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Stable Isotopic Variations in Precipitation in Southwest China

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ABSTRACT

This study analyzes the relationships of stable isotopes in precipitation with temperature, air pressure and humidity at different altitudes, and the potential influencing mechanisms of control factors on the stable isotopes in precipitation in Southwest China. There appear marked negative correlations of the δ^{18}O in precipitation with precipitation amount, vapor pressure and atmospheric precipitable water (PW) at the Mengzi, Simao and Tengchong stations on the synoptic timescale; the marked negative correlations between the δ^{18}O in precipitation and the diurnal mean temperature at 400 hPa, 500 hPa, 700 hPa and 850 hPa are different from the temperature effect in middle-high-latitude inland areas. In addition, the notable positive correlation between the δ^{18}O in precipitation and the dew-point deficit ΔT_{d} at different altitudes is found at the three stations. Precipitation is not the only factor generating an amount effect. Probably, the amount effect is related to the variations of atmospheric circulation and vapor origins. On the annual timescale, the annual precipitation amount weighted-mean δ^{18}O displays negative correlations not only with annual precipitation but also with annual mean temperature at 500 hPa. It can be deduced that, in the years with an abnormally strong summer monsoon, more warm and wet air from low-latitude oceans is transported northward along the vapor channel located in Southwest China and generates abnormally strong rainfall on the way. Meanwhile, the abnormally strong condensation process will release more condensed latent heat in the atmosphere, and this will lead to a rise of atmospheric temperature during rainfall but a decline of δ^{18}O in the precipitation. On the other hand, in the years with an abnormally weak summer monsoon, the precipitation and the atmospheric temperature during rainfalls decrease abnormally but the δ^{18}O in precipitation increases.

Key words: Southwest China, stable isotopes, precipitation, temperature, humidity, correlation

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1. Introduction

In the past few decades, stable isotopes of oxygen (^{18}O, as $H_2^{18}O$) and hydrogen (D, as $HD^{16}O$) in water bodies have seen increasing application in the studies of cloud physics, climatology, hydrology and palaeoclimatology (Dansgaard, 1964; Jouzel et al., 1997). Although found in very small concentrations in natural water, the stable isotopes respond to environmental variation very sensitively. Precipitation is an important link in the water cycle. The abundance of the stable isotopes in precipitation is closely associated with the weather process of generating condensation, initial conditions at the vapor origin and large-scale circulation patterns (Jouzel et al., 1997; Araguás-Aroguás et al., 1998).

The global survey for the stable isotopes in precipitation shows that there appears a marked positive correlation between stable isotopic ratio in precipitation and temperature in middle-high-latitude inland areas, which is called the temperature effect (Dansgaard, 1964). The temperature effect arises from the fact that the temperature during phase change controls the fractionation rate of stable isotopes in the atmosphere and in the precipitation (Dansgaard, 1964; Rozanski et al., 1992, 1997; Zhang et al., 2003). Using the temperature effect, the stable isotopic compositions in different sediments can be quantitatively

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recovered as a temperature proxy in middle-high latitudes (Yao et al., 1996; Yao, 1999; Thompson et al., 2000).

In monsoon regions, especially in Asian monsoon areas, the variation of precipitation shows a distinct isotope signature: the summer rains are usually more depleted in stable isotopes than the winter rains in the same area, regardless of the generally higher summer temperature. The function of the surface air temperature on the overall stable isotopic fractionation during precipitation is overshadowed by the effects controlled by precipitation amount. Dansgaard (1964) identified this phenomenon as the amount effect, which is the marked negative correlation between precipitation amount and stable isotopic ratio in precipitation. Several investigators have attempted to explain the observed isotopic characteristics in monsoon regions. Yapp (1982) conducted a simulation in order to interpret the generation of the amount effect. He thought that, in precipitation resulting from convective storm cells, the deeper the convection is, the more completely the vapor is "wrung out" of the air, the heavier the precipitation, and the closer to that of the vapor is the δ^{18}O in the precipitation (i.e., depleted in ^{18}O); conversely, the shallower the convection is, the less the vapor condenses, and thus the higher the δ^{18}O in precipitation is. Posmentier et al. (2004) used a simple model to show that a key climate variable that may be responsible for the observed isotopic patterns in the eastern Asia monsoon areas is the seasonal differences in atmospheric stability in a zone affected by Walker circulation in the zone of prevailing westerlies. Hoffmann and Heimann (1997) successfully simulated the winter-enriched and summer-depleted isotopic seasonality in Hong Kong using a GCM. However, since GCMs are very complex, the physical mechanisms responsible for the winter-enriched and summer-depleted isotopic seasonality were not isolated by the authors. Araguás-Araguás et al. (1998) thought that the generation of the amount effect can be qualitatively explained as the combined effect of the extent of the rainout process of deep convective clouds (being the major contributions to rainfall in this region) high precipitation intensity, recirculation of moisture, and isotope exchange processes beneath the cloud base.

Southwest China is located in a typical monsoon region. The origins of atmospheric moisture forming precipitation in the region are very complex, as well as other factors that have effects on precipitation. Vapor from the South China Sea, the Bay of Bengal, the Arabian Sea and air flows across the equator converge over Southwest China, being further transported into the middle-lower reaches of the Yangtze River and other East Asian regions. Therefore, the monsoon precipitation over the passage of water vapor transport is closely associated with the water vapor over Southwest China (Duan et al., 2000). The characteristics of water vapor along the channel are thus important for a better understanding of the physical underlying reasons for the anomalous droughts and floods in the Yangtze River Valley and East Asia, where precipitation amounts are closely related to the Asian monsoon, the subtropical-high system, the ocean-continent thermodynamic forcing and the dynamic effects of synoptic systems in mid latitudes (Xu et al., 2002). Nevertheless, the water vapor channel over Southwest China is deeply involved as it is the region where the monsoon systems over East Asia, South Asia and the Tibetan Plateau interact with each other.

The focus of the study, by a sampling survey conducted in Yunnan Province (the shading in Fig. 1) in Southwest China, and by the quantitative and qualitative analyses of the stable isotopic variations in precipitation on different timescales and the relationships with meteorological factors in the atmosphere, is on stable isotopic features in monsoon rains on synoptic, seasonal and annual timescales, and on the possible mechanisms of influence of different vapor origins and atmospheric circulations on stable isotopes in precipitation in the region.

2. Sampling stations and data

The precipitation sampling of stable isotopes and the regular weather observations of surface air temperature, precipitation amount, humidity and air pressure at 0200, 0800, 1400 and 2000 LST as well as radio-sounding observations of temperature, geopotential height and dew-point deficit, $\Delta T_{\rm d} (= T - T_{\rm d}$, where T is air temperature and $T_{\rm d}$ is dew-point temperature) at 400 hPa, 500 hPa, 700 hPa and 850 hPa at 0700 and 1900 LST were conducted daily at three national primary weather stations, Mengzi (23.23°N, 103.23°E, 1301.7 m MSL), Simao (22.40°N, 101.24°E, 1302.9 m MSL) and Tengchong (25.01°N, 98.30°E, 1648.7 m MSL), from February 2003 to February 2004.

During nearly one year of sampling, 161, 141 and 198 precipitation samples were obtained at Mengzi, Simao and Tengchong, respectively.

Kunming (25.01°N, 102.41°E, 1892 m MSL) is one of the Chinese sampling stations associated with the global survey network established by the International Atomic Energy Agency (IAEA) in co-operation with the World Meteorological Organization (WMO). There have been 13-year stable isotopic survey records from 1986 to 2001 (1993 to 1995 are missing) at Kunming. The monthly stable isotopic ratios in precipitation and relational weather data at the surface and at different standard isobaric surfaces are available from

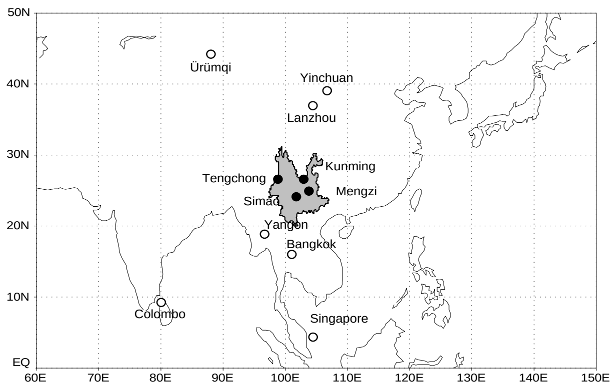

Fig. 1. Map showing the distribution of sampling stations (solid dots) in Yunnan (shaded), Southwest China.

IAEA/WMO (2001) and from the Chinese National Meteorological Center.

In addition, three GPS (Global Positioning System) vapor receivers (CRS1000, Switzerland) were installed respectively at the Mengzi, Simao and Tengchong stations in order to reveal the relationship between stable isotopes in precipitation and atmospheric humidity. The water vapor in the atmosphere can be retrieved, once an hour, by the GPS receivers. Compared with radiosondes, the retrieved water vapor has a high precision of 1–2 g cm^{-2} (Bevis et al., 1992). The vapor monitoring was conducted from February to November, September, and November 2003, respectively, at Mengzi, Simao, and Tengchong, ending at these times due to failures of the three instruments.

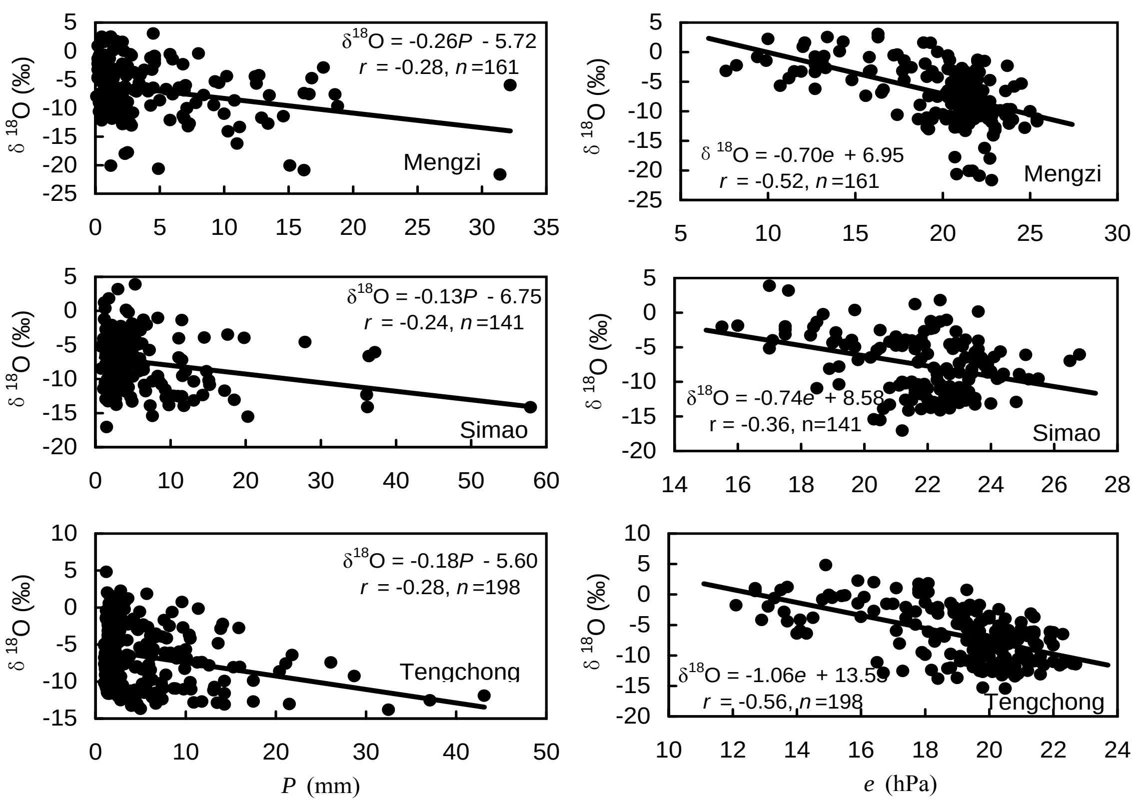

Fig. 2. Scatter plots of δ^{18}O in precipitation versus precipitation P and vapor pressure e, during the sampling from February 2003 to February 2004 at Mengzi, Simao and Tengchong on the synoptic timescale. (The linear regression equations with correlation coefficient r and sample size n are shown in each panel).

All precipitation samples collected in the rain gauges were sealed in plastic bottles and kept in a freezing tank, and were finally measured for their stable oxygen isotopic ratios by the Delta-Plus mass spectrometer (made in Germany), at the Key Laboratory of Ice Core and Cold Regions Environments, Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences. The measured ratio of the stable oxygen isotopes in the samples, ^{18}O/^{16}O, is expressed in parts per thousand of their deviation relative to Standard Mean Ocean Water (SMOW). The quantity δ^{18}O is defined by the following equation,

$$\delta^{18}O = \left[\frac{R_{\text{sample}}}{R_{\text{SMOW}}} - 1\right] \times 1000 \,, \tag{1}$$

where R_{sample} and R_{SMOW} stand for the isotopic ratio R in the water sample and in the Standard Mean

Ocean Water, respectively. The measurement precision is ±0.1‰.

The precipitation amount weighted-mean δ^{18}O in precipitation, $\overline{\delta^{18}\mathrm{O}}$, is defined as

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

where, $\delta^{18}\mathrm{O}_i$ and P_i are the stable oxygen isotopic ratio and the corresponding precipitation amount.

3. Variations of the δ^{18}O in precipitation

3.1 *Variations of the δ^{18}O in precipitation on the synoptic timescale*

3.1.1 *Correlations of the δ^{18}O in precipitation with precipitation and vapor pressure at the surface*

Figure 2 shows that there are marked negative correlations of the δ^{18}O in precipitation with precipitation amount and vapor pressure at Mengzi, Simao and Tengchong. All correlations are significant at the 0.01 confidence level, showing the marked amount effect and humidity effect. It can be found, at the three stations, that the negative correlation between the δ^{18}O in precipitation and vapor pressure is more notable, demonstrating that the atmospheric humidity has an important impact on the evaporation enrichment function of stable isotopes in precipitation. However, the panels in Fig. 2 show all the weak dependence of the stable isotopic ratio on precipitation and vapor pressure. The possible causation is thought to be the stochastic influence of various controlled factors, e.g. condensation altitudes, wind speed, atmospheric stability and humidity, and vapor sources, etc. On the synoptic timescale, these factors greatly influence the fractionation of stable isotopes in the vapor, and lead to large amplitudes in the stable isotopic variation in precipitation.

3.1.2 *Correlation between the δ^{18}O in precipitation and the precipitable water*

Precipitable water in the atmosphere, PW, is defined as the height of the liquid-water column transformed by all vapor in a vertical air column. The value of PW stands for the aggregation of vapor in the atmosphere and reflects the humidity condition of the whole atmosphere. The temporal variations, at one hour intervals, of PW deduced by signals from GPS vapor receivers at three stations are displayed in Fig. 3 (solid lines). Some data gaps are due to failures of the three GPS vapor receivers.

From Fig. 3, the seasonal variation of the δ^{18}O in precipitation shows a reversed pattern to that of the PW. In the dry season, from February to the middle of May, the PW in the atmosphere is relatively low, but the δ^{18}O in precipitation is relatively high. With the onset of the summer monsoon in the middle of May, the PW increases rapidly and remains at a relatively high level; contrarily, the δ^{18}O in precipitation decreases markedly and remains at a relatively low level. After late September, with the southward expansion of the Mongolia High, the PW values at the three stations decrease successively and quickly, but the δ^{18}O values in precipitation increase. The linear regression equations of the δ^{18}O (units: ‰) in precipitation against the PW (units: mm)in the atmosphere are given below,

$$\delta^{18}\mathrm{O} = \begin{cases} -0.36\mathrm{PW} + 6.40 \\ \quad (\text{at Mengzi}, r = -0.58, n = 148) \\ -0.44\mathrm{PW} + 11.34 \\ \quad (\text{at Simao}, r = -0.50, n = 76) \\ -0.33\mathrm{PW} + 5.46 \\ \quad (\text{at Tengchong}, r = -0.55, n = 149) \end{cases} \tag{3}$$

in which r is the correlation coefficient, and n is the sample size. Compared with the amount effect, the atmospheric humidity condition, i.e. the PW, has a more marked correlation with the stable isotopes in precipitation.

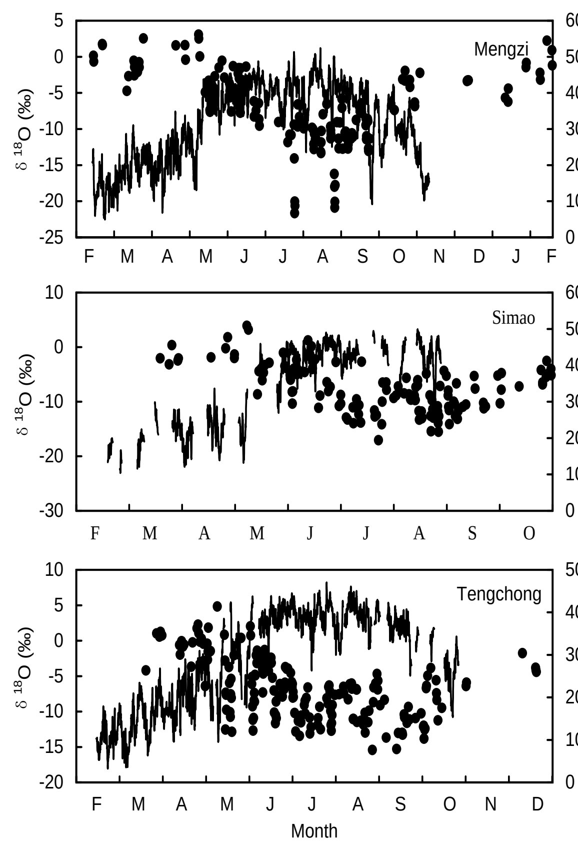

Fig. 3. Temporal variations of the δ^{18}O in precipitation (dots) and the atmospheric precipitable water (PW, sold lines, retrieved by GPS vapor receivers) during the sampling from February 2003 to February 2004 at Mengzi, Simao and Tengchong.

Although there is a good relationship between the δ^{18}O in precipitation and the PW in the atmosphere, such a result cannot justify a claim that the PW is the main and direct control factor impacting the δ^{18}O in precipitation. However, in view of the consistency of timing in the seasonal variation of the δ^{18}O and the PW, it can be deduced that both the δ^{18}O in precipitation and the PW in the atmosphere change against the same atmospheric circulation background and are governed by some similar geographical and meteorological factors. In the dry season, controlled by dry air mass, accompanied with less vapor transport and less precipitation, and because of the stronger evaporation enrichment in the raindrops below cloud, the stable isotopic compositions in precipitation are enriched; in the rainy season, influenced by warm and wet air currents, with more vapor transport and heavier precipitation, and because of the stronger rainout of stable isotopes in vapor on the way as well as weaker evaporation enrichment in the raindrops below cloud, the stable isotopic compositions in precipitation are depleted.

3.1.3 *Correlations of the* δ^{18}O *in precipitation with temperature, air pressure and humidity at different altitudes*

Precipitation and the precipitation process are controlled by factors such as the circulation pattern and vapor conditions. Accordingly, the magnitude of the δ^{18}O in precipitation should be influenced by the meteorological factors at different altitudes. Table 1 gives the correlations of the mean δ^{18}O in precipitation versus diurnal mean temperature T, diurnal mean geopotential height H and diurnal mean dew-point deficit $\Delta T_{\rm d}$ at 400 hPa, 500 hPa, 700 hPa and 850 hPa. The δ^{18}O values for the calculation in Table 1 have been converted into the diurnal amount weighted-means by Eq. (2). These data in the table stand for the variations of the correlations between the δ^{18}O in precipitation and the control parameters in the sampling period. The 850-hPa isobaric surface is close to the local surface at the Mengzi and Simao stations because of their high altitude, but it is under the surface altitude at the Tengchong station.

Usually, the relationship between stable isotopes and temperature is of most concern. However, in Table 1, the stable isotopes in precipitation demonstrate the marked negative temperature effect at four altitudes, especially at upper levels (approximate condensation levels). Such a result is very different from the situation in middle-high-latitude inland regions where the temperature effect is dominant (Dansgaard, 1964; Araguás-Araguás et al., 1998; Posmentier et al., 2004). In fact, this kind of negative correlation between stable isotopes in precipitation and temperature is unrelated to the fractionation mechanism of stable isotopes, and the temperature here does not represent condensation temperature.

Among all correlations between the δ^{18}O in precipitation and the geopotential heights H at different isobaric surfaces, the negative ones at 400 hPa and at 500 hPa are most notable. It is known that, according to the evolvement of weather systems, the lower the H at upper levels, the stronger the vertical convection in the atmosphere, the heavier the rainfall, and thus the lower the δ^{18}O in precipitation, owing to the amount effect; on the other hand, the higher the H at upper levels, the weaker the vertical convection in the atmosphere, the less the rainfall, and thus the higher the δ^{18}O in precipitation. Below 700 hPa, the marked negative correlation between the δ^{18}O and H disappears at all three stations. The above result illuminates that circulation patterns at upper levels have a marked impact on stable isotopic variations in precipitation.

In Table 1, the dew-point deficits $\Delta T_{\rm d}$ at different isobaric surfaces have all marked positive correlations with the δ^{18}O in precipitation. It is known that the $\Delta T_{\rm d}$ in the atmosphere reflects the atmospheric humidity conditions. Corresponding to a greater $\Delta T_{\rm d}$, the atmosphere is drier, and thus the stable isotopic compositions in precipitation are more enriched because of the stronger evaporation enrichment function of heavy isotopes in a falling raindrop; contrarily, corresponding to smaller $\Delta T_{\rm d}$, the stable isotopic compositions in precipitation maintain lower levels as they

Table 1. Correlations of the δ^{18}O in precipitation versus temperature T, geopotential height H and dew-point deficit $\Delta T_{\rm d}$ at different isobaric surfaces (the values in parentheses are correlation coefficients).

	400 hPa			500 hPa			700 hPa			850 hPa		
Station	$d\delta/dT$	$d\delta/dH$	$d\delta/d\Delta T_{\rm d}$	$d\delta/dT$	$d\delta/dH$	$d\delta/d\Delta T_{\rm d}$	$d\delta/dT$	$d\delta/dH$	$d\delta/d\Delta T_{\rm d}$	$d\delta/dT$	$d\delta/dH$	$d\delta/d\Delta T_{\rm d}$
Mengzi	−0.73	−0.05	0.52	−0.79	−0.06	0.46	−1.28	0.03	1.19	−0.41	0.04	1.09
n=89	(−0.55)	(−0.54)	(0.49)	(−0.56)	(−0.46)	(0.46)	(−0.43)	(0.16)	(0.40)	(−0.32)	(0.33)	(0.39)
Simao	−1.17	−0.07	0.68	−1.37	−0.06	0.69	−1.45	0.04	2.08	−0.71	0.05	1.62
n=90	(−0.58)	(−0.51)	(0.60)	(−0.58)	(−0.33)	(0.57)	(−0.45)	(0.21)	(0.53)	(−0.20)	(0.26)	(0.50)
Tengchong	−0.72	−0.06	0.46	−0.84	−0.08	0.65	−1.12	−0.03	1.34			
n=121	(−0.52)	(−0.63)	(0.39)	(−0.52)	(−0.60)	(0.44)	(−0.53)	(−0.16)	(0.45)			

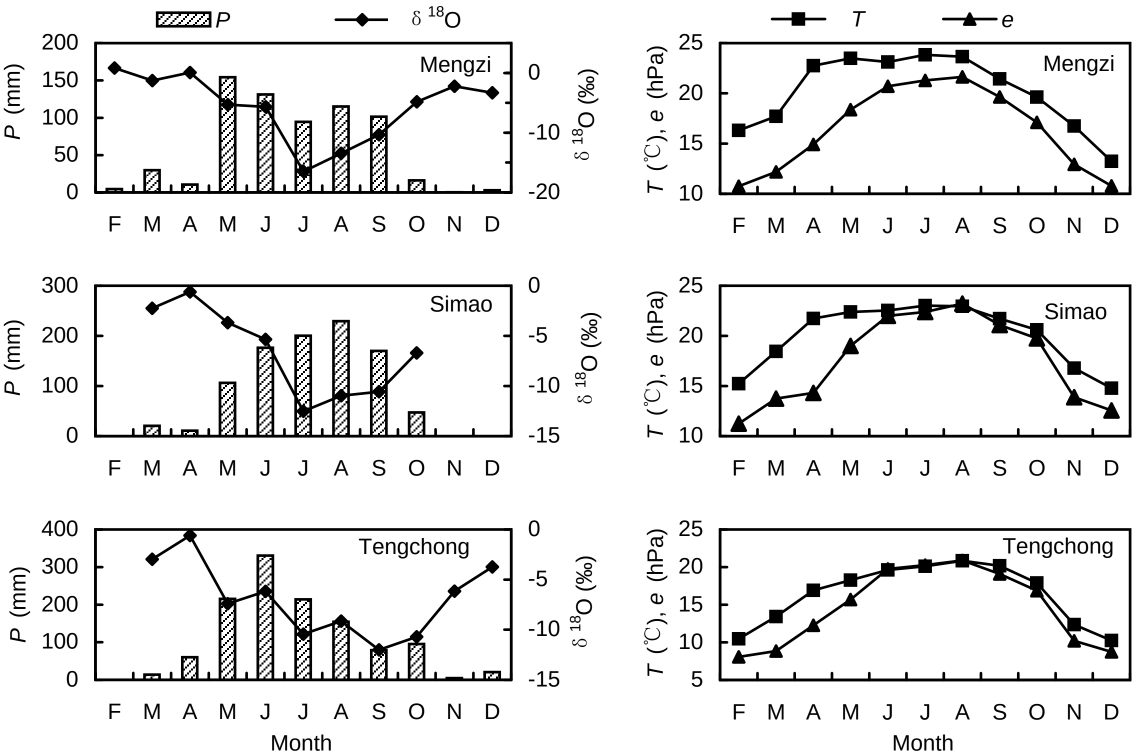

Fig. 4. Seasonal variations of the monthly amount weighted-mean δ^{18}O [based on Eq. (2)], monthly precipitation, monthly mean temperature and monthly mean vapor pressure during the sampling from February 2003 to February 2004 at Mengzi, Simao and Tengchong.

condense because of the weaker evaporation enrichment function in a falling raindrop. Consequently, when the summer monsoon prevails, a warm and wet marine air mass with great unstable energy easily forms a strong convection system, and consequently the δ^{18}O in precipitation decreases; when the winter monsoon prevails, the cold and dry continental air mass with a stable stratification forms very little precipitation, so, the δ^{18}O in precipitation increases. This is why the δ^{18}O in precipitation shows the marked negative correlation with temperature and the marked positive correlation with $\Delta T_{\rm d}$ in Yunnan.

The above analyses provide the qualitative interpretation for the seasonal variations of stable isotopes in precipitation observed in monsoon regions.

3.2 *The seasonal variations of the δ^{18}O in precipitation*

By calculating the amount weighted-mean values of stable isotopic ratios in precipitation according to Eq. (2), the monthly mean δ^{18}O values in precipitation are available at Mengzi, Simao and Tengchong. Figure 4 gives the seasonal variations of the monthly amount weighted-mean δ^{18}O, the monthly precipitation, the monthly mean temperature and the monthly mean vapor pressure at Mengzi, Simao and Tengchong.

It can be seen that the relatively high δ^{18}O values always appear in the dry season, with the maximum δ^{18}O in April before the rainy season. With the onset of the rainy season in May, the precipitation and atmospheric humidity increase remarkably, meanwhile, the δ^{18}O values in precipitation decrease rapidly. However, the minimum δ^{18}O does not correspond with the maximum precipitation, which occurs in July (Mengzi and Simao) and September (Tengchong). By this token, precipitation is not the only factor generating the amount effect. Probably, the amount effect is related to the variations of atmospheric circulation (Posmentier et al., 2004) and vapor origins (Araguás-Araguás et al., 1998; Jouzel et al., 1997). These variations will influence the vapor transport, and ultimately, the variation of stable isotopes in the vapor.

For example, in the rainy season, the stable isotopic compositions in surface ocean water are depleted gradually because more and more fresh water from land is poured continuously into the oceans and because the intense rainfall with relatively light δ^{18}O dilutes the marine δ^{18}O pool. Consequently, this induces a corresponding depletion of stable isotopes in the vapor and precipitation. The minimum δ^{18}O in surface ocean water, vapor and rainfall probably appears in late summer or early fall. Moreover, during the end of the rainy season when the vapor transported from low-latitude oceans is obviously weakened, the vapor forming rainfall originates more from terrestrial transpiration. Although the precipitation has started to decrease, the stable isotopic ratios in precipitation still keep decreasing, instead of increasing, because the vapor provided by transpiration originates from the former rainfall

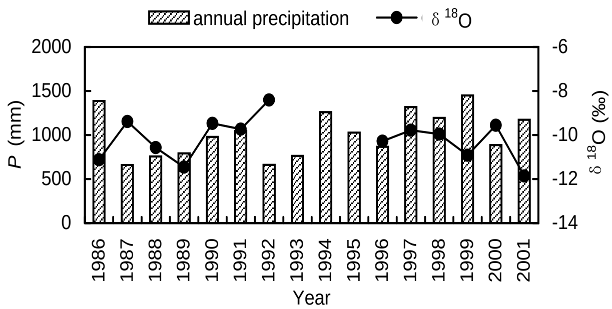

Fig. 5. Interannual variations of the annual amount weighted-mean δ^{18}O and annual precipitation at Kunming. The survey records of the δ^{18}O are from GNIP (IAEA/WMO, 2001), but absent from 1993 to 1995.

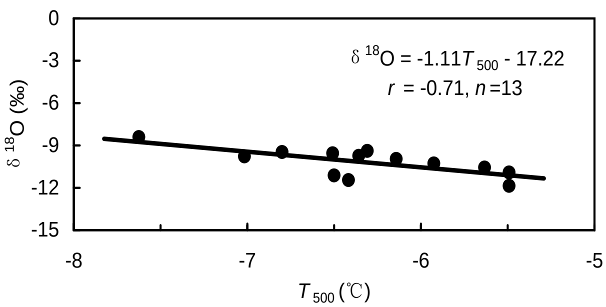

Fig. 6. Scatter plot of the annual amount weighted-mean δ^{18}O versus the annual mean temperature at 500 hPa at Kunming. [The data are from GNIP (IAEA/WMO, 2001). The linear regression equation with correlation coefficient r and sample size n is shown.]

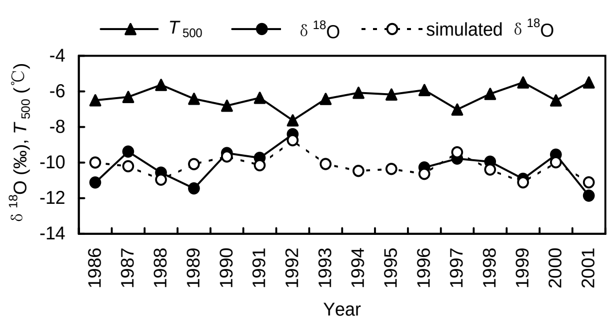

Fig. 7. Interannual variations of the annual amount weighted-mean δ^{18}O and annual mean temperature T_{500} at Kunming. The survey records of the δ^{18}O are from GNIP (IAEA/WMO, 2001), but absent from 1993 to 1995.

with low stable isotopic ratios. As is known, the transpiration process does not produce stable isotopic fractionation. As they are fractionated during their second condensation, the stable isotopes are depleted further.

3.3 *Annual variation of the* δ^{18}O *in precipitation*

The stable isotopic variation on the annual timescale can be used as an important means to recover palaeo-climatic and palaeo-environmental records in different sediments. Because the sampling records at the Mengzi, Simao and Tengchong stations are less than a year long, Kunming, with a relatively longer sampling series in the region (IAEA/WMO, 2001), is selected as the basic station in order to analyze the interannual variations of the δ^{18}O in precipitation.

The analyses find that, except for the marked amount effect, the monthly δ^{18}O in precipitation at Kunming also exhibit the marked negative correlations with the monthly mean temperature at the surface and at 500 hPa and with the geopotential height at 500 hPa, and the marked positive correlations with the dew-point deficit at the surface and at 500 hPa (Zhang et al., 2004; 2005). These features are similar to those at Mengzi, Simao and Tengchong.

On the annual timescale, although there are no marked correlations versus the annual mean temperature and the mean dew-point deficit at the surface, the annual amount weighted-mean δ^{18}O at Kunming displays the negative correlation versus annual precipitation, with a correlation coefficient of -0.44, significant at the 0.1 confidence level, showing that the magnitude of the annual mean δ^{18}O in precipitation may reflect, to a certain degree, the intensity of precipitation or the strength of the monsoon in the region. Figure 5 gives the variations of the annual amount weighted-mean δ^{18}O and the annual precipitation at Kunming.

Unlike the situation at the surface, the marked negative correlation appears between the annual amount weighted-mean δ^{18}O and annual the mean temperature at 500 hPa, with the correlation coefficient significant at the 0.01 confidence level (Fig. 6). On average, the lower mean annual δ^{18}O in precipitation matches the case with more precipitation, and higher value matches the case with less precipitation. However, as is known, the magnitude of precipitation cannot interpret the variation of δ^{18}O in precipitation.

Figure 7 gives the interannual variations of the annual amount weighted-mean δ^{18}O and the annual mean temperature at 500 hPa at Kunming. It can be seen that the two variations are opposite.

The amount weighted-mean δ^{18}O series are analyzed using the stepwise regression method. Among all factors mentioned above that were prepared for selection at the surface and at 500 hPa, only T_{500} is selected in the regression equation:

$$\delta^{18}\mathrm{O} = -1.11T_{500} - 17.22 \quad r = -0.71 \,. \tag{4}$$

This is a linear regression equation between δ^{18}O and T_{500}. The variation, simulated by Eq. (4), of the annual mean δ^{18}O (dashed line in Fig. 7) has a very good consistency with the actual mean δ^{18}O.

As in the above analyses, the simulation shows the negative correlations in the δ^{18}O with temperature and precipitation amount, whether on the synoptic timescale (Table 1) or on the monthly timescale (Fig. 4), which characterizes the stable isotopic variation in precipitation in the monsoon regions. Furthermore, on the annual timescale, such a feature is still notable, just like that shown in Figs. 5 and 6. This means that, in a year of plentiful precipitation, the stable isotopes in precipitation show low values, but in a year of scarce precipitation, high values are shown.

It needs to be pointed out that the temperature in the upper levels stands for, to a certain degree, the physical features of the air mass. In the years with an abnormally strong summer monsoon, more warm and wet air from low-latitude oceans is transported northward along the vapor channel located in Southwest China and produces abnormally strong rainfall on the way. Meanwhile, the abnormally strong condensation process that is produced will release more condensed latent heat into the atmosphere, and this will lead to the rise of atmospheric temperature during the rainfall but to a decline of the δ^{18}O in the precipitation. On the other hand, in the years with an abnormally weak summer monsoon, the precipitation and the atmospheric temperature during the rainfall decrease abnormally, but the δ^{18}O in the precipitation increases.

4. Discussions and conclusions

Forced by meteorological factors, the stable isotopes in precipitation have quasi-periodic variations (Rozanski et al., 1997; Yao et al., 1996). These variations are either consistent with that of temperature, showing up as the temperature effect (Jouzel et al., 1987, 1997; Zhang et al., 2004), or inverse to that of precipitation amount, showing up as the amount effect (Jouzel et al., 1987, 1997; Zhang et al., 2001, 2004). However, the occurrence of these effects does not mean that temperature or precipitation are the direct and only factors.

In low-middle-latitude oceans and monsoon regions, stable isotopes in precipitation are markedly depleted when the summer monsoon prevails but markedly enriched when the winter monsoon prevails. Such a seasonal distribution of stable isotopes in precipitation is considered to be associated with vapor origins (Wei et al., 1982; Wei and Lin, 1994; He et al., 2000). Wei et al. (1982) thought that the summer precipitation in coastal locations has low ^{18}O values because the moisture reaching the coast has gone through a long distance transport. Significant precipitation has fallen above the ocean, leaving the remaining vapor depleted in ^{18}O. In this case, why would the precipitation during the winter monsoon be enriched in stable isotopes in the same location? Some other researchers thought, according to qualitative analyses, that the stable isotopic ratios in vapor from a marine air mass have relatively low values and those from a continental air mass have relatively high values (He et al., 2000; Thompson et al., 2000). Below, two sets of statistical data may be helpful for comprehending there facts. The global isotopic survey found that the amount weighted-mean δ^{18}O values in the period from November to April are -6.84‰, -3.89‰, -3.94‰ and -4.85‰ respectively at Singapore, Colombo, Bangkok and Rangoon located in the places of origin of the water vapor in Southwest China (see Fig. 1), where the vapor forming precipitation is mainly from low-latitude oceans. However, in Northwest China, where the continental air mass controls precipitation throughout the whole year, the stable isotopes in precipitation are deeply depleted in the same period. For example, the amount weighted-mean δ^{18}O values from November to April are -16.18‰, -14.4‰ and -9.44‰ respectively at Ürümqi, Lanzhou and Yinchuan (Zhang et al., 2002, 2004). These values are distinctly lower than those in the vapor channel of Southwest China.

By a simulation, the δ^{18}O in initial condensate of vapor evaporated from low-latitude oceans is about -1.50‰ on average (Zhang et al., 2003). Actually, the precipitation δ^{18}O values at most sampling stations in low latitudes are all lower than this value (Zhang et al., 2004), showing that the precipitation is not the product of the initial condensation. In a convective system, the initially condensed droplets can be taken to the upper level by strong updraft. By collision with small drops condensed at low temperature and with low δ^{18}O, small drops become big ones that fall out of the cloud base and form rainfall. The stronger (weaker) the convection, the longer (shorter) the condensation process from small drops to great ones, the greater (smaller) the size of raindrops leaving the cloud base, the heavier (lighter) the rainfall, and thus the lighter (heavier) the stable isotopic compositions in the precipitation.

Furthermore, the stable isotopes in raindrops still undergo fractionation during the fall from the cloud base to the surface (Zhang et al., 2005). In a humid atmosphere, the evaporation enrichment in falling raindrops is light. The stable isotopic compositions in precipitation reaching the surface stays at a low level; however, in a dry atmosphere, the strong evaporation enrichment in falling raindrops causes the stable isotopes in precipitation reaching the surface to be enriched, which is an important reason why the maximum values of stable isotopic ratios in precipitation usually appear in dry regions or during the dry season in the global distribution.

To sum up, the following conclusions are made:

(1) At the Mengzi, Simao and Tengchong stations located in Southwest China, the δ^{18}O in precipitation has marked negative correlations versus precipitation amount, vapor pressure and atmospheric precipitable water on the synoptic timescale, showing that the amount effect and the humidity effect are notable in this region.

(2) Marked negative correlations appear between the δ^{18}O in precipitation and the diurnal mean temperature at 400 hPa, 500 hPa, 700 hPa and 850 hPa, which is different from the temperature effect in middle-high-latitude inland regions. In addition, the notable positive correlations between the δ^{18}O in precipitation and the dew-point deficit at different altitudes are found at Mengzi, Simao and Tengchong. The upper-level circulation patterns have a marked impact on stable isotopic variations in the precipitation.

(3) Precipitation is not the only factor generating the amount effect. Probably, the amount effect is associated with the variations of atmospheric circulation and vapor origins.

(4) On the annual time timescale, the annual amount weighted-mean δ^{18}O at Kunming displays marked negative correlations not only with annual precipitation but also with annual mean temperature at 500 hPa. The properties of the air mass are possible factors driving the stable isotopic variation in precipitation.

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