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Simultaneous monitoring of stable oxygen isotope composition in water vapour and precipitation over the central Tibetan Plateau

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Abstract

This study investigated the daily δ^{18} O variations of water vapour ($\delta^{18}O_v$) and precipitation ($\delta^{18}O_{p}$) simultaneously at Nagqu on the central Tibetan Plateau for the first time. The data show that the δ^{18} O tendencies of water vapour coincide strongly with those of associated precipitation. The δ^{18} O values of water vapour affect those of precipitation 5 not only on the same day, but also for the following several days. In turn, the δ^{18} O values of precipitation also affect those of water vapour. Hence, there exists an interaction between $\delta^{18}O_v$ and $\delta^{18}O_p$, and the interaction decreases gradually with time. During the entire sampling period, the variations of $\delta^{18}O_v$ and $\delta^{18}O_n$ at Naggu did not appear dependent on temperature, but did seem significantly dependent on the joint contribu-10 tions of relative humidity, surface pressure, and precipitation amount. In addition, the δ^{18} O changes in water vapour and precipitation can be used to diagnose different atmospheric trajectories, especially the influences of the Indian monsoon and convection. Moreover, intense activities of the Indian monsoon and convection may cause the enrichment of $\delta^{18}O_p$ relative to $\delta^{18}O_v$ at Nagqu (on the central Tibetan Plateau) to differ 15 from that at other stations on the northern Tibetan Plateau. These results indicate that the effects of different moisture sources, including the Indian monsoon and convection currents, need be considered when attempting to interpret paleoclimatic records on the central Tibetan Plateau.

20 **1** Introduction

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The Tibetan Plateau is a natural laboratory for studying the influences of different moisture sources, which include polar air masses from the Arctic, continental air masses from central Asia, and maritime air masses from the Indian and Pacific Oceans (Bryson, 1986), and for reconstructing paleoclimate variations (An et al., 2001). The stable oxygen isotope (δ^{18} O) provides an important tracer for understanding atmospheric moisture cycling, especially by using the δ^{18} O records in all three phases of water (Dans-



gaard, 1964; Lee et al., 2005). Oxygen isotopes also act as important indicators for reconstructing paleoclimates by using their records preserved in ice cores (Thompson et al., 2000), speleothems (Cai et al., 2010), tree rings (Treydte et al., 2006; Liu et al., 2014), and lake sediments (Zech et al., 2014). Variations of δ^{18} O result from isotope fractionation processes that may be influenced by temperature, rainout, amount effects, and different moisture sources (Dansgaard, 1964; Jouzel and Merlivat, 1984; Rozanski et al., 1992).

To better understand atmospheric moisture transport to the Tibetan Plateau and surrounding regions, the Chinese Academy of Sciences (CAS) established an observation network in 1991 to continually survey δ^{18} O variations in precipitation on the Plateau (the Tibetan Plateau Network of Isotopes in Precipitation, TNIP) (Zhang et al., 1995; Tian et al., 2001; Yu et al., 2008; Yao et al., 2013). Previous studies have shown that δ^{18} O variations in precipitation on the southern Tibetan Plateau differ distinctly from those on the northern Tibetan Plateau (Tian et al., 2003; Yu et al., 2008; Yao et al., 2013). In addition, many scientists have investigated the roles of various climatic fac-

- tors, especially the Asian monsoon's influence on δ^{18} O in precipitation (Aizen et al., 1996; Araguás-Araguás et al., 1998; Posmentier et al., 2004; Vuille et al., 2005; Yu et al., 2014a). Recent studies have also investigated δ^{18} O in river water (Bershaw et al., 2012), lake water (Yuan et al., 2011), and plant water (Zhao et al., 2011; Yu
- et al., 2014b). In comparison, only a few studies have focused on δ^{18} O from water vapour over the Tibetan Plateau (Yatagai et al., 2004; Yu et al., 2005; Kurita et al., 2008; Yin et al., 2008). Moreover, a gap exists in the studies regarding the interaction of the δ^{18} O values from water vapour and precipitation, and on the δ^{18} O enrichment between water vapour and precipitation over the Tibetan Plateau (In this study, the "en-
- ²⁵ richment" was defined as the difference of the δ^{18} O values of precipitation ($\delta^{18}O_p$) and vapour ($\delta^{18}O_v$), $\Delta\delta^{18}O = \delta^{18}O_p - \delta^{18}O_v$). An improved understanding of $\delta^{18}O$ as tracers of water movement in the atmosphere and as indicators of climate change requires detailed knowledge of the isotopic compositions in all three phases of water (Lee et al., 2005). In contrast to liquid or solid precipitation, measurements of $\delta^{18}O$ in water vapour



can be taken across different seasons and synoptic situations, and are not limited to rainy days (Angert et al., 2008). Hence, δ^{18} O in water vapour has become an important issue in the fields of paleoclimatology, hydrology (lannone et al., 2010), and ecology (Lai et al., 2006), especially for understanding different moisture sources.

- ⁵ With this background, we launched a project in the summers of 2004 and 2005 to collect simultaneous water vapour and precipitation samples at Nagqu (31°29′ N, 92°04′ E, 4508 m a.s.l.) on the central Tibetan Plateau (the first such study), despite the difficultly of collecting water vapour samples at this high elevation. Based on the δ^{18} O data sets from these samples, this paper discusses the interaction between the δ^{18} O values in water vapour and precipitation, considers the effects of various meteorological pa-
- rameters on the δ^{18} O of water vapour and precipitation, and attempts to explain the relationships between the isotopic compositions of samples and atmospheric trajectories.

2 Sampling sites, materials, and methods

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¹⁵ The Nagqu station lies in the middle of a short grass prairie, in a sub-frigid, semi-humid climate zone between the Tanggula Mountains and the Nyainqentanglha Mountains (Fig. 1). The annual average temperature at this station was recorded as -2°C, with an annual mean relative humidity of 50%, and average annual precipitation of 420 mm. Most of the rainfall at this site occurred during May through August and included about 20 77% of the annual precipitation.

This study collected water vapour samples at Nagqu during the periods of August– October 2004 and July–September 2005. On the basis of previous study, if the condensation temperature falls below -70 °C, the sampling method diminishes the correction factor (-0.07 ‰) to below the typical error value quoted for ¹⁸O analyses by modern mass spectrometers (Schoch-Fischer et al., 1984). Our study extracted water vapour cryogenically from the air, by pumping it slowly through a glass trap immersed in ethanol, which was continuously maintained as low as -70 °C with a set of electric



cryogenic coolers driven by a compressor (Yu et al., 2005). Thus the captured water vapour should faithfully reflect the water vapour in the atmosphere and minimize fractionation during the sampling. Moreover, the cold trap was made in a linked-ball shape to increase the surface area for condensation (Hübner et al., 1979), and to en-

- ⁵ sure complete removal of all the water vapour, in order to avoid isotope fractionation during sampling (Gat et al., 2003). In addition, duplicate analyses were conducted to ensure the trapping did minimize fractionation. The validity of the cold trap operation was rechecked by connecting an extra glass trap to the outlet of the original trap; within which no visible condensed vapour was found, reconfirming the validity of the water
- ¹⁰ vapour sampling method. A flow meter controlled the air flow rate. Air was drawn at a rate of about 5 L min⁻¹ (Gat et al., 2003) for about 24 h through a plastic tube attached to the rooftop of the Nagqu station (the height of the roof is about 6 m). At the end of each sampling, the two ends of the cold trap were sealed, and the samples melted at room temperature. Water was mixed across the trap before decanting it into
- ¹⁵ a small vial and sealing. One sample of about 10 mL was collected each day. In addition, rainfall from each precipitation event at the Nagqu Meteorological Station (close to the vapour sampling site) was collected immediately and sealed in clean and dry plastic bottles. A total of 153 water vapour samples and 90 precipitation samples were collected. All the samples were stored below –15 °C until analyzed. During the samples are plane metaorological parameters, such as temperature at 1.5 m temperature.
- ²⁰ pling period, some meteorological parameters, such as temperature at 1.5 m, temperature near ground, relative humidity, surface pressure, and precipitation amount were recorded.

The Key Laboratory of Tibetan Environment Changes and Land Surface Processes, Institute of Tibetan Plateau Research (Chinese Academy of Sciences, Beijing) performed the measurements of the oxygen isotopic compositions of all samples, using a MAT-253 mass spectrometer, with a precision of 0.2 parts per mil (‰) for the oxygen isotope ratios (δ^{18} O). The H₂O-CO₂ isotopic exchange equilibration method was adopted for the oxygen isotope ratios (δ^{18} O) measurements. This study expresses the measured oxygen isotope ratios (δ^{18} O) as parts per mil (‰) of their deviations, relative



to the Vienna Standard Mean Ocean Water (VSMOW). Unfortunately, deuterium data at Nagqu were not available for this project.

To identify the moisture transport paths and interpret δ^{18} O variability further in the time series, our study determined 120 h back trajectories for air parcels during the entire sampling period, using the NOAA HYSPLIT model (Draxler and Rolph, 1998) and NCEP reanalysis data sets (available at: ftp://arlftp.arlhq.noaa.gov/pub/archives/ reanalysis). The origin of air masses as diagnosed from the back trajectory analysis appears to approximate the moisture source direction for the water vapour and for the precipitation at the study site (Guan et al., 2013). The trajectories originated at 1000, 2000, and 3000 m above ground level (a.g.l.), respectively.

3 Results and discussion

3.1 Interaction and enrichment between δ^{18} O of water vapour and precipitation

Figure 2 displays the temporal changes of δ^{18} O in water vapour (δ^{18} O_v) and in precipitation (δ^{18} O_p) at Nagqu. Clearly, the trends of δ^{18} O_v closely approximate those of δ^{18} O_p (Fig. 2a and b). A strong positive relationship existed between δ^{18} O_v and δ^{18} O_p during the entire sampling period of 2004–2005 (δ^{18} O_v = 0.72 δ^{18} O_p – 14.43, r = 0.81, n = 86, p < 0.01) (Fig. 2c). Moreover, the positive correlations between δ^{18} O_v and δ^{18} O_p, whether in 2004 (δ^{18} O_v = 0.73 δ^{18} O_p – 14.39, r = 0.81, n = 42, p < 0.01), or in 2005 (δ^{18} O_v = 0.71 δ^{18} O_p – 14.85, r = 0.78, n = 44, p < 0.01), show similarities (Fig. 2c). The condensation of water vapour results in the observed precipitation. Hence, water vapour plays a key role in all precipitation events. As a result, the isotopic composition of the water vapour has a direct effect on that of the precipitation. Similar close relationships between δ^{18} O_v and δ^{18} O_p also exist at Heidelberg (Jacob and Sonntag, 1991) and at Ankara (Dirican et al., 2005). The isotopic composition of water vapour not only affects that of precipitation on the same day, but also affects that



of precipitation for several days thereafter. As shown in Table 1, the isotopic composition of water vapour correlated positively with that of precipitation over the following three days, with correlation coefficients of 0.48, 0.45, and 0.33 (within a 0.01 confidence limit), respectively. Nevertheless, the correlation coefficients decreased gradually with

- time. In particular, the correlation coefficient for the fourth day decreased to as low as 0.28, and only within a 0.05 confidence limit (Table 1). In addition, the slope decreased gradually from 0.90 to 0.31 over five days (Table 1). Because water vapour provided the primary moisture source for the precipitation, these isotopic exchanges had an effect on the vapour with which the raindrop equilibrates (Angert et al., 2008). During
- the rain event, water vapour rapidly interacts with raindrops and tends to move toward isotopic equilibrium (Deshpande et al., 2010). Thus, these exchanges were particularly significant at the same day, but gradually weakened over the four days after the initial rainfall event. On the other hand, precipitation influences water vapour at the local scale. As the raindrop falls, the content of the raindrop will contribute to the ambient
- ¹⁵ water vapour, due to the re-evaporation effect. As a result, the isotopic composition of raindrops also contributes to that of the ambient water vapour. Even as the raindrops fall, the isotopic composition of the residual water vapour changes because of a "rainout effect". Consequently, the isotopic composition of precipitation has a feedback effect on that of the water vapour. We show that the isotopic composition of precipitation
- ²⁰ affects that of water vapour, not only on the same day, but also for the next four days, resulting in correlation coefficients of 0.69, 0.64, 0.59, and 0.41 (within a 0.01 confidence limit), respectively (Table 2). Clearly, the correlation coefficients and the slopes also decrease gradually as over time, with the correlation coefficient for the fifth day decreasing even further (as low as 0.35) and correlated only within a 0.05 confidence
- ²⁵ limit (Table 2). Correspondingly, the slopes decreased gradually from 0.72 to 0.34. This may result from surface water evaporation from recent precipitation contributing to the isotopic composition of the local water vapour in the days following the rainfall event. Apparently, there exists an interaction between the δ^{18} O values of water vapour and of precipitation, and the interaction decreases gradually over time. Pfahl et al. (2012)



also found a microphysical interactions between rain drops and water vapour beneath the cloud base exists by using model.

Compared with the $\delta^{18}O_v$ values, the $\delta^{18}O_p$ values experienced significant enrichment at Nagqu in 2004 and 2005. Furthermore, the enrichment of $\delta^{18}O_p$ relative to $\delta^{18}O_v$ ($\Delta\delta^{18}O = \delta^{18}O_p - \delta^{18}O_v$) in 2004 (8.2%) was similar to that in 2005 (8.2%), even though the sampling period in 2004 differed from that in 2005. The average enrichment at Nagqu in 2004–2005 was 8.2%. In comparison, the average enrichment of $\delta^{18}O_p$ relative to $\delta^{18}O_v$ at the Delingha station (37°22′ N, 97°22′ E, 2981 m; see Fig. 1) on the northern Tibetan Plateau ($\Delta\delta^{18}O = 10.7\%$) (Yin et al., 2008), was lower. This resulted from more intense Indian monsoon and convection activities at Nagqu compared with those at Delingha. Due to the combined impact of the intense Indian

- monsoon and convection activities, the summer $\delta^{18}O_p$ values at Nagqu were more depleted than those at Delingha (Yu et al., 2008). As a consequence, the $\Delta \delta^{18}O$ value at Nagqu fell below that at Delingha. Further south, the enrichment of $\delta^{18}O_p$ relative
- ¹⁵ to $\delta^{18}O_v$ at the Bay of Bengal (Fig. 1) was 8.6% (Midhun et al., 2013) (similar to that at Nagqu), where the Indian monsoon exceeds the intensity of that at Nagqu, even though the oceanic moisture does not rise to the same degree as at Nagqu. We note that the enrichment of $\delta^{18}O_p$ relative to $\delta^{18}O_v$ at the Nagqu station differs from that at the northern station (Delingha), but resembles that at the southern station (Bay of Bengal), apparently because of its unique location, which is affected by both the Indian monsoon and convection. The next section discusses the influences of those activities on water vapour/precipitation $\delta^{18}O$ changes in detail.

3.2 The effects of meteorological and environmental factors on δ^{18} O of water vapour and precipitation

²⁵ A number of meteorological parameters affect the δ^{18} O variations of water vapour and precipitation. In particular, different processes dominate the relative humidity variations in different regions, resulting in different isotope ratios in the water vapour (Noone,



2012). The data from Palisades (USA) show that stable isotopic compositions of water vapour correlate positively with relative humidity (White et al., 1984). Wen at al. (2010) also found that a positive correlation exists between water vapour δ^{18} O and relative humidity at Beijing (China). At a North Greenland site, both diurnal and intra-5 seasonal variations show strong correlations between changes in local surface humidity and water vapour isotopic composition (Steen-Larsen et al., 2013). In addition, the water vapour δ^{18} O trends from Bermuda Islands (the North Atlantic) also resemble those of relative humidity (Steen-Larsen et al., 2014). Interestingly, the tendencies of $\delta^{18}O_v$ and $\delta^{18}O_p$ in our study oppose those of relative humidity (Fig. 3). Hence, at Naggu the δ^{18} O values of water vapour and of precipitation correlate negatively 10 with relative humidity (RH): ($\delta^{18}O_v = -0.20$ RH – 12.07, r = -0.45, n = 153, p < 0.01; $\delta^{18}O_{p} = -0.28RH + 2.79$, r = -0.36, n = 90, p < 0.01). Moreover, the tendencies of $\delta^{18}O_v$ and $\delta^{18}O_n$ in our study clearly differed from those of surface temperature at 1.5 m or ground temperature at 0 m during the entire sampling period (Fig. 3). No positive correlation was found between the δ^{18} O values and temperature. Thus, the 15 changes in the δ^{18} O values of water vapour and precipitation did not depend on the changes in temperature, and did not experience a "temperature effect". However, on the northern Tibetan Plateau, the δ^{18} O composition of water vapour and precipitation correlated positively with temperature (Yin et al., 2008). A positive correlation between the isotope record of water vapour and temperature (T) was also found at Heidelberg 20 (Germany), western Siberia, southern Greenland, and Minnesota (USA) (respectively, Schoch-Fischer et al., 1984; Bastrikov et al., 2014; Bonne et al., 2014; Welp et al., 2008). Clearly, the relationships between $\delta^{18}O - T$ and $\delta^{18}O - RH$ at our station differ from those at other stations. Those and the δ^{18} O depletion during the summer monsoon period (Fig. 3a and f) may reflect the influences of the Indian monsoon (Yu et al., 25 2008) and increasing convection (Tremoy et al., 2012). Due to an uplift effect of the massive mountains (such as the Himalayas), warm oceanic moisture transported by the Indian monsoon from the Indian Ocean onto the Tibetan Plateau rises to very high elevations, where very low temperatures prevail (Tian et al., 2003; Yu et al., 2008). This



rise results in the more depleted δ^{18} O values recorded in summertime water vapour and precipitation at Naggu. Moreover, the intense convection raises the oceanic moisture to higher elevations. Hence, the convection effect for the oceanic moisture increases the more depleted δ^{18} O in water vapour and precipitation in our study region (Yu et al., 2008). However, during the monsoon period, the corresponding surface air 5 temperature and the summer rainfall greatly exceed those during the pre-monsoon and post-monsoon periods (Fig. 3). Accordingly, an inverse correlation exists between δ^{18} O in water vapour/precipitation and surface air temperatures and rainfall, respectively, indicating the lack of a "temperature effect" on δ^{18} O in water vapour/precipitation in this study region. Furthermore, the δ^{18} O trends coincide with surface pressure (Pres) dur-10 ing the entire sampling period ($\delta^{18}O_v = 1.11$ Pres – 681.88, r = 0.41, n = 153, p < 0.01; $\delta^{18}O_{\rm p} = 1.09$ Pres – 658.73, r = 0.34, n = 90, p < 0.01). High precipitation amounts correspond to depleted isotope compositions of water vapour and precipitation, and low precipitation amounts correspond to enriched isotope compositions. Specifically, the isotope compositions of precipitation exhibit greater enrichment when there has been 15 no rainfall (P[precipitation amount] = 0) (Fig. 3a, f, e and j). This demonstrates that precipitation amount also affects the δ^{18} O variations of water vapour and precipitation at Nagqu. During precipitation events, the water vapour generally maintains a state of equilibrium with falling raindrops (Lee et al., 2006). During heavy precipitation events, the isotope ratios of water vapour and condensate decrease as satu-20 rated air rises, because of continued fractionation during condensation (Gedzelman and Lawrence, 1982); and the δ^{18} O values of precipitation tend to become more depleted (Fig. 3a and f). Correspondingly, heavily depleted δ^{18} O values of residual water vapour occur, due to the rainout effect. During periods without precipitation, water vapour deviates far from saturation, i.e., may exhibit low relative humidity. In these cir-25 cumstances, the δ^{18} O values of water vapour become highly enriched (Fig. 3a and f). To further reveal the relationships between the δ^{18} O values and various meteorological parameters, our study modeled δ^{18} O as a function of temperature, relative humidity, surface pressure, and precipitation amount, using a simple multiple regression model.



Using a *stepwise* method and based on the output of this model, the variable of temperature was excluded. The function can be expressed as:

 $\delta^{18}O_v = -502.80 - 0.11RH + 0.82Pres - 0.28P$ (*p* for RH, Pres, and *P* is 0.001, 0.000, 0.000, respectively; *F* = 28.276, *F*_a = 5.709, *F* > *F*_a, *a* = 0.001)

 $\delta^{18}O_p = -580.66 - 0.18RH + 0.98Pres - 0.26P$ (*p* for RH, Pres, and *P* is 0.022, 0.001, 0.002, respectively; *F* = 15.249, *F*_a = 5.932, *F* > *F*_a, *α* = 0.001).

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The multiple correlation coefficients (*R*) between all of the independent variables (relative humidity, surface pressure, and precipitation amount) and the dependent variables ($\delta^{18}O_v$ and $\delta^{18}O_p$) are 0.60 and 0.56; and the F-statistics are significant at the

¹⁰ 0.001 and 0.001 levels, respectively. In brief, the δ^{18} O changes in water vapour and precipitation at Nagqu relate closely to the joint contributions of relative humidity, surface pressure, and precipitation amount.

In addition, the land surface characteristics and processes such as evaporation and transpiration may also have affected the isotopic ratios of the water vapour. During dry periods, the land surface dries due to evapotranspiration; and the moisture in soil and grass (characterized by relatively enriched isotopic values) evaporates into the atmosphere. Therefore, the isotopic ratio of the water vapour becomes relatively enriched (Fig. 3a and f). That is why the isotope compositions of water vapour become more enriched during days with no rainfall, compared to during days with rainfall. Dur-

- ²⁰ ing heavy rain events, however, local evapotranspiration is extremely weak (Huang and Wen, 2014), because clouds and precipitation cool the surface and moisten the bound-ary layer, leading to high relative humidities (Fig. 3c and h) (Aemisegger et al., 2014). Therefore, effects of local evapotranspiration on the changes in water vapour δ^{18} O can be ignored during the such rainy periods, and the corresponding δ^{18} O values in water
- ²⁵ vapour become more depleted (Fig. 3a and f). On cessation of the rain, clouds clear, the ground heats up again, and the relative humidity decreases, partly due to warming, partly due to reduced humidity (Aemisegger et al., 2014). In that case, local evapotranspiration will contribute to the changes in water vapour δ^{18} O, which will quickly return to relatively enriched values (Fig. 3a and f) (Deshpande et al., 2010). Another



short-term study of Kurita et al. (2008), not far from this study area, also demonstrated that water vapour increased gradually, accompanied by an increased contribution of evapo-transpired water that had relatively enriched isotopic values.

3.3 δ^{18} O changes in water vapour and precipitation related to different atmospheric trajectories

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Synoptic weather circulation (especially atmospheric trajectories) strongly affects the variations of stable isotopic compositions of water vapour and precipitation (Strong et al., 2007; Pfahl and Wernli, 2008; Deshpande et al., 2010; Guan et al., 2013). This study used the NOAA HYSPLIT model to calculate 120 h back trajectories of air parcels for each day of the entire sampling period. Figure 4 shows a subset of the results of the atmospheric trajectories. The results of 12 July, 6 August, 26 August, and 5 September 2005, represent the weak monsoon, the active monsoon, the late monsoon, and the post-monsoon period conditions, respectively. During the weak monsoon period, moisture over Naggu at 1000 m a.g.l. appears to derive predominantly from the coastal

- regions of Bengal in the south, which might have been transported earlier by the Indian monsoon and lingered there. In this way, the coastal regions of Bengal act as a moisture reservoir during the weak monsoon period. Clearly, moisture from 2000 m and 3000 m a.g.l. recycles from the westerlies (which are associated with enriched surface waters that re-evaporate and with evaporated surface water under lower humidity
- ²⁰ conditions), and this contributes to the moisture over Nagqu during the weak monsoon period (Fig. 4a). Therefore, $\delta^{18}O_v$ and $\delta^{18}O_p$ values show relative enrichment (such as –17.8 and –14.7 ‰ observed on 12 July 2005) (Fig. 2b).

Compared to the weak monsoon period (Fig. 4a), the contribution of moisture from the westerlies and regional circulation decreased during the active monsoon period

²⁵ (Fig. 4b) (the specific humidity fells to $2 g kg^{-1}$ over Nagqu). Due to the dominant Indian monsoon circulation during this period, most moisture at the 1000 m a.g.l. of the trajectories came from this direction. As a result, the specific humidity over Nagqu from this pathway increased to $7 g kg^{-1}$ (Fig. 4b). In addition, the trajectories of the 2000 m a.g.l.



airflow came from the southern slope of the Himalayas (Fig. 4b). The moisture from both of those two paths were uplifted by the high mountains. Moreover, convection over the Tibetan Plateau often occurs in the region between the two major east-west mountain ranges, the Nyainqentanglha Mountains and the northern Himalayas (Fuji-

- ⁵ nami et al., 2005). As mentioned above, intense convection over the Tibetan Plateau, combined with uplift caused by the high mountains, causes oceanic moisture to rise to very high elevations. Obviously, convection of marine and continental air masses not only causes isotopic variations of water vapour (Farlin et al., 2013), but also significantly affects the isotopic composition of the precipitation (Risi et al., 2008). In particular, the
- ¹⁰ period of time over which convection significantly affects the isotopic composition of precipitation relates to the residence time of water within atmospheric reservoirs (Risi et al., 2008). Hence, an interaction exists between the isotopic composition of water vapour and precipitation. This results in the more depleted δ^{18} O values of water vapour and precipitation at Nagqu, such as -32.1 and -21.7% on 6 August 2005 (Fig. 2b).
- ¹⁵ The corresponding maximum precipitation amount of 25.9 mm over Nagqu was observed during this sampling period in 2005 (Fig. 3j). Purushothaman et al. (2014) also reported highly depleted nature of water vapour at Roorkee (north India) during rainy period, due to intense Indian monsoon.

Although moisture over Nagqu derived from the Bay of Bengal decreased during the late monsoon period, some of the trajectories continued to originate in the coastal regions. Figure 4c details one selected event on 26 August 2005, during which the trajectories came from the coastal regions of west India (near the Arabian Sea). The specific humidity over Nagqu from those pathways decreased to 2–6 g kg⁻¹, compared with those during the active monsoon period. The moisture from those paths were uplifted by the high mountains, via the Indian continent, and also contributed to the relatively depleted δ^{18} O values of water vapour and precipitation (-32.6, -25.0‰) (Fig. 2b).

Trajectories after the rainy season (such as 5 September 2005; accompanying the Indian monsoon retreat) show that all of the moisture had been recycled from the conti-



nent (Purushothaman et al., 2014): (1) moisture from the regional circulation dominated the moisture sources in the study area, and (2) moisture from the westerlies also affected the Nagqu region (Fig. 4d). During this period, no contributions from the Bay of Bengal or the coastal regions of Bengal/west India appear to have significantly enriched

⁵ δ¹⁸O values of water vapour (such as -17.5% on 5 September 2005) (Fig. 2b). During the dry season, the specific humidity over Nagqu from those pathways decreased below 3 g kg⁻¹, and isotopic re-equilibration of rain droplets with surrounding water vapour appear to affect the δ¹⁸O variations of precipitation (Sturm et al., 2007). Consequently, the δ¹⁸O values of precipitation increased rapidly during the post-monsoon period (to -10.4‰) (Fig. 2b).

In summary, during the summer period, moisture over the Nagqu region of the central Tibetan Plateau originates primarily from the southern portion of the Tibetan Plateau, as well as the southern slope of the Himalayas, the coastal regions of Bengal/west India, and the Bay of Bengal; all of which are strongly influenced by the Indian monsoon and convection. In contrast, convection on the northern Tibetan Plateau is weaker than that on the central Tibetan Plateau, and the westerlies prevail on the northern Tibetan Plateau, almost without any influence of the Indian monsoon (Tian et al., 2003; Yu et al., 2008). Different moisture sources cause different effects on the δ^{18} O values of water vapour and precipitation at the two stations of Nagqu and Delingha, located on the central and northern Tibetan Plateau, respectively. This results in different δ^{18} O characteristics of water vapour and precipitation from the central and northern Tibetan Plateau and may explain the different δ^{18} O characteristics of ice cores from the central and northern Tibetan Plateau. For example, the δ^{18} O record preserved in

the Dunde ice core from the northern Tibetan Plateau provides a reasonable proxy of summer temperature (Thompson et al., 1989), while the δ^{18} O record in the Tanggula ice core from the central Tibetan Plateau shows no correlation between average δ^{18} O values and temperature, probably due to the influence of the Indian monsoon (Joswiak et al., 2010). Accordingly, our findings indicate that the influences of different moisture sources and the activities of the Indian monsoon and convection may be signifi-



cant when reconstructing paleoclimate variations on the central and northern Tibetan Plateau. Certainly, the ice core (or other proxy) δ^{18} O records do not reflect day-to-day changes of δ^{18} O in water vapour/precipitation. In order to disprove the presence of a temperature effect over the central Tibetan Plateau, multiple years of data and data that span the entire year will be needed from future studies. Hence, the authors have

⁵ that span the entire year will be needed from future studies. Hence, the authors have launched a new project to survey a longer time series of isotopic compositions of water vapour and precipitation (δ^{18} O and δ D), which should provide greater confidence in our findings and gain better understanding of the links between the water vapour and precipitation δ^{18} O/ δ D values and paleoclimatic records.

10 4 Conclusions

This study represents the first simultaneous water vapour and precipitation δ^{18} O time series for the central Tibetan Plateau. In the study region of Nagqu, the isotopic composition of water vapour has a direct relationship to that of precipitation. In turn, the isotopic composition of precipitation provides a feedback effect on that of water vapour. Hence, an interaction between the δ^{18} O values of water vapour and precipitation clearly exists, and evidence shows that the interaction decreases gradually over time. The $\delta^{18}O_v$ and $\delta^{18}O_p$ variations at Nagqu appear mainly controlled by joint influences of relative humidity, surface pressure, and precipitation amount, but did not demonstrate a "temperature effect". Moreover, the different $\delta^{18}O$ characteristics of water vapour and precipitation at Nagqu appear to relate to different atmospheric tra-

- jectories, especially involving the influences of the Indian monsoon and convection. The enrichment of $\delta^{18}O_p$ relative to $\delta^{18}O_v$ at Nagqu (on the central Tibetan Plateau) is similar to that at the southern station (Bay of Bengal), but differ from that at the northern station (Delingha), due to intense Indian monsoon and convection activities.
- ²⁵ These results may explain the different δ^{18} O characteristics obtained from ice cores from the central and the northern Tibetan Plateau. Our findings presented here may provide a basis for reinterpretation of the δ^{18} O records in ice cores from the central



Tibetan Plateau, and suggest that the impacts of different moisture sources, the Indian monsoon, and convection activities all need to be considered.

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 Colorado, USA (ftp://arlftp.arlhq.noaa.gov/pub/archives/reanalysis). Some meteorological data were provided by the Climatic Data Center, National Meteorological Information Center, China Meteorological Administration.

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Table 1. Correlations between $\delta^{18}O_v$ and $\delta^{18}O_p$ at Nagqu. The x and y represent $\delta^{18}O_v$ and
$\delta^{18}O_p$ during the same day (Day _n), and the y_1, y_2, y_3 , and y_4 show $\delta^{18}O_p$ in the following first
day $(Day_{n+1}), \dots (Day_{n+2}), \dots$, and the following fourth day (Day_{n+4}) , respectively.

$\delta^{18}O_v - \delta^{18}O_p$	Linear regression	Slope	R^2	r	п	p
Day _n – Day _n	y = 0.90x + 6.9	0.90	0.65	0.81	86	< 0.01
Day _n – Day _{n+1}	$y_1 = 0.55x - 2.9$	0.55	0.23	0.48	84	< 0.01
Day _n – Day _{n+2}	$y_2 = 0.49x - 4.5$	0.49	0.20	0.45	84	< 0.01
Day _n – Day _{n+3}	$y_3 = 0.36x - 8.1$	0.36	0.11	0.33	83	< 0.01
$Day_n - Day_{n+4}$	$y_4 = 0.31x - 9.7$	0.31	0.08	0.28	82	< 0.05



Table 2. Correlations between $\delta^{18}O_p$ and $\delta^{18}O_v$ at Nagqu. The *x* and *y* represent $\delta^{18}O_p$ and $\delta^{18}O_v$ on the same day (Day_n), and the y_1, y_2, y_3, \ldots , and y_5 represent $\delta^{18}O_v$ in the following first day (Day_{n+1}), ... (Day_{n+2}), ... (Day_{n+3}), ..., and the following fifth day (Day_{n+5}), respectively.

$\delta^{18}O_p$ - $\delta^{18}O_v$	Linear regression	Slope	R^2	r	п	р
Day _n – Day _n	y = 0.72x - 14.5	0.72	0.65	0.81	86	< 0.01
$Day_n - Day_{n+1}$	$y_1 = 0.61x - 16.4$	0.61	0.47	0.69	86	< 0.01
Day _n – Day _{n+2}	$y_2 = 0.62x - 15.9$	0.62	0.41	0.64	85	< 0.01
Day _n – Day _{n+3}	$y_3 = 0.57x - 16.7$	0.57	0.35	0.59	82	< 0.01
Day _n – Day _{n+4}	$y_4 = 0.38x - 20.2$	0.38	0.17	0.41	83	< 0.01
$Day_n - Day_{n+5}$	$y_5 = 0.34x - 20.8$	0.34	0.12	0.35	85	< 0.05





Interactive Discussion



Figure 2. Temporal changes of δ^{18} O in water vapour (δ^{18} O_v) and precipitation (δ^{18} O_p) and the enrichment of δ^{18} O_p relative to δ^{18} O_v at Nagqu in 2004 (**a**) and 2005 (**b**), respectively, and the relationships between δ^{18} O_p of precipitation and δ^{18} O_v of water vapour at Nagqu (**c**). Note that in Panel (**c**), the values in 2004 are shown as pink open circles; the values in 2005 shown as green solid dots.











Figure 4. Back trajectories calculated by HYSPLT at 1000 (red lines), 2000 (blue lines), and 3000 m (green lines) a.g.l. on 12 July, 6 August, 26 August, and 5 September 2005, representing the conditions during the weak monsoon **(a)**, the active monsoon **(b)**, the late monsoon **(c)**, and the post-monsoon **(d)** periods, respectively, over the Nagqu station. Note that changes in specific humidity (gkg^{-1}) along the air parcel pathways are also shown.

