Long-term trends in the variation of the δD and $\delta^{18}O$ of precipitation and in humidity during the Holocene in the Indian and East Asian monsoonal regions

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Symbols:

$\delta^{18}O_p$ and δD_p : Oxygen and hydrogen isotopes of precipitation; δD_n : hydrogen isotopes of lacustrine/marine *n*-alkanes/*n*-alkanoic acid; $\delta^{18}O_c$: oxygen isotopes of lacustrine authigenic carbonate; $\delta^{18}O_s$: oxygen isotopes of stalagmite.

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42 **ABSTRACT**

43 This paper aims to achieve an improved understanding of the long-term change trends
44 of precipitation δD and $\delta^{18}\text{O}$ values (δD_p and $\delta^{18}\text{O}_p$) in the Asian monsoonal region

and their relationship with the corresponding humidity trends during the Holocene. To do this we first review the observed modern spatial pattern of summer precipitation distribution in the East Asian summer monsoon (EASM) region under different EASM intensities, and the relationship between modern observed $\delta^{18}\text{O}_p$ values and corresponding precipitation amounts on monthly and inter-annual timescales in the EASM and Indian summer monsoon (ISM) regions. Second, we compare Holocene lacustrine and marine compound-specific hydrogen isotopic records of *n*-alkanes/*n*-alkanoic acid (δD_n), lacustrine authigenic carbonate and cave stalagmite oxygen isotopic records ($\delta^{18}\text{O}_c$, $\delta^{18}\text{O}_s$) from the Asian monsoonal region, all of which are closely related to δD_p and $\delta^{18}\text{O}_p$ variations. The results demonstrate that in both the ISM and EASM regions, all of these isotopic records exhibit roughly similar long-term characteristics, *i.e.* they were all more negative during the early-Holocene and early mid-Holocene (ca. 11-6 ka B.P.; B.P. means before present, present = 1950 AD), and then became more positive towards the late-Holocene. Third, we compare representative paleo-humidity records from the Asian monsoonal region; the results confirm that, in the ISM region, a humid interval occurred in the early-Holocene and early mid-Holocene (ca. 11-6 ka B.P.) and subsequently the climate became more arid towards the late-Holocene. This indicates an enhanced ISM during the early-Holocene and early mid-Holocene (ca. 11-6 ka B.P.), and an ISM of decreasing intensity towards the late-Holocene. On a Holocene orbital scale, both $\delta^{18}\text{O}_p$ and δD_p appear to be controlled by an "amount effect" in the ISM region, consistent with the region's inter-annual modern $\delta^{18}\text{O}_p$ data. This evidence indicates that both $\delta^{18}\text{O}_p$ and δD_p

paleo-records are significantly related to paleo-humidity in the ISM region. In contrast, Holocene humidity variations in the EASM region exhibit clear spatial differences: a humid mid-Holocene interval (ca. 8-3 ka B.P.) occurred in southern and northern China, but an arid interval from ca. 7-3 ka B.P. occurred in central China, in the middle reaches of the Yangtze River. Based on precipitation distribution patterns under different EASM intensities in the EASM region over the past few decades, we conclude that EASM intensity was enhanced during the mid-Holocene (ca. 8-3 ka B.P.). Relative to the ISM intensity, the response of EASM intensity to summer insolation was relatively slow. In the EASM region the relationship between climate and $\delta^{18}\text{O}_p$ and δD_p is more complex, consistent with analyses of regional inter-annual modern $\delta^{18}\text{O}_p$ data. This evidence demonstrates that both $\delta^{18}\text{O}_p$ and δD_p paleo-records cannot be used directly as paleo-humidity (i.e. precipitation amount or EASM intensity) indicators in the EASM region. Further comparison and analyses demonstrate that the coupled variations in west-east Equatorial Pacific temperature gradients and the West Pacific subtropical high (WPSH) played an important role in determining EASM intensity during the Holocene.

Keywords: precipitation isotopes, stalagmite $\delta^{18}\text{O}$ record, lacustrine $\delta^{18}\text{O}$ and δD record, Indian summer monsoon, East Asian summer monsoon, Holocene

1. Introduction

Proxy records based on past variations in precipitation isotopes, including both

hydrogen and oxygen isotopes (δD_p , $\delta^{18}O_p$), have played an extremely important role in paleoclimatic studies. Over the past few decades, ice-core δD and $\delta^{18}O$ records from polar regions (accepted as paleotemperature indicators, [North Greenland Ice Core Project members, 2004](#); [Jouzel *et al.*, 2007](#)) and $\delta^{18}O$ records from marine sediments (accepted as indicators of sea water $\delta^{18}O$ and bottom water temperatures but used to infer changes in global ice volume and therefore as indirect indicators of global mean temperature, [Imbrie *et al.*, 1984](#); [Pisias *et al.*, 1984](#); [Martinson *et al.*, 1987](#); [Elderfield *et al.*, 2012](#)) have been widely accepted as the benchmark for paleoclimatic comparisons. Similarly, precipitation δD records preserved in the leaf wax of terrestrial plants in tropical regions have been used to reconstruct historical changes in local humidity or precipitation amount ([Schefuß *et al.*, 2005, 2011](#); [Tierney *et al.*, 2008](#)). Initial analyses of global monthly observation data for $\delta^{18}O_p$ demonstrate that $\delta^{18}O_p$ values are controlled by a "temperature effect" (shown as a positive correlation with temperature) in high latitude polar regions, but by an "amount effect" (shown as a negative correlation with precipitation amount) in low latitude tropical regions. In mid-latitudes, $\delta^{18}O_p$ exhibits a complex relationship with climate, and may be simultaneously influenced by both "effects" ([Dansgaard, 1964](#)).

Over the past two decades, the historical evolution of EASM intensity has been widely discussed. In the early 1990s, based mainly on reviews and summaries of Chinese data from the 1970s onwards, the interval from 8.5-3 ka B.P. has been termed the "mid-Holocene Megathermal" ([Shi *et al.*, 1994](#)). A "Holocene Optimum", defined by peak EASM rainfall, was then determined to be asynchronous in eastern China,

with an apparent southward-retreating trend (An *et al.*, 2000). However, subsequent evidence from stalagmite oxygen isotopes ($\delta^{18}\text{O}_s$, Dykoski *et al.*, 2005; Wang *et al.*, 2005; Dong *et al.*, 2010) and other records did not support this asynchronous pattern of EASM evolution, but rather it suggested a synchronous pattern of EASM evolution, with an intensified EASM during the interval from ca. 10-5 ka B.P. (Zhou *et al.*, 2007). Recent data from loess deposits in northern China indicates that an intensified EASM occurred from ca. 8-3 ka B.P. (Lu *et al.*, 2013). Nonetheless, stalagmite and peat bog data from the mid-reaches of the Yangtze River in central China indicate an arid interval from ca. 7-3 ka B.P. (Xie *et al.*, 2013).

The dating of speleothems has reached a high degree of precision and accuracy and high resolution records of oxygen isotope from stalagmites from monsoonal China, considered as indicators of EASM intensity, have been widely reported (Wang *et al.*, 2001, 2005, 2008; Yuan *et al.*, 2004; Zhang *et al.*, 2008) and discussed (Cheng *et al.*, 2005; Maher, 2008; Rao *et al.*, 2015; Liu Z *et al.*, 2014). Discussion has focused on the factors that may affect Chinese $\delta^{18}\text{O}_s$ values, such as ISM intensity (Pausata *et al.*, 2011; Yang *et al.*, 2014; Liu *et al.*, 2015), different vapor source ratios (Clemens *et al.*, 2010), and a "circulation effect" (Tan, 2009, 2011, 2014; Tan and Nan, 2010; Caley *et al.*, 2014). The Chinese $\delta^{18}\text{O}_s$ record may not therefore be an unambiguous indicator of EASM intensity.

Evidence from modern plants (Smith and Freeman, 2006; Liu and Yang, 2008; Kahmen *et al.*, 2013), surface soils (Jia *et al.*, 2008; Rao *et al.*, 2009, 2011; Tipple and Pagani, 2013) and surface lacustrine sediments (Huang *et al.*, 2004; Sachse *et al.*,

2004; Polissar and Freeman, 2010; Garcin *et al.*, 2012), has demonstrated that compound-specific hydrogen isotopes of long-chain *n*-alkanes (δD_n) derived from terrestrial higher plants record variations in δD_p , while mid- and short-chain δD_n data derived from aquatic plants and lower organisms (such as alga) record variations in lake water δD (Sachse *et al.*, 2012). It has been widely acknowledged that lacustrine authigenic carbonate oxygen isotopes ($\delta^{18}O_c$) also record variations in lake water $\delta^{18}O$ (Zhang *et al.*, 2011). Most lakes are mainly fed by precipitation within their catchments, via river water, spring water and melt water, meaning that the δD and $\delta^{18}O$ of the lake water is linked directly or indirectly to the δD_p and $\delta^{18}O_p$ in their catchments. This is the case in the absence of the significant modification of lake waters by environmental processes, such as evaporation. Essentially, therefore, the indicative significance of lacustrine δD_n and $\delta^{18}O_c$ as possible indicators of δD_p and $\delta^{18}O_p$ respectively, and the indicative significance of $\delta^{18}O_s$ as an indicator of $\delta^{18}O_p$, constitute the physical basis for the direct comparison of lacustrine δD_n and $\delta^{18}O_c$ records and $\delta^{18}O_s$ records, because the co-variation of δD_p and $\delta^{18}O_p$ is widely observed (Craig, 1961). This is very important for monsoonal margin regions since these areas are sensitive to climate change; however, high quality Holocene $\delta^{18}O_s$ records are rarely reported, and long-term observations of δD_p and $\delta^{18}O_p$ are rarely conducted.

Both the Holocene evolution of EASM intensity and the climatic significance of Chinese $\delta^{18}O_s$ records are still debated. However, lacustrine and marine sediments provide data that can be directly compared to $\delta^{18}O_s$ records. Therefore in order to

better understand the Holocene evolution of δD_p and $\delta^{18}O_p$ in the Asian monsoonal region and its relationship to climatic change, in the present study we (i) review modern $\delta^{18}O_p$ data from both the ISM and EASM regions and modern summer precipitation distribution in the EASM region under different EASM intensities; (ii) compare lacustrine and marine δD_n records from seven sites, and $\delta^{18}O_c$ records from eight sites, with $\delta^{18}O_s$ records from five sites in the Asian monsoonal region (**Fig. 1**); and (iii) consider the relationships between δD_p and $\delta^{18}O_p$ and climate in these two sub-monsoonal systems, and the most important factors that may have affected EASM intensity.

2. Review

2.1. Modern EASM intensity and spatial precipitation modes in eastern China

In the Asian monsoonal region, the most important feature of the long-term trends of Holocene $\delta^{18}O_s$ records is their spatial consistency (e.g. [Yang et al., 2014](#); [Dong et al., 2015](#); [Liu et al., 2015](#)). However, the inconsistent spatial distribution of summer precipitation in the EASM region is also widely known, described colloquially as “flood in the south and drought in the north”, or “drought in the south and flood in the north”. This phenomenon is one of the most important reasons for questioning the fidelity of the Chinese $\delta^{18}O_s$ record as an indicator of EASM intensity or the amount of monsoonal precipitation.

A recent analysis of high spatial resolution ($0.025^\circ \times 0.025^\circ$) gridded data based on observational data from 1951 to 2012 from 753 meteorological stations in China

demonstrate an increasing trend of summer (JJA, June, July and August) precipitation in southern China, but a decreasing trend in northern China (Wu *et al.*, 2014; Liu *et al.*, 2015). This indicates an overall retreating trend of EASM intensity during the study interval, i.e. the southward movement of the rain belt.

In fact, this dipolar pattern of summer precipitation in monsoonal China was suggested previously, by Ding *et al.* (2008), as a result of the analysis of observational data from 1951 to 2004 from 740 surface meteorological stations in China. More importantly, using 1971–2000 climatological means from the 740 meteorological stations as the background, Ding *et al.* (2008) analyzed the spatial distribution of average summer (JJA) precipitation in China during the intervals of 1951–1978 and 1979–1992. The results demonstrated that when the EASM was more intense in 1951–1978, average summer (JJA) precipitation increased in southern and northern China, but decreased in the middle and lower reaches of the Yangtze River. In other words, from southern to central China, and then to northern China, summer precipitation exhibited a mode of spatial variability which can be represented as "+-+" (Fig. 2). However, under the weaker EASM background of 1979–1992, the average summer (JJA) precipitation decreased in southern and northern China, but increased in the areas between the middle and lower reaches of the Yangtze and Yellow River valleys. In other words, from southern to central China, and then to northern China, summer precipitation exhibited a "-+-" mode of spatial variability (Fig. 2; Ding *et al.*, 2008).

In our view a more detailed documentation of the spatial representation of the

“+-+” and “++-” modes of summer precipitation in East China is of significant importance for paleoclimatic studies, since individual paleoclimate records from the region are obtained from specific locations and therefore may not be representative of larger regions.

2.2. Monthly and inter-annual correlations between modern precipitation $\delta^{18}O_p$ and precipitation amount in the ISM and EASM regions

In order to assess the paleoclimatic significance of the Chinese $\delta^{18}O_s$ records, the modern $\delta^{18}O_p$ data from monsoonal China have been analyzed. The interpretation of the Chinese $\delta^{18}O_s$ record as an indicator of EASM intensity is based mainly on the observation that the $\delta^{18}O_p$ data in southern China are generally more negative during the summer monsoon seasons (Cheng *et al.*, 2005). However, another study of modern $\delta^{18}O_p$ data from meteorological stations in China, provided by the Global Network of Isotopes in Precipitation (GNIP), demonstrated no consistently significant correlations between the $\delta^{18}O_p$ data and either precipitation amount or temperature, on both monthly and inter-annual timescales (Johnson and Ingram, 2004). Moreover, spatial co-variability analysis demonstrated that highly consistent covariation of the precipitation amount of the annual rainy season in Eastern Asia only occurred over a limited spatial extent, within 500 km. This distance is much less than the distance (>1000 km) between the relevant caves with $\delta^{18}O_s$ records which exhibit highly similar patterns of variation. This casts considerable doubt on the reliability of Chinese $\delta^{18}O_s$ records as an indicator of monsoonal precipitation amount (Dayem *et*

al., 2010).

More recently, modern observational $\delta^{18}\text{O}_p$ and climate data from the GNIP database for eight meteorological stations in the Asian monsoonal region were used for monthly and inter-annual correlation analyses between $\delta^{18}\text{O}_p$ and precipitation amount (Li *et al.*, 2015). All eight stations have relatively abundant observation data. The Kozhikode (KK), Mumbai (MB) and New Delhi (ND) stations in India, and the Kunming (KM) station in southwestern China, are located in the ISM region; the other four stations, in Hong Kong (HK), Nanjing (NJ), Zhengzhou (ZZ) and Shijiazhuang (SJZ) in eastern China, are located in the EASM region. The spatial distribution of these eight stations is shown in **Fig. 1**. Additional information about the stations, including location, modern climate and data source years, is listed in **Table 1**.

On a monthly timescale, there are inconsistent and opposing south-north correlations between $\delta^{18}\text{O}_p$ and precipitation amount in both the ISM and EASM regions (**Fig. 3**). In the EASM region, the correlation between monthly mean $\delta^{18}\text{O}_p$ and monthly mean precipitation is negative for HK and NJ in southern China, insignificantly negative at ZZ in central China, but insignificantly positive at SJZ in northern China. In the ISM region, the correlation between monthly mean $\delta^{18}\text{O}_p$ and monthly mean precipitation is positive at the KK and MB sites in the south, but negative at the ND and KM stations in the north (**Fig. 3**).

On an inter-annual timescale, the results exhibit a relatively significant negative correlation between the weighted mean annual $\delta^{18}\text{O}_p$ (weighted by monthly

precipitation amount) and annual precipitation amount for the four stations in the ISM region (**Fig. 3**; [Li et al., 2015](#)). This correlation indicates that modern δD_p and $\delta^{18}O_p$ values in the ISM region are controlled by an inter-annual "amount effect"; however, the correlations between the weighted mean annual $\delta^{18}O_p$ and annual precipitation amount exhibit internal differences for the four stations in the EASM region (**Fig. 3**; [Li et al., 2015](#)). Fundamentally, the negative correlation between weighted mean annual $\delta^{18}O_p$ and annual precipitation amount is insignificant at the HK site, in southernmost China, which has the longest observation period and the largest quantity of observation data (**Table 1**). At the NJ and ZZ sites in the central EASM region, the negative correlation between weighted mean annual $\delta^{18}O_p$ and annual precipitation amount is also insignificant (however, due to the small amount of available data, the correlation at these two sites is not as reliable). At the SJZ site, in the northern part of the EASM region, the negative correlation between weighted mean annual $\delta^{18}O_p$ and annual precipitation amount is relatively insignificant (**Fig. 3**; [Li et al., 2015](#)). All of these results indicate no consistent inter-annual "amount effect" for modern δD_p and $\delta^{18}O_p$ values in the EASM region.

In our view the inter-annual correlations between modern $\delta^{18}O_p$ and precipitation amount are better reference points for the interpretation of regional paleo-records related to past δD_p and $\delta^{18}O_p$ variations. This is because the monthly isotopic signals of precipitation are likely to be significantly smoothed by the "buffering effect" of lake water (for lacustrine records) and soil water (for stalagmite records).

2.3. Comparison of lacustrine and marine δD_n and $\delta^{18}O_c$ and stalagmite $\delta^{18}O_s$ records

Subsequently, for convenience in discussion and description and for consistency with previous work (e.g. Shi *et al.*, 1994; Lu *et al.*, 2013), we use the following sub-divisions of the Holocene: early-Holocene - 11-8 ka B.P., mid-Holocene - 8-3 ka B.P. and late-Holocene - 3-0 ka B.P. In addition, the mid-Holocene is further divided into an early (8-6 ka B.P.) and late (6-3 ka B.P.) mid-Holocene.

In order to achieve a comprehensive comparison we compiled almost all of the available lacustrine and marine δD_n data from the Asian monsoonal region, including records from six lakes and one marine sediment core (SO188-342KL; Contreras-Rosales *et al.*, 2014) in the northern Bay of Bengal (**Fig. 1**). Detailed information about the seven sites is listed in **Table 2**. The temporal resolution of the data from Paru Co (Bird *et al.*, 2014) and Koucha Lake (Aichner *et al.*, 2010a) are relatively lower than those from the other sites. The sediment chronologies for Gonghai Lake (Chen *et al.*, 2015; Rao *et al.*, 2016a), Nam Co (Mügler *et al.*, 2010; Kasper *et al.*, 2015) and Qinghai Lake (An *et al.*, 2012; Thomas *et al.*, 2016) are highly reliable (**Table 2**); however, the chronology for Koucha Lake is relatively weak, based on only 4 AMS ^{14}C ages obtained from pollen extracts (Aichner *et al.*, 2010a).

For Paru Co (Bird *et al.*, 2014), Nam Co (Mügler *et al.*, 2010; Günther *et al.*, 2015) and Bengal Bay (Contreras-Rosales *et al.*, 2014), the δD_n data are from nC_{29} alkanes, a typical component of the leaf wax of terrestrial higher plants. For Qinghai Lake (Thomas *et al.*, 2016), the δD_n data are from nC_{28} alkanic acid, also a typical

component of the leaf wax of terrestrial higher plants. For Gonghai Lake (Rao *et al.*, 2016a), the δD_n data are the weighted average values of nC_{27} , nC_{29} and nC_{31} alkanes, all typical components of the leaf wax of terrestrial higher plants. All of these records are likely to reflect local δD_p variations in their catchments.

For Koucha Lake (Aichner *et al.*, 2010a), the δD_n data are weighted average values of nC_{23} and nC_{25} alkanes, typical components of aquatic plants, especially submerged and floating-leaved aquatics (Ficken *et al.*, 2000; Aichner *et al.*, 2010a), and therefore are indicative of the lake water δD . For Genggahai Lake (Rao *et al.*, 2014), the δD_n data are weighted average values of nC_{23} , nC_{25} , nC_{27} , nC_{29} and nC_{31} alkanes (Rao *et al.*, 2016b). Genggahai is a small, shallow lake fed principally by underground springs rather than by an obvious inflowing river (Qiang *et al.*, 2013). Based on comparison of the molecular distribution of *n*-alkanes extracted from sediment core samples from Genggahai and those of modern aquatic plants presently dominant in Genggahai Lake (Qiang *et al.*, 2013) reported from other lakes on the Tibetan Plateau (e.g. Aichner *et al.*, 2010b), it has been determined conclusively that both mid- (nC_{23} , nC_{25}) and long-chain (nC_{27} , nC_{29} and nC_{31}) alkanes in the Genggahai Lake sediments were dominantly sourced from aquatic plants, with the long-chain components partially derived from emergent plants (Rao *et al.*, 2016b). Therefore, the weighted average δD_n value of nC_{23} , nC_{25} , nC_{27} , nC_{29} and nC_{31} alkanes from Genggahai is also an indicator of the lake water δD . Monthly observational data, spanning an interval of two years, from spring water, river water and lake water in the Genggahai Lake catchment (Jin *et al.*, 2015), demonstrate that the lake water isotopes

of Genggahai Lake were mainly controlled by the precipitation isotopes in the catchment and evaporation from the lake surface. Further analysis indicates that since the last deglacial the Genggahai δD_n record is an indicator of local δD_p in the catchment, and that the short-term positive excursions evident in the record may reflect lake surface evaporation (Rao *et al.*, 2016b). Due to the close proximity of Genggahai Lake and Koucha Lake, we speculate that the weighted average values of nC_{23} and nC_{25} alkanes from Koucha Lake, as an indicator of the lake water δD , is also an indicator of the local δD_p on a long timescale.

The original data from Koucha Lake (Aichner *et al.*, 2010a), Paru Co (Bird *et al.*, 2014) and Qinghai Lake (Thomas *et al.*, 2016) were provided by the authors. In the case of the Bay of Bengal (Contreras-Rosales *et al.*, 2014) and Nam Co (Mügler *et al.*, 2010; Günther *et al.*, 2015), the relevant data were digitized from the figures in the original references. The Nam Co record is composed of two separately reported records of the intervals of ca. 23.7-7.5 ka B.P. (Günther *et al.*, 2015) and ca. 7.2-0 ka B.P. (Mügler *et al.*, 2010). There is an absence of data in the Nam Co record from ca. 3.9-1.5 ka B.P. (Mügler *et al.*, 2010), but the other δD_n records are continuous.

We collected $\delta^{18}O_s$ records from five caves (Fig. 1) for comparison with the lake and marine sediment records. Compared to the lake and marine sediment δD_n records, the temporal resolution of the $\delta^{18}O_s$ data is generally much higher. In addition, the chronologies of all of the $\delta^{18}O_s$ records are highly reliable (Table 2). The $\delta^{18}O_s$ record from Heshang Cave (Hu *et al.*, 2008) covers the past ca. 9.5 ka, and that from Qunf Cave (Fleitmann *et al.*, 2003, 2007) covers the past ca. 10.6 ka. The data for particular

intervals are lacking in the records from Qunf Cave (ca. 2.7-1.3 ka B.P., [Fleitmann et al., 2003](#)), Sanbao Cave (ca. 7.6-6.6 ka B.P., [Dong et al., 2010](#)) and Lianhua Cave (ca. 4.9-4.4 ka B.P., [Dong et al., 2015](#)); however, the records from Dongge Cave ([Dykoski et al., 2005](#)) and Heshang Cave ([Hu et al., 2008](#)) are continuous.

In addition to lake and marine sediment δD_n and stalagmite $\delta^{18}O_s$ records, lacustrine $\delta^{18}O_c$ records have been widely accepted as indicators of lake water $\delta^{18}O$ variations. A similar principle is involved as in the case of the recording of lake water δD variations by mid- and short-chain δD_n records derived from lacustrine aquatic plants and lower organisms. In 2011, [Zhang et al.](#) analyzed and compared ten Holocene lacustrine $\delta^{18}O_c$ records from the Asian monsoonal region ([Zhang et al., 2011](#)). Of these, eight lacustrine $\delta^{18}O_c$ records covered the entire Holocene, and the other two span part of the Holocene. These ten lakes are located in the monsoonal margins of the Tibetan Plateau, the northern Chinese Loess Plateau and southwestern China (**Fig. 1**). Five lakes are on the Tibetan Plateau: Selin Co ([Gu et al., 1993](#); [Morinaga et al., 1993](#); [Kashiwaya et al., 1995](#)) and Ahung Co ([Morrill, 2004](#); [Morrill et al., 2006](#)) on the central-southern Tibetan Plateau; and Koucha Lake ([Mischke et al., 2008](#)), Donggi Cona ([Mischke et al., 2010](#)) and Qinghai Lake ([Lister et al., 1991](#); [Liu et al., 2007](#)) are on the northeastern Tibetan Plateau. Three lakes lie on the northern margin of the Chinese Loess Plateau in northern China: Hamaertai Lake ([Wei et al., 1995](#)), Hetongchahan Nur ([Chen et al., 1995](#)) and Bahan Nur ([Guo et al., 2007](#)). The remaining two lakes in southwestern China are Qilu Lake and Xingyun Lake ([Hodell et al., 1999](#)). The eight normalized lacustrine $\delta^{18}O_c$ records covering the entire

Holocene cited directly in the present study are shown in **Fig. 4**. The methods of data processing and the chronology constraints of these records are detailed in the original reference (Zhang *et al.*, 2011).

To minimize the effects of short-term events, and therefore to highlight the long-term trends and facilitate a direct comparison with the normalized lacustrine $\delta^{18}\text{O}_c$ records, all of the selected lacustrine and marine δD_n and stalagmite $\delta^{18}\text{O}_s$ records were processed further. All of the records were smoothed with a running average using a 500-year window then the averaged data were rescaled to a range from 0 (most positive isotopic value) and 1 (most negative isotopic value). The final results are illustrated in **Fig. 4**. Correspondingly, the summer insolation (monthly mean values for June at 30 °N, Laskar *et al.*, 2004, 2011) was normalized for comparison with the isotopic records (**Fig. 4**).

The records in **Fig. 4** provide comprehensive coverage of the ISM and EASM regions with most of the lacustrine records mainly from the monsoonal margins and most of the stalagmite records from core monsoonal areas (**Fig. 1**). Considering the significant differences in e.g., altitude, lake area, watershed area, climatic conditions, and vegetation cover of the selected lakes, the records should have significant regional representativeness.

Although the lacustrine and marine δD_n records (**Fig. 4A**), stalagmite $\delta^{18}\text{O}_s$ records (**Fig. 4B**) and lacustrine $\delta^{18}\text{O}_c$ records (**Fig. 4C**) are geographically widespread (**Fig. 1**), they all exhibit significant and consistent long-term trends, *i.e.* overall they were all more negative during the early-Holocene and early

375 mid-Holocene (ca. 11-6 ka B.P.), and became more positive towards the late-Holocene
 376 (**Fig. 4**). There are, of course, several significant differences in the records, such as the
 377 positive excursions around ca. 8 ka B.P. in the Gonghai δD_n record and in the nearby
 378 Lianhua Cave $\delta^{18}O_s$ record (**Fig. 4**), which may be related to their northerly locations
 379 (**Fig. 1**). In the lacustrine and marine δD_n records (**Fig. 4A**), the early-Holocene
 380 abrupt positive trend of the Koucha Lake record, which makes this record appear
 381 atypical compared to the other records, may result from the lower data resolution and
 382 less reliable chronology. In the stalagmite $\delta^{18}O_s$ records (**Fig. 4B**), the early-Holocene
 383 negative trend of the Qunf Cave record appears to lag those of the other records, and
 384 may result from the fact that the record only spans the past ca. 10.6 ka ([Fleitmann *et al.*, 2003, 2007](#)). The detailed differences between the records may result from the
 385 following reasons: 1) small differences in the chronological constraints of the various
 386 geological archives; *e.g.* it is well-known that the chronological constraints of
 387 stalagmites are superior to those of lacustrine and marine sediments; 2) the different
 388 sampling and time resolutions of the records; *i.e.* some of the lacustrine and marine
 389 δD_n record resolutions are relatively lower (**Fig. 4**); 3) different isotopic effects
 390 produced by differences in the relative masses of hydrogen and oxygen isotopes
 391 causing greater changes in the amplitude of the lacustrine and marine δD_n records
 392 compared to the stalagmite $\delta^{18}O_s$ records (**Fig. 4**); and 4) secondary influences on the
 393 proxies from factors other than δD_p and $\delta^{18}O_p$ variations; *e.g.*, long-chain δD_n data
 394 derived from terrestrial higher plants are potentially influenced by C_3/C_4 relative
 395 abundances ([Chikaraishi and Naraoka, 2003](#); [Smith and Freeman, 2006](#)) and by
 396

different vegetative forms (grasses, shrubs and trees, Liu *et al.*, 2006).

Although potentially influenced by the aforementioned factors, the similarity of the long-term trends in all of the lake and marine δD_n , lake $\delta^{18}O_c$ and stalagmite $\delta^{18}O_s$ records unquestionably demonstrates that they are primarily driven by past δD_p and $\delta^{18}O_p$ variations, and that they are reliable indicators of the long-term evolution of δD_p and $\delta^{18}O_p$, respectively. As a whole, these lacustrine and marine δD_n , lacustrine $\delta^{18}O_c$ and stalagmite $\delta^{18}O_s$ records (**Fig. 4**) demonstrate that in both the ISM and EASM regions, and in both the marginal and core monsoonal regions, Holocene δD_p and $\delta^{18}O_p$ values exhibit similar long-term trends, *i.e.* they were overall more negative during the early-Holocene and early mid-Holocene (ca. 11-6 ka B.P.), and then became more positive towards the late-Holocene (**Fig. 4**).

2.4. Evolution of humidity in the ISM region

There are numerous paleoclimatic records from the ISM region and we selected four representative Holocene records which apparently reflect variations in humidity or precipitation amount in order to investigate the long-term trends of Holocene humidity evolution in the region. Because paleo-humidity and/or precipitation amounts in the ISM region are mainly controlled by summer monsoonal precipitation and/or intensity, the relevant paleo-records we selected reflect the evolution of ISM intensity during the Holocene. The locations of these records are shown in **Fig. 1**, and detailed site information is listed in **Table 2**. The records are briefly described as follows:

1) α -cellulose carbon isotopic ($\delta^{13}\text{C}$) records from Hongyuan peat bog in the northeastern Tibetan Plateau. The α -cellulose $\delta^{13}\text{C}$ records from this site, including α -cellulose $\delta^{13}\text{C}$ records from both bulk peat samples and single plant species (*Carex mulieensis*), are mainly influenced by ISM precipitation amount, with more negative α -cellulose $\delta^{13}\text{C}$ values occurring during periods of enhanced ISM precipitation (Hong *et al.*, 2003, 2005; Fig. 5A).

2) Lake level variations from Genggahai Lake on the northeastern Tibetan Plateau. This record is based on assemblages of aquatic plants and co-existent encrustations and mollusc shells. The lake is fed principally by underground spring water and there are no obvious inflow rivers, previous studies having demonstrated that the Genggahai lake levels were mainly controlled by ISM precipitation amount (Qiang *et al.*, 2013; Fig. 5B).

3) Lake level variations from Lake Ngangla Ring Tso on the southwestern TP. The record is based on numerous dating results from lacustrine paleo-banks (53 radiocarbon and 8 U/Th series ages derived from tufa and beach cement, respectively, and 19 ages from shells and aquatic plant material). The lake levels were also principally controlled by ISM precipitation amount (Hudson *et al.*, 2015; Fig. 5C).

4) Quantitative paleoprecipitation reconstruction results from Xingyun Lake in southwestern China. This record is based on lacustrine pollen assemblages (Chen *et al.*, 2014; Fig. 5D), and the area is traditionally considered to be influenced by ISM intensity (An *et al.*, 2011).

Except for the Xingyun record which only spans the past ca. 8.5 ka, the other

three records span the entire Holocene (**Fig. 5**). In general, the chronologies of the four records are reliable, especially for Hongyuan and Ngangla Ring Tso (**Table 2**). All of the records demonstrate that precipitation was greatest in the ISM region during the early-Holocene and early mid-Holocene (ca. 11-6 ka B.P.), when the highest lake levels, most negative α -cellulose $\delta^{13}\text{C}$ data and most humid climate occurred (**Fig. 5**).

The α -cellulose $\delta^{13}\text{C}$ records from Hongyuan peat bog ([Hong *et al.*, 2003, 2005](#); **Fig. 5A**), are supported by subsequent analysis of the degree of peat humification at the same site ([Wang H *et al.*, 2010](#)). Previously, a synthesis of a large number of paleoclimatic records from the ISM region also demonstrated that the most humid climate and strongest ISM intensity occurred during the early Holocene (defined as 10-7 ka B.P., [Wang Y *et al.*, 2010](#)) -- largely consistent with the representative paleo-humidity records we selected ([Chen *et al.*, 2014](#); [Qiang *et al.*, 2013](#); [Hong *et al.*, 2003, 2005](#); [Hudson *et al.*, 2015](#); **Fig. 5**). An enhanced ISM intensity (most humid climate) in the early-Holocene and early mid-Holocene (ca. 11-6 ka B.P.), becoming more arid towards the late-Holocene, has also been supported by recent results from Qinghai Lake ([Liu X *et al.*, 2014](#); [Liu *et al.*, 2016](#)) on the northeastern Tibetan Plateau, close to Genggahai Lake (**Fig. 1**), and from Lake Lonar ([Prasad *et al.*, 2014](#)) in central India.

Considering all of the aforementioned evidence related to Holocene paleo-humidity evolution in the ISM region, the long-term evolution history of paleo-humidity in the ISM region can be summarized as follows: the entire ISM region was overall humid during the early-Holocene and early mid-Holocene (ca. 11-6

ka B.P.), and then became more arid towards the late-Holocene, with no significant spatial differences evident in the entire ISM region. This indicates that the ISM intensity was enhanced during the early-Holocene and early mid-Holocene (ca. 11-6 ka B.P.), and then weakened towards the late Holocene.

2.5. Evolution of humidity in the EASM region

There are numerous Holocene paleoclimatic records from the EASM region; however, their spatial distribution is non-uniform, with most of the records coming from northern China, with only a few from central and southern China. Results from northern Chinese sandy lands and deserts (Yang *et al.*, 2003, 2008, 2010, 2011, 2013; Zhao *et al.*, 2007), and from lake sediments (Xiao *et al.*, 2009; Zhai *et al.*, 2011; Hao *et al.*, 2014), demonstrate that regional humid climate intervals occurred in the mid-Holocene (usually defined as 9-8 to 4-3 ka B.P.). These results are consistent with a recent synthesis of a series of regional paleoclimatic records which demonstrate that the most humid climate in north-central China occurred in the mid-Holocene (Wang H *et al.*, 2010).

In the present paper, we selected a recent quantitative annual mean precipitation reconstruction, based on high-resolution pollen assemblages from a well-dated sediment core from Gonghai Lake (Chen *et al.*, 2015; **Table 2**), to represent the Holocene paleo-humidity history in northern China. The record reveals a relatively humid mid-Holocene (ca. 8-3 ka B.P., **Fig. 6A**) in northern China. This is also supported by magnetic susceptibility and organic $\delta^{13}\text{C}$ data from the Luochuan

(35°43'N, 109°25'E; **Fig. 1**), Yulin (38°20'N, 109°44'E; **Fig. 1**), and Xunyi (35°02'N, 108°13'E) sections in the central Chinese Loess Plateau (Lu *et al.*, 2013); by a synthesis of multiple dating and stratigraphic identification results from the Chinese Loess Plateau and the four major sandy lands (the Mu Us, Otindag, Horqin and Hulun Buir sandy lands) located on the modern monsoonal margins (Wang *et al.*, 2014); and by desert distribution and vegetation cover reconstructions in northern China (Li *et al.*, 2014). All of these results demonstrate that the basic Holocene paleo-humidity history in northern China consists of, as represented by the Gonghai record (**Fig. 6A**), a relatively arid early-Holocene (ca. 11-8 ka B.P.) and late Holocene (ca. 3-0 ka B.P.) and a relatively humid mid-Holocene (ca. 8-3 ka B.P.).

Holocene paleoclimatic records from the middle and lower reaches of the Yangtze River in central China are relatively rare. We selected results from the Dajiuhu peat bog and stalagmite from Heshang Cave (**Fig. 1**), from the middle reaches of the Yangtze River, to represent the Holocene paleo-humidity evolution of the area (Xie *et al.*, 2013). The mass accumulation rates of hopanoids from Dajiuhu peat bog were adopted as indicators of water level variations in the peatland. As biomarkers for aerobic microbes, hopanoid mass accumulation rates decrease with higher water levels, and *vice versa* (Pancost *et al.*, 2003; Jaatinen *et al.*, 2007). Similarly, a magnetic proxy (ARM/SIRM – the ratio of anhysteretic remanent magnetization to saturation isothermal remanent magnetization) from stalagmites from nearby Heshang Cave was used to represent the ratio of fine magnetic particles to total ferrimagnetic particles. With increased precipitation, more coarse magnetic particles infiltrate the

cave, resulting in a fall in the ARM/SIRM ratio, and *vice versa*. The stalagmite ARM/SIRM record from Heshang Cave therefore also reflects local humidity/precipitation variations (Xie *et al.*, 2013). The Dajiuhu peat bog (Fig. 6B) and Heshang stalagmite (Fig. 6C), in the middle reaches of the Yangtze River, demonstrate that precipitation decreased and the climate became more arid during the interval from ca. 7-3 ka B.P., and that precipitation increased and the climate became more humid during the other intervals of the Holocene (ca. 11-7 ka B.P. and ca. 3-1 ka B.P.; Xie *et al.*, 2013).

Holocene paleoclimatic records from southern China are relatively rare. We selected a compound-specific carbon isotopic record, consisting of long-chain (nC_{29} , nC_{31}) n -alkanes derived from terrestrial higher plants ($\delta^{13}C_{n\text{-alkanes}}$), from the sediments of Huguangyan Lake (a maar lake and the most southerly site, closest to the South China Sea; Fig. 1), to represent local paleo-humidity evolution (Jia *et al.*, 2015). Due to Huguangyan's low latitude location and relatively high local temperatures, any greater relative abundance of C_4 plants indicated by more positive long-chain $\delta^{13}C_{n\text{-alkanes}}$ values implies a generally more arid local climate, and *vice versa*. More importantly, differences between nC_{31} and nC_{29} $\delta^{13}C_{n\text{-alkanes}}$ data ($\Delta\delta C_{31-29}$, i.e., $\delta^{13}C_{31}-\delta^{13}C_{29}$) in the Huguangyan sediments would probably be a more sensitive indicator of local humidity, because: 1) independent and separate $\delta^{13}C_{n\text{-alkanes}}$ of long-chain (nC_{29} , nC_{31}) n -alkanes are primarily an indicator of the relative abundance of C_3/C_4 plants, and therefore may not only be influenced by relative humidity (Rao *et al.*, 2012); and 2) observational evidence from modern higher plants demonstrates that

$\Delta\delta_{31-29}$ data are, on average, more negative in evergreen plants in tropical-south
subtropical areas ($-0.9 \pm 1.3\text{‰}$) than in deciduous plants in northern
subtropical-temperate areas ($+0.2 \pm 1.0\text{‰}$) (Jia *et al.*, 2015). The Huguangyan
 $\Delta\delta C_{31-29}$ record (**Fig. 6D**) demonstrates that the local mid-Holocene (ca. 8-3 ka B.P.)
climate was relatively humid (Jia *et al.*, 2015), and this finding is supported by the
variation of the percentages of *Artemisia* from the same lake (Wang *et al.*, 2007).

The aforementioned results from northern, central and southern China,
especially the selected representative records (**Fig. 6**) all of which have highly reliable
age control (**Table 2**), are significantly different to the results from the ISM region,
clearly demonstrating the occurrence of spatial differences in Holocene
paleo-humidity evolution in the EASM region. Taken as a whole, the arid
paleoclimatic intervals in southern and northern China occurred in the early-Holocene
(ca. 11-8 ka B.P.) and late-Holocene (ca. 3-0 ka B.P.), but in the major part (ca. 7-3 ka
B.P.) of the mid-Holocene in central China (**Fig. 6**). Thus, from southern to central
China, and then to northern China, climatic humidity exhibited a spatial "-+-" mode
during the early-Holocene and late-Holocene, but a spatial "+-+" mode during the
mid-Holocene (**Fig. 6**). Indeed, the arid interval of ca. 7-3 ka B.P. in the middle
reaches of the Yangtze River does not correspond precisely to the humid interval of ca.
8-3 ka B.P. in northern and southern China. This may result from differences in the
climatic sensitivity and representativeness of the locations where the records came
from, because previously reported results have demonstrated that highly consistent
co-variation of precipitation amount in Eastern Asia only occur within a distance of

500 km (Dayem *et al.*, 2010). In summary, all of the results indicate a more complex pattern of evolution and, therefore, a more complex series of mechanisms responsible for driving EASM intensity, compared to ISM intensity.

3. Discussion

3.1. The relationship between δD_p , $\delta^{18}O_p$ and humidity in the ISM and EASM regions during the Holocene

The aforementioned comparisons demonstrate a consistent long-term evolutionary trend in δD_p and $\delta^{18}O_p$ values in both the ISM and EASM regions, as well as in the marginal and core monsoonal regions, *i.e.* a more negative early-Holocene and early mid-Holocene interval (ca. 11-6 ka B.P.), with values becoming more positive towards the late-Holocene (**Fig. 4**). Nonetheless, in the Asian monsoonal region, monthly correlations between $\delta^{18}O_p$ and precipitation amount cannot be adopted as a reference point for the interpretation of the relevant paleo-records. Essentially, this is because of the "reservoir effect" of both lake water (as the direct isotopic source for lacustrine δD_n and $\delta^{18}O_c$) and soil water (as the direct isotopic source for stalagmite $\delta^{18}O_s$). The isotopic signals in both lake and soil water have undergone relatively long-term buffering, meaning that the monthly isotopic signals of lake and soil water are smoothed. Recent observational isotopic data from local precipitation and cave dripwater at two sites in southern France clearly demonstrate that dripwater isotopes exhibited no significant variation over the entire observation period of more than 10 years (Genty *et al.*, 2014). This study

demonstrates that variations in monthly $\delta^{18}\text{O}_p$ values are largely smoothed out in soil water, and this is also likely to be the case for lake water. In addition, in the Asian monsoonal region, the consistent long-term evolution of Holocene δD_p and $\delta^{18}\text{O}_p$ (**Fig. 4**), allied with the spatial differences in monthly correlations between modern $\delta^{18}\text{O}_p$ and precipitation amount in both ISM and EASM regions (**Fig. 3**), unquestionably demonstrates that monthly correlations between $\delta^{18}\text{O}_p$ and precipitation amount cannot be adopted as reference points for the interpretation of regional Holocene paleo-records that relate to δD_p and $\delta^{18}\text{O}_p$ variations.

On an inter-annual timescale, the relatively significant negative correlations between the weighted mean annual $\delta^{18}\text{O}_p$ and annual precipitation amount for the four stations in the ISM region (**Fig. 3**; [Li et al., 2015](#)) indicate that modern δD_p and $\delta^{18}\text{O}_p$ values in the ISM region are controlled by an inter-annual "amount effect". This is consistent with the aforementioned δD_p and $\delta^{18}\text{O}_p$ values being overall more negative (**Fig. 4**) during the humid early-Holocene and early mid-Holocene (**Fig. 5**) in the ISM region. In other words, both modern inter-annual and ISM Holocene orbital δD_p and $\delta^{18}\text{O}_p$ values are controlled by this "amount effect". Thus, the paleo-records that record δD_p and $\delta^{18}\text{O}_p$ variations in the ISM region are strongly indicative of paleo-humidity, *i.e.* they can be adopted as a humidity indicator for paleoclimatic reconstruction.

The correlations between the weighted mean annual $\delta^{18}\text{O}_p$ and annual precipitation amounts for the four stations in the EASM region (**Fig. 3**; [Li et al., 2015](#)) exhibit significant spatial differences, indicating no consistent inter-annual "amount

effect” in the EASM region. These inter-annual correlations are consistent with the
aforementioned δD_p and $\delta^{18}O_p$ values being overall more negative (**Fig. 4**) during the
early-Holocene and early mid-Holocene interval (ca. 11-6 ka B.P.), although there are
obvious spatial differences in paleo-humidity in the EASM region (**Fig. 6**). Therefore,
on both modern inter-annual and Holocene orbital timescales, δD_p and $\delta^{18}O_p$ values in
the EASM region exhibit a complex relationship with humidity. In other words, the
paleo-records related to δD_p and $\delta^{18}O_p$ variations in the EASM have no significance in
terms of paleo-humidity and they cannot be used for paleoclimatic reconstruction
directly.

Nonetheless, previously-reported model results for paleoclimate and
precipitation isotopes ([Liu Z *et al.*, 2014](#)) have supported the view that stalagmite
 $\delta^{18}O_s$ records from southern China can indicate EASM intensity. The climate
simulation results for summer precipitation exhibit a decrease from early (11-9 ka BP)
to late Holocene in northern China, but an increase from early to late Holocene in
southern China ([Liu Z *et al.*, 2014](#)), seemingly in contrast to the aforementioned
geological records from both northern and southern China (**Fig. 6**). Recent studies,
also using model simulations, do not support the conclusion that stalagmite $\delta^{18}O_s$
records are indicators of summer monsoon intensity ([Caley *et al.*, 2014](#)). The control
mechanisms governing δD_p and $\delta^{18}O_p$ variations in the EASM region therefore remain
unclear and further study is required.

3.2. Enhanced EASM intensity during the mid-Holocene

Our comparisons clearly demonstrate spatial differences in Holocene paleo-humidity variations in the EASM region, *i.e.* from southern to central China, and then to northern China, during the early-Holocene (ca. 11-8 ka B.P.) and late-Holocene (ca. 3-0 ka B.P.), represented by a "-+-" spatial mode; however, a "+-+" mode occurred during the mid-Holocene (ca. 8-3 ka B.P.; **Fig. 6**). If the spatial distribution of summer precipitation and corresponding EASM intensity for the modern intervals of 1951-1978 and 1979-1992 (**Fig. 2**; [Ding *et al.*, 2008](#)) are taken as analogues, EASM intensity would have been enhanced during the mid-Holocene, but would have decreased during the early-Holocene and late-Holocene (**Fig. 2**). Interestingly, results from Paleoclimate Modeling Intercomparison Project (PMIP) simulations have demonstrated that EASM intensity strengthened during the mid-Holocene (6 ka B.P. time slice; [Jiang *et al.*, 2013a](#)). PMIP simulations also demonstrate that, as a whole, annual precipitation and humidity increased in the EASM region during the mid-Holocene ([Jiang *et al.*, 2013b](#)). More importantly, within the EASM region during the mid-Holocene, southern and northern China became more humid, but the areas between the middle and lower reaches of the Yangtze and Yellow River valleys became more arid ([Jiang *et al.*, 2013b](#)) – entirely consistent with the modern climatic characteristics ([Ding *et al.*, 2008](#)) and with Holocene paleo-humidity spatial variations in the EASM region (**Fig. 6**). This not only demonstrates that regional EASM precipitation varied spatially in response to differences in EASM intensity, but more importantly it further demonstrates that EASM intensity was stronger during the mid-Holocene, and weaker during the

early-Holocene and late-Holocene.

Northern Hemisphere summer insolation (represented by the monthly averaged insolation in June at 30 °N, [Laskar *et al.*, 2004, 2011](#)) exhibits a long-term decreasing trend from the early-Holocene to the late-Holocene (**Figs. 4, 5, 6**). Therefore, the enhanced ISM intensity during the early-Holocene and early mid-Holocene (ca. 11-6 ka B.P.), with limited phase lag relative to summer insolation, responded more rapidly to summer insolation (**Fig. 5**). However, the enhanced EASM intensity during the mid-Holocene (ca. 8-3 ka B.P.), which exhibited an apparent phase lag to summer insolation, responded more slowly to summer insolation (**Fig. 6**). All of these results indicate the occurrence of more complex control mechanisms of EASM intensity.

It has been recognized that when the west-east Equatorial Pacific thermal gradient steepens (La-Niña status), the West Pacific subtropical high (WPSH) weakens and withdraws northeastwards. Conversely, when the west-east Equatorial Pacific thermal gradient falls (El-Niño status), the WPSH strengthens and extends southwestwards ([Tan, 2009, 2011](#)). Normally, the belt of maximum rainfall in the EASM region is located ca. 5-10 ° to the north of the WPSH. Therefore, we can reasonably speculate that coupled variations in the west-east Equatorial Pacific thermal gradient and the WPSH potentially modulate EASM intensity variations, and thus the spatial distribution of precipitation in eastern China. When the West Pacific Warm Pool (WPWP) is warmer, more water vapor is generated from the warmer ocean surface, and subsequently a weakened and northeastwardly-retreating WPSH, acting as a huge "pump", draws water vapor from the WPWP and transports it further

north, bringing plentiful rainfall to northern China. When the WPWP is cooler, less water vapor is generated from the cooler ocean surface, and a strengthened and southwestwardly-extending WPSH blocks the transportation of water vapor to northern China, resulting in the maximum rainfall belt covering the middle and lower reaches of the Yangtze River (**Fig. 2**).

Mean SST records obtained from seven marine sediment cores in the WPWP and, from two marine sediment cores in the eastern Equatorial Pacific, clearly highlight an enhanced west-east Equatorial Pacific thermal gradient during the mid-Holocene (ca. 8-3 ka B.P., **Fig. 6**; [Koutavas and Joanides, 2012](#)), indicating an overall La-Niña status roughly corresponding to the enhanced mid-Holocene EASM intensity (**Fig. 6**). This suggests that the west-east Equatorial Pacific thermal gradient probably played an important role in modulating variations in Holocene EASM intensity. Paleoclimate simulations suggest that the intensified mid-Holocene EASM was caused by an enhanced land-sea thermal contrast and sea level pressure gradient during that interval ([Jiang et al., 2013a](#)). New paleoclimate simulations also suggest that internal forcing, such as that caused by Equatorial Pacific SSTs, is critical in driving monsoonal precipitation variations in the Asian monsoonal region ([Caley et al., 2014](#)).

The Holocene paleoclimate results from Huguangyan Lake in southern China ([Jia et al., 2015](#)) are largely synchronous with stalagmite $\delta^{18}\text{O}_s$ records ([Denniston et al., 2013](#)) from northern Australia (also located west of the WPWP, but in the Southern Hemisphere). Both regions were humid during the mid-Holocene. New

lacustrine Holocene paleoclimate results from the Northern and Southern Hemisphere
 outer-tropical Andes are also largely synchronous: both regions were arid during the
 mid-Holocene (Polissar *et al.*, 2013). An arid mid-Holocene is also supported by
 studies from Hawaii in the eastern Equatorial Pacific (Uchikawa *et al.*, 2010). It
 seems that Holocene paleo-humidity variations do not exhibit a south-north "seesaw
 effect" in the western and eastern margins of the Pacific, but rather a west-east
 "seesaw effect" (Jia *et al.*, 2015). Newly-reconstructed mean annual SSTs based on
 fossil mollusk $\delta^{18}\text{O}$ values from the Peruvian coast (Carré *et al.*, 2014) indicate a
 colder mid-Holocene interval. These results not only demonstrate that an abnormal
 mid-Holocene west-east Equatorial Pacific thermal gradient existed, but also that this
 abnormal thermal gradient profoundly affected the Holocene climate in both the
 western and eastern sides of Pacific, including the EASM intensity.

4. Conclusions

In this paper, we have made a comprehensive comparison of representative
 records of Holocene δD_p , $\delta^{18}\text{O}_p$ and humidity variations in the Asian monsoonal
 region. In combination with modern $\delta^{18}\text{O}_p$ data and the observed spatial distribution
 of precipitation in eastern China during periods of differing EASM intensity, we draw
 the following conclusions:

- 1). In both the ISM and EASM regions, and in both marginal and core monsoonal
 regions, δD_p and $\delta^{18}\text{O}_p$ values were overall more negative during the
 early-Holocene and early mid-Holocene (ca. 11-6 ka B.P.), and then became more

positive towards the late-Holocene.

- 2). In the ISM region the climate was humid during the early-Holocene and early mid-Holocene (ca. 11-6 ka B.P.), and then became more arid towards the late-Holocene, without apparent spatial differences. This indicates enhanced ISM intensity during the early-Holocene and early mid-Holocene. In the EASM region, the climate was humid during the mid-Holocene (ca. 8-3 ka B.P.) in southern and northern China, but humid during the early-Holocene and late-Holocene in central China. This indicates enhanced EASM intensity during the mid-Holocene.
- 3). In the ISM region, on both modern inter-annual and Holocene orbital timescales, δD_p and $\delta^{18}O_p$ values are controlled by a precipitation "amount effect", indicating that paleo-records related to δD_p and $\delta^{18}O_p$ variations reflect paleo-humidity in the ISM region. In the EASM region, on both modern inter-annual and Holocene orbital timescales, δD_p and $\delta^{18}O_p$ values are not controlled by an "amount effect", suggesting that paleo-records related to δD_p and $\delta^{18}O_p$ variations in the EASM region are not paleo-humidity indicators. Further studies of the mechanisms governing δD_p and $\delta^{18}O_p$ in the EASM region are required.
- 4). Relatively speaking, ISM intensity responds more quickly to summer insolation than EASM intensity, indicating more complex control mechanisms for the latter. Coupled variations in the west-east Equatorial Pacific thermal gradient and the WPSH potentially played an important role in modulating EASM intensity.

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Figure and table captions

Fig. 1. Locations of the sites cited in the text. Green dots are lake sites: NRT, Ngangla

Ring Tso; SL, Selin Co; NC, Nam Co; AC, Ahung Co; PC, Paru Co; KC, Koucha Lake; DC, Donggi Cona; QH, Qinghai Lake; GGH, Genggahai Lake; HM, Hamaertai Lake; HT, Hetongchahan Nur; BN, Bahan Nur; GH, Gonghai Lake; XY, Xingyun Lake; QL, Qilu Lake; HGY, Huguangyan Lake. Black squares are peat bog sites: HY, Hongyuan; DJH, Dajiuhu. Purple triangles are loess profiles: YL, Yulin; LC, Luochuan. Red pentangles are cave sites: QF, Qunf; DG, Dongge; HS, Heshang; SB, Sanbao; LH, Linahua. Blue diamonds are modern meteorological stations: KK, Kozhikode; MB, Mumbai; ND, New Delhi; KM, Kunming; HK, Hongkong; NJ, Nanjing; ZZ, Zhengzhou; SJZ, Shijiazhuang. The blue dot is the marine sediment core from the Bay of Bengal (BB). The bold black line indicates the modern northern limit of the Asian summer monsoon ([Chen et al., 2008](#)). The dashed red line is the boundary between the ISM and EASM regions, and the blue arrows represent the approximate trajectories of the ISM and EASM.

Fig. 2. Patterns of summer (JJA) rainfall (mm) averaged from 1951-1978 (A) and for 1979-1992 (B) relative to the 1971–2000 climatological mean based on observational data from 740 surface stations in China. Solid lines (dashed lines) denote positive (negative) percentages from the mean. Shaded areas denote positive departures. A and B are modified from [Ding et al. \(2008\)](#). C, the overall La-Niña status and humid climate in the western margins of the Pacific; D, the overall El-Niño status and humid climate in the margins of the Pacific. This figure demonstrates the important role of a coupled Equatorial Pacific

thermal gradient and the West Pacific subtropical high in the EASM intensity and corresponding rainfall patterns in eastern China.

Fig. 3. Linear correlations between precipitation amount and precipitation $\delta^{18}\text{O}$ on both monthly and inter-annual timescales (original data were downloaded from the website of the GNIP) for four meteorological stations in the ISM region and four stations in the EASM region (Li *et al.*, 2015). The locations of the stations are shown in Fig. 1 (codes identical as to Table 2).

Fig. 4. Comparison of selected and normalized lacustrine and marine δD_n records (A), stalagmite $\delta^{18}\text{O}_s$ records (B) and lacustrine $\delta^{18}\text{O}_c$ records (C). The lacustrine $\delta^{18}\text{O}_c$ records (C) are directly cited from Zhang *et al.* (2011). The grey bar represents the interval of the early-Holocene and early mid-Holocene (ca. 11-6 ka B.P.) with the most negative isotopic values. The grey arrow represents the positive trend of isotopic variation. For further comparison, normalized summer insolation (Laskar *et al.*, 2004, 2011) is shown as a solid red line.

Fig. 5. Comparison between Northern Hemisphere summer insolation (solid red line, Laskar *et al.*, 2004, 2011) and selected typical Holocene humidity records from the ISM region. A, α -cellulose $\delta^{13}\text{C}$ records of bulk samples and *Carex mulieensis* from Hongyuan peat bog (Hong *et al.*, 2003, 2005); B, lake level record from Genggahai Lake (Qiang *et al.*, 2013); C, lake level record from Ngangla Ring Tso (Hudson *et al.*, 2015); D, paleo-rainfall record based on pollen assemblages from Xingyun Lake (Chen *et al.*, 2014). All of these records exhibit a humid interval in the early-Holocene and early mid-Holocene

1210 (ca. 11-6 ka B.P.) in the ISM region that corresponds to enhanced summer
 1211 insolation (grey vertical bar), and an arid trend towards the late-Holocene that
 1212 corresponds to decreased summer insolation (grey arrows).

1213 **Fig. 6.** Comparison of selected representative Holocene humidity records from the
 1214 EASM region: A, paleoprecipitation record based on pollen assemblages from
 1215 Gonghai Lake ([Chen et al., 2015](#)); B, hopanoid mass accumulation rates from
 1216 Dajiuhu peat bog ([Xie et al., 2013](#)); C, stalagmite ARM/SIRM record from
 1217 Heshang Cave ([Xie et al., 2013](#)); D, $\Delta\delta C_{31-29}$ record from Huguangyan Lake
 1218 ([Jia et al., 2015](#)); and west-east Equatorial Pacific thermal gradient record (E,
 1219 [Koutavas and Joanides, 2012](#)) and Northern Hemisphere summer insolation
 1220 record (F, [Laskar et al., 2004, 2011](#)). The blue bars represent the humid
 1221 interval of the mid-Holocene (ca. 8-3 ka B.P.) in northern and southern China.
 1222 The yellow bar indicates the arid interval of ca. 7-3 ka B.P. in central China.
 1223 The green bar indicates the overall La-Niña status during the mid-Holocene
 1224 (ca. 8-3 ka B.P.).

1225 **Table 1** Detailed information for the eight meteorological stations.

1226 **Table 2** Detailed information for the isotopic and paleo-humidity study sites cited in
 1227 the text.

1228

1229 **Figures and Tables (applied separately):**

Table 1

Site	Code	Location	Altitude (m asl)	MAT (°C)	MAP (mm)	Source years for $\delta^{18}\text{O}_\text{p}$ data
Kozhikode	KK	11°15'N, 75°46'E	20	27.96	3041.7	1997-2007
Mumbai	MB	18°54'N, 72°49'E	10	27.33	1984.11	1961-1966,1973-1977
New Delhi	ND	28°34'N, 77°12'E	212	25.29	771.42	1962-1967,1969-1988,1991-1995,2001-2005,2007-2008
Kunming	KM	25°01'N, 102°41'E	1892	15.52	993.33	1986-1992,1996-2003
Hongkong	HK	22°19'N, 114°10'E	66	22.98	2290.97	1961-1964,1974-1980, 1982-2009
Nanjing	NJ	32°10'N, 118°10'E	26	14.33	1176	1988-1992
Zhengzhou	ZZ	34°43'N, 113°39'E	110	13.57	490.38	1986-1987,1989-1991
Shijiazhuang	SJZ	38°02'N, 114°25'E	80	13.99	553.7	1986-1991,1996-2000, 2002-2003

Table 2

Table 2

Site	Code	Type	Location	Altitude (m asl)	Details about proxy	Reference(s)	Details about age control	Reference(s)
Lacustrine and marine δD as indicator of δD_p								
Gonghai Lake	GH	lake sediments	38°54'N, 112°14'E	1860	weighted mean values of nC_{27} , nC_{29} and nC_{31} alkanes	Rao <i>et al.</i> , 2016a	25 AMS ^{14}C ages (terrestrial plant macrofossils) with 20 < 12 ka	Chen <i>et al.</i> 2015
Genggahai Lake	GGH	lake sediments	36°11'N, 100°06'E	2860	weighted mean values of nC_{23} , nC_{25} , nC_{27} , nC_{29} and nC_{31} alkanes	Rao <i>et al.</i> , 2014, 2016b	12 AMS ^{14}C ages (aquatic macrophyte remains) with 10 < 12 ka	Qiang <i>et al.</i> , 2013;
Koucha Lake	KC	lake sediments	34.0°N, 97.2°E	4540	weighted mean values of nC_{23} and nC_{25} alkanes	Aichner <i>et al.</i> , 2010	4 AMS ^{14}C ages (pollen extracts)	Aichner <i>et al.</i> , 2010
Paru Co	PC	lake sediments	29.78°N, 92.35°E	4845	values of nC_{29} alkanes	Bird <i>et al.</i> , 2014	7 AMS ^{14}C ages (6 carbonized grass fragments and 1 oogonia)	Bird <i>et al.</i> , 2014
Nam Co	NC	lake sediments	30°33'N, 90°50'E	4722	values of nC_{29} alkanes (with hiatus of ca. 1.5-3.9 ka)	Mügler <i>et al.</i> , 2010; Günther <i>et al.</i> , 2015	21 AMS ^{14}C ages (bulk sediment samples)	Mügler <i>et al.</i> , 2010; Kasper <i>et al.</i> , 2015
Qinghai Lake	QH	lake sediments	37°18'N, 99°00'E	3194	values of nC_{28} alkanolic acid	Thomas <i>et al.</i> , 2016	38 AMS ^{14}C ages (33 TOC from bulk sediments and 5 <i>Ruppiaceae</i> seeds)	An <i>et al.</i> , 2012
Bengal Bay	BB	marine sediments	19°58'N, 90°2'E	0	values of nC_{29} alkanes	Contreras-Rosales <i>et al.</i> , 2014	9 AMS ^{14}C ages (mixed planktic framiniferal species) with 5 < 12 ka	Contreras-Rosales <i>et al.</i> , 2014
Stalagmite $\delta^{18}O$ as indicator of $\delta^{18}O_p$								
Lianhua Cave	LH	stalagmite	38°10'N,	1200	$\delta^{18}O_s$ of samples LH4, LH5	Dong <i>et al.</i> , 2015	27 U/Th ages (15 from LH4, 8	Dong <i>et al.</i> , 2015

			113°43'E		and LH9 (with hiatus of ca. 4.4-4.9 ka)		from LH5, 4 from LH9)	
Heshang Cave	HS	stalagmite	30°27'N, 110°25'E	294	$\delta^{18}\text{O}_s$ of sample HS-4	Hu <i>et al.</i> , 2008	21 U/Th ages	Hu <i>et al.</i> , 2008
Sanbao Cave	SB	stalagmite	31°40'N, 110°26'E	1900	$\delta^{18}\text{O}_s$ of sample SB43 (with hiatus of ca. 6.6-7.6 ka)	Dong <i>et al.</i> , 2010	24 U/Th ages	Dong <i>et al.</i> , 2010
Dongge Cave	DG	stalagmite	25°17'N, 108°5'E	680	$\delta^{18}\text{O}_s$ of sample D4	Dykoski <i>et al.</i> , 2005	45 U/Th ages with 35 < 12 ka	Dykoski <i>et al.</i> , 2005
Qunf Cave	QF	stalagmite	17°10'N, 54°18'E	650	$\delta^{18}\text{O}_s$ of sample Q5 (hiatus of ca. 1.3-2.7 ka)	Fleitmann <i>et al.</i> , 2003	18 U/Th ages	Fleitmann <i>et al.</i> , 2003
Typical paleo-humidity records from ISM region								
Hongyuan	HY	peat bog	32°46'N, 102°30'E	3466	Peat cellulose $\delta^{13}\text{C}$ as indicator of humidity	Hong <i>et al.</i> , 2003	15 AMS ^{14}C ages (peat plant celluloses)	Hong <i>et al.</i> , 2003
Genggahai Lake	GGH	lake sediments	36°11'N, 100°06'E	2860	lake-level reconstruction based on macrofossil assemblages	Qiang <i>et al.</i> , 2013	12 AMS ^{14}C ages (aquatic macrophyte remains) with 10 < 12 ka	Qiang <i>et al.</i> , 2013
Ngangla Ring Tso	NRT	lake shorelines	31.6°N, 83.1°E	4800	Lake-level reconstruction based on shoreline dates	Hudson <i>et al.</i> , 2015	53 AMS ^{14}C and 8 U/Th ages (tufa and beach cement), 19 AMS ^{14}C ages (shell and aquatic plant material)	Hudson <i>et al.</i> , 2015
Xingyun Lake	XY	lake sediments	24°20'N, 102°47'E	1772	quantitative precipitation reconstruction based on pollen assemblages	Chen <i>et al.</i> , 2014	8 AMS ^{14}C ages (plant macrofossils and charcoal)	Chen <i>et al.</i> , 2014
Typical paleo-humidity records from EASM region								
Gonghai Lake	GH	lake sediments	38°54'N,	1860	quantitative precipitation	Chen <i>et al.</i> , 2015	25 AMS ^{14}C ages (terrestrial	Chen <i>et al.</i> , 2015

			112°14'E		reconstruction based on pollen assemblages		plant macrofossils) with 20 < 12 ka	
Dajiuahu	DJH	peat land	31°28'N, 110°00'E	1700	hopanoid accumulation rate as indicator of humidity	Xie <i>et al.</i> , 2013	14 AMS ¹⁴ C ages (bulk peat samples) with 12 < 12 ka	Xie <i>et al.</i> , 2013
Heshang Cave	HS	stalagmite	30°27'N, 110°25'E	294	stalagmite ARM/SIRM as indicator of precipitation amount	Xie <i>et al.</i> , 2013	21 U/Th ages	Hu <i>et al.</i> , 2008
Huguangyan Lake	HGY	maar lake	21°9'N, 110°17'E	23	difference between nC_{31} and nC_{29} $\delta^{13}C_{n\text{-alkanes}}$ as indicator of humidity	Jia <i>et al.</i> , 2015	13 AMS ¹⁴ C ages (5 bulk organic and 8 plant debris)	Jia <i>et al.</i> , 2015

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6