The ice-core record: climate sensitivity and future greenhouse warming

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The prediction of future greenhouse-gas-induced warming depends critically on the sensitivity of Earth's climate to increasing atmospheric concentrations of these gases. Data from cores drilled in polar ice sheets show a remarkable correlation between past glacial–interglacial temperature changes and the inferred atmospheric concentration of gases such as carbon dioxide and methane. These and other palaeoclimate data are used to assess the role of greenhouse gases in explaining past global climate change, and the validity of models predicting the effect of increasing concentrations of such gases in the atmosphere.

INFRARED-ABSORBING gases in the Earth's atmosphere raise the mean global surface temperature from −18 °C, which it would be in the absence of an atmosphere, to ~15 °C. This trapping of infrared energy emitted by the Earth's surface is known as the greenhouse effect[1-4]. The largest effect is from water vapour, with a significant contribution from carbon dioxide and smaller contributions from ozone, methane, nitrous oxide and, in the past few decades, anthropogenic chlorofluorocarbons (CFCs). The naturally occurring greenhouse effect makes our planet habitable, but there is growing concern about the future of the Earth's climate because of increasing atmospheric concentrations of CO_2, CH_4, N_2O and CFCs, which are largely a result of man's activities. This radiative forcing (the change of the planetary radiation balance) is just over 4 W m^{-2} for a doubling of the CO_2 concentration (from, say, 300 to 600 p.p.m.v.). The increase in atmospheric CO_2, CH_4, N_2O and CFCs that have occurred since the beginning of the industrial revolution results in a forcing of just over 2 W m^{-2} (ref. 5); thus the anthropogenic greenhouse forcing has already reached a level of about half of the doubled-CO_2 radiative forcing, the canonical large forcing used in theoretical climate studies.

The radiative forcing resulting from doubled atmospheric CO_2 would increase the surface and tropospheric temperature by $\Delta T_e \approx 1.2$ °C if there were no feedbacks in the climate system (see Box 1). But there are many possible feedbacks which may occur in response to an initial forcing. For example, a warming climate is likely to alter the amount of water vapour in the atmosphere, the cloud properties, snow cover and sea ice[1-4,6]. Doubled-CO_2 sensitivity experiments using the most comprehensive climate models, the three-dimensional general circulation models (GCMs), yield equilibrium warmings between 1.9 and 5.2 °C[6-11]. The uncertainty in climate sensitivity obtained from GCMs may, however, be even greater than this. A recent comparison of 14 GCMs showed about a threefold variation in climate sensitivity, attributable mainly to differences in the cloud feedbacks[12]. Some simple one-dimensional climate models suggest essentially no net feedback[3]. Thus the net feedback factor f (the amplification of the no-feedback radiative-equilibrium temperature change) is uncertain in the large range $f \approx 1$–4 (Box 1).

One implication of the uncertainty in climate sensitivity is an even greater uncertainty in the climate response time. The time required for the climate system to approach a new equilibrium temperature after a change of radiative forcing varies at least linearly with f[6,13] because the feedbacks come into play in response to the change of temperature, not in response to the climate forcing. A consequence of the large uncertainty in the response time, which may be anywhere from a decade to more than a century, is that the observed global temperature trend of the past century does not provide a very stringent empirical limit on climate sensitivity; indeed, allowing for the possibility of small unknown climate forcings, such as changes of tropospheric aerosols and solar irradiance, and for unforced climate variability, the temperature trend of the past century does not provide a useful bound on climate sensitivity[6,14,15].

A more promising source of empirical information on global climate sensitivity is provided by the changes in Earth's climate during the past few hundred thousand years because the global temperature change was very large, ~4–5 °C[16,17], the climate change was linked with changes of greenhouse-gas concentration (which by themselves resulted in a direct radiative forcing of ~2 W m^{-2}) and the changes were maintained for a sufficiently long time for equilibrium to be achieved, unlike the present greenhouse-gas changes. Based on these considerations there have already been attempts[6,19] to estimate climate sensitivity from conditions during the Last Glacial Maximum, 18,000 years before present (18 kyr BP).

Box 1 Equilibrium temperature, feedback processes and $2 \times CO_2$ experiments

Earth's atmosphere is heated by short-wavelength solar radiation and cooled by the emission of long-wavelength radiation to space. The planetary radiative energy budget per unit area, N, can be written as the difference of these terms

$$N = \frac{1-\alpha}{4} S - \varepsilon\sigma T_s^4 \qquad (1)$$

where S is the solar constant (1,370 W m^{-2}), α the planetary reflectivity or albedo (0.3), ε the effective emissivity of the atmosphere, σ the Stefan-Boltzmann constant and T_s the surface temperature[65]. At equilibrium ($N=0$) and with no atmosphere this yields $T_e = 255$ K. Earth's surface temperature is actually 288 K, primarily because of the presence of the atmosphere and the present greenhouse warming is therefore ~33 °C.

If the atmospheric structure and all other factors remain fixed, the response of the system to an increase of radiative forcing ΔQ would be a change in T_e necessary to restore radiative equilibrium, $\Delta T_e = T_e \Delta Q/(1-\alpha)S$ leading to $\Delta T_e \approx 0.3\,\Delta Q$. The radiative forcing for a doubling of the CO_2 concentration is $\Delta Q \approx 4$ W m^{-2} corresponding to $\Delta T_e \approx 1.2$ °C.

The surface temperature changes, ΔT_s, calculated in recent GCM studies for doubled CO_2[6-11] are shown in the table. Here we have adopted the feedback terminology of Hansen *et al.*[6] in which a net feedback factor f is defined by assuming that the radiative forcing for doubled CO_2 is ~4 W m^{-2} and hence $\Delta T_s \approx 1.2\,f$. The feedbacks that have been found to be significant in these models (water vapour, clouds, sea ice and snow cover) can be described as fast feedbacks, because they respond rapidly to climate change and are thus important on decadal timescales.

Recently, ice cores (see Box 2) from the Greenland[20,21] and Antarctic[22,23] ice sheets and, in particular, the Vostok record[24-30], which spans a full glacial-interglacial cycle[26], have allowed documentation of the relationship between greenhouse radiative forcing (CO_2 and CH_4) and climate over a period of large climate changes. These new data support the orbital theory of ice ages, but they are also directly relevant to the problem of greenhouse gases and climate sensitivity. Here we compare information about climate sensitivity derived from ice cores with the GCM results. Although a full understanding of the causes and sequences of climate fluctuations remains elusive, much information is revealed from comparison of glacial-interglacial palaeotemperature data with concurrent changes in the planet's energy balance that are linked with atmospheric CO_2 and CH_4 concentrations.

The ice-core data suggest that greenhouse gases have had a significant role in explaining the magnitude of past global temperature changes and indicate a net feedback for fast processes of $f \approx 3$, consistent with a climate sensitivity of ~3-4 °C for a doubled atmospheric CO_2. Studies of palaeoclimate changes therefore provide a clue to help validate estimations of the climate impact of increasing greenhouse gases.

Ice-core data

Ice-core data are unique in providing, in the same samples, access to both climate and climate forcings (Box 2). The climate information includes an estimate of temperature changes at the atmospheric level where the snow formed and at the surface, and the information about climate forcing includes data on the amount of aerosols in the atmosphere and the atmospheric chemical composition. So far, there are only two deep ice cores from Greenland (Camp Century[20] and Dye 3[21]) and three from Antarctica (Byrd[22], Dome C[23] and Vostok[24-30]) which provide long-term climate and environmental records extending over the last ice age. Only Vostok provides a complete undisturbed series over the last climate cycle. The principle findings obtained from these deep ice cores have been recently reviewed[31].

The Vostok temperature record is reconstructed from the continuous deuterium profile measured along the core[27]. Here we refer to the atmospheric temperature change, ΔT_a, just above the inversion layer (Box 2) because this is the parameter that is most directly accessible from the snow deuterium content, which depends on the temperature of formation of the precipitation (such an interpretation is well supported by a theoretical approach) and because this temperature is certainly more relevant for characterizing the global atmosphere as the existence of a strong inversion is restricted to central Antarctica. The record (Fig. 1*b*) shows the last two glacial-interglacial transitions with atmospheric temperature changes of ~6 °C. The last ice age is characterized by three minima, around 20, 60 and 110 kyr BP, separated by slightly warmer episodes (interstadials).

The three Antarctic ice cores yield remarkably similar isotope-derived temperature records over the last ~60 kyr (ref. 32). The surface warmings associated with the last deglaciation are quite comparable with values slightly below 10 °C and the Vostok record can be considered to be representative of a large area in Antarctica. The deglaciation warming for Dye 3 is similar to the one observed in Antarctica (whereas the Camp Century core suggests a somewhat larger change) and in fact, the main features of the Vostok climate record are of global significance, at least qualitatively speaking. This significance is suggested by the comparison with the global ice volume changes derived from isotope measurements in deep-sea sediments (during glacial periods the accumulation of isotopically depleted ice over the continents leads to a measurable isotope enrichment of the ocean). In particular the SPECMAP record of marine ^{18}O content[33], which essentially represents global ice volume (Fig. 1*c*), and the Vostok temperature record corresponds very closely (correlation coefficient, $r^2 = 0.87$) down to 110 kyr BP[27]. Further back there is a discrepancy in the start and in the duration of the previous interglacial which can probably be explained by dating uncertainties for this part of the record (see discussion below).

The close association between greenhouse gases and glacial-interglacial climate changes was first revealed from the analysis of air trapped in Greenland and Antarctic cores showing that the atmospheric CO_2 content was ~190-200 p.p.m.v. during the Last Glacial Maximum compared with an average close to 270-280 p.p.m.v. during the Holocene[34-37]. The detailed reconstruction from Byrd[38] for the period between 50 and 15 kyr fully confirmed the association between cold climate and low atmospheric CO_2 levels (Fig. 2*a*). The Vostok data confirmed the existence of a strong CO_2-climate relationship over a full climate cycle[24]. This CO_2 record exhibits large changes between two levels that are centred near 190-200 and 260-280 p.p.m.v., with the low and high levels associated with full glacial and interglacial conditions, respectively (Fig. 1*a*). Moreover, despite some noticeable differences, there is a remarkable correlation ($r^2 = 0.79$) between atmospheric CO_2 and temperature change throughout the record.

Such a correlation between climate change and radiatively active gases also exists for methane. Analysis of Dye 3 and Byrd samples showed an increase in the CH_4 concentration from 0.35 p.p.m.v. for the Last Glacial Maximum to Holocene values of 0.65 p.p.m.v.[39], and a similar increase parallelling the tem-

Box 2 Ice cores, climate and atmospheric data

Although we cannot expect to find an ideal record of climate, ice is a rather close approximation to it as summarized in the table below[68].

CLIMATE FACTORS AND QUANTITIES MEASURED IN ICE CORES

Atmosphere	Ice core
Temperature	D/H, $^{18}O/^{16}O$
Precipitation	D/H, $^{18}O/^{16}O$, ^{10}Be
Humidity	D/H, $^{18}O/^{16}O$
Aerosols	
natural (continents, sea, volcanoes, biosphere)	chemicals (Al, Ca, Na, H, SO_4, NO_3)
man-made	SO_4, NO_3, Pb, radioactive fallout particles
Circulation	
Gases: natural and man-made	O_2, N_2, CO_2, CH_4, N_2O

Palaeotemperature reconstruction is based on the present correlations between the ratios of ^2H (D) and ^1H and of ^{18}O and ^{16}O in the snow and the temperature conditions just above the inversion layer, where the precipitation is formed, and at the surface. These are correlated because of the fractionation processes that take place during the atmospheric water cycle and, although they depend on several parameters, lead to linear isotope-temperature relationships in both polar ice sheets[69,70]. The validity of using the present observed relationship for palaeotemperature reconstruction is supported by various evidence including atmospheric isotope models[71]. Although there is no doubt that the concentration of a species in the air is reflected in snow deposits at the site, quantitative estimation of atmospheric concentrations suffers from an incomplete understanding of the deposition processes, such as the relative contribution of 'wet' and 'dry' processes. In contrast, relating gas concentrations obtained from ice-core bubbles to the atmospheric value is quite straightforward. Owing to the air-enclosure process in the ice, however, the air is isolated from the atmosphere well after the precipitation has been deposited; in a site of low accumulation like Vostok the correction ranges from 3 to 6 kyr with an uncertainty of up to ~1 kyr[24,25].

Adequate site selection is required to avoid disturbance of the atmospheric record by ice flow. Current deep-drilling projects to obtain long-term records include US and European efforts in central Greenland; hopefully this will supplement the few ice cores available now and provide the awaited counterpart of the Vostok Antarctic record needed for a better understanding of glacial-interglacial forcings and feedbacks.

perature change was revealed from the Vostok ice for the transition between the penultimate ice age and the last interglacial[29]. These results are now fully confirmed[30] by the detailed and continuous CH_4 profile obtained along the Vostok core (Fig. 1*c*), which also exhibits four well marked maxima during the glacial period. These maxima occur at the time of relatively warm interstadials and although there are marked differences between the CH_4 and CO_2 record, the correlation with the temperature record is quite similar ($r^2 = 0.78$ in the case of CH_4).

Superimposed on these long-term trends, rapid and large changes have been documented both for CO_2 and CH_4. In the time interval between 40 and 30 kyr a series of rapid CO_2 variations (Fig. 2*b*), occurring in a century or less between low values in the range 180-200 p.p.m.v. and higher values in the range 240-260 p.p.m.v., were found first in Dye 3[40]. The existence of these rapid events, associated with large changes in surface temperature (5-6 °C) and in the dust content, has been confirmed in the Camp Century core but has not been identified in the Antarctic Byrd core, a difference not yet satisfactorily explained. There is also a clear indication of rapid climate change at the end of the last deglaciation extensively documented from isotope and dust studies on Greenland cores[41]. This so-called Younger Dryas cooling dated at ~11-10 kyr BP is well marked in North Atlantic and adjacent regions. Although there is also a weak cooling in Antarctica the dating is not yet sufficiently accurate to allow it to be firmly related to the Younger Dryas. Recent Vostok measurements[30] show that the Antarctic cooling was associated with a large decrease in CH_4 concentration (Fig. 1*c*), as is also suggested in the data from the Dye 3 core[39].

Reliable N_2O data can also be obtained from ice cores. Existing results suggest that N_2O concentrations during the Last Glacial Maximum were possibly smaller but not drastically different from those of the recent period[42]. We can therefore estimate the direct radiative forcing exerted by all naturally occurring greenhouse gases (except changes in water vapour and possibly in ozone). In terms of radiative forcing such changes in the concentration of greenhouse gases are important because, for example, of the logarithmic increase with CO_2 and CH_4 concentrations[43]. From glacial to interglacial this forcing increases by $\sim 2\ \mathrm{W\,m^{-2}}$ (ΔT_e of ~0.7 °C), half of that corresponding to a CO_2 doubling from 300 to 600 p.p.m.v.

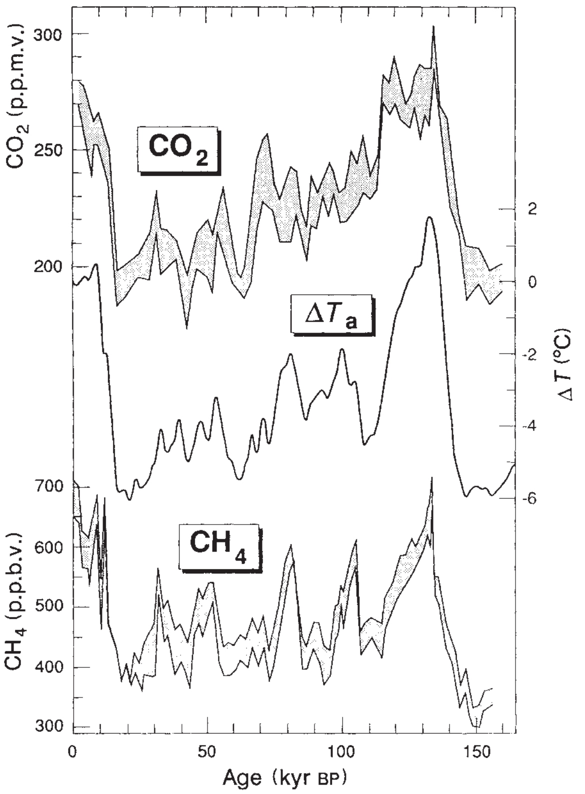

FIG. 1 Variation during the last climate cycle, as derived from measurements along the 2,083-m Vostok ice core, of the CO_2 atmospheric concentration[24], the atmospheric temperature change over Antarctica[27], and the CH_4 atmospheric concentration[30]. For CO_2 and CH_4 the envelope shown has been plotted taking into account the different sources of uncertainty whereas the temperature record is a smoothed curve.

Ice-core data and the theory of ice ages

The long Vostok ice core offered a unique opportunity to examine a continental record of the link between a time series for atmospheric temperature and the astronomical theory of the ice ages. Among the orbital parameters the obliquity of the Earth's axis (period of ~41 kyr) and the precession of the equinoxes (periods of 23 and 19 kyr) are the most important as they strongly affect the distribution of available energy between latitudes and seasons[44]; for example, the variation of the summer insolation at 65 °N, a key latitude in the astronomical theory of the ice ages, is $\sim 60\ \mathrm{W\,m^{-2}}$ over the last climate cycle. Spectral analysis shows that the Vostok temperature record is indeed dominated by a strong obliquity component and influenced to a lesser degree by the precession[27]. Aside from the interval of ~100 kyr between interglacials, the main features of the Vostok temperature record show that the minima are in good agreement with those of the annual insolation received at the Vostok latitude (78° S) and that there are also similarities with the 65° N summer insolation. These results further support the role of orbital forcing in Pleistocene glacial-interglacial cycles demonstrated by deep-sea sediment records[45]. The orbital forcing is, however, relatively weak when considered on an annual globally averaged basis (the total insolation received by Earth has varied by $<0.7\ \mathrm{W\,m^{-2}}$ over the past 160 kyr). The amplification of this forcing, the observed dominant 100-kyr cycle and the synchronized termination of the main glaciations and their similar amplitude in the Northern and Southern Hemispheres cannot easily be explained despite developments including the nonlinear response of ice sheets to orbital forcing[46]. The discovery of significant changes in climate forcing linked with the composition of the atmosphere has led to the idea that changes in the CO_2[24,28,47-49] and CH_4[29,30] content have played a significant part in the glacial-interglacial climate changes by amplifying, together with the growth and decay of the Northern Hemisphere ice sheets, the relatively weak orbital forcing and by constituting a link between the Northern and Southern Hemisphere climates.

The observed changes in CO_2 and CH_4 imply modifications in their sources and sinks that probably involve very different processes such as ocean circulation and marine production for CO_2 (see ref. 50 for a review) and fluxes of emission from natural wetlands for CH_4[29,30]. The mechanisms behind these modifications are not yet fully understood, especially those involving CO_2, but we note that the orbital frequencies are present in both of these Vostok series with, in particular, a strong precessional signal[24,30]. This suggests that the changes are in some way orbitally driven and, even if the orbital forcing is not the only important effect, this supports the idea that orbital changes are one initial cause of the ice ages.

Further, the astronomical theory cannot easily explain the rapid events recorded in ice cores. Rather, although the mechanisms are still unknown, rapid changes could be connected to a flip-flop mechanism in the North Atlantic ocean, perhaps a turning on and off of the North Atlantic current[51]. We note, however, that the correspondence between cold episodes and low CO_2 concentration (Fig. 2*b*) and conversely between warmest periods and high concentrations seems to hold true during the last ice age and possibly during the Younger Dryas event, as suggested by the CH_4 data (Fig. 1*c*). There is therefore some indication that greenhouse forcing also participated in the amplitude of these temperature changes.

Ice-core data and the greenhouse effect

The glacial–interglacial changes include not only the fast feedback processes associated with water vapour, cloud properties, snow cover and sea ice but also longer-term processes such as those linked with slow changes in several boundary conditions and in the atmospheric composition, which help maintain the ice age cold relative to the interglacial (Box 1). We now derive information about the role of these slow changes from the Vostok temperature series and compare it with other palaeodata and climate forcings. Using the data on the direct radiative forcing associated with changes in the concentration of greenhouse gases, we derive information about the role of fast feedback processes. This does not require a solution of the 'chicken and egg' problem, that is, we do not have to address fully the question of causes of the glacial–interglacial cycles and of the sequence of possible forcing factors[28,52]. For example, whether the temperature changes lead or lag the changes in CO_2 or CH_4 concentrations is not relevant for the study of fast feedbacks. There is, however, the important constraint that the planet must be in near radiation balance with space. This may require several thousand years and we therefore cannot use this approach to address the question of transients or rapid events, which may occur in a few decades[41]. Within these limits, we may assume that greenhouse gases have contributed to the glacial–interglacial temperature change through their direct radiative forcing associated with fast feedback processes. As discussed below, such an assumption is well supported by the GCM experiments described in refs 49 and 53 and we can take advantage of the long records of climate forcings and climate outputs, which reflect equilibrium conditions, to evaluate (or at least reasonably bracket) the role of greenhouse-gas radiative forcing in determining past climate.

Genthon *et al.*[28] followed such an approach in performing a multivariate analysis of the Vostok temperature record and three climate inputs. This approach decomposes the temperature variance into a part that can be accounted for by the inputs plus a residual, which is due to noise not related to these inputs. The relative contribution of each input is calculated so that the noise is minimized in a least-squares sense[54]. To be able to partition properly the output series as a combination of the input series, the input series must be independent. For climate forcing, Genthon *et al.*[28] used CO_2, a Southern Hemisphere component represented by the local (78° S) annual insolation (Fig. 3*d*) and a Northern Hemisphere component derived either from the 65° N July insolation or from the $\delta^{18}O$ marine record, a proxy of the Northern Hemisphere ice volume (Fig. 3*c*). We have used this approach and extended it to account for the following points.

First, such a multivariate analysis implicitly assumes that the system can be described by its steady-state behaviour. This is not the case for a climate system affected by nonlinearities associated with the growth and decay of the ice sheets. To account for this, we consider cases in which the ice volume reconstructed from the $\delta^{18}O$ record is taken as a proxy of the Northern Hemisphere forcing. This proxy combines the orbital insolation forcing and the nonlinearities mentioned above. We can therefore assume that it represents a large part of the slow feedbacks associated with the changes in boundary conditions, all the more because the increased area of ice sheets provides a large forcing ($\sim$3 W m^{-2} mainly owing to albedo effects but also to topographic changes).

Second, we account for changes in both CO_2 and CH_4 to evaluate the combined greenhouse radiative forcing. For this purpose we use the CO_2 and CH_4 Vostok records (including data from other cores would not result in significant changes, except for rapid events not considered here) and calculate the associated direct radiative forcing through the formulae given by Hansen *et al.*[43]. For CH_4, this direct radiative forcing was augmented by a factor of $\sim$2 because of chemical feedbacks linked with the increase of stratospheric water vapour[30]. The total greenhouse radiative forcing expressed in terms of equilibrium temperature change (with no climate feedback) given in Fig. 3*b* shows a glacial–interglacial amplitude of $\sim$0.7 °C.

Third, we found it useful to explore a wider range of possible atmospheric radiative forcings such as those resulting from changes in the aerosol loading[18,55] and in the amount of cloud condensation nuclei (CCN). A strong increase (up to a factor of 30) in the dust fallout has been recorded for the Last Glacial Maximum both in Greenland and Antarctic cores, as illustrated in Fig. 2*c* for Dome C[56]. It has been claimed that the resulting increase in the optical depth of the atmospheric aerosols could make a significant contribution to the cooling during this period[18]. This is still uncertain because it is difficult to evaluate the global aerosol loading using data coming mostly from polar ice cores. Also, a change in the optical depth and thus in direct radiative forcing depends strongly on the dust optical properties, which are not known. Analysis of the Vostok core[57] showed a strong dust increase at the end of the penultimate ice age and during the cold interstadial centred around 60 kyr BP, which also indicates a relation between cold periods and high aerosol loading. The Vostok dust record is very spiky, however, and not simply related with the smoother temperature record. There is no dust peak during the cold period around 110 kyr and the correlation ($r^2 = 0.36$) with the temperature record is much lower than the corresponding one for greenhouse gases. It has recently been proposed that the number concentration of CCN, mainly sulphate particles produced by the oxidation of dimethylsulphide emitted from the ocean, influences the albedo of marine stratus cloud and hence global climate[55]. In this hypothesis, a change in marine productivity may eventually act as a climate forcing through the induced change in CCN concentration. Using the Vostok non-sea-salt sulphate as a proxy for CCN concentration, Legrand *et al.*[58] suggested that this forcing may have participated in the glacial cooling. The correlation between non-sea-salt sulphate and the Vostok temperature is relatively high ($r^2 = 0.63$) but it must be noted that the role of CCN and the CCN–sulphate link are not yet firmly established.

TABLE 1 Results from recent doubled-CO_2 models

Study	Source	ΔT_s	f
Goddard Institute for Space Studies	ref. 6	4.2	3.5
National Center for Atmospheric Research	ref. 7	4.0	3.3
Geophysical Fluid Dynamics Laboratory	ref. 8	4.0	3.3
UK Meteorological Office	ref. 9	5.2	4.3
Oregon State University	ref. 10	2.8	2.3
UK Meteorological Office	ref. 11	1.9	1.6

Most results are adapted from ref. 1; an extreme experiment[11] in which cloud cover and cloud radiative properties depend strongly on temperature cloud water content is also included. The values of ΔT_s obtained by the GCMs, 1.9–5.2 °C, are substantially larger than the values of ΔT_e, indicating that the modelled processes significantly amplify the direct greenhouse forcing. More general discussions of the equilibrium climate sensitivity relevant to short timescales fall in the range $\Delta T_s = 1.5$–5.5 °C[1–3,14,15]. Thus the uncertainty in f seems to be $f \approx 1$–4. There are other potentially important feedbacks, which are not investigated by the GCM studies because they are treated as either fixed boundary conditions or as specified climate forcings. Many of these could be classified as slow feedbacks which may be significant on long timescales, such as glacial–interglacial climate change. Examples include changes of land ice, floating ice shelves[66] vegetation cover[6], and ocean circulation[67]. Possible atmospheric changes include the abundance of greenhouse gases, dust[18] and cloud condensation nuclei[55], including the impact of these on optical properties of clouds[18].

Finally, we reexamined the previously noted influence of the dating[28]. Indeed, the multivariate analysis must be done with all time series having a common timescale. This is not a problem for the temperature, non-sea-salt sulphate and dust Vostok series and, within the uncertainty associated with the correction due to the air–ice difference (Box 2), for the greenhouse forcing. There is, however, a noticeable dating discrepancy between the Vostok and ice volume records before 110 kyr, which has been shown to affect significantly the results of the multivariate analysis performed by Genthon *et al.*[28]. We tested numerous redatings of the oldest part of the Vostok series accounting for the recent interpretation of the Vostok dust record, which suggests that the marine and Vostok records are roughly in phase before 110 kyr[57]. A new multivariate analysis was then performed using five forcing factors: greenhouse gases, dust and non-sea-salt sulphate concentrations, ice volume and local insolation. The results may be summarized as follows:

There is no case for which the contribution of the greenhouse effect is below 40%. The lowest calculated value is 42% when the Vostok record is put exactly in phase with the marine $\delta^{18}O$ record, with a similar contribution from the ice volume (45%) and with contributions of $\leqslant$5% for each of the three other climate inputs. With this redating, Genthon *et al.*[28] obtained a CO_2 contribution of 35%. Indeed, for all analyses performed, adding CH_4 leads to a ~10% increase over the effect derived using only CO_2. This may be related[59] to the higher correlation of the Vostok temperature with the combined greenhouse ($r^2 = 0.84$) than with either CO_2 ($r^2 = 0.79$) or CH_4 ($r^2 = 0.78$).

Within the limits of dating by comparison of the Vostok dust record with marine records, the greenhouse contribution never exceeds 65%, with the sum of greenhouse and Northern Hemisphere forcing being generally over 80%. Generally >90% of the variance of the Vostok temperature is explained by the five climate inputs.

The greenhouse contribution falls in the top of this 40–65% range (between 55 and 65%) when the ice volume record of Labeyrie *et al.*[60] is used, instead of the SPECMAP curve, to represent the Northern Hemisphere forcing. This alternative approach was followed because the Labeyrie *et al.* curve probably represents the ice volume change better than the more generally used SPECMAP record, because it is corrected for the part of the isotope signal that comes from the change in ocean temperature. Also, there may be, at least for CO_2[24], some link between sea level (or ice volume) and the change in the atmospheric concentration and using different ice-volume curves is one way to estimate the bias introduced in the multivariate analysis by two of the input series (greenhouse and ice volume) not being fully independent.

To summarize, the contribution of greenhouse gases to the Vostok temperature change can be bracketed between a lower estimate of 40% and a higher estimate of 65%. This range is somewhat uncertain, however, owing to the nonlinearities linked with the growth and decay of the Northern Hemisphere ice sheet, all the series not having a common timescale and the possibility that there may have been other important forcings that we have not taken into account. We have attempted to account for the first two factors. In addition, the introduction of a new climate input, which must correspond to a process having a potential climate input, will significantly alter the result of the multivariate analysis only if this new series is well correlated with the Vostok temperature and independent of the other inputs. One example of such a forcing is that linked with vegetation change through its influence on the albedo, but model results[43,49], show that the climate role of the vegetation is relatively minor in maintaining the glacial climate. Although we cannot rule out another forcing, it is difficult to imagine another meeting these three requirements. So, within the limits of our multivariate analysis, ~50 ± 10% is a reasonable estimate for the overall contribution of the greenhouse gases to the Vostok temperature change over the last climate cycle. This means that ~3 °C of the 6 °C in the glacial–interglacial change at Vostok could be attributed to the greenhouse effect. This result agrees well with the GCM simulations for the Last Glacial Maximum performed by Broccoli and Manabe[49]. Using a model with prescribed cloud cover, they reduced the CO_2 concentration from 300 to 200 p.p.m.v. in experiments with and without changes in other boundary conditions. The CO_2 change alone, which in terms of radiative forcing roughly corresponds to the combined CO_2 and CH_4 glacial–interglacial change measured in ice cores, results in a temperature drop of 2–2.5 °C in central east Antarctica. With variable cloud, however, the GCM sensitivity is increased by 31%[19] implying a change of ~3 °C over this part of Antarctica. Although based on one particular GCM and experiment, this agreement between two independent approaches is worth noting.

Ice-core data and climate sensitivity

Despite the large geographical significance of the Vostok temperature record, it can be argued that looking at one particular site or area is too restrictive. Other palaeoclimate data and climate simulations are indeed helpful in examining the problem

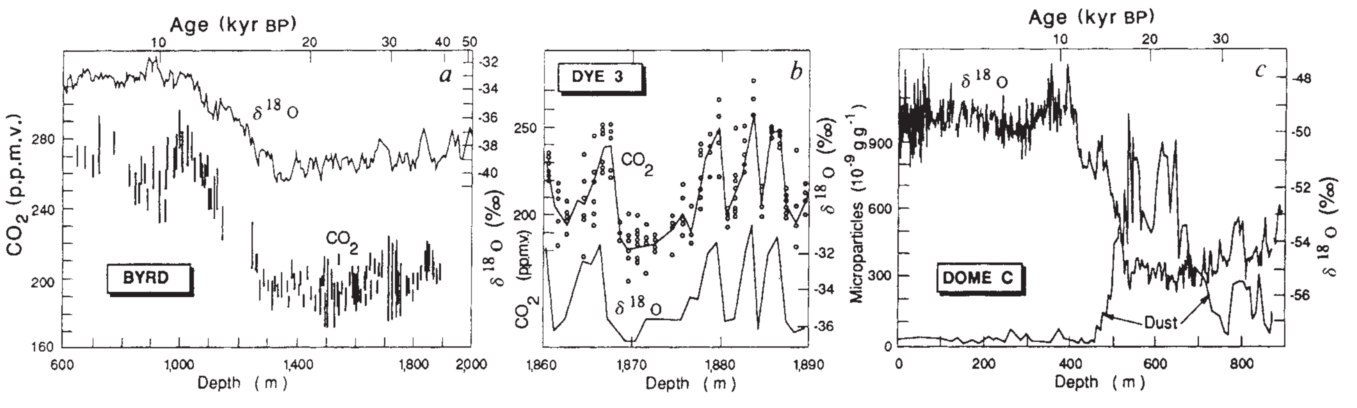

FIG. 2 This panel shows key data obtained from the Byrd (West Antarctica), Dye 3 (Greenland) and Dome C (East Antarctica) cores. The results are given as a function of depth. The corresponding timescale (upper abscissa) is also given, except for Dye 3, for which the 30-m ice increment corresponds to approximately 10 kyr between 40 and 30 kyr. The $\delta^{18}O$ record has been reported for each of these three cores ($\delta^{18}O(‰) = \{[(^{18}O/^{16}O)_{sample}/(^{18}O/^{16}O)_{standard}] - 1\} \times 1{,}000$ where the standard is standard mean ocean water. *a*, Byrd: A detailed CO_2 record covering the last glacial period and part of the Holocene[38], *b*, Dye 3: Rapid CO_2 changes (circles indicate the results of single measurements and the upper solid line connects the mean value for each depth), adapted from ref. 40, c, Dome C: The Dome C dust record showing a strong increase of dust during the Last Glacial Maximum[56].

of greenhouse gases and climate sensitivity in a global perspective.

First, we note that owing to the general polar amplification of climate changes, the Vostok inferred glacial-interglacial warming of ~6 °C over Antarctica is consistent with a globally averaged value of ~4-5 °C; for example, in doubled-CO_2 GCM experiments this amplification is, on average, ~50% for high-latitude continental regions[3]. This indicates that we can extend the conclusion that the greenhouse effect resulted in ~50% of the temperature change in the last climatic cycle to be a global estimate. This is supported by the similarities observed between the Vostok temperature and palaeorecords of global character; the multivariate analysis shows that generally >80% of the explained Vostok temperature variance is due to climate forcings of global significance (ice volume and greenhouse gases). Similarities between the Vostok and other palaeotemperature records further support this conclusion. For example, the Vostok temperature is well correlated with the surface temperature at the Indian Ocean site RC11-120[34] (with $r^2 = 0.66$ for the period before 110 kyr for which redating uncertainties are not critical). Replacing the Vostok temperature record by the RC 11-120 record as climate output in the multivariate analysis left a contribution of the greenhouse gases to the explained variance of ≥50%.

The approach used here suggests that greenhouse gases contributed ~2 °C to the global warming between glacial and interglacial periods. Estimating climate sensitivity then simply consists in dividing the estimated global warming associated with the greenhouse forcing by the corresponding ΔT_e (0.7 °C from glacial to interglacial) to get the net feedback factor. We stress

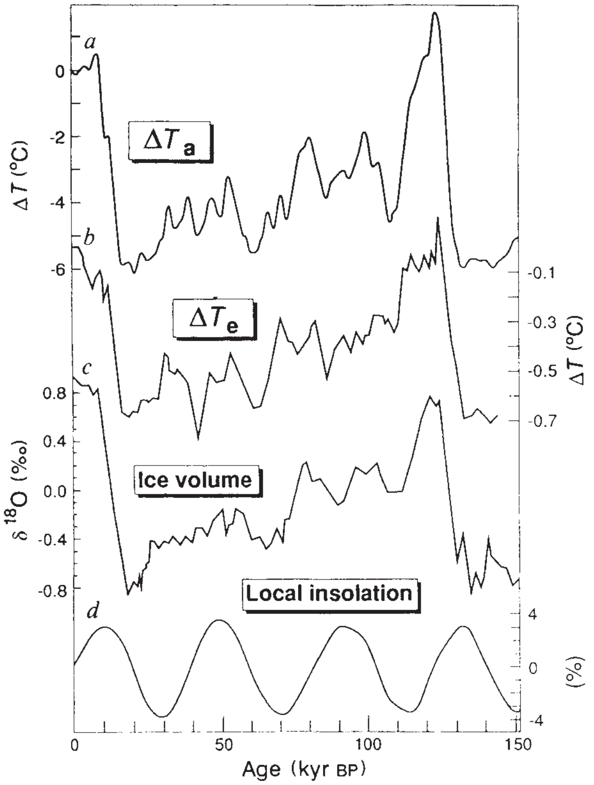

FIG. 3 Time series of the Vostok climate record and of climate forcings used in the multivariate analysis: *a*, atmospheric temperature change over Antarctica (Vostok record), ΔT_a; *b*, the direct greenhouse radiative forcing accounting for CO_2 and CH_4 variations, ΔT_e; *c*, $\delta^{18}O$ SPECMAP record taken as a proxy of the change in ice volume (normalized value from ref. 33); *d*, percentage change in total insolation at the Vostok latitude (78° S) during the entire year. Vostok curves (*a* and *b*) have been redated for the period before 110 kyr in such a way to put the Vostok temperature and the marine records in phase.

the independence of this method with respect to that based on GCM results. Although GCM simulations are useful in providing evidence that the regional results have a global significance, this use does not make our reasoning circular.

Second, it is of interest to compare the relative contribution of the greenhouse forcing as independently derived from the multivariate analysis and from GCMs. These results, only available for the GFDL and GISS models, indicate that a significant part of the glacial-interglacial warming may, on a global scale, be attributed to the greenhouse effect. In the Broccoli and Manabe experiments[49] the CO_2 increase from 200 to 300 p.p.m.v. and associated fast feedbacks, accounts for ~40% of the global glacial-interglacial warming. These experiments also showed that most of the remaining warming is due to the retreat of the Northern Hemisphere ice sheet in agreement with our multivariate analysis. Using the GISS model, Rind *et al.*[53] recently mentioned that an average warming of 2.5 °C would result from a CO_2 increase from 230 to 300 p.p.m.v. This corresponds to at least ~50% of the glacial-interglacial warming predicted by the GISS model for various simulations of the Last Glacial Maximum[6]. Such agreement supports the belief that the net effect of the fast feedback processes which amplify the direct greenhouse radiative forcing, or at least their combined effect, are realistically evaluated in these two models (or in models having similar climate sensitivity).

To the extent that atmospheric feedback processes are independent of the detailed causes of climate changes but governed by the overall climate response, the same feedback processes would operate now as during the Last Glacial Maximum[18]. Applying the net feedback factor of three deduced from ice-core data to the present climate may be done without further assumptions (at least within the limits of uncertainty of our method for estimating f). For greenhouse model experiments, climate sensitivity depends primarily on fast feedback processes. Analysis of ice-core results and palaeodata over a full glacial-interglacial cycle suggests that a warming induced by doubled CO_2 concentrations of 3-4 °C ($f \approx 3$) may be a realistic value which, although being in the middle of the range of values inferred from the GCM experiments (Box 1), corresponds to a relatively high climate sensitivity.

Concluding remarks

Ice-core records have been important in determining that climate and greenhouse-gas concentrations were intensely interactive. Palaeoclimate reconstructions are now recognized as an important element in climate studies because they: (1) allow us to assess the degree of natural variability and put observed changes in present climate in a broader perspective; (2) assist in understanding the causes and mechanisms of climate change (by relating climate and forcing factors, for example); (3) contribute to validating models through the comparison of outputs with empirical data sets; (4) may potentially be used in predicting geographical patterns of future climate based on past analogues.

We have approached the evaluation of climate sensitivity to greenhouse gases from the point of view of both the modellers and palaeoclimatologists. Despite the relative agreement of the two approaches there is still much research to be done and our assertion about the irrelevance of fully addressing the causes of glacial-interglacial cycles for estimating climate sensitivity must not be misunderstood. We attach the utmost importance to a full understanding of the physical, chemical and biological processes by which subtle changes in insolation are amplified to induce long-term changes in global climate. For this, a knowledge of the sequence of events and the exact timing of forcings and of the climate responses in various parts of the Earth system is essential. This includes acquisition of new well dated series from land, ocean and ice records to document fully the changes that have affected our planet over the last climate cycle; obtaining a globally averaged temperature record would improve our multivariate approach.

As far as ice cores are concerned, there is still a lot of information directly relevant to the interaction of greenhouse gases and climate buried in both Antarctic and Greenland polar ice caps. Obtaining Northern Hemisphere ice records of climate, greenhouse gases and other climate forcings over a full glacial-interglacial cycle is one important objective for a better understanding of the past complex interactions between Northern and Southern Hemispheres on this timescale. This objective is part of the GRIP (Greenland Ice Core Project) and GISP II (Greenland Ice Sheet Project) now being conducted in north central Greenland by European and American scientists. These drillings are expected to reach the bedrock (the ice thickness is 3.2 km) in 1992 and to cover the last climate cycle and hopefully more. These cores will allow further documentation of the rapid climate changes discussed here. With a snow accumulation higher than at Vostok they should also allow a better determination of the relative timing (phase lag) of climate and greenhouse forcing. Plans are also being developed for further drilling in Antarctica. Drilling at Vostok was interrupted at a depth of 2,546 m but Soviet drillers are starting a new hole with the aim of reaching the bedrock, that is, obtaining a 3.5-km ice core. Several other nations are also planning deep drilling in Antarctica (Australia, France, Japan, USA, for example). Scientists and operating agencies should develop a plan to organize an array of shallow, intermediate and deep ice cores ensuring both proper geographical coverage and coverage of relevant timescales at appropriate resolution. These plans should aim to obtain records of climate and forcing factors but also address other problems such as the interaction between climate and ice sheets and the role of the ice-sheet mass balance in sea level changes.

A strong interaction between data acquisition and interpretation and modelling of palaeoclimate and palaeoenvironment changes including the atmosphere, ocean, cryosphere, hydrosphere and land processes and various biogeochemical cycles is also needed. As well as GCMs, which produce steady-state simulations, a hierarchy of time-dependent models ranging from simple or even conceptual[46–48,61,62] to more complex two-dimensional models[63] can certainly provide information on how the various processes combine to produce observed climate changes. For this, understanding mechanisms that cause and are associated with rapid climate change is crucial. The effort in these directions must be well supported and the fact that the study of glacial-interglacial cycles constitutes one of the streams of the IGBP (International Geosphere Biosphere Program) project on Global Changes of the Past (PAGES) is very positive. Further studies of climate change will narrow the uncertainties in our present approach of estimating climate sensitivity. A modelling effort coupled with interpretation of palaeodata provides, along with the satellite data on radiative forcing[64], perhaps our best hope in the near future for predicting the impact of increasing greenhouse gases on climate. □

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ACKNOWLEDGEMENTS. We thank J. M. Barnola, J. Chappelaz and C. Genthon for discussions and J. Kutzbach, S. Manabe, J. Mitchell, D. Rind and S. Schneider for comments. We thank our Soviet colleagues in the Soviet-French collaboration on the Vostok core, particularly N. I. Barkov, V. M. Kotlyakov, Y. S. Korotkevich and V. M. Petrov.