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Validity of the temperature reconstruction from water isotopes in ice cores

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Abstract. Well-documented present-day distributions of stable water isotopes (HDO and H218O) show the existence, in middle and high latitudes, of a linear relationship between the mean annual isotope content of precipitation (δD and δ18O) and the mean annual temperature at the precipitation site. Paleoclimatologists have used this relationship, which is particularly well obeyed over Greenland and Antarctica, to infer paleotemperatures from ice core data. There is, however, growing evidence that spatial and temporal isotope/surface temperature slopes differ, thus complicating the use of stable water isotopes as paleothermometers. In this paper we review empirical estimates of temporal slopes in polar regions and relevant information that can be inferred from isotope models: simple, Rayleigh-type distillation models and (particularly over Greenland) general circulation models (GCMs) fitted with isotope tracer diagnostics. Empirical estimates of temporal slopes appear consistently lower than present-day spatial slopes and are dependent on the timescale considered. This difference is most probably due to changes in the evaporative origins of moisture, changes in the seasonality of the precipitation, changes in the strength of the inversion layer, or some combination of these changes. Isotope models have not yet been used to evaluate the relative influences of these different factors. The apparent disagreement in the temporal and spatial slopes clearly makes calibrating the isotope paleothermometer difficult. Nevertheless, the use of a (calibrated) isotope paleothermometer appears justified; empirical estimates and most (though not all) GCM results support the practice of interpreting ice core isotope records in terms of local temperature changes.

1. Introduction

Important relationships have been uncovered between the global distributions of δD and δ18O in modern precipitation and certain climatic variables [Craig, 1961; Dansgaard, 1964]. (The concentrations of HDO and H218O, the two stable isotopic forms of water, are expressed in per mil units with respect to standard mean ocean water (SMOW) as δD or δ18O.) Of primary interest to the paleoclimatologist is the linear relationship between annual values of δD and δ18O and mean annual temperature at the precipitation site that is observed at middle and high latitudes. In order to use the isotope signal as a paleothermometer, the present-day spatial isotope/surface temperature relationship δ = aTs + b defined over a certain region (δ stands for either δD or δ18O of the precipitation and Ts for the surface temperature; the isotope/surface temperature slope is a = dδ/dTs) is generally assumed to hold in time throughout the region, i.e., it is assumed that the spatial and temporal slopes are similar (hereafter "temporal" applies to the relationship describing the variation of isotopic contents with temperature through different climates over time, at a single geographic location). A so-called "modern analogue method" is thus used, similar to that adopted in most other approaches for reconstructing paleoclimates. Of course, the fact that present-day isotope concentrations and local temperature are strongly correlated, as illustrated in Figure 1 for Greenland and Antarctica, does not validate this critical assumption. Such factors as the evaporative origin and the seasonality of precipitation can also affect δD and δ18O, and if these factors change markedly under different climates, the accuracy of the isotope paleothermometer is reduced. The possibility of a difference between the temporal and spatial δTs relationships must therefore be seriously examined. There is, indeed, growing evidence that a difference between temporal and spatial δTs relationships does exist and appears to be time dependent, thus complicating the use of water stable isotopes as a paleothermometer.

The temporal δTs relationships can be tested in several ways. An empirical calibration can be inferred by direct comparison of the stable isotope composition of the upper snow
layers and the rather short instrumental temperature records [Dansgaard et al., 1975; Jouzel et al., 1983; Robin, 1983; Morgan, 1985; Aristarain et al., 1986; Peel et al., 1988; Peel, 1992; Rozanski et al., 1992; Shuman et al., 1995; White et al., this issue]. Much longer but lower-resolution comparisons can be carried out using independent, colocated paleotemperature estimates, which may be inferred from temperature profiles measured in ice sheets [Dahl-Jensen and Johnsen, 1986; Cuffey et al., 1994, 1995; Johnsen et al., 1995] and boreholes [Beltrami and Taylor, 1995], or from groundwaters which allow estimates of changes both in temperature (from the noble gas content) and in the water isotopic contents [Stute et al., 1992]. An alternative approach is the modeling of isotope distributions in precipitation, made possible by our knowledge of the physical laws governing isotopic fractionation at the various stages of the atmospheric water cycle and which in fact allows a direct comparison between the spatial and temporal δ/Ts relationships, based on model simulations for different climatic periods. Atmospheric general circulation models (GCMs) are well suited for this approach.

In this paper we review what can be inferred about the temporal δ/Ts relationships, first from empirical estimates and second from the modeling approach. The focus is on polar areas and more specifically on Greenland where the Greenland Ice Core Project (GRIP) and the Greenland Ice Sheet Project 2 (GISP2) drillings have allowed such empirical estimates to be obtained from paleothermometry [Cuffey et al., 1995; Johnsen et al., 1995] and have motivated dedicated published [Charles et al., 1994, 1995] and unpublished (shortly discussed hereafter) modeling experiments. We then examine the reasons why the two approaches may differ, as appears to be the case for central Greenland.

2. Empirical Estimates of the δ/Ts Relationships

Ideally, to estimate the δ/Ts relationship, records of both surface temperature and isotopic contents of the precipitation should span over a relatively long period at the same site. South Pole station, where temperatures are available since the International Geophysical Year (IGY) in 1958 for both surface and atmospheric levels, is the only polar site fulfilling this requirement. From a comparison with well-dated isotopic profiles covering the 1957–1978 period, mean annual and maximum deuterium values were shown to be correlated with the corresponding mean annual and summer temperatures, but, on the other hand, winter temperature and deuterium minima are poorly correlated [Jouzel et al., 1983].

To extend the approach to other sites, indirect estimates of the temperature variation at those sites must be used. Five different approaches have been employed that apply to different timescales: (1) use of temperature recorded in the vicinity of the sampling site allowed extension of the comparison over the period of instrumental observations; (2) high-resolution isotope profiles and automatic weather stations (AWS) and satellite microwave brightness temperature trends have been compared at the GISP2 site for a 3-year period spanning from 1987 to 1990; (3) for various periods of the Holocene, temperature shifts can be estimated from the changing percentage of melt layers in ice cores; (4) the isotopic thermometer can be calibrated against the borehole temperature profile over a wide range of timescales (centuries to tens of millennia); and (5) additional information can be retrieved from the temperature-dependent change in snow accumulation at a given location.

2.1. Comparison of δ and Instrumental Temperature Records

Many processes control the isotopic composition of individual snowfall events. These are factors that directly condition the isotopic composition of an air mass as it moves from the source region toward the deposition site, and include the influence of changing atmospheric circulation which also determines the effective location of the moisture source. These factors tend to dominate the regional mean stable isotope
distribution, but also dominate the low-frequency climate changes observed at individual sites. Superimposed on this is the influence of a series of statistically controlled processes which in the short term introduce high-frequency noise mainly due to local spatial variability in the snowfall deposition, but can also introduce significant biasing as snowfall does not occur randomly. Also potentially important is the fact that the temperature during precipitation events may not be representative of the annual mean temperature. The effects of some of these factors can be reduced by smoothing, while others present a systematic problem.

Local variations in the spatial pattern of snow accumulation lead to short-term differences in the time series of annually averaged parameters at adjacent drilling sites, the so-called "deposition noise" (the uncorrelated variance of two adjacent $\delta$ time series), which has been estimated for several sites in Greenland [Robin, 1983; Fisher et al., 1985] and in Antarctica [Benoit et al., 1982; Peel, 1992], using data from duplicate cores. It is therefore possible to obtain a crude estimate of the length of averaging needed to reduce the deposition noise to a level needed to detect a given magnitude of "climatic" signal. For Antarctica (where cores have been obtained from sites ranging over more than a factor of 20 in accumulation rate), the $\delta$ noise is strongly inversely related to accumulation rate; also the amplitude of the annual cycle is a major factor [Fisher et al., 1985], increasing with increasing continentality in Antarctica. The situation is rather less clear-cut in Greenland, although the observed noise levels appear broadly in line with those observed at sites with comparable accumulation rate in Antarctica. In practice, this means that high levels of correlation between unsmoothed replicate annual time series can be achieved in coastal areas of Antarctica [Robin, 1983; Peel et al., 1988; Grootes et al., 1990; Peel, 1992] and in central Greenland [Fisher et al., 1985; Clausen et al., 1988], including Summit [Steffensen et al., 1997], where the accumulation rate is greater than about 20 cm water/yr. A further source of high-frequency noise, "definition noise," arises because the stratigraphic horizons separating successive snow accumulation years do not fall on a constant calendar date [Peel, 1992]. For the higher-accumulation sites, where only 3–5 years averaging is needed to reduce the noise level to the equivalent of $-0.2^\circ C$ temperature shift, it is possible to make a direct comparison of the recent isotopic record with mean annual temperatures recorded in the vicinity, in an effort to validate directly the theoretically determined isotope/surface temperature gradients suitable to interpret time trends. Comparative studies have been carried out at several sites in Antarctica including Law Dome [Morgan, 1985], in the Antarctic Peninsula [Aristarain et al., 1986; Peel et al., 1988; Peel, 1992], and in Greenland, at Milcent [Robin, 1983], Crete [Dansgaard et al., 1975], and in the GRIP and GISP2 area [White et al., this issue].

Unfortunately, such studies can only provide a limited test; the records are often short: In Antarctica there are few data prior to IGY of 1957–1958. In addition, the nearest available weather stations are usually sited near the coast, and typically lie several hundred kilometers away from sites suitable for ice core drilling. In the Antarctic Peninsula region, temperatures recorded at weather stations lying along the east and west coasts correlate very weakly [Peel, 1992; Jones et al., 1993]. Many of the locations suitable for deeper drilling are sited near the ice divide of the peninsula, which is affected by climatic influences from both sides. Consequently, the levels of correlation achieved between time series of $\delta$ and $T_s$ tend to be considerably weaker than the spatial correlation of mean annual parameters [Robin, 1983]. In the Antarctic Peninsula, where cores from five locations across the region have been compared with temperature records from several stations that have operated since IGY, the isotopic data typically account for about 30% of the variance in the unsmoothed annual temperature series, although this rises to $-50\%$ for 5-year smoothed series, where the effects of any small dating error, together with deposition noise in the ice core data, are minimized. This level of correlation is similar ($r^2 = 0.37$, $r$ being the correlation coefficient) to that obtained between Casey Station temperatures and a site on Law Dome some 30 km distant [Budd and Morgan, 1977]. In Greenland, there are significant differences between temperature records from the east coast and the west coast [Dansgaard et al., 1975] which are still evident in 30-year smoothed records. The isotopic records from the interior of Greenland do not appear to follow consistently the temperature variations recorded at either east coast or west coast stations; e.g., annual $\delta$ values at Milcent [Robin, 1983] were significantly correlated ($r^2 = 0.32$) at the 5% level with $T_s$ at Jacobshaven for the period 1904–1933, but not during the adjacent 30-year periods. This behavior may reflect the alternating modes of the North Atlantic Oscillation, which is associated with temperature anomalies of opposite sign to the east and west of the crest [Barlow et al., 1993]. White et al. [this issue] produced a stacked isotopic record from the averaging of six individual annually resolved isotopic records obtained in the GRIP/GISP2 area which versus the average coastal temperature leads to a slope of 0.29%°C, which is roughly half that of the spatial slope of 0.67%°C [Dansgaard, 1964; Johnsen et al., 1989; see also Figure 1]. These authors noted that this difference may be real or perhaps due to the long distance and altitude difference between the ice core site and the meteorological stations.

Despite these limitations, it should be noted that the $\delta/T_s$ gradients derived from simple regression of the $\delta$ and instrumental temperature series fall consistently (generally in the range 40–50% but up to a factor of 2) below the geographically determined ratios [Dansgaard et al., 1975; Robin, 1983; Arista-
caused by diffusion [Johnsen, 1977; Whillans and Grootes, 1985], this definitely oversamples the record. Snow pit profiles typically sampling 3 years (2 m) were constructed near the GISP2 and GRIP sites over several consecutive years, again providing sufficient samples for accurate analyses.

Detailed curve matching of isotope-depth records from the snow pits to temperature-time records from the AWS and satellite sensors shows that major features are present in both curves [Shuman et al., 1995]. This includes summer minima and winter maxima, the late autumn or early winter secondary warm peak, and some other events, for four or more match points per year. Assuming that this excellent match implies causation, then timelines may be drawn in the snow pit data. The validity of this assumption can then be checked. The timelines plus density profile data allow estimation of the seasonal distribution of snow accumulation. The pattern produced [Shuman et al., 1995] is quite similar to that of the seasonal distribution of precipitation based on atmospheric measurements and models [Bromwich et al., 1993; Shuman et al., 1995], indicating that the snow pit records are being interpreted properly. Diffusive smoothing of the annual signal [Johnsen, 1977] is evident over the 3 years sampled in each snow pit. Restricting attention to the first year only, the data produce a calibration of 0.51%/°C (0.34–0.98%/°C with 95% confidence). Using data from all 3 years of each snow pit yielded 0.46%/°C (Figure 2), so the snow pit diffusion is not extreme from 1-3 years. This calibration is dominated by the large summer-to-winter signal. However, from additional data not detailed by Shuman et al. [1995], the winter-only and especially the summer-only match point data yield a slope that appears to be positive, although this analysis lacks high confidence.

These short-term comparisons show that the isotopic ratios of accumulated snow do reflect temperatures with some accuracy, especially for comparisons from summer to summer and probably from winter to winter, and thus from year to year. Some noise is, of course, present. The calibration is in the neighborhood of 0.5%/°C for oxygen isotopic ratios [Shuman et al., 1995; also Shuman et al., this issue], i.e., ~25% lower than the observed present-day spatial slope for Greenland (0.67%/°C [Johnsen et al., 1989]).

2.3. Use of Melt Layers

Another empirical technique provides support for the isotopic calibration. Frequency of occurrence of melting in usually cold regions is related to the temperature [Herron et al., 1981; Koerner and Fisher, 1990]. At one site, Agassiz Ice Cap in the Canadian Arctic, isotopic analysis has been performed on an ice core penetrating the early Holocene period, and Koerner and Fisher [1990] were able to make an independent estimate of the overall temperature shift during the period 8.5 kyr B. P. to the present from the changing percentage of melt layers in the core. These authors suggested that summer temperatures have decreased by ~2.0°C since the early Holocene, which is consistent in that case with what is inferred using the present-day spatial slope observed in Greenland.

Observations of the GISP2 core show significant variations in melt frequency over time, consistent with cooling from middle to late Holocene [Alley and Anandakrishnan, 1995]. Calibration of this signal was attempted in two ways. First, the melt frequency from site A, Greenland, which is about 2°C warmer than GISP2 today and has more melt than during the mid-Holocene at GISP2 [Alley and Koci, 1990], places limits on GISP2 temperatures. A probabilistic interpolation to the GISP2 mid-Holocene melt frequency gives a cooling from middle to late Holocene of 1-2°C. Alternatively, the daily mean temperatures and their standard deviation can be estimated from automatic weather station data. Assuming normally distributed temperatures, the recent melt frequency then can be related to this standard deviation, and the shift in mean temperature required to give the mid-Holocene frequency estimated. This gives the same temperature change from middle to late Holocene as does the spatial calibration. Of course, these calibrations involve many assumptions. The largest may be that the change in melt frequency is related to a change in mean temperature rather than to a change in temperature variability. The observation that melt frequency decreases toward the
present is consistent with the results of the borehole-calibrated isotopic thermometer (described below), which also indicates cooling toward the present.

If one accepts the melt frequency and calibrated isotopic thermometers (see below), then together they provide further information. Melt is an indicator of summertime temperatures, whereas borehole-calibrated isotopes record mean annual temperature. (Note that melt features are sparse everywhere in the core, at < 1% of the ice, and so would not have affected the borehole temperature profiles.) Similarity of Holocene summertime and mean annual temperature changes implies similar winter and summer temperature trends, with both cooling from middle to late Holocene. Insolation changes over Greenland from orbital (Milankovitch) causes would have produced summertime cooling but wintertime warming or stability over this time, so these observations imply processes other than insolation [also Alley and Anandakrishnan, 1995].

2.4. Borehole Thermometry

Near-surface (~10 m) temperatures of ice sheets lacking abundant surface melt features are similar to (within roughly 1 to 2°C) and primarily controlled by mean annual temperature [e.g., Paterson, 1994]. Changes in air temperature propagate into the ice by diffusive heat flow and by ice flow, while the record is smoothed by the diffusive heat flow. The temperature profile through an ice sheet thus provides a record of past air temperature, modified by heat diffusion and ice flow, and by the small and well-characterized heat generation from ice deformation. The profile is readily measured in a borehole into or through the ice [e.g., Gundestrup et al., 1994; Cuffey et al., 1994, 1995]. The thermal properties of ice and firm are well known. Uncertainties introduced by lack of knowledge of ice flow range from tiny to huge in different regions of ice sheets. Availability of excellent dating at the Greenland summit, combined with the central location on a rather stable ice sheet, allows accurate calculation of ice flow effects.

Because of heat diffusion, conversion of the depth temperature record of a borehole to a surface temperature history involves some difficulties. The borehole temperature record of old, high-frequency events is lost entirely. More than one possible climate history would produce a given measured borehole profile, although a range of other plausible surface temperature histories can be excluded. Several approaches exist to interpret borehole temperature profiles and compare them to isotopic records. One can conduct a formal inversion for the surface temperature over time, using various rules to choose the “best” history among the compatible ones [e.g., MacAyeal, 1995]. Rules may include constraints on the rapidity or magnitude of allowable surface temperature change, or a requirement that the surface temperature history more or less closely match an independently estimated temperature history. The derived surface temperature history then can be compared to isotopic records [e.g., Dahl-Jensen and Johnsen, 1986]. Care must be used to avoid circularity, if an isotopically based history is used to constrain the inversion.

A different, and very powerful, technique was suggested by Paterson and Clarke [1978]. This involves calibrating the isotopic paleothermometer against the borehole profile. One assumes that isotopic ratios record surface temperature, but that the calibration is not known. The resulting relation is used to convert the history of isotopic values to a history of surface temperatures, which is used as the upper boundary condition of a time-dependent model of heat and ice flow in the ice sheet to calculate a modern borehole temperature profile. Comparison of the calculated and measured borehole profiles (typically in a least squares sense) then provides a test of the assumptions. Typically, the initial guesses will not provide a good fit between the calculated and observed borehole temperature profiles. An inverse procedure then is used to adjust the constants in the calibration. If a good fit is obtained, then one has established that the isotopic record provides a good thermometer using the assumed calibration form, and one has calibrated that thermometer. If a good fit is not obtained, then either the procedure has failed in finding the optimal correlation, or the isotopic ratios do not provide a good thermometer. A “good” fit is one that is much closer than could be obtained from a random “isotopic” record with time series characteristics similar to those of the observed record. In practice, isotopic records containing thousands of data points have been filtered through ice and heat flow models with two or three adjustable parameters to match borehole temperature curves which cannot be matched as well in a curve-fitting exercise without introducing many more free parameters, demonstrating that “good” fits are obtained. Notice that if the borehole temperature profile is close to steady state, and if the isotopic record lacks significant fluctuations at frequencies that would be expected to perturb the borehole profile significantly, then only part of a calibration (case of a constant surface temperature) is possible.

This exercise has been independently conducted for the GISP2 [Cuffey et al., 1995] and GRIP [Johnsen et al., 1995] fluid-filled deep boreholes and also for temperature profiles from a 216-m, air-filled borehole near the GISP2 core [Alley and Koci, 1990; Cuffey et al., 1992, 1994]. The calculations at shallow depths are almost entirely insensitive to ice flow uncertainties. They are primarily sensitive to changes in surface temperature over the most recent few centuries. Analyses of the deep holes were restricted to the fluid-filled section, which did not extend to the surface. Thus they are largely insensitive to surface temperature changes over the most recent century, whereas they are sensitive to changes through the glacial maximum and beyond. The two teams assumed different forms for the calibration curves: Cuffey et al. [1995] assumed a linear relation between isotope and temperature but allowed some time variations, while Johnsen et al. [1995] kept this linear coefficient constant but allowed for an additional term proportional to δ². They, however, ended up with similar leading results: (1) the isotopic ratios are an excellent paleothermometer, over all time intervals studied; and (2) the calibration is different for recent or short times, i.e., for the last few centuries to millennia, (~0.5‰/°C and 0.6‰/°C at GISP 2 and GRIP, respectively, for δ¹⁸O) than for older or longer times (~0.33‰/°C and 0.23‰/°C, respectively). Notice that Beltrami and Taylor [1995] recently inferred a slightly lower value for the recent past (~0.42‰/°C) in comparing, for the last 2000 years, the δ¹⁸O in the Agassiz Ice Cap (Canadian Arctic) and the borehole temperature in a well close to this ice cap.

Figure 3 shows the GRIP and GISP2 temperature records then derived from the δ profiles, over the last 100 kyr. As a result of the high sensitivity for glacial conditions (3°C/‰ or higher, i.e., about twice or more that derived using the spatial gradient of 0.67‰/°C), the isotopic profiles indicate a temperature increase from glacial maximum to Holocene higher than 20°C and up to 25°C at Summit, taking into account the lapse rate effect of thickening associated with accumulation increase at the end of the ice age and the change in seawater δ¹⁸O (as
based on the fact that in the presence of a temperature gradient, gases become fractionated due to thermal diffusion [Severinghaus et al., 1996]. Preliminary analyses of air from GISP2 yield an inferred Younger Dryas firm temperature that is colder than the value inferred from isotopes in the ice using the present-day spatial slope, but agrees with borehole temperature--based estimates. Application of this method to the Dansgaard-Oeschger events may reveal whether they were associated with larger temperature changes in central Greenland than inferred using the present-day spatial slope.

Interestingly, the approach consisting of calibrating the isotopic record using the temperature profile has been independently, and with a slightly different method, followed for the Vostok deep core in central east Antarctica [Salamatin et al., 1997] with the supplementary assumption that there is an additional precession signal in the variations of the surface temperature (estimated by paleothermometry) with respect to the variations in the inversion temperature (derived from the snow isotopic content, see below). This approach suggests that long-term (glacial-interglacial) temporal and spatial isotope/temperature slopes are more comparable in central Antarctica (within 30% or less) than in Greenland but, again, leads to higher glacial-interglacial temperature changes than initially thought.

### 2.5. Information Inferred From Accumulation Change

A further empirical technique aiming to provide some temperature information, classically used in many studies, is to correlate temperature and snow accumulation [e.g., Lorius et al., 1985; Ritz, 1992; Jouzel et al., 1993]. These authors calculated the precipitation rate continuously along the Vostok cores as the product of its present value and the ratio of the derivatives of the saturation vapor pressure for the time of precipitation (derived from the isotopically derived temperature change) and for present conditions. This is the formulation that is directly derived from a simplified unidimensional model, neglecting possible changes in atmospheric circulation and in the local horizontal temperature gradient above the area of precipitation. Surprisingly, support for this very simple one-dimensional model of estimating paleoprecipitation rate came from the measurement of $^{10}$Be, a cosmogenic isotope produced by the interaction of cosmic rays with the upper atmosphere. Assuming a constant $^{10}$Be deposition flux (which at best is a first-order assumption valid only over long periods [Yiou et al., 1985; Raisbeck et al., 1987; Mazaud et al., 1994]), accumulation rate changes can be deduced which overall show a quite remarkable agreement with those derived from the isotopic profile [Jouzel et al., 1989; Genthon et al., 1994]. More impressive, and less subject to various assumptions, is the comparison between the accumulation rate at GISP 2 (directly inferred from the annual layer thickness) and the isotopic profile, as illustrated in Figure 4 for the period between 18 and 10 kyr B. P. [Kapsner et al., 1995; Stuiver et al., 1995]. Though less detailed at the moment, the comparison between the accumulation and the temperature derived from the isotope record in the GRIP core is also impressive [Dahl-Jensen et al., 1993]. There is thus no question that snowfall includes a thermodynamic term. Indeed, except for the middle to late Holocene, for which Cuffey and Clow [this issue] have noted a strong negative correlation between snow accumulation and temperature, those two parameters covary on all submillenial timescales studied and over longer timescales in the ice age.

However, it is equally true that snowfall depends on atmospheric circulation. Indeed, we see time variations in the correlation between snow accumulation and isotopic ratio illustrated in Figure 8, such large present-day/glacial maximum (LGM) temperature differences are predicted in some Greenland grid points, but note that these model results are not corrected for the model change in topography). The calibration also accounts for the change for the effect of changing ice elevation in central Greenland. One consequence is that although the isotopic records are flat, the temperature records show a postglacial warming between 10 and 8 kyr B. P., which as for other high-latitude records [Koerner and Fisher, 1990; Ciais et al., 1992] occurred before the hypsithermal climatic optimum (6 kyr B. P.).

Notice that paleothermometry does not allow direct calibration of the isotopic changes at such rapid climatic reorganizations as the termination of the Younger Dryas cold event, or the onset and termination of the stadial phases of the Dansgaard/Oeschger oscillations. As shown by Firestone [1995], because of heat diffusion the temperature signal of the Younger Dryas event is not detectable in the borehole temperature record.

This limitation does not hold true for a new method using a temperature change indicator in the gas phase. This method is based on the fact that in the presence of a temperature gradient, gases become fractionated due to thermal diffusion [Severinghaus et al., 1996].

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**Figure 3.** Calculated temperature change at the GRIP and GISP2 sites over the last 100 kyr. The GRIP record (adapted from Johnsen et al. [1995]) is calculated using the second-order equation derived by these authors and corrected for surface elevation changes and for the changing composition of seawater. The GISP2 record is adapted from Cuffey et al. [1995] using a calibration of 0.5‰/°C for the Holocene and of 0.33‰/°C before (with no adjustment in source water composition). For this comparison the GISP2 profile has been placed on the GRIP timescale developed by Dansgaard et al. [1993] in using correspondences that can be easily established using similarities between the two isotopic profiles.
Dahl-Jensen et al. [1993]). These variations are consistent with the hypothesis that the major climatic shifts involve reorganizations of the atmospheric circulation, such that the storm track reaches Greenland during interglacial and interstadial times, but trends elsewhere during glacial and stadial times. Nonetheless, the observed changes in snowfall are difficult to explain by circulation changes alone [Fawcett et al., 1995, 1997], suggesting that there are large temperature changes at the large isotopic jumps, as expected. Indeed, using a temporal slope of 0.33‰/°C [Cuffey et al., 1995] instead of 0.53‰/°C [Kapsner et al., 1995] would lower the sensitivity of snow accumulation rate to temperature change to ~6‰/°C for the entire GISP2 data set (thus reinforcing the role of thermodynamics with respect to atmospheric circulation as driving changes in accumulation). We note here that Stuiver et al. [1995] inferred $\delta/\Theta$ gradients at GISP2 for the Bölling-Allerod and the Preboreal periods (0.28‰/°C and 0.42‰/°C, respectively) assuming thermodynamic control of accumulation and observed isotope accumulation relationships. However, shifting storm tracks over Greenland are consistent with our understanding of the North Atlantic system and limit the use of accumulation as a thermometer (note, however, that major circulation changes will inevitably be accompanied or caused by significant temperature changes). Conclusive tests for such storm track shifts are not yet available from the Antarctic and should be sought, but simpler meteorological conditions suggest that Antarctic storm track shifts will be less important and that accumulation could be a better thermometer than for Greenland.

3. Model Estimates of the $\delta/\Theta$ Relationships

Water isotopes have been incorporated into a hierarchy of models including dynamically simple Rayleigh-type distillation models, two-dimensional models and atmospheric general circulation models (GCM). In this section, we first examine what are the main climate parameters or processes that influence the worldwide distribution of isotopes in precipitation, through a short description of simple models. We then review the current state of development of isotopic GCMs and their performance for present-day climate and experiments performed for the climate of the last glacial maximum (LGM, ~20,000 years ago).

3.1. Fractionation Processes and Simple Rayleigh-Type Models

Fractionation processes take place at most of the phase changes of the water during its atmospheric cycle. Two processes have to be considered. First, the saturation vapor pressures of HDO and H$_2$O are slightly lower than that of H$_2$18O. As a result, the condensed phase (either liquid or solid) is, at equilibrium, isotopically enriched with respect to the vapor phase. The fractionation coefficients (i.e., the ratios of D/H or 18O/16O in the condensed and in the vapor phases), are practically equal to the ratio of the saturation vapor pressures of the corresponding molecules and depend only on the temperature and on the phase change considered. A second isotopic effect, “the kinetic effect,” is due to the fact that molecular diffusivities in air are lower for isotopically heavier molecules than for H$_2$18O, which induces an additional effect for non-equilibrium processes (evaporation and condensation). The equilibrium isotopic effect is 8–10 times higher for deuterium than for oxygen 18, whereas the kinetic effects are of the same order. In a relative sense the kinetic effect is thus much more important for H$_2$18O than for HDO, which is one of the reasons why it is interesting to analyze both isotopic species (see Jouzel [1986, and references therein] for a review).

A Rayleigh model [Dansgaard, 1964] considers the isotopic fractionation occurring in an isolated air parcel traveling from an oceanic source toward a polar region. The condensed phase is assumed to form in isotopic equilibrium with the surrounding vapor and to be removed immediately from the parcel. Under these assumptions, the isotope content of this precipitation is a unique function of the initial isotope mass and water vapor mass within the air parcel and of the water vapor mass remaining when the precipitation forms. The parcel’s water vapor content is proportional to the saturation vapor pressure, a function of temperature and phase change, and is inversely proportional to the air pressure. Thus, in this simple model, the isotope content of precipitation depends only on initial isotopic concentrations of the vapor (i.e., when the first precipitation forms) and on the initial and final condensation temperatures and air pressures. This model quite satisfactorily explains the main features of the global distribution of isotopes in precipitation, namely, its seasonal and spatial characteristics, the observed relationships with local temperature or precipitation amount, and the strong link between $\delta D$ and $\delta^{18}O$ [Craig, 1961; Dansgaard, 1964; Friedman et al., 1964]. Further developments of these Rayleigh-type models focused on how to estimate the initial isotopic concentrations of the vapor as a function of sea surface conditions [Merlivat and Jouzel, 1979] and how to account for the kinetic fractionation associated with snow formation [Jouzel and Merlivat, 1984; Ciais and Jouzel, 1994]. Figure 5 shows the combined influence on $\delta$ of the sea surface temperature, $T_s$, and of the temperature of formation of the precipitation, $T_c$. The existence of large drops which are out of isotopic equilibrium prevents the application of these simple models to convective systems, but isotopic
cloud models in which the presence of large drops is taken into account have now been developed [Jouzel et al., 1980; Federer et al., 1982; Gedzelman and Arnold, 1994]. They are, however, limited to the study of idealized clouds and cannot account for the complexity of large convective systems, such as those occurring in tropical and equatorial regions.

Despite such limitations, simple isotopic models are appropriate to explain the main characteristics of $\delta D$ and $\delta^{18}O$ in precipitation, at least in middle and high latitudes where the precipitation is not predominantly produced by large convective systems. Indeed, their ability to correctly simulate the present-day temperature-isotope relationships in those regions has been the main justification of the standard practice of using the present-day spatial slope to interpret the isotopic data in terms of records of past temperature changes. Notice that, at least for Antarctica (Figure 5), data and simple models agree only with respect to the temperature of formation of the precipitation, estimated by the temperature just above the inversion layer [Robin, 1977], and not with respect to the surface temperature, which owing to a strong inversion is much lower. From this figure, the combined influence of condensation ($T_c$) and source ($T_w$) temperature changes can be written $\Delta \delta^{18}O = 1.1 \Delta T_c - 0.55 \Delta T_w$ (where $\Delta$ stands for the differences between two different climates) [Aristarain et al., 1986]. Thus one can easily see that using the spatial slope as a surrogate of the temporal slope strictly holds true only if the characteristics of the source have remained constant through time. Interestingly, one can note that a simultaneous change in the source and in the precipitation site temperatures results in a temporal slope lower than the spatial one (by a factor of 2).

This example illustrates how the temporal slope used by paleoclimatologists may differ from the modern spatial slope. However, one can think of many other reasons not taken into account in a simple Rayleigh-type model which in the real and very complex world can lead to differences between those two slopes and thus bias the standard interpretation of paleodata. Furthermore, such a simple model cannot adequately account for the complexity of dynamical and microphysical processes leading to the formation of individual precipitation events, or for the changes in ocean surface characteristics, in surface topography, and in atmospheric circulation associated with important climatic changes, such as the transition between the last glacial maximum and the Holocene. Water isotope cycles have been incorporated into atmospheric GCMs which already account for the complexity of the atmospheric processes leading to the formation of precipitation at these different timescales.

### 3.2. Isotope Modeling With GCMs

Atmospheric general circulation models simulate the time evolution of various atmospheric fields (wind speed, temperature, surface pressure, specific humidity) discretized over a global grid, through the integration of the basic equations: hydrostatic equation of motion, thermodynamic equation, mass continuity equation, and water vapor transport equation. The incorporation of the HDO and $H_2^{18}O$ cycles into such models is relatively straightforward. It consists in following the two stable isotopes through every stage of the water cycle in accounting for isotopic fractionations, including both equilibrium and kinetic effects, at every change of phase, i.e., during surface evaporation, atmospheric condensation, and reevaporation of precipitation. Joussaume et al. [1984] pioneered this isotopic GCM approach, producing global isotopic fields for present-day January climate with the GCM of the Laboratoire de Météorologie Dynamique (LMD, Paris) using a low-resolution version of this model. Jouzel et al. [1987] generated a full annual cycle of isotope fields with the $8^{\circ} \times 10^{\circ}$ NASA
Jouzel, 1993; Jouzel et al., 1994; Charles et al., 1994, 1995; Mapping and Prediction (CLIMAP), 1981. A further advantage of the ECHAM [Hoffmann and Heimann, 1993, 1997] which is the Hamburg version of the European Centre for Medium-Range Weather Forecast GCM. Beyond these modern climate simulations aiming at examining how isotopic GCMs are able to reproduce the observed isotopic distributions, such modeling efforts have a common objective, which is the reconstruction of paleoclimatic isotopic fields to help in the interpretation of paleodata. Such studies focusing on the last glacial maximum have been performed with the LMD [Joussaume and Jouzel, 1993], the GISS [Jouzel et al., 1994; Charles et al., 1994, 1995] and the ECHAM [Hoffmann and Heimann, 1997] GCMs.

Figure 6 displays the annual average $\delta^{18}O$ content of precipitation as predicted with the GISS model, whereas the corresponding contour plot was obtained from International Atomic Energy Agency (IAEA) observations and complementary data [Jouzel et al., 1987]. A common feature, also shared by the two other models, is the clear latitudinal pattern with $\delta^{18}O$ decreasing as one approaches the poles. The predicted $\delta^{18}O$ field compares well with observations, though the predicted values in middle latitudes are slightly too low. In contrast, polar values are slightly overestimated, at least over Greenland. Producing too high polar values is a defect which is indeed much more pronounced both in the LMD and in the ECHAM model. For tropical and equatorial sites the isotopic content of the precipitation is controlled mostly by precipitation amount and not by temperature; this observed behavior is well reproduced in the three models. In the lower temperature range the models correctly simulate the observed linear spatial relation between $\delta^{18}O$ and temperature. The observed and predicted gradients are within 10% except for the ECHAM model, which predicts a slightly lower value (see Jouzel et al. [1996] for a review). Furthermore, the observed linear relationship between $\delta$D and $\delta^{18}O$ is well reproduced also. Model performances are also generally satisfactory with respect to the prediction of the seasonal isotopic distributions [Jouzel et al., 1987; Hoffmann and Heimann, 1993, 1997]. This was not at all the case over Greenland in the $8 \times 10$ GISS simulation, but this defect is much less pronounced with the $4 \times 5$ GISS model simulation, indicating that the lack of a $\delta^{18}O$ seasonal cycle over Greenland is due to too low a resolution, giving Greenland air masses an excessively maritime character [Jouzel et al., 1987].

Overall, the ability of isotopic GCMs to reproduce the main characteristics of $\delta$D and $\delta^{18}O$ in precipitation has been judged sufficient to start simulations of paleoclimatic fields to help in the interpretation of paleodata. The studies [Joussaume and Jouzel, 1993; Jouzel et al., 1994; Charles et al., 1994, 1995; Hoffmann and Heimann, 1997] focus on the last glacial maximum (LGM) because the glacial climate is very different from the current climate and the LGM boundary conditions are sufficiently well known [Climate: Long-Range Investigation, Mapping and Prediction (CLIMAP), 1981]. A further advantage of the LGM is that isotope paleodata are available for this period in both polar and temperate regions, allowing partial validation of model results through either $\delta$D or $\delta^{18}O$. Comparisons with available paleodata [Joussaume and Jouzel, 1993; Jouzel et al., 1994] suggest that GCMs reproduce LGM isotope concentrations reasonably well and are adequate to examine the key question that paleoclimatologists have to address, i.e., are the present-day isotope/temperature spatial slope and the temporal slope similar? Illustrated in Figure 7a is the GISS model distribution of temporal slopes calculated by dividing the difference in precipitation $\delta^{18}O$ by the difference in surface temperature after correction for the change in the ocean $\delta^{18}O$ and application of a simple smoothing algorithm to the simulated temperature and $\delta^{18}O$ fields. The salient feature is the first-order agreement between the temporal slope and the present-day global spatial slope of 0.6%°C over the northern ice sheets, where the slope is within 0.2%°C of this spatial slope [Jouzel et al., 1994]. Although there are high differences in some regions, for example, 50% in North America, the overall similarity between temporal and present-day model-derived spatial slopes led these authors to suggest that spatial slopes seem on average to be adequate surrogates for temporal slopes. Over the ice sheets, relative errors are of the order of 30% or less.

Jouzel et al. [1994] also noted that in the GISS GCM experiments temporal slopes are generally lower than spatial slopes but did not infer any conclusion from this result as far as interpretation of paleodata is concerned. One reason is that this characteristic was not shared by all regions; instead the temporal slope is higher than the spatial slope over east Antarctica. More important was the fact that the two simulations then available (LMD model and coarse grid version of the GISS model) showed differences on a regional basis [Joussaume and Jouzel, 1993] and clearly not too much weight could be given to a conclusion which is model dependent. Rather, Jouzel et al. [1994] emphasized one result which is common to both models, i.e., that over polar ice sheets, model-predicted temporal and spatial slopes are, for a given model, within 30%. Such a cautious approach is indeed justified by the comparison of the two GISS model simulations now available (coarse and fine grid). Preliminary analysis (Figure 7b) shows that the fine grid version predicts higher temporal slopes over Greenland than the coarse grid version (0.80%°C over Greenland instead of 0.43%°C). There is presently no clear explanation for this difference but, again, the spatial and temporal slopes are internally consistent (the present-day predicted spatial slope over Greenland is 0.89%°C with the $4 \times 5$ version of the GISS model and 0.48%°C with the $8 \times 10$ version). Recent modern and LGM simulations [Hoffmann and Heimann, 1997] performed with the isotopic version of the ECHAM model (2.8° × 2.8°) present interesting information in this respect. Over central Greenland the present-day spatial gradient is 0.57%°C, whereas the temporal slope is 0.51%°C. On the one hand, the general statement that present-day and temporal slopes are within 30% of each other over polar ice caps holds true. On the other hand, we have here a second example of the temporal slope being lower than the spatial slope (16%). It is also noted that the temporal slope is indeed highly variable with values weaker in central Greenland (0.45%°C, relatively close to that inferred from paleothermometry) than in north Greenland.

The growing empirical evidence showing that temporal slopes are consistently lower than spatial slopes for century to millenia changes, and much lower for glacial-interglacial changes, now led us to examine, in the light of both simple
Figure 6. Global distribution of $\delta^{18O}$ in precipitation for the present-day climate (adapted from Jouzel et al. [1987]). The bottom map is derived from observations, while the top map is produced from a 3-year simulation of the NASA GISS model.
models and GCMs and in focusing on the Greenland area, how various factors can influence the temporal slope.

4. Factors Possibly Influencing the δ/Δ relationships

The review of empirical estimates of the δ/Δ relationships brings compelling arguments indicating that the temporal slope, in a given site or region, is lower than the present-day spatial slope in this region. The temporal slope appears consistently closer to the spatial slope during the recent period than for the ice age climate. In this latter case the difference can reach a factor of 2 or even more, at least over Greenland. From simple models we have pointed out that part of the difference may be due to a simultaneous change between the condensation and source temperatures. Other factors, most of them already mentioned in this review, are currently cited as potentially influencing the distribution of water isotopes between two different climates, and thus the temporal slope. Numerous authors have drawn attention to the possible influence of a change in the precipitation distribution during the year [Robin, 1983; Steig et al., 1994]. The fact that the precipitation is intermittent implies that the isotope record is, in any event, only a discrete and possibly biased sampling of the temperature record (the temperature during precipitation events may not be representative of the annual mean temperature) is also cited. In the polar regions, for example, this is likely to undersample the extreme cold periods, frequently associated with a temperature inversion. One can also think of the influence of modifications in cloud microphysical properties affecting isotopic fractionations [Fisher, 1991]. Finally, we should keep in mind that isotopic changes record cloud temperature [Cuffey et al., 1995], whereas empirical estimates, e.g., borehole thermometry, give access to the temperature at the Earth surface.

Broadly speaking, those factors fall into two categories de-
pending on whether they relate to the fact that parameters other than the temperature of formation of precipitation influence its isotopic content (source characteristics, microphysical processes) or to the bias that is inherently introduced in the way that the temperature record is sampled by snowfall, related to the seasonality and intermittency of precipitation formation and to the difference between atmospheric and surface temperatures. We successively examine their potential influence, largely using information derived from simulations of water isotope distribution [Jouzel et al., 1994] and of changes in moisture source for Greenland [Charles et al., 1994, 1995] performed with the GISS model. In this section we thus focus on large glacial-interglacial changes.

4.1. Origin of the Precipitation

The GCM approach is particularly well suited to examining the link between the evaporative origin of a precipitation mass and its isotope content. Water evaporating from a well-defined source region on the Earth's surface can be "tagged" in the GCM and followed through the atmosphere until it precipitates. Through this approach, the relative contributions of many different evaporative regions to a given region's precipitation can be quantified exactly. Joussauume et al. [1986] determined the evaporative contribution of 10 global divisions to local continental precipitation in the LMD model. The GISS model was used to determine the sources of local precipitation in the northern hemisphere [Koster et al., 1986], and Koster et al. [1992] followed both the H2O and the HDO coming from a given source (defined by sea surface temperature) during simulations of July climate with the GISS model. Their results show that the deuterium content of Antarctic precipitation decreases as the temperature Te of the evaporative source for the water increases, by about the amount predicted by simple Rayleigh-type models, which suggests that meteorological conditions are relatively simple over Antarctica.

The situation appears to be more complex over Greenland (see, for example, Newell and Zhu [1994]). Charles et al. [1994] performed a similar experiment with the 4 × 5 version of the GISS model, focusing on Greenland precipitation and defining the evaporative source regions geographically rather than according to sea surface temperature. As was found for Antarctica, several evaporative source regions contribute to the precipitation at a given Greenland site, and the isotope contents of the different contributions vary significantly. For example, moisture from the North Pacific source arrives at the Greenland coast with a δ18O value roughly 15% lower than its North Atlantic counterpart (a difference comparable to that predicted for Antarctica when comparing the contributions from the warm and intermediate sources). Charles et al. [1994] attributed the lower δ18O to the fact that North Pacific moisture is advected along a much colder path before reaching Greenland; the tagging by large regions makes the comparison with simple models less straightforward than for Antarctica, but a recent simulation in which Greenland precipitation is tagged with respect to the oceanic temperature indicates a clear dependency of the isotopic content of the precipitation with respect to this parameter. As an extreme example, Charles et al. [1994] point out that a δ18O anomaly of 7‰ would be generated at a Greenland site if no climate change (including temperature) other than a shift from a pure North Atlantic contribution to an even mixture of North Atlantic and Pacific moisture occurred.

This example has opened a debate on the relative extents to which local temperature changes and changes in the evaporative sources of precipitation define the isotope shifts recorded in Greenland ice cores. The example, however, must not be misinterpreted. As noted by Charles et al. [1994], the GCM results do not suggest that such extreme moisture source changes for central Greenland actually occurred. Indeed, the present-day/LGM change in the contribution of North Pacific moisture is only 2%, not 50% as in the above example. A simple calculation that accounts for the simulated changes in moisture sources shows that in the GCM, even a 30% difference between the isotope contents of Pacific-derived and Atlantic-derived water translates into only a 1‰ or 2‰ net present-day/LGM change, which is still relatively small with respect to the 5–12‰ shifts recorded in Greenland cores.

4.2. Microphysical and Atmospheric Processes

Besides changes in moisture origin (and concurrently in atmospheric circulation), one can think of many other factors potentially influencing the isotopic distribution in polar regions. One such example concerns the kinetic fractionation at snow formation from the vapor phase [Jouzel and Merlivat, 1984; Ciais and Jouzel, 1994]. Fisher [1991] noted that the temperature at which this nonequilibrium process starts, i.e., where there are no longer liquid droplets in the cloud, depends on the "cleanliness" of the air. Also, supersaturation levels that prevail at snow formation depend on cloud condensation nuclei (CCN) concentration. One can thus speculate [Fisher, 1991] that present-day/glacial changes in impurity loading, well-documented in ice cores [e.g., Alley et al., 1995], could modify the importance of the isotopic kinetic effect. In agreement with simple model results, GCM simulations [Jouzel et al., 1991] confirm that the isotopic content of polar snow is influenced by supersaturation changes. However, sensitivity to supersaturation is much stronger for deuterium excess [Jouzel et al., 1991] and one can hope, in examining variations of this latter parameter in different ice cores from coastal and inner sites, either to demonstrate that impurity loading has no influence on isotopic distributions, or, at least, to estimate its maximum influence. Other factors related to the formation of the precipitation have less significant influence on the isotopic content of polar snow [Jouzel et al., 1991]. Finally, note that isotopic GCMs implicitly account for the relative importance of isobaric and adiabatic cooling, the influence of which has been pointed out from simple model results [Dansgaard, 1964; Grootes and Stuiver, 1987].

4.3. Seasonality of the Precipitation

As recently noted by Steig et al. [1994], the deposition by precipitation of any atmospheric constituent that exhibits large seasonal changes and relatively small long-term changes will be sensitive to the seasonal timing of the precipitation. In fact, the average δ18O in a region is more logically related to the precipitation-weighted temperature than to the mean annual temperature Ta in the region: If all the precipitation occurs during warm summer months, for example, the "annual δ18O" will naturally reflect a temperature warmer than Ta. This consideration is supported by observations of δ18O in central Greenland over the past century, which are found to track the precipitation-weighted temperature more closely than Ta at Jakobshavn (a west Greenland coastal site). If major changes in seasonality occur between climates, such as a shift from summer-dominated to winter-dominated precipitation, the impact on the isotope signal could be large.
Simulations performed with the 4 × 5 resolution GISS isotope model [Charles et al., 1995] reveal virtually no systematic change in the seasonal timing of Greenland precipitation between the ice age and present-day climates. As a result, simulated isotope shifts over Greenland do not correlate significantly better with shifts in "precipitation-weighted temperature" than with shifts in annual temperature ($r^2$ increases from 0.76 to 0.78). Therefore, at least for these two climates over Greenland, the GISS model results do not support Steig et al.'s suggestion that changes in seasonality have a primary influence on isotope shifts [Charles et al., 1995]. This example, though geographically limited (the situation may indeed be different for other regions), serves to illustrate how isotope GCMs can be used to examine the influence of seasonality on isotope content.

Following another approach, Fawcett et al. [1995, 1997] came to a different conclusion. Using the seasonal patterns of temperature and precipitation produced in experiments they performed with the Global Environmental and Ecological Simulations of Interactive Systems (GENESIS)/NCAR model (in which water isotopes are now being incorporated) and assuming that colder temperatures produce lighter isotopes, these authors found that changes in seasonality would have a large effect on the isotopic thermometer. They showed that the low value of the temporal slope derived from borehole temperature profiles is consistent with changing seasonality, in which wintertime snowfall is more important in warm times than in cold, and noted that GCMs support this view.

Interestingly, if such a role of seasonality was confirmed and the isotope changes calibrated accurately by models, differences between temperature estimates obtained from isotopes and from paleothermometry would provide further climatic information (the isotopic temperature is weighted by the seasonality of snowfall whereas the borehole temperature is a mean annual indicator). However, model experiments (using, preferably, model versions in which water isotopes are incorporated) should be repeated with different GCMs and boundary conditions to clearly assess if seasonality changes have a major or minor influence on the isotopic thermometer.

4.4. Cloud–Surface Temperature Difference

Beyond the seasonality of precipitation formation, two other factors must also be considered when interpreting the isotopic signal recorded in polar precipitation [Robin, 1983; Peel et al., 1988; Fisher, 1992]. First, it is the temperature during the precipitation events that is imprinted in the isotopic signal. Second, the formation of an inversion layer of cold air up to several hundred meters thick over polar ice sheets (primarily due to radiative loss of heat by the surface) makes the temperature of formation of precipitation warmer than the temperature at the surface of the ice sheet. Inversion forms under a clear sky except in summer months, but even in winter it is destroyed rapidly if thick cloud moves over a site [Robin, 1983]. Thus surface temperature during precipitation events may then be warmer than the average cloud temperature. As a consequence of precipitation intermittency and of the existence of an inversion layer, the isotope record is only a discrete and biased sampling of the surface temperature and even of the temperature at the atmospheric level where the precipitation forms. Current interpretation of paleodata implicitly assumes that this bias is not affected by climate change itself. However, White et al. [1995] recently pointed out that if the frequency of precipitation differs between present-day and LGM climates, this would affect the difference between the temperature during precipitation events and its annual average.

As the isotope signal gives, in any case, a better access to atmospheric than to surface temperatures, the Vostok isotope profile is currently interpreted in terms of atmospheric temperature changes [Lorius et al., 1990; Jouzel et al., 1993]; translation in terms of surface temperature changes is valid only if the present-day linear relationship between the strength of the inversion and the surface temperature at the site is kept unchanged for glacial conditions. Instead, temperature changes inferred from GRIP and GISP2 isotopic profiles have been related to the surface (largely because there is no sufficient information on the strength of the inversion for present-day climate). Still, one can imagine that the inversion there was much stronger for glacial than for present-day conditions, which would offer a way to reconcile the isotopic thermometer and the paleothermometry approach (it cannot be excluded that temperature were ~10°C cooler in the atmosphere and ~20°C cooler at the surface). The existence of an inversion layer over Greenland is well captured by the GISS model (present-day January temperatures in the second layer are more than 10°C warmer than at the surface), and there are indications of a stronger inversion for the LGM climate at least in most of the Greenland grid boxes. Further interpretation is difficult because there are large (and unrealistic) changes in the model topography over Greenland between those two climates which can themselves have a strong influence on the vertical temperature profiles. Interestingly, Cuffey and Clow [this issue] suggest that out of the ~20°C cooling inferred between ice age and modern conditions by Cuffey et al. [1995], 5°C can be attributed to ice age strengthening of the inversion.

5. Discussion and Conclusion

In the light of simple model results (Figure 5), it is tempting to explain the fact that the temporal slope is lower than the spatial slope, as inferred from paleothermometry in central Greenland, at least in part, by a cooler source of Greenland precipitation for glacial than for present-day climate. Presently available GCM experiments do not, however, allow direct testing of this intuitively reasonable assumption as the experiment in which precipitation origin is tagged with respect to the source temperature has not yet been run for glacial conditions. The relatively good agreement between GISS model and paleothermometry results noted by Cuffey et al. [1995] provided strong encouragement for further exploring this assumption. Preliminary analyses of the GISS 4 × 5 simulation data suggest, however, that this agreement may depend on model resolution. We need to verify this resolution dependence and, if it truly exists, understand it; only then can we use the models to assess the possible influence of glacial-interglacial changes in moisture source on temporal slopes in Greenland. A complementary way to assess the role of the origin of the precipitation is to exploit the deuterium excess signal, which is clearly influenced by the characteristics (temperature and humidity) in the moisture source regions [Jouzel et al., 1982; Johnsen et al., 1989; Dansgaard et al., 1989]. The excess profiles will be available soon along the full length of the GRIP and GISP2 cores, and they should tell us if, as we can expect, the average source temperature was colder during glacial than during interglacial periods (to the extent that the influence of other factors potentially influencing the deuterium excess of polar snow, such as changes in cloud microphysical processes, could be evaluated).
Jouzel et al. [1994] examined, with the 8 x 10 GISS model, the in 80 against the concurrent change in local temperature on the glacial-interglacial isotope shifts by plotting the change in precipitation, or in the strength of the inversion layer (or ob-
ject but could help in testing the suggestion of White et al. [1995] relative influences of local temperature and other parameters temperature signal in the isotopic record. To illustrate this, of each of these factors taken individually. However, we have strong evidence that taken together, the influence of these various factors does not mask the local temperature signal in the isotopic record. To illustrate this, Jouzel et al. [1994] examined, with the 8 x 10 GISS model, the relative influences of local temperature and other parameters on the glacial-interglacial isotope shifts by plotting the change in $\delta^{18}O$ against the concurrent change in local temperature over the Laurentide Ice Sheet region. The same plot is given for the grid squares comprising the combined Laurentide and Greenland Ice Sheets by Jouzel et al. [1996]. This shows a strong correlation ($r^2 = 0.72$) between $\Delta \delta^{18}O$ and $\Delta T_s$, which indicates that over 70% of the simulated $\Delta \delta^{18}O$ changes are accounted for by local temperature changes alone (note that the strength of the correlation gives an indication of the capacity to extract temperature changes from isotopic series but contains, by itself, no information on the value of the temporal slope). This suggests that climatic changes in other parameters, such as evaporative source, precipitation seasonality, and inversion strength (but the model also accounts for the intermittency of precipitation events and for the large changes in topography and atmospheric circulation) may explain less than 30% of the isotope signal. We give (Figure 8) the regression analysis of the 4 × 5 GISS model results over Greenland alone. It is even more significant ($r^2 = 0.76$), despite the smaller range of temperature differences.

Among the processes that we have invoked when pointing out that the temperature recorded by isotopes may differ from the mean annual surface temperature, our scientific community focuses on the changes in seasonal timing of the precipitation. This should be explored further both in examining new data sets and through a modeling approach [Charles et al., 1995]. In this context it would also be important to examine thoroughly the capacity of models to correctly reproduce both the seasonal timing of precipitation fallout and its isotopic content. Note that as for seasonality, the bias linked with precipitation intermittency could be easily evaluated in a GCM (in comparing the weighted average temperature during precipitation events with the annual average). This has not been done yet but could help in testing the suggestion of White et al. [1995] that a drastic change in frequency distribution may affect the temporal slope. We have drawn attention, here, to the possible changes in the strength of the inversion layer and would like to put some emphasis on this assumption because of the difference in atmospheric circulation associated with the presence of the Laurentide Ice Sheet. Also, temperatures above the inver-
sion layer are, for polar areas, probably of larger geographical significance than surface temperature, and the former is probably a more significant parameter for study of past climates.

To sum up, several explanations may be offered to explain why temporal slopes are lower than spatial slopes over Green-
land and why this discrepancy was apparently much greater during the last glacial. Most probably, this difference is due to changes either in the moisture origin or seasonality of the precipitation, or in the strength of the inversion layer (or ob-
viously to a combination of these changes). Unfortunately, we are not yet able (even in a model world) to assess the influence of each of these factors taken individually.

However, we have strong evidence that taken together, the influence of these various factors does not mask the local temperature signal in the isotopic record. To illustrate this, Jouzel et al. [1994] examined, with the 8 x 10 GISS model, the relative influences of local temperature and other parameters on the glacial-interglacial isotope shifts by plotting the change in $\delta^{18}O$ against the concurrent change in local temperature.

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As modelers, we are disappointed not to have been able to predict that temporal and spatial slopes differ for glacial-interglacial changes on which model experiments have focused, and not even to be, now, in a position to clearly assess why they differ. This is largely because beyond the fairly strong link between precipitation isotopic content and local temperature that they have in common, the various isotopic models show differences as far as temporal slopes are concerned. These inconsistencies between isotopic models must be resolved (or at least understood). With this in mind, the isotopic GCM groups are now promoting an intercomparison of their models. Ten-year simulations of several different climates (present-day, glacial, Little Ice Age, Climatic Optimum, 6 ky B. P. and 2CO2 climates) will be compared. In addition to the three models presently available (LMD-Paris, NASA/ GISS–New York and ECHAM-Hamburg with, in each case, coarse and fine grid versions), isotopes are now being incorporated into the GENESIS/NCAR. We are still optimistic that this already fruitful modeling approach will, in the near future, provide a solid basis for the climatic interpretation of isotope paleodata over all timescales.

Acknowledgments. This work is a contribution to the Greenland Ice Core Project (GRIP) and to the Greenland Ice Sheet Project 2.
Aristarain, A. J., J. Jouzel, and C. Lorius, A 400 years isotope record


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