LONG-TERM VARIATIONS IN INSOLATION AND THEIR EFFECTS ON CLIMATE, THE LLN EXPERIMENTS

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Abstract. Used to test the Milankovitch theory over the last glacial-interglacial cycles, the Louvain-la-Neuve two-dimension Northern Hemisphere climate model shows that orbital and CO_2 variations induce, in the climate system, feedbacks sufficient to generate the low frequency part of the climatic variations over the last 200 kyr. Initiation and termination of glacial cycles cannot indeed be explained without invoking both the fast feedbacks associated with atmospheric processes (water vapor, cloud, snow and sea ice) and the slower feedbacks associated with coupling to other parts of the climate system, in particular the land ice-sheet buildup and disintegration. This model shows that long-term changes in the Earth's orbital parameters lead to variations in the amount of solar radiation received at the top of the atmosphere, which in turn act as a pacemaker for climatic variations at the astronomical frequencies, through induced albedo-temperature and greenhouse gases-temperature feedbacks. Spectral analysis of the Northern Hemisphere global ice volume variations simulated under both insolation and CO_2 forcings reproduces correctly the relative intensity of the peaks at the orbital frequencies as seen in SPECMAP data. Except for variations with time scales shorter than 5 kyr, the simulated long-term variations of total ice volume are comparable to that reconstructed from deep-sea cores. For example, the model simulates glacial maxima of similar amplitudes at 134 kyr BP and 15 kyr BP, followed by abrupt deglaciations.

Key words: Astronomical theory, CO_2 forcing, insolation, modelling past climates

1. Transient Models of the Climate System

Since the publication of the papers by Hays et al. (1976) and Berger (1977), a number of modelling efforts have attempted to explain the relation between astronomical forcing and climatic change (Berger, 1995). Most of these modelling studies have focused on the origin of the 100-kyr cycle, the most important cycle found in proxy records of the ice volume. These studies have been motivated by the recognition that the amount of insolation perturbation at the 100-kyr period is not enough to cause a climate change of ice-age magnitude. Specific explanations for the large 100-kyr variance generally fall into two classes. In the first case, orbital variations are essential to ice-volume fluctuations, but nonlinear interactions in the air-sea-ice system modify the signal. In the second case, ice-volume fluctuations result from inherently nonlinear interactions in the air-sea-ice system, with orbital variations serving only to phase-lock the variations, at the preferred time scales.

For models where the ice-volume fluctuations are primarily driven by orbital forcing, Imbrie et al. (1992) showed that at the 23-kyr and 41-kyr periods, ice volume responds linearly to orbital forcing, but at 100 kyr the effect is nonlinear

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(Imbrie et al., 1993). In particular, the 100-kyr power can be generated by transmission of 19-kyr and 23-kyr periods through a nonlinear system (Wigley, 1976), producing substantial power in both harmonics and subharmonics.

Other approaches to the 100-kyr cycle focus on nonlinear interactions between ice-sheet accumulation and ablation, ice-sheet flow, elastic lithosphere, and viscoelastic mantle (e.g. Pollard, 1983). In contrast, results from a model incorporating ice sheets, but with different representations of the physical processes, suggest that the external orbital forcing, together with internal forcing, drives the observed 100-kyr cycle. Le Treut et al. (1988) modeled such internally generated glacial-interglacial fluctuations which, when orbitally forced, produce characteristic Milankovitch periods, including harmonics (e.g., 10 kyr) and subharmonics (e.g., 100 kyr). Recent dynamic models developed by Maasch and Saltzman (1990) are also particularly interesting because they are able to reproduce, with some success, both the 100-kyr cycle and the transition from dominant 41-kyr ice volume fluctuations prior to 900 kyr BP to the dominant 100-kyr ice volume fluctuations after 900 kyr BP. The latter transition may also reflect the existence of "multiple equilibrium" states, whereby slowly changing boundary conditions can cause an abrupt transition in the climate state.

Although these models are based on parameters which are considered to be physically plausible, they are all highly simplified. None of them can be considered as a fully realistic representation of the whole climate system. What these models do confirm is that the response to orbital forcing is nonlinear and that it involves some elements of internal forcing. Whether the external orbital forcing drives the internal forcing, phase-locks the oscillations of an internally driven system, or acts as a pace-maker for the free oscillations of an internally driven system remains, however, an open question.

As a consequence, the discussion of how the climate system responds to orbital forcing requests the construction of a physically realistic model of the time-dependent behaviour of the coupled climate system, including the atmosphere, the oceans, the cryosphere, the lithosphere and the biosphere. Unfortunately, such a model is likely to be too complex, given the constraints of computing power and speed and our lack of knowledge in the biogeochemistry of the climate system, in particular. This is why a transient simulation of global climatic change at the thousands of years time scale with more simple 2-D coupled models of the climate system is now tentatively used for testing the astronomical theory of paleoclimates (Berger et al., 1990a; Deblonde and Peltier, 1991).

It was suggested earlier (Berger, 1979) that the time-evolution of the latitudinal distribution of the seasonal pattern of insolation is the key factor driving the behavior of the climate system while the complex interactions between its different parts amplify this orbital perturbation. That dynamical behaviour of the seasonal cycle suggests that time-dependent coupled climate models might be able to test whether or not the astronomical forcing can drive the long-term climatic variations.

Such a climate model (designated by LLN, Louvain-la-Neuve) which links the Northern Hemisphere atmosphere, ocean mixed layer, sea ice, ice sheets and continents, has been validated for the present-day climate. It is fully documented in Gallée et al. (1991). The atmosphere-ocean model is asynchronously coupled to a model of the three main Northern Hemisphere ice sheets and their underlying bedrock in order to assess the influence of several factors (processes and feedbacks), including the snow-surface albedo over the ice sheets, upon ice-age simulations using astronomically-derived insolation and CO_2 data from the Vostok ice core as forcings (Berger et al., 1990a; Gallée et al., 1992).

2. Insolation Forcing over the Last 220 kyr

June insolation at 65° N (Figure 2) is very often used as a guideline for the analysis of climatic changes and, in particular, for ice volume changes. According to the Milankovitch theory (1941; Berger, 1988; Milankovitch, 1995), these latitude and time are indeed the most sensitive ones for explaining the ice-sheets waxing and waning. Moreover, using this particular latitude does not bias the conclusions because for a given month, the insolation changes are in phase for all latitudes, except those which are close to the polar night (where the energy received from the Sun is very weak). Actually, the insolation changes are in phase for all latitudes where precession dominates (Berger et al., 1993a). It must be stressed here that in the LLN 2-D model experiments described in the next section, insolations for all latitudes from the equator to the North Pole are used at each time step of the simulation (i.e. every three days).

As we are primarily interested in the climate response of the LLN 2-D model over the last glacial- interglacial cycles, let us first concentrate upon the insolation behaviour over the last 220 kyr BP. The 65° N June irradiation is maximum around 220 kyr BP ($\sim$550 W m^{-2}), then decreases to $\sim$460 W m^{-2} over the next 10,000 years and increases again up to 535 W m^{-2} at 199 kyr BP. At 215 kyr BP the value is 516 W m^{-2}; this is the date of isotopic event 7.3, it means the interglacial peak of isotopic stage 7 in the stack oxygen-isotope record of SPECMAP (Martinson et al., 1987). It is important to note that from 244 to 190 kyr BP, i.e. during the whole isotopic stage 7, the amplitude of the insolation variations is maximum, reaching more than 20% per 10,000 years, in particular at 65° N in June, a record over the last one million years. This is due to the maximum value of the eccentricity peaking at about 0.05 and a maximum variation of obliquity (Figure 1). Between 232 and 231 kyr BP, the obliquity reaches a minimum value of 22°, eccentricity is equal to 0.042 and summer solstice occurs at aphelion, which leads to an absolute minimum insolation of 434 W m^{-2} at 65° in June. Twelve thousand years later (at 219 kyr BP), $e = 0.048$, $\epsilon \sim 24°$ and summer solstice is at perihelion leading to a maximum of insolation of 550 W m^{-2} in June at 65° N, a record which occurs only three times over the last million years at 579, 219 and 128 kyr BP.

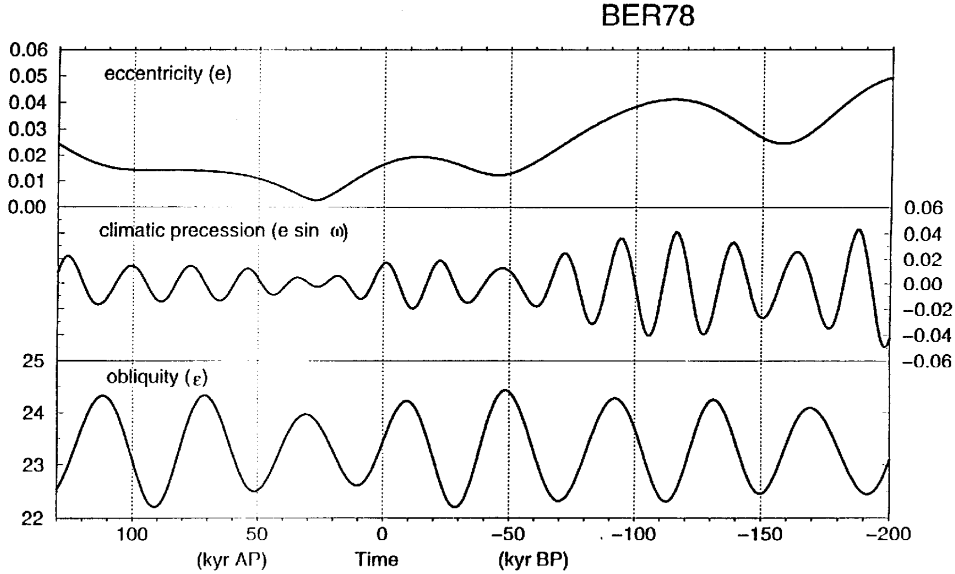

Figure 1. Long-term variations of the astronomical parameters (eccentricity, climatic precession and obliquity) over the last 200 kyr and the next 130 kyr (Berger, 1978).

A quick look to this 65° N June insolation over the last two glacial-interglacial cycles (Figure 2) shows easily that from 220 kyr BP to 70 kyr BP (∼ isotopic event 4.23), this insolation (and also the insolation of the other months and latitudes) is characterized by large variations, except from 165 to 135 kyr BP (Berger, 1978). Over the last glacial-interglacial cycle, the 65° N insolation starts to decrease at 133, 128 and 122 kyr BP for March, June and September respectively. Summer insolation therefore peaks around 125 kyr BP in high northern latitudes. The following minimum is reached 11 to 12 kyr later. For June, insolation decreases of almost 20% from 546 W m^{-2} to 440 W m^{-2}. These latitudes and months are characterized by a strong precession signal whereas, in December, obliquity dominates the spectrum at this latitude close to the polar night (Berger et al., 1993a), the 65° N December insolation maxima and minima occuring alternatively at 147, 130, 115 and 89 kyr BP. From 140 to 70 kyr BP (it means about during the whole isotopic stage 5), the insolation variations are almost as large as during isotopic stage 7. The seasonal and latitudinal changes in solar radiation caused by orbital changes are approximately twice as large at 125 kyr BP as at 9 to 6 kyr BP. This is because the eccentricity of the Earth's orbit was significantly larger then (0.04), so that the Earth-Sun distance was significantly smaller at perihelion (by 3%), and because perihelion passage occurred in northern summer around 125 kyr BP, whereas it occurs now in northern winter. In the Northern Hemisphere high latitudes, summer radiation was increased by more than 50 W m^{-2} (12–13%)

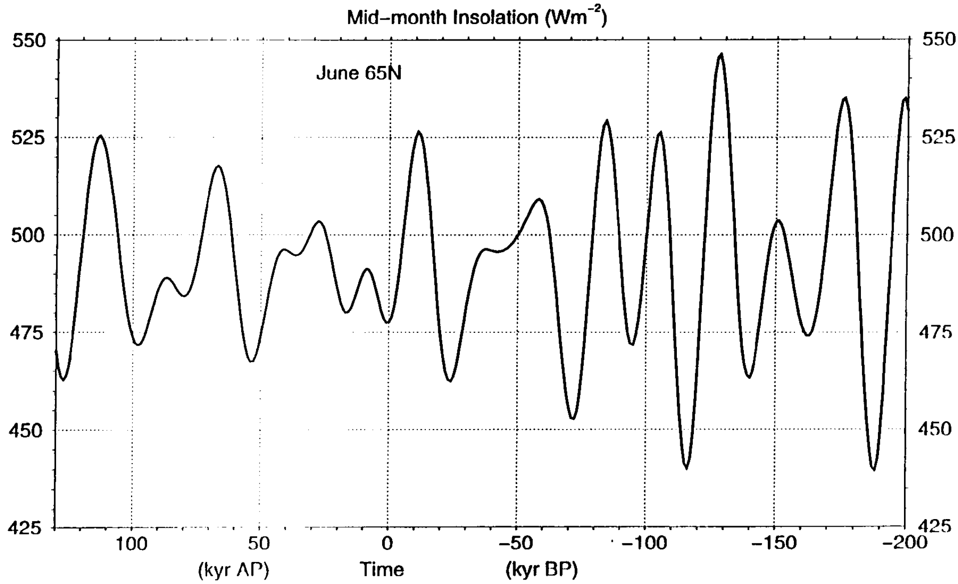

Figure 2. June mid-month insolation at 65° N over the last 200 kyr and the next 130 kyr (Berger, 1978).

compared to present, winter radiation was slightly decreased and the annual average insolation was increased by 3 to 4 W m^{-2}.

After 70 kyr BP, the amplitude of the insolation variations become smaller and between 60 and 34 kyr BP, the 65° N June insolation does even not change very much. It is also a time when the precessional period almost disappears creating a long cycle between the two minima of 70 and 20 kyr BP. A more or less similar behaviour will happen between 17 and 55 kyr AP.

3. Response of the LLN Model to the Insolation and CO_2 Forcings

A first set of 2-D modelling experiments (Berger et al., 1990a; Gallée et al., 1992) showed that the variations in the Earth's insolation alone are sufficient to induce feeedbacks in the climate system (Berger et al., 1990b, 1992, 1993b, 1993c), which amplify the direct radiative impact and generate large climatic changes. Sensitivity experiments to constant CO_2 concentrations (Gallée et al., 1992; Berger et al., 1996) show that the ice volume simulated by this version of the LLN 2-D model compares quite well with the geological reconstructions, provided this concentration is about 210 ppmv. In particular, the saw-thoothed shape of the ice volume curve is well reproduced. It seems therefore that CO_2 variations are not absolutely required to be taken into account to generate a rough 100-kyr cycle, which confirms the Hays et al. (1976)'s idea that the orbital forcing acts as a pacemaker of the ice ages (because of the sensitivity of the LLN model to the atmospheric CO_2 concentration, less and

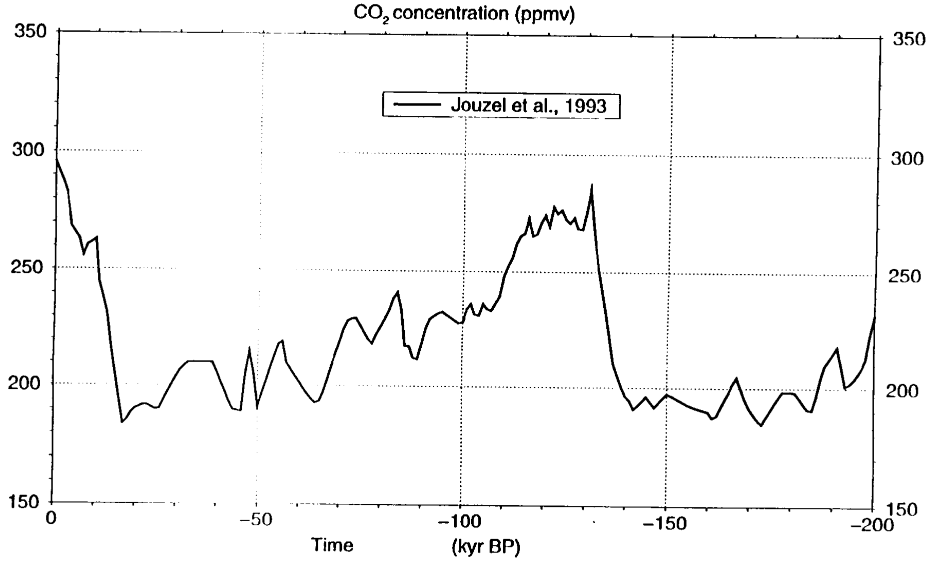

Figure 3. Atmospheric CO_2 concentration from 200 kyr BP to present. This CO_2 concentration has been reconstructed from the Vostok ice core by Jouzel et al. (1993) using the new EGT time scale.

less ice is simulated for constant concentrations higher than about 260 ppmv. In such cases, the model does not reproduce the 100-kyr cycle any more).

However, as the CO_2 concentration has varied in the past, it is important to see how the response of the LLN 2-D model to the orbitally induced forcing alone is modified by using the reconstructed variable CO_2 as an additional forcing (Gallée et al., 1993; Berger et al., 1996).

Using the Berger (1978) insolation and the Jouzel et al. (1993) CO_2 forcings (Figure 3) the integration was started 200 kyr BP with no Northern Hemisphere ice sheets; this seems to be a reasonable assumption according to sensitivity experiments described in Berger et al. (1996). Figure 4 shows that the Northern Hemisphere ice volume starts to grow only after 10 kyr. It reaches a first maximum of 39 $\times$ 10^6 km^3 at 180 kyr BP followed by a partial deglaciation leading to a minimum of 14 $\times$ 10^6 km^3 of ice at 170 kyr BP. Actually the Greenland ice sheet reaches its maximum size at $\sim$170 kyr BP, but the northern american and the Eurasian ice sheets reach their minimum size and start to increase again up to the penultimate glacial maximum of 133 kyr BP. At this maximum, the Northern Hemisphere ice volume amounts to 47 $\times$ 10^6 km^3, just about the same as at the Last Glacial Maximum. All the Northern Hemisphere ice sheets start to melt and totally disappear at 126 kyr BP, what is unrealistic according to GRIP (1993) which shows that the Greenland ice sheet has survived during the last interglacial.

It is interesting to note that the Northern Hemisphere ice sheets are melting totally three times in the LLN experiments: between 126 and 117 kyr BP, 100 and 97 kyr BP and 83 and 74 kyr BP. Although this is not realistic, it can be explained

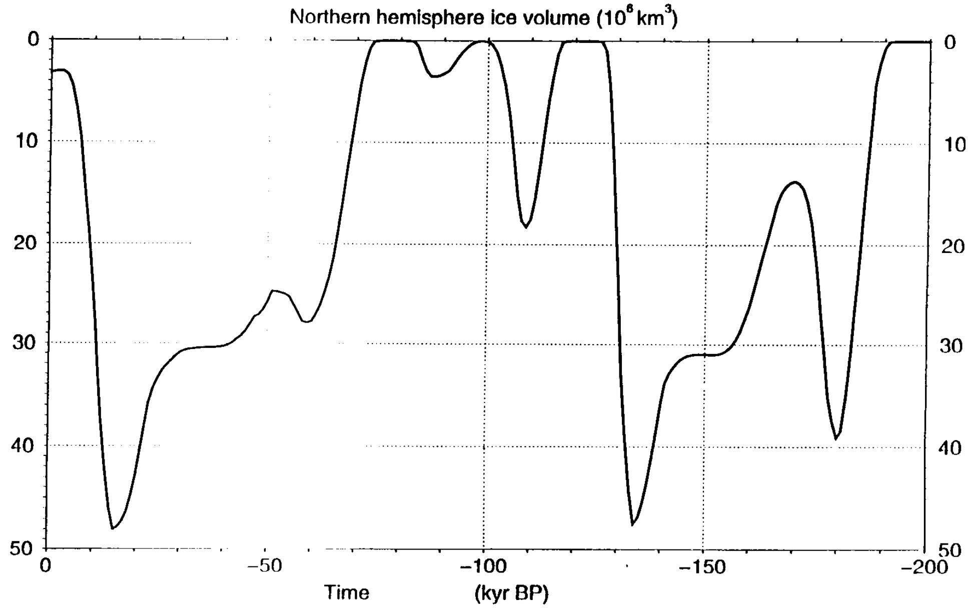

Figure 4. Simulated ice volume of the Northern Hemisphere using the LLN 2-D NH climate model forced by insolation and CO_2 (Berger and Loutre, 1996).

by the large insolation occuring at these times and the large CO_2 concentration between 130 and 115 kyr BP which continues to play an important role in the behaviour of the ice sheets over the next 30 to 40 kyr.

At 70 kyr BP, the ice sheets start to form quite rapidly to lead to a first ice volume maximum at 59 kyr BP. This is followed by a weak melting of $\sim 5 \times 10^6$ km^3 leading to a minimum in the simulated ice volume at 50 kyr BP. The ice sheets are then starting to grow again, slowly to about 30 kyr BP and then more rapidly to the Last Glacial Maximum where they amount to 47×10^6 km^3 of ice at 15 kyr BP. Finally, the model simulates a deglaciation from 15 kyr to 3 kyr BP leaving only the Greenland ice sheet in the Northern Hemisphere with roughly 3×10^6 km^3 of ice. Since 3 kyr BP, the ice volume is shown to increase slightly reaching 3.2×10^6 km^3 today which represents about the present value. Moreover, our model simulation is showing that the climate has slightly cooled since the peak of the Holocene. A similar observation is clearly visible, over the whole 10,000 years, in the Greenland temperature record published by Johnsen et al. (1995; see their Figure 2).

Finally, integration of the model in the future for different natural CO_2 scenarios shows that the atmospheric CO_2 concentration must decrease below 250 ppmv for the ice sheets to grow in the Northern Hemisphere over the next 50 kyr (Berger et al., 1996).

4. The Past, Analogue for the Future?

For the future 130 kyr, analysis of the insolation shows that the present interglacial might last particularly long (50 kyr). The small insolation variation over this period is indeed quite exceptional, more or less similar situations occuring only 5 times over the last 3 million years (Berger and Loutre, 1996). On the point of view of insolation only, the Eemian can therefore hardly be taken as an analogue of the next thousands of years, as it is often assumed. Moreover, forced by insolation and a natural CO_2 scenario based upon the Jouzel CO_2 reconstruction for the past, the LLN 2-D model simulates a Northern Hemisphere ice volume which remains almost the same as today up to 50 kyr AP (Berger et al., 1996). The next glacial maximum is reached at 100 kyr AP and is followed by a partial, but significant, melting of the ice sheets which peaks at 120 kyr AP. Sensitivity experiments and other scenarios, including global warming scenarios resulting from man's activities, have demonstrated that the non-reappearance of the ice sheets before 50 kyr AP is a robust feature of the LLN model. This kind of results confirms how much it will be difficult to find a satisfactory geologic analogue for the future and even more for a future greenhouse warming, as already claimed by Mitchell (1990) and Crowley (1990).

However, the geological record can provide valuable information about certain processes operating in the climate system (Crowley, 1993a). A very good first-order agreement between high net CO_2 and warm climates has been found for most of the last 540 million years; agreement between models and observations also arises when comparing estimated CO_2 variations during this Phanerozoic (Crowley, 1993b). But the use of the geologic record to calibrate model sensitivity for increase in CO_2 is somewhat more contentious. Uncertainty in the data and in the processes involved to explain past climate changes suggests that paleoclimate data do not constrain the sensitivity to doubled CO_2 much more than the present IPCC range in uncertainty.

Although the seasonal and non synchronous nature of the Pleistocene warm periods seems to be mostly accounted for by Milankovitch variations, related feedbacks and changes in the hydrosphere and cryosphere, this result does not question the paradigm of climate sensitivity defined as the ratio of global mean radiative forcing to global mean temperature change, the cornerstone of the IPCC projections of future climate change from humanity's greenhouse gases and aerosols (Houghton et al., 1996).

Some critics of anthropogenic global warming have indeed addressed this question on grounds that the annual and global mean radiative forcing from the astronomical theory during the ice ages was near zero and yet the climate clearly changed. Actually the two problems are conceptually totally different. On the one hand, the radiative forcing on climate related to the anthropogenic enhancement of the greenhouse effect involves primarily the longwave terrestrial radiation of the Earth, absorbed and then re-emitted by a number of trace gases in the atmosphere above. It is a global phenomenon where the latitudinal and seasonal components of the energy balance do not play a major role as in the Milankovitch theory. On the other hand, the glacial-interglacial cycles are brought about by the long-term variations of the latitudinal and seasonal distributions of solar radiation which is absorbed by the Earth. Then, slow feedbacks of the system cause the buildup of ice sheets and the decrease of CO_2 during glacial cycle initiations, and the reverse during terminations. It means that the climate system is consequently forced by global mean albedo and greenhouse gas changes, the global mean atmospheric surface temperature responding with the appropriate climate sensitivity. In other words, it is not the global mean insolation forcing that causes the ice ages, but changes in the insolation distribution over the annual cycle and with latitude that change planetary albedo and greenhouse gas concentration, making the ice ages consistent with the IPCC paradigm of global mean forcing and climate sensitivity.

5. The LLN Simulated and SPECMAP Reconstructed Ice Volumes

The stacked, smoothed oxygen-isotope record of SPECMAP (Imbrie et al., 1984; Martinson et al., 1987; Figure 5) provides a unique record of isotopic variations over the last 800 kyr which can be compared to the simulated continental ice volume changes. However, our simulations are only providing variations of the Northern Hemisphere continental ice volume. In the comparisons, we must therefore keep in mind that we do not have any reconstruction of the Southern Hemisphere ice volume (mainly Antarctica) and that a large number of hypotheses subtend the model.

Despite of these deficiencies, the overall timing of ice volume changes over the last 200 kyr is quite well reproduced, but discrepancies in the magnitude of the ice volume can be observed. The largest discrepancy is probably the too large ice melting simulated by the model around 170 kyr BP, although the ice volume maximum at 182 kyr BP seems to be well captured by the model (Figure 4). The large values of insolation around 175 kyr BP induce this melting of the Northern Hemisphere ice sheets. Either it is an important deficiency of the model or we have to look for a significant change in the Southern Hemisphere continental ice at that time. At the end of stage 6, the simulated glacial maximum occurs at 135 kyr BP while the δ^{18}O ice volume maximum occurs at 151 kyr, but we must accept that the δ^{18}O ice volume remains large from 156 to 133 kyr BP. The model simulates very well the transitions between isotopic stages 6 and 5 and between the isotopic substages 5e (Eemian interglacial) and 5d. This is due mainly to the insolation changes (these features are also well reproduced under a constant CO_2 forcing – Berger et al., 1996), but it is reinforced by important changes in the CO_2 concentration in the present experiment. Although the timing of what may correspond to isotopic substages 5c, 5b and 5a is quite well reproduced (a coincidence which is certainly reinforced by the astronomical tuning of the SPECMAP curve), the amount of ice

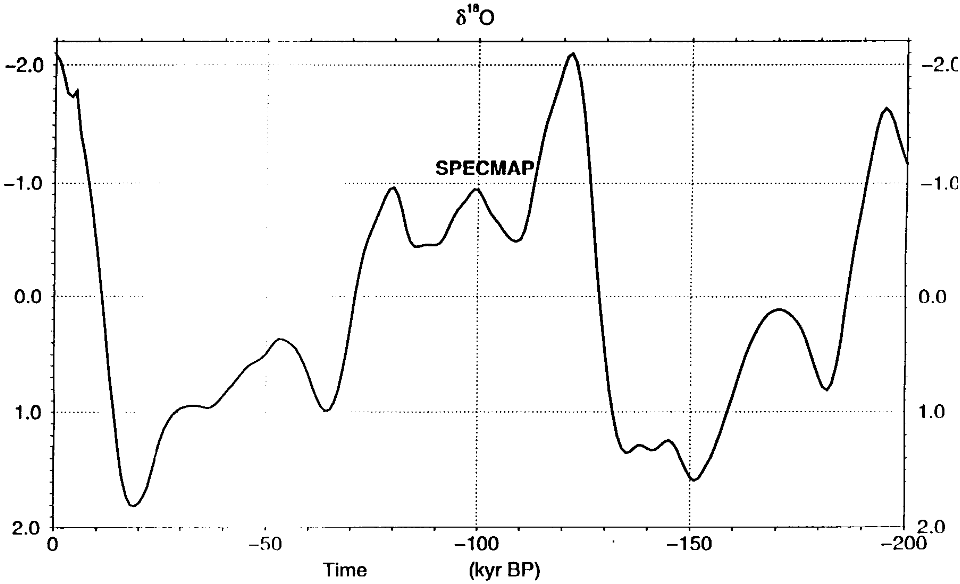

Figure 5. The stacked, smoothed oxygen-isotope SPECMAP record over the last 200 kyr (Imbrie et al., 1984; Martinson et al., 1987).

is here again much less in the simulation than in SPECMAP, this discrepancy being more or less similar to the one occuring at isotopic substage 6.5, 170 kyr ago. Moreover, the δ^{18}O ice volume minimum around 90 kyr BP lasted for about 10 kyr, contrarily to the simulations, and the simulated ice volume at stages 5a and 4 is lagging behind the δ^{18}O curve by 6 kyr. The delay to induce isotopic stage 4 might be explained, at least partly, by the fact the LLN model does not take into account stratospheric dust. In particular, the Toba volcanic explosion occured around 73.5 kyr BP (Rampino and Self, 1992), a time which coincides also with large insolation and CO_2 decreases. Therefore this huge eruption might have accelerated the rapid ice accumulation in the Northern Hemisphere, a possibility which is now under investigation with the LLN model (Halenka et al., 1996).

Finally, stage 3 is well reproduced although, as for the 170 kyr BP interstadial, there is not enough ice according to SPECMAP. The simulated Last Glacial Maximum is lagging behind SPECMAP by 4 kyr, a lag which is related to the late CO_2 minimum in the Vostok curve.

As far as the ice volume maxima are concerned, SPECMAP (Figure 5) shows two extremes located at the Last Glacial Maximum ($\sim$20 kyr BP) and at stage 6 ($\sim$151 kyr BP), with the LGM being slightly the largest. There are also two secondary maxima, roughly of the same size: isotopic substage 6.6 (183.3 kyr BP) and stage 4 (60.4 kyr BP). The simulation shown in Figure 4 mimicks pretty well these maxima, except for the late arrival of stage 6 and an ice volume larger at stage 6.6 than at stage 4.

Being given all the hypotheses used in the modelling experiments, the results may be considered as surprisingly good. Using the reconstructed CO_2 concentrations (Figure 4) improves the constant CO_2 simulations (reproduced in Figure 4 of Berger et al., 1996) if SPECMAP (Figure 5) is used as a standard for comparison. For example, the amplitude of the last and penultimate glacial maxima better fits SPECMAP, but it is mainly the southward extent of the individual Northern Hemisphere ice sheets and the air temperature which are better represented when using the reconstructed CO_2 instead of a constant CO_2 concentration (Gallée et al., 1992).

The relative success of these modelling experiments in reproducing the last glacial-interglacial cycles allows to question whether we may identify the most important processes and parts of the Earth system to be included to quantitatively model the ice ages. In order to answer this very important but difficult question models involving different processes will have to perform more transient simulations for reconstructing the last hundreds of thousands of years. Although a large amount of climatic processes and sub-systems are already represented in this model, some are only very roughly parameterized and others – which might play a role – are still missing. In particular, an important limitation is that the model covers the Northern Hemisphere only. As, seasonally speaking, the Milankovitch forcing is often out of phase between the two hemispheres, it remains fundamental to question how, in the real ice ages, the responses of the two are synchronized. A tentative answer will only be possible when the model will be extended to the whole Earth (Dutrieux, 1997). In the meantime, we may guess that the Southern Hemisphere responses first following the early change of insolation in this hemisphere. As this hemisphere is mainly oceanic, the response would be damped, quite slow and delayed well after insolation changes become sensible in the Northern Hemisphere. The large continental fraction of this hemisphere would be responsible for a large response there which would rapidly drive the climate change over the whole Earth, including the Southern Hemisphere through the ocean circulation and changes in the composition of the Earth’s atmosphere (Manabe and Broccoli, 1985; Broccoli and Manabe, 1985).

It is nevertheless intriguing that this model with all its hypotheses and limitations reproduces quite well the low frequency part of the Northern Hemisphere climatic variations reconstructed from geological data over the last few hundreds of thousands of years and, in particular, simulates steadily a recurrent 100-kyr cycle of variable length (Gallée et al., 1993; Berger et al., 1996).

6. Conclusions

A series of experiments have been made with the LLN 2-D Northern Hemisphere climate model to test its sensitivity to both the astronomical and the CO_2 forcings. It must be stressed that this model is based upon a series of hypotheses and contain many parameterizations. Clouds and the hydrological cycle are oversimplified and the heat transport by the middle and deep oceans are parameterized. Despite these weaknesses, the model succeeds to simulate the long-term variations of the Northern Hemisphere ice sheets with an acceptable confidence.

Being the response of a simple model, all these results have to be confirmed by other more sophisticated models. This will be done, at least with an improved and extended version of the LLN model which will consider, in particular, both hemispheres, three oceanic basins and a more realistic hydrological cycle. Dust originating from natural sources at the geological time scale must also be included both in the troposphere and in the stratosphere. Peltier and Marshall (1995) and Li et al. (1996) have indeed shown the possible influence of terrigenous dust on climate during the ice ages; but huge amounts of volcanic dust might also play an important role in accelerating the ice-sheet growth as, for example, between isotopic stages 5 and 4 (Halenka et al., 1996).

Finally, a model of the terrestrial and oceanic components of the carbon cycle has to be coupled interactively to the LLN 2-D model. The present version of the LLN model is indeed forced by both insolation and observed CO_2 atmospheric concentration. But actually, atmospheric CO_2 variations are also apparently paced by orbital insolation variations (Yiou et al., 1991), through feedback links that may involve ocean biology, circulation and alkalinity pumps. Modelling past and future CO_2 variations (as it has been tentatively done by Saltzman and Verbitsky, 1994) will therefore request future research to solve the theoretical problem of why atmospheric CO_2 changes the way it does during glacial-interglacial transitions.

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