Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica

J. R. Petit†, J. Jouzel†, D. Raynaud†, N. I. Barkov‡, J.-M. Barnola†, I. Basile†, M. Bender‡, J. Chappellaz†, M. Davis‡, G. Delaygue†, M. Delmotte†, V. M. Kotlyakov‡, M. Legrand†, V. Y. Lipenkov‡, C. Lorius†, L. Pépin†, C. Ritz†, E. Saltzman† & M. Steienard†

The recent completion of drilling at Vostok station in East Antarctica has allowed the extension of the ice record of atmospheric composition and climate to the past four glacial–interglacial cycles. The succession of changes through each climate cycle and termination was similar, and atmospheric and climate properties oscillated between stable bounds. Interglacial periods differed in temporal evolution and duration. Atmospheric concentrations of carbon dioxide and methane correlate well with Antarctic air-temperature throughout the record. Present-day atmospheric burdens of these two important greenhouse gases seem to have been unprecedented during the past 420,000 years.

The late Quaternary period (the past one million years) is punctuated by a series of large glacial–interglacial changes with cycles that last about 100,000 years (ref. 1). Glacial–interglacial climate changes are documented by complementary climate records of ice cores, largely derived from deep sea sediments, continental deposits of flora, fauna and loess, and ice cores. These studies have documented the wide range of climate variability on Earth. They have shown that much of the variability occurs with periodicities corresponding to the precession, obliquity and eccentricity of the Earth’s orbit. But understanding how the climate system responds to this initial orbital forcing is still an important issue in palaeoclimatology, in particular for the generally strong 100,000-year (100-kyr) cycle.

Ice cores give access to palaeoclimate series that includes local temperature and precipitation rate, moisture source conditions, wind strength and aerosol fluxes of marine, volcanic, terrestrial, cosogenic and anthropogenic origin. They are also unique with their entrapped air inclusions in providing direct records of past changes in atmospheric trace-gas composition. The ice-drilling project undertaken in the framework of a long-term collaboration between Russia, the United States and France at the Russian Vostok station in East Antarctica (78°S, 106°E, elevation 3,488 m, mean temperature −55°C) has already provided a wealth of such information for the past two glacial–interglacial cycles4–11. Glacial periods in Antarctica are characterized by much colder temperatures, reduced precipitation and more vigorous large-scale atmospheric circulation. There is a close correlation between Antarctic temperature and atmospheric concentrations of CO₂ and CH₄ (refs 5, 9). This discovery suggests that greenhouse gases are important as amplifiers of the initial orbital forcing and may have significantly contributed to the glacial–interglacial changes12–16. The Vostok ice cores were also used to infer an empirical estimate of the sensitivity of global climate to future anthropogenic increases of greenhouse-gas concentrations15.

The recent completion of the ice-core drilling at Vostok allows us to considerably extend the ice-core record of climate properties at this site. In January 1998, the Vostok project yielded the deepest ice core ever recovered, reaching a depth of 3,623 m (ref. 17). Drilling then stopped ~120 m above the surface of the Vostok lake, a deep subglacial lake which extends below the ice sheet over a large area18, in order to avoid any risk that drilling fluid would contaminate the lake water. Preliminary data17 indicated that the Vostok ice-core record extended through four climate cycles, with ice slightly older than 400 kyr at a depth of 3,310 m, thus spanning a period comparable to that covered by numerous oceanic1 and continental2 records.

Here we present a series of detailed Vostok records covering this ~400-kyr period. We show that the main features of the more recent Vostok climate cycle resemble those observed in earlier cycles. In particular, we confirm the strong correlation between atmospheric greenhouse-gas concentrations and Antarctic temperature, as well as the strong imprint of obliquity and precession in most of the climate time series. Our records reveal both similarities and differences between the successive interglacial periods. They suggest the lead of Antarctic air temperature, and of atmospheric greenhouse-gas concentrations, with respect to global ice volume and Greenland air-temperature changes during glacial terminations.

The ice record

The data are shown in Figs 1, 2 and 3 (see Supplementary Information for the numerical data). They include the deuterium content of the ice (δD atm, a proxy of local temperature change), the dust content (desert aerosols), the concentration of sodium (marine aerosol), and from the entrapped air the greenhouse gases CO₂ and CH₄, and the δ¹⁸O of O₂ (hereafter δ¹⁸O atm) which reflects changes in global ice volume and in the hydrological cycle19. (δD and δ¹⁸O are defined in the legends to Figs 1 and 2, respectively.) All these measurements have been performed using methods previously described except for slight modifications (see figure legends).

The detailed record of δD atm (Fig. 1) confirms the main features of the third and fourth climate cycles previously illustrated by the coarse-resolution record27. However, a sudden decrease from interglacial-like to glacial-like values, rapidly followed by an abrupt return to interglacial-like values, occurs between 3,320 and 3,330 m.
In addition, a transition from low to high CO₂ and CH₄ values (not shown) occurs at exactly the same depth. In undisturbed ice, the transition in atmospheric composition would be found a few metres lower (due to the difference between the age of the ice and the age of the gas²⁰). Also, three volcanic ash layers, just a few centimetres apart but inclined in opposite directions, have been observed—10 m above this δD excursion (3,311 m). Similar inclined layers were observed in the deepest part of the GRIP and GISP2 ice cores from central Greenland, where they are believed to be associated with ice flow disturbances. Vostok climate records are thus probably disturbed below these ash layers, whereas none of the six records show any indication of disturbances above this level. We therefore limit

**Figure 1** The deuterium record. Deuterium content as a function of depth, expressed as δD (in ‰ with respect to Standard Mean Ocean Water, SMOW). This record combines data available down to 2,755 m (ref.13) and new measurements performed on core 5G (continuous 1-m ice increments) from 2,755 m to 3,350 m. Measurement accuracy (1σ) is better than 1 ‰. Inset, the detailed deuterium profile for the lowest part of the record showing a δD excursion between 3,320 and 3,330 m. δD,(‰) = [(δDsample/δDstandard) - 1] × 1,000.

**Figure 2** Vostok time series and ice volume. Time series (GT4 timescale for ice on the lower axis, with indication of corresponding depths on the top axis and indication of the two fixed points at 110 and 390 kyr) of: a, deuterium profile (from Fig. 1); b, δ¹⁸O atm profile obtained combining published data¹¹,²⁵ and 81 new measurements performed below 2,760 m. The age of the gas is calculated as described in ref. 20; c, seawater δ¹⁸O (ice volume proxy) and marine isotope stages adapted from Bassinot et al.²⁶; d, sodium profile obtained by combination of published and new measurements (performed both at LGGE and RSMAS) with a mean sampling interval of 3–4 m (ng g⁻¹ or p.p.b); and e, dust profile (volume of particles measured using a Coulter counter) combining published data¹⁰,¹³ and extended below 2,760 m, every 4 m on the average (concentrations are expressed in μg g⁻¹ or p.p.m. assuming that Antarctic dust has a density of 2,500 kg m⁻³). δ¹⁸O atm (‰) = [(¹⁸O/¹⁶Ogas/¹⁸O/¹⁶Ostandard) - 1] × 1,000; standard is modern air composition.
the discussion of our new data sets to the upper 3,310 m of the ice core, that is, down to the interglacial corresponding to marine stage 11.3.

Lorius et al.4 established a glaciological timescale for the first climate cycle of Vostok by combining an ice-flow model and an ice-accumulation model. This model was extended and modified in several studies12,13. The glaciological timescale provides a chronology based on physics, which makes no assumption about climate forcings or climate correlation except for one or two adopted control ages. Here, we further extend the Extended Glaciological Timescale (EGT) of Jouzel et al.12 to derive GT4, which we adopt as our primary chronology (see Box 1). GT4 provides an age of 423 kyr at a depth of 3,310 m.

Climate and atmospheric trends

Temperature. As a result of fractionation processes, the isotopic content of snow in East Antarctica (δD or δ18O) is linearly related to the temperature above the inversion level, T inv, where precipitation forms, and also to the surface temperature of the precipitation site, T s (with ΔT s = 0.67ΔT inv, see ref. 6). We calculate temperature changes from the present temperature at the atmospheric level as ΔT inv = (ΔδD, inv − 8Δδ18O, inv)/9, where Δδ18O, inv is the globally averaged change from today’s value of seawater δ18O, and 9‰ per °C is the spatial isotope/temperature gradient derived from deuterium data in this sector of East Antarctica21. We applied the above relationship to calculate ΔT inv. This approach underestimates ΔT inv by a factor of ~2 in Greenland22 and, possibly, by up to 50% in Antarctica23. However, recent model results suggest that any underestimation of temperature changes from this equation is small for Antarctica24,25.

To calculate ΔT s from δD, we need to adopt a curve for the change in the isotopic composition of sea water versus time and correlate it with Vostok. We use the stacked δ18Osw record of Bassinot et al.26, scaled with respect to the V19-30 marine sediment record over their common part that covers the past 340 kyr (ref. 27) (Fig. 2). To avoid distortions in the calculation of ΔT s linked with dating uncertainties, we correlate the records by performing a peak to peak adjustment between the ice and ocean isotopic records. The δ18Osw correction corresponds to a maximum ΔT s correction of ~1 °C and associated uncertainties are therefore small. We do not attempt to correct ΔT s either for the change of the altitude of the ice sheet or for the origin of the ice upstream of Vostok; these terms are very poorly known and, in any case, are also small (<1 °C).

The overall amplitude of the glacial–interglacial temperature change is ~8 °C for ΔT s (inversion level) and ~12 °C for ΔT inv, the temperature at the surface (Fig. 3). Broad features of this record are thought to be of large geographical significance (Antarctica and part of the Southern Hemisphere), at least qualitatively. When examined in detail, however, the Vostok record may differ from coastal28 sites in East Antarctica and perhaps from West Antarctica as well. Jouzel et al.12 noted that temperature variations estimated from deuterium were similar for the last two glacial periods. The third and fourth climate cycles are of shorter duration than the first two cycles in the Vostok record. The same is true in the deep-sea record, where the third and fourth cycles span four precessional cycles rather than five as for the last two cycles (Fig. 3). Despite this difference, one observes, for all four climate cycles, the same ‘sawtooth’ sequence of a warm interglacial (stages 11.3, 9.3, 7.5 and 5.5), followed by increasingly colder interstadial events, and ending with a rapid return towards the following interglacial. The

Figure 3 Vostok time series and insolation. Series with respect to time (GT4 timescale for ice on the lower axis, with indication of corresponding depths on the top axis) of: a, CO2; b, isotopic temperature of the atmosphere (see text); c, CH4; d, δ18O atm; and e, mid-June insolation at 65°N (in W m−2) (ref. 3). CO2 and CH4 measurements have been performed using the methods and analytical procedures previously described18. However, the CO2 measuring system has been slightly modified in order to increase the sensitivity of the CO2 detection. The thermal conductivity chromatographic detector has been replaced by a flame ionization detector which measures CO2 after its transformation into CH4. The mean resolution of the CO2 (CH4) profile is about 1,500 (950) years. It goes up to about 6,000 years for CO2 in the fractured zones and in the bottom part of the record, whereas the CH4 time resolution ranges between a few tens of years to 4,500 years. The overall accuracy for CH4 and CO2 measurements are ±20 p.p.b.v. and 2–3 p.p.m.v., respectively. No gravitational correction has been applied.
The coolest part of each glacial period occurs just before the glacial termination, except for the third cycle. This may reflect the fact that the June 65°N insolation minimum preceding this transition (255 kyr ago) has higher insolation than the previous one (280 kyr ago), unlike the three other glacial periods. Nonetheless, minimum temperatures are remarkably similar, within 1°C, for the four climate cycles. The new data confirm that the warmest temperature at stage 7.5 was slightly warmer than the Holocene, and show that stage 9.3 (where the highest deuterium value, ~414.8‰, is found) was at least as warm as stage 5.5. That part of stage 11.3, which is present in Vostok, does not correspond to a particularly warm climate as suggested for this period by deep-sea sediment records. As noted above, however, the Vostok records are probably disturbed below 3,310 m, and we may not have sampled the warmest ice of this interglacial. In general, climate cycles are more uniform at Vostok than in deep-sea core records. The climate record makes it unlikely that the West Antarctic ice sheet collapsed during the past 420 kyr (or at least shows a marked insensitivity of the central part of East Antarctica and its climate to such a disintegration).

The power spectrum of $\Delta T_t$ (Fig. 4) shows a large concentration of variance (37%) in the 100-kyr band along with a significant concentration (23%) in the obliquity band (peak at 41 kyr). This strong obliquity component is roughly in phase with the annual insolation at the Vostok site. The variability of annual insolation at 78°S is relatively large, 7% (ref. 3). This supports the notion that annual insolation changes in high southern latitudes influence Vostok temperature. These changes may, in particular, contribute to the initiation of Antarctic warming during major terminations, which (as we show below) herald the start of deglaciation.

There is little variance (11%) in $\Delta T_t$ around precessional periodocities (23 and 19 kyr). In this band, the position of the spectral peaks is affected by uncertainties in the timescale. To illustrate this point, we carried out, as a sensitivity test, a spectral analysis using the control points provided by the $\delta^{18}$O atm record (see Table 1). The position and strength of the 100- and 40-kyr-spectral peaks are unaffected, whereas the power spectrum is significantly modified for periodocities lower than 30 kyr.

**Insolation.** $\delta^{18}$O atm strongly depends on climate and related properties, which reflect the direct or indirect influence of insolation. As a result, there is a striking resemblance between $\delta^{18}$O atm and mid-June insolation at 65°N for the entire Vostok record (Fig. 3). This provides information on the validity of our glaciological timescale.
(see Box 1). The precessional frequencies, which do not account for much variance in $\Delta T$, are strongly imprinted in the $\delta^{18}O_{\text{atm}}$ record (36% of the variance in this band, Fig. 4). In addition, the remarkable agreement observed back to stage 7.5 between the amplitude of the filtered components of the mid-June insolation at 65°N and $\delta^{18}O_{\text{atm}}$ in the precessional band holds true over the last four climatic cycles (not shown). As suggested by the high variance of $\delta^{18}O_{\text{atm}}$ in the precessional band, this orbital frequency is also reflected in the Dole effect, the difference between $\delta^{18}O_{\text{atm}}$ and $\delta^{18}O_{\text{snow}}$, confirming results obtained on the last two climatic cycles.18,20

**Aerosols.** Figure 2 shows records of aerosols of different origins. The sodium record represents mainly sea-salt aerosol entrained from the ocean surface, whereas the dust record corresponds to the small size fraction ($\sim 2 \mu m$) of the aerosol produced by the continent. The extension of the Vostok record confirms much higher fallout during cold glacial periods than during interglacials. Concentrations range up to 120 ng g$^{-1}$, that is, 3 to 4 times the Holocene value, for sea-salt. For dust, they rise from about 50 ng g$^{-1}$ during interglacials to 1,000–2,000 ng g$^{-1}$ during cold stages 2, 4, 6, 8 and 10. The sodium concentration is closely anti-correlated with isotopic temperature ($r^2 = -0.70$ over the past 420 kyr). The power spectrum of the sodium concentration, like that of $\Delta T$, shows periodicities around 100, 40 and 20 kyr (Fig. 4). Conditions prevailing during the present day austral winter could help explain the observed glacial-interglacial changes in sodium. The seasonal increase of marine aerosol observed in the atmosphere and snow at the South Pole in September31 corresponds to the maximum extent of sea ice; this is because the more distant source effect is compensated by the greater cyclic activity, and by the more efficient zonal and meridional atmospheric circulation probably driven by the steeper meridional (ocean–Antarctica) temperature gradient. These modern winter conditions may be an analogue for glacial climates, supporting the apparent close anti-correlation between sodium concentration and temperature at Vostok.

The extension of the Vostok dust record confirms that continental aridity, dust mobilization and transport are more prevalent during glacial climates, as also reflected globally in many dust records (see ref. 10 and references therein). The presence of larger particles in the Vostok record, at least during the Last Glacial Maximum, indicates that the atmospheric circulation at high southern latitudes was more turbulent at that time. Lower atmospheric moisture content and reduced hydrological fluxes may also have contributed significantly (that is, one order of magnitude) to the very large increases of dust fallout during full glacial periods because of a lower aerosol-removal efficiency.

Unlike sodium concentration, the dust record is not well correlated with temperature (see below) and shows large concentrations of variance in the 100- and 41-kyr spectral bands (Fig. 4). The Vostok dust record is, in this respect, similar to the tropical Atlantic dust record of de Menocal29 who attributes these spectral characteristics to the progressive glaciation of the Northern Hemisphere and the greater involvement of the deep ocean circulation. We suggest that there also may be some link between the Vostok dust record and deep ocean circulation through the extension of sea ice in the South Atlantic Ocean, itself thought to be coeval with a reduced deep ocean circulation.16,20 Our suggestion is based on the fact that the dust source for the East Antarctic plateau appears to be South America, most likely the Patagonian plain, during all climatic stages of the past 420 kyr (refs 35, 36). The extension of sea ice in the South Atlantic during glacial times greatly affects South American climate, with a more northerly position of the polar front and the belt of Westerlies pushed northward over the Andes. This should lead, in these mountainous areas, to intense glacial and fluvial erosion, and to colder and drier climate with extensive dust mobilization (as evidenced by glacial loess deposits in Patagonia). Northward extension of sea ice in the South Atlantic during glacial times greatly affects South American climate, with a more northerly position of the polar front and the belt of Westerlies pushed northward over the Andes. This should lead, in these mountainous areas, to intense glacial and fluvial erosion, and to colder and drier climate with extensive dust mobilization (as evidenced by glacial loess deposits in Patagonia). Northward extension of sea ice also leads to a steeper meridional temperature gradient and to more efficient poleward transport. Therefore, Vostok dust peaks would correspond to periods of increased sea-ice extent in the South Atlantic Ocean, probably associated with reduced deep ocean circulation (thus explaining observed similarities with the tropical ocean dust record).30

**Greenhouse gases.** The extension of the greenhouse-gas record shows that the main trends of CO$_2$ and CH$_4$ concentration changes are similar for each glacial cycle (Fig. 3). Major transitions from the lowest to the highest values are associated with glacial–interglacial transitions. At these times, the atmospheric concentrations of CO$_2$ rises from 180 to 280–300 p.p.m.v. and that of CH$_4$ rises from 320–350 to 650–770 p.p.b.v. There are significant differences between the CH$_4$ concentration change associated with deglaciations. Termination III shows the smallest CH$_4$ increase, whereas termination IV shows the largest (Fig. 5). Differences in the changes over deglaciations are less significant for CO$_2$. The decrease of CO$_2$ to the minimum values of glacial times is slower than its increase towards interglacial levels, confirming the sawtooth record of this property. CH$_4$ also decreases slowly to its background level, but with a series of superimposed peaks whose amplitude decreases during the course of each glaciation. Each CH$_4$ peak is itself characterized by rapid increases and slower decreases, but our resolution is currently inadequate to capture the detail of millennial-scale CH$_4$ variations. During glacial inception, Antarctic temperature and CH$_4$ concentrations decrease in phase. The CO$_2$ decrease lags the temperature decrease by several kyr and may be either steep (as at the end of interglacials 5.5 and 7.5) or more regular (at the end of interglacials 9.3 and 11.3). The differences in concentration–time profiles of CO$_2$ and CH$_4$ are reflected in the power spectra (Fig. 4). The 100-kyr component dominates both CO$_2$ and CH$_4$ records. However, the obliquity and precession components are much stronger for CH$_4$ than for CO$_2$. The extension of the greenhouse-gas record shows that present-day levels of CO$_2$, CH$_4$ (~360 p.p.m.v. and ~1,700 p.p.b.v., respectively) are unprecedented during the past 420 kyr. Pre-industrial Holocene levels (~280 p.p.m.v. and ~650 p.p.b.v., respectively) are found during all interglacials, while values higher than these are found in stages 5.5, 9.3 and 11.3 (this last stage is probably incomplete), with the highest values during stage 9.3 (300 p.p.m.v. and 780 p.p.b.v., respectively).

The overall correlation between our CO$_2$ and CH$_4$ records and the Antarctic isotopic temperature9,16,20 is remarkable ($r^2 = 0.71$ and 0.73 for CO$_2$ and CH$_4$, respectively). This high correlation indicates that CO$_2$ and CH$_4$ may have contributed to the glacial–interglacial changes over this entire period by amplifying the orbital forcing along with albedo, and possibly other changes. The direct radiative forcing corresponding to the CO$_2$, CH$_4$ and N$_2$O changes is the largest CO$_2$ change, which occurs between stages 10 and 9, implies a direct radiative warming of

---

**Table 1 Comparison of the glaciological timescale and orbitally derived information**

| Depth (m) | Insolation maximum (kyr) | Age GT4 (kyr) | Difference (kyr) |
|----------|--------------------------|--------------|-----------------|
| 305      | 11                       | 10           | 1               |
| 900      | 58                       | 57           | 1               |
| 1,213    | 84                       | 83           | 1               |
| 1,528    | 105                      | 105          | 0               |
| 1,863    | 128                      | 128          | 0               |
| 2,110    | 151                      | 150          | 1               |
| 2,350    | 176                      | 179          | -3              |
| 2,530    | 199                      | 203          | -4              |
| 2,680    | 220                      | 222          | -2              |
| 2,788    | 244                      | 239          | 5               |
| 2,863    | 265                      | 256          | 10              |
| 2,972    | 293                      | 282          | 11              |
| 3,042    | 314                      | 301          | 13              |
| 3,119    | 335                      | 322          | 13              |

Control points were derived assuming a correspondence between maximum 65°N mid-June insolation and $\delta^{18}O_{\text{atm}}$, mid-transitions. Age GT4 refers to the age of the gas obtained after correction for the gas-age/ice-age differences.20
Glacial terminations and interglacials

Our complete Vostok data set allows us to examine all glacial commencements and terminations of the past 420 kyr. We can consider, during the terminations, \( \delta^{18}O_{\text{atm}} \) tracks \( \delta^{18}O_{\text{ice}} \) with a lag of \( \sim 2 \) kyr (ref. 11), the response time of the atmosphere to changes in \( \delta^{18}O_{\text{ice}} \). \( \delta^{18}O_{\text{atm}} \) can thus be taken as an indicator of the large ice-volume changes associated with the deglaciations. Broecker and Henderson\(^4\) recently supported this interpretation for the last two terminations and discussed its limitations. Our extended \( \delta^{18}O_{\text{atm}} \) record indeed reinforces such an interpretation, as it shows that the amplitudes of \( \delta^{18}O_{\text{atm}} \) changes parallel \( \delta^{18}O_{\text{ice}} \) changes for all four terminations. \( \delta^{18}O_{\text{ice}} \) changes are similar for terminations I, II and IV (1.1–1.2‰) but much smaller for termination III (\( \sim 0.6‰ \)). The same is true for \( \delta^{18}O_{\text{atm}} \) (1.4–1.5‰ for I, II and IV, and 0.8‰ for III).

A striking feature of the Vostok deuterium record is that the Holocene, which has already lasted 11 kyr, is, by far, the longest stable warm period recorded in Antarctica during the past 420 kyr (Fig. 5). Interglacials 5.5 and 9.3 are different from the Holocene, but similar to each other in duration, shape and amplitude. During each of these two events, there is a warm period of \( \sim 4 \) kyr followed by a relatively rapid cooling and then a slower temperature decrease (Fig. 5), rather like some North Atlantic deep-sea core records\(^4\). Stage 7.5 is different in all respects, with a slightly colder maximum, a more spiky shape, and a much shorter duration (7 kyr at mid-transition compared with 17 and 20 kyr for stages 5.5 and 9.3, respectively). This difference between stage 7.5 and stages 5.5 or 9.3 may result from the different configuration of the Earth’s orbit (in particular concerning the phase of precession with respect to obliquity\(^5\)). Termination III is also peculiar as far as terrestrial aerosol fallout is concerned. Terminations I, II and IV are marked by a large decrease in dust; high glacial values drop to low interglacial values by the mid-point of the \( \Delta T_{\text{ice}} \) increase. But for termination III, the dust concentration decreases much earlier, with low interglacial values obtained just before a slight cooling event, as for termination I (for which interglacial values are reached just before the ‘Antarctic Cold Reversal’).

Unlike termination I, other terminations show, with our present resolution, no clear temperature anomalies equivalent to the Antarctic Cold Reversal\(^6\) (except possibly at the very beginning of termination III). There are also no older counterparts to the Younger Dryas \( \text{CH}_4 \) minimum\(^4\) during terminations II, III and IV given the present resolution of the \( \text{CH}_4 \) record (which is no better than 1,000–2,000 yr before stage 5).

The sequence of events during terminations III and IV is the same as that previously observed for terminations I and II. Vostok temperature, \( \text{CO}_2 \) and \( \text{CH}_4 \) increase in phase during terminations. Uncertainty in the phasing comes mainly from the sampling frequency and the ubiquitous uncertainty in gas-age/ice-age differences (which are well over \( \pm 1 \) kyr during glaciations and terminations). In a recent paper, Fischer et al.\(^4\) present a \( \text{CO}_2 \) record, from Vostok core, spanning the past three glacial terminations. They conclude that \( \text{CO}_2 \) concentration increases lagged Antarctic warmings by \( 600 \pm 400 \) years. However, considering the large gas-age/ice-age uncertainty (1,000 years, or even more if we consider the accumulation-rate uncertainty), we feel that it is premature to infer the sign of the phase relationship between \( \text{CO}_2 \) and temperature at the start of terminations. We also note that their discussion relates to early deglacial changes, not the entire transitions.

An intriguing aspect of the deglacial \( \text{CH}_4 \) curves is that the atmospheric concentration of \( \text{CH}_4 \) rises slowly, then jumps to a maximum value during the last half of the deglacial temperature rise. For termination I, the \( \text{CH}_4 \) jump corresponds to a rapid Northern Hemisphere warming (Bölling/Allerød) and an increase...
in the rate of Northern Hemisphere deglaciation (meltwater pulse IA)\footnote{A}. We speculate that the same is true for terminations II, III and IV. Supportive evidence comes from the $\delta^{18}O_{a} \delta$ curves. During each termination, $\delta^{18}O_{a}$ begins falling rapidly, signalling intense deglaciation, within 1 kyr of the CH$_4$ jump. The lag of deglaciation and Northern Hemisphere warming with respect to Vostok temperature warming is apparently greater during terminations II and IV ($\sim$9 kyr) than during terminations I and III ($\sim$4–6 kyr). The changes in northern summer insolation maxima are higher during terminations II and IV, whereas the preceding southern summer insolation maxima are higher during terminations I and III. We speculate that variability in phasing from one termination to the next reflects differences in insolation curves\footnote{C} or patterns of abyssal circulation during glacial maximum. Our results suggest that the same sequence of climate forcings occurred during each termination: orbital forcing (possibly through local insolation changes, but this is speculative as we have poor absolute dating) followed by two strong amplifiers, with greenhouse gases acting first, and then deglaciation enhancement via ice-albedo feedback. The end of the deglaciation is then characterized by a clear CO$_2$ maximum for terminations II, III and IV, while this feature is less marked for the Holocene.

Comparison of CO$_2$ atmospheric concentration changes with variations of other properties illuminates oceanic processes influencing glacial–interglacial CO$_2$ changes. As already noted for terminations I and II\footnote{D}, the sequence of climate events described above rules out the possibility that rising sea level induces the CO$_2$ increase at the beginning of terminations. On the other hand, the small CO$_2$ variations associated with Heinrich events\footnote{E} suggest that the formation of North Atlantic Deep Water does not have a large effect on CO$_2$ concentrations. Our record shows similar relative amplitudes of atmospheric CO$_2$ and Vostok temperature changes for the four terminations. Also, values of both CO$_2$ and temperature are significantly higher during stage 7.5 than during stages 7.1 and 7.3, whereas the deep-sea core ice volume record exhibits similar levels for these three stages. These similarities between changes in atmospheric CO$_2$ and Antarctic temperature suggest that the oceanic area around Antarctica plays a role in the long-term CO$_2$ change. An influence of high southern latitudes is also suggested by the comparison with the dust profile, which exhibits a maximum during the periods of lowest CO$_2$. The link between dust and CO$_2$ variations could be through the atmospheric input of iron\footnote{F}. Alternatively, we suggest a link through deep ocean circulation and sea ice extent in the Southern Ocean, both of which play a role in ocean CO$_2$ ventilation and, as suggested above, in the dust input over East Antarctica.

**New constraints on past climate change**

As judged from Vostok records, climate has almost always been in a state of change during the past 420 kyr but within stable bounds (that is, there are maximum and minimum values of climate properties between which climate oscillates). Significant features of the most recent glacial–interglacial cycle are observed in earlier periods of the last glacial period. Nature 326, 273–277 (1987).

1. Imbrie, J. et al. On the structure and origin of major glacial cycles. 1. Linear responses to Milankovich forcing. Palaeoceanography 7, 701–738 (1992).
2. Tzedakis, P. C. et al. Comparison of terrestrial and marine records of changing climate of the last 500,000 years. Earth Planet. Sci. Lett. 150, 171–176 (1997).
3. Berger, A. L. Long-term variations of daily insolation and Quaternary climate change. J. Atmos. Sci. 35, 2362–2367 (1978).
4. Lorius, C. et al. A 150,000-year climatic record from Antarctic ice. Nature 346, 591–596 (1985).
5. Barnola, J. M., Raynaud, D., Korotkevich, Y. S. & Lorius, C. Vostok ice core provides 160,000-year record of atmospheric CO$_2$. Nature 329, 408–414 (1987).
6. Jouzel, J. et al. Vostok ice core: a continuous isotope temperature record over the last climatological cycle (160,000 years). Nature 329, 402–408 (1987).
7. Raisbeck, G. M. et al. Evidence for two intervals of enhanced $\delta^{13}$C deposition in Antarctic ice during the last glacial period. Nature 326, 273–277 (1987).
8. Legrand, M., Lorius, C., Barboux, N. I. & Petiev, V. N. Vostok (Antarctic ice core): atmospheric chemistry changes over the last climatological cycle (160,000 years). Atmos. Environ. 22, 317–331 (1988).
9. Chappellaz, J., Barnola, J.-M., Raynaud, D., Korotkevich, Y. S. & Lorius, C. Ice core record of atmospheric methane over the past 160,000 years. Nature 327, 131–136 (1990).
10. Petit, J.-R. et al. Paleoclimatological implications of the Vostok core dust record. Nature 343, 56–58 (1990).
11. Sowers, T. et al. 135 kyr AR5 Vostok—SPECMAP common temporal framework. Palaeoceanography 8, 737–776 (1993).
12. Jouzel, J. et al. Extending the Vostok ice-core record of palaeoclimate to the penultimate glacial period. Nature 364, 407–412 (1993).
13. Jouzel, J. et al. Climatic interpretation of the recently extended Vostok ice-core record. Clim. Dyn. 12, 513–521 (1996).
14. Gentien, C. et al. Vostok ice core: climatic response to CO$_2$ and orbital forcing changes over the last climatological cycle. Nature 329, 414–418 (1987).
15. Lorius, C., Jouzel, J., Raynaud, D., Hansen, J. & Le Treut, H. Greenland warming, climate sensitivity and ice core data. Nature 347, 139–145 (1990).
16. Raynaud, D. et al. The ice record of greenhouse gases. Science 259, 926–934 (1993).
17. Petit, J.-R. et al. Four climatic cycles in Vostok ice core. Nature 367, 359–360 (1997).
18. Kapsa, A. P., Ridley, J. K., Robin, G. D. Q., Siegert, M. J. & Zotikov, I. A. A large deep freshwater lake beneath the ice of central East Antarctica. Nature 381, 684–686 (1996).
19. Bendle, S., et al. Past temperatures directly from the Greenland ice sheet. Science 282, 268–271 (1998).
20. Barnola, J. M., Pimenta, P., Raynaud, D. & Korotkevich, Y. S. CO$_2$ climate relationship as deduced from the Vostok ice core: a re-examination based on new measurements and on a re-evaluation of the air dating. Tellus B 43, 83–91 (1991).
21. Lorius, C. & Merlivat, L. in Isotopes and Impurities in Snow and Ice. Prot. the Grenoble Symp. Aug/Sept. 1975 127–137 (Pubs. 118, IASHP, 1975).
22. Dahl-Jensen, D. et al. Past temperatures directly from the Greenland ice sheet. Science 282, 268–271 (1998).
23. Hoffmann, G., Masson, V. & Jouzel, J. Stable water isotopes in atmospheric general circulation models. Hydrol. Processes (in the press).
24. Bassinot, F. C. et al. The astronomical theory of climate and the age of the Brunhes-Matuyama magnetic reversal. Earth Planet. Sci. Lett. 126, 91–108 (1994).
25. Shackleton, N. J., Imbrie, J. & Hall, M. A. Oxygen and carbon isotope record of East Pacific core V19-36: implications for the formation of deep water in the late Pleistocene North Atlantic. Earth Planet. Sci. Lett. 60, 233–244 (1983).
26. Stagl, G. et al. Synchronous climate changes in Antarctica and the North Atlantic. Science 282, 95–98 (1995).
27. Howard, W. A warm future in the past. Nature 388, 418–419 (1997).
28. Malini, B., Faullid, D., Jouzel, J. & Raynaud, D. The Dole effect over the last two glacial-interglacial cycles. J. Geophys. Res. (in the press).
29. Legrand, M. & Delmas, R. J. Formation of HCl in the Antarctic atmosphere. J. Geophys. Res. 93, 7153–7168 (1987).
30. Yung, Y. K., Lee, T., Chung-Ho & Shek, Y. T. Dust: diagnostic of the hydrological cycle during the last glacial maximum. Science 271, 962–963 (1996).
33. de Menocal, P. Plio-Pleistocene African climate. Science 270, 53–59 (1995).
34. CLIMAP. Seasonal Reconstructions of the Earth’s Surface at the Last Glacial Maximum (Geol. Soc. Am., Boulder, Colorado, 1981).
35. Basile, I. et al. Patagonian origin dust deposited in East Antarctica (Vostok and Dome C) during glacial stages 2, 4 and 6. Earth Planet. Sci. Lett. 166, 573–589 (1997).
36. Basile, I. Origin des Aérosols Volcaniques et Continentaux de la Carotte de Glace de Vostok (Antarctique). Thesis, Univ. Joseph Fourier, Grenoble (1997).
37. Leuenberger, M. & Siegenthaler, U. Ice-age atmospheric concentration of nitrous oxide from an Antarctic ice core. Nature 360, 449–451 (1992).
38. Ramstein, G., Serafini-Le Treut, Y., Le Treut, H., Forichon, M. & Joussaume, S. Cloud processes associated with past and future climate changes. Clim. Dyn. 14, 233–247 (1998).
39. Berger, A., Loutre, M. F. & Galle`e, H. Sensitivity of the LLN climate model to the astronomical and CO2 forcings over the last 200 ky. Clim. Dyn. 14, 815–829 (1998).
40. Weaver, A. J., Eby, M., Fanning, A. F. & Wilke, E. C. Simulated influence of carbon dioxide, orbital forcing and ice sheets on the climate of the Last Glacial Maximum. Nature 390, 847–853 (1998).
41. Broecker, W. S. & Henderson, G. M. The sequence of events surrounding termination II and their implications for the causes of glacial interglacial CO2 changes. Palaeoceanography 13, 552–564 (1998).
42. Cortijo, E. et al. Ermian cooling in the Norwegian Sea and North Atlantic ocean preceding ice-sheet growth. Nature 372, 446–449 (1994).
43. Chappellaz, J. et al. Synchronous changes in atmospheric CH4, and Greenland climate between 40 and 8 kyr in: Nature 366, 443–445 (1993).
44. Fischer, H., Wahlen, M., Smith, J., Mastroianni, D. & Deck, B. Ice core records of atmospheric CO2 around the last three glacial terminations. Science 283, 1712–1714 (1999).
45. Stauffer, B. et al. Atmospheric CO2 concentration and millennial-scale climate change during the last glacial period. Nature 392, 59–61 (1998).
46. Martin, J. H. Glacial-interglacial CO2 change: The iron hypothesis. Palaeoecography 5, 1–13 (1990).
47. Paillard, D., Labeyrie, L. & Yiou, P. Macintosh program performs time-series analysis. Eos 77, 379 (1996).
48. Ritz, C. Un Modèle Thermo-mécanique d’évolution Pour le Bassin Glaciaire Antarctique Vostok-Glacier Byrd: Sensibilité aux Valeurs de Paramètres Mal Connu. Thesis, Univ. Grenoble (1992).
49. Blunier, T. et al. Timing of the Antarctic Cold Reversal and the atmospheric CO2 increase with respect to the Younger Dryas event. Geophys. Res. Lett. 24, 2683–2686 (1997).
50. Waelbroeck, C. et al. A comparison of the Vostok ice deuterium record and series from Southern Ocean core MD 88-770 over the last two glacial-interglacial cycles. Clim. Dyn. 12, 113–123 (1995).
51. Blunier, T. et al. Asynchrony of Antarctic and Greenland climate change during the last glacial period. Nature 394, 739–743 (1998).
52. Bender, M., Malazert, B., Orchado, J., Sowers, T. & Jouzel, J. High precision correlations of Greenland and Antarctic ice core records over the last 100 kyr in: The Role of High and Low Latitudes in Milennial Scale Global Change (eds Clark, P. & Webb, R.) (AGU Monogr., Am. Geophys. Union, in press).

Supplementary information is available on Nature’s World-Wide Web site (http://www.nature.com) or as paper copy from the London editorial office of Nature.

Acknowledgements. This work is part of a joint project between Russia, France and USA. We thank the drillers from the St Petersburg Mining Institute, the Russian, French and US participants for field work and ice sampling, and the Russian Antarctic Expeditions (RAE), the Institut Français de Recherches et Technologies Polaires (IFRTP) and the Division of Polar Programs (NSF) for the logistic support. The project is supported in Russia by the Russian Ministry of Sciences, in France by PNEDC (Programme National d’Etudes de la Dynamique du Climat), by Fondation de France and by the CEC (Commission of European Communities) Environment Programme, and in the US by the NSF Science Foundation.

Correspondence and requests for materials should be addressed to J.R.P. (e-mail: petit@glaciog.ujf-grenoble.fr).

436