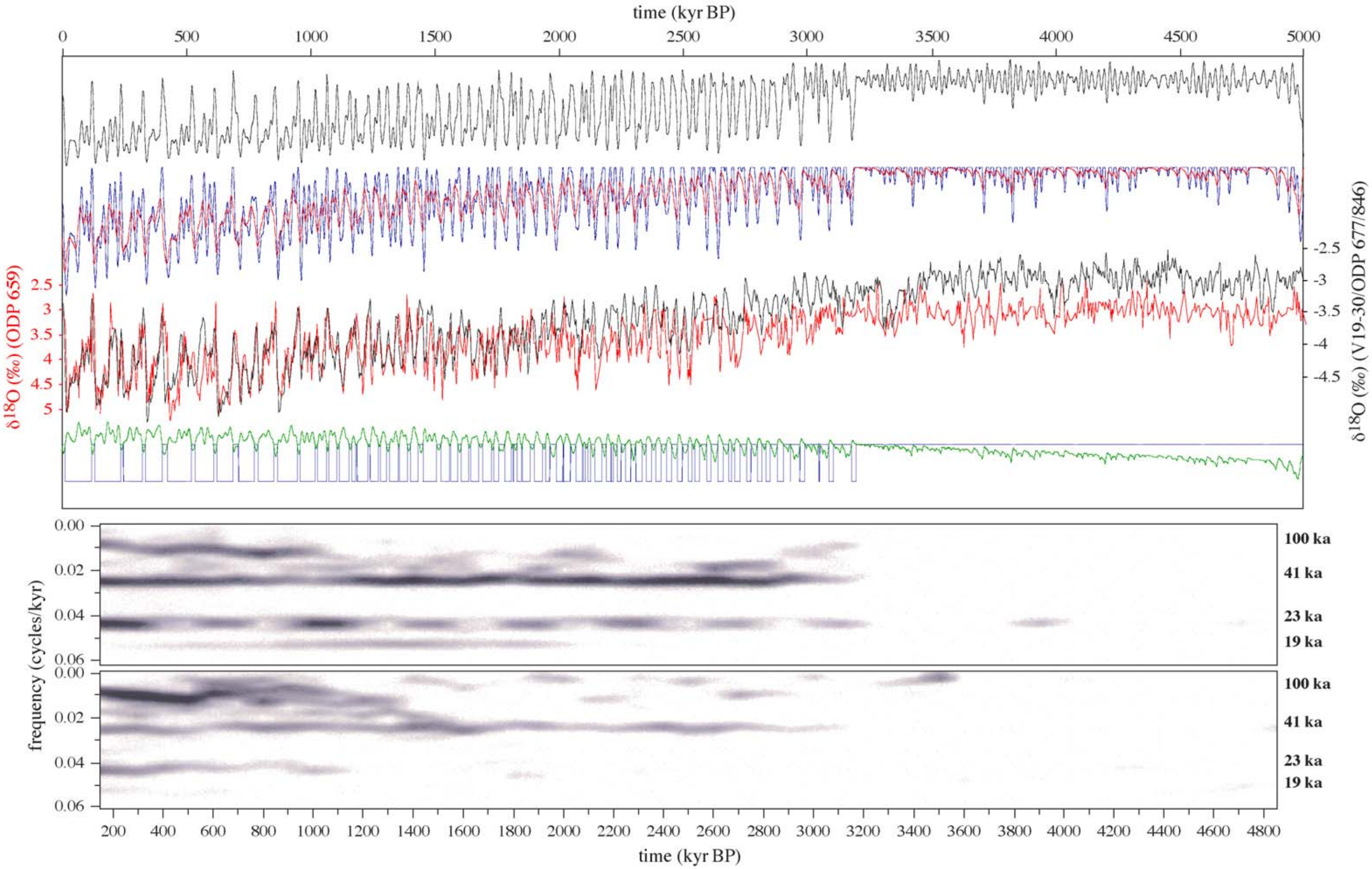

Fig. 4. Model results over the last 5 Myr, same parameters as in Fig. 2 but using a small drift on F ($F=aV-bA-cI_{60}+d+kt$, with t the time and k=8.5 10^{-5} kyr^{-1}). Top, the simulated CO_2 levels (black line). Below, the simulated ice volume (blue line), the Antarctica area (red line). Below, sea-level curves from planktonic foraminifera δ^{18}O: red, from [21]; black, from [22]. Bottom, the critical F parameter (green) and the state of the ocean (blue): glacial or interglacial. Below, the time-frequency analysis of the modeled ice volume. Bottom, the time-frequency analysis of the data [22]. The transition from 23 to 41 ka oscillations occurs around 3 Myr BP. The transition from 41 to 100 ka ones occurs around 1 Myr BP.

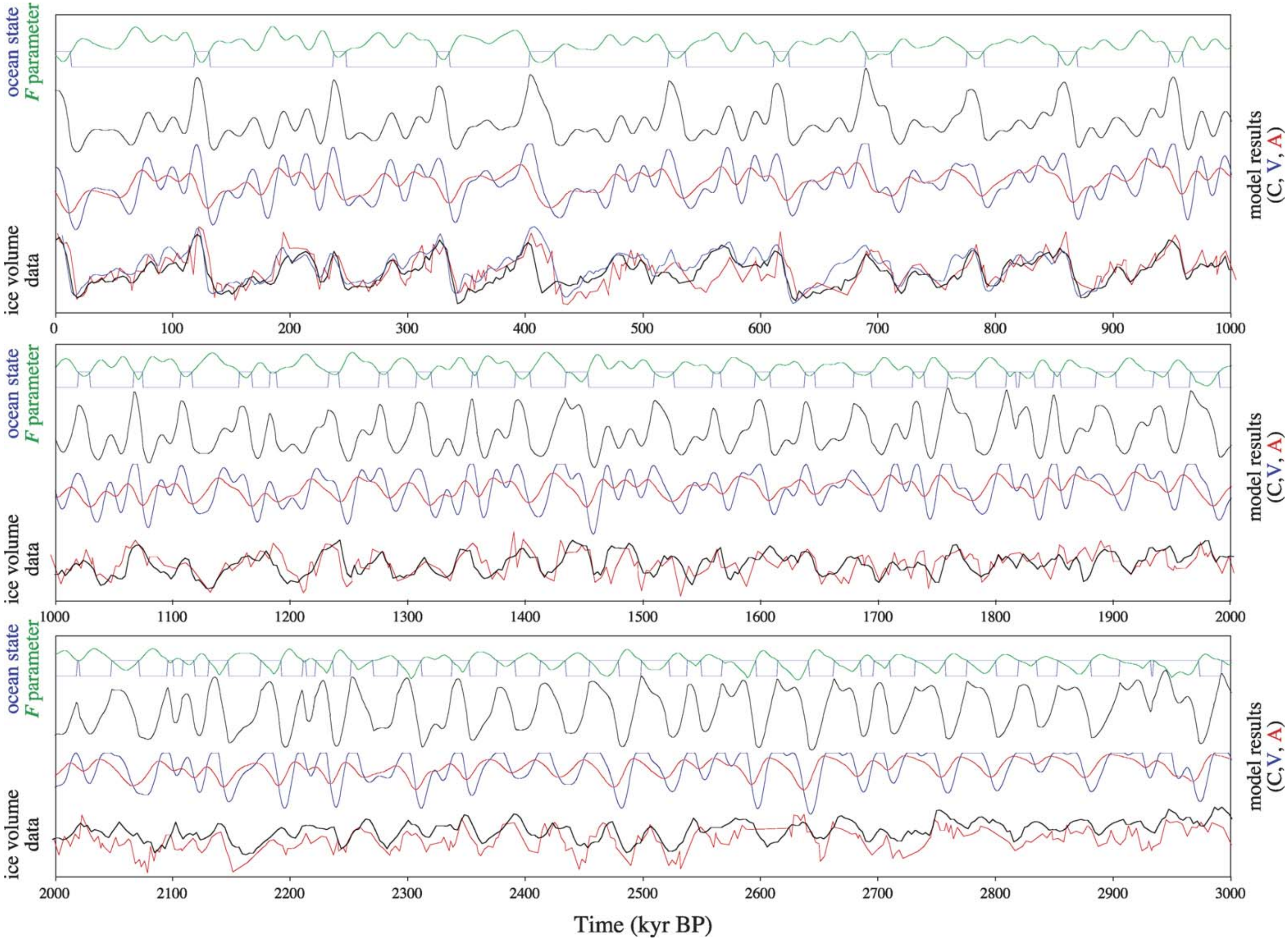

Fig. 5. Same as Fig. 4 but with more details. Top between 0 and 1 Myr BP, middle between 1 and 2 Myr BP, bottom between 2 and 3 Myr BP. Almost all glacial–interglacial transitions are located at the correct time. Almost all glacial–interglacial cycles have the right shape.

encouraging, and the underlying physical mechanisms should therefore be studied carefully in the future with more sophisticated models. This model turns out to generate a self-sustained relaxation oscillation when there is no insolation forcing (I_{65} and I_{60}=0). For the parameter values listed in Table 1, this oscillation has a periodicity of 132 kyr, as can be shown on Fig. 3.

An interesting result of our conceptual model concerns the more remote past and the frequency changes of the main climatic variations. By introducing a small drift in the oceanic critical parameter F over the last million years, the model reproduces most glacial–interglacial cycles over this period (Figs. 4 and 5). Starting from 5 Myr BP, the model first experiences rapid (23 ka) oscillations, while the ocean stays in a permanent «interglacial» mode without efficient CO_2 sequestration. In this first part, the main nonlinearity of the system being inactive, the model response reflects mostly the insolation forcing. Around 3 Myr BP, the drift in F induces switches between the two ocean states together with obliquity changes (41 ka oscillations). The so-called «mid-Pleistocene» transition occurs afterwards around 1 Myr BP, where the dominant periodicity becomes 100 ka. In contrast to our previous conceptual models [2,4], termination VI (MIS 14 to 13, about 530 kyr BP) is now predicted at the correct time. The most sensitive time period is the mid-Pleistocene transition itself, where the shape of the oscillations is sometimes slightly different when compared to records. Introducing a drift directly on the CO_2, as often mentioned before [2,23], does not produce, in our model, such frequency shifts. This slow drift in «bottom water formation efficiency» from the late Pliocene into the Pleistocene could be possibly linked to a slow change in the geometry of the area around Antarctica due to glacial erosion of the Antarctic continent or to continental drift and plate tectonics.

The match between this idealized model and the geological data is rather good. The clue of this success is in the identification of the largest nonlinearity of the system: an ocean-CO_2 switch based on the vertical density structure of the deep ocean. A similar argument has been recently and independently suggested for the global cooling of the late Pliocene [25]. Previous models [2–4] were quite successful when assuming a priori a multistable climate system. But the identification of the physical processes involved provides a large step forward, as illustrated by the better success of this new model. Clearly, this success is not a proof that our mechanism is causing glacial–interglacial cycles, but we hope that it will promote some necessary further studies in order to validate or invalidate this new theory.

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References

[1] J.D. Hays, J. Imbrie, N.J. Shackleton, Variations in the Earth's orbit: pacemakers of the ice ages, Science 194 (1976) 1121–1132.

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

[3] D. Paillard, Glacial cycles: toward a new paradigm, Reviews of Geophysics 39 (2001) 325–346.

[4] F. Parrenin, D. Paillard, Amplitude and phase modulation of glacial cycles from a conceptual model, Earth and Planetary Science Letters 214 (2003) 243–250.

[5] L.D. Labeyrie, J.C. Duplessy, P.L. Blanc, Variations in mode of formation and temperature of oceanic deep waters over the past 125,000 years, Nature 327 (1987) 477–482.

[6] C. Waelbroeck, L. Labeyrie, E. Michel, J.C. Duplessy, J.F. McManus, K. Lambeck, E. Balbon, M. Labracherie, Sea-level and deep water temperature changes derived from benthic foraminifera isotopic records, Quaternary Science Reviews 21 (2002) 295–305.

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

[8] J.-J. Pichon, L. Labeyrie, G. Bareille, M. Labracherie, J. Duprat, J. Jouze, Surface water temperature changes in the high latitudes of the Southern Hemisphere over the last glacial–interglacial cycle, Paleoceanography 7 (1992) 289–318.

[9] J.-C. Duplessy, L. Labeyrie, C. Waelbroeck, Constraints on the ocean oxygen isotopic enrichment between the Last Glacial Maximum and the Holocene: paleoceanographic implications, Quaternary Science Reviews 21 (2002) 315–330.

[10] E. Michel, L. Labeyrie, J.-C. Duplessy, N. Gorfti, M. Labracherie, J.-L. Turon, Could deep subantarctic convection feed the world deep basins during the last glacial maximum? Paleoceanography 10 (1995) 927–942.

[11] B.B. Stephens, R.F. Keeling, The influence of Antarctic sea ice on glacial–interglacial CO2 variations, Nature 404 (2000) 171–174.

[12] R.F. Keeling, B.B. Stephens, Antarctic sea ice and the control of Pleistocene climate instability, Paleoceanography 16 (2001) 112–131.
[13] J.F. Adkins, K. McIntyre, D.P. Schrag, The salinity, temperature and $\delta^{18}O$ of the glacial deep ocean, Science 298 (2002) 1769–1773.
[14] R. Stouffer, S. Manabe, Equilibrium response of thermohaline circulation to large changes in atmospheric CO_2 concentration, Climate Dynamics 20 (2003) 759–773.
[15] D. Paillard, M. Ghil, H. Le Treut, Dissolved organic matter and the glacial–interglacial pCO_2 problem, Global Biogeochemical Cycles 7 (1993) 901–914.
[16] L. Bopp, K.E. Kohfeld, C. Le Quéré, O. Aumont, Dust impact on marine biota and atmospheric CO_2 during glacial periods, Paleoceanography 18 (2003) 1046.
[17] C. Ritz, V. Rommelaere, C. Dumas, Modeling the evolution of Antarctic ice sheet over the last 420,000 years: implications for altitude changes in the Vostok region, Journal of Geophysical Research, D: Atmospheres 106 (2001) 31,943–31,964.
[18] G.H. Denton, T.J. Hughes, Reconstructing the Antarctic ice sheet at the Last Glacial Maximum, Quaternary Science Reviews 21 (2002) 193–202.
[19] J.B. Anderson, S.S. Shipp, A.L. Lowe, J.S. Wellner, A.B. Mosola, The Antarctic ice sheet during the Last Glacial Maximum and its subsequent retreat history: a review, Quaternary Science Reviews 21 (2002) 49–70.
[20] F.C. Bassinot, L.D. Labeyrie, E. Vincent, X. Quidelleur, N.J. Shackleton, Y. Lancelot, The astronomical theory of climate and the age of the Brunhes–Matuyama magnetic reversal, Earth and Planetary Science Letters 126 (1994) 91–108.
[21] R. Tiedemann, M. Sarnthein, N.J. Shackleton, Astronomic timescale for the Pliocene Atlantic $\delta^{18}O$ and dust records of Ocean Drilling Program site 659, Paleoceanography 9 (1994) 619–638.
[22] N.J. Shackleton, A. Berger, W.R. Peltier, An alternative astronomical calibration of the lower Pleistocene timescale based on ODP site 677, Transactions of the Royal Society of Edimburgh. Earth Sciences 81 (1990) 251–261.
[23] M.E. Raymo, The timing of major climate terminations, Paleoceanography 12 (1997) 577–585.
[24] J. Laskar, The chaotic motion of the solar system: a numerical estimate of the size of the chaotic zones, Icarus 88 (1990) 266–291.
[25] D.M. Sigman, S.L. Jaccard, G.H. Haug, Polar ocean stratification in a cold climate, Nature 428 (2004) 59–63.