Lessons learned from the modern monsoon applied to interpretation of paleoclimate records

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Abstract

Variability in oxygen isotope ratios collected from speleothems in Chinese caves are often interpreted as proxies for variability of precipitation and related to strengthening and weakening of the southeast Asian monsoon, and, in some cases, the Indian monsoon. Using modern data to evaluate the validity of these interpretations, we find that annual and rainy season precipitation totals in each of central China, south China, and east India do not correlate well with those in the other areas. The short distances over which observed precipitation amounts correlate with one another does not support, though also cannot by itself refute, the idea that apparently synchronous variations in $\delta^{18}O$ values at widely spaced caves in China show variations in monsoon strength. We also evaluate connections between climate variables and $\delta^{18}O$ values using available instrumental measurements of $\delta^{18}O$ values in precipitation. These data, from stations in the Global Network of Isotopes in Precipitation (GNIP), show that monthly $\delta^{18}O$ values generally do not correlate well with either local precipitation or local temperature, and the degree to which monthly $\delta^{18}O$ values do correlate with precipitation or temperature varies from station to station. Cave speleothems, however, may record interannual, rather than monthly, variations in climate, but station records of $\delta^{18}O$ values in precipitation are not sufficiently long to establish whether significant correlations exist between average $\delta^{18}O$ values and either temperature or amount of precipitation. In the few locations that do show significant correlations between $\delta^{18}O$ values and precipitation amount, we estimate the change in precipitation required to account for variability in $\delta^{18}O$ values in speleothems from Hulu and Dongge caves if $\delta^{18}O$ values are controlled by the precipitation amount effect. We find that differences on the order of at least 50% in mean annual precipitation are required to explain the $\delta^{18}O$ variations on orbital time scale, which is implausibly large and inconsistent with prior GCM results. Thus, we conclude that variations in $\delta^{18}O$ values in Chinese cave speleothems primarily reflect changing source regions of the precipitation or changing pathways between the moisture source and the paleorecord site.

Keywords: monsoon; paleoclimate; oxygen ixotope ratios; Asia; precipitation; amount effect

1. Introduction

To infer climate conditions in the past from a paleoclimate record, one typically assumes that the primary signal in the record represents a single climate variable such as precipitation amount or temperature. Oxygen isotope records obtained from cave speleothems in China, for example, are usually used as a proxy for precipitation amount, or 'monsoon strength' (e.g., Cai et al., 2006; Cheng et al., 2006; Dykoski et al., 2005; Wang et al., 2008; Yuan et al., 2004). Apparent correlations between $\delta^{18}O$ records at sites on the order of 1000 km apart are used to infer that changes in the isotopic records are indicative of regional climate changes, which, in the case of the China records, is interpreted as strengthening and weakening of the East Asian monsoon.

Variability of oxygen isotope ratios (expressed as $\delta^{18}O$ values) on orbital time scales at three caves in China is ~ 5 ‰ at Hulu cave (32.5°N, 119.1°E) (Yuan et al., 2004), ~ 5 to 6‰ at Dongge cave (25.3°N, 108.1°E) (Wang et al., 2001), and ~ 4‰ at Xiaobailong cave (24.2°N, 103.3°E) (Cai et al., 2006) (Fig. 1). In comparison, modern $\delta^{18}O$ values of precipitation obtained from Global Network of Isotopes in Precipitation (GNIP) stations (IAEA/WMO, 2004) in eastern China show seasonal ranges ~ 6 to 10‰ (Fig. 1), and differences in mean-annual values of ~2 to 4‰. Because the magnitudes of variation in $\delta^{18}O$ values in the modern seasonal data and the paleorecords are similar, we use the modern data to examine spatial and temporal correlations between $\delta^{18}O$ values and two variables that are related to 'monsoon strength': precipitation and temperature.

The appeal of isotope ratios as proxies for precipitation or temperature is understandable. To reduce a paleoclimate signal to the variability of one climate variable allows a straightforward interpretation. Modern and presumably past climates, however, are not so simple; isotope ratios depend on the temperature, precipitation rate, and horizontal and vertical distance from the moisture source. They are also a function of the moisture recycling on the continents (e.g., Gat, 1996), precipitation rate and raindrop size (e.g., Lee and Fung, 2008), and the atmospheric circulation – the agent that transports moisture from source to precipitation site (e.g., Dansgaard, 1964; Lee et al., 2007; Rozanski et al., 1992). In essence, to understand the variability of an oxygen isotope ratio signal we need to know how atmospheric processes affect isotopic ratios in precipitation, and which of these processes have the largest influence on the isotope signal.

Oxygen isotope ratios recorded in cave speleothems are also affected by processes that occur between the time when the precipitation reaches the ground and when the oxygen atoms are incorporated in the speleothem. For example, retention of water in the soil may serve to average $\delta^{18}O$ values over several years (Vaks et al., 2003), and variation of cave temperature (Johnson et al., 2006a) may lead to variation in fractionation during the speleothem formation process. Vegetation and aquifer conditions may also affect speleothem formation (Fairchild et al., 2006). Thus, our ability to demonstrate that a paleoclimate record is a signal of atmospheric processes and to interpret that record depends on an understanding of the hydrology of the cave and the aquifer above it (e.g., Fairchild et al., 2006).

To our knowledge, there has been only one study of the isotopic composition of dripwater in any Chinese cave to assess the degree to which fractionation and mixing may occur on the oxygen's path from precipitation to speleothem. Johnson et al. (2006a) study monthly resolved data from a stalagmite from Heshang Cave (30.4°N, 110.4°E), which lies ~ 600 km northeast of Dongge and ~ 1000 km southwest of Hulu, two caves where long times series of $\delta^{18}O$ values have been derived from speleothems. Johnson et al. (2006a) find that both precipitation and temperature influence $\delta^{18}O$ values in cave drip water and speleothems on a monthly to yearly timescale. At the opposite extreme, the continuous sampling method used to sample the speleothem along its growth axis may average out any time resolution shorter than several years to hundreds of years (e.g., Cai et al., 2006; Cheng et al., 2006; Hu et al., 2008; Yuan et al., 2004).

In summary, many processes in both the atmosphere and the aquifer influence $\delta^{18}O$ values in cave speleothems. In a step toward understanding what such values imply for paleoclimate, we focus on two questions regarding interpretation of these records: How large an area does a paleorecord describe? and: Can we demonstrate that precipitation amount (or monsoon strength) is the primary influence on these $\delta^{18}O$ values? To answer the former, we examine the spatial extent of correlations of precipitation and temperature of cave locations with the rest of Asia. To answer the latter question, we test for correlation between $\delta^{18}O$ values and precipitation or temperature, and we estimate precipitation in the past assuming that precipitation amount is the main influence on $\delta^{18}O$ values. Although the magnitude of seasonal and interannual variation in modern data is comparable to the difference between high and low values of $\delta^{18}O$ recorded in cave speleothems, we do recognize that the time scales that we consider may be too short to allow us to examine all processes that affect $\delta^{18}O$ values, and that other tools, such as GCM experiments, may be required for paleoclimatic interpretation of such records.

2. Spatial extent of variability

The paucity of paleoclimate records and the large investments in time and money required to obtain them make it appealing to apply climatic inferences from a given record to as large an area as possible. Authors of studies of this nature typically apply interpretations to the entire East Asian monsoon region (e.g., Cheng et al., 2006; Kelly et al., 2006; Wang et al., 2008; Yuan et al., 2004), and some strive to link the East Asian monsoon to the Indian monsoon (e.g., Cai et al., 2006). In this section we establish that variability in modern data do not support this generalization, at least on interannual to decadal timescales.

Regions impacted by monsoon precipitation, such as northern India and southeast China, receive large amounts of precipitation, even in the annual mean (Fig. 2). We test whether variations in annual precipitation are similar across these broad monsoon impacted regions in China and India by correlating annual mean precipitation and temperature from the NCAR/NCEP reanalysis data set (e.g., Kalnay et al., 1996) at sites near Hulu, Dongge, and Dandak (East India) caves, where $\delta^{18}O$ records have been collected from cave speleothems (Sinha et al., 2007; Wang et al., 2001; Yuan et al., 2004), with the same variables at all other points in Asia. We carried out the same correlations using data from the ECMWF ERA-40 data set, and found patterns similar to those we describe below. Annual (January to December) mean precipitation, which eliminates the seasonal march in precipitation from south to north in eastern China, correlates positively and significantly over relatively small spatial scales (Fig. 3, left column). The spatial scale of significant correlation is ~ 500 km near Hulu cave and slightly near larger Dongge cave (Fig. 3c). Thus, a strong monsoon near one cave does not imply a strong monsoon at another. Precipitation on the east coast of India correlates with precipitation over the whole of northern India, but hardly at all with anywhere in China (Fig. 3e). The lack of significant correlation between precipitation near Hulu cave with either Dongge cave or East India, and none between the latter two sites either, suggests that the Indian and southeast Asian monsoon systems are broadly separate (e.g., Fasullo and Webster, 2003; Maher, 2008; Wang and Fan, 1999; Webster et al., 1998), and that the processes that affect monsoon variability in eastern China seem to behave differently in the north and south (e.g., Lee et al., 2008).

Mean annual temperature covaries over a broader region than does precipitation (Fig. 3, right column). The temperature near Hulu cave correlates positively and significantly with temperature along eastern China and north of the Tibetan plateau (Fig. 3b). Temperature near Dongge cave covaries with temperature in southern China, northern India, and north of the Tibetan plateau (Fig. 3d). Temperature in eastern India correlates positively with that across India and southeastern Asia (Fig. 3f). Correlations made using rainy season mean temperature show similar patterns. Thus, a paleorecord that is assumed to be a proxy for temperature may describe a larger region than one assumed to be a proxy for precipitation. If variability of $\delta^{18}O$ values at Hulu and Dongge caves were the response to variability in temperature, the positive correlation seen in the orbital time-scale variability in $\delta^{18}O$ values in these two places would be consistent with variability in modern climate.

Although we do not obtain significant correlations over a broad area for precipitation, it does not necessarily follow that that wide-spread covariance in precipitation did not occur in the past climates, which responded to different insolation forcing than modern climate. The degree of spatial covariance in precipitation might be tested using GCM simulations for past climates such as those made for paleoclimate modeling intercomparison project (PMIP) (Joussaume and Taylor, 1995), which, to our knowledge, has not been published.

3. Seasonality of modern precipitation and temperature

The seasonality of climate in India is characterized by wet summers associated with the Indian monsoon and relatively dry winters. Moisture that falls as summertime monsoon precipitation in India is transported from the Bay of Bengal by a succession of storms, known as Bengali depressions (e.g., Gadgil, 2003). These may result from differential heating between Asia and the Indian Ocean, or from movement of the intertropical convergence zone (ITCZ) over the region (Gadgil, 2003).

In contrast, seasonality of present-day precipitation in eastern China varies from south to north (Fig. 1). Precipitation rates are maximum in the spring in southeast China (see stations Guilin, Hong Kong, Liuzhou, and to a lesser degree Fuzhou in Fig. 1), but are maximum in mid- to late summer farther north (see Nanjing and Shijiazhuang, Fig. 1). This south to north progression of high precipitation rates follows the path of the Meiyu front, a warm, humid, and convective subtropical frontal system that is related to the subtropical high-pressure system over the western Pacific Ocean (Zhou et al., 2004 and references therein). The front stretches northeast to southwest over southeast China, extends as far west as ~ 105°E and as far north ~ 35° N (Zhou et al., 2004) (Fig. 1). Only two of the stations we examine lie outside the Meiyu front region: Shijiazhuang is north of the northernmost location of the front, and Kunming is west of the region affected by frontal dynamics (Fig. 1). Stations at Guiyang and Zunyi are on the western edge of the Meiyu front region, and receive maximum precipitation rates in the early summer rather than in the spring due to the NE-SW orientation of the front (Fig. 1). Winds associated with Meiyu frontal precipitation are generally from the south.

Thus the term ‘monsoon’ applied to southeast China region is somewhat misleading; the majority of the precipitation that falls in southeast China results from frontal dynamics rather than differential land-sea heating or the northward progression of the ITCZ. Coastal stations Hong Kong and Fuzhou receive high precipitation rates again in the summer after the Meiyu front has moved northward. These high precipitation rates are associated with easterly winds (not shown) and may be a result of local differential land-sea heating. We stress, however, that the majority of the precipitation in southeast China is associated with frontal dynamics and convergence of the large-scale circulation.

The difference in the seasonality of precipitation and in the atmospheric dynamics that delivers it to the continent supports the lack of correlation between either temperature or precipitation in southeast China with that of India in the modern climate (Fig. 3). Comparisons of proxies of Indian monsoon and southeast China precipitation contain the underlying assumption that in past climates, broad regions are responding to some forcing in the same way, which is not supported by the analysis above.

4. Correlation of $\delta^{18}O$ values with precipitation and temperature

4.1 Monthly correlations

We wish to test the assumption that $\delta^{18}O$ values are controlled either by the amount of precipitation or by local temperature. To do so, we use data from GNIP stations (IAEA/WMO, 2004) in eastern China (Fig. 1) to calculate correlations for monthly and for 12-month and 24-month running average values of $\delta^{18}O$, temperature, and precipitation. Although modern $\delta^{18}O$ data is limited to as few as 5 years at some stations with a maximum of 35 years at Hong Kong, we expect that if robust relationships between $\delta^{18}O$ values and precipitation or temperature exist, even these short term modern records should show systematic correlations with climate variables. Correlations on the monthly time scale contain information on present-day atmospheric variability. Correlations using one- or two-year running average data may better reflect the atmospheric variability as it is recorded in cave speleothem records, as the latter reflects a smoothed version of $\delta^{18}O$ values in precipitation due to the retention time in the soil above a cave (e.g., Johnson et al., 2006a; Vaks et al., 2003). We also report correlations between anomalies (differences between monthly values and the corresponding average monthly value) of the same variables, to remove correlations associated with the seasonal cycle. In the remainder of this section, we show that where significant correlations exist, monthly correlations between $\delta^{18}O$ values and temperature or precipitation vary from station to station and explain < 50 % of the variance in all cases. In general, temperature is better anticorrelated with $\delta^{18}O$ values than is precipitation. Correlations between 12- and 24-month running averages of the variables are generally not significant, leading us to urge caution to those who assume that $\delta^{18}O$ values in speleothems are a proxy for the amount of climatological precipitation.

On a seasonal cycle, temperature and $\delta^{18}O$ values covary (anti-phased) at most sites. Temperature is maximum in summer and $\delta^{18}O$ values are smallest in the late summer to early fall (Fig. 1). Values of $\delta^{18}O$ are generally less negative in the wintertime. Precipitation does not covary with $\delta^{18}O$ values throughout southern China, where maximum precipitation occurs in springtime, and $\delta^{18}O$ values reach a minimum in late summer (Fig. 1).

For the few stations that show a statistically significant relationship, monthly correlations between $\delta^{18}O$ values and precipitation amount are negative (Table 1). Plots of $\delta^{18}O$ values versus monthly precipitation (Fig. 4) indeed show large scatter at most sites. Correlations statistically significant from zero are found only at Guiyang, Hong Kong, Kunming, and Zunyi. Correlation coefficients between monthly anomalies of $\delta^{18}O$ values and monthly anomalies in precipitation are also negative, but are significantly different from zero only at Hong Kong.

Fig. 4 shows scatter plots of the monthly averaged values of $\delta^{18}O$ versus temperature for all stations. Where correlations are significant (see Table 1), temperature is negatively correlated with $\delta^{18}O$, except at Shijiazhuang, which lies north of the Meiyu front region and is unaffected by monsoon precipitation (Fig. 1). In contrast, monthly anomalies of $\delta^{18}O$ values and temperature are positively correlated where the correlation is significant, at Kunming and Shijiazhuang, the two stations unaffected by the Meiyu front. These differences between correlations of raw monthly data and those with the seasonal cycle removed suggest that the seasonal cycle contains much of the information in the $\delta^{18}O$ signal. Thus we suspect that the correlations between $\delta^{18}O$ values and temperature result from correlations of each variable with some other seasonally varying variable such as insolation or the large-scale atmospheric circulation. If this is the case, it need not be local temperature that determines $\delta^{18}O$ values, but instead some other independent process that affects both temperature and the value of the $\delta^{18}O$ in the precipitation. Local temperature is thus an indicator of – not the cause of – changes in processes elsewhere and the latter determine the $\delta^{18}O$ that is being precipitated over China. We also note that just as $\delta^{18}O$ values in the paleorecords decrease with summer insolation and hence presumably increasing local temperature (Cai et al., 2006; Wang et al., 2001; Yuan et al., 2004), modern $\delta^{18}O$ values decrease with increasing temperature, which is opposite the prediction from Rayleigh fractionation that $\delta^{18}O$ values increase with increasing temperature (e.g., Dansgaard, 1964). The lack of agreement between trends in modern $\delta^{18}O$ values and Rayleigh fractionation has been observed globally in both observations and model results (e.g., Brown et al., 2008; Lee et al., 2007), and suggests that if temperature influences $\delta^{18}O$ values, then Rayleigh fractionation is not the dominant process by which that occurs.

All sites in our study receive most of their precipitation in the spring and/or summer seasons, which means that monthly average temperature and precipitation are positively correlated. To test whether the lack of independence between precipitation and temperature affects the correlations above, we calculate partial correlation coefficients for the monthly mean time series, which remove the influence of either temperature or precipitation (e.g., Arkin and Colton, 1970). For example, the partial correlation $\rho(\delta^{18}O,T, P)$ is the correlation between $\delta^{18}O$ values and temperature with the effect of the correlation between temperature and precipitation removed. Where significant, partial correlation coefficients have the same sign and tend to be slightly smaller in magnitude than the correlation coefficients (Table 1), suggesting that correlation between temperature and precipitation affects correlations between $\delta^{18}O$ values and temperature or precipitation by only a small amount.

4.2 Interannual correlations

For comparison to paleoclimate records, correlations between longer time intervals may be more appropriate. Therefore we calculate correlations between 12-month and 24-month running average values of $\delta^{18}O$ with corresponding averages of precipitation and temperature. Note that in calculating 12- and 24-month averages of $\delta^{18}O$ values, we use the monthly values of $\delta^{18}O$ weighed by the amount of the precipitation that fell that month and denoted by $\delta^{18}O_w$. To assess statistical significance, we use an effective degrees of freedom n - 2 where n is the number of years of data for 12-month averages and half that number for 24-month averages. Because of the limited amount of data available, no correlations are significant at the 95% confidence level. Only the correlations between 12- and 24-month running average values of $\delta^{18}O_w$ and temperature at Hong Kong are significant at the 80 % confidence level (r = -0.30 (12-month average), r = -0.45 (24-month average)), which is suggestive at best. Thus the available instrumental observations provide evidence that neither supports nor denies the normal paleoclimate interpretation. Beyond waiting for longer timeseries of isotope measurements to become available, another approach to build confidence in the correct climatic interpretation of the speleothem record may be climate model studies that test how $\delta^{18}O$ values in precipitation respond to various atmospheric processes (e.g., Bony et al., 2008; Lee et al., 2007; Lee and Fung, 2008; Risi et al., 2008).

Although the statistics do not allow for definitive conclusions to be made, they are suggestive of correlations that are consistent with the global scale findings of Rozanski et al. (1992). They find that anomalies of 12-month average $\delta^{18}O$ values are positively correlated with temperature at mid-latitudes, but that there is little relationship between the either temperature or precipitation at Hong Kong, which is consistent with results from other tropical regions that they examined and are affected by a monsoon. The two coastal sites that receive both springtime precipitation related to the Meiyu front dynamics and summertime precipitation presumably related to typical monsoon differential heating (Hong Kong and Fuzhou) may fit with Rozanski et al.'s (1992) tropical sites, and inland sites may fit with their mid-latitude sites.

5. Discussion

Monthly correlations suggest that variations in $\delta^{18}O$ values generally correlate better with temperature than with precipitation. At all stations except Shijiazhuang, $\delta^{18}O$ values are negatively correlated with temperature, so that rainwater is isotopically lighter when temperature is higher (summer). This is the same sign sense as the orbitally induced changes in paleoclimate records: $\delta^{18}O$ values are more depleted during warmer times (e.g., Cai et al., 2006; Wang et al., 2001; Yuan et al., 2004), but opposite to that predicted by Rayleigh fractionation: $\delta^{18}O$ values increase with temperature (e.g., Dansgaard, 1964). For northern and western stations (Guiyang, Kunming, Nanjing, Shijiazhuang, and Zunyi), maximum temperature and depleted $\delta^{18}O$ values also correspond to the maximum precipitation rate (Fig. 1). Locations in southeast China such as Guilin and Liuzhou, however, receive their maximum precipitation rates in the springtime. Minimum $\delta^{18}O$ values and maximum temperature occur in the summer at the southeast China sites, and $\delta^{18}O$ is more negatively correlated to temperature than precipitation. Because correlations between monthly anomalies of $\delta^{18}O$ values and precipitation or temperature are small and, with a few exceptions, insignificant, we infer that much of the variation in $\delta^{18}O$ values results from seasonal variation of some process and does not reflect local temperature or precipitation.

Johnson and Ingram (2004) perform multiple regression analyses on 3 years of data from most of the same GNIP stations in China. They deduce that $\delta^{18}O$ values are influenced primarily by precipitation in areas that are affected by the summer monsoon, and by temperature in areas not affected by the summer monsoon. Although partial correlation coefficients (Table 1) do support their conclusion that precipitation contributes to the $\delta^{18}O$ signal in Hong Kong (which is influenced by the summer monsoon), and to some extent at Shijiazhuang, which is north of the monsoon region (Fig. 1; Table 1). At these and other sites, however, the partial correlation coefficients relating monthly values of $\delta^{18}O$ to temperature are larger in magnitude than those for precipitation (Table 1). Thus, if cave records provide information on a monthly scale, our results from the modern day climate imply that the $\delta^{18}O$ signal in cave records from central China is indicative of non-local changes (such as changes in source regions of the precipitating water, changes in the source or pathway between moisture source and the precipitating site) that on orbital time scales are due to changes in insolation which also causes temperature changes over China that are passively correlated with co-located $\delta^{18}O$ values.

Cave speleothems do not appear to record monthly variations in precipitation. Precipitated water may be retained and mixed in the soil layer for one or more years before it seeps into a cave (e.g., Vaks et al., 2003). Consequently an understanding of 12-month or 24-month running averages of modern station data would be ideal to interpret paleoclimate records from cave speleothems. Unfortunately, station records are not long enough to obtain significant results with the exception of the Hong Kong record, which suggests that $\delta^{18}O_w$ values and temperature are negatively correlated. In general, there is insufficient data available to provide support for the claim that the $\delta^{18}O_w$ record obtained from a cave stalagmite is a simple a proxy for monsoon strength.

We can, however, perform a scale analysis to explore whether it is plausible for local precipitation changes to explain the orbital time scale changes in $\delta^{18}O$ in the cave records. For the sake of argument, let us assume that $\delta^{18}O$ values are a valid proxy for precipitation, and ask: How much must annual precipitation change to produce the amplitude of $\delta^{18}O$ values in the paleorecords? The maximum amplitude of the orbital timescale swings of $\delta^{18}O$ values from Dongge cave is ~ 4 - 5 ‰ at ~ 130 ka (Yuan et al., 2004). The amplitude of variability in modern $\delta^{18}O$ values is ~ 7 – 8 ‰, and for comparison the most recent minimum $\delta^{18}O$ value is ~ -9 ‰ at ~ 9 ka and the most recent maximum $\delta^{18}O$ value is ~ -5 ‰ at ~ 15ka (Yuan et al., 2004).

We estimate weighted annual $\delta^{18}O$ values for hypothetical past climates with mean annual precipitation and seasonal amplitudes different from those day. We write monthly precipitation as the sum of the annual average plus the monthly anomaly:

$$P(t) = f_o P_o + f' P'(t) \qquad (1)$$

where f_o and f' are weights on the annual mean and amplitude of seasonal variability, respectively. For the modern day, $f_o = f' = 1$. For a climate where mean annual precipitation is larger (smaller) than present, $f_o > 1$ ($f_o < 1$). For a climate with wetter summers and drier winters (stronger monsoon) than present, $f' > 1$, and for a climate with less seasonal variability (less monsoonal) than present, $f' < 1$.

We want to test the effect of varying the annual mean with f_o and the seasonal amplitude with f', assuming that $\delta^{18}O$ values are controlled by the amount of precipitation. We determine empirical relationships between monthly precipitation and monthly average $\delta^{18}O$ values for each station using the station data. We fit $\delta^{18}O$ values as a function of precipitation (Fig. 4) with straight lines, and use those lines to define the relationship between $\delta^{18}O$ values and precipitation:

$$\begin{aligned} \delta_o &= aP_o + b \\ \delta'(t) &= aP'(t) \end{aligned} \tag{2}$$

where $a = \Delta\delta^{18}O / \Delta P$ is the slope of the best fit line and b is its y-intercept. Admittedly, this method is crude – the modern data shows so much scatter that a linear fit may not be reasonable (Fig. 4), but we proceed nonetheless and consider stations where monthly values of $\delta^{18}O$ and precipitation are significantly correlated: Guiyang, Hong Kong, Kunming, and Zunyi, noting that others have used linear regressions to estimate changes in precipitation inferred from $\delta^{18}O$ records. For example, Johnson et al. (2006b) deduce that an 80% decrease in precipitation is needed to explain a 3 ‰ reduction in $\delta^{18}O$ values in a record from Wanxiang Cave, which is north of the northern limit of the Meiyu front, and they too go on to argue that the $\delta^{18}O$ record at this cave cannot be accounted for by the amount effect or by changes in temperature.

Assuming the seasonal cycle can be described with a cosine function,

$$P'(t) = P_a \cos(t) \tag{3}$$

where P_a is the maximum precipitation anomaly, the $\delta^{18}O$ value adjusted for this approximation to the amount effect is then:

$$\delta_{ae}(t) = \frac{\int_0^{2\pi} P\delta dt}{\int_0^{2\pi} P dt}. \tag{4}$$

Using (1) and (2) to substitute for $P(t)$ and δ_o and integrating yields:

$$\delta_{ae} = af_o P_o + b + \frac{af'^2 P_a^2}{2f_o P_o} \tag{5}$$

Then the difference between a past climate state and the modern is:

$$D = aP_o\left[(f_o - 1) + \frac{1}{2}\left(\frac{f'^2}{f_o} - 1\right)\right]. \tag{6}$$

To simplify matters, we have assumed that $P_a = P_o$ in (6) (i.e., modern precipitation is P = $P_o(1 - \cos(\underline{t}))$, and require that $|f'/f_o| \leq 1$ for positive values of precipitation. Note that the difference between past and modern climates is not dependent on *b* because we have assumed that the relationship between $\delta^{18}O$ value and precipitation is invariant with time. Using modern data to find *a* and P_o for each station, we plot *D* as a function of f_o holding $f' = 1$ (Fig. 5a) and *D* as a function of f' holding $f_o = 1$ (Fig. 5b). The minimum f_o and maximum f' values are limits beyond which the absolute value of the monthly precipitation anomaly in the driest months of the year is larger than the annual mean, resulting in negative precipitation for the month. We also calculated *D* using the modern day seasonal cycle instead of a cosine function, and the resulting curves differ only slightly in shape from those plotted in Fig. 5. Thus variation in the shape of the seasonal cycle has little effect on *D*.

To decrease the $\delta^{18}O$ value by 1 ‰ (the approximate difference between $\delta^{18}O$ values of present-day and 9 ka), the mean annual precipitation must be ~1.5 times larger than present at Kunming and Zunyi, and as much as ~2 times larger than present at Hong Kong (Fig. 5a). In this formulation, changing the amplitude of the seasonal cycle cannot cause a decrease of the $\delta^{18}O$ value by 1 ‰ given the upper limit of f' (Fig. 5b). Kutzbach (1981) estimates ~ 10 % increase in summertime precipitation and ~ 5 % increase in annually averaged precipitation amount between modern day and 9 ka using general circulation model experiments. GCM ensemble results from the PMIP2 experiment show no change in annual mean precipitation from 6 ka to present (Braconnot et al., 2007). Our calculations above suggest that if the amount effect is responsible for the difference in $\delta^{18}O$ values between the two times, the change in precipitation amount must be much larger (Fig. 5a). Sustained differences of 50% or more between present-day and modern annual precipitation seem unlikely, and thus this calculation suggests that insofar as the modern dependence of $\delta^{18}O$ values on precipitation applies to paleoclimate, monsoon strength cannot be the explanation for the large variations in $\delta^{18}O$ values in speleothems in China.

Finally, the station-specific amount effect *a* calculated above is similar to that calculated by Bony et al. (2008) using a simple column-integrated model over tropical ocean and by Lee et al. (2008) using atmospheric GCM with an isotope model (Lee et al., 2007). For a relatively modest change in $\delta^{18}O$ of 2 ‰, what is the change in precipitation predicted by the models and our empirical relationship for Hong Kong (the most tropical stations analyzed here)? For Hong Kong's average precipitation rate of 196 mm/month, a 2 ‰ change in $\delta^{18}O$ values implies a change in precipitation of ~ 68 % estimated from

Bony et al.'s (2008) $\delta^{18}O$/precipitation relationship, ~ 100 % estimated from Lee et al. (2007) model, and ~ 225 % from the GNIP Hong Kong data above. Such large changes in precipitation are unlikely, which provides another argument against the assumption that precipitation amounts are the dominant control on $\delta^{18}O$ values in instrumental or paleoclimate records.

6. Conclusions

Although $\delta^{18}O$ values from cave records in China are often interpreted as records of variations in Asian monsoon strength, modern station data offer little support for the idea that variations in $\delta^{18}O$ values reflect variations in precipitation. In fact, correlations between monthly data suggest that temperature variations correlate better with variations in $\delta^{18}O$ values. Monthly $\delta^{18}O$ values correlate negatively with temperature – the same sign of the correlation between $\delta^{18}O$ values in the cave records and the amplitude of insolation ($\delta^{18}O$ values are more depleted during warmer times) – and are opposite to that expected from Rayleigh fractionation ($\delta^{18}O$ values increase with temperature). Monthly anomalies of $\delta^{18}O$ values, however, do not generally correlate with monthly anomalies of temperature, indicating that much of the covariance between $\delta^{18}O$ values and temperature is contained in the seasonal cycle. Thus we propose that variation of $\delta^{18}O$ values and temperature on the seasonal time scale primarily reflect independent processes that are both regulated by changes in insolation (e.g., local insolation directly regulates local temperature, and global insolation gradients – correlated with local insolation – affects the source regions and pathways of the $\delta^{18}O$ as it is delivered to the local site).

Cave speleothems, however, record a $\delta^{18}O$ signal of precipitation averaged over several years because precipitation must percolate through the soil before reaching the cave and being precipitated on the stalagmite. Modern station data averaged over 12 or 24 months do not show significant variations between $\delta^{18}O_w$ values and temperature or precipitation at most stations. It is possible that such correlations do exist, but that they cannot be inferred on the basis of the short duration of the observations.

Finally, if we assume that the $\delta^{18}O$ values from a paleoclimate record are a proxy for precipitation amount, then we can make a crude estimate of the difference in precipitation between present day and times in the past. Our calculations show that for a ~ 1‰ increase in $\delta^{18}O$ values, the difference between modern and 9 ka $\delta^{18}O$ values at Dongge and Hulu caves (e.g., Wang et al., 2001; Yuan et al., 2004), annual precipitation 9 ka would be at least 1.5 times that of today. Calculations using atmospheric general circulation models estimate this difference to be much smaller, around 10%. Thus paleoclimate records may not be a proxy for one single variable, but a combination of atmospheric processes.

Though not a perfect analogue for the past, the modern instrumental record and the patterns of modern climate variability offer insights on the causes of past changes in environmental conditions. In the context of interpreting the records of Asian speleothems, the modern record does not support the conventional paleoclimatic interpretation that $\delta^{18}O$ values reflect precipitation amount. In fact, correlations between monthly data suggest that temperature variations correlate better to variations $\delta^{18}O$ values, although neither precipitation nor temperature explains a majority of the variance. In light of these results, we suspect that other processes, such as re-evaporation, storm type, and variations in atmospheric circulation have more of an influence on $\delta^{18}O$ values than do precipitation or temperature.

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References

Arkin, H., Colton, R.R., 1970. Statistical Methods, Harper & Row, New York.

Bony, S., Risi, C., Vimeux, F., 2008. Influence of convective processes on the isotopic composition ($\delta^{18}O$ and δD) of precipitation and water vapor in the tropics: 1. Radiative-convective equilibrium and the Tropical Ocean-Global Atmosphere-Coupled Ocean-Atmosphere Response Experiment (TOGA-COARE) simulations, J. Geophys. Res., 113, 10.1029/2008JD009942.

Braconnot, P., et al., 2007. Results of PMIP2 coupled simulations of the Mid-Holocene and Last Glacial Maximum – Part 1: experiments and large-scale features, Climate of the Past, 3, 261-277.

Brown, D., Worden, J., Noone, D., 2008. Comparison of atmospheric hydrology over convective continental regions using water vapor isotope measurements from space, J. Geophys. Res., 113, doi:10.1029/2007JD009676.

Cai, Y., An, Z., Cheng, H., Edwards, R.L., Kelly, M.J., Liu, W., Wang, X., Shen, C.-C., 2006. High-resolution absolute-dated Indian Monsoon record between 53 and 36 ka from Xiaobailong Cave, southwestern China, Geology, 34, 621-624.

Cheng, H., Edwards, R.L., Wang, Y.J., Kong, X.G., Ming, Y.F., Kelly, M.J., Wang, X.F., Gallup, C.D., Liu, W.G, 2006. A penultimate glacial monsoon record from Hulu Cave and two-phase glacial terminations, Geology, 34, 217-220.

Dansgaard, W., 1964. Stable isotopes in precipitation, Tellus, 16, 436-486.

Dykoski, C.A., Edwards, R.L., Cheng, H., Yuan, D., Cai, Y., Zhang, M., Lin, Y., Qing, J., An, Z., Revenaugh, J., 2005. A high-resolution, absolute-dated Holocene and deglacial Asian monsoon record from Dongge Cave, China, Earth Planet. Sci. Lett, 233, 71-86.

Fairchild, I.J., Smith, C.L., Baker, A., Fuller, L., Spötl, C., Mattey, D., McDermott, F., E.I.M.F, 2006. Modification and preservation of environmental signals in speleothems, Earth Sci. Rev., 75, 105-153.

Fasullo, J., Webster, P.J., 2003. A hydrological definition of Indian Monsoon onset and withdrawal, J. Climate, 16, 3200-3211.

Gadgil, S., 2003. The Indian monsoon and its variability, Annu. Rev. Planet. Sci., 31, 429-467.

Gat, J.R., 1996. Oxygen and hydrogen isotopes in the hydrologic cycle, Annu. Rev. Earth Planet. Sci., 24, 225-262.

Hu, C., Henderson, G.M., Huang, J., Xie, S., Sun, Y., Johnson, K.R., 2008. Quantification of Holocene Asian monsoon rainfall from spatially separated cave records, Earth Planet. Sci. Lett., 266, 221-232.

IAEA/WMO, 2004. Global Network of Isotopes in Precipitation, The GNIP Database. Accessible at: http://isohis.iaea.org

Johnson, K.R., Ingram, B.L., 2004. Spatial and temporal variability in the stable isotope systematics of modern precipitation in China: implications for paleoclimate reconstructions, Earth Planet. Sci. Lett., 220, 365-377.

Johnson, K.R., Hu, C., Belshaw, N.S., Henderson, G.M., 2006a. Seasonal trace-element and stable-isotope variations in a Chinese speleothem: The potential for high-resolution paleomonsoon reconstruction, Earth Planet. Sci. Lett., 244, 221-232.

Johnson, K.R., Ingram, B.L., Sharp, W.D., Zhang, P.Z., 2006b. East Asian summer monsoon variability during Marine Isotope Stage 5 based on speleothem $\delta^{18}O$ records from Wanxiang Cave, central China, Palaeogeogr. Palaeoclimatol. Palaeoecol., 236, 2-19.

Joussaume, S., Taylor, K.E., 1995. Status of the Paleoclimate Modeling Intercomparison Project (PMIP), in Proceedings of the first international AMIP scientific conference, WCRP Report, 425-430.

Kalnay, E., et al., 1996. The NCEP/NCAR 40-Year Reanalysis Project, Bull. Am. Meteorol. Soc., 77, 437–472.

Kelly, M. J., Edwards, R.L., Cheng, H., Yuan, D., Cai, Y., Zhang, M., Lin, Y., An, Z., 2006. High resolution characterization of the Asian Monsoon between 146,000 and 99,000 years B.P. from Dongge Cave and global correlation of events surrounding Termination II, Palaeogeogr. Palaeoclimatol. Palaeoecol., 236, 20-38.

Kutzbach, J.E., 1981. Monsoon climate of the early Holocene: Climate experiment with the earth’s orbital parameters for 9000 years ago, Science, 214, 59-61.

Lee, E., Chase, T.N., Rajagopalan, B., 2008. Seasonal forecasting of East Asian summer monsoon based on oceanic heat sources, Int. J. Climatol., 28, 667-678.

Lee , J.-E., Fung, I., DePaolo, D.J., Henning, C.C., 2007. Analysis of the global distribution of water isotopes using the NCAR atmospheric general circulation model, J. Geophys. Res, 112, doi:10.1029/2006JD007657.

Lee, J.-E., Fung, I., 2008. “Amount effect” of water isotopes and quantitative analysis of post-condensation processes, Hydrol. Process., 22, 1-8.

Legates, D.R., Willmott, C.J., 1990. Mean seasonal and spatial variability in gauge-corrected, global precipitation, Int. J. Climatol., 10, 111-127.

Maher, B., 2008. Holocene variability of the East Asian summer monsoon from Chinese cave records: a re-assessment, The Holocene, 18, 861-866.

Risi, C., Bony, S., Vimeux, F., 2008. Influence of convective processes on the isotopic composition ($\delta^{18}O$ and δD) of precipitation and water vapor in the tropics: 2. Physical interpretation of the amount effect, J. Geophys. Res, 113, doi:10.1029/2008JD009943.

Rozanski, K., Araguás- Araguás, L., Gonfiantini, R., 1992. Relation between long-term trends of oxygen-18 isotope composition of precipitation and climate, Science, 258, 981-985.

Sinha, A., Cannariato, K.G., Stott, L.D., Cheng, H., Edwards, R.L., Yadava, M.G., Ramesh, R., Singh, I.B., 2007. A 900-year (600 to 1500 A.D.) record of the Indian summer monsoon precipitation from the core monsoon zone of India, Geophys. Res. Lett., 34, doi:10.1029/2007GL030431.

Vaks, A., Bar-Matthews, M., Ayalon, A., Schilman, B., Gilmour, M., Hawkesworth, C.J., Frumkin, A., Kaufman, A., Matthews, A., 2003. Paleoclimate reconstruction based on the timing of speleothem growth and oxygen and carbon isotope composition in a cave located in the rain shadow in Israel, Quaternary Research, 59, 182-193.

Wang, B., Fan, Z., 1999. Choice of south Asian summer monsoon indices, Bull. Amer. Metero. Soc., 80, 629-638.

Webster, P. J., Magaña, V.O., Palmer, T. N., Shukla, J., Tomas, R.A., Yanai, M., Yasunari, T., 1998. Monsoons: Processes, predictability, and the prospects for prediction, J. Geophys. Res., 103, 14,451-14,510.

Wang, Y. J., Cheng, H., Edwards, R.L., An, Z.S., Wu, J.Y., Shen, C.-C., Dorale, J. A., 2001. A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China, Science, 294, 2345-2348.

Wang, Y.J., Cheng, H., Edwards, R.L., Kong, X.G., Shao, X.H., Chen, S.T., Wu, J.Y., Jiang, X.Y., Wang, X.F., An, Z.S., 2008. Millennial- and orbital-scale changes in the East Asian monsoon over the past 224,000 years, Nature, 451, 1090-1093.

Yuan, D.X., Cheng, H., Edwards, R.L., Dykoski, C.A., Kelly, M.J., Zhang, M.L., Qing, J.M., Lin, Y.S., Wang, Y.J., Wu, J.Y., Dorale, J.A., An, Z.S., Cai, Y.J., 2004. Timing, duration and transition of the last interglacial Asian Monsoon, Science, 304, 575-578.

Zhou, Y, Gao, S., Shen, S.S.P., 2004. A diagnostic study of formation and structures of the Meiyu Front System over East Asia, J. Meteor. Soc. Japan, 82, 1565-1576.

Table caption

Table 1: Correlation coefficients r and partial correlation coefficients ρ calculated from nine GNIP stations for $\delta^{18}O$ values and precipitation P and temperature T. Sample size is n. Coefficients that are significant at the 95% confidence interval are printed in bold italics. A reduced degrees of freedom of $(n/3)-2$ is used in the monthly correlations to account for autocorrelation in the records, which is significant for one or two month lags.

Figure captions

Figure 1: Elevation map of China and surrounding areas with locations of GNIP stations used in this study. Dongge, Hulu, Heshang, and Xiaobailong cave locations are marked with black dots. Insets show seasonal cycle of temperature (red lines, units of °C, left axis), precipitation (blue lines, units of cm/month, left axis), and $\delta^{18}O$ values (black lines, units of ‰, right axis). Dashed line indicates approximate northern limit of Meiyu front (Zhou et al., 2004).

Figure 2: Annual mean (1920-1980) precipitation rate (mm/day) over southeast Asia from the Legates Surface and Ship Observation of Precipitation dataset (Legates and Willmott, 1990). Black contour lines denote elevations of 0 m and 2000 m. Note high precipitation rates along the Himalayan front and in southeast China resulting from Indian and East Asian monsoon activity.

Figure 3: Spatial correlation of annual mean precipitation (left) and temperature (right) between a given site (top: Hulu cave, middle: Dongge cave, bottom: East India) and the rest of Asia. Correlation coefficient is shown in filled contours, and correlations significant at a 95% confidence interval are outlined by the black contour. Precipitation and temperature from NCAR/NCEP Reanalysis (Kalnay et al. 1996).

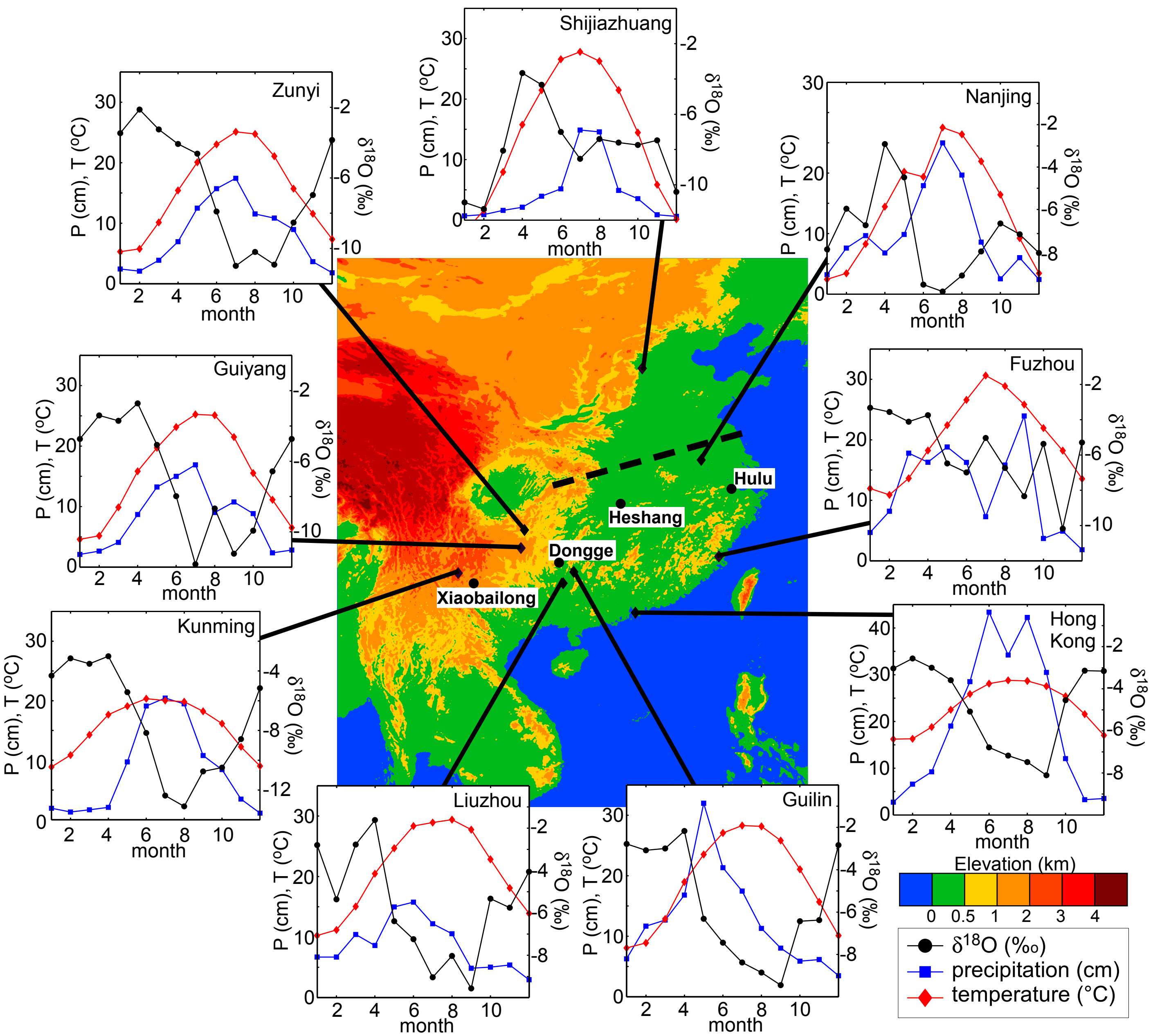
Figure 4: Monthly total precipitation (mm, squares) and monthly mean temperature (°C, diamonds) versus monthly mean $\delta^{18}O$ values (‰) for stations at (a) Fuzhou, (b) Guilin, (c) Guiyang, (d) Hong Kong, (e) Kunming, (f) Liuzhou, (g) Nanjing, (h) Shijiazhuang, and (i) Zunyi. Linear regressions used in the calculations in the Discussion are shown in black lines for stations at Guiyang, Hong Kong, Kunming, and Zunyi.

Figure 5: Annual average weighted $\delta^{18}O$ value relative to modern, D, as a function of (a) f_o for $f' = 1$ and (b) f' for $f_o = 1$ calculated using equation (6) relative to modern values. Dotted line (marked 'G') is calculation for station at Guiyang, dashed line (marked 'HK') is Hong Kong, dot-dashed line (marked 'K') is Kunming, and solid line (marked 'Z') is Zunyi. Values for a and P_o in equation (2) for the stations shown are: a = -0.025, P_o = 80.4 (Guiyang); a = -0.0081, P_o = 196.1 (Hong Kong); a = -0.03, P_o = 83.0 (Kunming); and a = -0.030, P_o = 81.5 (Zunyi). Units of a and P_o are ‰/mm/month and mm/month, respectively.

Table 1: Correlation coefficients

station	monthly average			monthly anomaly			monthly average partial correlations		
	$r(\delta^{18}O,P)$	$r(\delta^{18}O,T)$	n	$r(\delta^{18}O,P)$	$r(\delta^{18}O,T)$	n	$\rho(\delta^{18}O,P,T)$	$\rho(\delta^{18}O,T,P)$	n
Fuzhou	-0.35	-0.38	71	-0.36	-0.08	71	-0.29	-0.33	71
Guilin	-0.20	***-0.72***	92	-0.18	0.00	92	0.09	***-0.71***	92
Guiyang	***-0.48***	***-0.57***	58	-0.31	0.11	58	-0.22	-0.40	58
Hong Kong	***-0.61***	***-0.67***	276	***-0.33***	-0.03	276	***-0.36***	***-0.47***	276
Kunming	***-0.61***	***-0.44***	152	-0.09	***0.23***	152	***-0.48***	-0.11	152
Liuzhou	-0.37	***-0.55***	45	-0.42	0.37	45	-0.27	-0.50	45
Nanjing	-0.45	-0.25	58	0.06	-0.07	58	-0.39	0.02	58
Shijiazhuang	-0.09	***0.38***	146	-0.20	***0.30***	146	***-0.35***	***0.49***	146
Zunyi	***-0.56***	***-0.65***	70	-0.34	0.07	70	-0.25	***-0.44***	70

Correlation coefficients that are significant at the 95% confidence level are shown in bold italics
degrees of freedom = (n/3) - 2



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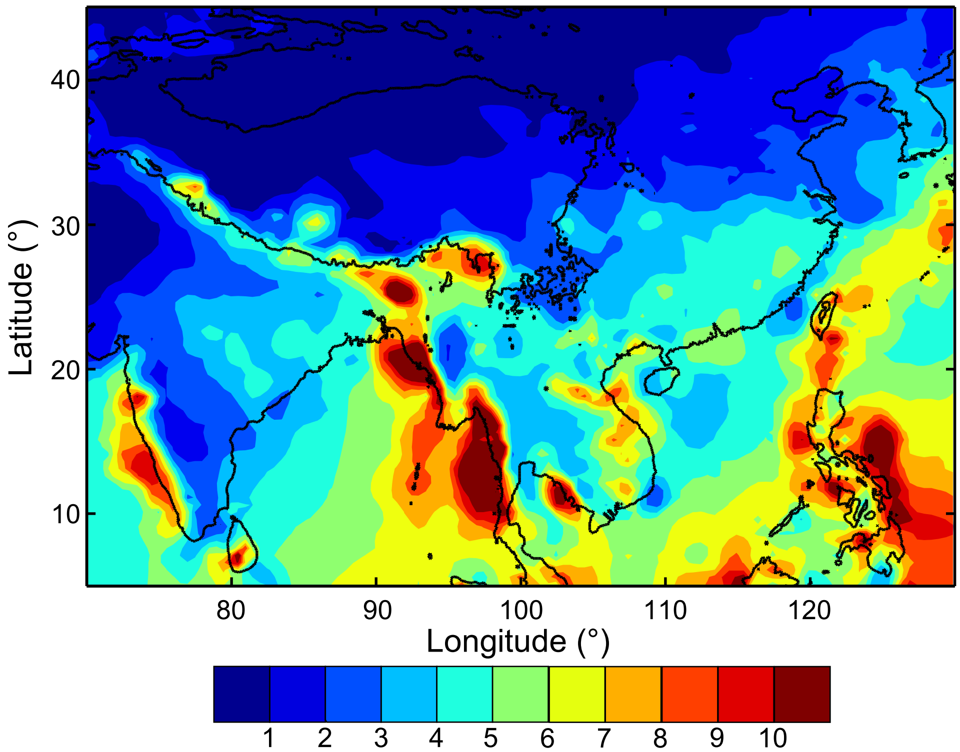

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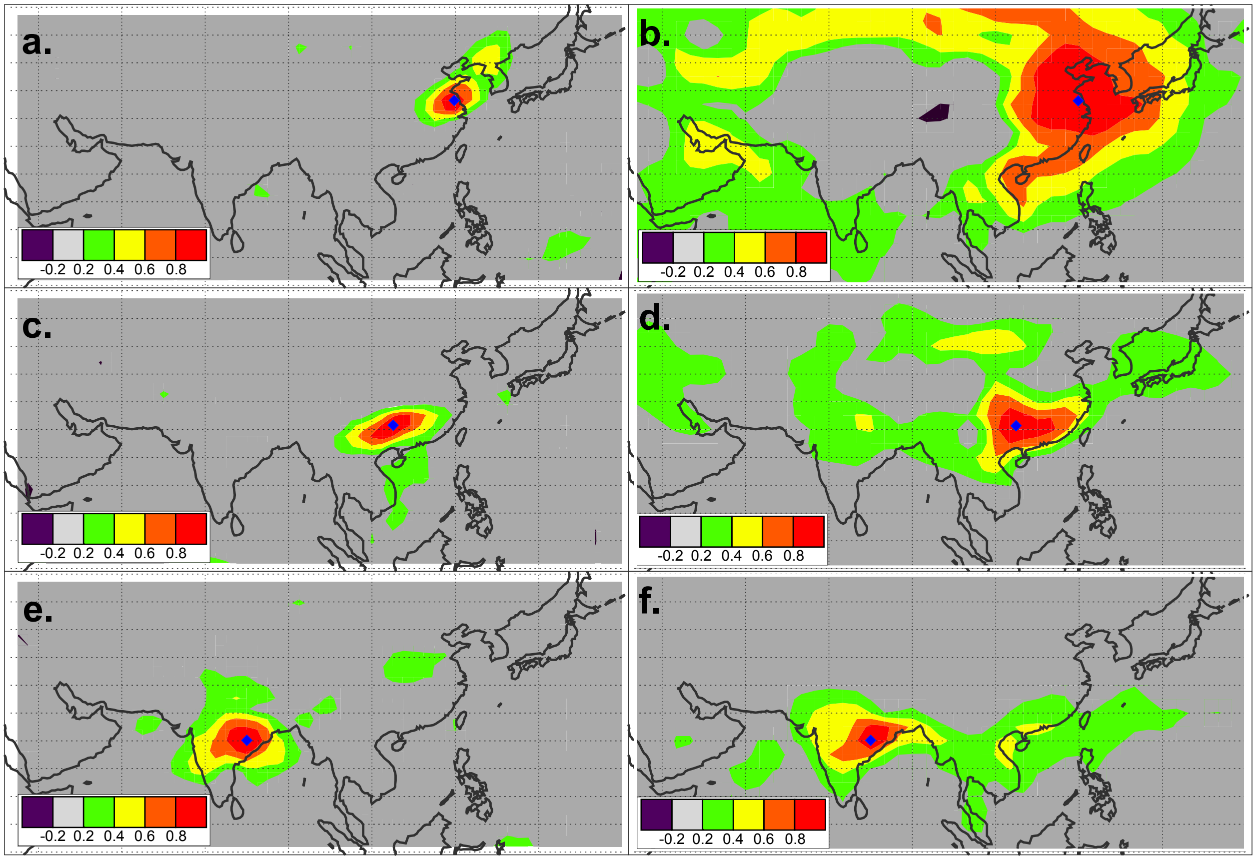

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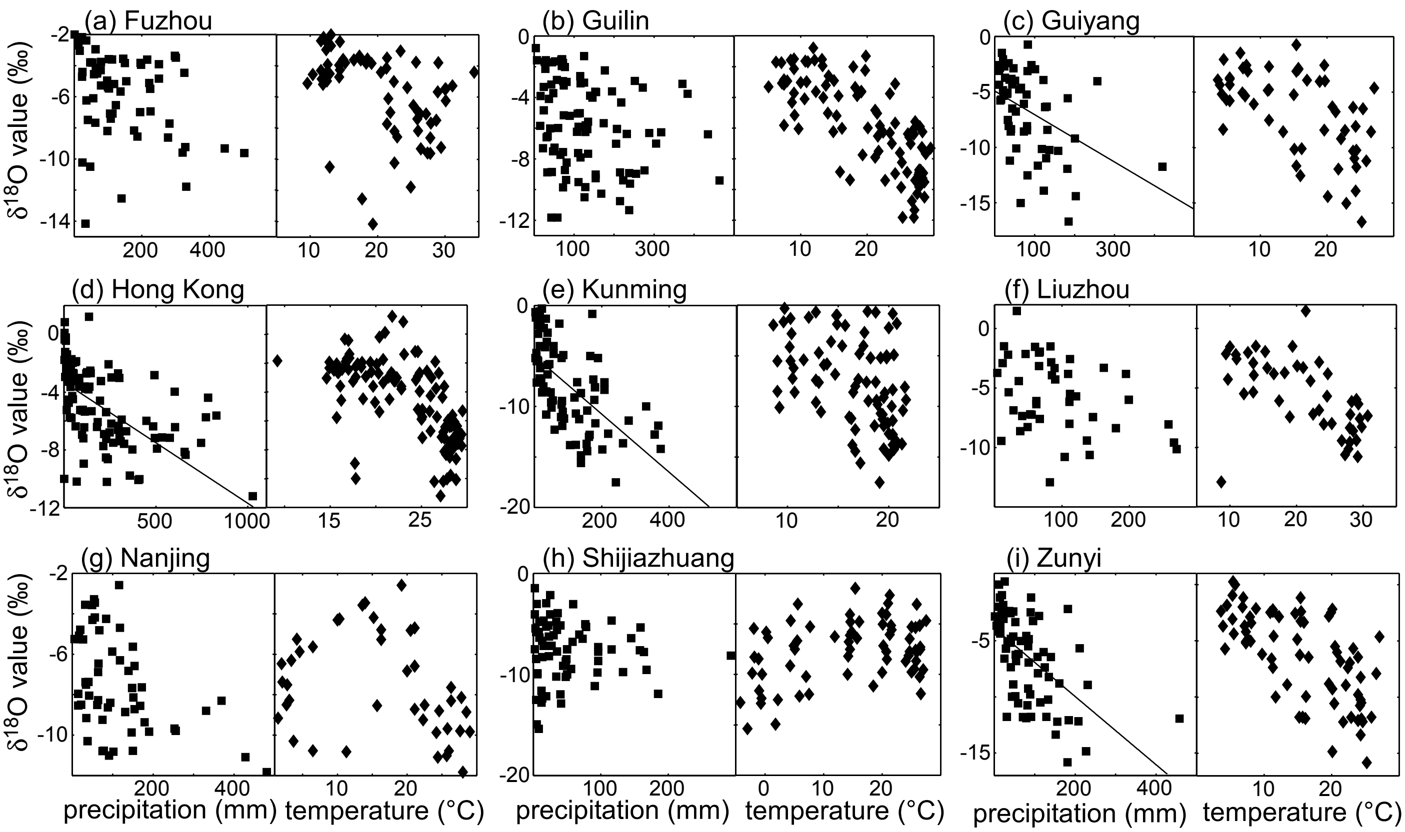

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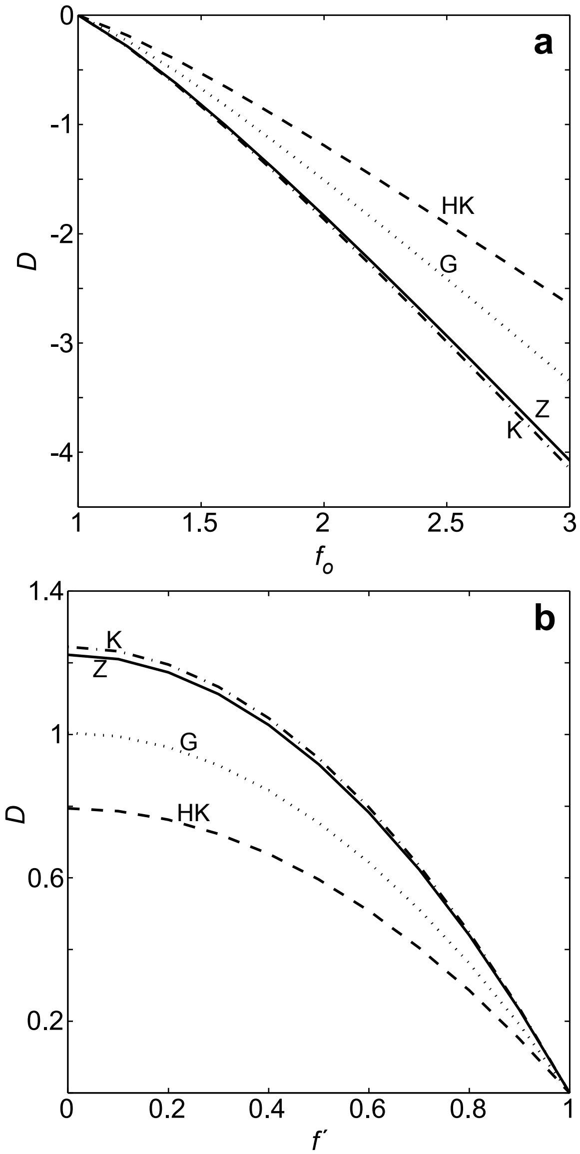

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