Modification of three ice-core $\delta^{18}O$ records from an area of high melt

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ABSTRACT. Stable oxygen isotope ratios ($\delta^{18}O$) of three shallow ice cores (extending back to 1963) from Ürümqi glacier No. 1 at the headwater of Ürümqi river, Tien Shan, northwest China, were used to test the relationship between $\delta^{18}O$ and contemporaneous surface air temperature (T_a). The ice cores were dated using the seasonal stable-isotopic signals, and seven insoluble particulate β-activity horizons associated with known nuclear tests. Although a strong positive relationship exists between $\delta^{18}O$ in precipitation and T_a at our study site, this relationship is not preserved between the annually averaged ice-core $\delta^{18}O$ records and the local temperature due to post-depositional modification. These results indicate that the processes forming the ice-core chemical records in areas of high melt must be understood before the $\delta^{18}O$ record can be confidently interpreted as a climatic indicator.

1. INTRODUCTION

Ice-core chemical records from areas experiencing high melt have attracted more and more attention for paleoclimatic construction because of their widespread occurrence in the world (Koerner, 1997; Grumet and others, 1998; Vehviläinen and others, 2002). However, such valuable ice-core records might be more-or-less modified by post-depositional processes. Hou and Qin (2002) compared the chemical profiles from two snow pits sampled successively at the same spot on Ürümqi glacier No. 1 (UG1) at the head of Ürümqi river, Tien Shan, northwest China, and reported the striking effect of post-depositional processes on the chemical profiles. This phenomenon is also evident for an ice core drilled from Far East Rongbuk Glacier in the Himalaya (Hou and others, 2002).

Historically, three independent precipitation-sampling experiments for $\delta^{18}O$ were performed at the Daxigou meteorological station (43.05° N, 86.49° E; 3539 m a.s.l.; shown as DMS in Fig. 1) located near UG1. The daily mean $\delta^{18}O$ values of precipitation samples collected from 9 July to 17 August 1981 became more negative with a reduction in daily mean air temperature (Watanabe and others, 1983). For all the precipitation events sampled from June 1995 to June 1996, a strong temporal relationship was also found between the monthly $\delta^{18}O$ in precipitation and monthly T_a, with linear fits as high as 0.95‰ °C^{-1} ($R = 0.98$; $p < 0.001$; Hou and others, 1999). This was further confirmed by Yao and others (1999) based on continuous precipitation sampling from May 1997 to August 1998, with a linear fit of 0.92‰ °C^{-1} ($R = 0.96$; $p < 0.001$) for monthly averages.

During May 1996, two shallow ice cores (6.03 and 6.81 m long, ~2 km away from DMS) were recovered from the accumulation zone of UG1 at an elevation of 4040 m, and another 14.08 m core was recovered near the previous drilling sites in October 1998 (shown as TS-1 , TS-2 and TS-3 in Fig. 1). In this paper, $\delta^{18}O$ of these shallow ice cores was used to examine the relationship between $\delta^{18}O$ and the contemporaneous surface air temperature recorded at DMS which has been in continuous operation since 1959.

The study site is surrounded by large deserts on the north, south and east sides. The nearest sea is located >3000 km away, so the region is dominated by classic continental climate conditions. The weather conditions in the study area are influenced by the blocking effect of the Qinghai–Tibetan Plateau. Since the jet stream maintains its west-to-east orientation along the Tien Shan on the northern side of the Qinghai–Tibetan Plateau (Reiter, 1981), it carries moisture originating from the Atlantic Ocean and/or the Mediterranean Sea to the study site (Li, 1991). The annual average surface air temperature and precipitation at DMS are –5.3°C and 440.6 mm w.e., respectively, for the period 1959–96 (Chinese Meteorological Administration, unpublished data).

2. METHODOLOGY

The ice-core samples were processed in the field by scraping with a clean stainless knife to obtain a contamination-free center sample. The TS-1 , TS-2 and TS-3 ice cores were cut into disks roughly at 3, 5 and 4 cm intervals, respectively. Samples were then transferred into polyethylene bags, melted at about 20°C and poured into pre-cleaned high-density polyethylene bottles. The tops were sealed in wax to avoid evaporation or diffusion. Bottled samples were transported to the Laboratory of Cryosphere and Environment, and kept in a cold room at –20°C until $\delta^{18}O$ analyses were performed with a Finnigan MAT-252 Spectrometer (precision 0.05‰). The results are expressed as the relative deviation of the heavy isotope content of Standard Mean Ocean Water (SMOW).

The β-activity samples are from the TS-3 core, with sample lengths ranging from 0.52 to 0.61 m for the upper 5.69 m section, and from 0.26 to 0.34 m below 5.69 m. The sample masses vary from 0.56 to 1.93 kg. The radionuclides are concentrated from the melted samples onto cation exchange filters after adding concentrated HCl (0.333 mL kg^{-1}). Each sample was filtered twice. Afterwards the filters were dried and measured for the activity of insoluble particulates in a gas-flow proportional counter (Dibb, 1992). Since retention of activity depends upon the total quality of the insoluble matter (Picciotto and Wilgan, 1963), we measured the insoluble matter of each sample by combusting the filter, and then divided the bulk activity measurement of each

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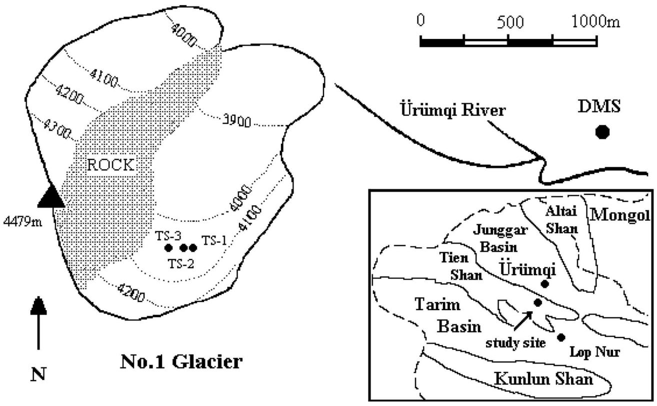

Fig. 1. Sketch map showing our study site and the locations of the Daxigou meteorological station (DMS) and the ice-core drilling sites (TS-1, TS-2 and TS-3).

sample by its corresponding insoluble matter quantity. This might eliminate the influence of the dust layers occurring at seven places from top to bottom: 0–0.06, 0.62–0.64, 4.60–4.64, 5.03–5.10, 6.89–6.97, 8.50–8.59 and 13.00–13.17 m, respectively. The β-activity profiles are plotted with and without the normalization to insoluble matter (Table 1; Fig. 2).

3. ICE-CORE DATING

The β-activity, major-ionic and $\delta^{18}O$ profiles of the cores are shown in Figure 2, together with the yields of the atmospheric nuclear tests conducted at Lop Nur (~440 km south of our study site) before 1980 (Norris and others, 1994). The close proximity and similar geographic environment of the test area and our study site, along with the regional atmospheric circulation system (Li, 1991), create favorable conditions for the artificial nuclear fallout to reach UG1. For both non-normalized and normalized β-activity profiles, the β-activity peaks at 12.92–13.24 m and at 3.46–3.98 m are evident. We suspect that the peak at 12.92–13.24 m is a 1963 reference horizon, because 57% of the total atmospheric nuclear-test yields (or some 244 Mt) were concentrated in the 16 month period September 1961–December 1962. The peak at 3.46–3.98 m is very likely associated with the Chernobyl nuclear power station accident in April 1986, as UG1 and the accident site are located close together in central Asia. Moreover, the Chernobyl horizon is also evident in the snow layers of the Greenland ice sheet (Dibb, 1989). In addition to the β-activity peaks of regional or hemispherical importance, five peaks of the normalized β-activity profiles from 10.35 to 6.57 m are likely to correspond to the five atmospheric nuclear tests at Lop Nur between 1967 and 1976. However, only two peaks are evident in the non-normalized β-activity profiles, and we attribute this to dilution by the high concentration of insoluble dust.

Based on the multiple β-activity horizons, the ice cores were further dated by counting the distinct seasonality of $\delta^{18}O$ and major ions (Fig. 2). At our study site, the $\delta^{18}O$ of summer precipitation is enriched, while winter precipitation is depleted (Hou and others, 1999; Yao and others, 1999), and high atmospheric chemical loading is observed during spring due to the transportation of continental dust stirred up by the frequent dust storms in northwest China (Sun and

Table 1. Non-normalized and normalized β-activity values of the TS-3 ice core

Top depth of each sample	Bottom depth of each sample	Water mass of each sample	Non-normalized β activity	Insoluble matter of each sample	Normalized β activity
m	m	kg	cph kg^{-1}	µg	cph µg^{-1}
0.00	0.52	1.93	163.95	65.20	4.84
0.52	1.12	0.85	146.82	36.50	3.42
1.12	1.70	0.91	56.04	54.40	0.94
1.70	2.24	0.93	71.35	32.20	2.05
2.24	2.85	0.92	71.09	34.70	1.88
2.85	3.46	0.89	124.72	35.50	3.13
3.46	3.98	0.75	152.00	15.00	7.60
3.98	4.54	0.88	73.64	19.50	3.32
4.54	5.10	0.74	109.46	27.00	3.00
5.10	5.69	0.95	70.11	19.40	3.43
5.69	5.97	0.79	82.03	11.60	5.59
5.97	6.31	0.67	68.06	5.70	8.00
6.92	7.20	0.56	54.64	8.90	3.44
7.20	7.49	0.87	39.31	5.80	5.90
7.49	7.76	0.63	76.19	8.00	6.00
7.76	8.04	0.70	72.00	6.60	7.64
8.04	8.35	0.80	74.25	13.80	4.30
8.35	8.66	0.98	91.84	49.30	1.83
8.66	8.92	0.97	39.59	13.70	2.80
8.92	9.23	0.66	47.27	15.10	2.07
9.23	9.49	1.00	46.80	13.80	3.39
9.49	9.77	1.00	42.00	10.40	4.04
9.77	10.05	1.00	48.60	21.10	2.30
10.05	10.35	0.78	48.46	10.30	3.67
10.35	10.62	1.00	67.80	22.30	3.04
10.62	10.91	1.00	61.80	28.00	2.21
10.91	11.17	1.00	52.80	19.00	2.78
11.17	11.45	1.00	31.20	9.40	3.32
11.45	11.73	1.00	36.60	12.10	3.02
11.73	12.03	1.00	32.40	20.50	1.58
12.03	12.35	0.85	76.24	31.80	2.04
12.35	12.62	1.00	40.80	10.70	3.81
12.62	12.92	1.00	28.80	9.30	3.10
12.92	13.24	1.00	152.40	24.70	6.17
13.24	13.51	0.90	59.33	20.70	2.58
13.51	13.80	1.00	66.00	33.00	2.00

others, 1998). This dating is supported by the seasonality of the major organic and inorganic ions of the TS-3 core (Lee and others, 2003). We notice a large $\delta^{18}O$ valley occurring in the 1.05–3.64 m section of the TS-3 core, which matches the nearly vertical crack cutting through from one side to the other from 1.38 to 3.86 m. Lee and others (2002) suggest that the continuous low level of $\delta^{18}O$ might be caused by the post-depositional changes from the crack. Therefore, we exclude the $\delta^{18}O$ values from the 1.05–3.86 m section of the TS-3 core, even though the annual variation might be still evident in the 3.56–3.84 m section (Fig. 2).

4. ICE-CORE $\delta^{18}O$–T_a RELATIONSHIP

We calculate the annual mean $\delta^{18}O$ value of each year by averaging all sample $\delta^{18}O$ values between two adjacent annual $\delta^{18}O$ minima, and plot the annually averaged $\delta^{18}O$ against the contemporaneous annual mean temperature and the mean temperature of the ablation season (May–September) in Figure 3. It is apparent that the ice cores did

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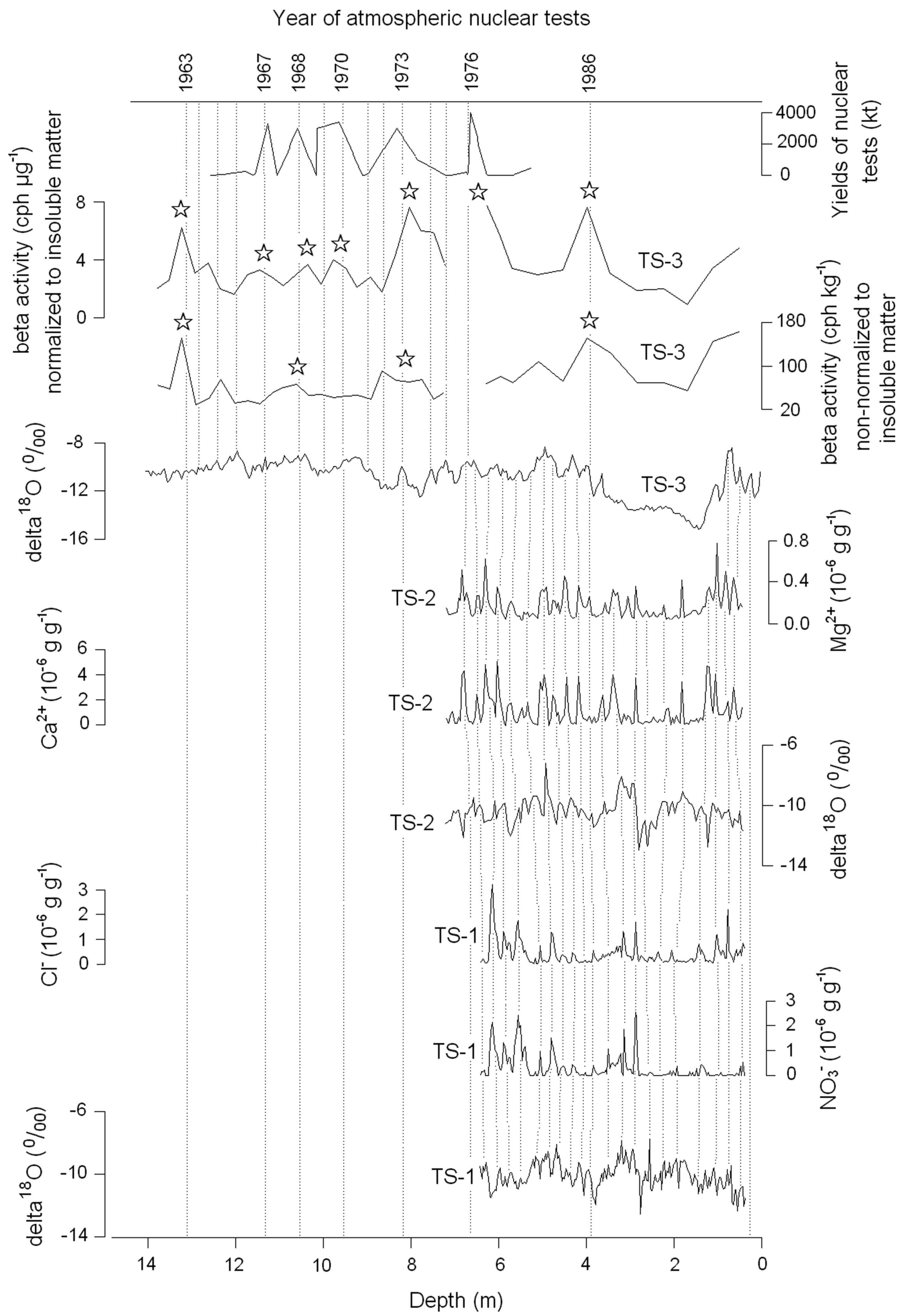

Fig. 2. The annual layers in the ice cores are indicated by the light dashed lines that include the seven radioactivity horizons of 1963, 1967, 1968, 1970, 1973, 1976 and 1986, respectively, indicated by the stars.

not preserve the strong positive relationship observed between $\delta^{18}O$ in precipitation and T_a (Watanabe and others, 1983; Hou and others, 1999; Yao and others, 1999). We calculate the correlation coefficients between the annual mean $\delta^{18}O$ and the annual mean temperature, the mean temperature of the ablation season, summer (June–August) and winter (December–February) at DMS. All the coefficients are negative (Table 2), implying that the ice-core $\delta^{18}O$ records are inversely correlated with annual mean temperature and the mean temperature of the ablation season. The disappearance of a positive $\delta^{18}O$–T_a relationship in the ice cores, as was expected from the precipitation results, might be attributed to post-depositional modification.

As suggested by Yao and others (1996), the air-temperature data reflect an equal weighting of monthly temperatures, while the ice-core $\delta^{18}O$ record is skewed toward wet-season precipitation. Thus, it is unrealistic to expect the average $\delta^{18}O$ from an annual layer in an ice core to record the annual average air temperature of the corresponding year. In the dry snow and firn layers, the isotopic homogenization is caused by recrystallization of the grains via the vapor phase, and diffusion in the vapor phase also causes considerable interstratificial mass exchange (Dansgaard and others, 1973). In low-accumulation areas, the seasonal $\delta^{18}O$ oscillations are simply missing due to redistribution by snowdrifting or lack of winter (or summer) snow, which introduces small-scale and local depositional noise (Fisher and others, 1983). When snow particles are in contact with meltwater, isotopic fractionation takes place at the interface. The effect of snowmelt percolation on the $\delta^{18}O$ records was examined in

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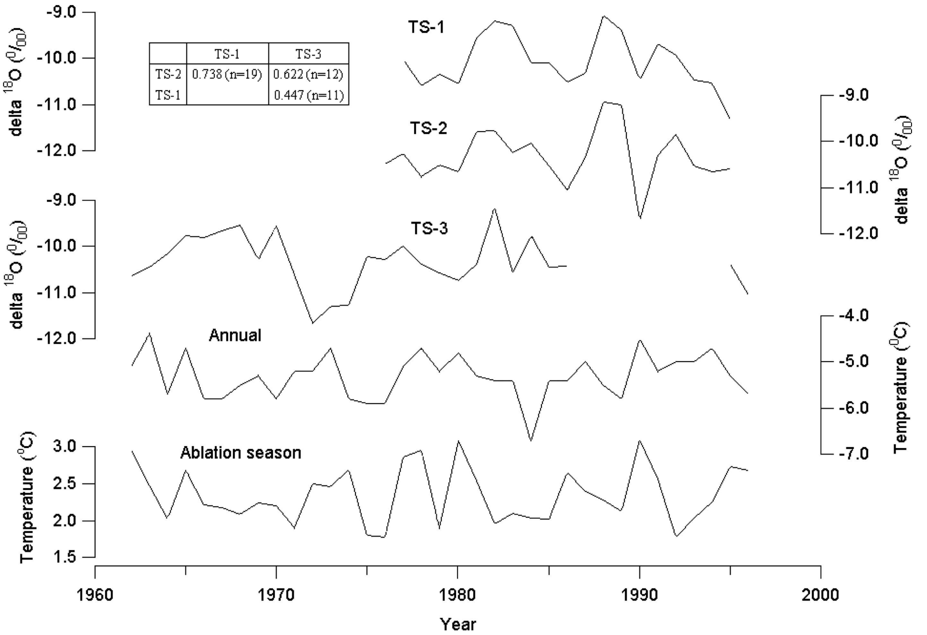

Fig. 3. The annual mean δ18O profiles of the TS-1, TS-2 and TS-3 ice cores, compared with the annual mean temperature and the mean temperature of the ablation season (May–September) recorded at DMS.

detail by comparing the δ18O profiles of a set of successive snow pits sampled at UG1 near our drilling site during the 1996 ablation season (Hou and others, 1999).

Annual average net accumulation rates constructed from TS-1, TS-2 and TS-3 are 284, 275 and 357 mm w.e., respectively, while the annual average precipitation at DMS is 440.6 mm for the period 1959–96. The precipitation amount at the ice-core drilling sites should be larger than that at DMS owing to increase of precipitation with elevation (Wang and Zhang, 1985). Therefore, at least one-fifth of the accumulation has been removed as meltwater runoff. As a result, the seasonal δ18O oscillations in precipitation rapidly diminish in the presence of percolating meltwater (Arnason, 1969). The δ18O values of the precipitation samples collected at DMS from June 1995 to June 1996 (Hou and others, 1999) and from May 1997 to August 1998 (Yao and others, 1999) range from –38.24‰ to 0.97‰ and from –30‰ to near zero, respectively. However, the δ18O values of the ice-core samples have much smaller ranges: –12.57‰ to –7.74‰ for TS-1; –12.96‰ to –7.20‰ for TS-2; and –12.79‰ to –8.32‰ for TS-3 (excluding the 1.05–3.86 m section).

Koerner (1997) concluded that the melting in early summer will affect the very negative winter/spring snow, which lies near the surface. The meltwater then percolates down to refreeze within, or at the base of, the current annual snowpack. Continued melting may result in the run-off of the less negative snow deposited during the early winter/fall period, leaving a snowpack that is very negative. Therefore, we speculate that the fraction of annual ablation against bulk precipitation will increase during warm summers and, as a result, the winter precipitation with more negative δ will increase its fractional contribution to the annual net accumulation. If our hypothesis is correct, then a lower annual δ18O value is expected when averaging the very negative winter snowpack and the remaining summer snow. In contrast, cool summers will preserve a thicker layer of the less negative summer snow at the surface, giving a disproportionately warm δ value for the annual layer (Koerner, 1997). Therefore, we suggest that this scenario could result in an inverse relationship between the ice-core δ18O and temperature records in areas of warm temperatures and high melt.

Other post-depositional processes (e.g. evaporation or condensation) can further complicate the ice-core δ18O–T_a relationship, resulting in the low correlation coefficients between the annual mean δ18O and temperature as shown in Table 2. A quantitative analysis of the effects of all the possible post-depositional processes on the final ice-core δ18O records is beyond the scope of the current paper, but it is reasonable to conclude that post-depositional processes have the potential to compromise or even obliterate the climatological information preserved in the ice-core δ18O records from areas experiencing high melt.

5. CONCLUDING REMARKS

Ice-core δ18O records from areas of high melt might be modified by post-depositional processes. Although such effects are unlikely to have occurred during glacial stages, they could become important during interglacials or interstadials with temperatures as warm as (or warmer than) at

Table 2. Correlation coefficients between the annual mean δ18O and annual ablation season (May–September), summer half-year (April–September) and winter half-year (October–March) average temperatures

	Annual	Ablation season	Summer	Winter
TS-1 ($n = 19$)	–0.375	–0.439*	–0.376	–0.075
TS-2 ($n = 20$)	–0.496*	–0.488*	–0.309	–0.159
TS-3 ($n = 27$)	–0.294	–0.348	–0.196	–0.004

Notes: *n* stands for the sample numbers. * indicates significant at $p = 0.05$.

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present. For instance, the $\delta^{18}O$ increased to its maximum (most enriched) of the whole Guliya ice core during 16.5–13.5 kyr (Thompson and others, 1997). Our results necessitate a better understanding of the processes that govern the formation of ice-core $\delta^{18}O$ records from areas of high melt before the present-day relationship of $\delta^{18}O$ in precipitation and temperature can be confidently applied for paleo-reconstruction. In addition, our results suggest that there is an urgent need to recover the valuable ice-core records from these mountainous ice fields before their valuable paleohistories are completely destroyed by enhanced snowmelt that will occur if temperatures continue to rise.

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