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Effects of changes in moisture source and the upstream rainout on stable isotopes in summer precipitation – a case study in Nanjing, East China

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In the Asian monsoon region, variations in the stable isotopic composition of speleothems have often been attributed to the "amount effect". However, an increasing number of studies suggest that the "amount effect" in local precipitation is insignificant or even non-existent. To explore this issue further, we examined the variability of daily stable isotopic composition (δ^{18} O) in summer precipitation of 2012–2014 in Nanjing, East China. We found that δ^{18} O was not significantly correlated with local rainfall amount, but could be linked to changes in the location and rainout processes of precipitation source regions. Our findings suggest that the stable isotopes in precipitation could signal the location shift of precipitation source regions in the intertropical convergence zone (ITCZ) over the course of the monsoon season. As a result, changes in moisture source location and upstream rainout effect should be taken into account when interpreting the stable isotopic composition of speleothems in the Asian monsoon region.

1 Introduction

The "amount effect" refers to the observed negative correlation between the isotopic composition in precipitation and rainfall amount. It was first put forward by Dansgaard (1964), and is generally observed in low-latitude regions (Araguás-Araguás et al., 1998). Based on this relationship, stable isotopic records obtained from low-latitude regions are often used for paleohydroclimate reconstructions (e.g., Cruz et al., 2005, 2009; Partin et al., 2007; Tierney et al., 2008; Sano et al., 2012). However, some recent studies suggest that the "amount effect" is insignificant or even non-existent in low-latitude monsoon areas. For example, Conroy et al. (2013) found spatial and temporal examples of precipitation—isotope mismatches across the tropical Pacific, indicating that factors beyond the "amount effect" influence precipitation isotope variability. They compared 12 isotope-equipped global climate models to assess the distribution of

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simulated stable isotopic variability. Model simulations support observations in the western tropical Pacific, showing that monthly δ^{18} O are correlated with large-scale, not local, precipitation (Conroy et al., 2013). Peng et al. (2010) also found no significant correlation between precipitation amount and $\delta^{18}{\rm O}$ values in the western Pacific monsoon region near Taiwan. They suggest that moisture sources of diverse air masses with different isotopic signals are the main factor controlling the precipitation isotopic characteristics. Breitenbach et al. (2010) observed no empirical amount effect at their study site in the monsoonal northeast India. They identified a strong trend towards lighter isotope values over the course of the summer monsoon, with lowest δ^{18} O and δ D values in late monsoon season, with a temporal offset between the highest rainfall and the most negative δ^{18} O. Other observations (Lawrence et al., 2004: Kurita et al., 2009) show that at marine island stations, even short-term (daily or event-based) isotopic variations are independent of local precipitation intensity, but linked to the rainout process in the surrounding regions. Some ice core studies also suggest that records of precipitation δ^{18} O in ice cores of the Indian monsoon region do not match the local precipitation amount. For example, Pang et al. (2014) found a significant correlation between δ^{18} O records from the East Rongbuk ice cores and summer monsoon rainfall along the southern slope of the Himalayas, whereas no significant correlation was found between the δ^{18} O records and accumulation rates (an indicator of local precipitation). This suggests that summer monsoon precipitation δ^{18} O over the high Himalayas is controlled by the upstream rainout over the entire southern slope of the Himalayas rather than local precipitation processes.

Stable oxygen isotopes in speleothems are widely used for paleoclimate reconstructions. Recently, stable oxygen isotopes measured in cave speleothems from China have received much attention: e.g., Hulu Cave (Wang et al., 2001), Dongge Cave (Yuan et al., 2004; Dykoski et al., 2005; Kelly et al., 2006), Sanbao Cave (Wang et al., 2008; Cheng et al., 2009), Heshang Cave (Hu et al., 2008), Wanxiang Cave (Zhang et al., 2008), Buddha Cave (Paulsen et al., 2003), and Dayu Cave (Tan et al., 2009) (Fig. 1a). However, the interpretation of these stable isotope records in

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speleothems remains controversial. Some researchers used the stable isotope records from stalagmites in monsoonal east China as proxies for precipitation amount (Hu et al., 2008; Tan et al., 2009; Cai et al., 2010). Paulsen et al. (2003) showed that short-term (< 10 years) variations in δ^{18} O in stalagmites from Buddha Cave reflect changes in precipitation amounts, but longer-term (> 50 years) δ^{18} O variations indicate changes in air temperature. Other studies suggest that $\delta^{18}{\rm O}$ indicates changes in the ratio of summer to winter precipitation, which they refer to as "monsoon intensity" (Wang et al., 2001, 2008; Yuan et al., 2004; Dykoski et al., 2005; Kelly et al., 2006; Zhang et al., 2008; Cheng et al., 2009). Dayem et al. (2010) reported that annual and rainy season precipitation totals in each of central China, south China, and east India have correlation length scales of ~ 500 km, shorter than the distance between many speleothem records that share similar long-term time variations in δ^{18} O values. Thus, the short correlation distances do not support the idea that apparently synchronous variations in δ^{18} O values at widely spaced (> 500 km) caves in China are due to variations in annual precipitation amounts. Most of the above-mentioned studies indicate that the variations of δ^{18} O in speleothems from the Asian summer monsoon region are not controlled by the local precipitation amount.

Recent studies have revealed the importance of variability in moisture sources (Peng et al., 2010; Xie et al., 2011) and large-scale convective activities (Vimeux et al., 2011; Tremoy et al., 2012; Kurita, 2013; Moerman et al., 2013) in controlling precipitation δ^{18} O in monsoon regions. Strong convection at source regions tends to produce more precipitation, causing heavy isotopes to preferentially condense from vapor, leading to lower values of downstream precipitation δ^{18} O. In addition, the location of moisture source determines the distance that water vapor has to travel, hence affects the precipitation δ^{18} O. Soderberg et al. (2013) found that the variability of the isotopic composition of individual rain events in central Kenya could be partly explained by the distance traveled of air mass over land. Therefore, the rainout effect at the water vapor source areas and upstream regions should have a significant influence on stable isotopes in precipitation in downstream regions.

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In the Asian monsoon regions, moisture sources for summer precipitation often lie in the strong convection areas within the intertropical convergence zone (ITCZ). The variability of ITCZ position and intensity could therefore affect precipitation δ^{18} O in these regions. Using the outgoing longwave radiation (hereafter OLR) as a tracer for deep tropical convection (Wang et al., 1997), the ITCZ position and strength could be identified (Gu and Zhang, 2002). In the East Asia-West Pacific region, the onset of the Asian summer monsoon corresponds to the northward movement of the ITCZ to an area between 5 and 25° N (Ding, 2007), and brings with it large amount of convective precipitation (Ananthakrishnan et al., 1981). In this study, we examined in detail how summer precipitation δ^{18} O related to changes in the position and intensity of moisture sources within ITCZ, using the daily δ^{18} O data at Nanjing in summer (June-September) during 2012-2014, the daily OLR data, and relevant meteorological data. According to monthly long term means of Nanjing precipitation for the years 1981-2010 from China Meteorological Data Sharing Service System (http://cdc.nmic.cn/home.do), the proportion of summer monsoon precipitation (June-September) at Nanjing accounts for 54.8 % of its annual precipitation. We focused our analysis on summer because in the Asian monsoon region precipitation concentrates in summer, and hence the annual precipitation-weighted mean of the isotopic composition in precipitation is dominated by the isotopic value in summer precipitation. The results of this study could help with the interpretation of stable isotopic paleoclimate records in this region.

Study area

General atmospheric circulation

Nanjing is located on the lower reaches of the Yangtze River, surrounded by low hilly terrain with an average altitude of 26 m (Fig. 1a). The mean annual air temperature is 16°C and the average annual precipitation is 1106 mm. Located close to the Tropic

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of Cancer, this area has a strong seasonal climate (Fig. 1b), with a distinct seasonal reversal of wind and alternation of dry and rainy periods. In the winter, the air masses over Nanjing mainly originate from the high pressure system over Mongolia in the North and West (Fig. 1a). In the summer, the city is under the influence of both the East Asian summer monsoon and the Indian summer monsoon (Fig. 1a).

2.2 Intraseasonal variations in the Asian Summer Monsoon

With the onset of the summer monsoon, the warm and moist air masses from the south collide with cold northerly air masses in east China, forming a quasi-stationary rain belt known as Meiyu. The weather systems developed at the Meiyu front provide the majority of summer precipitation in this region, with the enhanced moisture transport from the South China Sea (SCS) and the Bay of Bombay (BOB) (Ding. 1992). The Meiyu system also includes the Baiu in Japan (Saito, 1995) and the Changma in Korea (Oh et al., 1997). Meiyu starts in southern China between April and May, moving to the middle part of eastern China (Yangtze and Huai He River Basins) in May and July, and to northern China in July and August, bringing with it consistent rainfall. Baiu occurs from mid-June to mid-July in Japan (Saito, 1995), and Changma takes place from the end of June to the end of July when the rain belt shifts northward to Korea (Oh et al., 1997). The retreat of the Asian summer monsoon is observed earliest in East Asia, and occurs very rapidly, taking only a month or less to retreat from northern to southern China. In early September, the leading edge of the summer monsoon quickly withdraws southward to the northern part of the SCS and remains there, marking the end of the summer monsoon in East Asia (Ding, 1992, 2007).

2.3 Moisture sources of summer precipitation

To determine the probable source regions of the air masses influencing our study area, we generated backward trajectories based on the Hybrid Single-Particle Lagrangian Integrated Trajectories (HYSPLIT) of the Air Resources Laboratory of the US National

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Oceanic and Atmospheric Administration, based on the data generated by the Global Data Assimilation System (GDAS) (ftp://arlftp.arlhq.noaa.gov/pub/archives/gdas1). Although backward trajectory analysis only reflects the synoptic situation and is only an approximation of the general origin of an air mass, this approach has been widely used in studies of moisture transport (Brimelow et al., 2005; Perry et al., 2007; Sodemann and Stohl, 2009; Drumond et al., 2011). In our study, the vapor source trajectory was simulated for each precipitation event from June to September, and cluster analysis was applied to the trajectories. The moisture transport paths were identified using the HYSPLIT back trajectory model combined with NCEP reanalysis at 12 h time steps back to 11 days at 1500 ma.g.l. (about 850 hPa), as water vapor transport is usually concentrated in the middle and lower troposphere (Bershaw et al., 2012). The total spatial variance (TSV) (Fig. 2a) was used to identify the optimum number of clusters. Rapid growth in TSV occurred when the number of clusters fell below six, therefore six clusters were retained as the final simulated cluster trajectories.

The simulation suggests that in summer, Nanjing was dominated by the influence of several major moisture sources: the Bay of Bengal (BOB), the South China Sea (SCS), the western Pacific, and the northern inland areas (Fig. 2b).

3 Sampling and isotope measurements

Using a deep open-mouthed container, precipitation samples were collected on days with precipitation greater than 0.1 mm from September 2011 to December 2014 with the exception of January–April 2013. Immediately after collection, the samples were poured into 100 mL polyethylene bottles and sealed tightly for storage in a freezer. Only precipitation isotope compositions (δ^{18} O) for samples from June to September (i.e., summer) were analyzed and discussed in this study due to the concentration of annual precipitation in summer in the study area.

The δ^{18} O of these samples were measured using a Picarro L2120-I wavelength scanned-cavity ring down spectroscopy (WS-CRDS) system with an analytical δ^{18} O

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The stable isotopic ratio was calculated as:

$$\delta = \left[\frac{R_{\text{sample}}}{R_{\text{reference}}} - 1 \right] \cdot 1000\% \tag{1}$$

where R is the ratio of the composition of the heavier to lighter isotopes in water ($^{18}\text{O}/^{16}\text{O}$ for $\delta^{18}\text{O}$, or D/H for δ D), and the reference is the Vienna Standard Mean Ocean Water standard. Each sample was measured eight times, with the first five measurements discarded in order to eliminate the effect of memory. The mean value of the last three measurements was taken as the test result.

4 Results

4.1 δ^{18} O variations in summer precipitation

Stable isotopes in our precipitation samples fluctuated between -14.8 and 2.3%, and were more depleted in summer than in the other seasons (Fig. 3a). In 2012, after a sudden decrease on 6 June, the precipitation δ^{18} O remained low, reaching a minimum (-14.8%) on 14 July. The δ^{18} O values increased in early August, and decreased again in late August. In early September, δ^{18} O in precipitation became enriched (Fig. 3b). In 2013, precipitation δ^{18} O decreased suddenly on 7 June, then increased slowly until it peaked (-4.0%) on 22 August. The stable isotope composition was depleted in late August and reached a minimum (-13.8%) on 7 September. In late September, δ^{18} O in precipitation was enriched (Fig. 3c). In 2014, δ^{18} O in precipitation decreased on 1 June and slightly increased afterward until it was depleted again in July. It started to increase in early August. From late August to early September, δ^{18} O in precipitation remained depleted, but became enriched since late September (Fig. 3d).

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We divided the summer into 5 distinct stages (Fig. 3), based on the temporal patterns δ^{18} O variations, together with the intraseasonal variations in the Asian summer monsoon and Meiyu (see Sect. 2.2). Stage 1 started with the sudden decrease in δ^{18} O in early June, which is generally considered as an indicator for the onset of the summer monsoon (e.g., Tian et al., 2001; Vuille et al., 2005; Yang et al., 2012). Stage 2 covered the Meiyu period. The start dates of Meiyu in 2012–2014 were 26 June, 23 June, and 25 June respectively, based on the observations by the Jiangsu Provincial Meteorological Bureau. Stage 3 was characterized by relatively high precipitation δ^{18} O, whereas in stage 4, δ^{18} O remained low. Stage 5 marked the return of δ^{18} O values to enriched state. The 5 stages were delineated in Fig. 3b–d.

4.2 The amount effect of δ^{18} O in precipitation

The amount effect refers to the observed negative correlation between precipitation isotopic composition and precipitation amount. The most discussed mechanism for the amount effect is that high precipitation rates increase relative humidity, hence decrease evaporation. As evaporation serves to enrich heavy isotopes, its reduction leads to more depleted precipitation isotopic signatures. Moreover, high relative humidity also inhibits re-evaporation of local water to feed back into the precipitation. As local water is usually more enriched in heavy isotopes, its diminished input also leads to more depleted precipitation isotopic composition. Here we investigated if the amount effect could be clearly observed from our data. We performed separate correlation analyses between precipitation δ^{18} O and precipitation amount, relative humidity and the evaporation ratio defined as evaporation divided by precipitation (E/P). Results are shown in Fig. 4. There was a weak negative correlation between precipitation δ^{18} O and precipitation amount in 2013 (Fig. 4b). In addition, precipitation δ^{18} O became more depleted with increased relative humidity (Fig. 4e) and decreased evaporation ratio (Fig. 4h). This seems to suggest that the amount effect was present in the 2013 data. However, no significant correlation was observed in 2012 and 2014. This implies that

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the local amount effect is not the dominant factor controlling the precipitation isotopic composition, and that other factors could contribute to its variations at our study site.

5 Discussion

To explore the possible influence of ITCZ intensity and position on δ^{18} O in summer precipitation in Nanjing, a composite analysis of OLR was performed for each stage (Fig. 5). Low OLR values correspond to cold and high clouds associated with enhanced convection, and a negative relationship is generally observed between OLR and convection intensity (Wang et al., 1997). Therefore, a composite analysis of OLR could help establish the location and intensity of deep convections associated with ITCZ, which serve as moisture sources for the monsoon precipitation in Nanjing. It was also necessary to establish the moisture transport for each stage in order to link the source regions with our study area and determine the distance, as both could potentially influence the precipitation δ^{18} O. To achieve this, we calculated the vertically integrated mean water vapor transport for each stage, using the daily NCEP/NCAR reanalysis data (Fig. 6).

In stage 1, the abrupt decrease of $\delta^{18}O$ indicated the onset of the Asian summer monsoon, with strong ITCZ convections in the BOB and the SCS (Fig. 5a, f and k). The summer monsoon brought huge amount of moisture from both regions (Fig. 6a, f and k). The isotope fractionation that occurred during the strong convection and the transport process lightened the stable isotopes in water vapor, resulting in the abrupt decrease of $\delta^{18}O$ in precipitation in Nanjing.

In stage 2, the ITCZ intensity and location in 2012 did not change significantly from stage 1 (Fig. 5b), and δ^{18} O remained low. The extreme negative δ^{18} O on 14 July was due to the continuous local rainfall from 12 to 14 July, further depleting δ^{18} O in precipitation. In 2013, the ITCZ intensity did not change much in the BOB, but decreased significantly in the SCS and the low-latitude western Pacific Ocean (Fig. 5g). Weak convection reduced the rainout effect, and hence increased δ^{18} O in precipitation.

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In 2014 the ITCZ intensity increased in the SCS and the low-latitude western Pacific Ocean, but it did not change significantly in the BOB (Fig. 5l). At this stage, as the meridional water vapor transport to the north from the SCS increased (Fig. 6b, g and I), changes in convective activity in the SCS had a stronger influence on δ^{18} O in study 5 area precipitation. Strong convection in the SCS enhanced rainout effect, resulting in depleted δ^{18} O in precipitation in Nanjing.

In stage 3, the ITCZ intensity decreased in the BOB in both 2012 and 2013, but increased in the SCS and the low-latitude western Pacific Ocean. The center of strong convection propagated northward (Fig. 5c and h). Water vapor mainly originated from the SCS and the low-latitude western Pacific Ocean (Fig. 6c and h) for this stage. The relatively shorter transport distance resulted in higher δ^{18} O values in precipitation. In 2014, the ITCZ intensity was relatively low in the SCS and the low-latitude western Pacific Ocean (Fig. 5m). The water vapor came mainly from the adjacent seas (Fig. 6m). As a result, the relatively weak convection in the area and short transport distance enriched δ^{18} O in precipitation in Nanjing.

Stage 4 covered the late monsoon season. In 2012, in addition to increased convection strength in the west Pacific Ocean, the strong convective center also moved eastward, increasing the water transport distance to Nanjing. Both of these changes acted to deplete δ^{18} O in precipitation. Moreover, the moisture transport suggested that vapor from the BOB was also transported to Nanjing (Fig. 6d). The strong convection in the BOB and its long distance from the study site contributed to further deplete δ^{18} O in precipitation. In 2013 and 2014, the ITCZ intensity in the SCS and the western Pacific was relatively weak. However, both the moisture transport from the BOB (Fig. 6i and n) and the convective activity in the BOB was strong (Fig. 5i and n). In addition, the strong convective center in the BOB moves southward in stage 4 of 2013 (Fig. 5i) resulting in longer distance from Nanjing. The combination of these factors depleted the isotopic composition of precipitation in this stage for both 2013 and 2014. The time series of δ^{18} O in precipitation showed a clear trend of decreasing δ^{18} O values during the late monsoon period, while rainfall peaked earlier in the season. The depleted

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 δ^{18} O values in late monsoon season were also observed in the other monsoon areas. Pang et al. (2006) suggested that the low δ^{18} O values were caused by the recycling of monsoon precipitation in late monsoon season. Breitenbach et al. (2010), on the other hand, argued that the BOB freshwater plume, consisted of isotopically depleted rain water and snow melt water, diluted the BOB surface water δ^{18} O pool in late monsoon season. This contributed to the depleted δ^{18} O in precipitation. Our results suggest that the depleted precipitation δ^{18} O in the late monsoon season could result from the combination of increased convective activities and transport distance due to the retreat of the ITCZ southward in the BOB.

In stage 5, the Asian summer monsoon retreated and water vapor from the inland areas with a high stable isotopic composition was transported to Nanjing (Fig. 6e, j and o), enriching the δ^{18} O in precipitation. It is worth noting that the ITCZ intensity in the SCS and the low-latitude western Pacific Ocean strengthened in stage 5 of 2013 because of the super Typhoon Usagi. However, Nanjing was no affected due to its location at the edge of the Typhoon. At the time, the moisture in Nanjing came mainly from the northern inland areas and the adjacent seas in the northeast (Fig. 6e, j and o). Therefore, the stable isotopic composition of precipitation remained enriched.

The above observations seemed to suggest a close relationship between precipitation δ^{18} O and the convective activity in the moisture source regions. In order to further explore this relationship quantitatively, we performed a time-lagged spatial correlation analysis between precipitation δ^{18} O in Nanjing and the daily OLR time series. Results are shown in Fig. 7. Several patterns emerged from this analysis. For stage 1 and 4, there was a strong positive correlation between δ^{18} O and OLR in the BOB at 13 and 14 days before the rainfall (Fig. 7a and b). This supports the conclusion of previous studies that convective processes could have integrated impacts on water vapor over several days preceding precipitation events (Tremoy et al., 2012; Gao et al., 2013). For stage 2, our analysis showed a strong positive correlation between δ^{18} O and the OLR in the SCS at 5 and 6 days preceding the rainfall (Fig. 7c and d). This confirms

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the significant influence of convective intensity in the SCS on δ^{18} O in precipitation in Nanjing at stage 2. As this stage covered the Meiyu period, this result is largely in agreement with previous studies, which indicate that moisture for Meiyu precipitation mainly comes from the SCS (Simmonds et al., 1999; Ding et al., 2007). For stage 5, a positive correlation was observed between daily δ^{18} O in precipitation and OLR in the inland areas to the north and west at 7 and 8 days before the rainfall (Fig. 7e and f), suggesting that inland vapor contributed to δ^{18} O in precipitation after the monsoon withdrew. However, no significant correlation between δ^{18} O and OLR was found for stage 3. This could partially attributed to the shift of ITCZ location northward and eastward in 2012 and 2013, reducing the vapor transport distance. This could have played a more important role in determining the δ^{18} O values in precipitation in Nanjing than convective intensity.

6 Conclusions

We analyzed daily stable isotopic composition of summer precipitation in Nanjing in 2012–2014, and related it to OLR and water vapor transport data to identify the influence of ITCZ location and intensity on the stable isotopic composition of precipitation. At the onset of the summer monsoon (stage 1), vapor to our study site was mainly transported from the BOB, where the strong convection in the source area and its relatively long distance from our study area acted to reduce δ^{18} O in precipitation in Nanjing. During the Meiyu period (stage 2), water vapor came mainly from the SCS, and changes in ITCZ intensity in the SCS led to the variability of δ^{18} O in precipitation in Nanjing. The northward propagation of the ITCZ during the mid-monsoon season (stage 3) reduced the vapor transport distance, resulting in relatively enriched δ^{18} O. During the late monsoon period (stage 4), the ITCZ retreated to the BOB. The strong convection and relatively long-distance vapor transport again led to depleted δ^{18} O values in precipitation in Nanjing. Finally, when the monsoon withdrew (stage 5), vapor from the north and west inland areas contributed to the enriched δ^{18} O.

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Our study indicates that the changes in the ITCZ location and intensity were major factors affecting the stable isotopes in summer precipitation in Nanjing. Our results suggest that the stable isotopes in precipitation could signal a shift of precipitation source regions and ITCZ over the course of monsoon season. As a result, changes in moisture sources and upstream rainout effect should be taken into account when interpreting the stable isotopic composition of speleothems in the Asian monsoon region.

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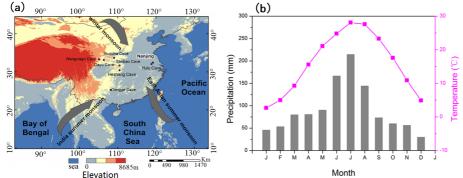


Figure 1. (a) Elevation map of China; the study site Nanjing is marked by a black star. Black dots indicate the cave locations mentioned in this study: Hulu, Dongge, Heshang, Sanbao, Wanxiang, Buddha, and Dayu. Grey arrows indicate the dominant circulation patterns over the region. (b) Monthly average temperature (T) and monthly average precipitation (P) for the years 1981–2010; data from the China Meteorological Data Sharing Service System.

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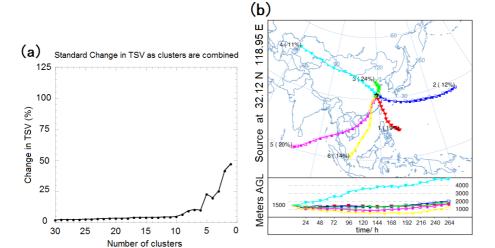


Figure 2. (a) Change in TSV (total spatial variance) as clusters are combined, and **(b)** spatial distribution of water vapor pathways. The percentage shows the frequency of each clustered backward trajectory.

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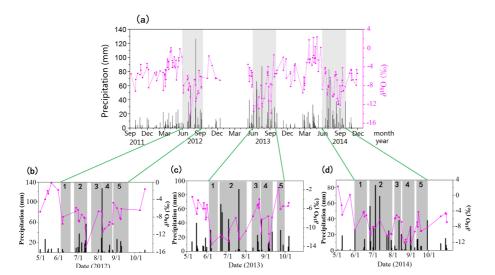


Figure 3. (a) Temporal variation of daily precipitation δ^{18} O and precipitation amount (P) from September 2011 to November 2014. The daily precipitation data were from the China Meteorological Data Sharing Service System. Note the missing data for δ^{18} O from January to April 2013, and for precipitation amount in November 2014. (**b–d**) Temporal variations in daily precipitation δ^{18} O and local precipitation amount from May to October in 2012 (**b**), 2013 (**c**), 2014 (**d**). The shaded bars represent different stages. In (**b**) for 2012, stage 1: 3–20 June; stage 2: 27 June–14 July; stage 3: 20 July–9 August; stage 4: 10–26 August; and stage 5: 27 August–20 September. In (**c**) for 2013, stage 1: 7–20 June; stage 2: 23 June–1 August; stage 3: 12–22 August; stage 4: 25 August–11 September; and stage 5: 20–30 September. In (**d**) for 2014, stage 1: 1–20 June; stage 2: 26 June–1 August; stage 3: 6–17 August; stage 4: 18 August–8 September; and stage 5: 12–30 September.

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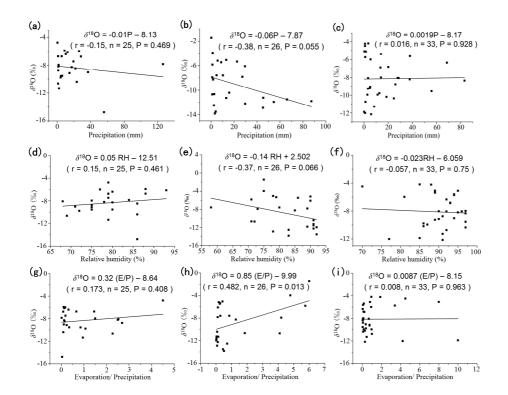


Figure 4. This figure shows: (1) correlation between δ^{18} O and precipitation amount in Nanjing from June to September for 2012 (a), 2013 (b) and 2014 (c); (2) correlation between δ^{18} O and relative humidity in Nanjing from June to September for 2012 (d), 2013 (e) and 2014 (f); (3) correlation between δ^{18} O and evaporation/precipitation in Naniing from June to September for 2012 (g), 2013 (h) and 2014 (i). Linear regression lines, correlation coefficient r and p values are also shown.

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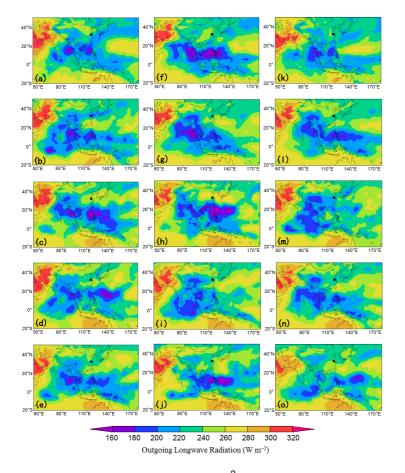


Figure 5. Composite results for average OLR (Wm⁻²) for stages 1 to 5. The convective activity is indicated by low values in OLR. Daily OLR data at 2.5° × 2.5° resolution were used (http: //www.cdc.noaa.gov). The study site of Nanjing is marked by a black dot. (a-e) Stages 1 to 5 in 2012; (f-j) stages 1 to 5 in 2013; (k-o) stages 1 to 5 in 2014.

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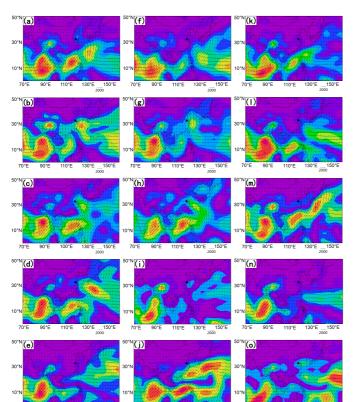


Figure 6. Vertically integrated mean water vapor transport (g cm⁻¹ s⁻¹). Different colors indicate the magnitude of the moisture flux vector. The study site of Nanjing is marked by a black dot. (a-e) standing for stages 1 to 5 in 2012; (f-j) for stages 1 to 5 in 2013; (k-o) for stages 1 to 5 in 2014.

0 600 800 1000 1200 1400 1600 1800 2000 Mean water vapor transport (g cm⁻¹ s⁻¹)

90°E 110°E 130°E __150°E

70°E 90°E 110°E 130°E __ 150°E

110°E 130°E __150°E

70°E

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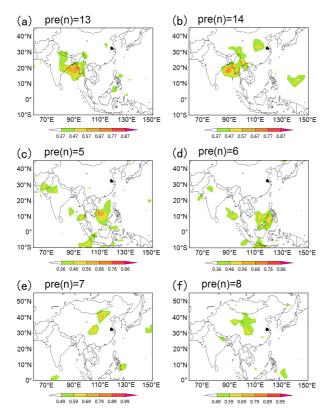


Figure 7. Spatial correlation between daily δ^{18} O in precipitation and OLR at n days prior to the rainfall date: spatial correlation between δ^{18} O in precipitation and OLR at 13 days (a) and 14 days (b) prior to the rainfall date for stages 1 and 4; spatial correlation between δ^{18} O in precipitation and OLR at 5 days (c) and 6 days (d) prior to the rainfall date for stage 2; spatial correlation between δ^{18} O in precipitation and OLR at 7 days (e) and 8 days (f) prior to the rainfall date for stage 5. For all maps, only areas with correlations significant at 0.05 level are shown. The study site Nanjing is marked with a black dot.

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