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Hemispheric roles of climate forcings during glacial-interglacial transitions as deduced from the Vostok record and LLN-2D model experiments

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Abstract. The Vostok ice contains fingerprints of atmospheric greenhouse trace gases, Antarctic temperature, Northern Hemisphere temperature, and global ice volume/sea level changes during the last glacial-interglacial cycles and thus allows us to investigate the sequences of these climatic events, in particular during the transitions from full glacial to interglacial conditions. The use of the updated CO_2 record presented here and a reexamination of the sea level proxy confirm that the succession of changes has been similar through each of the four marked transitions found at Vostok. Antarctic air temperature and CO_2 increase in parallel and almost synchronously, while the rapid warmings over Greenland take place during the last half of their change and coincide with the marked decay in continental ice volume. The Vostok results thus emphasize a fundamental difference between South and North in terms of climate dynamics. Our results confirm the role of CO_2 as an important amplifier of the glacial-interglacial warming in the South. It appears also that the marked warming observed at high northern latitudes (lagging behind the CO_2 increase by several thousand years) is roughly synchronous with the decay of the northern ice sheets, suggesting a major role of climatic feedback due to this decay. Such a climatic scenario is supported by sensitivity experiments performed with the LLN 2-D model forced by the Northern Hemisphere insolation and CO_2. Model results indicate that the decay of the northern ice sheets and the Northern Hemisphere temperature depend primarily on the northern summer insolation. These results, nevertheless, could be affected if mechanisms specific to the Southern Hemisphere appear to play a major role in driving the Northern Hemisphere climate. The model also helps to constrain the time response of ice volume to insolation and CO_2 changes.

1. Introduction

Changes in insolation and in composition and circulation of the atmosphere, as well as continental and oceanic modifications, such as ice and vegetation distribution, ocean dynamics, are the basic forcings changing the climate of our planet. Such mechanisms are partially or totally involved in the climate variability observed at different timescales, from the Earth's orbital frequencies, modulating the latitudinal distribution of the insolation, to changes occurring at much higher frequencies. If the different forcings are generally reasonably identified, it is much more difficult to assess the role played by each one in the temporal variability of the past and present climate. In this respect the glacial-interglacial transitions appear privileged, due to their global nature and exceptional amplitudes.

The Vostok ice record offers a unique opportunity to investigate the sequence of changes during the last four glacial-interglacial transitions (named hereinafter T1 to T4, from the most recent to the oldest one) [*Petit et al.*, 1999]. The ice archives different properties, in particular the deuterium content of the ice (δD_{ice}), a proxy of the Vostok temperature. The entrapped air, encloses information on CO_2, CH_4, and ^{18}O of O_2 ($\delta^{18}O_{air}$) which can be considered as proxies of atmospheric greenhouse trace gases, temperature at high northern latitudes, and sea level/ice volume, respectively. We can thus investigate phase relationships between climate forcings, i.e., greenhouse gases and ice volume, and responses, i.e., temperatures of both hemispheres.

CO_2 and CH_4 are two major atmospheric greenhouse trace gases which are considered to have significantly contributed to the glacial-interglacial global warmings [*Lorius et al.*, 1990; *Berger et al*, 1998; *Weavers et al.*, 1998]. Past changes of CO_2 are essentially driven by the ocean, while changes in wetlands through the hydrological cycle are thought to be the main source of the modifications in atmospheric CH_4 before the anthropogenic impact.

Furthermore, a similar CH_4 event found in each of the four transitions can be used as a time marker of the glacial-interglacial warming in the North. Namely, one of the striking feature of the methane record is the marked jump observed in its increase during the second half of the four glacial-interglacial transitions (Figure 1). *Petit et al.* [1999] note that in the case of the last transition (T1) the CH_4 jump, which is a global signal (i.e., which occurs simultaneously

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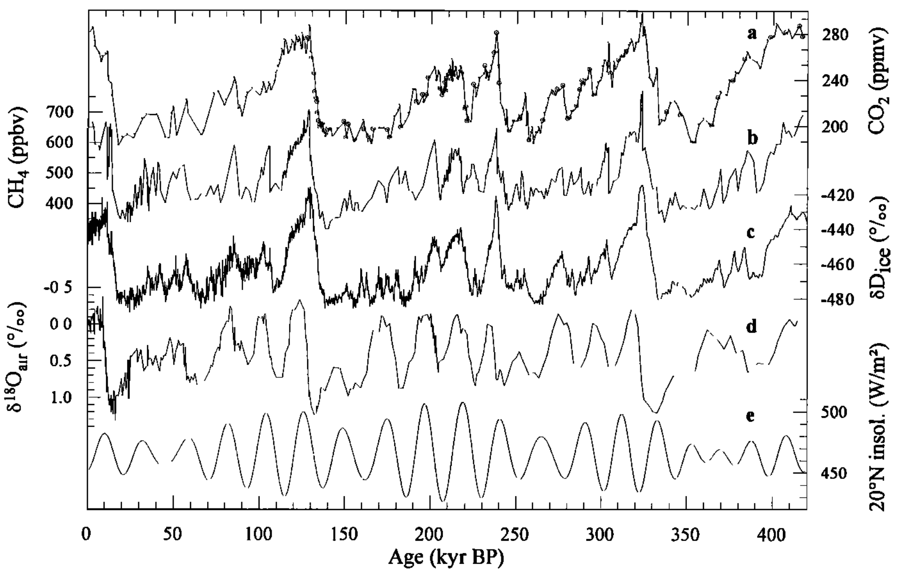

Figure 1. Records from the Vostok ice cores used in this paper (glaciological timescale [cf. Petit et al., 1999]): (a) CO_2 (ppmv), the record published by Petit et al. [1999] has been updated with the new measurements (open circles, this work); (b) CH_4 (ppbv) [Petit et al., 1999]; (c) isotopic temperature (δD_{ice} in ‰) [Petit et al., 1999]; (d) content in ^{18}O isotope of atmospheric O_2 [Petit et al., 1999] expressed in per mil unit ($\delta^{18}O_{air}$= $1000*[(^{18}O/^{16}O)_{sample}/(^{18}O/^{16}O)_{standard} - 1]$). The standard is the modern air composition; (e) summer insolation (JJA) at low latitude (20°N) (astronomical timescale).

over Greenland and Antarctica), is synchronous with the rapid and marked Northern Hemisphere warming (Bølling/Allerød), as recorded in the Greenland ice records (GRIP and GISP2). They speculate that the same feature is true for the three other transitions and propose to use these CH_4 jumps recorded in the Vostok ice as time markers of the prominent glacial-interglacial warmings in high northern latitudes.

The $\delta^{18}O_{air}$ tracks the changes in the ^{18}O content of seawater ($\delta^{18}O_{sw}$) [*Bender et al.*, 1985; *Sowers et al.*, 1991] and has been used, by *Broecker and Henderson* [1998] for the last two transitions and *Petit et al.* [1999] for the four transitions, as an indicator of the continental ice volume decay associated with the deglaciations.

Since the publication by *Petit et al.* [1999], we have increased the resolution of the CO_2 record. We also revisit here the use of variations in $\delta^{18}O_{air}$ at Vostok as a proxy of ice volume changes. We then discuss the sequence of events during the transitions in terms of climate dynamics and forcings both in the Northern and Southern Hemispheres. Finally, we investigate the response of the northern climate (temperature, ice volume) to changes in the latitudinal distribution of the insolation and CO_2, using the LLN 2-D (Northern Hemisphere) climate model.

2. Updating the Vostok CO_2 Record

The Vostok data used in the present paper are basically those published by *Petit et al.* [1999]. Nevertheless, for CO_2 a set of 75 new depth levels has been measured, using the Grenoble analytical setup described by *Sowers et al.* [1997] in the same way as by *Petit et al.* [1999]. The scattering of the results obtained on 20 replicate levels is about 1.6 ppmv as can be expected from the analytical uncertainty evaluated to be 3 ppmv [*Sowers et al.*, 1997]. These additional measurements are mainly in the 100-400 kyr B.P. interval (see Figure 1). They improve the time resolution of the profile and confirm it, especially during T2. The record is now based on 379 measurements (corresponding to 359 depth levels) and the mean temporal resolution for T1 to T4 is 1800, 600, 750, 1050 years, respectively. The temporal resolution of CO_2 measurements through T1 is still significantly lower, due to the difficulty to find suitable uncracked samples in the brittle zone of the Vostok core.

3. Revisiting the Use of $\delta^{18}O_{air}$ As an Ice Volume Proxy

Sea level changes during deglaciations are mainly due to waxing and waning of Northern Hemisphere ice sheets. Direct evidences of sea level change, as dated coral terraces [*Bard et al.*, 1990; *Fleming et al.*, 1998], are available for the last deglaciation and previous sea level high stands. Sea level changes can also be deduced from the oceanic records of $\delta^{18}O$ in benthic forams [*Imbrie et al.*, 1984; *Bassinot et al.*, 1994]. However, the problem of the comparison of these records with Vostok arises as they are expressed in different chronologies. To avoid these dating problems, it is possible to find a proxy of the sea level change in the set of Vostok records. The $\delta^{18}O_{air}$ has been considered as such a proxy and, as mentioned in section 1, has thus been used in the case of the four Vostok terminations by *Petit et al.* [1999]. However, the atmospheric ^{18}O is fractionated relatively to the oceanic ^{18}O during the photosynthesis-respiration processes and through the continental hydrological cycle (the Dole effect) [*Bender et al.*, 1994; *Mélières et al.*, 1997; *Malaizé et al.*, 1999], and the respective influences of sea level and hydrological cycle are still under debate. The comparison of the amplitude of glacial to interglacial changes of $\delta^{18}O_{air}$ and

$\delta^{18}O_{sw}$ (about 1‰ in the two cases) tends to attribute the major role to the ice volume during deglaciations [*Sowers et al*, 1991; *Sowers et al.*, 1993]. On the other hand, *Mélières et al.* [1997] have shown for marine isotopic stage 6.5 (about 175 kyr B.P.) that $\delta^{18}O_{air}$ exhibits a large variation, whereas $\delta^{18}O_{sw}$ stay relatively constant. This points out, in this case, the role of the hydrological cycle of continental waters on the variations of the $\delta^{18}O_{air}$, and the authors conclude that this influence may be major in the whole record.

We propose here to check how robust is $\delta^{18}O_{air}$ as an indicator of the timing of ice volume/sea level changes by decomposing the $\delta^{18}O_{air}$ signal into its two major components:

$$\delta^{18}O_{air} = \delta_{hydro} + \delta_{sl},$$

where δ_{hydro} represents the influence of the hydrological cycle on the $\delta^{18}O_{air}$ variations and δ_{sl} that of continental ice volume, i.e., sea level.

Following *Bender et al.* [1994] we assume that insolation at low latitude is the dominant factor governing the δ_{hydro}. We use two different methods to extract δ_{hydro} from the basic signal. First, taking the advantage of the obvious similarities between $\delta^{18}O_{air}$ and insolation (see Fig 1), we synchronize the Vostok and the insolation records by lining up the $\delta^{18}O_{air}$ extrema on the maxima and minima of JJA insolation at low latitude (20°N). Between these fixed points $\delta^{18}O_{air}$ ages are linearly interpolated. This imposes an artificial shift of the GT4 chronology of a mean value of about 7 kyr toward older ages. However, this shift is not constant and it generally increases with age. This points out either to a drift (Age_{insol}=1.03*Age_{GT4}- 1700), suggesting an overestimation of the accumulation rates of about 3% in the GT4 chronology, either to a discontinuity in the chronology around the stage 7.5 (240 kyr B.P.) as suggested by *Parrenin et al.* [this issue]. In the first case, the deviation from the drift is ±7 kyr, while in the second case, the mean shift is about 4 ±5 kyr until 240 kyr B.P. and 13.5 ±4 kyr before. The shift imposed by the synchronization procedure could either be due to a lag in the response of $\delta^{18}O_{air}$ to the insolation or to the choice of the summer insolation or, finally, to the glaciological timescale GT4. At this stage we do not want to discuss the validity of the GT4 chronology or the phase between $\delta^{18}O_{air}$ and insolation, the synchronization is only used to estimate the maximum of $\delta^{18}O_{air}$ variance explained by the insolation variations. Thus the linear regression between $\delta^{18}O_{air}$ and insolation is calculated and used to extract $\delta_{hydro}{}^{I}$ ($\delta_{hydro}{}^{I}$ =a*Insol + b). The correlation coefficient of this regression is r= 0.68. This means that 46% (r^2) of $\delta^{18}O_{air}$ variations can be statistically explained by changes in the low-latitude summer insolation.

The second way to extract δ_{hydro} (index P) is to filter the $\delta^{18}O_{air}$ record, after synchronization with the precession signal, with a band pass around the precession frequencies (between 18 and 24 kyr).

These approaches are probably simplistic because they are based on statistics and not on real physical processes; however, they probably provide upper limits for δ_{hydro}.

Removing each δ_{hydro} from the $\delta^{18}O_{air}$ signal, we then obtain $\delta_{sl}{}^{I}$ and $\delta_{sl}{}^{P}$. Assuming that the precession signal, which governs the 20°N insolation, is not dominant in the sea level, we associate these residuals to the sea level component of the $\delta^{18}O$. The two δ_{sl} records are shown over the four cycles on Figure 2, together with $\delta^{18}O_{air}$ and a sea level curve reconstructed by taking the deep dea sediment $\delta^{18}O$ record of *Bassinot et al.* [1994] as sea level signal. Our two δ_{sl} records are very similar (r^2=0.83) and show variations much closer to the sea level reconstruction than $\delta^{18}O_{air}$ does. We note that amplitudes of δ_{sl} variations are generally smaller than those of $\delta^{18}O_{air}$. The comparison between the sea level and the δ_{sl} curves is most meaningful over the last 150 kyr, during which the cross-dating problems are less critical. For this period the correlation coefficients (r^2) between the sea level reconstruction and the δ_{sl} signals are 0.56 (for $\delta_{sl}{}^{I}$) and 0.62 (for $\delta_{sl}{}^{P}$), although there are probably some differences between Vostok and marine chronologies. We should also keep in mind that $\delta^{18}O_{air}$ (and, consequently, δ_{sl})

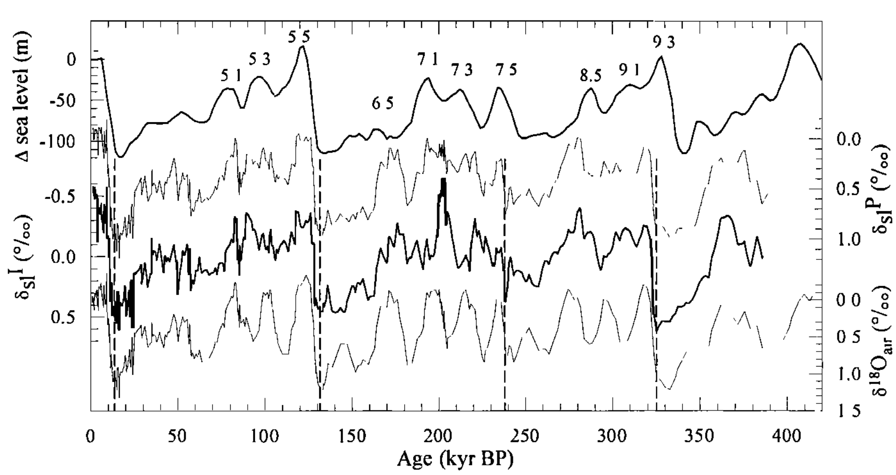

Figure 2. (top-bottom) Reconstructed sea level based on the $\delta^{18}O$ record from the oceanic core MD900963 [Bassinot et al., 1994], $\delta_{sl}{}^{P}$, and $\delta_{sl}{}^{I}$ signals (estimated contributions of sea level change to $\delta^{18}O_{air}$; see text), $\delta^{18}O_{air}$. The $\delta^{18}O$ signal of Bassiont et al. has been translated in term of sea level, by taking a sea leval 120 m lower than today during the last glacial maximum. Each profile is plotted against its own chronology: GT4 for δ_{sl} and $\delta^{18}O_{air}$ and a chronology derived from SPECMAP for the sea level changes. Vertical lines point out the major changes of δ_{sl} corresponding to the four glacial-interglacial transitions.

Table 1. Vostok Sequence of Climatic Events [a]

Transition 1	Transition 2	Transition 3	Transition 4
↗ CH_4: 18,900 ±1 200 years B.P.	↗ CH_4: 138 400 ±1 200 years B.P.	↗ CH_4: 245,500 ±800 years B.P.	↗ CH_4: 335,400 ±1700 years B.P.
↗ CO_2: 1200 ±3000 years	↗ CO_2: 200 ±700 years	↗ CO_2: 600 ±600 years	↗ δD: 1400 ±500 years
↗ δD: 1900 ±100 years	↗ δD: 400 ±100 years	↗ δD: 1500 ±200 years	↗ CO_2: 1800 ±1600 years
↗ CH_4 (second phase): 4700 ±900 years	↘ $\delta sl^{I,P}$: 6900 ±2000 years	↘ $\delta^{18}O_{atm}$: 2500 ±1600 years	↘ $\delta^{18}O_{atm}$: 5100 ±4000 years
↘ $\delta sl^{I,P}$: 5300 ±500 years	↘ $\delta^{18}O_{atm}$: 6900 ±2000 years	↘ $\delta sl^{I,P}$: 6900 ±1600 years	↘ $\delta sl^{I,P}$: 10,200 ±1000 years
↘ $\delta^{18}O_{atm}$: 5300 ±500 years	↗ CH_4 (second phase): 10,100 ±200 years	↗ CH_4 (second phase): 6900 ±400 years	↗ CH_4 (second phase): 11,200 ±800 years

[a] The sequences are described for the last four glacial-interglacial transitions. The date of the first event (increase of CH_4 concentration) is given in the GT4 timescale. Time lags between the other events and this first one are indicated. The evaluation of uncertainties is discussed in the text.

tracks $\delta^{18}O_{sw}$ with a lag of about 2 kyr [*Sowers et al.*, 1993], which is the response time of the atmosphere to changes in $\delta^{18}O_{sw}$.

When focusing on the glacial-interglacial transitions, the choice of δ_{sl} signal ($\delta_{sl}{}^{I}$ or $\delta_{sl}{}^{P}$) does not induce significant modifications in the timing and succession of the climatic events (Table 1). Furthermore, in the case of T1 and T2, δ_{sl} is also in phase with $\delta^{18}O_{air}$, while $\delta^{18}O_{air}$ seems to start to decrease prior to the change in δ_{sl} at the beginning of T4, and to a less extent, at the beginning of T3.

In conclusion, the decomposition of the $\delta^{18}O_{air}$ signal allows us to obtain a better Vostok proxy for ice volume/sea level changes, which will be used in the next sections. Because the use of $\delta_{sl}{}^{I}$ or $\delta_{sl}{}^{P}$ leads to the same result for sequencing the transitions, we will use hereinafter a unique notation: δ_{sl}.

In the case of the first deglaciation we can also compare the δ_{sl} record with the sea level record obtained by Bard and others [*Bard et al.*, 1990] on dated Barbados coral terraces. For this comparison we follow the chronology published by *Blunier et al.* [1998], which is based on the synchronization of the CH_4 antarctic ice record with the GRIP record. On this chronology the timing of the variations of $\delta^{18}O_{air}$ and δ_{sl} are not significantly different from the changes indicated by the Barbados data, and the major rise of the sea level linked with the MeltWater Pulse MWP-1A (about 14 kyr B.P.) corresponds to the beginning of the marked decrease of both δ_{sl} and $\delta^{18}O_{air}$.

4. Sequence of Climatic Events for the Last Four Glacial Terminations Recorded in the Vostok Ice

In this section we reexamine the sequence of events discussed previously by *Petit et al.* [1999] for the four transitions, taking into account the updated CO_2 record and δ_{sl} as an indicator of ice volume/sea level. The Vostok records (plotted on GT4 chronology) of δD_{ice}, CO_2, CH_4, and δ_{sl} are shown in Figure 3 for each transition. The δD_{ice} is measured on the ice itself, while the three other properties are analyzed on the entrapped air. As the air bubbles do not close at the surface of the ice sheet but at the firn-ice transition (at about 100 m below surface at Vostok), the extracted air is younger than the surrounding ice. The corresponding difference in age (Δage) is evaluated using a firn densification model [*Barnola et al.*, 1991]. The timing of the events is detailed in Table 1.

The uncertainties shown in Table 1 take into account the time resolution of each of the records and the inaccuracy in determining the start of the transitions. When comparing δD_{ice} with the other properties, we have to add to the values given in Table 1 an uncertainty of about ±600 years which

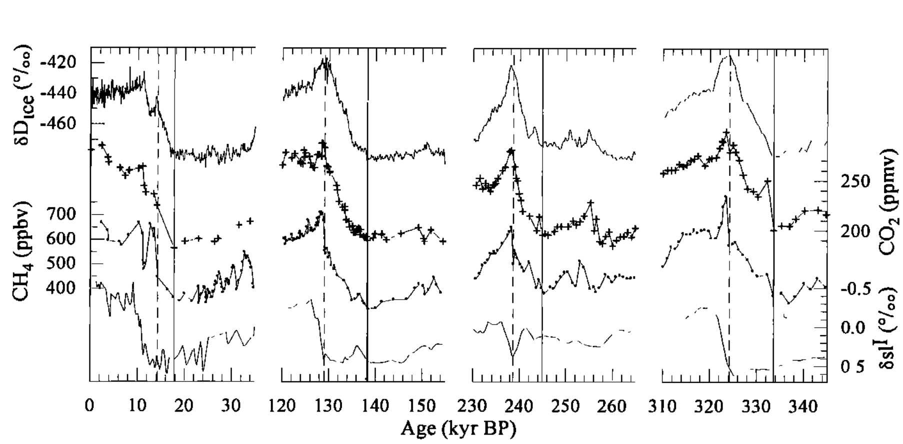

Figure 3. Comparison of Vostok records (GT4 chronology) during the last four glacial to interglacial transitions (T1 to T4). (top-bottom) Antarctic temperature (δD_{ice}), CO_2 content of the atmosphere, high northern latitude temperature (second phase of the CH_4 rise), and sea level changes ($\delta_{sl}{}^{I}$). Solid vertical lines correspond to the starts of the CO_2 increases and dashed lines to the warmings of the Northern Hemisphere.

corresponds to the uncertainty on the Δage in glacial conditions.

Our new analysis of the Vostok $\delta^{18}O_{air}$ signal as a proxy for ice volume changes confirms that through each of the last four glacial-interglacial transitions, at least half of the southern warming and of the CO_2 increase are achieved before the intense deglaciation of the northern ice sheets.

Another common feature of the four transitions is the difference in shape between the CO_2 and CH_4 increases, even if the atmospheric concentration of these two species start to rise synchronously. The CO_2 increase is almost in phase and parallel with the δD_{ice}, while the CH_4 concentration shows a marked enhancement in its growing rate during the second half of the transitions. Following *Petit et al.* [1999], we assume that these CH_4 jumps are synchronous with rapid glacial-interglacial warmings taking place in the Northern Hemisphere, as it is the case for T1. This suggests that the dynamics of glacial-interglacial warmings, and consequently the major temperature forcings, are systematically different (at least for the last four transitions) between North and South.

Thus the sequence of each of the four transitions at Vostok suggests a scenario where the greenhouse trace gases play a major role in contributing to the early warming in the South, while some mechanism delayed the northern warming. If the assumption that the CH_4 jumps reflect the rapid warming in the North is correct, this warming appears to roughly coincide, in each of the transitions, with a rapid stage of deglaciation in the North. Because of the resolution of the Vostok record and mainly of the response time of the $\delta^{18}O_{air}$ to the $\delta^{18}O_{sw}$ changes, the potential phasing between the warming in the North and the rapid stage of ice retreat would be difficult to assess.

5. Role of Greenhouse Trace Gases and Insolation

Thus, what we found in investigating the sequences through the Vostok transitions is a clear and systematic difference in the warming dynamics between North and South reflecting a different response to climatic forcings. The record also strongly suggests a dominant role of the greenhouse trace gases, especially CO_2, in contributing to the southern warming, but the time resolution is too low to estimate with confidence if the warmings at Vostok were initiated before or after the beginning of the CO_2 increases ±1 kyr (Table 1). Nevertheless, a CO_2 record has been recently obtained from the new EPICA core drilled at Dome C (central part of East Antarctica) with a remarkable high resolution [*Monnin et al.*, 2001], which indicates that most probably, the Antarctic temperature started to rise at the beginning of the last transition a few hundred years before CO_2. This recent record confirms previous results by *Neftel et al* [1988] and *Fischer et al.* [1999], suggesting a similar phasing between Antarctic temperature and CO_2. CH_4 starts also to increase in phase with CO_2 and the southern temperature, which suggests that tropical latitudes have been affected by a contemporaneous warming.

It is well documented and accepted that the climate system at the scale of the glacial-interglacial cycles responds to the orbital forcing. The direct comparison of insolation curves with the Vostok records is difficult because of the poor absolute dating of Vostok ice. Moreover, the climate system is most probably responding (maybe in a complex way) to the insolation forcing at all latitudes and time in the year. Therefore the choice of one target insolation curve for the comparison is critical. It is even more difficult because there is a phase lag of roughly 2 kyr between insolation curves for two consecutive months at given latitude (as far as the insolation signal is dominated by precession); so we will present next sensitivity experiences made with the LLN 2-D climate model in order to investigate the response of the climate system to the orbital and CO_2 forcings.

The LLN 2–D NH climate model [*Gallée et al.*, 1991] is a model of intermediate complexity, which was designed in order to understand the response of the climate system to the astronomical forcing. It links the atmosphere, the upper mixed layer of the ocean, the sea ice, the continents, the ice sheets, and their underlying lithosphere. The model used has clear limitations in that it considers only the Northern Hemisphere (there is no Southern Hemisphere), and it has no explicit representation of the thermohaline circulation. On the other hand, the LLN model is one of the few models able to simulate the climatic changes over hundreds of thousand years, and it can provide important information when performing sensitivity studies of climate over the last glacial-interglacial cycles. It has, indeed, shown its ability to simulate the global long-term climatic variability in good agreement with the paleorecords and, in particular, the changes in the Northern Hemisphere continental ice volume under both the insolation and the CO_2 forcings [*Berger et al*, 1998; *Li et al.*, 1998]. A major weakness of the model still lies in the very frequent melting of the Northern Hemisphere ice sheets, including the Greenland ice sheet, during the interglacials. However, this does not prevent the model from simulating correctly the glacial-interglacial cycles. From experiments over the last two climatic cycles it was concluded that the variations in the Earth's orbit and related insolation are driving the climatic changes during ice ages, and the CO_2 has an influence for shaping the 100 kyr cycle and for improving the simulated surface air temperature.

To date, special attention has been paid to the simulation of the last 200 kyr [*Berger et al.*, 1998]. Here the LLN-2D climate model is used to simulate the climatic changes over the Northern Hemisphere during the last 400 kyr. The exercise is of special interest since the insolation around 400 kyr, for instance, is very different from insolation around the last interglacial. The model is forced by computed insolation and CO_2 as reconstructed in this paper from the Vostok ice core. It must be kept in mind that two different timescales are thus mixed up in this simulation; that is, the insolation is computed with the astronomical timescale, while GT4 is used for the CO_2 concentration. Moreover, the time series for CO_2 concentration was interpolated with a time step of 1 kyr to match the time step of the slow part of the model. The response in terms of Northern Hemisphere mean annual temperature and continental ice volume is shown in Figure 4. The 60°N insolation at June solstice (further called insolation) is taken as a guideline although the model is forced by the seasonal and latitudinal distribution of insolation over the Northern Hemisphere. To free ourselves of choosing initial conditions (mainly ice sheet) at about 420 kyr B.P., the simulations were started at 575 kyr, which corresponds to MIS 15 (interglacial conditions). It was assumed that there were no Northern Hemisphere ice sheets

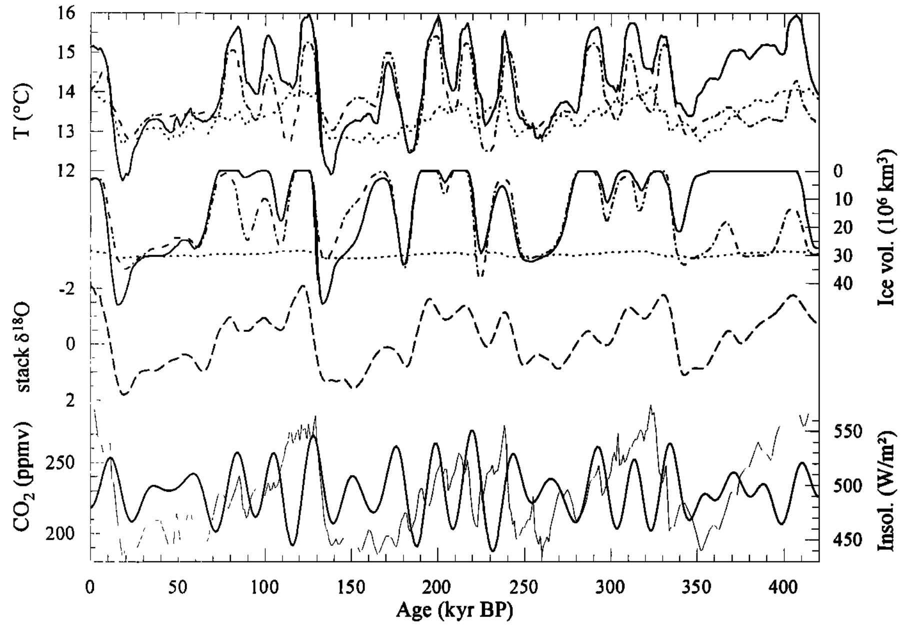

Figure 4. Variations over the last four glacial-interglacial cycles of the simulated Northern Hemisphere annual mean temperature (top) and continental ice volume (middle). The reference simulation (solid line) is forced by both the computed insolation and the reconstructed atmospheric CO_2 concentration. The results of two sensitivity experiments are also displayed for which either the CO_2 concentration is kept constant to 210 ppmv (dashed-dotted line) or the insolation distribution is kept to its present-day value (dotted line) over the four cycles. The stacked, smoothed oxygen-isotope record as a function of age in the SPECMAP timescale [Imbrie et al., 1984] is also plotted (dashed line; normalised scale on the left). This record is the average value of $\delta^{18}O$ variations in five deep-sea cores normalised to zero mean and unit standard deviation. The bottom panel displays the mid month insolation at 60°N in June (thick line) in June and the CO_2 concentration (thin solid line) used as forcings for the model.

at that time. Before 420 kyr, CO_2 concentrations based on regression between the Vostok CO_2 data and SPECMAP oxygen values [*Li et al.*, 1998] were used.

The climatic cycles simulated over the last 400 kyr match qualitatively very well with the SPECMAP record. However, the same discrepancy as in the last two glacial-interglacial cycles, i.e., the too frequent melting of the ice sheets during the interglacial, also appears. Moreover, an unusual feature shows up between 400 and 350 kyr B.P. This time interval is characterized by a very long interglacial, which does not seem to be recorded in Vostok data. Sensitivity experiments allowed us to identify a possible cause of such behavior. The CO_2 reconstruction used in the reference simulation remains larger than 250 ppmv over a longer time around 400 kyr B.P. than during the following interglacials. This confirms the conclusion already drawn by *Loutre and Berger* [2000b] that the amplitude of the simulated ice volume depends strongly on CO_2 concentration when insolation variations are small. However, we note that a long period of high CO_2 concentations around 400 kyr B.P. could be partly linked with the uncertainty in the Vostok chronology [*Petit et al.,* 1999]. This will be tested later on. In the meantime, sensitivity experiments were performed with GT4 for the CO_2 concentration and the astronomical timescale for the insolation.

Three simulations (Figure 4) are compared over the four cycles (400 kyr B.P. to present) in order to assess the relative importance of the insolation and the CO_2 forcing on the Northern Hemisphere continental ice volume and on the annual hemispheric temperature. The reference simulation (described above) is showing the response of the climate system to both forcings. In a second experiment, the insolation variations are the same as in the reference simulation, but the CO_2 concentration is kept constant at glacial level (210 ppmv); note that if we keep the CO_2 concentration close to interglacial values, the model is unable to build large ice sheets. In the third simulation the insolation is kept constant to its present-day distribution, and the CO_2 is evolving according to the Vostok reconstruction.

The experience at constant CO_2 (Figure 4) indicates changes in simulated ice volume similar to those of the paleorecord and only slightly lower than in the reference. Moreover, the annual mean Northern Hemisphere temperature also exhibits a time-dependent pattern similar to the reference one although some difference can be noticed, in particular in the amplitude of the variations, which are smaller and seem to oscillate between stable bounds. This confirms the influence of CO_2 variations for improving the simulated surface air temperature [*Gallée et al.*, 1992].

On the other hand, CO_2 alone (i.e., fixed insolation distribution as in the work of *Loutre and Berger* [2000a]) is unable to generate any significant variation in continental ice volume at the glacial-interglacial timescale (Figure 4), although it is hard to say whether a more CO_2-sensitive model would react differently. Moreover, the Northern Hemisphere annual mean temperature has no large-amplitude variations anymore (lower than 1.5°C, while the reference and the experiment with constant CO_2 exhibit amplitudes of

~3°C), and the pattern of temperature changes simulated with fixed insolation distribution is much like the pattern of CO_2. It thus appears from these experiments that insolation is the main forcing able to drive the building and retreat of the northern ice sheets as well as the Northern Hemisphere temperature through the glacial-interglacial cycles. The reconstructed CO_2 concentration improves the simulated climate, mainly for the amplitude of the signal when the insolation variations are small [*Berger et al.*, 1998; *Loutre and Berger*, 2000b].

Two additional experiments were designed in order to test the importance of the chronology used for the CO_2 record. In these experiments the CO_2 concentration is made younger or older by 5 kyr. The simulated Northern Hemisphere continental ice volume over the four cycles is only very slightly affected by these changes in the CO_2 concentration. Its timing is hardly changed, and there are some small changes in its amplitude. The Northern Hemisphere annual mean temperature is undergoing more important changes, but the change in timing remains smaller for temperature than for CO_2. These experiments once again confirm the conclusions already drawn [*Gallée et al.*, 1992; *Berger et al.*, 1998]; that is, the insolation forcing is playing a major role to simulate the glacial-interglacial cycle, and the CO_2 concentration has a smaller influence, which can be felt mainly on the surface air temperature variations.

Finally, it would be important to investigate the climatic sequences during the transitions. However, as the uncertainty on the chronology GT4 used for the CO_2 reconstruction becomes large prior to 200 kyr, we will focus on general features derived from transitions 1 and 2. June insolation at 60°N is leading the Northern Hemisphere continental ice volume by a few thousand years. It also leads the Northern Hemisphere annual mean temperature, this signal having a tendency to be in phase with the atmospheric CO_2 concentration (there is only a small phase difference between CO_2 and insolation during transitions 1 and 2).

6. Conclusions

The Vostok ice core provides a unique climatic archive, containing proxy indicators of atmospheric temperature, greenhouse gases and ice volume, and covering the last four glacial-interglacial cycles. Having such a multiproxy record measured on the same core greatly reduces the uncertainties in relative dating and facilitates the study of the succession of climatic events and the investigation of leads and lags.

The use of the updated CO_2 record and a reexamination of the sea level Vostok proxy confirm that the succession of changes has been similar through each of the last four glacial-interglacial transitions. Antarctic air temperature and CO_2 increase in parallel and almost synchronously, while the rapid warming over Greenland takes place during the last half of these increases and coincides with the marked decay in continental ice volume. This is in agreement with the records of oceanic sediment cores showing that the summer sea surface temperature of the Southern Ocean leads the changes in Northern Hemisphere ice sheet volume during the major glacial-interglacial transitions [*Labeyrie et al.*, 1996, and references therein]. The record also suggests strongly a dominant role of the greenhouse trace gases, especially CO_2, in contributing to the early southern warming and that the northern warming was delayed through some mechanism.

On the other hand, we have performed new experiments over the four climatic cycles with the LLN-2D NH model, and we have tested its sensitivity to insolation and CO_2 changes in terms of ice sheet and temperature responses. From these experiments it appears that the building and retreat of the northern ice sheets, as well as the Northern Hemisphere mean annual temperature changes, depend primarily on the insolation forcing over the four last glacial-interglacial cycles. The climatic role of CO_2 in the Northern Hemisphere is to improve the amplitude of the simulated climate signal, especially when insolation variations are small. Unfortunately, it is difficult to infer from these experiments what the response time is of the system to the insolation and CO_2 forcings, although the results indicate that for the four transitions the model response is delayed compared to the June insolation at 60°N.

Of course, the LLN model, which has no Southern Hemisphere, cannot be used to investigate the possible effects of any North-South climatic connections. Nevertheless, by combining the LLN-model simulations and the Vostok sequence of climatic events during the transitions, we propose a scenario where CO_2 largely contributes to the southern warming (well supported by the synchronism and parallelism between δD_{ice} and CO_2), and the northern warming is delayed by a few thousand years and coincides with the ice sheet collapse, which is essentially caused by the northern summer insolation.

Our results do not allow us to state exactly the respective roles of the northern and southern summer insolation in initiating the change from glacial-interglacial mode. This is mainly due to the absence of a common chronology between the Vostok record and the orbital control on insolation, and also to the difficulty to choose the most relevant target insolation curves. Although the earlier temperature signal in the South, in phase with CO_2, may suggest a dominant role of the southern insolation, which would be in agreement with the conclusion reached by *Henderson and Slowey* [2000] for the penultimate deglaciation (T2), the more traditional view of the northern insolation forcing the deglaciations cannot be ruled out. We can advance the hypothesis that low latitudes could drive a quasi-simultaneous change in both hemispheres, through for instance the orbital control on the tropical climate [*Clement et al.*, 1999], and that the warming in the North was counterbalanced during several thousand years by some oceanic process in relation with a progressive melting of the northern ice. This scenario would be supported by the fact that methane, taken as a climatic indicator of tropical regions [*Chappellaz et al.*, 1990], also changed earlier during the transitions and that sea level started probably also to increase well before the large melt water pulse during T1 [*Bard et al.*, 1996].

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