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Stable-isotopic composition of precipitation over the northern slope of the central Himalaya

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ABSTRACT. Stable-water-isotope data (δD and δ^{18}O) from three groups of samples (fresh-snow and snow-pit samples collected on Qomolangma (Mount Everest) and Xixabangma during field seasons 1997, 1998 and 2001, and precipitation samples collected at Tingri station during summer 2000) are presented and used to survey the isotopic composition of precipitation over the northern slope of the central Himalaya. Multi-year snow-pit samples on Qomolangma have a local meteoric water-line (slope = 8) close to the global value. Deuterium excess ($d = \delta D - 8\delta^{18}O$) values at Tingri are much lower than those in fresh snow from Qomolangma, probably due to differences in moisture source and air-mass trajectories as well as local weather conditions. There is no obvious seasonal trend for d values in the Qomolangma region. A negative relationship exists between δ^{18}O and d values in both fresh snow on Qomolangma and precipitation at Tingri. Fresh-snow samples collected from different altitudes on Xixabangma allow us to investigate the altitude effect on δ^{18}O values in snow. Of four storm events, only one has an obvious altitude effect on δ^{18}O variation and a very low gradient of −0.1% per 100 m elevation.

1. INTRODUCTION

The main crest of the Himalaya forms the climatic boundary between a region dominated by the influence of the Indian summer monsoon to the south, and the relatively cold, dry and continental climate which characterizes much of the Qinghai–Tibetan Plateau to the north. The main moisture sources for the glaciers of the central Himalaya are the Indian and Pacific Oceans (Aizen and others, 1996). Major meteorological components of the Himalaya are the southwest monsoon and the Tibetan anticyclone during the summer, and the subtropical jet stream at other times (e.g. Bryson, 1986; Murakami, 1987; Tang, 1998). The isotopic composition (δ^{18}O and δD) of precipitation is influenced by the evaporation and condensation history of an air mass and is closely linked to climatic parameters such as surface air temperature, precipitation amount, and relative humidity of the atmosphere (e.g. Dansgaard, 1964; Rozanski and others, 1993). Investigating the link between the isotopic composition of precipitation and Himalayan moisture sources allows a more accurate interpretation of paleoclimate records from ice cores in the regions (e.g. Qin and others, 2000; Thompson and others, 2000).

Using data from the global network of isotopes in precipitation (GNIP), Araguás-Araguás and others (1998) defined the spatial and temporal variability of precipitation isotopes over Asia, with special emphasis on China. Recent isotopic data from precipitation over the Tibetan Plateau reported by Tian and others (2001) reveal the northward extent of the summer monsoon by spatial variations in deuterium excess (d, defined as $d = \delta D - 8\delta^{18}O$) (Dansgaard, 1964) of precipitation. Other previous work has revealed the isotopic composition of snow in high regions of the Himalaya (Wushiki, 1977; Wake and Stiévenard, 1995; Aizen and others, 1996; Kang and others, 2000). Here we report stable-isotopic data from fresh-snow and snow-pit samples collected on Xixabangma and Qomolangma (Mount Everest), as well as precipitation samples collected at the Tingri meteorological station on the northern slope of the central Himalaya (Fig. 1). All of the samples from Qomolangma and Tingri are used to define the local meteoric water-line (LMWL) and relationships between δ^{18}O and deuterium excess. Snow-pit samples from Xixabangma and Qomolangma are used to explore the seasonal variations of isotopic composition. Fresh-snow samples from Xixabangma are used to investigate the altitude effect on snowfall δ^{18}O.

2. METHODS

During August and September 1997, fresh-snow samples were collected at the camp (5800 m a.s.l.) and along the climbing route (5800–7000 m a.s.l.) in the Dasuopu glacier region on the northwest margin of Xixabangma (28°33′ N, 85°44′ E) (Fig. 1). In addition, a 5 m snow pit (snow pit 1) located in the relatively flat portion of Dasuopu glacier (7000 m elevation) was sampled.

During September and October 1998, fresh-snow samples were collected at elevations of 6300–6500 m in the East Rongbuk Glacier region on the northern slope of Qomolangma (Fig. 1). Meanwhile, a 3.3 m snow pit (snow pit 2; 6500 m elevation) was sampled. In May 2001, samples were collected from a 1.45 m snow pit (snow pit 3; 6500 m elevation) (Fig. 1). All

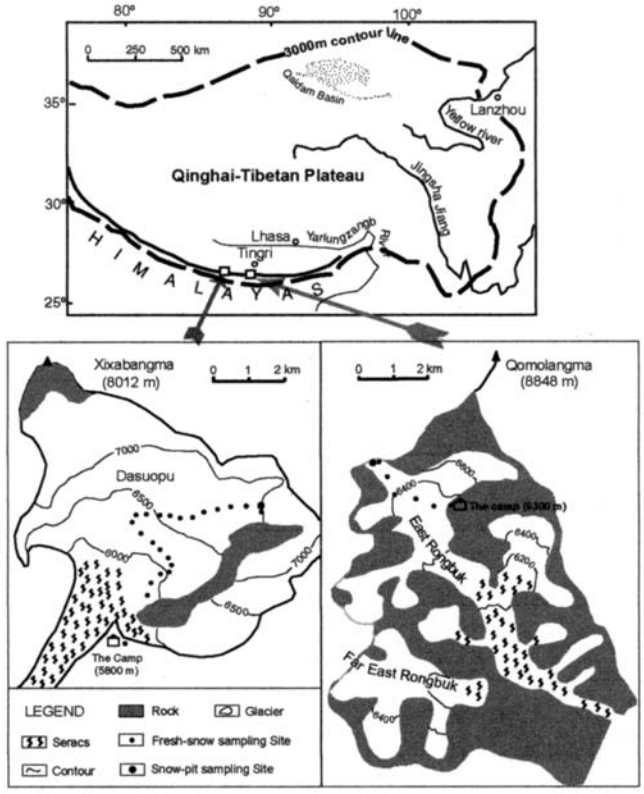

Fig. 1. Top: Location map of Xixabangma, Qomolangma (Mount Everest) and Tingri meteorological station. Bottom left: map of sampling sites of fresh snow and snow pits. Bottom right: map of northern slope of the central Himalaya.

samples were collected directly into large-mouth polyethylene bottles, kept frozen during transport, and stored in a freezer at −25°C prior to measurement of δ^{18}O and δD. The sampling interval for snow pits varies from 5 to 8 cm.

From July to September 2000, precipitation samples were collected at the Tingri meteorological station (28°35′ N, 86°37′ E; 4300 m a.s.l.) about 50 km from the Qomolangma base camp (Fig. 1). Samples were collected into a precipitation bucket. Rainfall samples were transferred directly from the bucket into polyethylene bottles. Snow and hail samples were transferred into plastic bags, and after melting were transferred into polyethylene bottles. Bottled samples were stored in a small freezer for about 1 year at a temperature of −10°C at Tingri and kept frozen for transport to a freezer at University of Maine.

Table 1 lists the details of sampling sites. Although the fresh-snow and precipitation samples collected during the summer seasons are from different years (Tingri in 2000, Qomolangma in 1998 and Xixabangma in 1997), our goal is to assess the isotopic composition of precipitation during

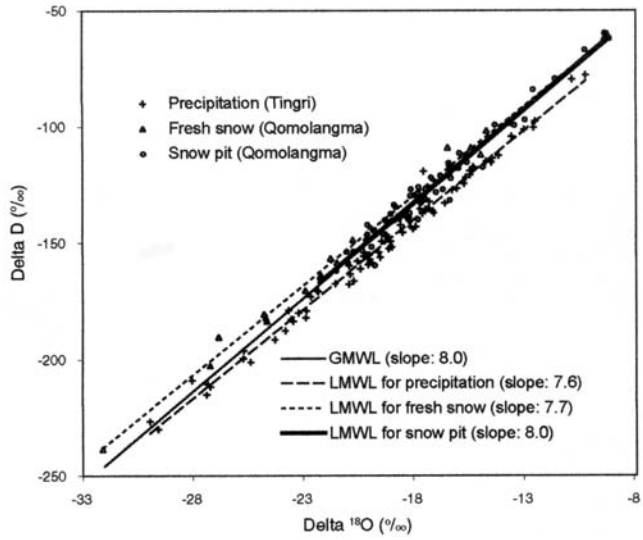

Fig. 2. The LMWL, calculated from precipitation samples at Tingri, fresh-snow and snow-pit samples on Qomolangma, compared with the GMWL ($\delta D = 8\delta^{18}O + 10$; Craig, 1961).

the summer season in this region generally, and to explore relationships between δ^{18}O and d. For snow-pit samples we focus primarily on the seasonal variations of δ^{18}O and d, and not spatial or temporal relationships between sites.

Samples were analyzed for stable-isotopic (δ^{18}O and δD) composition as follows: oxygen isotope ratios were measured via standard CO_2 equilibration techniques (e.g. Craig, 1961) with a Multiprep device coupled to a UG SIRA mass spectrometer (±0.05‰ precision); and hydrogen isotope ratios were measured via Cr reduction with a Eurovector elemental analyzer coupled to a Micromass Isoprime mass spectrometer (±0.5‰ precision; Morrison and others, 2001). All data are reported in standard delta (δ) notation vs Standard Mean Ocean Water (SMOW).

3. RESULTS AND DISCUSSION

3.1. Local meteoric water-line (LMWL)

The slopes of the LMWL calculated from individual precipitation events during the monsoon season and snow-pit samples appear in Figure 2. Samples from two snow pits on Qomolangma, each representing multi-year snow accumulation, have a slope of 8.0. This slope value is equal to the global meteoric water-line (GMWL) (Craig, 1961) and close to the value of 8.1 at Lhasa (Araguás-Araguás and others, 1998). The slope of summer precipitation in this region (7.6 for Tingri samples and 7.7 for Qomolangma samples) is slightly lower than the global value, and most of the precipitation samples from Tingri stay below the GMWL. The slope

Table 1. Sample description and isotopic composition of precipitation at three sites on the northern slope of the central Himalaya

	Tingri	*Qomolangma*		*Xixabangma*	
Location	28°35′ N, 86°37′ E	27°59′ N, 86°55′ E		28°23′ N, 85°44′ E	
Altitude	4300 m	5100–6500 m		5800–7000 m	
Sampling time	July–Sept. 2000	Sept.–Oct. 1998; May 2001 (for one snow pit)		Sept.–Oct. 1997	
Sample type	Precipitation	Fresh snow	Snow pits	Fresh snow	Snow pit
Number of samples	78	19	67	87	97
δ^{18}O (‰)*	−19.5 ± 4.2	−20.1 ± 5.5	−16.7 ± 3.1	−18.2 ± 4.5	−19.5 ± 5.4
δD (‰)*	−152.0 ± 32.3	−145.1 ± 42.4	−122.6 ± 25.0	–	–
d (‰)	+4.1 ± 4.0	+15.3 ± 4.1	+11.1 ± 4.1		

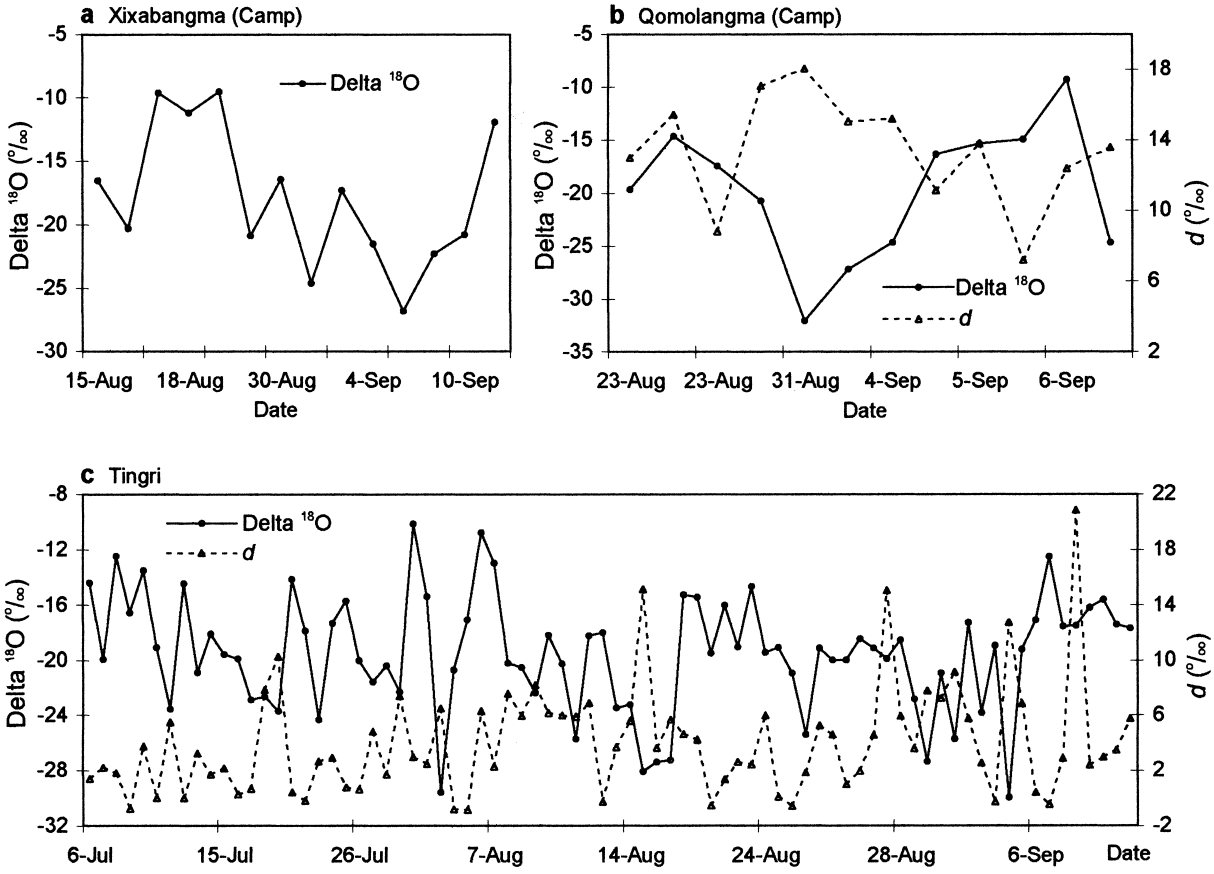

Fig. 3. Oxygen isotope composition and deuterium excess in daily snowfall/rainfall events during the summer monsoon season at: (a) the camp (5800 m a.s.l.) on Xixabangma; (b) the camp (6300 m a.s.l.) on Qomolangma; and (c) Tingri (4300 m a.s.l.).

of the LMWL from multi-year snow samples in the Qomolangma region is equal to the global value, supporting the use of deuterium excess definition ($d = \delta D - 8\delta^{18}O$) in extracting the second-order isotopic information.

3.2. Isotopic composition of fresh snow/precipitation during the summer monsoon season

Mean isotopic values (arithmetic average of all daily events) for all fresh-snow/precipitation samples collected from Qomolangma/Xixabangma and Tingri are shown in Table 1. $\delta^{18}O$ and d variations by date at the camps (5800 m on Xixabangma and 6300 m on Qomolangma) and Tingri are shown in Figure 3. During the summer monsoon season, the range of observed precipitation $\delta^{18}O$ values are similar over the study region (despite the different sampling year) and vary from around −10‰ to −30‰ in both the high-altitude regions and Tingri (Table 1). $\delta^{18}O$ values in fresh snow on Xixabangma are similar to published data from the summer season (Tian and others, 2001).

Average d in precipitation at Tingri (4.1‰) is close to the value of 5.5‰ in Tingri river water (Tian and others, 2001) and much lower than that in fresh snow from Qomolangma (Table 1). The obvious difference of precipitation d between higher-altitude glacier regions and lower-altitude sites was also observed by Tian and others (2001). Deuterium excess is a second-order isotopic parameter reflecting the non-equilibrium fractionation that occurs as moisture initially evaporates from the ocean. In this case, it is related to relative humidity and temperature over the evaporating surface, and wind speed in the source region of the atmospheric moisture. Additionally, d is associated with the air-mass trajectory, such as re-evaporation of the droplets and formation of ice crystals (e.g. Merlivat and Jouzel, 1979; Jouzel and others, 1982; Johnsen and others, 1989). Deuterium excess of snow may also vary in response to changes in the supersaturation of vapor with respect to the temperature of snow formation (Fisher, 1991; Petit and others, 1991). The difference in d between two sites may be caused by differences in moisture origins and air-mass trajectories, although Tingri is only 50 km north of Qomolangma. In the summer monsoon season, marine air masses from the Bay of Bengal move along the Brahmaputra–Yalongzangbo river (Fig. 1) and bring moisture to the Tingri region (Gao and others, 1985; Lin and Wu, 1990). During this long-distance transport, large amounts of precipitation are released due to the uplift of air masses, resulting in more negative $\delta^{18}O$ (more depletion caused by intense convection) (Wushiki, 1977; Tian and others, 1996). Considering the conditions typical for the semi-arid climate at Tingri (annual precipitation of 280 mm), the low d values may be due to partial evaporation of raindrops on the way from the cloud to the ground in an unsaturated atmosphere (Araguás-Araguás and others, 1998). On the other hand, air masses which are transported northwards from the Indian Ocean over the Indian subcontinent may travel through low passes in the Himalaya, allowing moisture to be transported to higher mountain regions (Yao, 1998). On this pathway, evaporation over the Indian subcontinent (with low relative humidity) may change the initial marine air-mass isotopic composition and result in high d in the resultant precipitation (Tian and others, 2001). When snow-formation effect is considered, d generally increases as temperatures decrease (Ciais and Jouzel, 1994). Thus, another possibility for high d in the glacier regions is the low cloud temperatures at these high altitudes.

During precipitation sampling at Tingri, measurements of

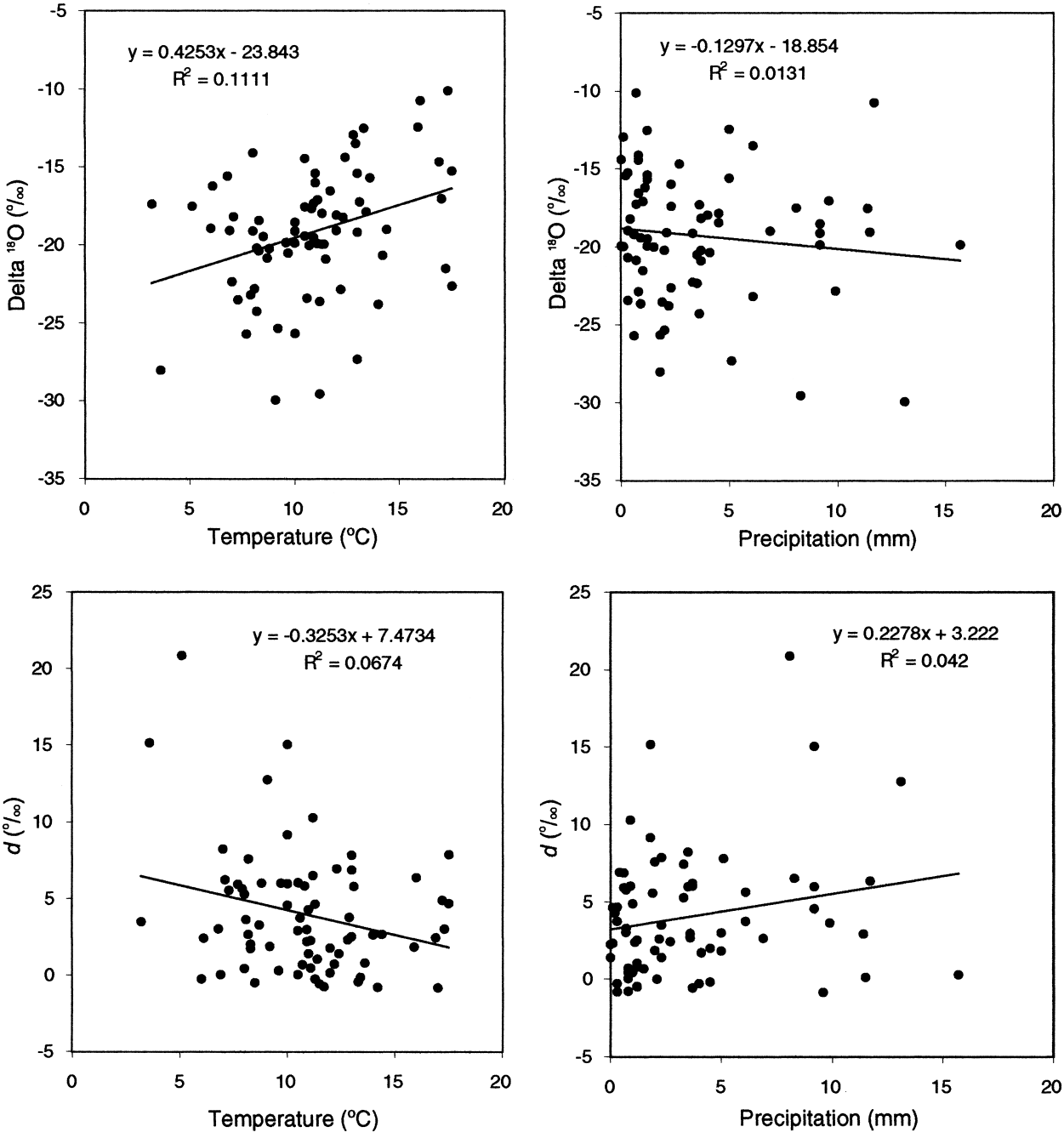

Fig. 4. Variations of $\delta^{18}O$ and deuterium excess in precipitation vs simultaneous air surface temperature and precipitation amount at Tingri during the summer monsoon season.

air temperature and precipitation amount were observed immediately after every event. There are no significant relationships between $\delta^{18}O$ and air temperature and precipitation amount for these daily storm events (Fig. 4). This indicates that simultaneous temperature and precipitation amount at the storm destination site are not the dominant factors for $\delta^{18}O$ in individual events. There are also no significant relationships between d and air temperature and precipitation amount. Previous studies have shown that monthly average $\delta^{18}O$ values are negatively related to monthly precipitation amount during the summer monsoon season in the southern Qinghai–Tibetan Plateau and Himalaya (Wushiki, 1977; Tian and others, 1996; Araguás-Araguás and others, 1998; Kang and others, 2000). Though our datasets are only for 3 months (July–September) at Tingri, this phenomenon still exists. A more negative monthly average $\delta^{18}O$ value (−20.2‰) which occurs in August corresponds to the largest monthly precipitation amount (141.2 mm) in 2000 at Tingri.

3.3. Seasonal variations of isotopic composition

Variations of $\delta^{18}O$ and d along depth profiles of the three snow pits sampled on the northern slope of the Himalaya are shown in Figure 5. Seasonal variations of $\delta^{18}O$ values in the snow profiles allow us to distinguish summer and winter layers, with more negative $\delta^{18}O$ values in summer and less negative values in winter. Several studies from tropical regions (e.g. Dansgaard, 1964; Rozanski and others, 1993), and even outside the tropics (e.g. Mount Logan in the Yukon Territory, Canada) (Holdsworth and others, 1991), have demonstrated correlation between monthly precipitation amount and isotopic content, with summer (maximum) precipitation characterized by more negative $\delta^{18}O$ values. Furthermore, at many tropical stations, mean monthly $\delta^{18}O$ is inversely correlated with monthly temperature, confirming the dominant role of the amount effect in controlling the observed seasonal variations of $\delta^{18}O$ in precipitation. Measurements of precipitation and

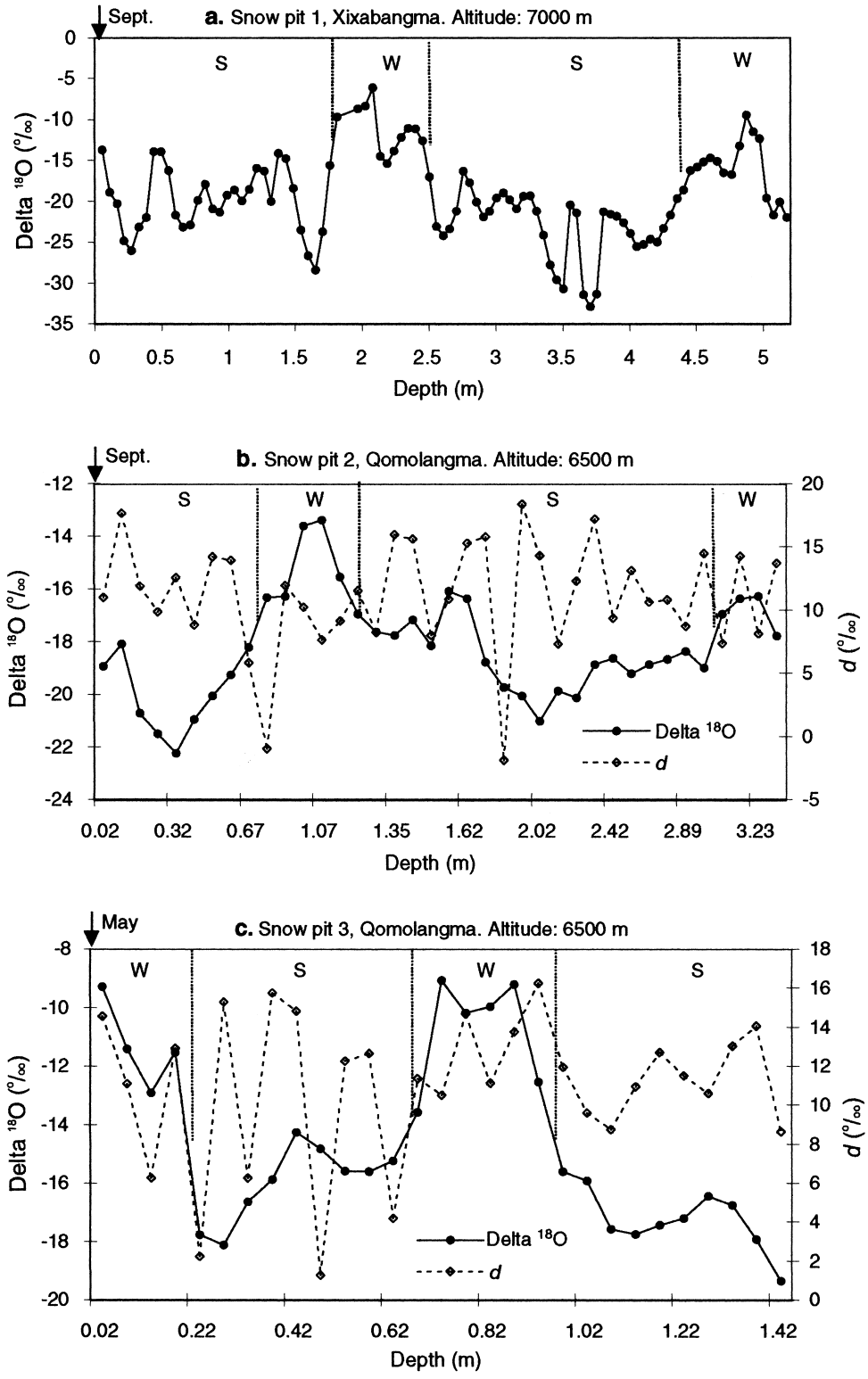

Fig. 5. Variations of δ^{18}O and deuterium excess vs depth in snow pits. (a) Snow pit 1 at 7000 m altitude on Xixabangma sampled in September 1997; (b) Snow pit 2 at 6500 m altitude on Qomolangma sampled in September 1998; (c) same as (b) but in May 2001. S represents the half-year of summer. W represents the half-year of winter. Vertical dashed lines represent the boundary of S and W.

snow-pit samples in the Himalaya and southern regions of the Tibetan Plateau clearly show that the amount effect dominates seasonal variations of δ^{18}O in precipitation (Wushiki, 1977; Tian and others, 1996; Araguás-Araguás and others, 1998; Kang and others, 2000). As for many tropical locations, more negative δ^{18}O values in snowfall on Xixabangma and Qomolangma occur during the summer monsoon period due to the strong removal of heavy-isotope species during the intense monsoon rains that spread from the Indian coast to the Himalaya. Greater depletion of ^{18}O occurs in the summer snow layers, and therefore the amount effect is clearly the primary control on the seasonal isotopic composition preserved in snow and ice layers on high-elevation glaciers on the northern slopes of the central Himalaya.

Deuterium excess values fluctuate around +11‰ for snow pits 2 and 3 on Qomolangma (Fig. 5b and c). Deuterium excess values are slightly higher in summer snow layers than in winter layers in snow pit 2 (Fig. 5b), but there are no clear seasonal trends in snow pit 3 (Fig. 5c). For both pits 2 and 3, fluctuations of *d*, ranging from around 0 to >+16‰, are much more frequent and dramatic than those of δ^{18}O. At New Delhi, south of the Himalaya, the lowest *d* values occur

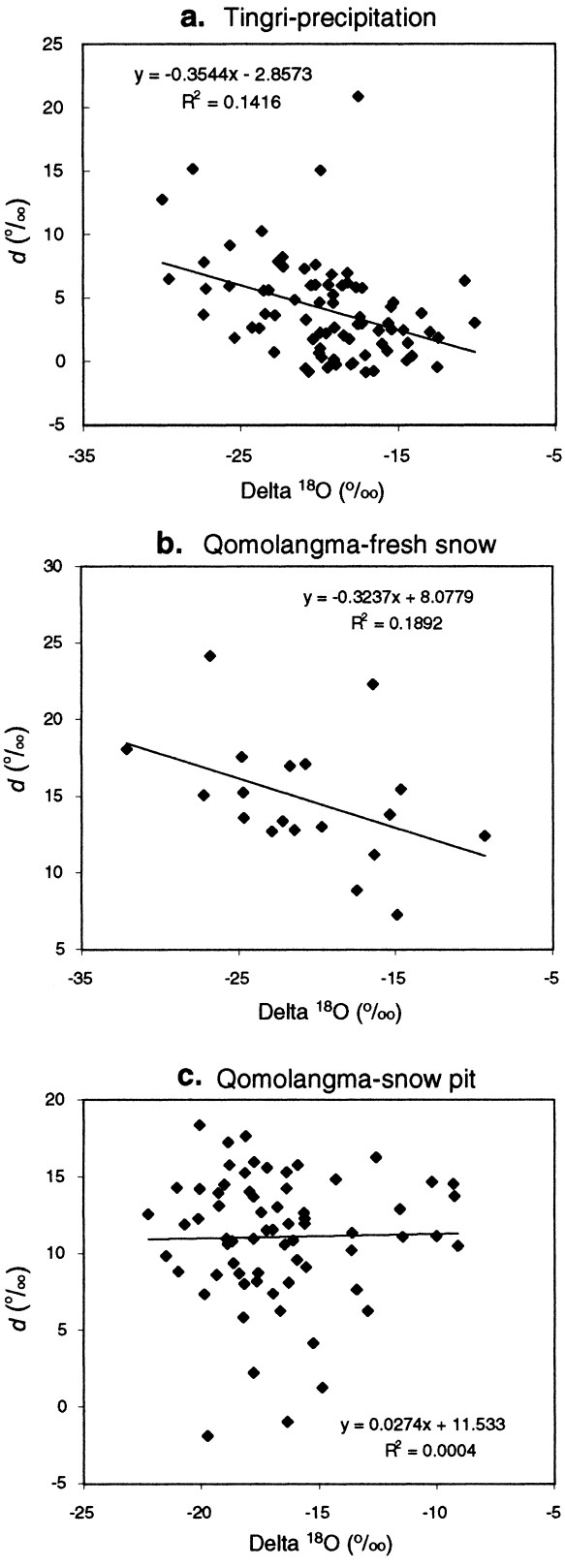

Fig. 6. Temporal variations of deuterium excess vs $\delta^{18}O$ in (a) precipitation at Tingri; (b) fresh snow on Qomolangma; and (c) snow pits on Qomolangma.

in the pre-monsoon season (May and June), and relatively stable higher values occur through the monsoon season (Araguás-Araguás and others, 1998). Inversely, at Lhasa, north of the Himalaya, lower d values occur during the monsoon season, and higher d values occur before and after the monsoon season (Tian and others, 2001). This likely indicates that the factors controlling precipitation d values vary in different regions on a seasonal basis. The complexity inherent in d variability, especially in mountainous regions (e.g. convection precipitation and seasonal changes in moisture source), may contribute to the lack of an obvious seasonal d trend for snow in the Qomolangma region.

3.4. Relationships between $\delta^{18}O$ and deuterium excess

A negative relationship exists between $\delta^{18}O$ and d values in summer precipitation on Qomolangma (correlation coefficient $r = 0.43$, significance level 97%) and Tingri ($r = 0.38$, significance level 99.9%) (Fig. 6). A previous study on snowfall samples from Xixabangma (Tian and others, 2001) has shown that when the $\delta^{18}O$ value is above −22‰, d is positively correlated with $\delta^{18}O$, whereas, when $\delta^{18}O$ is below −22‰, d is negatively correlated with $\delta^{18}O$. The difference between the two studies may reflect the different sampling time involved. Tian and others (2001) studied only about half a month from 29 July to 15 August 1996, whereas the studies at Tiangri lasted 3 months from 6 July to 20 September 2000. Alternatively, different local conditions may have been responsible. The model simulation and observation show that in the tropical convergence zone where amount effect dominates the isotopic composition in precipitation, d is a geographical minimum, with the most negative $\delta^{18}O$ value corresponding to a maximum precipitation amount (Fisher, 1990). However, this pattern does not appear in our observations. At Tingri, the decrease (more negative) in $\delta^{18}O$ values corresponds to an increase in d values in precipitation (Figs 3 and 6). This probably reflects the re-evaporation of raindrops during storm events en route, which is caused by local moisture under very dry conditions and accounts for less negative $\delta^{18}O$ values and low d values (Tian and others, 2001). No visible relationship exists between $\delta^{18}O$ and d for snow-pit samples on Qomolangma. This lack of an obvious relationship between $\delta^{18}O$ and d values may be due to more complicated factors related to snow d in different seasons, such as the changing of moisture sources (Aizen and others, 1996) and storm processes during the different seasons in the high-altitude mountain regions. These complexities could be clarified by further work with more isotope data from snow, and more associated meteorological observation in this region.

3.5. The altitude effect on fresh-snow $\delta^{18}O$ on Xixabangma

Four snowstorm events were sampled along the climbing route from the camp (5800 m) to the top of Dasuopu glacier (7000 m) (Fig. 1) on the northern slope of Xixabangma during the field season. $\delta^{18}O$ values in fresh snow are plotted vs altitude in Figure 7, and mean $\delta^{18}O$ values for each event are listed in Table 2. Only event 3 has a significant negative relationship between $\delta^{18}O$ and altitude ($r = 0.87$, significance level 97%), with a slope of approximately −0.1‰ ($\delta^{18}O$) per 100 m elevation. There is an increasing trend for $\delta^{18}O$ in fresh snow vs altitude in event 2, and no trends in events 1 and 4.

The altitude effect on the isotopic composition of precipitation has been investigated by many authors (e.g. Niewodniczański and others, 1981; Holdsworth and others, 1991). These studies have revealed a general decrease in $\delta^{18}O$ (or δD) values with increasing altitude that primarily reflects a decrease in condensation temperature as air masses are uplifted by topography over high mountains. A recent cyclone-based isotopic model shows that there is an elevation zone in which $\delta^{18}O$ is nearly constant due to a mixing of the isotopic signatures from the upper warm-air layer

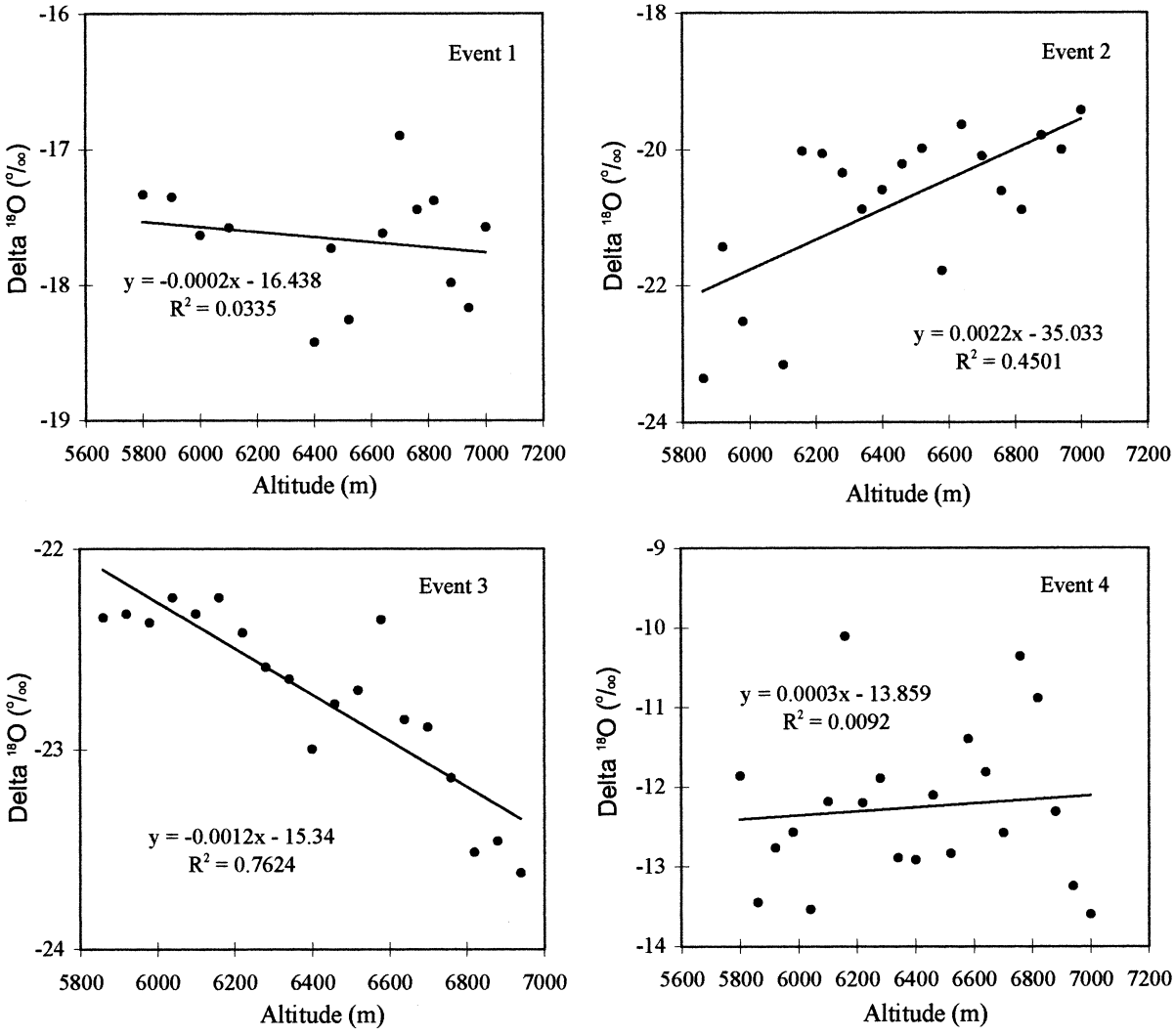

Fig. 7. Variations of $\delta^{18}O$ vs altitude for four snowfall events on Xixabangma during the summer monsoon season

and the lower cold layer (Holdsworth, 2001; Holdsworth and Krouse; 2002). Earlier work done by Niewodniczański and others (1981) shows that for $\delta^{18}O$ this effect has a gradient ranging from −0.6‰ to about −1.0‰ per 100 m elevation. Our result from event 3 reveals a gradient of −0.1‰ per 100 m, much smaller than previously published values. The lack of altitude effect for storm events 1, 2 and 4 suggests that secondary factors such as post-depositional changes in snow (i.e. melting, evaporation, snowdrift) and snowfall from a horizontal and high cloud formation may influence the altitude effect, even overcompensate, invert or enhance the effect (Moser and others, 1973; Niewodniczański and others, 1981). Our fresh-snow samples are unlikely to have been influenced by post-depositional changes, since the sampling was carried out several hours after the storms. Therefore, variations of $\delta^{18}O$ in fresh snow mainly depend on weather conditions in high mountains at the time of the storm, which may cause variability in the direction of air-mass movement. The other possibility for the lack of altitude effect is that snowfall originates from high clouds extending across the whole slope of the mountains.

Table 2. Mean $\delta^{18}O$ values for four snowfall events at Dasuopu glacier on the northern slope of Xixabangma

Event No.	*Date*	*Altitude* m	*Number of samples*	*Mean $\delta^{18}O$* ‰ (std dev.)
1	4 Sept.	5800–7000	17	−17.7 (0.4)
2	5 Sept.	5860–7000	19	−20.8 (1.2)
3	9 Sept.	5860–7000	19	−22.7 (0.4)
4	20 Sept.	5800–7000	21	−12.3 (1.0)

4. CONCLUSIONS

Data on the isotopic composition of precipitation are still very sparse over the Himalaya and the Qinghai–Tibetan Plateau, and our results extend the isotopic database in the high mountains of this region. Investigation of the isotopic composition in precipitation in reaction to the moisture sources over the Himalaya may allow us to interpret more accurately the paleoclimate records from future ice cores in the regions. The slope of the LMWL (8) for multi-year snow-pit samples on Qomolangma is close to the global value and the value at Lhasa (8.1), while samples from summer monsoon precipitation in the region have slightly lower values than the global value. During the summer monsoon season, $\delta^{18}O$ (and δD) values in precipitation are similar between high glacier regions and the lower site (Tingri). However, *d* values at Tingri are much lower than those in fresh snow from Qomolangma, probably due to differences in moisture sources and air-mass trajectories plus local weather conditions, which is consistent with the report from Xixabangma (Tian and others, 2001).

Seasonal variations of $\delta^{18}O$ in precipitation are mainly controlled by the amount effect over high glacier regions, which agrees with previous work in tropical regions, representing greater depletion of ^{18}O in the summer snow layers. There is a lack of seasonal trend for d values in the Qomolangma region, although the data from adjacent regions (Lhasa and New Delhi) show quite different patterns in the seasonal tendency of d in precipitation. Complex controls on d variability, especially in mountainous regions (e.g. convection precipitation and seasonal changes in moisture source), may contribute to the lack of an obvious seasonal trend for d in snow in the Qomolangma region. Clearly, more detailed work is needed to clarify the climatological significance of d in precipitation, and the use of d in Asian ice-core records.

At Tingri, there is a negative relationship between $\delta^{18}O$ and d values in summer precipitation (more negative $\delta^{18}O$ values correspond to increases in d values) which is slightly different from the findings at Xixabangma (Tian and others, 2001), probably due to the influence of the re-evaporation of raindrops during storm events. No obvious relationship exists between $\delta^{18}O$ and d in multi-year snow on Qomolangma, reflecting the complicated control on snow d in different seasons over high mountains.

Fresh-snow samples collected from different altitudes allow us to investigate the altitude effect on $\delta^{18}O$ values in snow. Of four storm events, only one has a significant altitude effect on $\delta^{18}O$ variation and a very low gradient of −0.1‰ per 100 m elevation. Additional factors, such as lack of air stratification due to variable directions of air-mass movement and/or high cloud covering the whole slope of the mountains, may result in a lack of altitude effect on $\delta^{18}O$ values.

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