Lessons learned from oxygen isotopes in modern precipitation applied to interpretation of speleothem records of paleoclimate from eastern Asia

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Abstract

Variability in oxygen isotope ratios collected from speleothems in Chinese caves is often interpreted as a proxy for variability of precipitation, summer precipitation, seasonality of precipitation, and/or the proportion of ^{18}O to ^{16}O of annual total rainfall that is related to a strengthening or weakening of the East Asian monsoon and, in some cases, to the Indian monsoon. We use modern reanalysis and station data to test whether precipitation and temperature variability over China can be related to changes in climate in these distant locales. We find 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 $\delta^{18}O$ values. Thus the short distances of correlation do not support, though by themselves cannot refute, the idea that apparently synchronous variations in $\delta^{18}O$ values at widely spaced (>500 km) caves in China are due to variations in annual

precipitation amounts. 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 amount or local temperature, and the degree to which monthly $\delta^{18}O$ values do correlate with them varies from station to station. For the few locations that do show significant correlations between $\delta^{18}O$ values and precipitation amount, we estimate the differences in precipitation amount that would be required to account for peak-to-peak differences in $\delta^{18}O$ values in the speleothems from Hulu and Dongge caves, assuming that $\delta^{18}O$ scales with the monthly amount of precipitation or with seasonal differences in precipitation. Insofar as the present-day relationship between $\delta^{18}O$ values and monthly precipitation amounts can be applied to past conditions, differences of at least 50% in mean annual precipitation would be required to explain the $\delta^{18}O$ variations on orbital time scales, which are implausibly large and inconsistent with published GCM results. Similarly, plausible amplitudes of seasonal cycles in amounts or in seasonal variations in $\delta^{18}O$ values can account for less than half of the 4-5‰ difference between glacial and interglacial $\delta^{18}O$ values from speleothems in China. If seasonal cycles in precipitation account for the amplitudes of $\delta^{18}O$ values on paleoclimate timescales, they might do so by extending or contracting the durations of seasons (a frequency modulation of the annual cycle), but not by simply varying the amplitudes of the monthly rainfall amounts or monthly average $\delta^{18}O$ values (amplitude modulation). Allowing that several processes can affect seasonal variability in isotopic content, we explore the possibility that one or more of the following processes contribute to variations in $\delta^{18}O$ values in Chinese cave speleothems: different source regions of the precipitation, which bring different values of $\delta^{18}O$ in vapor; different pathways between the moisture source and the paleorecord site along which exchange of ^{18}O between vapor, surface water, and condensate might differ; a different mix of processes involving condensation and evaporation within the atmosphere; or different types of precipitation. Each may account for part of the range of $\delta^{18}O$ values revealed by speleothems, and each might contribute to seasonal differences between past and present that do not scale with monthly or even seasonal precipitation amounts.

Keywords: monsoon; paleoclimate; oxygen isotope ratios; Asia; precipitation

1. Introduction

Oxygen isotopes measured in cave speleothems from China show systematic variations that are related to orbitally paced variations in insolation (e.g., Wang et al., 2001; Yuan et al, 2004; Zhang et al, 2008). Variability of the ratio of ^{18}O to ^{16}O in calcite (measured as $\delta^{18}O$ values) on orbital time scales at four 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), ~ 4‰ at Xiaobailong cave (24.2°N, 103.3°E) (Cai et al., 2006), and ~ 3‰ at Heshang cave, a shorter record covering only the past ~9500 years (Hu et al., 2008) (Fig. 1). Such variability in $\delta^{18}O$ values almost surely reflects differences in some aspect of precipitation in China over the same time scale. Logical arguments can be made that the isotopic composition of precipitation should depend on some of the following: the amount of local precipitation that occurs on timescales as short as individual rainstorms to as long as years, on temperature (possibly on similar timescales), on the source of water vapor and changes in its temperature, and on the path followed by the vapor including precipitation and evaporation along it. Essential to the interpretation of paleoclimate records is an understanding of which of the factors listed above are responsible for the $\delta^{18}O$ signals recorded in stalagmites. Our goal here it to improve that understanding.

Paleoclimate records collected from caves in the subtropics in Israel (e.g. Bar-Matthews et al., 2000, 2003), Oman (e.g., Burns et al., 2000; Fleitmann et al., 2003, 2004), India (Sinha et al. 2005, 2007), South America (Cruz et al., 2009), and Borneo (Partin et al., 2007) have been have been interpreted as proxies for local precipitation amount, and some assume the same ("amount of summer monsoon precipitation") for

China (e.g., Cai et al., 2010; Zhou et al., 2007). Others argue that isotopic variability does not imply differences in precipitation amount; rather it indicates changes in the ratio of summer to winter precipitation, which they refer to as 'monsoon intensity' (e.g., Cai et al., 2006; Cheng et al., 2006, 2009; Dykoski et al., 2005; Kelly et al., 2006; Wang et al., 2008; Yuan et al., 2004). Their logic is that $\delta^{18}O$ values in modern spring rainfall are less negative than those in modern summer rainfall. In the annual mean, more summer rainfall should lead to more negative annual weighted $\delta^{18}O$ values. Thus in interpreting a $\delta^{18}O$ record as a proxy for summer monsoon intensity, these authors implicitly assume that the same seasonal moisture sources and transport pathways have prevailed in the past, but their relative contributions to the annual average $\delta^{18}O$ values have varied.

Johnson and Ingram (2004) examine the relationship between $\delta^{18}O$ values measured in precipitation and the *in situ* temperature and precipitation. They regress $\delta^{18}O$ values against the annual cycle in local temperature and precipitation using data from three continuous years at 10 stations over China. They conclude that, to the extent the processes controlling $\delta^{18}O$ values in the modern climate are relevant to those in past climates, the $\delta^{18}O$ variations in the caves should be interpreted as a proxy record of a combination of temperature and precipitation. More recently, extending these results, Johnson et al. (2006b) argue that changes in monsoon intensity could contribute significantly to the orbital scale variations in the speleothem $\delta^{18}O$ values, but they go on to conclude that, most likely, the dominant process contributing to the orbital scale variations in the $\delta^{18}O$ in the cave records is changes in the pathway and processing of moisture from the evaporation source to the cave sites.

Currently, there is a widespread belief that the oxygen isotopes measure some aspect of the strength or intensity of the monsoon, but apparent differences in usage of words such as strength and intensity has complicated interpretations of such isotopic data in terms of variations in climate. As noted above, many equate monsoon intensity to the ratio of summer to winter rainfall amount that is local to the cave site, but the use of adjectives like "strong" or "weak" to describe paleo-monsoons, coupled with the explicit association of large negative $\delta^{18}O$ values with local summertime precipitation has led to some confusion within the paleoclimate community of how $\delta^{18}O$ values relate to past climate and what atmospheric feature(s) is implied when the term "monsoon" is used. In a recent example, Cheng et al. (2009, p. 249) define explicitly what they mean by "monsoon intensity," but in commenting on the paper, Severinghaus (2009) seems to ignore that definition and summarizes the work of Cheng et al. as "a record of past monsoon strength." We prefer to frame our analysis in terms of rainy and dry seasons, and in the discussion below comparing our modern climate analysis with paleoclimate record interpretation, we will use the term monsoon to refer to the seasonal *circulation* in China.

In eastern China, high-resolution records of oxygen isotopes from cave speleothems provide climate data back to 224 kyr (e.g., Wang et al., 2008). Speleothems from cave stalagmites record $\delta^{18}O$ values that are determined by the $\delta^{18}O$ value of the precipitation and by any fractionation that may occur in the aquifer and during calcite precipitation in the cave (Fairchild et al., 2006; Hendy, 1971; Johnson et al., 2006a; Vaks et al., 2003). We focus only on atmospheric process here, but call attention to detailed studies of the isotopic composition of modern dripwater in Chinese caves to assess the degree to which fractionation and mixing may occur on the oxygen's path from precipitation to speleothem, such as that of Johnson et al. (2006a).

In this study we present further analysis of the modern climate and isotopic precipitation data that supports the hypothesis that the orbital scale variability (as well as the stadial-interstadial differences) in cave $\delta^{18}O$ values in China is most likely due to a combination of processes that include differences in the $\delta^{18}O$ values in the source waters and in the pathways and processing of moisture transport en route to the cave site and to local differences in convective processes (and hence fractionation) but not in precipitation amount. We focus on two questions. (1) What is the spatial extent of covariability of temperature or precipitation? This pertains to the question: What is the spatial scale of a climate anomaly that would be captured by the proxy stalagmite $\delta^{18}O$ values? (2) Can we better quantify the influence of local temperature and precipitation amount on $\delta^{18}O$ values in precipitation in the modern climate? This provides a step toward answering: what could $\delta^{18}O$ values at a site represent in climates of the past? To answer (1), we examine the spatial extent of correlations of precipitation and temperature of cave locations with the rest of Asia.

Following Johnson and Ingram (2004), we answer question (2) by correlating $\delta^{18}O$ values with local precipitation and temperature. We then estimate precipitation in the past assuming that the main influence on $\delta^{18}O$ values on paleoclimate time scales arises from variations in precipitation amount, which we infer using correlations with modern monthly $\delta^{18}O$ values, but without necessarily ascribing such correlations to the "amount effect." Our analysis differs from Johnson and Ingram (2004) in that we use longer data sets, calculate correlations using both monthly, annual cycle, and annually averaged data, and we use more stations in the region impacted by the Meiyu front: the region commonly associated with the East Asian monsoon. If significant relationships between $\delta^{18}O$ values and precipitation or temperature do not exist in present-day data, relationships between those variables in the lower frequency paleorecords may still exist, but, if so, they suggest that the fundamental processes that are responsible for variability in the present-day climate are different from those in the distant past. Conversely, modern variability offers tests of the hypothesized explanations for variability in the cave $\delta^{18}O$ signals that, if they pass, can give support for such explanations.

Our approach is undoubtedly simplistic. Modern isotope ratios may depend not only on temperature, precipitation rate, and horizontal and vertical distance from the moisture source, but also on the moisture recycling on the continents (e.g., Gat, 1996), precipitation rate and raindrop size (e.g., Lee and Fung, 2008), and atmospheric circulation – the agent that transports moisture from source to precipitation site (e.g., Cobb et al., 2007; Dansgaard, 1964; Johnson et al., 2006b; Kelly et al., 2006; Lee et al., 2007; Rozanski et al., 1992; Wang et al., 2001). 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. Precipitation and temperature observations are easy to obtain and hence our first test is for covariability of these variables with $\delta^{18}O$ values. The lack of a significant relationship would indicate that the dominant control on $\delta^{18}O$ values is another process, or that no single dominant process, or simple set of processes, exists.

2. Spatial extent of modern climate variability

Tropical and mid-latitude regions of Asia, 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 coherent across broad regions in China and India by correlating annual mean precipitation and temperature at sites near Hulu, Dongge, and Dandak (East India) caves, where $\delta^{18}O$ records have been collected from cave speleothems (e.g., Sinha et al., 2007; Wang et al., 2001; Yuan et al., 2004), with precipitation and temperature at all other points in Asia. Temperature and precipitation are from the NCAR/NCEP reanalysis data set (e.g., Kalnay et al., 1996). We carried out the same analysis using data from the ECMWF ERA-40 data set (Uppala et al., 2005) and obtained similar results to those we describe below.

The annually averaged (January to December) precipitation, which eliminates the seasonal march in precipitation from south to north in eastern China, correlates positively and significantly over only relatively small spatial scales (Fig. 3, left column). The spatial scale of significant correlation is ~ 500 km near Hulu cave and slightly larger near Dongge cave (Fig. 3c). Thus, in modern climate, a wet year near one cave does not imply the same at the other cave. 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 that near Dongge cave or East India, as well as none between the latter two sites, suggests that processes that bring moisture to the Indian and southeast Asian monsoon regions are broadly separate (e.g., Fasullo and Webster, 2003; Wang and Fan, 1999; Webster et al., 1998), and that the processes that affect variability of precipitation in eastern China seem to behave differently in its northern and southern parts (e.g., Lee et al., 2008).

The annually averaged temperature covaries over a larger 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 averaged temperature show similar patterns. Thus based only on the modern record, one might expect that a local temperature record reflects variability over a larger region than does a local precipitation record.

Our analysis suggests that precipitation anomalies are not correlated over an area large enough to account for the high correlation in the cave $\delta^{18}O$ records on orbital time scales to precipitation through a local amount effect. Hence, for the coherence in the $\delta^{18}O$ values in cave records to reflect changes in local precipitation on orbital time scales, the response of the climate system to orbitally induced variation in insolation must be different from the processes responsible for the natural variability in seasonal and annual precipitation in the modern climate.

3. Seasonality of modern precipitation and temperature in eastern China

The seasonality of present-day precipitation in eastern China varies from south to north (Fig. 1). Precipitation rates are maximum in late spring and early summer in southeast China (stations Guilin, Hong Kong, Liuzhou, and to a lesser degree Fuzhou in

Fig. 1), but are maximum in mid- to late summer farther north (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 edge of the front, and Kunming is west of the region affected by frontal dynamics (Fig. 1). Stations at Guiyang and Zunyi, on the western edge of the Meiyu front region, receive maximum precipitation rates in the early summer rather than in the spring (Fig. 1). Low-level winds associated with Meiyu frontal precipitation are generally from the south. Coastal stations Hong Kong and Fuzhou receive high precipitation rates both as the Meiyu front impacts them in late spring to early summer and again in late summer after the Meiyu front has moved northward. These later high precipitation rates are associated with easterly winds (not shown) and may result, at least in part, from local differential land-sea heating. We stress, however, that the majority of the precipitation in southeast China is associated with frontal dynamics, convection, and convergence of the large-scale circulation.

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

We wish to test the hypothesis that $\delta^{18}O$ values in precipitation scale either with the amount of *in situ* precipitation or with *in situ* temperature. To do so, we use data from GNIP stations (IAEA/WMO, 2004) in eastern China (Fig. 1) to calculate correlations of monthly and of 12-month and 24-month running average values of $\delta^{18}O$ in precipitation with local 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 recorded in cave speleothems, as the latter reflect 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 less than 50% of the variance in all cases. In general, temperature is better (anti-) correlated with $\delta^{18}O$ values than is precipitation. Correlations between 12- and 24-month running averages of the variables, however, are generally not significant.

We recognize that the time scales that can be sampled with modern data are short compared with the integrated time sampled by a single measurement of $\delta^{18}O$ in calcite from a speleothem. Nevertheless, we are motivated by two views: first, amplitudes of proxies of climate variability in the paleorecord commonly are comparable to, if not smaller than, amplitudes of variability in modern monthly data; and second, the physical processes that fractionate ^{18}O during evaporation and condensation and the mechanisms by which the atmosphere transports it (the laws of physics) did not differ in the past, even if boundary conditions were different. An understanding of the modern record, therefore, is a prerequisite for interpreting the paleorecord, even where that understanding is quantitatively limited.

4.1 Monthly correlations

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 $\delta^{18}O$ generally then become less negative in the wintertime. Precipitation covaries with $\delta^{18}O$ values throughout southern China less well than does temperature, for maximum precipitation occurs in late spring to early summer, and $\delta^{18}O$ values reach a minimum in late summer (Fig. 1).

For the few stations that show a statistically significant relationship, monthly $\delta^{18}O$ values and precipitation amount are negatively correlated (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.

Figure 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 $\delta^{18}O$ values, except at Shijiazhuang, which lies north of the

Meiyu front region and is unaffected by Meiyu precipitation (Fig. 1). In contrast, monthly anomalies of δ^{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 δ^{18}O signal. Thus we suspect that the correlations between δ^{18}O values and temperature result from correlations of each variable with some other seasonally varying factor such as insolation, the large-scale atmospheric circulation, or precipitation type (e.g., convective storms versus drizzle). If this is the case, it need not be local temperature that determines δ^{18}O values, but instead some other independent process that affects both temperature and the value of the δ^{18}O in the precipitation. Local temperature is thus an indicator of – but not necessarily the cause of – changes in processes elsewhere, and the latter determine the δ^{18}O that is being precipitated over China. We also note that just as δ^{18}O values in the paleorecords decrease with increasing summer insolation (Cai et al., 2006; Wang et al., 2001; Yuan et al., 2004) and hence presumably with increasing local temperature, modern δ^{18}O values decrease with increasing temperature. This tendency, however, is opposite that expected from the temperature dependence in Rayleigh fractionation: ^{18}O passes more readily from vapor to condensate when the air is saturated and the temperature decreases (e.g., Dansgaard, 1964). The lack of agreement between trends in modern δ^{18}O values and expectations based on Rayleigh fractionation has been observed globally in both observations and model results (e.g., Brown et al., 2008; Lee et al., 2007), though a part of this poor agreement stems from re-evaporation of liquid water. In any case, other processes must conspire with Rayleigh fractionation to yield the recorded $\delta^{18}O$ values.

All sites in our study receive most of their precipitation in spring and/or summer, 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 correlations between temperature and precipitation affect correlations between $\delta^{18}O$ values and temperature or precipitation by only small amounts.

4.2 Interannual correlations

For comparison to paleoclimate records, correlations between longer time intervals may be more appropriate than monthly values. Therefore we calculate correlations between the 12-month and 24-month running average values of $\delta^{18}O$ and the 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 $\delta^{18}O$ weighted by the amount of the precipitation that fell during 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. 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 for 12-month averages, and r = -0.45 for 24-month averages), which is suggestive at best. Thus, the available instrumental record neither supports nor excludes the possibility of a relationship between local climate variables and $\delta^{18}O_w$ values in precipitation.

A recent record of $\delta^{18}O$ values from a cave speleothem does, however, show that $\delta^{18}O$ values covary with local temperature and precipitation amount for the past 50 years. Zhang et al. (2008) calculated correlation coefficients of $\delta^{18}O$ values measured in Wangxiang cave from 1950 to 2000 with 5-year running average precipitation (r = -0.64) and temperature (r = 0.8) from a weather station ~15 km from the cave (Zhang et al., 2008, Fig. S4). The magnitude of the interannual variations of the Wangxiang $\delta^{18}O$ values (~0.3‰) is small compared to monthly variations of GNIP data (~6-10‰) and orbital variations of paleoclimate data (~5‰). Note also that the correlation between $\delta^{18}O$ values and precipitation is of the same sign sense as the modern data, but that between $\delta^{18}O$ values and temperature is the opposite. If a paleoclimate record responds similarly to the recent part of the Wangxiang record, then it may be a good indicator of local precipitation. Because the magnitude of variation in the Wangxiang record is so much smaller than orbitally related variations, however, processes other than local precipitation amount must account for the orbitally induced variations in cave $\delta^{18}O$ values at this site. 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 to exploit 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).

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: rainwater is isotopically lighter when temperature is higher (in summer). This negative relationship is like that of orbitally induced changes in paleoclimate records, in that $\delta^{18}O$ values are more depleted during the warmer periods (e.g., Cai et al., 2006; Wang et al., 2001; Yuan et al., 2004), but opposite to that predicted by the temperature dependence in Rayleigh fractionation and to that observed by Zhang et al. (2008) in a modern speleothem record, which they attribute to global climate change. For northern and western stations (Guiyang, Kunming, Nanjing, Shijiazhuang, and Zunyi), maximum temperature and lighter (most negative) $\delta^{18}O$ values also correspond to the maximum precipitation rate (Fig. 1). Locations in southeast China such as Guilin and Liuzhou receive maximum precipitation in spring or early summer but minimum $\delta^{18}O$ values and maximum temperatures occur in late summer, so that $\delta^{18}O$ is more negatively correlated to temperature than to precipitation. The correlations between monthly anomalies of $\delta^{18}O$ values and precipitation or temperature, however, are small and, with a few exceptions, insignificant. Thus, we infer that much of the variation in $\delta^{18}O$ values results from seasonal variation of some process that may not depend directly on local temperature or precipitation.

Partial correlation coefficients (Table 1) show that precipitation contributes to the $\delta^{18}O$ signal in Hong Kong (which is influenced by a summer monsoon-like seasonal precipitation), and to some extent at Shijiazhuang, which is north of the monsoon region (Fig. 1; Table 1), in agreement with the analysis of Johnson and Ingram (2004). 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).

5.1 Simple scaling analysis

A relationship between precipitation amounts or seasonal differences of precipitation and $\delta^{18}O$ values can be tested using a simple scaling analysis that uses estimates based on linear regressions of monthly $\delta^{18}O$ values versus monthly precipitation amounts for stations with statistically significant correlations between them (see Appendix A). In agreement with Johnson et al. (2006b) and Kelly et al. (2006), we find that the difference between modern and last glacial maximum $\delta^{18}O$ values cannot be explained by reasonable differences in the annual amount of local precipitation. For example, to account for only 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), we find that the annually averaged precipitation at 9 ka would be at least 1.5 times greater than today (Appendix A).

Moreover, different seasonal amplitudes of precipitation amounts and of $\delta^{18}O$ values can account for part, but by no means all, of the 4-5‰ differences in $\delta^{18}O$ values in caves, or the 3‰ difference between present-day and Last Glacial Maximum values at Dongge cave (Dykoski et al., 2005). We address analysis of amplitude scaling in detail in Appendix A, but consider the following simple calculation. Suppose, first, that today $\delta^{18}O$ values averaged -6‰ during the 9 autumn, winter, and spring months and -10‰ during one 3-month season; second, suppose that roughly the same precipitation fell in the three summer months as during the other nine months, so that the mean annual $\delta^{18}O$ value were ~-8‰. (These values approximate those for Nanjing in Figure 1.) Now suppose that in the past either the 3 summer months' precipitation did not occur at all or it carried the same $\delta^{18}O$ values as the average for the other seasons; the resulting annual average $\delta^{18}O$ value would have been -6‰, only 2‰ different from that today. This, obviously, is an extreme consideration, given the complete elimination of all of the most negative $\delta^{18}O$ values, and that precipitation rates during some months when $\delta^{18}O$ values are most negative are relatively low. Following the reasoning above and the more rigorous calculations in Appendix A, we conclude that no plausible difference in the amplitude of seasonal cycle of precipitation amount or its monthly $\delta^{18}O$ values can account for even as much as half of variability of $\delta^{18}O$ values in the speleothems of China.

GCM calculations of past climates also yield smaller glacial-interglacial differences in $\delta^{18}O$ values than those reported in cave records (e.g., Hoffmann and Heiman, 1997, Hoffmann et al., 2000, Jouzel et al., 1994) (Appendix B). Hence, neither the GCM results nor the standard views of how $\delta^{18}O$ values vary in modern precipitation can account for the full range of glacial-interglacial differences in $\delta^{18}O$ values, though each can account for a part of that range.

5.2 Additional processes that may affect $\delta^{18}O$ values

The work of Johnson and Ingram (2004), Kelly et al. (2006), and Yuan et al. (2004), along with the analysis above, point to changes in atmospheric circulation and in moisture sources to explain the majority of the glacial-interglacial variation of $\delta^{18}O$ values of 4-5‰. Below we consider how glacial and interglacial $\delta^{18}O$ values might differ at a cave site by assuming $\delta^{18}O$ values differences due to changes at the moisture source, in the atmospheric circulation, and the storm type. The value of this exercise is not in the specific $\delta^{18}O$ values we list (which are only rough estimates), but rather in the approach: we hypothesize how atmospheric processes during past climates might have been different from those of today, and estimate the potential for those processes to contribute to the orbital scale variations in the cave records across China.

5.2.1. Moisture source

Ocean $\delta^{18}O$ values are enriched during glacial times, as lighter oxygen isotopes are preferentially sequestered in glaciers and ice sheets. Mean glacial ocean $\delta^{18}O$ values increase ~ 1‰ compared to interglacial ocean $\delta^{18}O$ values (e.g., Guilderson et al., 2001; Schrag et al., 2002). Sea surface temperature is reduced by ~ 2-3°C in the South China Sea during glacial periods (Oppo and Sun, 2005), however, and this may have led to a ~0.5‰ decrease of $\delta^{18}O$ values in vapor sourced from that area. Thus the enrichment of ^{18}O in vapor from the net of the ocean changes in glacial times is a relatively small ~ 0.5‰.

5.2.2. Shifts in atmospheric circulation

Because of decreased insolation during glacial times, when colder conditions prevailed especially at higher latitudes, the Meiyu front and the dynamics associated with it might not have migrated as far north as it does today. To estimate the effect of this southward contraction of the region affected by that front, we substitute summer $\delta^{18}O$ values for one station (Nanjing, for example) with those from a station farther north (Shijiazhuang, for example), and we calculate that mean annual $\delta^{18}O$ values would increase ~1‰ during glacial times. We offer this as one possible shift in atmospheric circulation, but as discussed below, others seem just as plausible.

5.2.3. Rainstorm type

Precipitation from convective storms has been measured to have lower $\delta^{18}O$ values than non-convective precipitation (e.g., Lawrence and Gedzelman, 1996; Lawrence et al., 2004; Risi et al., 2008; Scholl et al., 2009). For example, convective rainfall during the monsoon in Niger is ~ -2 to -6‰, whereas non-convective precipitation before the monsoon is ~ 0‰ (Risi et al., 2008). Similarly, low $\delta^{18}O$ values are also observed in China during the wet season (Fig. 1), when the area receives convective precipitation related to the Meiyu front. Notice that except for Shijiazhuang most of the months with the most negative $\delta^{18}O$ values are those with the largest mean monthly temperatures (Fig. 4). If during glacial times, surface temperatures were 5°C cooler than today, as studies elsewhere in the subtropics suggest (Stute et al., 1992, 1995), summer precipitation might have been denied the very low $\delta^{18}O$ values. Holmgren et al. (2003) suggested the fraction of precipitation from convection may vary on millennial time scales and this might account for isotopic differences in speleothems in South Africa. Similarly, we hypothesize that convective rainfall, which scales with radiative cooling of the atmosphere (e.g., Emanuel, 2007), is reduced during glacial times. Using the monthly average values, we calculate annual weighted mean $\delta^{18}O$ values in which we eliminate months when precipitation is likely convective, and replace those months with $\delta^{18}O$ values of months when precipitation is likely not convective. We estimate annual weighted mean $\delta^{18}O$ values without convective rainfall could be ~ 2‰ heavier than modern mean annual $\delta^{18}O$ values.

It appears that none of the processes suggested above can, alone, account for the 4-5‰ difference between glacial and interglacial $\delta^{18}O$ values, and even their sum of ~ 3.5‰ seems to be too small. Part of this difference could be made up by seasonal differences in precipitation amounts that do not depend on precipitation type, such as what others have called 'monsoon intensity' (e.g., Cai et al., 2006; Cheng et al., 2006, 2009; Dykoski et al., 2005; Kelly et al., 2006; Wang et al., 2008; Yuan et al., 2004). In any case, each of the estimates made above is approximate, and we offer them as examples of how differences between modern and paleoclimate might account for observed differences in $\delta^{18}O$ values. With better understanding of how modern $\delta^{18}O$ values are influenced by atmospheric circulation and mixing processes, hypothesized circulation changes may be rejected or supported.

5.3. Durations of seasons

The analysis above and in Appendix A has assessed how varying the amplitude of the seasonal cycle of precipitation could affect mean annual $\delta^{18}O$ values, which can be seen as an amplitude modulation of the seasonal cycle. The high spectral power in the precession band, however, raises the question of whether a better characterization of the orbital forcing would exploit frequency modulation. As Huybers (2006) has shown, Kepler's second law requires that the variation in durations of seasons play a key role in how the earth's climate responds to orbital forcing.

So, suppose that instead of the amplitudes of seasonal precipitation amounts and/or $\delta^{18}O$ values varied according to some seasonal cycle with a 1-year period, durations of seasons changed, and with them so did $\delta^{18}O$ values (e.g., Cheng et al., 2009). As an example, consider the case for Nanjing illustrated in Figure 1. Suppose that in early Holocene time, when climate was warmest, the winter jet moved north across Tibet not in May, as it does today (Schiemann et al., 2009), but in March. As an extreme example, if summer rains like those currently in June, July, and August and $\delta^{18}O$ values of -9‰ to -10‰, replaced those in April and May (with present-day values of -3‰ to -4‰), the mean annual $\delta^{18}O$ values would differ by ~1‰, consistent with the difference between early Holocene and present-day ^{18}O values in caves. Similarly, if during glacial times, the mid-latitude jet remained south of Tibet throughout much of the summer, and if present-day springtime $\delta^{18}O$ values characterized those of summer months, the calculated difference in mean annual $\delta^{18}O$ values between Last Glacial Maximum conditions and the present-day would be more than 3‰. This too is an extreme assumption for the seasonal differences in $\delta^{18}O$ values, but unpublished idealized General

Circulation Model calculations of K. Takahashi and Battisti (discussed briefly by Molnar et al. (2010)) suggest that winter and spring rain result from the mid-latitude jet passing south of Tibet and then accelerating over eastern China. Thus, the possibility that in glacial times the jet remained south of Tibet seems plausible.

Because the annual cycles of precipitation and of $\delta^{18}O$ values differ from station to station, we cannot argue that the differences in isotopes recorded in speleothems among present-day, early Holocene, and glacial times will be the same throughout China. In fact, it seems unlikely that a simple explanation of the kind offered above can account for the full range of variations in speleothems. The simple arguments given in the previous paragraph do suggest, however, that if different amplitudes of seasonal variations cannot account for more than small fraction of the amplitude of variability seen in speleothems, different durations of seasons may be an important factor that controls variability on orbital time scales.

6. Conclusions

Modern station data offer little support for the idea that monthly or annual variations in $\delta^{18}O$ values reflect variations in local precipitation on the same time scales, and 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) – but are opposite to that expected from the temperature dependence in Rayleigh fractionation. Monthly anomalies of $\delta^{18}O$ values, however, do not generally correlate well 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 infer that variation of $\delta^{18}O$ values and temperature on the seasonal time scale primarily reflect independent processes each of which is regulated by changes in insolation: local insolation directly regulates local temperature, and global insolation gradients, correlated with local insolation, affect 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. 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. Although one $\delta^{18}O$ record from 1953 to 2000 collected from Wanxiang cave (Zhang et al., 2008) correlates negatively with local 5-year average precipitation and positively with temperature, the magnitude of variation of $\delta^{18}O$ values on this timescale is much smaller than that on orbital timescales.

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 necessary to account for variability in $\delta^{18}O$ values in the paleoclimate record. 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%. In light of the results above, we, like others (e.g., Cai et al., 2006; Cheng et al., 2006; Dykoski et al., 2005; Johnson and Ingram, 2004, Kelly et al., 2006; Wang et al., 2008; Yuan et al., 2004) conclude that other processes – such as differences in re-evaporation, variations in atmospheric circulation, and variations in rainstorm type – have as much influence on orbital scale variability of $\delta^{18}O$ values in China as do precipitation amount or temperature. Similarly, we argue that different, but plausible, amplitudes of the seasonal cycle of precipitation amount and of $\delta^{18}O$ values can account for only a part, less than half, of the difference between glacial and present-day $\delta^{18}O$ values.

The climatic cause of isotope fluctuations in Chinese speleothem records and the nature of their link to the overall East Asian monsoon circulation remain open questions. Several differences between past and present-day climates may combine to account for the 3-6‰ amplitude of variability on orbital time scales. These include different values in the oceans from which water is evaporated, different sources of moisture and different pathways, different amounts of convective precipitation, which is highly depleted in ^{18}O, and different durations of seasons. Some of these processes may conspire together as separate and even independent consequences of differences in large- and small-scale atmospheric circulation, such as the seasonal cycle of the mid-latitude jet strength and position and summer radiative heating over China, that result directly from differing boundary conditions imposed by variations in isolation on a Milankovitch time scale and from the effects of high-latitude ice sheets.

Appendix A. Simple scaling relationship

We perform a scale analysis to explore whether local precipitation differences, arising either because of different amplitudes of annual precipitation or differing amplitudes of the seasonal cycle, can plausibly explain the differences in $\delta^{18}O$ in the cave records on orbital time scales. We assume that $\delta^{18}O$ values are a valid proxy for monthly precipitation amounts and use empirical relationships between monthly precipitation and monthly average $\delta^{18}O$ to ask: How much must annual precipitation amount or seasonality 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‰ near ~ 130 ka, 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 ~ 15 ka, a difference of ~4‰ (Yuan et al., 2004). For comparison the amplitude of variability in modern $\delta^{18}O$ values is ~ 7 – 8‰, and the maximum peak-to-peak difference in $\delta^{18}O$ values in the modern speleothem record from Wangxiang is only ~ 0.3‰ (Zhang et al., 2008).

We estimate 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 factors that scale 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 effects of different annual means f_o and the seasonal amplitudes f', assuming that $\delta^{18}O$ values scale with the monthly 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). The station-specific values of $a = \Delta\delta^{18}O/\Delta P$ calculated above, however, are similar to those calculated by Bony et al. (2008) using a simple column-integrated model for radiative-convective equilibrium over tropical ocean and by Lee et al. (2008) using atmospheric GCM with an isotope model that includes a dependence of isotopic content on precipitation amount during rainfall events (Lee et al., 2007).

We consider stations where monthly values of $\delta^{18}O$ and precipitation are significantly correlated: Guiyang, Hong Kong, Kunming, and Zunyi, noting that others also 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 go on to argue that dependences of $\delta^{18}O$ on the amount of precipitation or on local temperature cannot account for the $\delta^{18}O$ record at this cave. Similarly, Kelly et al. (2006) require > 95% differences from present-day in precipitation to explain differences in glacial and interglacial $\delta^{18}O$ values from Dongge cave.

If $P'(t)$ describes the seasonal cycle of precipitation, then the mean annual $\delta^{18}O$ value, weighted by seasonal variations in precipitation, is:

$$\delta_{ae}(t) = \frac{\int_0^{2\pi} P\delta dt}{\int_0^{2\pi} P dt}. \quad (3)$$

Suppose, first, that $P'(t)$ can be described with a cosine function,

$$P'(t) = P_a \cos(t) \quad (4)$$

where P_a is the maximum monthly precipitation anomaly, then with substitution of (1), (2), and (4), (3) becomes:

$$\delta_{ae} = af_o P_o + b + \frac{af'^2 P_a^{\,2}}{2 f_o P_o} \quad (5)$$

The difference between a past climate state and the modern (for which δ_{ae} in (5) is given by substituting $f_o = f' = 1$) is:

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

To simplify matters, we have assumed that $P_a = P_o$ in (6) (i.e., modern precipitation is $P = P_o(1 - \cos(t))$). We require that $| f'/f_o | \leq 1$ to ensure positive values of precipitation. Note that the difference between past and modern climates is not dependent on b, defined in (2), because we have assumed that the relationship between $\delta^{18}O$ value and precipitation is invariant with time. Using modern data to assign values to a and P_o for each station, we plot D as a function of mean annual precipitation amount f_o holding $f' = 1$ (Fig. 5a), D as a function of seasonal amplitudes of precipitation f' holding $f_o = 1$ (Fig. 5b), and D for the case where the mean annual and seasonal amplitude of precipitation vary proportionally: $f = f' = f_o$ (Fig. 5c). The minimum f_o and maximum f' values that we consider define limits beyond which the absolute value of the monthly precipitation anomaly in the driest months of the year would be larger than the annual mean, resulting in negative precipitation for the month. We also calculate D using the observed modern seasonal cycles (Figure 1) instead of a cosine function in (4), and the resulting curves differ only slightly in shape from those plotted in Fig. 5. Thus, insofar as modern scaling of $\delta^{18}O$ on monthly precipitation amounts applies, variations in the shape of the seasonal cycle have little effect on D.

The gray bands in Fig. 5a and 5c indicate minimum values of $\delta^{18}O$ values relative to modern values from Hulu and Dongge caves (Yuan et al., 2004). To decrease the $\delta^{18}O$ value by 1‰ (the approximate difference between $\delta^{18}O$ values of present-day and 9 ka), we estimate that 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 the formulation presented above, changing the amplitude of the seasonal cycle cannot cause a decrease of the $\delta^{18}O$ value by as much as 1‰, given the upper limit of f' (Fig. 5b). For the relative amount of summer to winter precipitation to increase sufficiently to call for a 1‰ decrease in the $\delta^{18}O$ values, mean annual precipitation must be >~1.4 times larger than present at Kunming and Zunyi, and >~1.6 times larger at Hong Kong (Fig. 5c).

We can also apply our analysis to the recent $\delta^{18}O$ record of Zhang et al. (2008) from Wangxiang cave, assuming the seasonal cycle of precipitation there is reasonably described by nearby stations Zunyi or Kunming. The maximum peak-to-peak difference in their $\delta^{18}O$ record is a decrease of 0.3‰ between 1998 and 1986, when precipitation increases from 320 mm/yr to 430 mm/yr (Zhang et al., 2008, Fig. S4). Using the simple scaling law above and parameters appropriate for Zunyi or Kunming, a decrease of 0.3‰ implies a precipitation rate of 1.2 times 320 mm/yr, or ~380 mm/yr (Fig. 5a, 5c). In the case of modern speleothem $\delta^{18}O$ records, therefore, the dependence on precipitation amount underestimates the observed difference in precipitation. At the opposite extreme, the small differences in $\delta^{18}O$ for relatively large variations of precipitation at Wangxiang calls for absurdly large glacial-interglacial variations. The 110 mm/yr annual rainfall difference correlated with a 0.3‰ in $\delta^{18}O$ values would require differences of >1000 mm in annual rainfall at speleothem sites, when present-day annual precipitation is ~ 480 mm (Zhang et al., 2008).

Appendix B. Comparison with General Circulation Model calculations

The failure of modern relationships between monthly $\delta^{18}O$ values and precipitation amount pose the question of the relationship between $\delta^{18}O$ values and climate variables on longer time scales differ from that for the present. GCMs can be used to examine this from their calculated amounts of either precipitation or $\delta^{18}O$ values vary in past climates. Several runs have been carried out with such tests in mind.

Using GCM experiments Kutzbach (1981) estimates ~ 10% greater summertime precipitation and ~ 5% greater annually averaged precipitation amount at 9 ka than in modern day. GCM ensemble results from the PMIP2 experiment show no significant change in annual mean precipitation from 6 ka to present (Braconnot et al., 2007). By comparison, our calculations above suggest that if a dependence on precipitation amount 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 these calculations suggest that insofar as the modern dependence of $\delta^{18}O$ values on precipitation applies to paleoclimate, summer precipitation amount cannot be the explanation for the large variations in $\delta^{18}O$ values in speleothems in China.

We note also that most GCM runs that include stable isotopes of water predict smaller differences between glacial and interglacial $\delta^{18}O$ values than those measured in cave records. Calculated differences in $\delta^{18}O$ values in China from GCM simulations are ~2‰ between the last glacial maximum and present (Hoffmann and Heiman, 1997, Hoffmann et al., 2000, Jouzel et al., 1994), which are more than two small times smaller than those observed in cave speleothems (Yuan et al., 2004). Recent simulations of $\delta^{18}O$ values over Holocene time by LeGrande and Schmidt (2009) do replicate approximately the ~1‰ difference in this period, but they attribute this difference largely to differences in vapor transport from the Pacific, not to precipitation amount or seasonal differences in it.

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Table caption

Table 1: Correlation coefficients *r* and partial correlation coefficients ρ calculated for $\delta^{18}O$ values and precipitation *P* and temperature *T* from nine GNIP stations. 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 cycles 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 South Asian 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, whose locations are shown in Figure 1. 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: Calculated annual average weighted $\delta^{18}O$ values relative to modern, *D*, as a function of (a) f_o for $f' = 1$, (b) f' for $f_o = 1$, and (c) $f = f_o = f'$ calculated using equation (6) relative to modern values. Dotted lines (marked 'G') are calculations for station at Guiyang, dashed lines (marked 'HK') are for Hong Kong, dot-dashed lines (marked 'K') are for Kunming, and solid lines (marked 'Z') are for Zunyi. Grey bands in (a) and (c) indicate minimum $\delta^{18}O$ values in the records from Dongge and Hulu caves (2004). 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.

Figure

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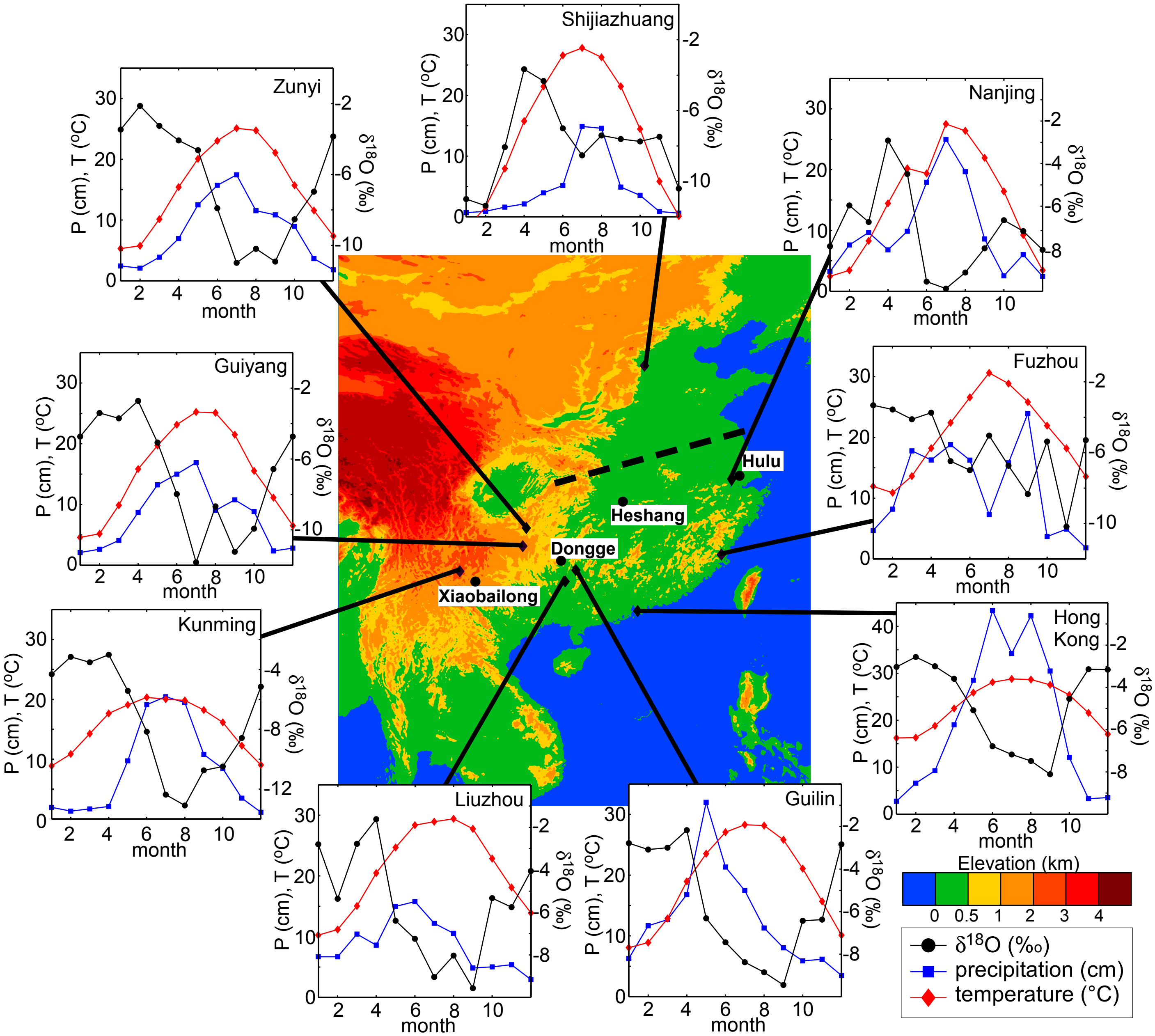

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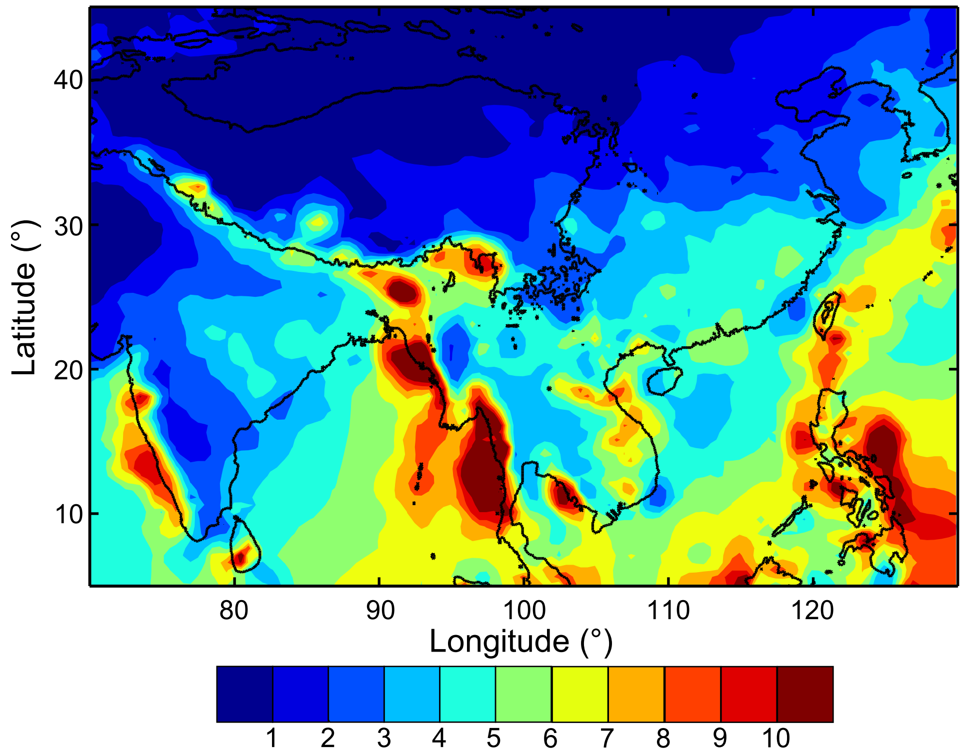

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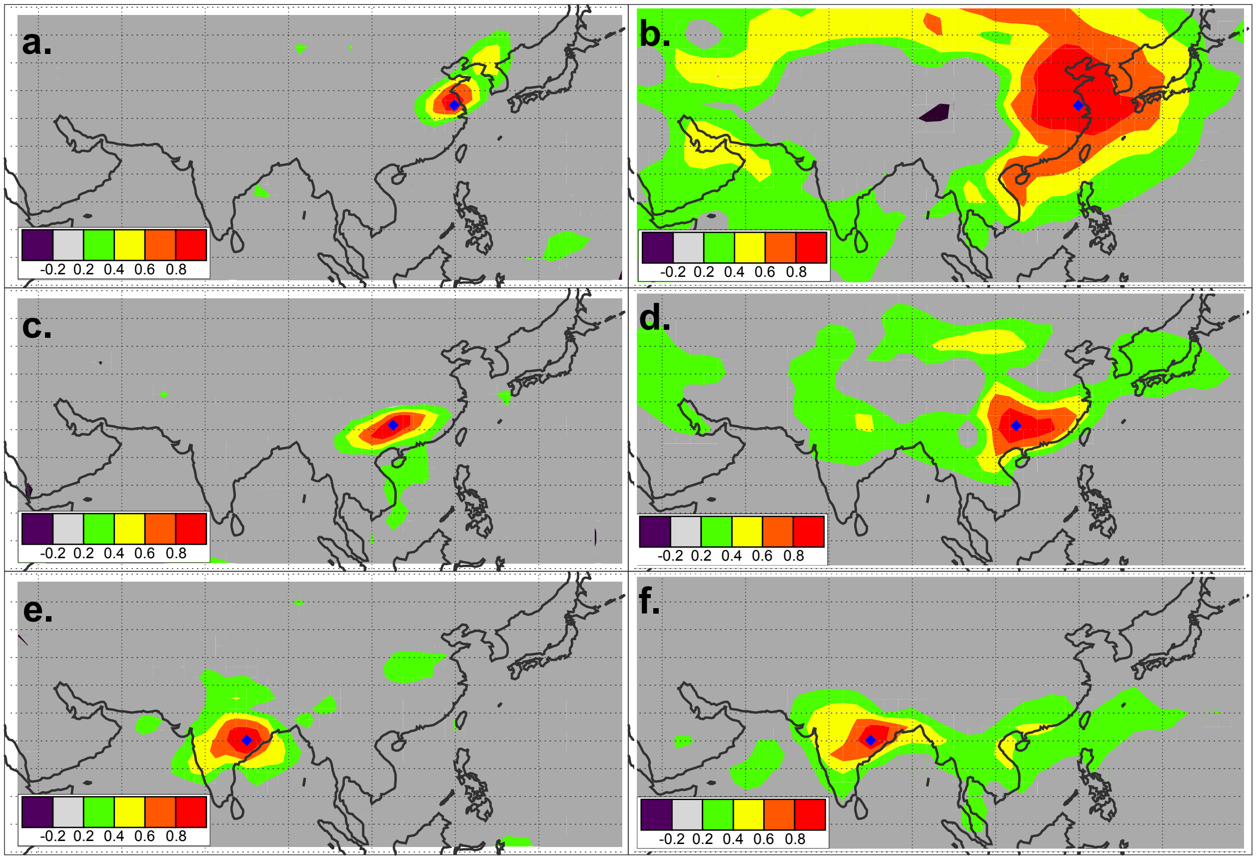

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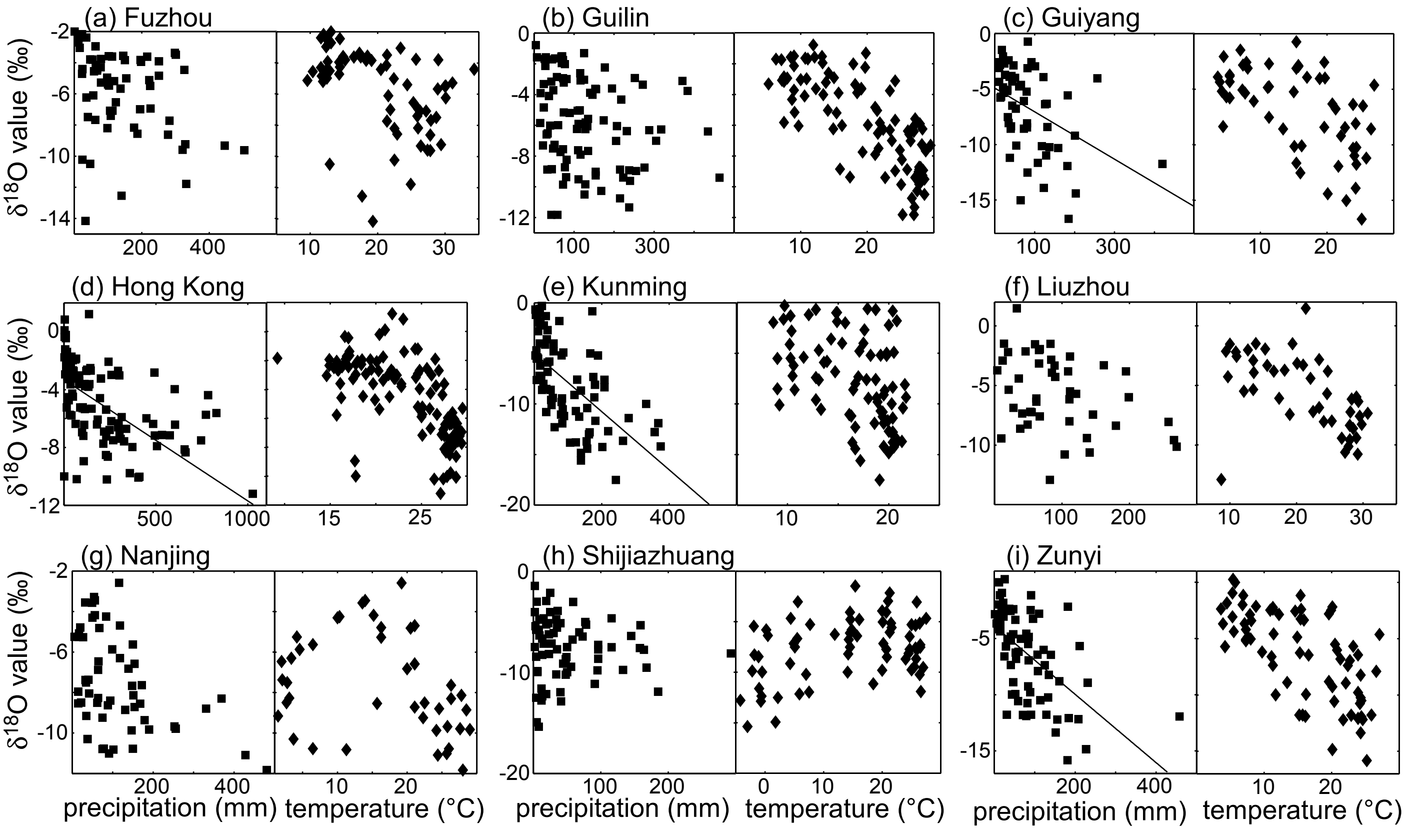

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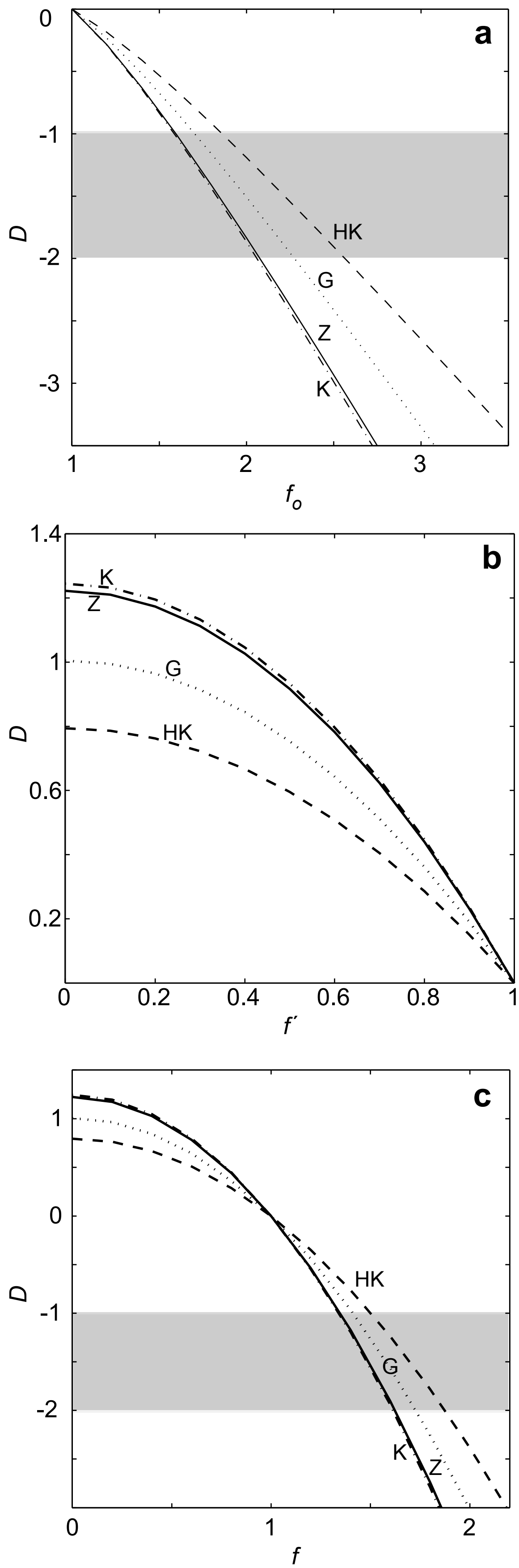

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Table
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station	monthly average		monthly anomaly		monthly average partial correlations		n
	r(δ^{18}O,P)	r(δ^{18}O,T)	r(δ^{18}O,P)	r(δ^{18}O,T)	ρ(δ^{18}O,P,T)	ρ(δ^{18}O,T,P)	
Fuzhou	-0.35	-0.38	-0.36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