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Water isotope variations in the snow pack and summer precipitation at July 1 Glacier, Qilian Mountains in northwest China

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This paper presents the stable isotope data of the snow pack and summer precipitation collected at the July 1 Glacier, Qilian Mountains in northwest China and analyses their relationships with meteorological factors. On an event scale, there is no temperature effect on the δ^{18}O values in the summer precipitation, whereas the amount effect is shown to be clear. By tracing the moisture transport history and comparing the precipitation with its isotopic composition, it is shown that this amount effect not only reflects the change in moisture trajectory, which is related to the monsoon activities, but is also associated with the cooling degree of vapor in the cloud, the evaporation of falling raindrops and the isotopic exchange between the falling drops and the atmospheric vapor. As very little precipitation occurs in winter, the snow pack profile mainly represents the precipitation in the other three seasons. There are low precipitation δ^{18}O ratios in summer and high ratios in spring and autumn. The Meteoric Water Line (MLW) for the summer precipitation is δD = 7.6 δ^{18}O + 13.3, which is similar to that at Delingha, located in the south rim of the Qilian Mountains. The MWL for the snow pack is δD = 10.4 δ^{18}O + 41.4, showing a large slope and intercept. The deuterium excess (*d*) of the snow pack is positively correlated with δ^{18}O, indicating that both *d* and δ^{18}O decrease from spring to summer and increase from early autumn to early spring. This then results in the high slope and intercept of the MWL. Seasonal fluctuations of *d* in the snow pack indicate the change of moisture source and trajectory. During spring and autumn, the moisture originates from continental recycling or rapid evaporation over relatively warm water bodies like Black, Caspian and Aral Seas when the dry westerly air masses pass over them, hence very high *d* values in precipitation are formed. During summer, the monsoon is responsible for the low *d* values. This indicates that the monsoon can reach the western part of the Qilian Mountains.

Tibetan Plateau, Qilian Mountains, July 1 Glacier, precipitation, snow, firn, stable isotopes, meteoric water line, deuterium excess

Many studies[1−6] have been carried out on the isotopes of the precipitation over the Tibetan Plateau since the early 1990s. These studies not only reveal the large-scale characteristics of the water cycle on the Tibetan Plateau[5], but also provide the proxy for the paliocli-matic interpretation by isotopic records preserved in ice

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cores[6−8]. However, because of the large territory, the processes of water cycling and isotopic fractionation are extremely complicated. Moreover, the present long-term observational sites are situated on the plateau surface and near the Qinghai-Tibet Highway. Data for the other areas of the plateau are very limited. This is especially true for the mountainous areas, for which, except for one or two places (e.g. the Mt. Qomolangma[9] and its adjacent Mt. Xixiabangma[10]), there are ^{18}O data from ice cores only. This impedes more precise and profound interpretations of the ice cores[11]. Therefore, further work is needed on the stable isotopes of precipitation on the Tibetan Plateau[5,6].

The Qilian Mountains, located at the northeast rim of the Tibetan Plateau, has developed many glaciers and thus is the main source of water resources for the Hexi Corridor. Since this area closely adjoins the inland arid and semiarid territories, it is important to the studies of regional water cycle, climate and environment, and thus was chosen as the right place where the first ice core was drilled in China[12,13]. Along with the ice core research, the precipitation collection for isotope research was initiated at the Delingha Meteorological Station, which is located in the south rim of the Qilian Mountains[2,5,14]. Observations[5,6] have shown that at the station, isotopic variations of the precipitation are controlled by the temperature effect, which is representative of the northern part of the Tibetan Plateau. Nevertheless, the altitude at Delingha (2981 m) is much lower than that of the Qilian Mountains (4000−5000 m on average), and the correlation of the isotope variations in precipitation between the two areas is unknown. Ice core records[11] from the Inilchek Glacier (5100 m) in the Tianshan Mountains, Kyrgyzstan, have displayed a slope of the local meteoric water line and mean deuterium excess value, which are significantly different from the global meteoric water line[15] values. This indicates the complexity of the fractionation of stable isotopes in the precipitation over central high Asia. Therefore, it is necessary to make observations on the stable isotopes in precipitation in the Qilian Mountains.

During 2002−2005, a series of fieldwork, including meteorological, hydrological, snow, glacier variation and mass balance observations, were conducted at the July 1 Glacier (97°45′E, 39°14′N (Figure 1)) in the north of the western Qilian Mountains. Part of the observed results have been published[16−18]. This paper analyses temporal variations of the stable isotopes (δ^{18}O and δD) in precipitation and firn of the glacier, and their relationships with the meteorological factors, using the data mainly obtained during the period of June to September 2002.

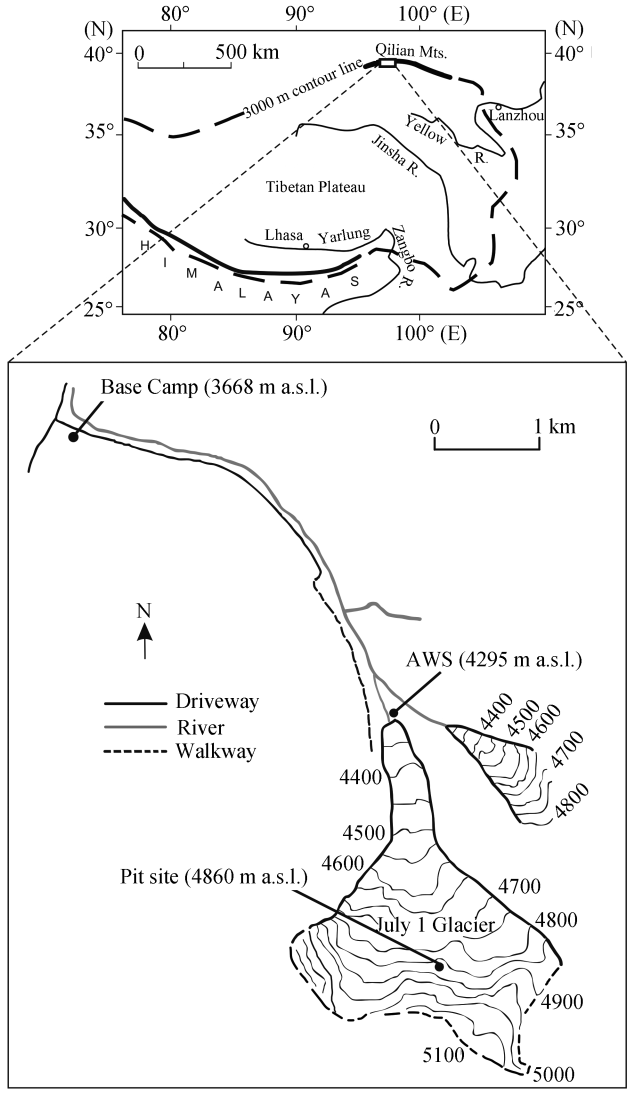

Figure 1 Location map for precipitation sampling, automatic weather station (AWS) and snow pit sites at July 1 Glacier.

1 Methods

In early June 2002, a base camp was established at the low reach of the July 1 Glacier valley (3668 m, Figure 1). An instrument shelter was set up at the camp, inside which a thermometer was placed to automatically record the air temperature. At the terminus of the glacier (4295 m), an automatic weather station was installed to record the wind speed, wind direction, air temperature, relative humidity, radiation and precipitation once every ten minutes. The precipitation of each event was collected and its time was recorded at the base camp. Also, the wind speed and direction were manually recorded several times a day. From June 16, 2002 onwards, successive snow pits were dug on the glacier and the entire depth profile at 4 cm increments was sampled at a site of 4860 m (Figure 1), about 200 m higher than the multi-year equilibrium line altitude[19]. The interval between each successive snow pit was 6—7 days. By July 17, 2002, a total of 6 pits had been dug however, on this day the firn depth decreased to 20 cm due to melting. Since there was no new snow on the profile surface of the latter four pits and the characteristics of their isotopic profiles at corresponding depths were similar[20], only the data of the first two pits are used in this analysis. On August 12, 2002, the new snow of 15 cm thickness was collected at the snow pit site. All the samples were analyzed for their oxygen-18 and deuterium ratios (with an accuracy of ±0.1‰ for δ^{18}O and ±1.0‰ for δD) at Hydrospheric Atmospheric Research Center, Nagoya University, Japan.

2 Results

2.1 Relationship between precipitation δ^{18}O and air temperature

As the water vapor condensation is controlled by temperature, the relationship between precipitation δ^{18}O and air temperature is used not only for ice core dating and paleoclimatic reconstructions[21], but also for tracing water vapor source and water cycling processes[6]. By comparing the air temperature records between the base camp and the terminus of the glacier, it is clear that their variations are highly accordant although the air temperature at the base camp is higher because of the “altitude effect”. From June 9, when the observations started, to about June 20, the air temperature increased rapidly. The temperature then fluctuated over the next few days until a maximum temperature was recorded on about July 25. The air temperature decreased in early August, and then towards the end of August, they rose again to the same high values recorded in late July. In early September, the temperature decreased regularly. The precipitation δ^{18}O of the base camp, showing no clear temporal variation trend, largely fluctuates with a maximum of 4.5‰ (June 26) and a minimum of −15.5‰ (July 15), giving a difference of 20‰.

Figure 2 shows the relationship of precipitation δ^{18}O with air temperature for all of the precipitation events in the whole observational period. It is seen from Figure 2 that the data points are scattered and no linear relationship between precipitation δ^{18}O and air temperature can be found, indicating that the water vapor source and trajectory were complicated, or that strong evaporation occurred during the rain drops falling from the cloud to the ground and its impact on the isotopic fractionation surpassed the temperature effect.

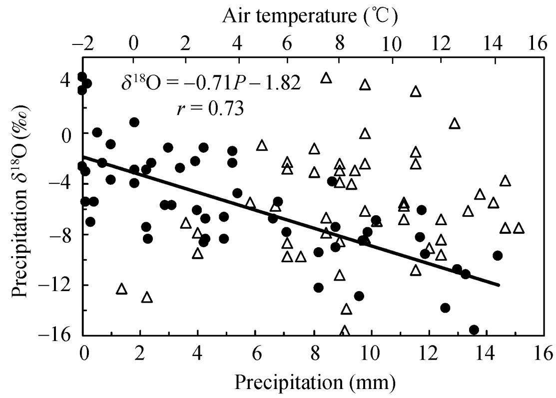

Figure 2 Summer precipitation δ^{18}O vs. air temperature (triangles) and precipitation (dots) at the base camp.

2.2 Precipitation amount effect

The amount effect is that the stable isotopic composition of meteoric water is negatively correlated with the precipitation amount[21,22]. Since the precipitation amount was not systematically measured at the base camp in 2002, the precipitation records by the AWS at the terminus of the glacier are adopted in this analysis. The elevation difference between these two sites is only 627 m, and their air temperature fluctuations followed a similar pattern. For the observed 56 precipitation events in the whole observational period, except for three when the precipitation was not captured by the AWS due to too small an amount, the duration of each event was consistent at the two sites. When looking at the wind directions between the two sites, their consistency is also evident. During the observational period, the wind directions of 43 precipitation events were manually recorded at the base camp. Of the 43 events, while no wind existed in 8 events, the wind directions in 30 events were identical at the two sites, being north or approximately north wind. Only in 5 events can the wind directions between the two sites exhibit some differences and be in the non-steady state, but during these 5 events, the wind speeds were weak. During the period of

May 26 to July 25, 2004, precipitation measurements were carried out at the base camp, and compared with those recorded by the AWS at the glacier terminus. The comparisons confirmed the co-variance of the precipitation in duration and amount of each event at the two sites, and the small altitude effect of the precipitation. In the observational period of 2004, the total precipitation were 152 mm and 131 mm at the base camp and the glacier terminus respectively[16]. Therefore, it is feasible to treat the precipitation data of the glacier terminus as those of the base camp in the analysis. Figure 2 presents the relationship between precipitation $\delta^{18}O$ and amount for all of the events, showing a clear amount effect. This effect is more evident during a single weather process. As the duration of the weather and the accumulated precipitation amount increases, the precipitation $\delta^{18}O$ of a single precipitation event becomes increasingly lower (Figure 3).

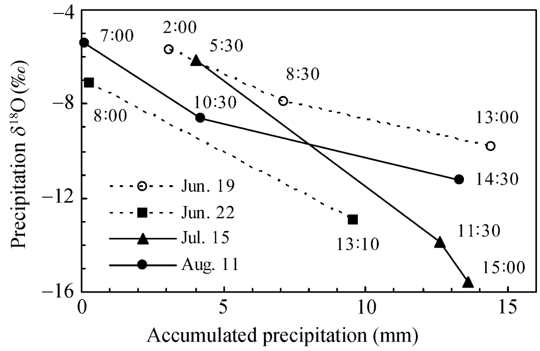

Figure 3 Temporal changes of precipitation $\delta^{18}O$ for the events in a single synoptic process, showing $\delta^{18}O$ decreasing with increasing duration and accumulated precipitation.

2.3 Fluctuation of precipitation $\delta^{18}O$ and moisture transport history

Previous studies[23] have shown that in the south of the Tibetan Plateau, the stable isotopic fluctuations of precipitation are closely related to the different moisture sources. The moisture transport history of the precipitation at the base camp in the observational period of 2002 was traced using the model devised by Tian et al.[23] and the meteorological data provided by NCAR/NCEP, including wind direction, wind speed, specific humidity and air temperature. The model, having a precision of 2.5°×2.5° (longitude×latitude), was assumed to have a particle moving in the wind field, then, based on the variations of the wind direction and speed, the location of the particle at a previous time can be determined. In the present study, the atmosphere is divided into six layers by air pressures: 600, 550, 500, 450, 400, 350 and 300 hPa, and the interval is chosen to be six hours. The outputs include the moisture trajectory, the air pressure changes of the trajectory and the specific humidity changes along the trajectory in ten days prior to a precipitation event. Figure 4 shows the modeled results in

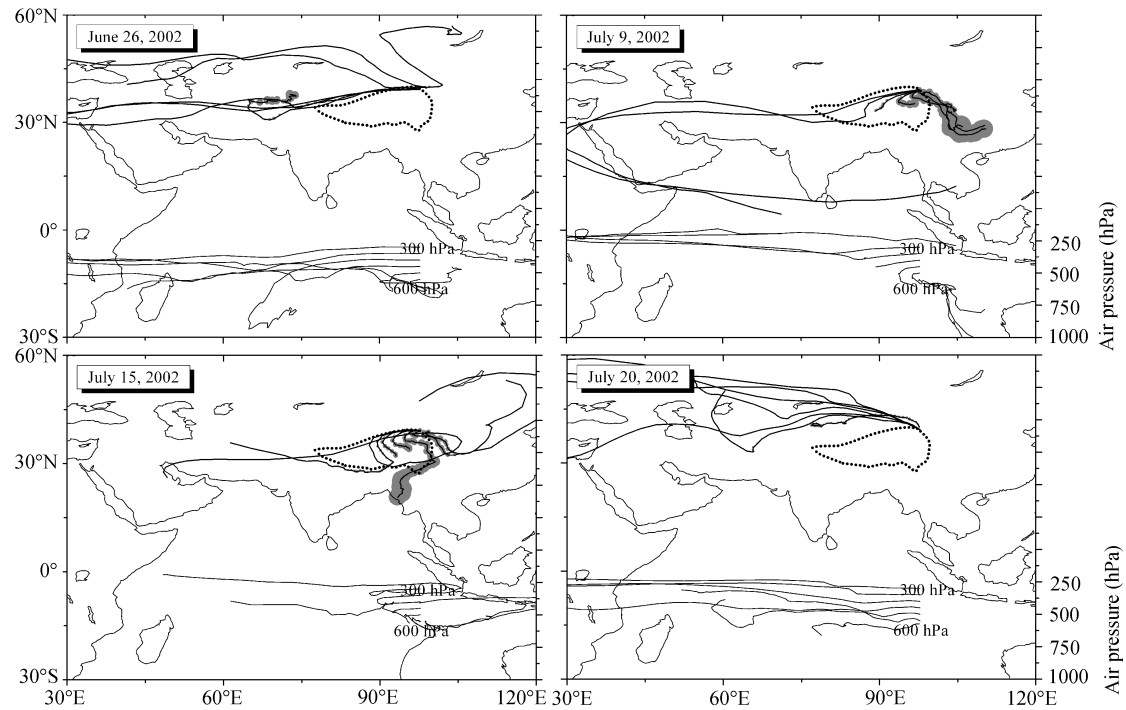

Figure 4 Traced moisture transport histories for four precipitation days (see text for details).

four precipitation days. In the figure, the curves on the upper part denote the moisture trajectories of the six layers in the previous ten days of the precipitation, and the size of the bubble areas represents the magnitude of the specific humidity of a layer. The curves on the lower part show the air pressure changes of the moisture trajectories of the layers.

The traced results of the moisture transport history of each precipitation day show that the moisture of the precipitation in the summer at the July 1 Glacier, except for a little amount originating from the west for a few days, all came from the southeast monsoon area of China and South Asia. Little moisture was transported from the north. For the stable isotopes in the precipitation, the higher isotopic ratios corresponded to the precipitation events without moisture or with little moisture transport by the west wind. The precipitation amount of these events was generally small. For example, the δ^{18}O values of the three precipitation events on June 14, July 19 and July 20, were −1.2‰, −2.2‰ and −2.4‰, respectively, corresponding to the amount of 4.2 mm, 3.9 mm and 2.4 mm respectively. Since this kind of precipitation originates from the evaporation of inland surface, the isotopic ratios are very high. When the amount of precipitation from an event of this kind is considerably small, its stable isotopic ratios become extraordinarily high. For example, the δ^{18}O values of the three precipitation events on June 26 (Figure 4), June 27 and July 4, were 4.5‰, 3.9‰ and 3.4‰, respectively and were the three highest values in the whole observational period in 2002. The precipitations of these three events were traced to be transported by the west wind, with the amount of each < 0.2 mm. The δ^{18}O values of the precipitation, of which the moisture was traced to be transported from the southeast monsoon area of China or South Asia, were obviously lower than that of the precipitation from the moisture transported by the west or local wind, with the former mainly in the range of −5—−10‰. For example, the precipitation δ^{18}O and amount on July 9 are −8.3‰ and 11.7 mm respectively, and the moisture was from southeast China (Figure 4). This range is much higher than that of most precipitation δ^{18}O values[23] in the southern Tibetan Plateau, but largely coincides with that of the average precipitation δ^{18}O values[24] at the GNIP main stations of eastern China (including Zhangye in the vicinity of the July 1 Glacier) during the monsoon period. Also, the range compares with that of monthly average values[5] of summer precipitation δ^{18}O at Delingha. These indicate that the air masses converge or exchange a large quantity of local water vapor during the course of their westward and northward transport. For instance, the precipitation δ^{18}O of −15.5‰ on July 15 was the lowest in the whole observational period. The moisture transport history (Figure 4) shows that the air masses containing moisture originating from the north coast of the Bay of Bengal first moved northward to the western part of Sichuan Province, China, and then turned to the northwest. After this, their specific humidity was largely reduced, and meanwhile, the middle and lower layers collected the water vapor from inside the plateau. Hence, while the stable isotopes of oxygen in the water vapor originating from the coast of the Bay of Bengal were extremely depleted after the long journey[23], the water vapor flux became small. When these air masses were mixed with the water vapor evaporated from inside the plateau, the precipitation δ^{18}O of the mixed air masses would thus be higher than that of the southern Tibetan Plateau in the monsoon season, but lower than that of the moisture purely from the inland of the plateau. To some extent, the lowest δ^{18}O of the day also reflects the amount effect associated with the cooling degree of vapor in the cloud, the evaporation of falling drops and the isotopic exchange between the falling drops and the environmental vapor[21,22], since the precipitation δ^{18}O for each of the three successive events in the day progressively decreased (Figure 3). The precipitation of the three events totaled 13.6 mm, being among the largest amounts of a single weather process in the observational period in 2002. Therefore, on the synoptic scale, the July 1 Glacier is still influenced by monsoons including the South Asia monsoon.

2.4 Stable isotopic profiles of firn

Figure 5 presents the density and isotopic profiles for the first (June 16, 2002) and second (June 23, 2002) firn pits at 4860 m. The firn pits were dug deep into the glacier ice. As the firn at the pit site was entirely melted from late July to early August (after this, there was short-term preservation of new snow in the melting season), and the melting season was observed to end during the period of late to middle September[16,17] in the 2002—2005 observations, all the snow at the pits should be deposited after the ending of the melting season in September 2001. The precipitation records from June 2002 to Au-

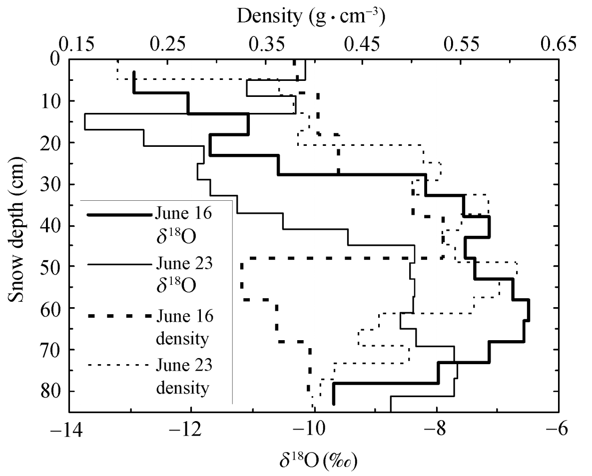

Figure 5 δ^{18}O and density profiles of the snow pack (4860 m) at the July 1 Glacier.

gust 2005, by the AWS, show that[16] there is almost no precipitation in November, December, January and February (calibrated monthly precipitation<1.2 mm), very little in October (<13.7 mm), March (<6.1 mm) and April (<12.5 mm), not much in September (21.3—37.4 mm), a large increase in May (39.7—92.0 mm), and much in June (66.5—89.7 mm), July (97.0—166.7 mm) and August (48.3—117.6 mm). Thus, the profiles of the firn pits mainly represent the precipitation of the autumn (September-October) in 2001 and the spring (March-May) in 2002, especially from May onwards. However, the water equivalent of the firn thickness of the pit on June 16 was 345.5 mm, matching the annual precipitation observed to be 344.9 mm in 2003 and 364.0 mm in 2004[16]. This may be due to snow drift, as the vicinity of the pit is relatively flat and therefore no avalanche can occur. It can be seen from the isotopic profile of June 16 (Figure 5) that from the lowest point to 25cm, the δ^{18}O increases; in the next 20 cm upwards, the δ^{18}O varies very little and after this, getting closer to the snow surface, it decreases gradually. This indicates that the precipitation δ^{18}O in the early spring is higher than in the early autumn, and it progressively decreases from the spring to the summer. The δ^{18}O profile on June 23 is similar to that of June 16, but the surface 10 cm thickness (three samples) is the new snow deposited during June 20—22, and no apparently lower δ^{18}O is seen in the bottom layer. This was due to the melting during June 16—19 and thus the meltwater was frozen to the superimposed ice in the original bottom layer. Moreover, 15 cm thick new snow, with a water equivalent of about 50 mm, was collected at the pit site on August 12, 2002 (it entirely melted before August 21). Its isotope ratios are −67.41‰ (δD) and −10.44‰ (δ^{18}O), which further demonstrates the lower stable isotopic values of the precipitation in the summer months.

2.5 Comparison of the meteoric water lines of precipitation and snow

On average, in global precipitation a linear relationship exists between the stable isotopes of oxygen and hydrogen: $\delta D = 8\delta^{18}O + 10‰$[15], termed global meteoric water line (GMWL). However, the stable isotopic ratios of a local region may represent different meteoric conditions and can be described by a local meteoric water line (LMWL) of a different slope and intercept. Thus, the LMWL and another related parameter—the deuterium excess, which is defined as $d = \delta D - 8\delta^{18}O$[21], are commonly used as indicators of water vapor source, humidity of the source, and kinetic conditions[21,25,26] in a number of fields including isotope hydrology. Figure 6 displays the MWL observed at the base camp. Its slope (7.6) and intercept (13.3) are close to the values (slope 8.3 and intercept 13.5[5]) observed in summer at Delingha. In Figure 6, owing to the amount effect, the dots of the higher isotopic ratios located in the upper segment, correspond to the light precipitation events or the initial stages of the heavy precipitation processes. Clark and Fritz[27] demonstrate that in arid or some temperate regions in summer, due to evaporation of falling

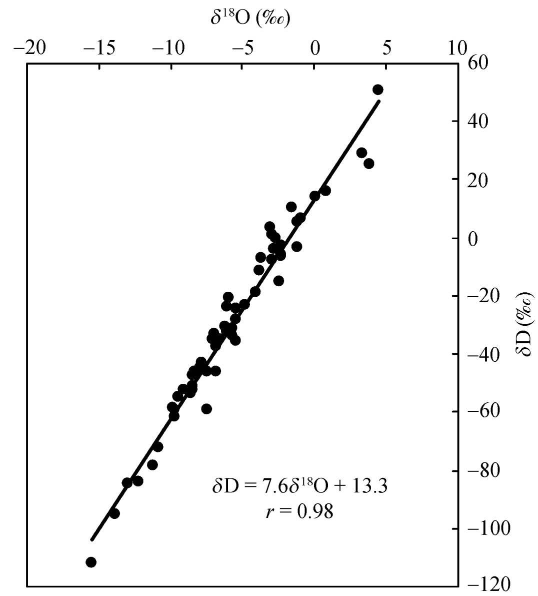

Figure 6 Meteoric water line of summer precipitation observed at the base camp.

rain drops, the amount effect would cause both the slope and intercept of the MWL to decrease. In Figure 6, the uppermost three dots represent three raining events from late June to early July, which rightly belonged to the dry period[16], and the precipitation for each of them is less than 0.2 mm. If the three dots are excluded, the slope and intercept of the new regression line would become 7.9 and 15.9, respectively, which is fairly close to the multiple-year observed values at Delingha (slope 8.4, intercept 15.0[5]). Figure 7 presents the δD–δ^{18}O relationships of the firn pack on June 16 and June 23. Both the slopes and intercepts of the lines for the two days are very high. The slope on June 16 reaches 10.4. Due to the refreezing effect[20], the slope on June 23 has largely decreased, but no less than 9.5. High slopes like these have rarely been reported. Three main post-deposition processes, evaporation, melting and refreezing may modify the slope of the MWL, but they do not cause it to increase. Evaporation[28] and refreezing[20] result in a slope decrease, and melting has no clear impact on it[20]. Therefore, the high slopes of the line for the firn pack reflect the high slope of the precipitation. This indicates a distinct difference between the slopes of the MWL in summer and in the whole year (mainly the spring, the early summer and autumn) at the July 1 Glacier.

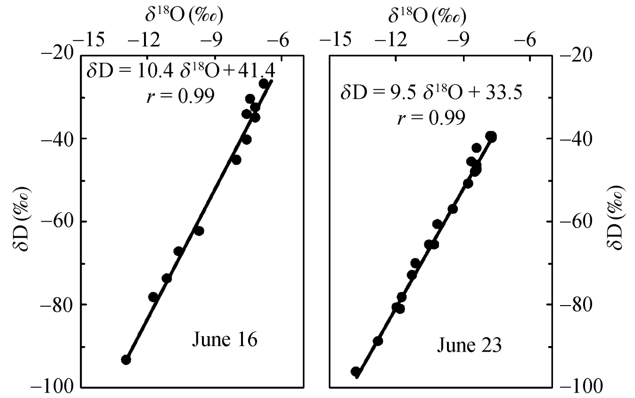

Figure 7 δD-δ^{18}O regression lines for isotopic profiles of the snow pack (4860 m) at the July 1 Glacier.

The above difference of the MWLs means differing values of deuterium excess, as is easily deduced from the definition of deuterium. Figure 8 gives the deuterium excess profiles for the two pits on June 16 and June 23 respectively. It is seen that in both profiles, the deuterium excess values are very high. They are higher not only than the monthly ratios of the multiple-year averaged precipitation over the Northern Hemisphere (<12‰[29]), but also than the monthly averaged ratios in the main precipitation months at Delingha (<about 15‰[5]). For the structure of the profiles, similar to the δ^{18}O profiles, the *d* values for the early spring are higher than those in the early autumn, and it also decreases from spring to summer. Compared with the profile of June 16, the *d* values in the middle and lower profiles of June 23 clearly decreased. This is due to the refreezing effect of the percolating meltwater[20]. The *d* values of the 10 cm new snow closest to the surface on June 23 are higher than those of the subsurface layer, but still much lower than the higher values (the highest 28.3‰) in the *d* profile of June 16. Also, the *d* value of the 15-cm new snow collected on August 12 is 16.11‰. These demonstrate the lower *d* value in summer, which is further confirmed by the fact that at the base camp the arithmetically averaged *d* of 56 precipitation events observed in summer is 15.4‰, and the weighted average is 17.0‰. This seasonality is consistent with that of the precipitation *d* at Zhangye. The multiple-year monthly averaged *d* values in every month of a year range approximately between 3 and 15‰[24], which, although being considerably lower than the *d* ratios of the firn pack (Figure 8) as a whole, shows high ratios in winter with low ones in summer. The decreasing of *d* from spring to summer is also similar to that at Delingha. Although generally the *d* of precipitation at Delingha is high in summer and low in winter , it clearly decreases from late spring to summer, which can be observed from the data by Tian et al.[5]. Figure 9 shows the *d*-δ^{18}O relationships for the firn pack in two days. It is clear, from Figure 9, that they are positively correlated. This means that from spring to summer, the *d* decreases with decreasing precipitation δ^{18}O, and that contrarily, from

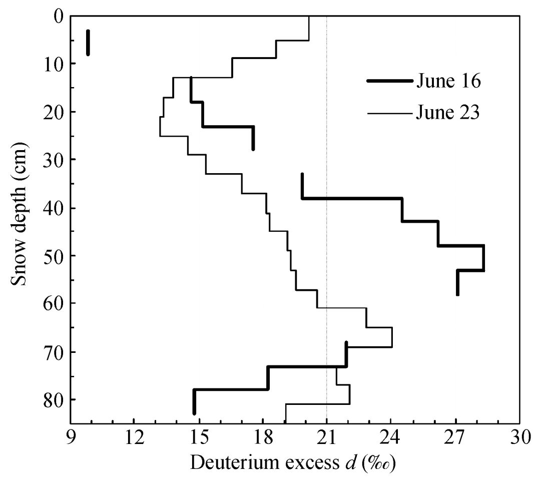

Figure 8 Deuterium excess profiles of the snow pack (4860 m) at the July 1 Glacier.

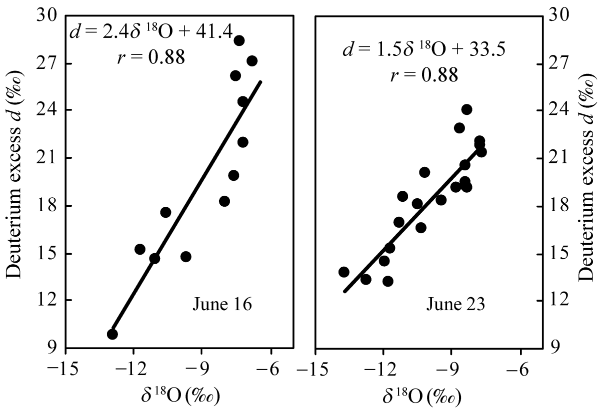

Figure 9 Deuterium excess d-$\delta^{18}O$ regression lines of isotopic profiles of the snow pack (4860 m) at the July 1 Glacier.

late autumn to early spring, the d increases with increasing precipitation $\delta^{18}O$, which consequently results in the high slope and intercept.

3 Discussion

Although the temperature effect on the isotopic composition of precipitation is evident on an event scale at Delingha[6] at the south rim of the Qilian Mountains, the multiple-year yearly[21] or monthly[24] averaged values are generally used in the analysis. While no temperature effect was observed at the July 1 Glacier, it can't be judged whether it exists or not on a long-term scale, as no winter data and only data of one summer season are available.

It is determined by the isotopic profile of the firn pack that the stable isotopic ratios of precipitation in summer are higher than in spring and autumn. This determination is based on the estimation that the deposited snow in the summer of 2001 at the pit site entirely melted. This estimation is inferred from the 2002—2005 observations. However, if the air temperature in 2001 summer was unusually low and thus most portion of the firn pack was preserved, then, contrary to this conclusion, the isotopic profile has higher ratios in summer and lower ratios in autumn and spring. Nevertheless, this possibility is extremely less because it would require a considerably low range of air temperature on a very large scale, and in fact, no abnormity of this kind has been reported. More important is that the stable isotopic ratios of the 10-cm (28.3 mm water equivalent on June 23, Figure 5) and 15-cm (about 50 mm water equivalent on August 12) new snow collected at the pit site are lower, which supports the determination of high isotopic ratios in spring and autumn and low ones in summer. However, this does not impede the existence of the temperature effect, because, if the stable isotopic ratios of winter precipitation are extremely low, in a range much lower than the differences between spring, autumn and summer, then they become high in summer and low in winter on a year scale as a whole. This is fully possible. At Delingha, the multiple-year monthly averaged $\delta^{18}O$ of precipitation in January reaches a ratio lower than −20‰[5]. Nevertheless, since there is little precipitation in winter and therefore it is difficult to be detected in a snow profile, this kind of temperature effect is meaningless for ice core interpretations. Instead, it is likely to lead to misinterpretations, namely, treating the higher isotope ratios of spring and autumn as those for summer precipitation, and then regarding the lower ratios of summer as those for the winter precipitation. Therefore, great care must be taken whilst interpreting the ice cores from the mountains on the Tibetan Plateau when limited precipitation data are available.

The above amount effect of the isotopes in precipitation is also on an event scale, although the monthly amount-weighted average values are commonly used in the analysis[21]. It is known, from the analysis of moisture transport history, that this amount effect includes two kinds of reasons. One is the differences of the moisture sources, which is similar to the season-based amount effect in the southern Tibetan Plateau and mainly related to the monsoon circulation[5,6]. Another is associated with the cooling degree of vapor in the cloud, the evaporation of falling drops and the isotopic exchange between the falling drops and the environmental vapor. As much precipitation occurs in summer and less in winter, on a year scale, the amount effect would not exist if the temperature effect is demonstrated, and vice versa.

The variations of d, along the depth profile of the firn pack, reflect the changes of the moisture sources. The $d/\delta^{18}O$'s of the precipitation in the southern Tibetan Plateau are very high before and after the monsoon season, but very low during the monsoon season, which indicates that higher d's are from continental moisture source and lower d's from a monsoon source[5,6,24]. The seasonal $d/\delta^{18}O$ variations in the firn pack at the July 1 Glacier are rightly similar to this, suggesting the change of moisture source. During winter, spring and autumn, the Qilian Mountains are controlled by the Mongolian

High[30], and the moisture originates from continental recycling or rapid evaporation over relatively warm water bodies like Black, Caspian and Aral Seas when the dry westerly-traveling air masses pass over them[11], and hence very high d values in precipitation are formed. During summer, the monsoon causes the low d values. This is coincident with the results of moisture transport tracing, and further demonstrates the monsoon characteristics of the climate in the Qilian Mountains. Although the above moisture transport tracing suggests the influences of both the southeast monsoon and Indian monsoon, probably the latter is only of synoptic significance and the climatic features are mainly related to the former, as it has been demonstrated by Tian et al.[5] that on a long-term basis, the Indian monsoon can only reach the Tanggula Range. In either case, they indicate that the westward extension of the monsoon at around 39°N is far western than the boundary determined by Araguas-Araguas et al.[24].

4 Conclusions

On an event scale, the temperature effect on the δ^{18}O values in precipitation at the July 1 Glacier is shown to be absent, while the amount effect is demonstrated to be clear. This amount effect not only reflects the change in moisture trajectory, which is related to the southeast monsoon and Indian monsoon activities, but is also associated with the cooling degree of vapor in the cloud, the evaporation of falling raindrops and the isotopic exchange between the falling drops and the environmental vapor. As very little precipitation occurs in winter, the firn pack profile mainly represents the precipitation in the other three seasons. Low precipitation δ^{18}O ratios in summer and high ratios in spring and autumn are seen. This suggests that when interpreting ice cores recovered from high mountainous areas in palioclimatic reconstructions, except the stable isotope data of modern precipitation, the seasonal distribution data of precipitation are required so as to avoid misinterpretation. The MWL for summer precipitation is δD = 7.6 δ^{18}O + 13.3, which is similar to that at Delingha, at the south rim of the Qilian Mountains. The δD–δ^{18}O line for the firn pack is δD = 10.4 δ^{18}O + 41.4, showing excessively high slope and intercept. The deuterium excess (d) is positively correlated with δ^{18}O in the firn pack, indicating that they both decrease from spring to summer, and increase from early autumn to early spring. This then results in the high slope and intercept. Seasonal fluctuations of d in the firn pack indicate the change of moisture source and trajectory. During spring and autumn, the moisture originates from continental recycling or rapid evaporation over relatively warm water bodies like Black, Caspian and Aral Seas when the dry westerly-traveling air masses pass over them, and hence very high d values in precipitation are formed. During summer, the monsoon causes the low d values. This also indicates that the monsoon can reach the western part of the Qilian Mountains.

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