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Holocene temperature variations inferred from six Antarctic ice cores

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ABSTRACT. We have reconstructed temperature changes over the past 15 000 years from ice-core data in Antarctica. We used measurements of the D/H isotope ratio in ice as a proxy of temperature for central sites (Vostok, Dome C and Komsomolskaya) as well as coastal sites (D47, D15 and D10). First, we examined the dating of each core and built up a common temporal framework for the ensemble of the data. Secondly, we addressed the problem of inferring small-amplitude temperature fluctuations from the isotope data, in the light of noise-generating mechanisms involved in snow deposition. Temperature was reconstructed so as to minimize distortion created by the sampling of ice cores in the field. The seven ice cores studied yield an average temperature curve which can be put in perspective with nearby paleoclimatic records. The early Holocene experienced climates warmer than today by 1–2°C. The late Holocene period shows more discernible, shorter-duration, temperature fluctuations, superimposed on a fairly stable “base-line” temperature.

INTRODUCTION

Although Antarctic ice cores have provided a wealth of paleoclimate data for the past 200 000 years, with the key advantage that ice cores are continuous paleoclimate recorders, little attention has been given as yet to the Holocene interval. During the past 10 000 years, climate proxies recorded in Antarctic ice all have fairly constant levels, with fluctuations typically one order of magnitude smaller than Pleistocene–Holocene changes. This is true for trace gases trapped in air bubbles, aerosols and dust content as well as for the concentration of stable isotopes D and ^{18}O in ice.

The primary objective of this paper is to discuss Holocene temperature changes from the isotope record of several ice cores. D/H and $^{18}\text{O}/^{16}\text{O}$ ratios in ice are linearly related to local temperature and this makes it possible to reconstruct the temperature on a quantitative

basis. However, the climatic signal contained in isotope data is embedded in noise related to snow-deposition processes. We need to separate the climatic information from the noise. We show in the following that only fluctuations of duration longer than ≈ 500 years are relevant for climate.

The ice cores we examined fall within two categories (Table 1; Fig. 1). “Deep ice cores” were drilled on the Antarctic plateau, where both the annual accumulation and ice motion are small. Such cores give access to very long climate records essentially exempt from ice-flow distortion. But, within the Holocene, deep ice cores offer only a limited time resolution. “Coastal ice cores” drilled along the margins of the ice sheet have greater accumulation rates and thus better time resolution. However, the regional ice flow is relatively fast and strongly affects the isotope profiles, as ice measured in the cores originates from upstream. Coastal ice cores integrate

Table 1

	Vostok 3G	Vostok 4G	Dome C	Komsomolskaya	D-47	D-15	D-10
Location (lat./long.)	72°28' S 106°48' E	72°28' S 106°48' E	74°39' S 124°10' E	74°05' S 97°29' E	67°23' S 154°03' E	66°42' S 139°46' E	66°42' S 139°46' E
Elevation (m)	3488	3488	3250	3499	1550	296	235
Temperature (°C)	−55.5	−55.5	−53.5	−52.6	−25.4	−16.2	−14.4
Accumulation (g cm ^{−2} a ^{−1})	2.3	2.3	3.4	6.4	26	43	13
Accumulation determination	Tambora (Legrand and Delmas, 1987)	Tambora (Legrand and Delmas, 1987)	Tambora (Legrand and Delmas, 1987)	β + stakes (pers. comm., M. Pourchet)	Tambora (pers. comm., M. Legrand)	β + stakes (Pettre and others, 1986)	β + stakes (Pettre and others, 1986)
Core length (m)	950	2546	930	850	870	345	310
Ice thickness (m)	≈3400	≈3400	≈3700	≈3000	≈1700	390	310
Surface-ice velocity (m s ^{−1})	≈5	≈5	≈0	–	26	≈10	≈10
Sampling	100 cm	100 cm in the Holocene	200 cm	200 cm every 5 m above 725 m	100 cm above 700 m	100 cm	1 m
	gap between 0 and 138 m	50 cm in the deglaciation	65 cm between 0 and 73 m	500 cm every 25 m below	50 cm below	10 cm above 50 m	Non- continuous
Glacial– Holocene boundary	260 m	260 m	400 m	≈680 m	Absent	Absent	≈200 m
Reference	Lorius and others, 1985	Jouzel and others, 1987b	Lorius and others, 1979; Jouzel and others, 1982	Nicolaiev and others, 1988; Ciais and others, 1992	This paper	This paper	Jouzel and Merlivat, 1977

past variations in climate over an extended area.

The dating of the deep ice cores from Vostok and Dome C has been based on ice-flow models over the last climatic cycle. For the Holocene, the dating needs to be examined in detail. Because Holocene ice represents a minor fraction of the total ice-sheet thickness, its chronology is dominated by accumulation rates, while such variables as bedrock topography or ice rheology are unimportant. The dating of coastal ice cores can also be addressed by models, which offer the advantage of being able to deconvolute the flow-related component from the climate-related component in the isotope profiles. Complications arise, however, at sites located on the ice-sheet margins because several distinct boundary conditions affect the ice dynamics. With the exception of D-10 and D-15, each core presented in this study is dated separately and is independent from other paleo-climate records.

1. ISOTOPE DATA

The isotope data of the deep and coastal ice cores are presented in Table 1 and Figure 2. Further information on the cores (drilling technique and core recovery) can be found in the references provided. The ice cores of D-47

and D-15, respectively, drilled in 1988 and 1983 by the French Polar Expeditions, are presented for the first time in this paper. The dating of these cores is discussed in section 2.2.

2. ICE-CORE CHRONOLOGY

2.1. Deep ice cores

Lack of seasonal variations in the isotope data at inland Antarctic sites

The snow deposition in Antarctica is much lower inland (≈5 g cm^{−2} year^{−1}) than on the coast (≈100 g cm^{−2} year^{−1}) (Bromwich, 1988). Therefore, the surface relief (sastrugi) and scouring by wind may erase the annual isotope signal. In addition, vapor-phase diffusion of HDO and H₂¹⁸O in the firn smooths short-length isotope events (Johnsen, 1983; Whillans and Grootes, 1985). There is no seasonal variation in the isotopes (nor in other species recorded in ice) at Vostok, Dome C or Komsomolskaya. The dating of these cores cannot rely on annual layer counting.

Ice-flow models within the Holocene period

The chronology of the deep ice cores is established by ice-flow models. The one-dimensional model of Dome C only accounts for vertical thinning under a constant strain rate.

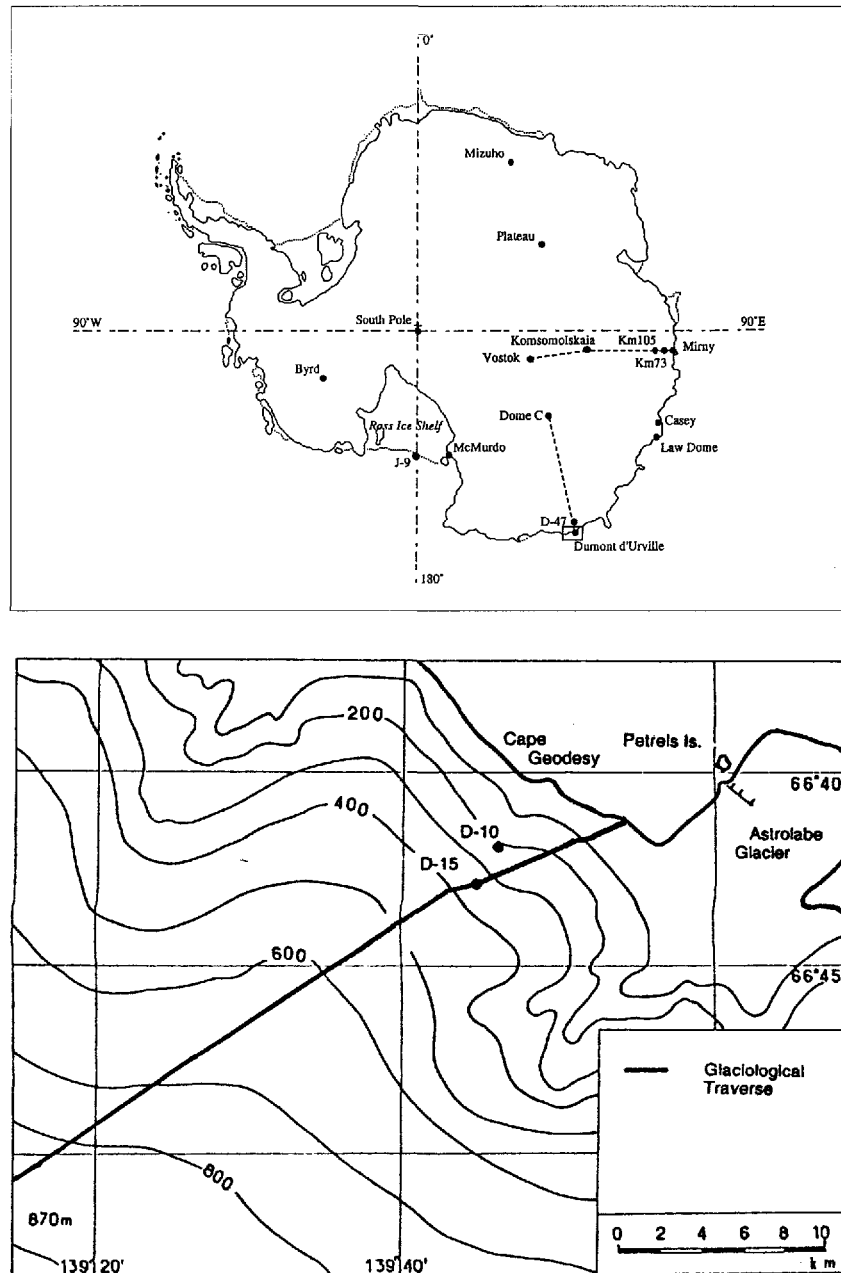


Fig. 1. Map of Antarctica with location of the drilling sites.

The two-dimensional model of Vostok accounts for horizontal flow (Lorius and others, 1985; Ritz, 1992). At Dome C and Vostok, because of the slow motion and great thickness of the ice sheet, the dating of the Holocene ice depends essentially on the accumulation rate. Currently, it is assumed that the amount of water vapor available for precipitation is controlled by the inversion temperature (T) (Lliboutry, 1965; Robin, 1977). The accumulation decreases during cold periods and increases during warm periods. A quantitative formulation of this argument is given by Equation (1) (Lorius and others, 1985):

$$b(z, t) = b(z, 0) \frac{\partial P_s / \partial T|_{T(t)}}{\partial P_s / \partial T|_{T(t=0)}}. \quad (1)$$

$b(z, t)$ = accumulation rate at depth z and time t ,
 $P_s(T)$ = saturation water-vapor pressure.

Using Equation (1), Jouzel and others (1989) dated

two ^{10}Be peaks measured both on Dome C and Vostok at 35 000 years BP and revised the maximum age of Dome C by 5000 years. They reached the conclusion that the dating over the last climatic cycle of both cores was approximately correct and only needed minor adjustments. When it comes to the Holocene period, where a more precise dating is needed, the validity of Equation (1) becomes questionable. Indeed, ice-flow models based on Equation (1) have yielded an artificial lag of 1500 years for the Vostok record with respect to Dome C within the deglaciation (Ciais and others, 1992). Recently, the Vostok dating has been refined using new accumulation measurements on the flowline upstream from the drilling site (Jouzel and others, 1993). In this study, we use the new EGT (extended glaciological time-scale) for Vostok which reduces the lag with Dome C during the deglaciation to 500 years.

Lacking reliable accumulation data for 1991, the earlier dating of Komsomolskaya was arbitrarily cor-

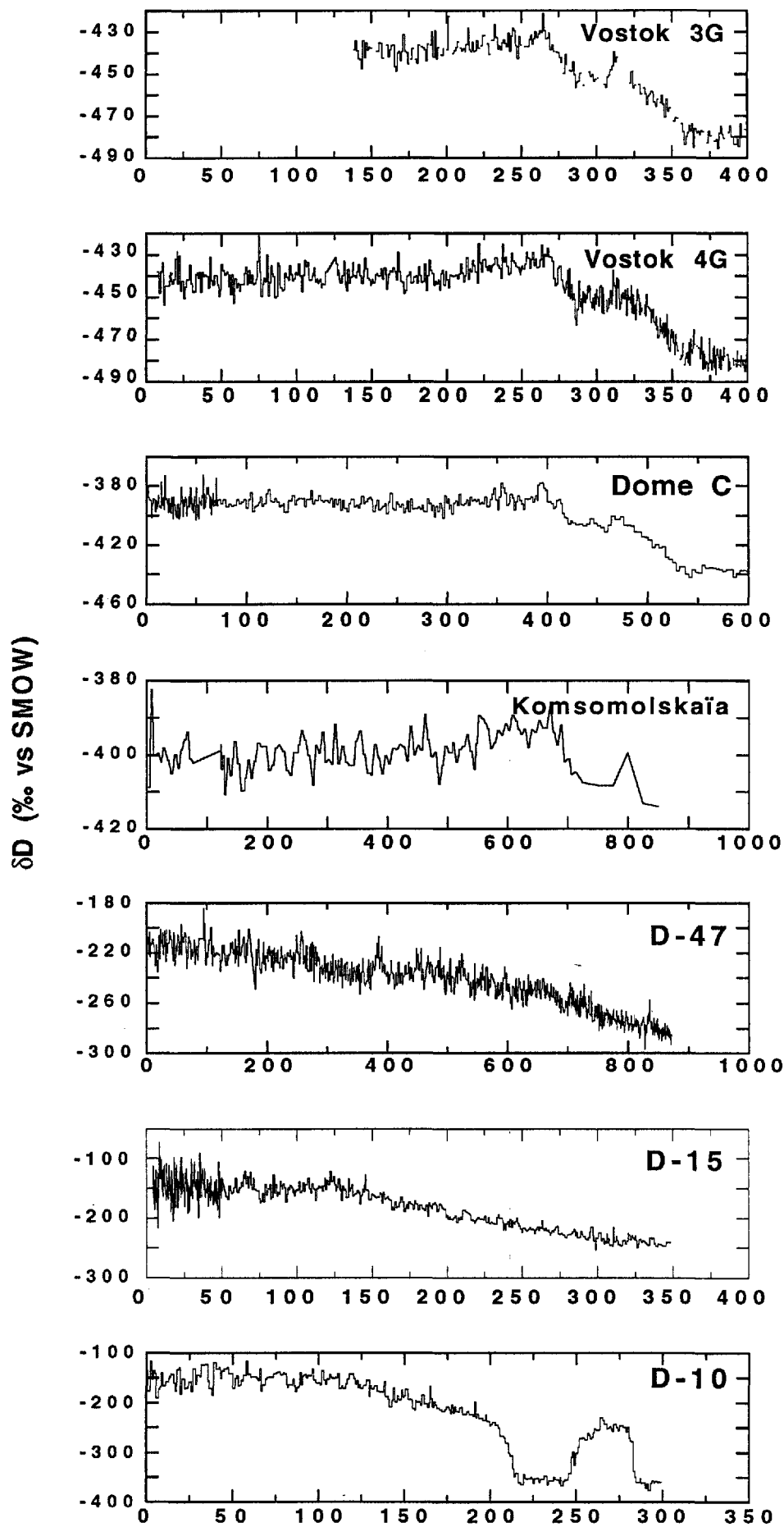


Fig. 2. Isotope profiles (vs depth) for seven Antarctic ice cores: coast ice cores are D-10, D-15 and D-47; deep ice cores are Vostok 3G and 4G, Dome C and Komsomolskaya. Note that only D-10 is drilled to the bedrock and shows the Glacial–Holocene climatic transition (the amplitude is more than three times higher than for Vostok or Dome C). Upper 400 m data are shown for Vostok and upper 600 m for Dome C.

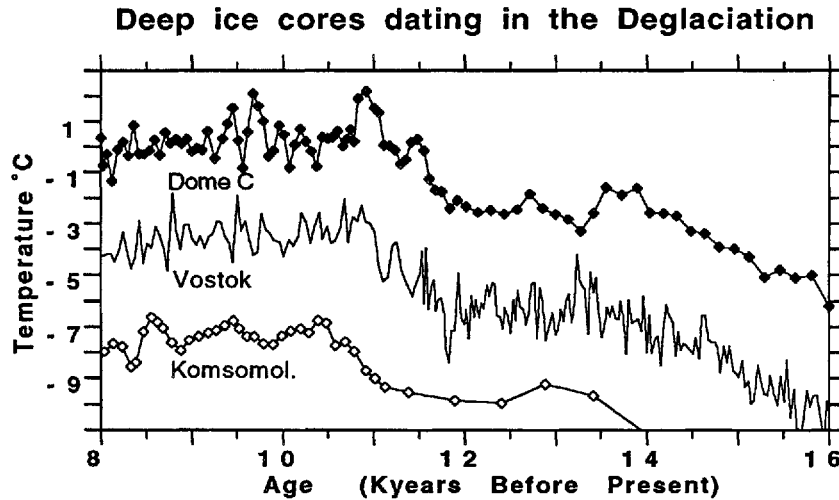


Fig. 3. Dome C, Vostok and Komsomolskaya isotope profiles between 8000 years BP and 18000 years BP. Each core is dated independently using a regional ice-flow modeling technique and modern accumulation-rate measurements (Table 1). Time-scale discrepancies do not exceed 500 years between the three profiles.

related with Dome C (Ciais and others, 1992). We use now an accumulation of $6.54 \text{ g cm}^{-2} \text{ year}^{-1}$ provided by M. Pourchet (β technique) in a Nye ice-flow model. The new Komsomolskaya chronology relates within 400 years to the chronology of Vostok (Fig. 3). The respective chronologies of Vostok, Dome C and Komsomolskaya are all in good agreement at the onset of the Holocene. Comparison of the δD profiles suggests an age difference that does not exceed 300 years at 11000 years BP.

2.2. Coast ice cores

D-47: ice-flow model and seasonal variations in δD

Dating of the D-47 ice core is complicated by the fast motion of the ice (decreasing trend of δD with depth) shown in Figure 2. The absence of characteristic discontinuity in δD shows that the core does not reach into the Pleistocene. The regional flow-boundary conditions are (1) the accumulation rate upstream of the core site, (2) the divergence of the flowlines, (3) the topography of the bedrock and of the ice-sheet surface, and (4) the ice rheology. We have adapted the two-dimensional model for Vostok (Ritz, 1992) to the flowline of D-47. Ice trajectories are back-calculated until the "origin" of the ice (upstream location where the ice was deposited) is determined. We obtain in that manner several age- Z curves that correspond to different assumptions for the boundary conditions. These curves can be very different. A detailed sensitivity study shows that boundary conditions (1) and (2) play the major role in the calculation of the age- Z curve (Ciais, 1991). This is probably because the ice of D-47 is distant enough from the bedrock for basal-flow complications not to occur. We used regional accumulations from the well-documented data set of β measurements by Petre and others (1986). We estimated the regional flow divergence from the surface elevations (Lorius and others, 1967; Drewry, 1983) and hypothesized no change in the divergence with depth.

Still, the uncertainty of the model boundary conditions yields age- Z curves that differ by almost a factor of 2 (Fig. 4). An appropriate constraint is raised by high-

resolution δD measurements on selected parts at discrete depths along the core (every 100 m). Each high-resolution δD series is a 3 m long ice segment sampled every 2 cm. From the preserved isotopic cycles (seasonal variations), we calculated a mean annual-layer thickness using the maximum entropy spectral-analysis method (Johnsen, 1978). Annual-layer thicknesses are used to constrain ice-flow model calculations and therefore allow us to select the exact age- Z relationship with an error of approximately 10% (Fig. 4). The age of the D-47 ice core is determined as 7500 years BP.

D-15 and D-10: ice-flow model and stratigraphic correlation

At D-10 and D-15, the reduced thickness of the ice sheet makes the bedrock topography and the rheology as important as the accumulation and divergence in determining the chronology. Also, large variations in the ice thickness could have occurred during the Holocene. For D-10, we forced the age- Z calculated by our two-dimensional regional model to put the termination of the δD discontinuity at 11000 years BP (according to the Dome C profile). D-15 does not reach into the Glacial. At the bottom of the core, δD (-245‰) is equal to the value observed for the onset of the Holocene at D-10. Because D-15 and D-10 are located 2 km apart and the present-day ice velocity is 10 m year^{-1} , we dated D-15 by fixing the bottom of the core to the onset of the Holocene.

3. ISOTOPE-TEMPERATURE-TRANSFER FUNCTION

3.1. Isotope-temperature relationship

Relation between δ and T at present and in the past

In cold ice-sheet regions, the average δ of surface snow is a linear function of local temperature. A regression of the data collected in Antarctica gives a value of $6.04\text{‰ } ^\circ\text{C}^{-1}$ for the δD - T gradient (Lorius and Merlivat, 1977; Qin Dahe and others, 1994). The observed decrease of δ with T in polar snowfall is due to the cumulative effect of

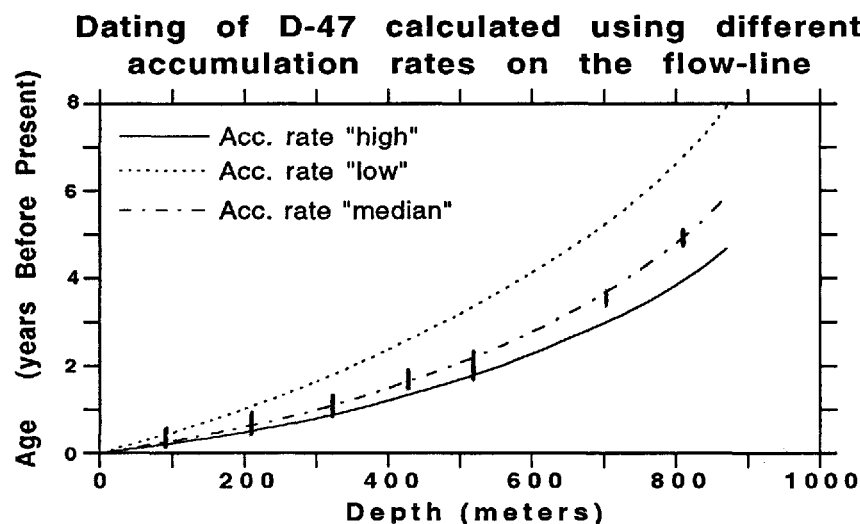


Fig. 4. Ice-flow model time-scales calculated for the D-47 ice core under different accumulation rates parameterization. Accumulation scenarios are based on measurements made on the D-47 flowline by Pettre and others (1986) ("median" case), idem plus standard deviation ("high" case), idem minus standard deviation ("low" case). Note the large sensitivity of the model to the accumulation-rate parameter. Vertical bars mark the isotopic constraints on the ages (see section 2.2).

condensation events that deplete the water vapor above the ice sheet in isotopes (Dansgaard, 1964). The stability of the δ - T gradient in the past can be tested using general circulation models where water isotopes can be run under realistic climate-boundary conditions (Joussau and others, 1985; Jouzel and others, 1987a). Joussau and Jouzel (1993) recently found that the δ - T gradient in Antarctica during the last glacial maximum was not very different from today. Dealing with stable climates during the Holocene, it is reasonable to use the observed present δ T gradient to reconstruct past temperatures. Complications arise in coastal regions where the δ - T relationship only prevails above 1000 m elevation (Lorius and Merlivat, 1977). D-47 is located above 1000 m and δ D along the core records temperature. For D-15 and D-10, one can see in Figure 2 that only the ice from upstream regions above 1000 m depth (now at a depth below 120 m in D-15) shows a characteristic decrease of δ D with depth. After our dating of the core, surface ice for which δ D is a dubious proxy of temperature is no older than 700 years. In this study, we do not discuss the climate of the last millennium but the fluctuations during the last 10 000 years, and the core of D-15 is a reliable indicator of temperature over this time period.

Biases in isotope records

We examine three systematic biases of the temperature information contained in the isotopes. (1) Changes in ice thickness and ice flow cause a displacement of the origin of the ice which may affect the isotope record. Over central Antarctica, Holocene variations in the thickness of ice are probably negligible (Vostok elevation did not change by more than 100 m over the last climatic cycle). Over coastal Antarctica, the ice-flow model we used to date D-47 was in agreement with the isotopic constraints on the dating only if elevation changes were less than 200 m (Ciais, 1991). At the edge of the ice sheet, the Holocene isotope records of D-15 and D-10 are certainly affected by changes in ice thickness (Groote and Stuiver, 1987). We attribute millennium-scale variations in δ D to temperature

changes, whereas the global trend of δ D with depth integrates past changes in thickness and ice flow. (2) Changes in temperature at the moisture source influence δ of Antarctic snowfall. A 1°C increase in SST has the same effect on δ D over central Antarctica as a local temperature decrease of 0.7°C (Koster and others, 1992). A systematic correction would have to account for past variations in the SST at the moisture source. However, central Antarctic moisture likely originates from sub-tropical latitudes (Petit and others, 1991; Koster and others, 1992) where SST shows little change in the past (CLIMAP, 1981). No SST correction in the past is applied to coastal cores. (3) Changes in the isotope composition of the surface ocean offset δ in Antarctic precipitation. This is corrected using the marine $\delta^{18}\text{O}_{\text{sea}}$ record (Labeyrie and others, 1987) scaled by a factor of 8 for deuterium (Craig and Gordon, 1965). Over the past 10 000 years, this corrects the Antarctic temperature by 0.2°C.

3.2. Noise in the isotope-temperature relationship

δ T noise and δ noise

There is a random dispersion on the δ - T transfer function that we call δ - T noise. Its variance, $\text{Var}(N_{\delta-T}(t))$, represents a threshold below which a fluctuation in δ is not significant in terms of temperature. Processes entering in δ - T noise are listed in Table 2 (after Robin (1979)). Lacking micro-meteorological data and surface-relief data at the drilling sites, it is very difficult to estimate directly $\text{Var}(N_{\delta-T}(t))$ from the variance of each process. One possibility is to examine two parallel cores at the same site. The isotope profiles measured on two such cores are not perfectly correlated, which defines a δ noise, $N_{\delta}(t)$. Generally, $\text{Var}(N_{\delta-T}(t))$ is larger than $\text{Var}(N_{\delta}(t))$, because it contains more processes (see Table 2). We have assumed that these extra processes entering in $\text{Var}(N_{\delta-T}(t))$ are much less important than the spatial irregularity in the deposition, given the very low accumulation observed in central Antarctica (Bromwich, 1988). Therefore, we make the approximation:

Table 2. Process entering in the δ noise and δ - T noise

δ noise Noise on δ inferred from two adjacent cores	δ - T noise Noise on the δ - T relationship
Irregular deposition	Irregular desposition
Precipitation scattering + wind effects	Precipitation scattering + wind effects
Isotope smoothing	Isotope smoothing
Summer melting	Summer melting
Core processing	Core processing
	Variations in the winter/ summer deposition
	Variations of T not recorded in δ such as Inversion strength + storms frequency
	Long-term variations in the δ - T relationship such as Moisture origin + air-masses trajectories

$$\text{Var}(N_{\delta-T}(t)) \approx \text{Var}(N_{\delta}(t)). \tag{2}$$

Isotopic noise at Vostok

The drilling of the adjacent cores 3G and 4G makes it possible to estimate directly $\text{Var}(N(t))$. To do so, we decomposed δ_i of each core as the sum of a climatic signal $\delta_i^c(t)$, common to both cores, plus random noise (Fisher and others, 1985):

$$\delta_i(t) = \delta_i^c(t) + N_i(t) \quad i = 3G, 4G. \tag{3}$$

Both cores, within a distance of 100 m, have the same deposition conditions, which yields:

$$\text{Var}(N_{3G}(t)) = \text{Var}(N_{4G}(t))$$

and from Equation (3) $\text{Var}(\delta_{3G}(t)) = \text{Var}(\delta_{4G}(t)). \tag{4}$

From the definition of the 3G-4G correlation coefficient (R_{3G-4G}), we then obtain

$$\text{Var}(N(t)) = (1 - R_{3G-4G})\text{Var}(\delta(t)). \tag{5}$$

Estimating $\text{Var}(N(t))$ from Equation (5) over the interval 0-15 000 years BP, we obtained a correlation that was not statistically significant, yielding a noise variance of 15‰ in δD , a value larger than the expected “climatic” variance during the Holocene (the deglaciation represents about 40‰). This shows that the short-term variability of temperature cannot be extracted from the noise on δ . Therefore, we must choose an arbitrary time-scale for the T fluctuations we want to study. Table 3 shows that the signal/noise ratio is larger than 1 only when periods shorter than 200 years are cut off (one core sample is ≈ 50 years). To study Holocene temperature changes, we adopted a cut-off period of 500 years (signal/noise = 4). This focuses our discussion on events typically of the same duration as the Little Ice Age. To translate this in terms of an error bar on the reconstructed temperature, we considered the δ - T relation (corrected from ocean composition):

$$T = \frac{\delta - 51 + \Delta\delta_{\text{sea}}}{6.04}. \tag{6}$$

$\Delta\delta_{\text{sea}}$ is the difference between the past ocean $\delta^{18}\text{O}$ and present-day SMOW. From the experimental error, $\text{Var}(\Delta\delta_{\text{sea}}) = 0.1\text{‰}$ (0.8‰ in δD). Equation (5) gives a noise variance of 2.2‰.

Table 3. Noise on time series of δD at Vostok

Low-pass cut-off period	$\sigma\delta =$ δD standard deviation			$\sigma N\delta =$ δ noise standard deviation		$\sigma NT =$ T noise standard deviation		Signal/noise ratio
years	‰			‰		°C		
	Core 3G	Core 4G	R_{3G-4G}					
50*	4.65	4.72	0.37	3.71		0.61		0.6
100	4.19	3.99	0.53	2.81		0.47		1.2
200	3.93	3.79	0.67	2.21		0.36		2.0
300	3.44	3.44	0.73	1.78		0.30		2.7
400	3.30	3.36	0.77	1.60		0.26		3.4
500	3.21	3.30	0.79	1.49		0.24		3.8
1000	2.55	2.98	0.82	1.17		0.19		4.6

* Original data sampling.

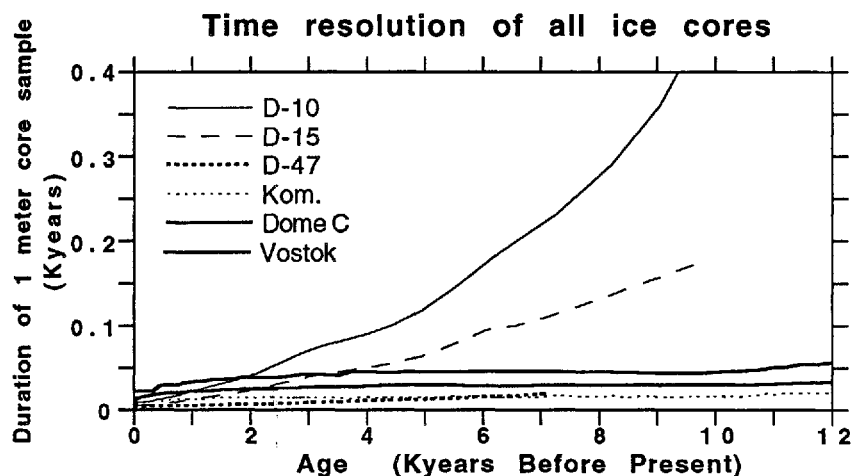


Fig. 5. Average number of years contained in a 1 m long sample for six Antarctic ice cores (i.e. derivative $\partial \text{age} / \partial z$). The smaller the number of years, the better the time resolution of the core. Note that coast ice cores have a better time resolution during the late Holocene period and deep ice cores during the early Holocene.

$$\begin{aligned} \text{Var}(N_T(t)) &= \frac{\text{Var}(N_\delta(t)) + \text{Var}(N_{\Delta\delta_{\text{sea}}}(t))}{6.04^2} \\ &= \frac{2.2 + 0.8}{6.04^2} = 0.082^\circ\text{C}^2. \end{aligned} \quad (7)$$

This turns into a standard deviation of 0.25°C which represents physically the threshold below which a variation in temperature derived from the isotope record is not significant when dealing with isotope data filtered at 500 years. This value is intuitively an upper estimate of the isotopic noise in Antarctica because Vostok has the lowest accumulation. At Dome C, based on the dispersion of δD on 20 surface cores, Ciais (1991) estimated an isotopic noise of 0.3°C (data filtered at 500 years). In coastal regions, one would expect the snow deposition to show less scatter than inland, because of larger accumulations. However, this may be counterbalanced by substantial wind scouring of freshly deposited snow (Wendler, 1989; Parish and Wendler, 1991). From the δD variability of surface cores, we provided a rough estimate of the noise variance at D-47 of $0.2\text{--}0.3^\circ\text{C}$ (Ciais, 1991).

4. HOLOCENE ANTARCTIC CLIMATE

4.1. Paleo-temperature reconstruction

Distortion due to ice-core sampling

The core-sampling increment for isotopes is constant with depth. Therefore, because of the vertical ice thinning, one sample represents more years at greater depths. The number of years contained in a given sample, which characterizes the time resolution of the core, is calculated from the age Z function, $A(Z)$, and the sampling increment, l , using Equation (9):

$$\Delta t = l \times dA/dZ. \quad (8)$$

Δt as a function of Z is shown for all cores in Figure 5. Deep ice cores have an almost constant Δt , because Holocene ice is high above the bedrock and does not undergo strong vertical thinning. In contrast, coastal ice

cores are characterized by a strong increase in Δt with depth. In the case of D-10, for instance, a 1 m long sample contains ≈ 2 years near the surface and ≈ 500 years at $Z = 200$ m. To present our records in a manner that is not distorted by ice-flow-induced thinning, the data were digitized to create an annually sampled time series, which was filtered to remove periods of less than the maximum value of Δt along the core. A cut-off period of 500 years was found to satisfy both requirements of the noise and of the sampling (Fig. 6).

4.2. Antarctic temperature trends

Average temperature curve

We observed that intervals characterized by colder temperatures can be correlated in most of the curves and we derived an ‘‘average temperature’’ curve. Temperature fluctuations do not exceed $1\text{--}2^\circ\text{C}$ in any

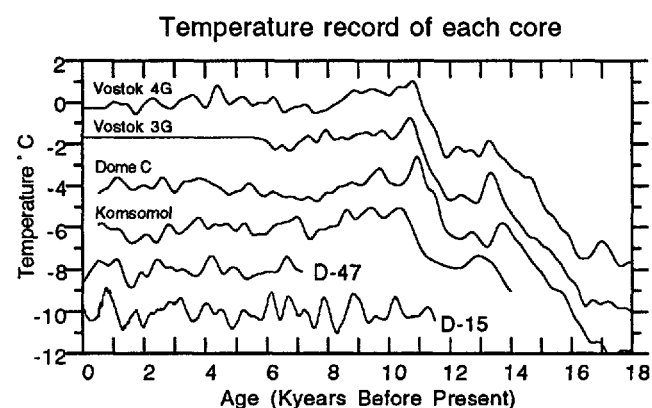


Fig. 6. Temperature reconstructed from the isotope profile of six Antarctic ice cores. Isotope data have been filtered in order to remove all periods shorter than 500 years to minimize the noise on the data. Temperature variations are expressed as the departure from the mean value over the past 5000 years. A constant offset has been added to each profile for display clarity. The D-10 reconstructed temperature shows fluctuations of magnitude $3\text{--}4^\circ\text{C}$ as a consequence of discontinuous sampling of the core and is not included.

core (large variations at D-10 are likely to be due to the non-continuous sampling of the core and D-10 is excluded from the average). From Figure 5, it is seen that the climate of the early Holocene is better resolved in the deep ice cores, whereas the late Holocene is better resolved in coastal ice cores. We have attributed to each time series a weight w_i proportional to its time resolution Δt_i as defined by Equation (8)

$$w_i = (\Delta t_i)^{-1} / \sum_{j=1}^6 (\Delta t_j)^{-1}. \quad (9)$$

The weighted average temperature, $\bar{T}(t)$, is then expressed by:

$$\bar{T}(t) = \sum_{j=1}^6 w_j T_j(t). \quad (10)$$

$\bar{T}(t)$ is shown in Figure 7. Considering a weighted average in place of a simple average does not change the timing of the T fluctuations, although it has the effect of changing their amplitude in such a manner that $\bar{T}(t)$ is dominated by Vostok and Dome C in the early Holocene, and by D-47 and D-15 in the late Holocene. Maximum temperatures (1°C warmer than today) prevail between 11 000 and 8000 years BP, defining an early Holocene optimum (Ciais and others, 1992). The most significant cooling is from 8000 to 5500 years BP. The late Holocene is characterized by a stable average temperature. A noticeably warmer climate episode occurred around 4000 years BP. An important cooling occurred between 2000 and 1000 years BP.

Comparison with the Byrd ice-core record, West Antarctica

The Byrd $\delta^{18}\text{O}$ (see for instance, Johnsen and others, 1972) shows a rapid increase of 1.5‰ at 4500 years BP, which is absent from the East Antarctic cores. Assuming a $\delta^{18}\text{O}$ - T gradient of $0.8\text{‰ }^\circ\text{C}^{-1}$, this increase translates into a regional warming of 2°C . A change in the ice thickness (thinning of 200 m) or in the moisture supply to West Antarctica is more likely to have occurred.

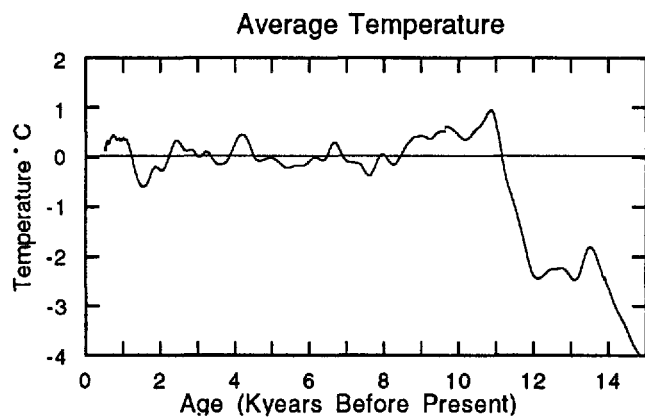


Fig. 7. Average temperature curve from six Antarctic ice cores (Fig. 6). Core D-10 is not included. A different weight has been given to each record according to the quality of its time resolution (Fig. 5). Temperature is expressed as the departure from the mean over the past 5000 years.

Early Holocene Antarctic climate and “Holocene optimum”

Evidence for maximum temperatures during the first 2000 years of the Holocene contradicts the commonly admitted picture of a worldwide mid-Holocene optimum (Lamb, 1977; Folland and others, 1990). If higher δD in Antarctica did not reflect warmer temperatures, then one would have to look for changes in (1) elevation and ice flow, (2) moisture source and (3) source composition. Our curves are already corrected from (3). Regarding point (2), isotopic models demonstrate that SST influences δD according to (Koster and others, 1992)

$$\Delta\delta = 6\Delta T_p - 4\Delta T_e. \quad (11)$$

T_e = sea-surface temperature, T_p = site temperature.

The SST in the Southern Ocean were warmer than today during the early Holocene, as inferred from the marine record (Hays and others, 1976; Labracherie and others, 1989). This would yield even warmer Antarctic temperatures in the early Holocene. Point (1) can be discussed qualitatively, provided that the central EAIS did not change by more than ± 200 m during the climatic transition (Huybrecht, 1990; Martinerie, 1990). Greater δD values during the early Holocene could be attributed to a change in elevation if the EAIS remains thinner than today up to 8000 years BP, and then thickens rapidly to reach its present characteristics. This would translate into a change in accumulation of 10 cm year^{-1} at around 8000 years BP, which is not compatible with the isotope data (Equation (1)). A step further would be to compare our Antarctic temperature record with the climate of high southern latitudes over the past 15 000 years. There are terrestrial and marine ^{14}C dated paleoclimate records in South America, New Zealand, Tasmania and in the Southern Ocean.

CONCLUSIONS

In the six ice cores examined in this work, the Holocene period is characterized by fairly constant δD values. Because the climate signal contained in the isotopes is embedded in a strong noise (irregular snow deposition), the data were filtered to minimize the signal/noise ratio. The average temperature curve shows warmer conditions than at present during the Holocene interval (11 000 and 9000 years BP) by about $1\text{--}2^\circ\text{C}$. The late Holocene period is characterized by shorter fluctuations, the most noticeable being a warm peak around 4500 years BP and a cold interval around 2000 years BP.

We have restricted the present study to the isotopes, a proxy primarily related to local changes. Variables of more global character (gases and chemical impurities) are also reconstructed from ice cores. They relate to past atmospheric composition and circulation, as well as to the emission by sources. Within the Holocene interval, the concentration of dust is fairly constant but CO_2 and CH_4 appear to be higher during the early Holocene, concomitant with the local maximum in temperature. A better knowledge of the Holocene from the ice-core record can be placed in perspective with an abundance of paleoclimate reconstructions worldwide. Hopefully, this

will help to explain why climate conditions have been so remarkably equable over the past 10 000 years in contrast to the large and rapid changes observed during Glacial and late-Glacial periods.

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