JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 101, NO. D23, PAGES 29,531–29,537, DECEMBER 27, 1996

Climatological significance of $\delta^{18}O$ in north Tibetan ice cores

Tandong Yao,[1] Lonnie G. Thompson,[2,3] Ellen Mosley-Thompson,[2,4] Yang Zhihong,[1] Zhang Xingping,[5] and Ping-Nan Lin[2]

Abstract. Oxygen isotopic ratios ($\delta^{18}O$) of precipitation samples collected over several years at three meteorological stations on the northern Tibetan Plateau were used to conduct the first investigation of the relationship between $\delta^{18}O$ and contemporaneous air temperatures (T_a). Inferring past temperatures from $\delta^{18}O$ measured in recently acquired Tibetan ice cores necessitates establishing whether a $\delta^{18}O$-T_a relationship exists. For each station a strong temporal relationship is found between $\delta^{18}O$ and T_a, particularly for monthly averages which remove synoptic-scale influences such as changes in condensation level, condensation temperature, and moisture sources. Moisture source is identified as a major factor in the spatial distribution of $\delta^{18}O$, but air temperature determines the temporal fluctuations of $\delta^{18}O$ at individual sites on the northern Tibetan Plateau. The 30-year records of annually averaged $\delta^{18}O$ from three different ice coring sites are not correlated significantly with contemporaneous air temperature records from their closest meteorological station (150 to 200 km). However, since 1960 the three air temperature records reveal a modest warming trend, while the three contemporaneous $\delta^{18}O$ records show a modest ^{18}O enrichment.

1. Introduction

Oxygen isotopic ratios ($\delta^{18}O$) have been used extensively in paleoclimatic reconstructions [*Dansgaard*, 1953; *Craig*, 1961; *Lorius et al.*, 1985; *Thompson et al.*, 1989, 1995a; *Yao et al.*, 1991; *Jouzel et al.*, 1994] as a proxy for air temperature. The relationship between $\delta^{18}O$ and air temperature (T_a) has been investigated most thoroughly in the polar regions. In Antarctica, various studies [*Aldaz and Deutsch*, 1967; *Lorius and Merlivat*, 1977; *Kotlyakov et al.*, 1982; *Jouzel and Merlivat*, 1984; *Mosley-Thompson et al.*, 1990; *Peel*, 1992] have demonstrated a strong linear relationship between $\delta^{18}O$ and T_a with slopes ranging from 0.76 to 0.92. In Greenland snow, *Dansgaard et al.* [1973] and *Johnsen et al.* [1989] obtained slopes from 0.62 to 0.67. Paleoclimatic reconstructions rely on the robustness of the assumption that the $\delta^{18}O$-T_a relationship can be used to reconstruct accurately the climatic record (e.g,. temperature trends) preserved in ice cores [*Jouzel et al.*,1983]. A recent study by *Joussaume and Jouzel* [1993] demonstrated that using currently measured $\delta^{18}O$-T_a relationships to reconstruct temporal variations in T_a from $\delta^{18}O$ may result in errors of 20% to 30% over Greenland and Antarctica.

Outside the polar regions these studies are more limited. A comprehensive study of the long-term $\delta^{18}O$-T_a relationship was conducted in midlatitude, low-elevation regions [*Rozanski et al.*, 1992] using the relatively long $\delta^{18}O$ and meteorological records from stations in the International Atomic Energy Agency (IAEA)/World Meteorological Organization (WMO) precipitation network. Three different changes in the $\delta^{18}O$-T_a relationship were identified: spatial changes, short-term changes, and long-term changes. Spatially, the $\delta^{18}O/T_a$ slope values varied from 0.71‰ per degree Celsius for high-latitude areas to virtually zero in the tropics. The relationship between long-term changes of $\delta^{18}O$ and T_a for a given location was found to be the most appropriate for paleoclimatic reconstructions purposes. Recent modeling efforts [*Joussaume and Jouzel*, 1987; *Jouzel et al.*,1987, 1991] have used global temperature observations to simulate the basic features of the spatial distribution of $\delta^{18}O$.

Siegenthaler and Oeschger [1980] reported a $\delta^{18}O/T_a$ slope of 0.56 from observations at five Swiss stations. *Jocob and Sonntag* [1991] suggested that different slopes may reflect different thermal regimes, with steeper slopes characterizing colder regions and lower slopes characterizing warmer regions. From a study of the factors controlling $\delta^{18}O$ in European precipitation, *Rozanski et al.* [1982] reported that continuing precipitation leads to an increasing depletion of heavier isotopes in the residual water vapor as it moves across Europe.

Recently, *Cuffey et al.* [1994] used central Greenland borehole temperatures, which are independent of the $\delta^{18}O$ history, to calibrate the paleothermometer or linear relationship, $\delta^{18}O=\alpha T+\beta$. Their results indicate that $\delta^{18}O$ does provide a "faithful" proxy for long-term average temperature at that site. Further, they demonstrated that temperature histories inferred from $\delta^{18}O$ are of higher temporal resolution than borehole temperature histories. Thus abrupt climatic changes can be extracted from $\delta^{18}O$ records but not borehole temperature records [*Cuffey et al.*, 1994, 1995]. They argue that it may be inappropriate to use a single or constant linear $\delta^{18}O$-T_a relationship to infer past climatic changes as coefficients depend upon factors that may vary through time. Their results support the contention by *Jouzel et al.* [1994] that $\delta^{18}O$ is a proxy of temperature and that better estimates of previous values of α may be obtained using a general circulation model which links atmospheric circulation and source temperature changes to a physical model of isotopic fractionation. It is important to consider that borehole temperatures (when reconstructed) reflect near-surface temperatures which are dominated strongly by the radiation regime. The original $\delta^{18}O$ signal is modified postdepositionally and ice core $\delta^{18}O$ results reflect temperatures at the atmospheric condensation level.

Despite the extensive work linking $\delta^{18}O$ in precipitation and climate in polar regions, for high elevation sites in the midlatitudes

[1]Lanzhou Institute of Glaciology and Geocryology, Chinese Academy of Sciences, Lanzhou, Gansu, China.
[2]Byrd Polar Research Center, Ohio State University, Columbus.
[3]Department of Geological Sciences, Ohio State University, Columbus
[4]Department of Geography, Ohio State University, Columbus.
[5]Department of Geography, Hunan Normal University, Changsha, Hunan, China.

Paper number 96JD02683.
0148-0227/97/96JD-02683$09.00

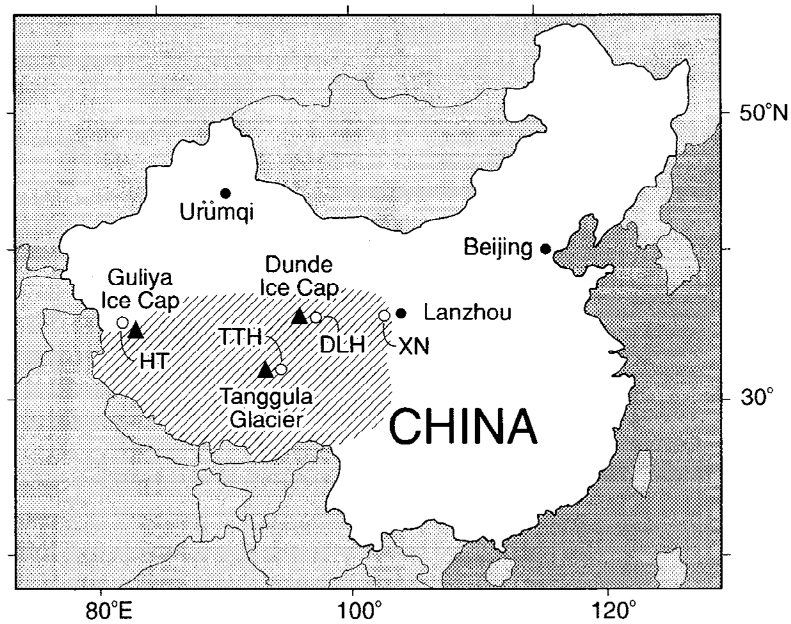

Figure 1. Shown are the meteorological stations (DLH, Delingha; TTH, Tuotuohe; and XN, Xining) on the Tibetan Plateau (shaded) where $\delta^{18}O$ samples and contemporaneous air temperature measurements were taken. Other locations mentioned in the text are also included.

and tropics the observations of contemporaneous $\delta^{18}O$ and air temperatures are very limited. Several ice cores drilled recently on the Tibetan Plateau for paleoclimatic reconstruction [*Yao et al.*, 1990, 1991, 1995; *Thompson*, 1992; *Yao and Thompson*, 1992; *Thompson et al.*, 1989, 1993, 1995b] necessitate a better understanding of the $\delta^{18}O$-T_a relationship.

2. Discussion

This paper reports the first study of the $\delta^{18}O$-T_a relationship on the northern part of the Tibetan Plateau which occupies over 2,000,000 km^2 with a mean elevation of 4500 m above sea level (asl) and is often called "the Earth's third pole. " The study is based on contemporaneous $\delta^{18}O$ and air temperature (T_a) observations at three meteorological stations (Figure 1): Xining (XN) (36°37′N, 101°45′E, 2261 m asl) and Delingha (DLH) (37°22′N, 97°22′E, 2981 m asl) on the northeastern side of the Plateau and Tuotuohe (TTH) (34°13′N, 96°26′E, 4533 m asl) near the center of the Tibetan Plateau. Precipitation (snowfall and rainfall) samples were collected for every precipitation event during the period of observation. Note that ≈ 80% of the annual precipitation on the Tibetan Plateau falls during the summer wet season. At DLH, sampling began in April 1991 while it began in September 1991 at TTH, and only 13 months of observations (September 1991 to September 1992) are available from the XN station.

At all stations, identical sampling procedures were used to collect the snow or rain from each precipitation event. To eliminate contamination and sample carryover, the plastic container used for collection was dried (if necessary) and cleaned with a brush between samples. After collection each sample was placed in a clean plastic bag, melted at about 20°C, and poured into high density polyethylene bottles, and the tops were sealed in wax to avoid evaporation or diffusion. The bottled samples were transported to the Lanzhou Institute of Glaciology and Geocryology, and kept in a cold room at -20°C until $\delta^{18}O$ analysis was performed using a Finnigan MAT-252 Spectrometer which has an accuracy of 0.05‰. Relevant meteorological conditions such as air temperature at the beginning and the end of each precipitation event, cloud type, wind speed, wind direction, and humidity were recorded contemporaneously for each precipitation sample.

Figures 2a, 2b, and 2c show the $\delta^{18}O$ and contemporaneous T_a for every precipitation event during the period of observation at each meteorological station. Figures 2d, 2e, and 2f show the $\delta^{18}O$ of each sample plotted against its corresponding T_a, demonstrating that $\delta^{18}O$ is more negative when T_a is cooler and is less negative when T_a is warmer (Table 1). The slope for each $\delta^{18}O$-T_a relationship is shown in Figures 2d, 2e, and 2f, and the regression equation is given in Table 1. At Delingha (DLH) the slope (0.67±0.047) of the $\delta^{18}O$-T_a relationship is similar to that for polar precipitation, but at Tuotuohe (TTH) and Xining (XN) the slopes are much smaller (0.36±0.057 and 0.30±0.072, respectively). Comparison of the three data sets (Figure 2) reveals that only at DLH is $\delta^{18}O$ always below -15‰ when T_a is cooler than -2°C, which produces a pronounced bimodal distribution in the data. Note (Figure 2f) only four precipitation events occur at XN when T_a is below 0°C. TTH is much higher and colder (Table 1) than either DLH or XN, and precipitation is more frequently in the form of snowfall. However, unlike at DLH, snowfall collected at TTH when T_a is less than -2°C frequently has $\delta^{18}O$ values more enriched than -15‰, indicating less ^{18}O depletion and suggesting a source of moisture more enriched in ^{18}O. The weaker seasonality of $\delta^{18}O$ in precipitation at TTH and XN (Figure 2) arises largely from the lack of winter (dry season) precipitation (see crosses in Figure 3) and results in smaller $\delta^{18}O$-T_a slopes. These results reveal no clear indication that the $\delta^{18}O$-T_a slope is less for warmer sites as suggested by *Jocob and Sonntag* [1991]. Air temperatures are coldest at the highest station, TTH, where the $\delta^{18}O$-T_a slope is less than at DLH. These data demonstrate that precipitation on the Tibetan Plateau follows the normal $\delta^{18}O$-T_a relationship, that is, more negative $\delta^{18}O$ is associated with cooler temperatures of condensation.

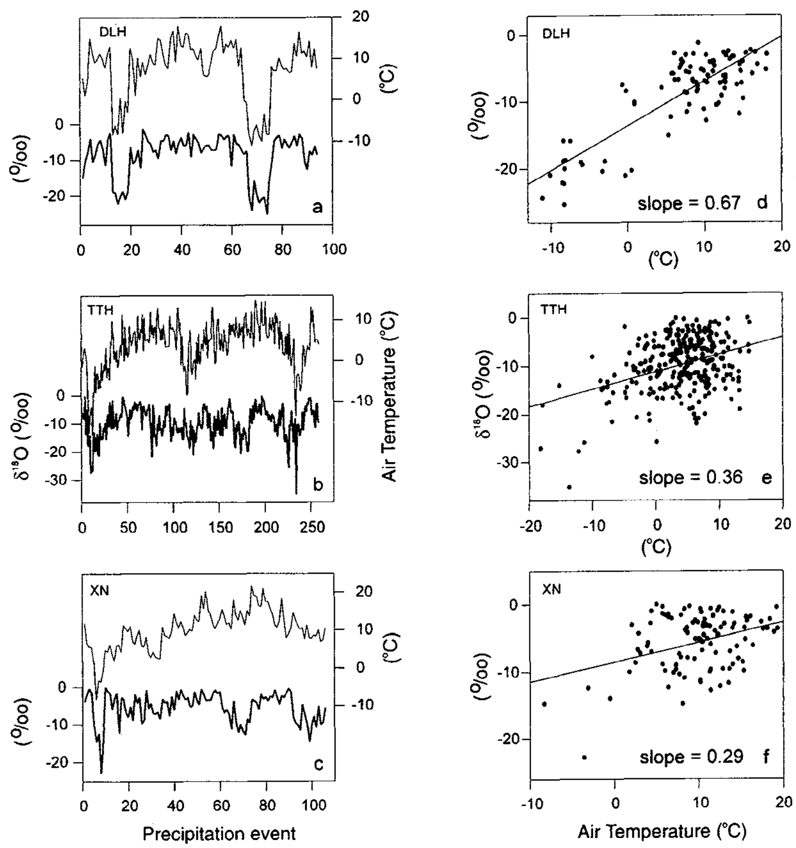

Figure 2. (a), (b), and (c) The $\delta^{18}O$ (dark line) and contemporaneous air temperature (light line) are shown for individual precipitation events at three meteorological stations on the Tibetan Plateau. (d), (e), and (f) The linear relationship between $\delta^{18}O$ and T_a (degrees Celsius) for individual precipitation events is weakly positive, and the accompanying statistics are given in Table 1.

Although the linear correlation of $\delta^{18}O$ and T_a for individual precipitation events is positive (Figures 2d, 2e, and 2f), variations in $\delta^{18}O$ exist that are not attributable to changes in T_a. Figures 2a, 2b, and 2c demonstrate that these types of variations occur more frequently at TTH and XN. The $\delta^{18}O$-T_a relationship also contains variance arising from different sources of moisture, varying air mass histories, and different temperature structures which control the condensation temperature [*O'Neill and Epstein*, 1966; *Friedman and Smith*, 1972; *Gedzelman and Lawrence*, 1982; *Kato*, 1982; *Covey and Haagenson*, 1984; *White and Gedzelman*, 1984; *Fisher and Alt*, 1985]. Another potentially important factor influencing the $\delta^{18}O$-T_a relationship is the amount effect [*Dansgaard*, 1964], which is more common in tropical regions and less important in midlatitudes [*Yurtserver and Gat*, 1981; *Yapp*, 1982; *Grootes et al.*, 1989].

A stronger relationship between $\delta^{18}O$ and T_a (Figure 3 and Table 1) emerges when individual precipitation events are aggregated into monthly averages. Averaging minimizes the influence of small-scale differences (discussed above), and as suggested by *Jouzel et al.* [1987], the $\delta^{18}O$-T_a relationship obtained this way should be more robust. Figures 3a, 3b, and 3c illustrate that $\delta^{18}O$ and T_a exhibit strong seasonal fluctuations with maximum $\delta^{18}O$ in summer and minimum $\delta^{18}O$ in winter. DLH shows the largest seasonal range (17.2‰) with a maximum in August (-3.3‰) and a minimum in February (-20.5‰). The coefficients of determination (R^2) for the $\delta^{18}O$-T_a relationship of monthly averages are much higher, all exceeding 50% (Table 1), than the R^2 values for the individual precipitation events. Undoubtedly, the strong seasonal cycle accounts for the higher R^2, but the scatter about the regression line coupled with the substantial reduction in the number of data points (degrees of freedom) results in higher errors for the slopes, particularly, for TTH and XN (Table 1). Although a seasonal cycle is present at TTH and XN, it is less pronounced largely because of the lack of winter precipitation. Again, the highest correlation ($R^2 = 0.86$) is for DLH where winter precipitation is most frequent. The slopes for TTH and XN (0.48±0.12 and 0.49±0.14, respectively) are consistent with those reported for Swiss alpine stations [*Siegenthaler and Oeschger*, 1980], but the slope for DLH (0.76±0.075) is more consistent with that of polar precipitation as discussed earlier.

The precipitation collection program is continuing at these meteorological stations to produce longer records necessary for study of the $\delta^{18}O$ relationship with temperature [e.g., *Rozanski et al.*, 1992]. The precipitation collection program began in 1991 so that the station $\delta^{18}O$ records are too short for meaningful comparison with the ice core $\delta^{18}O$ records. However, annually averaged ice core $\delta^{18}O$ records at three different sites may be compared with annually averaged surface temperatures (T_{an}) from their nearest meteorological stations. As previously mentioned, ice core $\delta^{18}O$ records are available from three sites (Figure 1) on the Tibetan Plateau. In the deep cores from Guliya (6710 m asl) and Dunde (5325 m asl), ice deposited during the Late Glacial Stage is distinguished easily from Holocene ice by ^{18}O depletion. Only a short core, 14 m long, was recovered from the Tanggula Glacier; however, it contains a 50-year record, and for all three cores, annual resolution is available in the upper part. Figure 4 shows the

Table 1. Linear Regressions Between $\delta^{18}O$ and Air Temperature for Both Individual Precipitation Events and Monthly Averages for Precipitation Collections at Three Meteorological Stations

Station	Average $\delta^{18}O$ ‰	Air Temperature °C	Individual Events: Regression Equation	Individual Events: Coefficient of Deterermination	Monthly Averages: Regression Equation	Monthly Averages: Coefficient of Determination
Delingha	-8.66	7.40	$\delta^{18}O$(‰)= (0.67±.047)T-13.59	R^2=0.69	$\delta^{18}O$(‰)= (0.76±.075)T-14.29	R^2=0.86
Tuotuohe	-9.80	3.91	$\delta^{18}O$(‰)= (0.36±.057)T-11.20	R^2=0.13	$\delta^{18}O$(‰)= (0.48±.12)T-11.09	R^2=0.45
Xining	-5.56	10.44	$\delta^{18}O$(‰)= (0.29±.072)T-8.51	R^2=0.14	$\delta^{18}O$(‰)= (0.49±.14)T-10.51	R^2=0.60

annually averaged $\delta^{18}O$ from these three ice cores (Tanggula, Dunde, and Guliya) and the contemporaneous annual surface temperatures (T_a) at the closest meteorological stations (Hotan (HT) located 200 km from Guliya, Tuotuohe (TTH) located 150 km from Tanggula, and Delingha (DLH) located 150 km from Dunde). It is evident that individual annual $\delta^{18}O$ and T_{an} values do not correlate significantly (Figure 4 and Tables 2a and 2b). The poor correlation for the individual $\delta^{18}O$ and T_{an} values results from various factors.

First, the temperature data reflect an equal weighting of monthly temperatures while the ice core $\delta^{18}O$ record is skewed toward wet season precipitation. As ≈ 80% of the precipitation falls in summer (Figure 3) and $\delta^{18}O$ is only recorded during precipitation, it is unrealistic to expect the $\delta^{18}O$ from an annual layer in a single ice core to record the annual average air temperature of the corresponding year. Second, after deposition, snowfall is subject to drifting and redeposition which introduces small-scale, local depositional noise. The original isotopic signal may be smoothed by isotopic diffusion within the upper firn layers. Even in the polar regions the

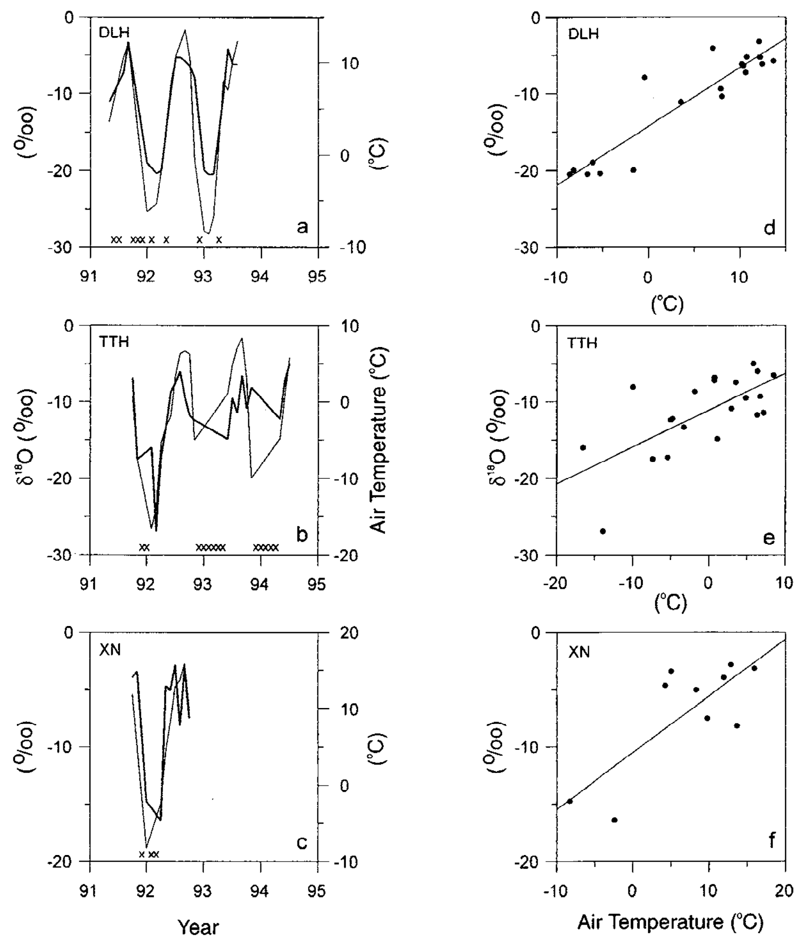

Figure 3. (a), (b), and (c) The monthly averages of $\delta^{18}O$ (dark line) and T_a (degrees Celsius) (light line) for individual precipitation events are shown. (d), (e), and (f) The linear relationship between monthly averages of $\delta^{18}O$ and T_a (degrees Celsius) are strongly positive (Table 1). Each cross indicates a month for which no precipitation occurred.

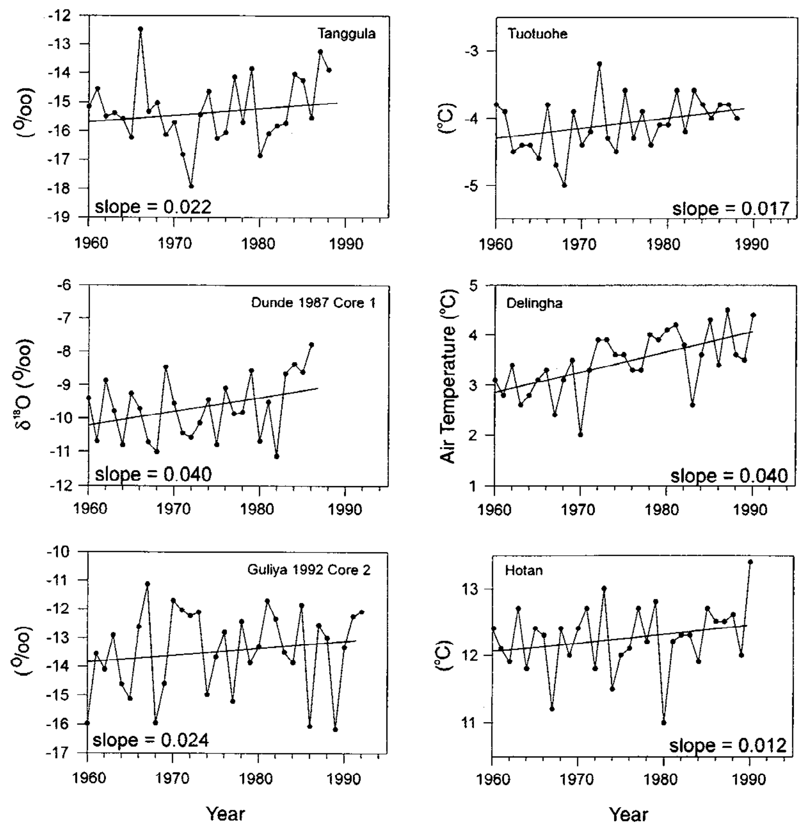

Figure 4. The annual $\delta^{18}O$ from cores on the Guliya, Tanggula, and Dunde ice caps are compared with correlative annual air temperatures since 1960 as measured at the closest meteorological stations (Figure 1). The linear regression and its slope (Table 2) are shown for each data set.

average $\delta^{18}O$ for a specific year in a single core is not well correlated with the annually averaged surface air temperatures at nearby stations. Third, these ice caps and glaciers are at least 150 km from their closest meteorological stations, and they sit at a much higher elevation than their closest station. These elevation differences (ice cap elevation - closest station elevation) are 5.3 km for Guliya, 2.3 km for Dunde, and 1.27 km for Tanggula. In addition, the meteorological stations occupy different physical environments than the ice caps. For example, Hotan is located in the desert off the Tibetan Plateau, and thus local meteorological conditions are likely to be different from those prevailing over the much higher (5.3 km) Guliya ice cap. As the length of the observation period increases, smaller-scale local influences should be dominated by larger-scale regional processes so that $\delta^{18}O$ trends from the ice core records and temperature trends at their closest station are more similar. Although the meteorological records are of limited length, the $\delta^{18}O$-T_{an} relationship is examined over the available record.

Figure 4 illustrates that snowfall on all three ice caps shows a slight ^{18}O enrichment (Table 2) from 1960 to the end of their respective records (1988 for Tanggula, 1987 for Dunde, and 1992 for Guliya). Over identical time intervals their closest stations also show modest warming trends (0.017°C yr^{-1} for TTH, 0.040°C yr^{-1} for DLH, and 0.012°C yr^{-1} for HT; Figure 4 and Table 2). Trend lines must be viewed cautiously as they are sensitive to unusually high or low values near the end points of the record.

The statistics in Table 2a confirm little statistical correlation between the annually averaged ice core $\delta^{18}O$ and T_{an} at their closest meteorological station. However, both Figure 4 and Table 2 confirm a linear $\delta^{18}O$ trend since 1960 indicating ^{18}O enrichment for all three ice caps (0.022±0.025‰ yr^{-1} for Tanggula, 0.040±0.022‰ yr^{-1} for Dunde, and 0.024±0.026‰ yr^{-1} for Guliya). Also, note that the greatest ^{18}O enrichment trend exists in Dunde snowfall, and the steepest temperature trend is recorded at its nearest meteorological station, Delingha. No further $\delta^{18}O$-T_{an} comparisons are possible because of the lack of statistical significance of the estimates (Table 2).

Finally, longer $\delta^{18}O$ records are available from the Dunde and Guliya ice caps; however, long meteorological records from the Tibetan Plateau are not available for comparison. *Thompson et al.* [1993, Figure 5] demonstrated that the 5-year running mean of the annual $\delta^{18}O$ averages from the Dunde ice cap was strongly correlated (R^2=0.25, significance is 99.9%) with the 5-year running mean of the Northern Hemisphere annual temperatures from 1895 -

Table 2a. Coefficient of Determination for Annually Averaged $\delta^{18}O$ From the Ice Cores versus Surface Air Temperatures at the Closest Meteorological Station

Record	R^2
Tanggula ($\delta^{18}O$) versusTuotuohe (T°C)	0.000
Dunde ($\delta^{18}O$) versus Delingha (T°C)	0.011
Guliya ($\delta^{18}O$) versus Hotan (T°C)	0.006

Table 2b. Linear Regression for Each of the Annually Averaged Ice Core Derived $\delta^{18}O$ Records and the Station Surface Temperature Records Shown in Figure 4

Record	Slope and Error	Error of the Estimate	R^2	Degrees of Freedom
Tanggula Glacier	0.022±0.025 ‰ yr^{-1}	1.16‰ yr^{-1}	0.02	27
Dunde Ice Cap	0.040±0.022 ‰ yr^{-1}	0.89‰ yr^{-1}	0.11	27
Guliya Ice Cap	0.024±0.026 ‰ yr^{-1}	1.54‰ yr^{-1}	0.03	31
Tuotuohe	0.017±0.008°C yr^{-1}	0.37°C yr^{-1}	0.13	27
Delingha	0.040±0.009°C yr^{-1}	0.48°C yr^{-1}	0.36	29
Hotan	0.012±0.009°C yr^{-1}	0.49°C yr^{-1}	0.05	29

1985. The long period of record (91 years) and smoothing eliminate much of the local-scale and short-term variability thus revealing the longer-term, more spatially extensive trends.

3. Conclusion

This paper presents the first observations of contemporaneous temperature and oxygen isotopic ratios for individual precipitation events on the Tibetan Plateau. The $\delta^{18}O$ composition of precipitation falling at these sites on the northern half of the Tibetan Plateau reflects seasonal variations in near-surface air temperatures. Owing to the seasonal nature of the precipitation and postdepositional modification of the seasonal signal, contemporaneous annual averages of ice core derived $\delta^{18}O$ and T_{an} from the nearest meteorological station (150 to 200 km away) are not significantly related. However, when viewed within a longer-term (multidecadal) perspective, the records show similar trends. Surface air temperatures since 1960 reveal modest warming trends, and the three ice core records indicate a modest ^{18}O enrichment. Obviously, longer records of contemporaneous $\delta^{18}O$ and near-surface air temperature for individual precipitation events will be necessary to explore their longer-term relationship. Such studies are of utmost importance if $\delta^{18}O$ records from Tibetan Plateau ice cores are to be used successfully for paleoclimatic reconstructions.

Acknowledgments. The efforts of the many individuals involved in the collection of the $\delta^{18}O$ samples and the meteorological observations are appreciated. The field work and the $\delta^{18}O$ analyses were supported by the Natural Science Foundation of China. Keith Henderson, Mary Davis and John Nagy contributed to the data analysis and graphics conducted at the Byrd Polar Research Center of The Ohio State University. This work was supported by grants from the National Science Foundation and the National Geographic Society. This is contribution number 979 of Byrd Polar Research Center.

References

Aldaz, L., and S. Deutsch, On a relationship between air temperature and oxygen isotope ratio of snow and firn in the South Pole region, *Earth Planet. Sci. Lett., 3*, 2667-2674, 1967.

Covey, C., and P.L. Haagenson, A model of oxygen isotope composition of precipitation: Implications for paleoclimate data, *J. Geophys. Res., 89*, 4647- 4655, 1984

Craig, H., Standard for reporting concentrations of deuterium and oxygen-18 in natural waters, *Science, 133*, 1833-1834, 1961.

Cuffey, K.M., R.B. Alley, P.M. Grootes, J.F. Bolzan, and S. Anandakrishnan, Calibration of the $\delta^{18}O$ isotopic paleothermometer for central Greenland, using borehole temperatures, *J. Glaciol., 40*, 341-349, 1994.

Cuffey, K.M., G.D. Clow, R.B. Alley, M. Stuiver, E. D. Waddington, and R.W. Saltus, Large arctic temperature change at the Wisconsin-Holocene glacial transition, *Science, 270*, 455-458, 1995.

Dansgaard, W., The abundance of O^{18} in atmospheric water and water vapour, *Tellus*, 5, 461-469, 1953.

Dansgaard, W., Stable isotopes in precipitation, *Tellus, 16*, 436-468, 1964.

Dansgaard, W., S.J. Johnsen, H.B. Clausen, and N. Gundestrup, Stable isotope glaciology, *Medd. øm Grönl., 197*, 1-53, 1973.

Fisher, D.A., and B.T. Alt, A global oxygen isotope model semi-empirically zonally averaged, *Ann. Glaciol., 7*, 117-124, 1985.

Friedman, I., and G.I. Smith, Deuterium content of snow as an index to winter climate in the Sierra Nevada area, *Science, 176*, 790-793, 1972.

Gedzelman, S.D., and J.R. Lawrence, The isotopic composition of cyclonic precipitation, *J. Appl. Meteorol., 21*, 1385-1404, 1982.

Grootes, P.M., M. Stuiver, L.G. Thompson, and E. Mosley-Thompson, Oxygen isotope ratio changes in tropical ice, Quelccaya, Peru, *J. Geophys. Res.*, 94,1187-1194, 1989.

Jocob, H., and C. Sonntag, An 8-year record of the seasonal variation of ^2H and 18O in atmospheric water vapour and precipitation at Heidelberg, Germany, *Tellus*, Ser. B., *43*, 291-300, 1991.

Johnsen, S.J., W. Dansgaard, and J.W.C. White, The origin of Arctic precipitation under present and glacial conditions, *Tellus, Ser. B., 41*, 452-468, 1989.

Joussaume, S., and J. Jouzel, Simulation of paleoclimatic tracers using atmospheric general models, in *Abrupt Climatic Change: Evidence and Implications*, edited by W. H. Berger and L. D. Labeyrie, *NATO ASI Ser. C, 216*, 369-381, 1987.

Joussaume, S., and J. Jouzel, Paleoclimatic tracers: An investigation using an atmospheric general circulation model under Ice Age conditions to water isotopes, *J. Geophys. Res., 98*, 2807-2830, 1993.

Jouzel, J., and L. Merlivat, Deuterium and oxygen 18 in precipitation: Modeling of the isotopic effects during snow formation, *J. Geophys. Res., 89*, 11,749-11,757, 1984.

Jouzel, J., L. Merlivat, J.R. Petit, and C. Lorius, Climatic information over the last century deduced from a detailed isotopic record in the south pole snow, *J. Geophys. Res., 88*, 2693-2703, 1983.

Jouzel, J., G. L. Russell, R. J. Suozzo, R. D. Koster, J.W.C. White, and W.S. Broecker, Simulations of the HDO and $H_2{}^{18}O$ atmospheric cycles using the NASA GISS General Circulation Model: The seasonal cycle for present-day conditions, *J. Geophys. Res., 92*, 14,739-14,760, 1987.

Jouzel, J., J. Suozzo, R.D. Koster, G.L. Russell, J.W. White, and W.S. Broecker, Simulations of the HDO and $H_2{}^{18}O$ atmospheric cycles using the NASA GISS General Circulation Model: Sensitivity experiments for present-day conditions, *J. Geophys. Res., 96*, 7495-7507, 1991.

Jouzel, J., C. Lorius, J.R. Petit, C. Ritz, M. Stievenard, P. Yiou, N.I. Barkov, V.M. Kotlyakov, and V. Lipenkov, The climatic record from Antarctic ice now extends back to 220 kyr BP, Long-term climatic variations, *Global Environ. Change, 22*, 213-237, 1994.

Kato, K., Oxygen isotopic composition and formation process of snow under anticyclone at Syowa Station, East Antarctica, *Mem. Natl. Inst. Polar Res. Spec Issue Jpn., 24*, 108-112, 1982.

Kotlyakov, V.M., F.G. Gordienko, N.I. Barkov, and E.S. Korotkevich, Isotopic studies of a core from Vostok station and their paleoglaciological interpretation, *Antarc. Comm. Rep., 19*, 45-53, 1985.

Lorius, C., and L. Merlivat, Distribution of mean surface stable isotope values in East Antarctica: observed changes with depth in a coastal area, in *Impurities in Snow and Ice, IAHS Publ. 118*, 127-137, 1977.

Lorius, C., J. Jouzel, C. Ritz, L. Merlivat, N.I. Barkov, Y.S. Korotkevich, and V.M. Kotlyakov, A 150,000-year climatic record from Antarctic ice, *Nature, 316*, 591-596, 1985.

Mosley-Thompson, E., L.G. Thompson, P.M. Grootes, and N. Gundestrup, Little Ice Age(Neoglacial) paleoenvironmental conditions at Siple Station, Antarctica, *Ann. Glaciol., 14*, 199-204, 1990.

O'Neill, J., and S. Epstein, A method for oxygen isotope analysis of milligram quantities of water and some of its applications, *J. Geophys. Res., 71*,4955-4961, 1966.

Peel, D.A., Ice core evidence from the Antarctic Peninsula region, in *Climate Since A.D. 1500*, edited by R.S. Bradley and P.D. Jones, pp. 549-571, Routledge, New York, 1992.

Rozanski, K., C. Sonntag, and K. Münnich, Factors controlling stable isotope composition of European precipitation, *Tellus, 34*, 142-150, 1982.

Rozanski, K., L. Araguas, and R. Gonfiantini, Relation between long term trends of oxygen 18 isotope composition of precipitation and climate, *Science, 258*, 981-985, 1992.

Siegenthaler, U., and H. Oeschger, Correlation of ^{18}O in precipitation with temperature and altitude, *Nature, 285*, 314-319, 1980.

Thompson, L.G., Ice core evidence from Peru and China, in *Climate Since A.D. 1500*, edited by R.S. Bradley and P.D. Jones, pp. 517-548, Routledge, New York, 1992.

Thompson, L.G., *et al.*, Holocene-Late Pleistocene climatic ice core records from Qinghai-Tibetan Plateau, *Science, 246*, 474-477, 1989.

Thompson, L.G., E. Mosley-Thompson, M. Davis, P.-N. Lin, T. Yao, M. Dyurgerov, and J. Dai., "Recent warming": Ice core evidence from tropical ice cores with emphasis upon Central Asia, *Global and Planetary Change Section, 7*, 145-156, 1993.

Thompson, L.G., E. Mosley-Thompson, M.E. Davis, P.-N. Lin, K.A. Henderson, J. Cole-Dai, J.F. Bolzan and K. -b. Liu, Late Glacial Stage and Holocene tropical ice core records from Huascarán, Peru, *Science, 269*, 46-50, 1995a.

Thompson, L.G., E. Mosley-Thompson, M.E. Davis, P.N. Lin, V. Mikhalenko, and J. Dai, A 1000 year ice core climate record from the Guliya Ice Cap, China and its relationship to global climate variability, *Ann. Glaciol.*, 21,165-181, 1995b.

White, J.W.C., and S.D. Gedzelman, The isotopic composition of atmospheric water vapor and the concurrent meteorological conditions, *J. Geophys .Res., 89*, 4937-4939, 1984.

Yao, T., and L.G. Thompson, Trends and features of climatic changes in the past 5000 years recorded by the Dunde ice core, *Ann. Glaciol., 16*, 21-24, 1992.

Yao, T., Z. Xie, X. Wu, and L.G. Thompson, The Little Ice Age as recorded by the Dunde Ice Cap, *Sci. Sin. Ser. B, 11*, 1196-1201, 1990.

Yao, T., Z. Xie, Q. Yang, and L.G. Thompson, Temperature and precipitation fluctuations since 1600 A.D. provided by the Dunde Ice Cap, China, in *Glacier-Ocean-Atmosphere Interactions, IAHS Publ., 208*, 1991.

Yao, T., L.G. Thompson, K. Jiao, E. Mosley-Thompson, and Z., Yang, Recent warming as recorded in the Qinghai-Tibet cryosphere, *Ann. Glaciol.*, 21, 196-200, 1995.

Yapp, C.J., A model for the relationships between precipitation D/H ratios and precipitation intensity, *J. Geophys. Res., 87*, 9614-9620, 1982.

Yurtserver, Y., and J. Gat, Atmospheric waters, in Stable isotope hydrology: Deuterium and oxygen -18 in the water cycle, *Tech. Rep. Ser. IAEA.*, 210,103-142, 1981.

T. Yao and Y. Zhihong, The Lanzhou Institute of Glaciology and Geocryology, Chinese Academy of Sciences, China, 730000 Lanzhou Gansu P.R. China. (e-mail: tdyao@ns.lzb.ac.cn)

P.-N Lin, E. Mosley-Thompson and L.G. Thompson, The Byrd Polar Research, Ohio State University, 108 Scott Hall, 1090 Carmack Road, Columbus, Ohio, 43210. (e-mail: plin@magnus.acs.ohio-state.edu; thompson.4@osu.edu; thompson.3@osu.edu)

Z. Xingping, Department of Geography, Hunan Normal University, Changsha, Hunan, China.

(Received July 24, 1995; revised July 24, 1996; accepted July 24, 1996.)