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Variations of δ^{18}O in Precipitation along Vapor Transport Paths

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ABSTRACT

Three sampling cross sections along the south path starting from the Tropics through the vapor passage in the Yunnan-Guizhou Plateau to the middle-low reaches of the Yangtze River, the north path from West China, via North China, to Japan under the westerlies, and the plateau path from South Asia over the Himalayas to the northern Tibetan Plateau, are set up, based on the IAEA (International Atomic Energy Agency)/WMO global survey network and sampling sites on the Tibetan Plateau. The variations, and the relationship with precipitation and temperature, of the δ^{18}O in precipitation along the three cross sections are analyzed and compared. Along the south path, the seasonal differences of mean δ^{18}O in precipitation are small at the stations located in the Tropics, but increase markedly from Bangkok towards the north, with the δ^{18}O in the rainy season smaller than in the dry season. The δ^{18}O values in precipitation fluctuate on the whole, which shows that there are different vapor sources. Along the north path, the seasonal differences of the mean δ^{18}O in precipitation for the stations in the west of Zhengzhou are all greater than in the east of Zhengzhou. During the cold half of the year, the mean δ^{18}O in precipitation reaches its minimum at Ürümqi with the lowest temperature due to the wide, cold high pressure over Mongolia, then increases gradually with longitude, and remains at roughly the same level at the stations eastward from Zhengzhou. During the warm half of the year, the δ^{18}O values in precipitation are lower in the east than in the west, markedly influenced by the summer monsoon over East Asia. Along the plateau path, the mean δ^{18}O values in precipitation in the rainy season are correspondingly high in the southern parts of the Indian subcontinent, and then decrease gradually with latitude. A sharp depletion of the stable isotopic compositions in precipitation takes place due to the very strong rainout of the stable isotopic compositions in vapor in the process of lifting over the southern slope of the Himalayas. The low level of the δ^{18}O in precipitation is from Nyalam to the Tanggula Mountains during the rainy season, but δ^{18}O increases persistently with increasing latitude from the Tanggula Mountains to the northern Tibetan Plateau because of the replenishment of vapor with relatively heavy stable isotopic compositions originating from the inner plateau. During the dry season, the mean δ^{18}O values in precipitation basically decrease along the path from the south to the north. Generally, the mean δ^{18}O in precipitation during the rainy season is lower than in the dry season for the regions controlled by the monsoons over South Asia or the plateau, and opposite for the regions without a monsoon or with a weak monsoon.

Key words: stable isotope, vapor transport path, temperature, precipitation

1. Introduction

Precipitation is an important link in the water cycle. The abundance of stable isotopic compositions in precipitation is closely related to the weather process of generating condensation and the initial conditions in vapor origins, and fluctuates with space and time (Dansgaard, 1953, 1964; Araguas et al., 1998). Investigation of stable isotopic change in the water cycle started in about 1950 (Dansgaard, 1953). The large-scale and organized collection of precipitation samples was initiated in 1961 (Dansgaard, 1964). The Inter-

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national Atomic Energy Agency (IAEA), in cooperation with WMO, sets up a global network with more than 550 sampling stations. In China, the number of stations attached to the global IAEA/WMO survey network has reached 30. The main objective in conducting the global survey program is to determine the atmospheric circulation patterns and global or local water cycle mechanisms, and consequently, to provide basic isotope data for the use of environmental isotopes in hydrometeorological investigations for water or vapor resources inventory, by analyzing the temporal and spatial variations of stable isotopes in precipitation (Dansgaard, 1964).

The achievements in stable isotope research in atmospheric precipitation are very plentiful. The temperature effect, namely the marked positive correlation between temperature and stable isotopic ratio in precipitation, appearing mainly in continental, middle-high latitudes, is recognized as one of the most important ones (Dansgaard, 1964; Zhang and Yao, 1994; Jouzel et al., 1997; Rozanski et al., 1997). According to the temperature effect, the stable isotopes in different sediments can be quantitatively recovered as a temperature proxy (Rozanski et al., 1997; Yao, 1999; Thompson et al., 2000). The temperature effect arises from the fact that the fractionation of stable isotopes in the atmosphere and in precipitation is mainly controlled by the phase temperature in the process of phase change (Dansgaard, 1964; Jouzel et al., 1997; Zhang et al., 2003). The amount effect, namely the marked negative correlation between precipitation amount and stable isotopic ratio in precipitation, mainly appears in coastal regions and islands in low-middle latitudes. The generation of the amount effect is related to strong convective phenomena (Dansgaard, 1964; Rozanski et al., 1997; Zhang et al., 2001).

Tracing the fluctuation of stable isotopic compositions in precipitation is significant for comprehending vapor sources of different regions. This study intends to deal with the possible impact of different vapor sources on regional hydrometeorological features by analyzing stable isotopic variations and the relationship of stable isotopes in precipitation with temperature and precipitation along three vapor transport paths. The three paths are composed of sampling cross sections starting from the Tropics through the vapor passage in the Yunnan-Guizhou Plateau to the middle-low reaches of the Yangtze River (abbreviated as the south path), from west China, via North China, to Japan under the westerlies (as the north path), and from South Asia over the Himalayas to the northern Tibetan Plateau (as the plateau path).

2. Data

Because few sampling stations, and some with a short sampling time period, are attached to the IAEA/WMO network in China and in Asia, less stations are selected. In this paper, Singapore, Kosamu, Bangkok, Yangon, Luang-Pra, Kunming, Guiyang, Zunyi, Chengdu, Changsha, Wuhan, and Nanjing are the 12 IAEA/WMO stations from south to north along the south path; Ürümqi, Zhangye, Lanzhou, Yinchuan, Xi'an, Baotou, Taiyuan, Zhengzhou, Shijiazhuang, Tianjin, Yantai, Pohang, Tokyo, and Ryōrizaki are the 14 IAEA/WMO stations from west to east along the north path; and Colombo, Warangal, Bombay, Karāchi, Shillong, New Delhi, Kathmandu, Nyalam, Kyangjin, Yala, Xixabangma, Tingri, Xigaze, Lhasa, Damxung, Nagqu, Amdo, Tanggula, Tuotuohe, Wudaoliang, Xining, and Delingha are the 22 stations from south to north along the plateau path selected as the basic stations. Of these in the plateau path, Colombo, Warangal, Bombay, Karāchi, Shillong, and New Delhi located in South Asia and Lhasa in China belong to the IAEA/WMO survey network, while the others were set up by the Chinese-Japanese-American scientists for conducting large-scale and multiple scientific experiments (Zhang et al., 2001, 1995a, 1995b, 2002; Tian et al., 1997). The geographic positions and basic information for all of the above sampling stations are shown in Fig. 1 and Table 1.

The stable isotopic ratios in precipitation are reported as monthly values at the stations. The stable isotopes in precipitation are sampled on the 15th of every month for the IAEA/WMO sampling stations while monthly temperature is calculated by calendar month. Therefore, the monthly values of stable isotopic ratios in precipitation actually stand for the mean of ratios from the 16th of the previous month to the 15th of the current month (IAEA/WMO, 2001). In view of the fact that most sampling stations on the Tibetan Plateau have poor data coverage (see Table 1), the mean annual temperature and precipitation at the sampling stations on the Tibetan Plateau in China are calculated according to the meteorological data from 1961 to 2000.

Here, the measured ratio of the stable oxygen isotope in precipitation is expressed as parts per thousand of its deviation relative to the Vienna Standard Mean Ocean Water (V-SMOW), defined by

$$\delta^{18}\mathrm{O} = (R_{\mathrm{s}}/R_{\text{V-SMOW}} - 1) \times 1000 \,, \tag{1}$$

where R_{s} and $R_{\text{V-SMOW}}$ stand for the stable isotopic ratio $^{18}\mathrm{O}/^{16}\mathrm{O}$ in the precipitation sample and in V-SMOW, respectively. The weighted mean of the $\delta^{18}\mathrm{O}$ in precipitation $\overline{\delta^{18}\mathrm{O}}$ is calculated according to the equation

$$\overline{\delta^{18}\mathrm{O}} = \sum P_i \delta^{18}\mathrm{O}_i / \sum P_i \,, \tag{2}$$

Table 1. Summary of basic statistic data for the selected sampling stations.

Station	Position	Altitude (m a.s.l.) (m)	Mean annual temperature (°C)	Mean annual precipitation (mm)	Weighted mean annual δ^{18}O ‰	Sampling period
Singapore	1.35°N, 103.9°E	32	26.3	2164.2	–7.29	1968–1976
Kosamu	9.28°N, 100.03°E	7	27.8	1530.8	–5.72	1979–1983
Bangkok	13.73°N, 100.5°E	2	28.2	1488.6	–6.60	1968–1997
Yangon	16.77°N, 96.17°E	20	27.2	2419.3	–4.56	1961–1963
Luang-Pra	19.88°N, 102.13°E	305	25.7	1286.5	–7.49	1961–1964
Kunming	25.01°N, 102.41°E	1892	15.4	1008.2	–10.26	1986–93; 1995–99
Guiyang	26.35°N, 106.43°E	1071	15.3	964.6	–8.33	1988–1992
Zunyi	27.7°N, 106.88°E	844	15.4	972.9	–8.32	1986–1992
Chengdu	30.67°N, 104.02°E	506	16.2	840.0	–7.05	1986–94; 1996–98
Changsha	28.12°N, 113.04°E	37	17.2	1271.0	–5.65	1988–1992
Wuhan	30.62°N, 114.13°E	23	18.1	1224.1	–6.67	1986–88; 1996–98
Nanjing	32.18°N, 118.18°E	26	14.7	1192.1	–8.41	1987–1992
Ürümqi	43.78°N, 87.62°E	918	7.6	286.1	–10.76	1986–92; 1995–98
Zhangye	38.93°N, 100.43°E	1483	7.8	135.2	–6.48	1986–92; 1995–96
Lanzhou	36.05°N, 103.88°E	1517	10.4	328.3	–6.62	1988–1992
Yinchuan	38.29°N, 106.13°E	1112	8.9	220.7	–6.62	1988–1992
Xi'an	34.3°N, 108.93°E	397	13.5	547.2	–7.41	1985–1992
Baotou	40.67°N, 109.85°E	1067	7.5	277.8	–7.82	1986–1992
Taiyuan	37.78°N, 112.55°E	778	10.3	415.0	–7.04	1986–1988
Zhengzhou	34.72°N, 113.65°E	110	14.2	574.9	–7.04	1986–1988
Shijiazhuang	38.02°N, 114.25°E	80	13.9	558.9	–8.05	1985–92; 1995–99
Tianjin	39.06°N, 117.1°E	3	14.2	513.5	–7.66	1988–1992
Yantai	37.53°N, 121.4°E	47	12.9	566.6	–7.24	1986–1991
Pohang	36.03°N, 129.38°E	6	9.5	1121.6	–7.76	1961–67, 1972–76
Tokyo	35.68°N, 139.77°E	4	15.5	1384.1	–7.30	1961–1979
Ryōri-zaki	39.02°N, 141.5°E	260	10.0	1280.8	–8.28	1979–1986
Colombo	6.54°N, 79.524°E	7	27.6	2304.1	–3.54	1986–88; 1992–95
Warangal	18.19°N, 79.44°E	550				Jul–Nov 1977
Bombay	18.9°N, 72.82°E	10	27.3	1984.8	–1.51	1960–1978
Karāchi	24.9°N, 67.13°E	23	25.7	195.1	–3.80	1961–1975
Shillong	25.57°N, 91.88°E	1598	16.8	2064.0	–4.91	1969–1978
New Delhi	28.58°N, 77.2°E	212	25.0	805.9	–5.86	1961–1996
Kathmandu	27.7°N, 85.37°E	1340	18.6	1259.5	–6.15	1993; 1995–96
Nyalam	28.18°N, 85.97°E	3810	3.5	652.0	–10.19	1997–1998
Kyangjin	28.2°N, 85.57°E	3880	1.6	649.2	–14.33	1993–1996
Yala	28.23°N, 85.6°E	5110				May–Oct 1996
Xixabangma	28.45°N, 85.75°E	5680				Jul–Sep 1993
Tingri	28.63°N, 87.08°E	4300	2.4	279.4		Jul–Sep 1993
Xigazê	29.25°N, 88.88°E	3836	6.4	440.9		Jul–Sep 1993
Lhasa	29.7°N, 91.13°E	3649	7.8	435.0	–15.24	1986–1992
Damxung	30.48°N, 91.1°E	4200	1.6	468.8		Jun–Sep 1993
Nagqu	31.48°N, 92.07°E	4507	–1.4	429.2		Jun–Sep 1993
Amdo	32.35°N, 91.1°E	4800	–2.9	431.8		Jun–Sep 1993
Tanggula	33.03°N, 90.03°E	5070				Jul–Sep 1993
Tuotuohe	34.22°N, 92.43°E	4533	–4.2	276.0	–9.20	1991–1992
Wudaoliang	35.35°N, 91.1°E	4645	–5.5	272.8		Jun–Aug 1993
Xining	36.62°N, 101.77°E	2261	5.9	373.0	–6.83	1991–1992
Delingha	37.37°N, 97.37°E	2981	3.7	164.7	–7.19	1991–1992

where $\delta^{18}O_i$ and P_i are the monthly δ^{18}O and the corresponding precipitation amount, respectively.

3. Spatial and temporal variations of δ^{18}O in precipitation

3.1 *Spatial variation of* δ^{18}O

The spatial distribution of the mean annual δ^{18}O is obtained by calculating the weighted mean using equation (2) for stations covering more than one calendar year (see the sixth column in Table 1)

Usually, the initial condensate in evaporated vapor has a comparatively high stable isotopic ratio, and the stable isotopes in vapor and in precipitation are depleted gradually with the generation of condensation and rainfall because heavy isotopes are preferentially condensed. As a result, the δ^{18}O in precipitation will decrease gradually along the vapor transport path (Dansgaard, 1964; Zhang et al., 2003). However, the actual distribution of the δ^{18}O in precipitation is very different from the ideal one owing to the influence of different vapor sources. The magnitude of stable isotopic ratios in precipitation at Singapore located at the south end of the south path is thought to stand for the mean situation of the vapor evaporated from the West Pacific or the South China Sea, to a certain degree (Araguas, 1998), because of its position. The weighted mean δ^{18}O in precipitation at Singapore is –7.29‰ lower than that at Colombo and Bombay located at the south end of the plateau path, which shows low δ^{18}O values in the vapor evaporated from the West Pacific or the South China Sea and higher δ^{18}O values in the vapor from the Arabian Sea or the Bay of Bengal.

On the southern Tibetan Plateau, the ^{18}O in precipitation is greatly depleted and its ratios reach the lowest parts in the plateau path because of the very strong rainout of the stable isotopic compositions in vapor in the process of lifting over the southern slope of the Himalayas (Zhang et al., 2002).

Along the north path, the minimum δ^{18}O appears at Ürümqi, which is the furthest station from the oceans. However, the mean δ^{18}O increases, instead of decreasing, with increasing longitude, which shows that, on one hand, the vapor with relatively high δ^{18}O has replenished the precipitation of these stations, and on the other hand, the stable isotopic enrichment in falling raindrops owing to evaporation is strengthened in the dry central part of the north path. In the east part of the north path, the mean δ^{18}O tends to decrease, which shows that the evaporation enrichment in falling raindrops is weakened in the relatively humid atmosphere under the influence of monsoon vapor from the oceans.

3.2 *Seasonal variation of* δ^{18}O *in precipitation*

The δ^{18}O in precipitation also exhibits seasonal variation like temperature, precipitation, and other hydrometeorological factors. Figures 2a, 2b, and 2c give the time cross sections of the mean δ^{18}O in precipitation corresponding to the 12 stations along the south path, the 14 along the north path, and the 12 with more than one calendar year on record along the plateau path, respectively.

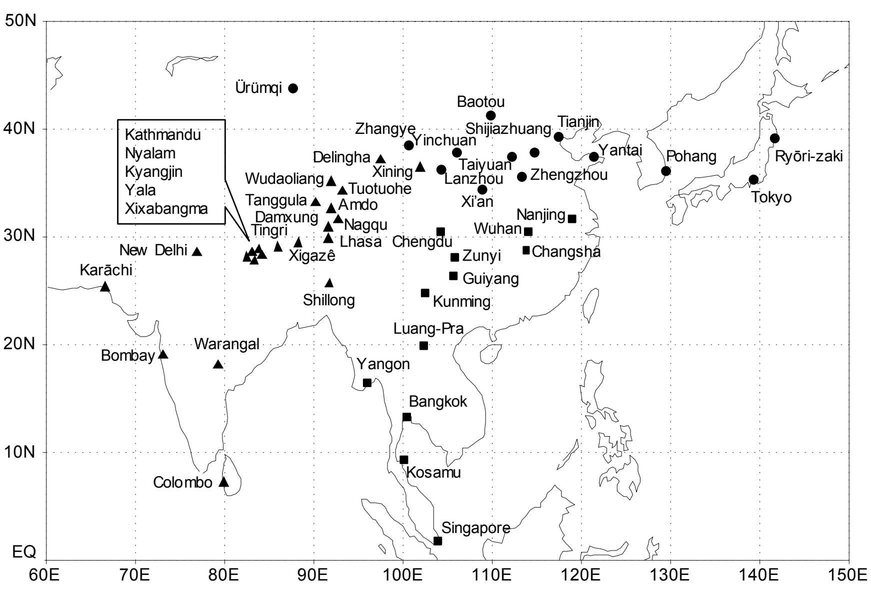

Fig. 1. Geographical distribution of the selected sampling stations.

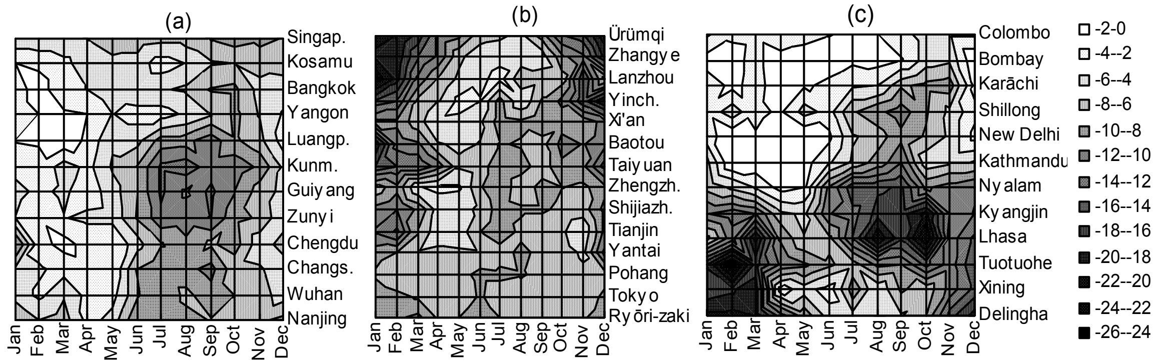

Fig. 2. Seasonal variations of the mean δ^{18}O in precipitation (a) along the south path, (b) along the north path, and (c) along the plateau path.

For the sampling stations along the south path, the maximum δ^{18}O in precipitation usually appears in spring before the rainy season, which is connected to the strong evaporation enrichment of raindrops in the falling process under the dry-hot synoptic condition in this season. The δ^{18}O in precipitation decreases continuously after the rainy season comes. It is noticeable that the minimum δ^{18}O does not appear in the month with the most heavy precipitation (figure omitted) but in the late rainy season. In the Tropics, the δ^{18}O in precipitation, with a small annual amplitude, shows a double humped pattern corresponding to the variations of the temperature and precipitation.

Along the north path, the high values of δ^{18}O usually appear in the warm season with the maximum in July or August, whereas the low values appear in the cold season with the minimum in December or January. The synchronism between the δ^{18}O in precipitation and temperature is very marked for the western stations with a great δ^{18}O annual amplitude. With increasing (eastern) longitude, the minimum δ^{18}O in precipitation ascends, but the maximum δ^{18}O descends, showing that the annual amplitude of the δ^{18}O in precipitation decreases gradually.

Along the plateau path, the seasonal variations of the δ^{18}O in precipitation have marked diversity for different stations. In the south of the Himalayas, the δ^{18}O in precipitation has a small annual amplitude with its maximum in the dry period, and it decreases after the rainy season starts and reaches the lowest point of the whole year in the fall; in the Himalayas and the southern Tibetan Plateau, two minimum δ^{18}O values appear in late winter to early spring and in fall, respectively. The generation of the former minimum is related to the intrusion of ocean vapor from low latitudes when the winter monsoon is weakened, but the later one is attributed to the very low δ^{18}O in vapor because of the possible minimum δ^{18}O values in ocean surface water in that season (Zhang et al., 2002). The maximum δ^{18}O in precipitation happens in April through June in conjunction with the dry and warm atmosphere (Zhang et al., 2002). On average, the δ^{18}O in precipitation basically remains at a low level during most of the year there compared with South Asia; from Tuotuohe to Delingha, the seasonal variation of the δ^{18}O in precipitation exhibits the continental pattern, namely keeping the synchronism of the seasonal variation of the δ^{18}O with that of temperature.

4. Relationship of the δ^{18}O in precipitation with temperature and precipitation

In this study, the period from May to October is prescribed as the warm half of the year (or the rainy season) and November to April as the cold half of the year (or the dry season).

4.1 *Correlations of the δ^{18}O with precipitation and temperature along the south path*

In order to compare the seasonal differences, the correlations of the δ^{18}O with precipitation and temperature under the annual, the warm half year (or the rainy season), and the cold half year (or the dry season) timescales are analyzed. The statistical results are listed in Table 2.

Under the annual scale, there is a marked amount effect for all sampling stations except Yangon and Changsha, showing that precipitation of these stations is influenced by the vapor from the oceans to different degrees (Dansgaard, 1964; Rozanski et al., 1997; Zhang et al., 2002). As for Yangon without the amount effect, its statistical confidence limit is poor because of the short sampling period (only 3 years) and the lack of precipitation during January through April in the dry season. In Changsha, although the seasonal variation of the δ^{18}O is similar to that of other Chinese stations along the south path, the precipitation is distributed mainly from January to early July (figure omitted). Such a distributional situation is linked to the frequent incursion of the southwest eddy. Although there is no amount effect under the annual scale, the negative temperature effect is notable at Changsha.

It can be seen that the lapse rates $d\delta^{18}O/dP$ at Kunming, Guiyang, Zunyi, and Chengdu located in southwest China are greater than that of other stations, which is attributed to the influence of altitude. If the same precipitation as that at low altitude is generated at high altitude, the wet air-parcel will undergo a greater temperature drop range because of the decrease of absolute humidity in the atmosphere. This means that the variation range in the $\delta^{18}O$ will also decrease, thus leading to a greater slope $d\delta^{18}O/dP$.

In the rainy season, all stations except Yangon, Luang-Pra, and Changsha have a marked amount effect.

In the dry season, the amount effect to a certain degree happens at the stations from Singapore to Zunyi. However, the temperature effect is notable at Changsha, Wuhan, and Nanjing, showing that the vapor sources and the influence factors in the rainy season are different from that in the dry season (Rozanski et al., 1997; Zhang et al., 2002).

Table 2. Correlations of the $\delta^{18}O$ in precipitation with precipitation and temperature along the south path.

Stations	Annual				Rainy season				Dry season			
	$\frac{d\delta^{18}O}{dP}$	r	$\frac{d\delta^{18}O}{dT}$	r	$\frac{d\delta^{18}O}{dP}$	r	$\frac{d\delta^{18}O}{dT}$	r	$\frac{d\delta^{18}O}{dP}$	r	$\frac{d\delta^{18}O}{dT}$	r
Singapore	–0.89	–0.39†	0.46	0.15	–1.32	–0.50†	1.25	0.35	–0.78	–0.35*	0.47	0.11
Kosamu	–1.05	–0.64†	0.21	0.10	–1.00	–0.42$^{\Delta}$	0.15	0.07	–1.18	–0.74†	1.94	0.40
Bangkok	–1.44	–0.64†	0.15	0.06	–1.07	–0.60†	1.09	0.48†	–2.29	–0.70†	–0.08	–0.02
Yangon	0.26	0.20	–0.08	–0.04	–0.22	–0.20	0.03	0.03	–1.46	–0.63	–1.82	–0.69
Luang-Pra	–1.29	–0.43*	–0.03	–0.02	–1.10	–0.36	0.96	0.49*	–3.12	–0.40	–0.11	–0.08
Kunming	–2.72	–0.58†	–0.45	–0.42†	–2.35	–0.54†	–0.71	–0.24$^{\Delta}$	–6.33	–0.65†	–0.7	–0.52†
Guiyang	–2.46	–0.48†	–0.28	–0.56†	–1.64	–0.32$^{\Delta}$	–0.71	–0.59†	–7.23	–0.71†	–0.44	–0.62†
Zunyi	–2.96	–0.57†	–0.34	–0.64†	–1.96	–0.39*	–0.64	–0.55†	–5.74	–0.62†	–0.48	–0.65†
Chengdu	–2.16	–0.55†	–0.23	–0.38†	–2.23	–0.65†	–0.28	–0.36†	5.53	0.14	–0.21	–0.14
Changsha	–0.11	–0.03	–0.17	–0.49†	0.29	0.07	–0.49	–0.61†	0.04	0.01	0.16	0.36*
Wuhan	–1.04	–0.36†	–0.07	–0.22	–1.31	–0.45*	–0.36	–0.49†	–1.15	–0.18	0.21	0.46*
Nanjing	–1.13	–0.45†	–0.07	–0.27*	–1.49	–0.57†	–0.32	–0.63†	–0.17	–0.04	0.33	0.61†

Note: 1. Symbols $\dagger, *$, and Δ represent where the correlation confidence limits exceed 0.01, 0.05, and 0.1, respectively.
2. Lapse rates $d\delta^{18}O/dP$ are all enlarged by 100 times (the same below).

Table 3. Correlations of the ^{18}O in precipitation with precipitation and temperature along the north path.

Stations	Annual				Warm half-year				Cold half-year			
	$\frac{d\delta^{18}O}{dT}$	r	$\frac{d\delta^{18}O}{dP}$	r	$\frac{d\delta^{18}O}{dT}$	r	$\frac{d\delta^{18}O}{dP}$	r	$\frac{d\delta^{18}O}{dT}$	r	$\frac{d\delta^{18}O}{dP}$	r
Ürümqi	0.41	0.87†	9.85	0.35†	0.39	0.67†	–2.00	–0.14	0.40	0.62†	10.83	0.37†
Zhangye	0.51	0.84†	19.73	0.48†	0.23	0.44†	0.25	0.02	0.56	0.66†	24.70	0.24
Lanzhou	0.45	0.70†	5.07	0.37*	0.21	0.32$^{\Delta}$	0.63	0.08	0.53	0.58*	19.57	0.33
Yinchuan	0.32	0.68†	9.17	0.46*	0.17	0.19	1.03	0.06	0.59	0.88†	41.59	0.85†
Xi'an	0.07	0.21$^{\Delta}$	–0.51	–0.07	0.01	0.01	–2.53	–0.33*	0.37	0.61*	7.89	0.46*
Baotou	0.20	0.55†	1.54	0.12	0.21	0.39*	–1.57	–0.17	0.37	0.50*	11.21	0.24
Taiyuan	0.10	0.18	–1.97	–0.32	–0.07	–0.10	–2.69	–0.53*	–	–	–	–
Zhengzhou	–0.01	–0.02	–1.83	–0.27*	–0.09	–0.12	–2.96	–0.43*	0.24	0.38$^{\Delta}$	1.24	0.09
Shijiazhuang	0.11	0.34†	–0.50	–0.12	–0.02	–0.04	–1.04	–0.35†	0.39	0.64†	–0.61	-0.02
Tianjin	0.10	0.33*	–0.56	–0.11	–0.15	–0.35$^{\Delta}$	–1.84	–0.59†	0.38	0.65†	6.52	0.29
Yantai	0.03	0.10	–0.84	–0.19	–0.14	–0.22	–1.17	–0.37*	0.28	0.40$^{\Delta}$	0.34	0.02
Pohang	–0.10	–0.36†	–0.84	–0.27†	–0.09	–0.14	–0.22	–0.09	–0.11	–0.22	–1.54	–0.33*
Tokyo	0.08	0.28†	–0.35	–0.13	0.06	0.11	–0.85	–0.37†	0.29	0.44†	–0.21	–0.05
Ryōri-zaki	0.11	0.38†	0.48	0.14	0.05	0.10	–0.51	–0.17	0.24	0.41†	0.51	0.10

Table 4. Correlations of the δ^{18}O in precipitation with precipitation and temperature along the plateau path.

	Annual				Rainy season				Dry season			
Stations	$\frac{d\delta^{18}O}{dP}$	r	$\frac{d\delta^{18}O}{dT}$	r	$\frac{d\delta^{18}O}{dP}$	r	$\frac{d\delta^{18}O}{dT}$	r	$\frac{d\delta^{18}O}{dP}$	r	$\frac{d\delta^{18}O}{dT}$	r
Colombo	–0.19	–0.18	1.43	$0.45^{\dagger}$	–0.23	$-0.35^{\triangle}$	0.13	0.04	–0.22	–0.18	2.50	0.37
Bombay	0.09	0.16	–0.05	–0.02	–0.06	–0.15	0.19	0.16	–0.80	–0.13	–0.37	–0.11
Karāchi	–0.91	–0.21	–0.17	$-0.27^{\triangle}$	–0.70	–0.21	–0.48	–0.20	–0.80	–0.07	–0.25	–0.22
Shillong	0.00	0.00	–0.47	-0.39^{*}	0.44	0.24	–2.52	-0.49^{*}	–0.19	–0.06	–0.28	–0.21
New Delhi	–1.83	$-0.47^{\dagger}$	–0.10	–0.14	–1.84	$-0.52^{\dagger}$	0.83	$0.38^{\dagger}$	–7.38	$-0.32^{\dagger}$	–0.23	-0.22^{*}
Kathmandu	–1.26	$-0.40^{\triangle}$	–0.55	$-0.58^{\dagger}$	–0.44	–0.14	–1.31	$-0.47^{\triangle}$	–10.72	$-0.70^{\triangle}$	–0.35	–0.38
Nyalam	–2.25	–0.17	–0.59	-0.52^{*}	–13.64	$-0.51^{\triangle}$	–2.48	$-0.89^{\dagger}$	0.32	0.06	–1.11	$-0.77^{\triangle}$
Kyangjin	–4.82	$-0.50^{\dagger}$	–0.57	-0.54^{*}	–6.13	$-0.65^{\dagger}$	–1.46	-0.73^{*}	–9.45	–0.45	0.99	0.54
Lhasa	–4.34	-0.33^{*}	–0.08	–0.05	–6.64	$-0.50^{\dagger}$	–0.23	–0.08	–27.59	–0.47	–1.93	$-0.91^{\triangle}$
Tuotuohe	15.72	0.64^{*}	0.66	$0.88^{\dagger}$	4.77	0.55	0.24	0.38	160.77	0.49	0.65	0.72
Xining	3.14	0.22	0.56	0.72^{*}	–6.12	$-0.81^{\triangle}$	–0.22	–0.23	77.01	0.73	0.57	0.57
Delingha	26.72	$0.80^{\dagger}$	0.63	$0.93^{\dagger}$	5.24	0.54	0.45	0.63	135.05	$0.80^{\triangle}$	0.69	0.73

4.2 *Correlations of the* δ^{18}O *with precipitation and temperature along the north path*

Table 3 gives the correlations of the δ^{18}O in precipitation with temperature and precipitation under the annual, the warm half of the year, and the cold half of the year timescales along the north path.

Under the annual scale, all stations except Taiyuan, Zhengzhou, and Yantai have a marked temperature effect, showing that the variations of the δ^{18}O in precipitation are generally influenced by the continental vapor (Dansgaard, 1964; Rozanski et al., 1997; Zhang et al., 2002; Tian et al., 1997). It can be seen that the $d\delta^{18}O/dT$ decreases with increasing longitude, showing the high sensitivity of the δ^{18}O on temperature descent.

In the warm half of the year, Ürümqi, Zhangye, Lanzhou, Baotou, and Tianjin stations characterize the marked continental climate and thus exhibit the temperature effect to a certain degree. However, the δ^{18}O at other stations is either unrelated to temperature or shows a marked amount effect.

In the cold half of the year, influenced by the wide-range continental cold-high pressure system, all stations except Pohang have a marked temperature effect with the $d\delta^{18}O/dT$ in West China greater than that in East China, and that in East China is greater than that in Japan.

4.3 *Correlations of the* δ^{18}O *with precipitation and temperature along the plateau path*

The record series of the 14 sampling stations listed in Table 4 all exceed one calendar year.

Under the annual timescale, and in South Asia, there is a notable amount effect at New Delhi and Kathmandu, and a negative temperature effect to a certain degree at Karāchi and Shillong. Influenced by the ITCZ, precipitation, temperature, and δ^{18}O in precipitation demonstrate the double humped pattern at Colombo, with the corresponding relations: in the dry season high temperature-high δ^{18}O and in the rainy season low temperature-low δ^{18}O. Therefore, a special temperature effect happens in the Tropics. At Kyangjin and Lhasa with the amount effect, the lapse rate $d\delta^{18}O/dP$ is great. However, there is a notable temperature effect at Tuotuohe, Xining, and Delingha on the northern Tibetan Plateau.

In the rainy season, in addition to the marked amount effect or the negative temperature effect at the stations from Shillong to Lhasa, the marked amount effect appears also at Xining, showing that its vapor sources for precipitation are probably from the oceans.

In the dry season, the amount effect appears only at New Delhi and Kathmandu. Three stations in the middle and northern Tibetan Plateau all exhibit the temperature effect to a certain degree.

5. Variations of δ^{18}O in precipitation along vapor transport paths

5.1 *Variations of* δ^{18}O *along the south path*

The south path that converges the vapor from the Arabian Sea, the Bay of Bengal, the South China Sea, and the West Pacific, is an important vapor passage for influencing the synoptic, climatic, and hydrological situations of Southwest China, East China, and even all of East Asia.

Figure 3 shows the variations of the mean δ^{18}O in precipitation, mean temperature, and mean precipitation in different seasons along the south path. In low latitudes, with the small seasonal difference of temperature and precipitation between two seasons, small seasonal differences of mean δ^{18}O appear at Singapore

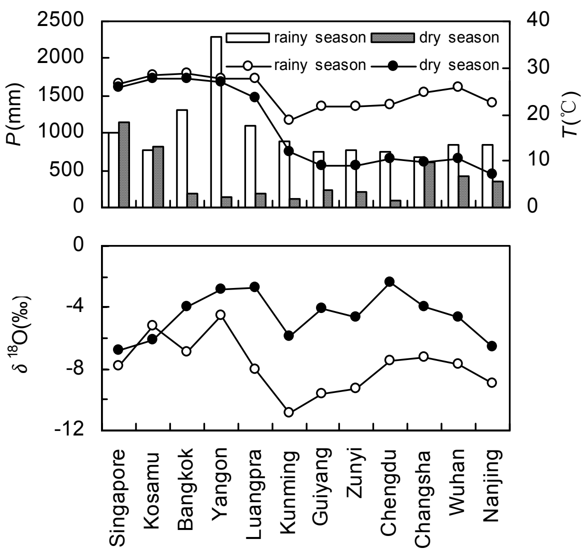

Fig. 3. Variations of the mean δ^{18}O, mean temperature and mean precipitation in different seasons along the south path.

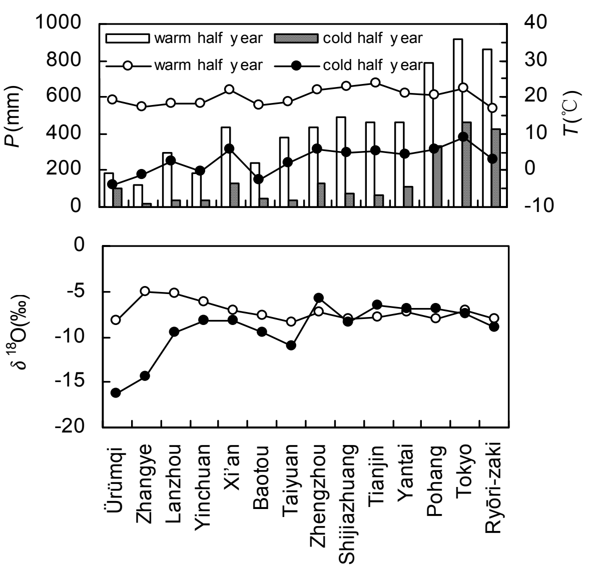

Fig. 4. Same as Fig. 3 but along the north path.

and Kosamu. After entering the Yunan-Guizhou Plateau, mean temperature decreases. It can be seen that the seasonal differences of mean temperature in middle latitudes is greater than in low latitudes. From Bangkok and towards the north, the seasonal differences of the δ^{18}O in precipitation increase markedly with the increase of the seasonal differences of mean precipitation, displaying as in the rainy season heavy precipitation-low mean δ^{18}O and in the dry season light precipitation-high mean δ^{18}O.

According to the Rayleigh fractionation model (Dansgaard, 1964; Jouzel et al., 1997), the ^{18}O in precipitation will be depleted gradually along the vapor transport path for the precipitation from identical vapor sources. However, whether in the rainy season or in the dry season, the δ^{18}O in precipitation does not show a monotone descent in Fig. 3, but a wavy variation, illustrating the different vapor sources. For example, the mean δ^{18}O at Kosamu, Bangkok, and Yangon in the rainy and dry seasons, and at Luang-Pra in the dry season, are all greater than that at Singapore. It can be inferred that the vapor for precipitation more likely originates from the Arabian Sea and the Bay of Bengal with high δ^{18}O values. That the minimum δ^{18}O appears at Kunming is related to its high altitude because a lot of vapor with relatively high δ^{18}O has been condensed into rainfall before arriving Kunming.

5.2 *Variations of* δ^{18}O *along the north path*

The weather and climate along the north path are influenced by westerly fluctuation on the whole. Figure 4 illustrates the latitudinal variations of the δ^{18}O in precipitation, mean temperature, and mean precipitation in the warm and cold halves of the year for the selected stations. Regardless of being in the warm half of the year or the cold half of the year, the increasing trend of precipitation and the seasonal differences of precipitation and temperature are all notable from the west to the east along the north path. It can be seen that the seasonal differences of δ^{18}O for the stations in the west of Zhengzhou are all greater than those in the east of Zhengzhou along the north path, displaying as in the warm half of the year high temperature-high mean δ^{18}O and in the cold half of the year low temperature-low mean δ^{18}O. However, the seasonal differences of the mean δ^{18}O in precipitation decrease with longitude and become minute from Zhengzhou towards the east.

As shown in Fig. 4, during the cold half of the year, the mean δ^{18}O in precipitation reaches its minimum at Ürümqi with the lowest temperature due to the wide, cold high pressure system over Mongolia, then increases gradually with increasing longitude and remains at roughly the same level from Zhengzhou towards the east. During the warm half of the year, precipitation increases markedly, influenced by the summer monsoon over East Asia, and thus the δ^{18}O in precipitation decreases due to the amount effect. Because the force of the summer monsoon is feeble in Xinjiang and Gansu under extreme continental climate, the ^{18}O values in precipitation are lower in the east than in the west. Therefore, two types of δ^{18}O variations along the north path are formed: in the west, whether in the warm half of the year or in the cold half of the year,

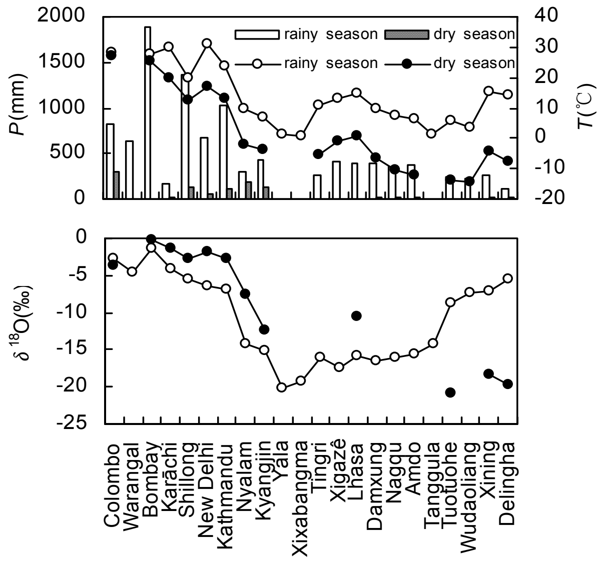

Fig. 5. Same as Fig. 3 but along the plateau path.

temperature is the major factor controlling the δ^{18}O in precipitation and the δ^{18}O in precipitation fluctuates with temperature; in the east, influenced by the monsoon over East Asia, the δ^{18}O in precipitation shows a marked temperature effect during the cold half of the year, but a marked amount effect during the warm half of the year.

5.3 *Variations of* δ^{18}O *along the plateau path*

The sampling stations along the plateau path stretch over South Asia and the Tibetan Plateau, two famous monsoon regions. Figure 5 gives the variations of the mean δ^{18}O in precipitation, mean temperature, and mean precipitation in the rainy season and in the dry season along the path. Because the sampling periods of most stations along the Qinghai-Xizang line are during the prevailing summer monsoon, June through September is regarded as the rainy season and December through March as the dry season in Fig. 5.

Whether in the rainy season or in the dry season, the δ^{18}O in precipitation, with small seasonal differences, shows relatively high values for 4 stations including Karāchi in the southernmost end of the path, because the vapor originates possibly from the Arabian Sea. With increasing latitude, the δ^{18}O decreases gradually, and a sharp depletion of the ^{18}O in precipitation takes place as the altitude rises above 3800 m a.s.l. (above sea level) due to the very strong rainout of the stable isotopic compositions in vapor in the process of lifting on the southern slope of the Himalayas, especially at Yala and Xixabangma where the mean δ^{18}O in the rainy season is −20.29‰ and −19.34‰, respectively, reaching the minimum in the cross section. With the decrease of altitude in the southern plateau, the δ^{18}O increases slightly, and is kept at a low level, but increases persistently with increasing latitude from the Tanggula Mountains to the northern Tibetan Plateau, and reaches another maximum at Delingha, at the north end of the plateau path. The situation in the dry season is different: the mean δ^{18}O values in precipitation basically decrease from South Asia over the Himalayas to the northern Tibetan Plateau, which is related to the climate being controlled by the cold, high pressure and the fact that the vapor source is unitary in the dry season.

It can be seen from Fig. 5 that the mean δ^{18}O in the rainy season is smaller than in the dry season for most stations in the south of the Himalayas, with great seasonal differences of precipitation and influence by the monsoon over South Asia. According to the analysis of sampling data limited in number, the mean δ^{18}O in the rainy season is smaller than in the dry season from Nyalam to the Tanggula Mountains, which is similar to the results under the monsoon systems over South Asia and East Asia. On the other hand, in the north of Tanggula Mountains, the mean δ^{18}O in the rainy season is greater than in the dry season, showing that the influence of the monsoon on the Tibetan Plateau precipitation is feeble here. Such a result is similar to that in the stations with extreme continental climate in the north path.

A survey has found that many water bodies have heavy stable isotopic compositions due to strong evaporation on the Tibetan Plateau where the evaporation amount is greater than the supply amount. According to the measurements, the mean δ^{18}O in water of Qinghai Lake is about +1.97‰ and that in Haiyanwan is about +2.89‰ (Zhang, 1994). Consequently, the replenishment of the vapor with heavy δ^{18}O would increase the δ^{18}O in precipitation. It can be concluded that the vapor evaporated from water bodies on the Tibetan Plateau is the basic reason that the δ^{18}O in precipitation increases with increasing latitude in the rainy season in the regions.

To sum up, the δ^{18}O in precipitation exhibits a marked temperature effect and its magnitude indicates the degree of temperature under continental climate. In the monsoon regions, the different variations characterize the δ^{18}O in precipitation in different seasons: under the summer monsoon, the magnitude of the δ^{18}O in precipitation, with a marked amount effect, indicates the intensity of the precipitation amount or the strength of the monsoon; under the winter monsoon, the magnitude of the δ^{18}O in precipitation, with a marked temperature effect, indicates the degree of the temperature.

6. Conclusions

Along the south path, the maximum δ^{18}O in precipitation usually appears in spring before the rainy season and the minimum in the late rainy season. The seasonal differences of mean δ^{18}O in precipitation are small in the stations located in low latitudes, but increase markedly from Bangkok towards the north. The δ^{18}O values in precipitation fluctuate on the whole. The appearance of the minimum δ^{18}O at Kunming is related to its high altitude. In the rainy season, all stations except Yangon, Luang-Pra, and Changsha exhibit a marked amount effect. In the dry season, the amount effect to a certain degree happens at the stations from Singapore to Zunyi. However, the temperature effect is notable at Changsha, Wuhan, and Nanjing.

Along the north path, the high δ^{18}O values usually appear in the warm season, whereas the low values usually appear in the cold season. The seasonal differences of the mean δ^{18}O for the stations in the west of Zhengzhou are all greater than in the east of Zhengzhou. In the warm half of the year, except Ürümqi, Zhangye, Lanzhou, Baotou, and Tianjin that exhibit the temperature effect to a certain degree, the δ^{18}O at other stations is either unrelated to temperature or shows as a marked amount effect. Markedly influenced by the summer monsoon over East Asia, the δ^{18}O values in precipitation are lower in the east than in the west. During the cold half of the year, all stations except Pohang experience a marked temperature effect. The mean δ^{18}O in precipitation reaches its minimum at Ürümqi with the lowest temperature due to the wide, cold high pressure over Mongolia, then increases gradually with longitude, and remains at roughly the same level from Zhengzhou towards the east.

Along the plateau path, the seasonal variations of the δ^{18}O in precipitation show marked diversity for different stations. In the south of the Himalayas, the maximum δ^{18}O in precipitation appears in the dry period and the minimum appears in fall. In the Himalayas and the southern Tibetan Plateau, the δ^{18}O in precipitation basically remains at a low level during most of the year. From Tuotuohe to Delingha, the seasonal variation of the δ^{18}O in precipitation remains synchronized with the seasonal variation of temperature. On average, the mean δ^{18}O values in precipitation are correspondingly high in the southern parts of the Indian subcontinent, then gradually decrease with latitude. A sharp depletion of the stable isotopic compositions in precipitation takes place due to the very strong rainout of the stable isotopic compositions in vapor in the process of lifting over the southern slope of the Himalayas. The δ^{18}O in precipitation is kept at a low level from Nyalam to the Tanggula Mountains during the rainy season, but persistently increases with increasing latitude from the Tanggula Mountains to the northern Tibetan Plateau because of the replenishment of vapor with relatively heavy stable isotopic compositions originating from water bodies on the inner plateau. The situation during the dry season is different: the mean δ^{18}O values in precipitation basically decrease along the path from south to north.

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