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Holocene moisture and East Asian summer monsoon evolution in the northeastern Tibetan Plateau recorded by Lake Qinghai and its environs: A review of conflicting proxies

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ABSTRACT

Climatic and environmental changes in the northeastern Tibetan Plateau are controlled by the Asian summer monsoon (ASM) and the westerlies, two key circulation components of the global climate system which directly affect a large human population and associated ecosystems in eastern Asia. During the past few decades, a series of Holocene palaeoclimatic records have been obtained from sediment cores from Lake Qinghai and from various other geological archives in the surrounding area of the northeastern Tibetan Plateau. However, because of uncertainties regarding the sediment chronologies and the climatic significance of the proxies used, the nature of Holocene climatic changes in the region remains unclear and even controversial. Here we review all major classes of the published data from drilled cores from Lake Qinghai, as well as other evidence from lakes and aeolian deposits from surrounding areas, in order to reconstruct changes in moisture patterns and possible summer monsoon evolution in the area during the Holocene. Combining the results of moisture and precipitation proxies such as vegetation history, pollen-based precipitation reconstruction, aeolian activity, lake water depth/lake level changes, salinity and sediment redness, we conclude that moisture and precipitation began to increase in the early Holocene, reached their maximum during the middle Holocene, and decreased during the late Holocene - similar to the pattern of the East Asian summer monsoon (EASM) in northern China. It is clear that the region experienced a relatively dry climate and weak EASM during the early Holocene, as indicated by relatively low tree pollen percentages and fluctuating pollen concentrations; generally low lake levels of Lake Qinghai and the adjacent Lake Hurleg and Lake Toson in the Qaidam Basin; and widely distributed aeolian sand deposition in the Lake Qinghai Basin and the nearby Gonghe Basin to the south, and in the eastern Qaidam Basin to the west. We argue that the ostracod $\delta^{18}O$ record, which is widely used as a proxy of effective moisture and summer monsoon intensity in lake sediments, at least in Lake Qinghai, and which exhibits light values in the early Holocene and heavier values thereafter, cannot be used to reflect the strength of the EASM or the intensity of monsoon precipitation - as is also the case for leaf wax $\delta^{2}H$ records. Rather, we argue that as is the case of the Chinese speleothem $\delta^{18}O$ record, which also is often interpreted as an EASM proxy, it reflects variation in the $\delta^{18}O$ of precipitation. Overall, we suggest that the EASM significantly affected precipitation in the northeastern Tibetan Plateau during the Holocene; and that it increased in

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strength during the early Holocene, reached a maximum during the middle Holocene and decreased during the late Holocene.

1. Introduction

Understanding the evolution of the Asian summer monsoon (ASM) is an important issue in earth science (Kutzbach, 1981; An et al., 1991; An, 2000; Ding et al., 2002; Yuan et al., 2004; Kutzbach et al., 2008), especially for the present interglacial. Moreover, changes in the ASM have significant economic and societal implications for the regions within its influence through changes in water resource availability. There is evidence that weak summer monsoon events during the Holocene significantly impacted Neolithic culture (An et al., 2005; Dong et al., 2012) and the later development of civilization in China (Zhang et al., 2008). Currently, variability of the ASM, especially extreme variations, can cause crop failures and flooding which affect almost two-thirds of the world's population (Webster et al., 1998). Thus it is clearly important to investigate the variability of the ASM during the Holocene, and to explore its underlying forcing mechanisms, in order to improve our ability to predict regional and global climatic changes.

The northeastern Tibetan Plateau (Fig. 1) is located at a climatic junction where the ASM and the westerlies interact strongly; thus the area is climatically sensitive and ideally located for palaeoclimatic studies (An et al., 2012; Cheng et al., 2013). Over the past few decades, a range of environmental and climatic reconstructions have been conducted in the northeastern Tibetan Plateau based on various climatic archives, such as lake sediments (cf., Shen et al., 2005b; Zhao et al., 2007; An et al., 2012; Cheng et al., 2013; Qiang et al., 2013b), ice cores (Thompson et al., 1988), tree rings (Yang et al., 2014a) and aeolian deposits (cf., Lu et al., 2011; Liu et al., 2012; Qiang et al., 2013a, 2016). Among these archives, lake sediments possess the advantages that they are widely distributed on the northeastern Tibetan Plateau and can potentially provide high-resolution and continuous climate proxies for the entire Holocene.

Lake Qinghai, an alpine lake, is not only the largest lake in the region but also in China (An et al., 2012). The site is a major archive of information on environmental and climatic changes, including the uplift process of the Tibetan Plateau and landform evolution, on times scales ranging from recent to Plio-Pleistocene (An et al., 2006b, 2012; Colman et al., 2007; Fu et al., 2013). It is one of the sites chosen for investigation by the International Continental Scientific Drilling Program (ICDP). Geological and geomorphologic investigations of Lake Qinghai date back to the mid-20th century (Sun, 1938; Shi et al., 1958; Chen et al., 1964; Lanzhou Institute of Geology, 1979). During the past thirty years there have been more than five investigations of the Holocene lake sediment record of Lake Qinghai which presented climatic reconstructions from the

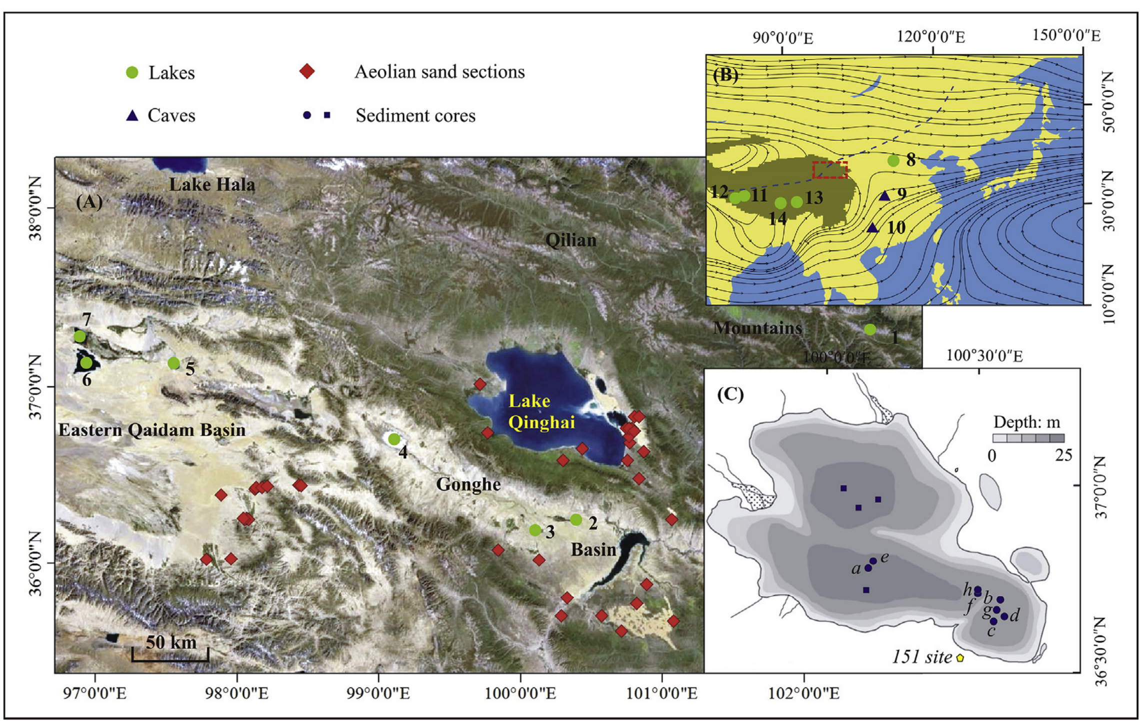

Fig. 1. Remote sensing image of the northeastern Tibetan Plateau (A) showing the locations of aeolian sand sections and lakes in the studied basins. 1- Lake Luanhaizi, 2- Lake Dalianhai, 3- Lake Genggahai, 4- Chaka Salt Lake, 5- Lake Gahai, 6- Lake Toson, 7- Lake Hurleg. The upper inserted map (B) shows June-July-August (JJA) mean atmospheric flow fields at 700 hPa from 1971 to 2000 based on the National Centers for Environmental Prediction/National Center for Atmospheric Research Reanalysis (Kalnay et al., 1996). Areas above 3000 m a.m.s.l. are shaded in dark green. The square indicates the study area of the northeastern Tibetan Plateau shown in A. The dashed line shows the modern monsoon limit (modified from Chen et al., 2008). 8- Lake Gonghai, 9- Sanbao Cave, 10- Dongge Cave, 11- Bangong Co, 12- Lake Tso Moriri, 13- Ahung Co, 14- Selin Co. The lower inserted map (C) shows the bathymetry of Lake Qinghai (modified from Colman et al., 2007). Solid dots show the locations of the long cores (*a-h*; details in Table 1), while the squares indicate the location of short cores obtained by Zhang et al. (2003), Henderson et al. (2003), Henderson (2004) and Liu et al. (2006, 2008), and yellow pentagon represents the location of archaeological site 151. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

following sediment cores: QH85-14B, 16A and 14C (Du et al., 1989; Zhang et al., 1989a, 1989b; Lister et al., 1991; Yu and Kelts, 2002); QH-2000 (Liu et al., 2002; Shen et al., 2005a, 2005b); QH1 (Guo et al., 2002; Shi et al., 2003); QH-2005 (Wang et al., 2011); LQDP05-1F and 1A, which were used to create a composite record 1Fs (An et al., 2012; Jin et al., 2015); QH07 (Liu et al., 2014a, 2014b); QH-2011 (Wang et al., 2014b, 2015b); and 2C (Zhou et al., 2016). Several short cores have also been used to reconstruct centennial- and millennial-scale climate changes: QING-6, 10 (Zhang, 2001; Zhang et al., 2003; Henderson et al., 2003; Henderson, 2004); QHN3/01 (Liu et al., 2006, 2008); QH0407-C-1, 2 (Xu et al., 2006a, 2006b, 2007a); and QH00A (Shen et al., 2001; Zhang et al., 2002) (Fig. 1C). Several long drilling cores extending back to the early Pliocene have also been obtained from shore areas around Lake Qinghai, such as Erlangjian and Yilangjian, in order to investigate long-term environmental changes (Yuan et al., 1990; An et al., 2006a; Fu et al., 2013).

Currently, more than one hundred papers have been published based on the results from lake cores and exposed sections around Lake Qinghai with most of them focusing on Holocene climatic and environmental changes. However, the inferred patterns of Holocene climate change and ASM evolution reconstructed from various environmental proxies from different cores, or sometimes even from the same core, are often inconsistent. For example, the climate of the northeastern Tibetan Plateau has been suggested to have been the wettest, with the strongest ASM, during the early Holocene, as inferred from carbonate or ostracod shell oxygen isotope $\delta^{18}O$ values (Wei and Gasse, 1999; Lister et al., 1991; Liu et al., 2007; Wang et al., 2011; An et al., 2012) and other geochemical data such as Rb, Sr (Jin et al., 2015), total organic carbon (TOC) and carbonate content (An et al., 2012; Liu et al., 2014a, 2014b), and leaf wax δ^2H (δ^2H_{wax}) (Thomas et al., 2016). However, other researchers have suggested that the early Holocene climate was relatively dry, which was possibly related to a weak ASM, as inferred from other proxies such as high shoreline deposits (Liu et al., 2015b), pollen assemblages (Du et al., 1989; Shen et al., 2005b), and geochemical evidence for lake level changes (cf. Zhang et al., 1994; Liu et al., 2013b; Wang et al., 2014b). It is clearly important to valuate and clarify these contradictory interpretations of environmental and ASM evolution for Lake Qinghai and the surrounding region during the Holocene. Consequently, in the present paper we review all of the major relevant published data; our main focus is on the differences between ostracod $\delta^{18}O$ ($\delta^{18}O_{ostr}$) records and the other proxy records, specifically relating to lake level variations of Lake Qinghai and vegetation changes and aeolian activity in the northeastern Tibetan Plateau. In the final part of the paper we discuss the most likely scenario of regional environmental changes and the evolution of the East Asian summer monsoon (EASM) during the Holocene. In order to provide a basis for comparison of the various data sets, all of the ages referenced in the text were calibrated to calendar years based on the original carbon reservoir effect estimates and we use ka (thousand years), either as calibrated radiocarbon ages or as optically stimulated luminescence (OSL) ages, unless otherwise indicated.

2. Study area

2.1. Background information on the studied basins

The study area is located on the northeastern Tibetan Plateau (Fig. 1B), with an elevation varying mainly from 2500 to 4500 m above mean sea level (a.m.s.l.). The topography is complex and there are numerous sub-basins, including Lake Qinghai Basin, Gonghe Basin and Qaidam Basin, many of which are occupied by lakes. In term of the location of some of the other major lakes, Lake Dalianhai, Lake Genggahai and Chaka Salt Lake are distributed in Gonghe Basin; Lake Gahai, Lake Hurleg and Lake Toson are distributed in the eastern Qaidam Basin; and Lake Luanhaizi is situated in the eastern part of the Qilian Mountains (Fig. 1A).

Mean annual precipitation and temperature are variable over the study area, with respective values of 100–200 mm and ~4 °C in Delingha Basin (eastern Qaidam Basin), about 310 mm and ~3.7 °C in Gonghe Basin, and about 360 mm and -0.7 °C over Lake Qinghai (Qiang et al., 2013a). Compared to the adjacent basins, precipitation is higher and temperature is lower in the mountainous areas. For example, the mean annual temperature is 0 °C, and annual precipitation is ~500 mm around Lake Luanhaizi (Herzschuh et al., 2005).

In spring and winter, strong westerlies or north- northwesterly winds prevail in the area (Wünnemann et al., 2012; Qiang et al., 2013a), and the maximum wind speed in spring in Gonghe Basin, for example, can reach 40 m/s (Qiang et al., 2013a). At the present time, aeolian activity is prevalent in the Qaidam Basin, Gonghe Basin and in the area around Lake Qinghai, resulting in broad dune fields in these basins.

The Qaidam Basin and Gonghe Basin are characterized by steppe, steppe desert and desert vegetation dominated by Chenopodiaceae, *Artemisia*, *Ephedra*, *Nitraria*, and Compositae (Zhao et al., 2007; Cheng et al., 2013). However, alpine meadows or steppe plant communities dominate the present-day vegetation surrounding high altitude lakes. Conifer forest is located at altitudes between 2600 and 3200 m a.m.s.l. in the surrounding mountains, mainly distributed in the Qilian Mountains (Herzschuh et al., 2005).

2.2. Dominating climate systems in the Lake Qinghai Basin and the surrounding areas

The ASM and the westerlies are the two major circulation systems over a vast area of eastern Asian. The ASM can also be divided into two subsystems: the Indian summer monsoon (ISM) and the EASM systems. Traditionally, changes in the seasonality of wind and in precipitation (i.e., precipitation amount) are used to define the spatiotemporal variation of the summer monsoon intensity or strength (Li and Zeng, 2002; Wang and Ding, 2008). Since it is difficult to infer the summer monsoonal wind intensity from geological archives, interpretation of the intensity of the Chinese summer monsoon, especially the EASM, emphasizes the rainfall amount in northern China: stronger EASM circulation carries more water vapor from the tropical Pacific and Indian Oceans, resulting in higher precipitation over northern China, with the opposite effect occurring with weak EASM circulation (Zhou et al., 2009b; Liu et al., 2014c; Chen et al., 2015b).

Dominated by the ASM and the westerlies, the study area is sensitive to global climate change. The distribution of the modern isohyets indicates that the bulk of the moisture transport onto the Tibetan Plateau is from the southeast (An et al., 2012), although it lies close to the modern limit of penetration of the ASM (Chen et al., 2008) (Fig. 1B). An et al. (2012) found that ASM dominated Lake Qinghai area during the Holocene, while westerly wind dominated the region during the last glaciation. It is suggested that the Tanggula Mountains, which lie about 500 km south of Lake Qinghai, mark the northern boundary of ISM propagation (Tian et al., 2001). Most of the moisture is transported from eastern Asia by direct penetration of the monsoon and by continental recycling (Thomas et al., 2016), and most of the precipitation (~70%) falls during the summer months. Xu et al. (2007a) suggested that variation in the Drought/Flood index of Xining, near Lake Qinghai, over the past 500 years, are synchronous with those of the regions in northern China where the EASM dominates. Therefore, the precipitation in north China, including the Lake Qinghai region, can potentially be used as indicator of EASM intensity.

3. Materials and methods

3.1. Sediment cores from Lake Qinghai and their chronology

Lake Qinghai is China's largest inland saline lake. It can be divided into three sub-basins: northern, southern and eastern (see Fig. 1C). The deepest parts in the three sub-basins are around 25 m. Several drilling programs in Lake Qinghai have been undertaken over the last few decades. Cores QH85-14B, 14C and 16A were drilled in the southern and eastern sub-basins (Fig. 1C) in 1985, in the course of the first international scientific drilling project in Lake Qinghai to drill through Holocene sediments into sediments of the last glacial. The results of measurements of the following proxies for these cores have been published: total nitrogen (Chen et al., 1990); pollen assemblages (Du et al., 1989; Kong et al., 1990); ostracod $\delta^{18}O$ and $\delta^{13}C$, Mg/Ca, and Sr/Ca (Zhang et al., 1989b, 1994; Lister et al., 1991); carbonate content (Kelts et al., 1989; Chen et al., 1990); and carbonate $\delta^{18}O$ ($\delta^{18}O_{carb}$) (Yu and Kelts, 2002). The reservoir effect on the ^{14}C dates from bulk organic sediment or seeds from these cores was presumed to be minimal (Zhang et al., 1989b; Lister et al., 1991; Yu and Kelts, 2002). After 2000, the results of many proxy measurements from core QH-2000 were also published, including pollen assemblages (Liu et al., 2002), carbonate content (Liu et al., 2003a), grain size (Liu et al., 2003b) as well as other proxies (Shen et al., 2005a). The chronology for core QH-2000 was not corrected for the reservoir effect until Shen et al. (2005b) published an 18-ka environmental reconstruction which used a reservoir effect of 1039 years to correct the ^{14}C ages of sedimentary organic matter. Based on this reservoir effect, the results of measurements of new proxies, such as redness (Ji et al., 2005), $\delta^{18}O_{ostr}$ (Liu et al., 2007) and the abundance of anoxygenic phototrophic bacteria (Ji et al., 2009), were published. Core QH-2005, with a length of 5.23 m, was drilled in the eastern sub-basin to the northeast of core QH-2000 in a water depth of 24 m (Wang et al., 2011). The age model for this core is mainly based on 6 bulk organic matter ^{14}C dates and 2 lignin ^{14}C dates with assumed reservoir ages of 728–2490 years. The values of $\delta^{18}O_{ostr}$ and ostracod body length, and redness and grain size of the sediments, were used to reconstruct environmental changes since 13.5 ka BP (Wang et al., 2011). Core QH07, 3.55-m-long and extending back to 20 ka BP, was drilled at the center of the eastern sub-basin of Lake Qinghai (the same sub-basin from which cores QH85-16A, QH-2000 and QH-2005 were obtained). A uniform carbon reservoir correction factor of 700 years was used to establish the core chronology (Liu et al., 2014a, 2014b). Recently, a detailed record of $\delta^{18}O_{ostr}$, $CaCO_3$ content, and total organic carbon from 18.61-m-long core 1Fs, obtained by the ICDP and spanning the last 32 ka, was published (An et al., 2012); as well as a Holocene record of $\delta^{13}C$ of bulk organic matter ($\delta^{13}C_{org}$) (Liu et al., 2013b), $\delta^{2}H$ and $\delta^{13}C$ records of leaf wax ($\delta^{13}C_{wax}$) (Thomas et al., 2014, 2016), grain size (Liu et al., 2016b), geochemistry (Jin et al., 2015) and long chain alkenones (Wang et al., 2015c). An et al. (2012) and Zhou et al. (2014) used linear regressions on three separate sections of the core to correct for the reservoir effect; ^{14}C corrections of 135, 1143 and 2523 years were obtained for the three intervals, with increasing sediment depth. For core 2C, reservoir ages of two separate sections are 140 and 2800 years respectively (Zhou et al., 2016). In addition, for 580-cm-long core QH-2011 drilled from the eastern sub-basin, Wang et al. (2014b, 2015b) used a uniform 996-year reservoir age to establish a chronology and investigate the distribution of archaeal lipids. However, based on a 538-year reservoir effect calibration, Yang et al. (2015) used the DNA of microbial functional aerobes to reconstruct past environmental conditions. Therefore, most of these core chronologies are based on only a few ^{14}C measurements of bulk organic matter without consideration of variations in the accumulation rate of the sedimentary sequences, with the exception of cores 1Fs and 2C (Zhou et al., 2014, 2016).

Determination of the reservoir age of Lake Qinghai is complex (Zhou et al., 2014), as can be seen from Table 1. A detailed and accurate chronology is critical not only for evaluating the rate of environmental changes, but also for understanding the leads and lags of climatic events and their forcing mechanisms. However, the reliability of dating lacustrine organic matter samples is subject to large temporal and spatial uncertainties because of variations in the ^{14}C reservoir effect (Godwin, 1951; Broecker and Walton, 1959; MacDonald et al., 1991; Grimm et al., 2009; Zhou et al., 2009a; Hou et al., 2012). Yu et al. (2007) used a box model based on observed data to investigate the reservoir effect in Lake Qinghai and the results indicated that both the dissolved inorganic and organic carbon pool have reservoir ages of about 1500 years for the pre-nuclear-weapons-testing era. In addition, investigations of modern surface sediments, inflowing river water and lake water were also conducted to estimate the reservoir effect of Lake Qinghai (e.g., Henderson et al., 2010; Jull et al., 2014), and the results indicated a ^{14}C age of ~660 years (Henderson et al., 2010). Therefore, although all of the sediment core chronologies from Lake Qinghai which are based on bulk organic carbon dates have incorporated a reservoir correction, they still suffer from the problem of age uncertainty.

In the present paper we aim to provide a coherent summary of the broad pattern of environmental changes in the Lake Qinghai area and of the EASM evolution during the Holocene; however, we are not concerned with changes on millennial or centennial time scales. Problems of the uncertainty of the chronologies of the various sediment cores should not affect the broad trends of change during the Holocene. It is important to note that the following review uses the previously-published chronologies.

3.2. Environmental proxies from lake sediments and landforms

3.2.1. Environmental proxies from lake sediments

Climate proxies from lake sediments can be mainly divided into two groups. One group includes pollen assemblages and terrestrial biomarkers such as $\delta^{2}H_{wax}$ (or δD) and $\delta^{13}C_{wax}$, which can directly reflect changes in terrestrial environments related to climate changes. Vegetation history inferred from pollen assemblages may be a more direct indicator of climatic changes, and especially moisture change and EASM evolution, in semi-arid areas such as the northeastern Tibetan Plateau (Liu et al., 2002; Shen et al., 2005b; Cheng et al., 2013). The second group includes such widely used proxies as the $\delta^{18}O$ of carbonate and ostracod shells and $\delta^{13}C_{org}$, and sediment chemical content (TOC, carbonate, elemental concentrations and their ratios). These proxies may not directly reflect climatic changes but rather are more likely to reflect changes in the physical, chemical and biological status of the lake, which may be influenced by multiple factors, including climate change.

Modern pollen studies in the Lake Qinghai basin indicate that pollen assemblages in each vegetation belt are significantly correlated with vegetation types (Shang et al., 2009). Modern pollen in the lake surface sediments mainly come from the vegetation in the entire drainage basin and the distribution of modern pollen in Lake Qinghai tends to be similar in most parts of the lake (Shang et al., 2009).

Table 1
Details of the Holocene sediment cores from Lake Qinghai referenced in the present paper.

Cores	Location	Core depth/cm	Water depth/m	Period/ka BP	Dating materials	^{14}C offset/years	Derivation	References
a. QH85-14B	Southern Basin	550	26	0–14.5	1 algal and 4 *Ruppia* seed samples	0	Assumed no offset	Kelts et al., 1989; Lister et al., 1991
b. QH85-16A	Eastern Basin	522	25	0–~14	7 bulk organic matter samples	439 (Colman et al., 2007)	No calibration	Zhang et al., 1989b
c. QH-2000	Eastern Basin	795	22.3	0–18	10 bulk organic matter samples	1039	Core-top intercept	Shen et al., 2005b
d. QH-2005	Eastern Basin	523	24	0–13.5	6 bulk organic matter samples; 2 lignin samples	728–2490	Interpolation	Wang et al., 2011
e. 1Fs	36°48′40.7″N, 100°08′13.5″E	1861	>20	0–32	52 bulk organic matter samples; 6 Ruppiamaritima seed samples; 7 plant remains	135; 1143; 2523	Intercept in each section	An et al., 2012; Zhou et al., 2014
f. QH07	36°43′36.7″N, 100°29′28.1″E	355	24.2	0–20	15 bulk organic matter samples; 2 Ruppiaceae seed samples	700	Average from Henderson et al., 2010	Liu et al., 2014b
g. QH-2011	36°39′34″ N, 100°35′37″ E	580	24	0–18	6 bulk organic matter samples	996; 538	Average from Shen et al. (2005b), Wang et al. (2011); linear regression	Wang et al., 2014b; Yang et al., 2015
h. 2C	36°43′37.8″ N, 100°29′28″ E	2400	>20	0–43	44 bulk organic matter samples; 1 Ruppiamaritima seed samples; 3 plant remains	140; 2800	Intercept in each section	Zhou et al. (2016)

For isotope proxies, the oxygen isotope composition of authigenic carbonate is mainly controlled by temperature and the isotopic composition of the lake water, and the latter is a function of long-term variations in the precipitation/evaporation ratio (Craig, 1965; Leng and Marshall, 2004). The hydrogen isotope composition of terrestrial plant leaf waxes is widely used to reconstruct past precipitation δ^2H (δ^2H_p); it is mainly influenced by precipitation source area, evaporative enrichment of rainfall, and condensation temperature (Lee et al., 2012; Thomas et al., 2016). Plants in the Lake Qinghai catchment likely use growing season rainfall as their main source water (Thomas et al., 2016), and therefore, δ^2H_{wax} mainly reflects growing season moisture source changes, with potential additional effects (Thomas et al., 2016). Terrestrial plant leaf wax carbon isotopes are controlled by the type of terrestrial vegetation and can be used to estimate C4 plant abundance (Thomas et al., 2014).

Lacustrine organic matter carbon isotopes are dominated by the type or sources of organic matter: terrestrial plants, including C3 and C4 plants, and aquatic plants (Meyers and Lallier-Vergès, 1999). The contribution of these different organic matter sources also affects the variation of the TOC and TN content. The trends of changes in C/N ratios can be used to determine the relative contributions of autochthonous and allochthonous organic matter in lake sediments (Meyers and Lallier-Vergès, 1999; Lamb et al., 2006). Aquatic organic matter from lacustrine algae commonly has C/N ratios between 4 and 10, while terrestrial organic matter from vascular plants has values higher than 20 (Meyers and Lallier-Vergès, 1999). Grain size parameters of lake sediments can be viewed as approximate indicators of hydrological conditions, and some of them are also sensitive to lake level changes (Xiao et al., 2009). Sediment redness is related to the sedimentary iron oxide content which appears to be derived from nearby red beds or loess deposits and is transported by fluvial processes to the lake (Ji et al., 2005). Most of the geochemical records of biomarkers reflect the chemical and biological status of the lake water (Wang et al., 2014b, 2015a).

Water depth is a basic property for describing the status of a lake and determined by many factors, including topography and catchment characteristics, climate and water management practices. Variations in the water level of closed-basin lakes usually reflect the balance of precipitation, water input by inflowing streams and surface runoff input, and evaporation. Therefore, lake level is an important source of information about past changes in hydrology and effective moisture and has been widely used to reconstruct changes in atmospheric circulation and precipitation patterns during the late Quaternary (e.g., Street-Perrott and Harrison, 1985; Morrill, 2004).

3.2.2. Past aeolian activity inferred from a probability density function (PDF) of OSL dates

Aeolian deposits are widely distributed in the Lake Qinghai Basin, Gonghe Basin and Qaidam Basin (Qiang et al., 2013a) and many sections from aeolian deposits have been investigated in Gonghe Basin, the eastern Qaidam Basin and in the area of Lake Qinghai (Lu et al., 2011; Li, 2012; Liu et al., 2012; Yu and Lai, 2012, 2014; Liu, 2013; Liu et al., 2013a; Qiang et al., 2013a, 2016; Yu et al., 2014; Lu et al., 2015). In the present paper, we collected 93 aeolian sand ages from 39 sections in aeolian deposits in order to produce a synthesis of aeolian activity in the northeastern Tibetan Plateau during the Holocene. We used a probability density function (PDF) to extract palaeoenvironmental information from the ages of the aeolian deposits. This approach can be used to produce a histogram for an ensemble of ages and has already been used successfully in previous studies (e.g., Singhvi et al., 2001; Lai et al., 2009; Yu and Lai, 2012; Qiang et al., 2013a). Age clusters in PDF plots may be an artifact of variations in sampling density from aeolian sections, as well as reflecting genuine variations in the rate of accumulation of aeolian deposits (Singhvi et al., 2001). However, careful and systematic choice of sampling density, combined with high density sampling, can minimize this potential source of error (Telfer and Thomas, 2007). Thus, the large number of OSL ages collected here should genuinely reflect variations in aeolian deposition in the region.

4. Results and discussion

4.1. Lake level variations of Lake Qinghai during the Holocene

Lake Qinghai has a complex drainage system due to the topographic heterogeneity of its catchment and is sensitive to climate changes. The history of water level changes of Lake Qinghai since the last deglacial has been reconstructed from its sediment record and from palaeoshorelines (Lister et al., 1991; Zhang et al., 1994; Yu, 2005; Liu et al., 2013b, 2015b; Wang et al., 2014b). Based on the evidence of the occurrence of aragonite nodular layers and the presence of seeds, joints, and stems from *Ruppia* in core QH85-14B, Yu (2005) found that the early Holocene lake level was about 20 m lower than today and that the highest Holocene lake level occurred from ~5.5 ka BP to 4 ka BP with a level about 10 m higher than today (Fig. 2D). This result is similar to that of a previous lake-level reconstruction of Lister et al. (1991) which gave a simple model based on seismic reflection stratigraphy, lake sediment records and shoreline terraces. A recent study of the stable isotopic composition of total organic carbon indicates that the carbon isotopic composition of organic material in the surface sediments was primarily controlled by aquatic plant species and that the $\delta^{13}C$ values of TOC can be used to indicate variations in lake level (Liu et al., 2013b). As shown in Fig. 2B, the $\delta^{13}C_{org}$ results from Lake Qinghai sediment core 1Fs indicate a level about 10 m lower than present during the early to middle Holocene (~12–5 ka BP), but a high level during the late Holocene (~3–0 ka BP) (Liu et al., 2013b). A similar scenario was presented by Wang et al. (2014b) using the relative abundance of thaumarchaeol in Lake Qinghai sediment core QH-2011 (Fig. 2C). This result strongly indicates a very low lake level during the early Holocene and that the level subsequently followed a transgressive trend. All of the foregoing lake level reconstructions are based on sediment proxies and mainly depend on the distribution of certain aquatic plants which are sensitive to water-level change.

The palaeoshorelines of a lake are direct geomorphic evidence of past changes in water level and hence potentially climate (Reheis et al., 2014). Previous investigations of the geomorphology of the Lake Qinghai Basin indicate that there are about 10 palaeostrandlines around the lake with an elevation some 10–15 m higher than the modern lake level; they are especially well-developed on the south and southeastern sides under the influence of northwesterly winds (Yuan et al., 1990). A significant amount of chronological work has been done on these palaeoshorelines (e.g., Liu et al., 2011b, 2015b), and based on shoreline investigations and OSL dating Liu et al. (2015b) proposed that the level of Lake Qinghai was 9.1 m higher than present at 5.1 ka BP; and that the level declined continuously during the past 2 ka. No high lake shoreline was found during the early Holocene, indicating that the level at that time was generally low (Fig. 2A).

The aforementioned reconstructions of Holocene lake level changes, from both the direct evidence of palaeoshorelines and the indirect evidence from sediment-based proxies, are supported by evidence of changes in lake water salinity. Zhang et al. (1989b, 1994) used Sr/Ca data from ostracods from core QH85-16A to reconstruct salinity changes in Lake Qinghai. The record indicates that the salinity was very high before ~7 ka BP and then fluctuated and decreased from 7 ka BP to ~3 ka BP, before gradually increasing from ~3 ka BP to the present. A study by Li and Liu (2014) also indicated that halophilic species of ostracods from core 1F were abundant during the early Holocene. Further salinity-based water level reconstruction results from Zhang et al. (1994) indicate that the lake level was low and fluctuated frequently during the early Holocene, was at its highest level at ~6 ka BP, was stable and relatively high from 6 to 3 ka BP, and thereafter declined to its present level (Fig. 2E).

In summary, the lake level record reconstructed from both sediment cores and geomorphic evidence demonstrates that there was no high water level stage of Lake Qinghai during the early Holocene, that the highest lake shoreline occurred during the middle Holocene or a little later, and that the lake level regressed continuously since ~2 ka BP. Coincidentally, the results of OSL dating and stratigraphic interpretation from Lake Toson in the eastern Qaidam Basin (Fig. 1A) suggest that the highest Holocene lake level occurred at 5.4 ka BP (Fan et al., 2012). In the case of Lake Hurleg, closed to Lake Toson, no beach deposits were found dating to the early Holocene and the highest lake-level period occurred at 5.0–4.7 ka BP (Fan et al., 2014). Nevertheless, it should be noted that the water level of Lake Genggahai in Gonghe Basin was high during the early Holocene, as indicated by sediment-based proxies (Qiang et al., 2013b). This high lake-level phase at the site may result from unusual local topographic conditions of Lake Genggahai and needs to be further investigated, because Lake Genggahai comprises two lakes and water in Upper Genggahai would drain into Lower Genggahai during intervals of higher lake level. In addition, biomarker-based proxies from a sediment core from Lake Gahai indicated that the water level was low during 7–2 ka BP (He et al., 2014), which appears to be in conflict with evidence of high palaeoshorelines records from lakes in the eastern Qaidam Basin. However, this result needs to be carefully reassessed and the sediment core chronology reconsidered because most of the ages are not in stratigraphical order; in addition, the curve (Fig. 6e) was wrongly cited in the paper by He et al. (2014), which reduces the credibility of the data. In general, with the exception of Lake Genggahai, there is significant evidence for lower-than-present lake-levels in the northeastern Tibetan Plateau during the early Holocene; and for higher-than-present lake-levels during the middle Holocene or a little later.

The variation of lake level of Lake Qinghai primarily reflects the balance of water input and evaporation loss; however, human activity may also have had an effect on the lake level during the past few decades (Zhang et al., 2011a). The monthly variation of the water level during the period from 1961 to 2002 indicates that the lake level begins to increase in May and reaches its highest status during September, and then decreases (Li et al., 2005). This highlights the important role of summer precipitation in controlling the water level variations of Lake Qinghai. A lower lake level and higher water salinity during the early Holocene are usually ascribed to intensive evaporation as a result of high local temperature (Liu et al., 2013b; Li and Liu, 2014; Wang et al., 2014b). However, this interpretation is inconsistent with the ostracod isotope records discussed in Section 4.3; therefore, Thomas et al. (2016) proposed that the increase in summer plant productivity could have removed water from the Lake Qinghai catchment, resulting in lower river runoff and lower lake level during the early Holocene. Lake Qinghai was very shallow (Jin et al., 2015) and loess-like silty sediment was deposited during the late Pleistocene (Kelts et al., 1989; Yu, 2005). The salty silty sediment may have been distributed on some parts of the flat lake bed as indicated by the increased abundance of salt-tolerant plants during that period (Thomas et al., 2014). The increased precipitation during the early Holocene may have caused rising lake level which led to the dissolution of the salt in the salty silty sediment, thus probably causing an increasing in lake water salinity to some extent. In contrast, the higher inflow during the middle Holocene may have caused a decrease of the salinity of the lake water and further increased the lake level.

4.2. Holocene aeolian activity recorded by aeolian deposits in the northeastern Tibetan Plateau

Aeolian deposits are widely distributed in the northeastern

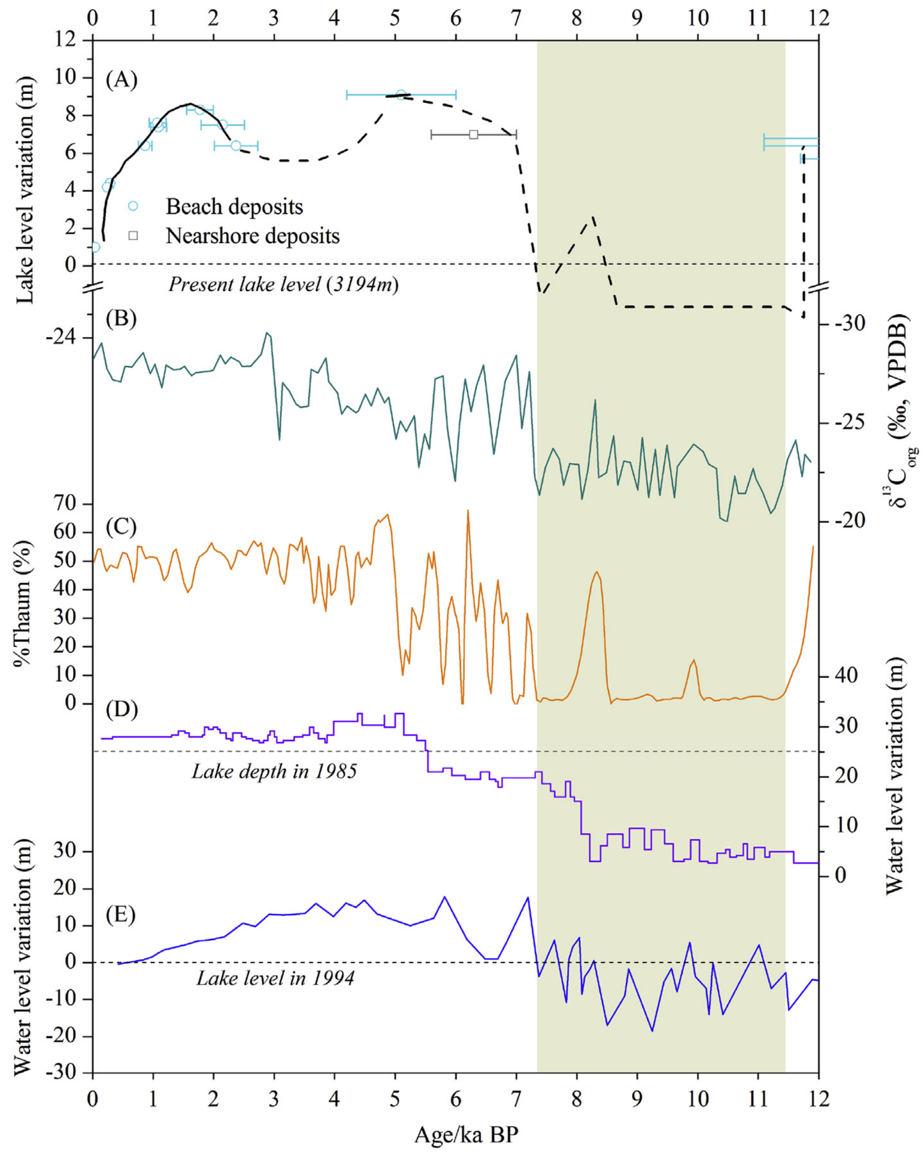

Fig. 2. Water-level change history of Lake Qinghai based on evidence from sediments cores and study of palaeoshorelines. (A) reconstructed shoreline history for Lake Qinghai for the last ~12 ka, plotted against age estimates for shoreline elevations (Liu et al., 2015b); (B) variation of $\delta^{13}C$ of sedimentary total organic matter of sediment core 1Fs from Lake Qinghai (Liu et al., 2013b). More negative values represent higher lake levels; (C) relative abundance of thaumarchaeol in Lake Qinghai sediments (Wang et al., 2014b); (D) reconstructed water depth from sediment core QH85-14B from Lake Qinghai based on the occurrence of aragonite nodular layers and the presence of seeds, joints, and stems of *Ruppia* (Yu, 2005); (E) record of water level variations based on ostracod Sr/Ca value (Zhang et al., 1994).

Tibetan Plateau. They are generally composed of well-sorted aeolian sand, palaeosols and/or loess (Lu et al., 2011; Qiang et al., 2013a) and the lithological variations can potentially reflect changes in sedimentary environments and climatic conditions. For example, the accumulation of aeolian sand usually reflects an arid climate with low effective moisture (Lu et al., 2011; Qiang et al., 2013a; Yu and Lai, 2014); and palaeosols are developed under a relatively humid climate when aeolian activity was weaker or even non-existent. The latter conditions usually reflect an increased ASM (An et al., 2000; Yu and Lai, 2014; Lu et al., 2015). The OSL dating method has been successfully applied to aeolian sediments and can provide reliable ages for sedimentary sequences (Lu et al., 2011; Qiang et al., 2013a; Yu and Lai, 2014). Therefore, in addition to lacustrine sediments, aeolian deposits are an independent source of information on moisture changes.

The age distribution of aeolian sand development is illustrated in Fig. 3A and B. It is clear that a major peak in aeolian sand accumulation occurred during ~10–7.5 ka BP, indicating enhanced aeolian deposition and strong aeolian activity, mainly during the early Holocene. The lower probability values indicate that weak aeolian activity occurred after ~7.5 ka BP; and increased probability values and thus renewed aeolian activity after ~1.5 ka BP. The precipitation reconstruction based on tree rings exhibits distinct fluctuations in the region after 3.5 ka BP and there was no decreasing trend after 1.5 ka BP (Yang et al., 2014a). Therefore, the increasing aeolian activity after ~1.5 ka BP may not be related to climate change and may instead be a consequence of intensified human activity, including grazing and agricultural development (Zhang et al., 1988; Yu et al., 2014; Chen et al., 2015a; Wu et al., 2016). In addition, mid-late Holocene pollen records from Lake Genggahai indicate that the percentage of Brassicaceae pollen increased significantly after around 1.5 ka BP (Liu et al., 2016a, Fig. 3C), which may have been caused by cultivation. A lower probability distribution of aeolian sand accumulation during the last deglaciation does not necessarily imply weak aeolian activity; rather, it may reflect intense aeolian activity resulting in frequent reworking of aeolian sand due to strong winds and low effective humidity during the last glacial, as suggested by previous studies (Sun et al., 2007; Liu et al., 2012; Xu et al., 2015). Alternatively it may indicate that few aeolian sections were investigated and limited OSL data were obtained during that period. There is no doubt that aeolian sand was widely deposited around the northeastern Tibetan Plateau mainly during the early Holocene, as suggested by several researchers (e.g., Lu et al., 2011; Qiang et al., 2013a, 2016).

As mentioned previously, palaeosol development is also an indicator of climate changes. However, because of the relatively small number of palaeosol ages, the PDF plot of palaeosols may be unreliable; moreover, there is no common signal of palaeosol development between the studied basins. For example, Qiang et al. (2013a, 2016) suggested that palaeosol occurred primarily during the late Holocene in Gonghe Basin. Soil development occurred at most sites in the Lake Qinghai Basin during the middle to late Holocene (Lu et al., 2011). For the eastern Qaidam Basin, palaeosols developed during the early and mid-Holocene (Yu and Lai, 2014). In addition, the age of a palaeosol may not correspond to the age of pedogenesis, but rather to the age of previously deposited materials (Lai et al., 2012). Therefore, the ages of palaeosols cannot provide additional information about regional environmental change in the northeastern Tibetan Plateau.

The relationship between aeolian activity and regional environmental change is complex. There are several factors which can potentially affect migration and deposition of sands. The development of vegetation cover, acting as a sediment trap, plays an important role in sand mobility and deposition (Stauch et al., 2012; Stauch, 2015; Qiang et al., 2016). Wind strength, sand supply and soil formation are important additional factors potentially affecting sand mobility and deposition (Mason et al., 2009; Qiang et al., 2016). In addition, Stauch (2015) suggested that the strengthening of aeolian accumulation on the Tibetan Plateau depends on the previous climate state. For the northeastern Tibetan Plateau, the cold and dry climate during the last glacial and the availability of fluvial sediments could have provided sufficient materials for aeolian sand deposition during the early Holocene. Although the strength of westerlies decreased significantly during the Holocene compared to the last glacial in the Lake Qinghai area, the westerlies during the early Holocene were still stronger than during the middle and late Holocene (An et al., 2012) and could have enhanced regional aeolian activity. In addition, the development of soil and vegetation recovery during the Holocene, especially after the early Holocene, prevented aeolian sand activity. However, the most popular view is that when the regional climate becomes arid and/or there is a low effective moisture, aeolian activity and the corresponding sand deposits prevail over a large area (Lu et al., 2011; Qiang et al., 2013a).

4.3. Ostracod $\delta^{18}O$ records from Lake Qinghai

Studies of Quaternary ostracods from the Tibetan Plateau date back several decades ago and several investigations have been conducted on the northeastern Tibetan Plateau (Kelts et al., 1989; Lister et al., 1991; Mischke, 2012). In addition to species assemblages and shell chemistry, the stable isotopic composition of ostracod shells provides important information for the reconstruction of regional environmental change. Although influenced by several factors such as lake temperature, $\delta^{18}O$ of the lake water, and disequilibrium and vital effects (Wei and Gasse, 1999; Leng and Marshall, 2004; Henderson et al., 2010), the $\delta^{18}O_{carb}$ and $\delta^{18}O_{ostr}$ values of lake sediments in semi-arid to arid regions are often considered as a proxy for the regional water balance between the precipitation in a catchment and lake surface evaporation, and consequently as an indicator of regional effective moisture (Jones et al., 2006; Mischke, 2012).

During the past three decades, five $\delta^{18}O_{ostr}$ records spanning the entire Holocene have been obtained from sediment cores from Lake Qinghai (Fig. 1C). In spite of the problems of differing carbon reservoir ages used in these studies, the four $\delta^{18}O_{ostr}$ records illustrated in Fig. 4A–D demonstrate a generally similar pattern,

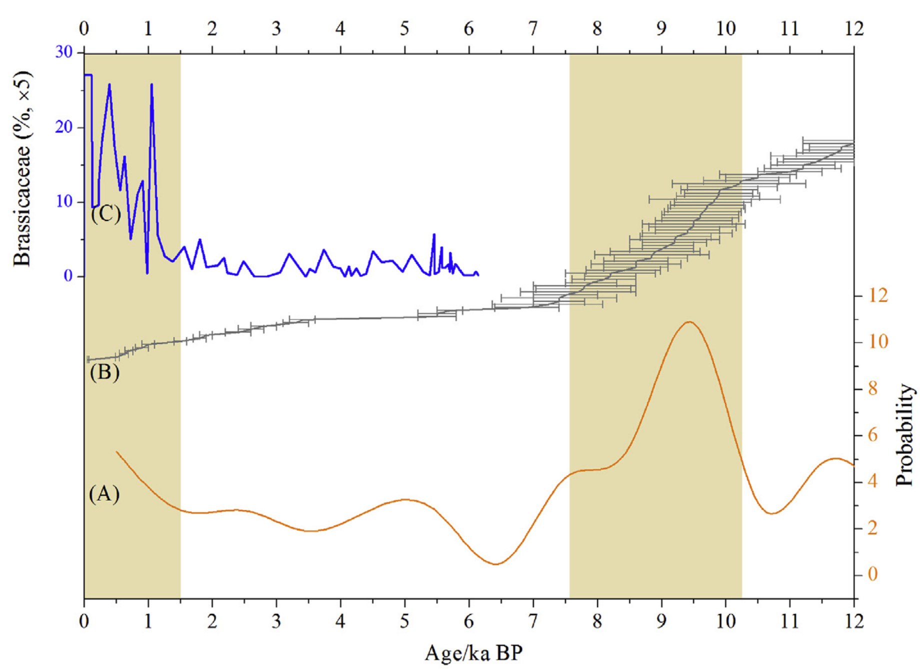

Fig. 3. Probability density function plot for OSL ages of aeolian sand sections from the northeastern Tibetan Plateau (A and B), and Brassicaceae pollen percentages from Lake Genggahai (C; Liu et al., 2016a). The shaded areas indicate the two periods of increased frequency of aeolian sand deposition between ~10–7.5 ka BP and since ~1.5 ka BP.

with depleted $\delta^{18}O_{ostr}$ values during the early to middle Holocene and enriched $\delta^{18}O_{ostr}$ values during the late Holocene. The oxygen isotopic composition of ostracods in Lake Qinghai has been used to indicate changes in palaeotemperature (Zhang et al., 1989b, 1994), effective humidity (Lister et al., 1991), and/or effective precipitation and monsoonal patterns (Wei and Gasse, 1999; Henderson et al., 2003; Liu et al., 2007; An et al., 2012). The $\delta^{18}O_{ostr}$ record from core QH85-16A has been suggested to reflect relative changes in temperature (Zhang et al., 1989b, 1994). However, Lister et al. (1991) interpreted the $\delta^{18}O_{ostr}$ record from core QH85-14B as indicating that effective humidity (evaporation/input water ratio) controls the $\delta^{18}O$ values of the lake water and ostracod shells. However, in the study of authigenic carbonate $\delta^{18}O$ in Lake Qinghai by Liu et al. (2008), the variations of $\delta^{18}O$ are interpreted as reflecting long-term mean isotopic values resulting from changes in moisture source in the region. Based on recent sediment cores from Lake Qinghai, it is accepted that the oxygen isotopic composition of ostracod shells can be used as a proxy for the effective precipitation or the strength of the ASM (Henderson et al., 2003; Liu et al., 2007; An et al., 2012). As suggested by Liu et al. (2007) and An et al. (2012), the depleted $\delta^{18}O_{ostr}$ values during the early Holocene indicate increased effective precipitation, implying strengthened monsoon intensity. In contrast, the notably enriched $\delta^{18}O_{ostr}$ values in the late Holocene reflect decreased monsoon intensity (Liu et al., 2007; An et al., 2012).

However, the prevailing view that ostracod $\delta^{18}O$ in sediments of Lake Qinghai reflect effective precipitation and summer monsoon intensity (Liu et al., 2007; An et al., 2012) is contradicted by other records. The records of depleted $\delta^{18}O_{ostr}$, interpreted as indicating intensive precipitation and strong monsoon intensity during the early Holocene, do not accord with coeval records of low water level of Lake Qinghai (Fig. 2) and other lakes in the northeastern Tibetan Plateau, and with the widely distributed aeolian sand deposits discussed above (Fig. 3). Several authors argue that the low lake level and intense aeolian activity during the early Holocene was caused by high evaporation under conditions of high temperature (Liu et al., 2013b; Qiang et al., 2013a); however, this remains in conflict with observations of the most negative $\delta^{18}O_{ostr}$ values during the early Holocene (Lister et al., 1991; Liu et al., 2007; An et al., 2012) because strong evaporation may lead to more positive $\delta^{18}O_{ostr}$ values. In addition, lake sediment proxies indicating that lake salinity was generally high during the early Holocene (Zhang et al., 1994; Li and Liu, 2014) are also in conflict with suggestions of highest effective humidity and strong summer monsoon inferred from the strongly negative $\delta^{18}O_{ostr}$ values at that time (Lister et al., 1991; Liu et al., 2007; An et al., 2012). Furthermore, pollen assemblages (Du et al., 1989; Liu et al., 2002; Shen et al., 2005b) indicate that regional vegetation changed from open forest-grassland and forest-grassland during the early Holocene, to forest during the middle Holocene, to forest-grassland and open forest-grassland during the late Holocene, indicating high precipitation and a strong summer monsoon during the middle Holocene (Fig. 6). Moreover, the sediment redness record also indicates that the strongest summer monsoon occurred during the middle Holocene (Ji et al., 2005; Wang et al., 2011). Therefore, it is possible that the $\delta^{18}O_{ostr}$ records from Lake Qinghai do not in fact reflect precipitation, summer monsoon and effective humidity. Thus, the significance of the $\delta^{18}O_{ostr}$ records from Lake Qinghai needs to be reassessed.

Temperature and the isotopic composition of the host water mainly control the oxygen isotope composition of authigenic carbonate (Craig, 1965; Leng and Marshall, 2004). However, the temperature dependence of equilibrium isotopic fractionation between carbonate and water is ca. -0.24‰ per °C (Craig, 1965) and the Holocene temperature variations are within the range of 2–3 °C (Fig. 6I; Shi et al., 1992; Marcott et al., 2013; Liu et al., 2014d), which cannot reasonably explain the observed ~2–7‰ of $\delta^{18}O_{ostr}$ variation during the Holocene which is obtained from sediment records from Lake Qinghai. The variation of the oxygen isotopic composition of biogenic carbonate such as ostracod shells, therefore, is mainly controlled by variations in the oxygen isotopic composition of the lake water, which is further influenced by the oxygen isotopic composition of regional moisture, direct precipitation inputs, river water flowing into lake and lake surface evaporation intensity related to temperature (Zhang et al., 2011b).

It is notable that the interpretation of the $\delta^{18}O_{ostr}$ records as reflecting the intensity of the summer monsoon or monsoonal precipitation was based on the assumption that the lake water oxygen isotopic composition results from the isotopic composition of the input water, i.e., regional rainfall (Liu et al., 2007; Henderson et al., 2010; An et al., 2012). Evaporation-induced isotopic enrichment of the water of a closed lake could potentially cause the long-term increase of $\delta^{18}O$ values of lake water during the Holocene (Wang et al., 2011), however this effect is limited in the case of Lake Qinghai. Leaf wax $\delta^{2}H$ is not affected by lake water evaporation, and therefore it is a more direct indicator of precipitation isotopes. In addition, the co-variation of $\delta^{2}H$ and $\delta^{18}O$ of modern precipitation is widely observed (Craig, 1961), as well as in the Lake Qinghai region (Wu et al., 2014). The $\delta^{18}O$ values of ostracod and the $\delta^{2}H$ values of leaf wax from Lake Qinghai exhibit the same trend of variation during the Holocene (see section 4.4 for more details), thus indicating that isotope composition of precipitation controls the isotopic characteristics of lake water and terrestrial water. Therefore, it can be accepted that variations in the $\delta^{18}O$ of biogenic carbonate in Lake Qinghai may mainly reflect changes in the $\delta^{18}O$ of the regional precipitation ($\delta^{18}O_{prep}$), which is also the basic assumption used for interpreting the $\delta^{18}O$ of biogenic carbonate in Lake Qinghai as a proxy of monsoon intensity (Liu et al., 2007; An et al., 2012). In fact, the biogenic carbonate $\delta^{18}O$ records from Lake Qinghai during the Holocene (Fig. 4A–D) exhibit the same trends as the stacked $\delta^{18}O_{carb}$ record (Fig. 4E) from 10 lakes across monsoonal China (Zhang et al., 2011b). Within the limits of the error resulting from chronological uncertainties in the Holocene sediment cores from Lake Qinghai, the Holocene $\delta^{18}O_{ostr}$ records (Fig. 4A–D) are also similar to the Chinese speleothem $\delta^{18}O$ ($\delta^{18}O_{cave}$) records (Fig. 4F and G; Dykoski et al., 2005; Dong et al., 2010), due to the fact that all of the carbonate $\delta^{18}O$ records (Fig. 6A–G) reflect changes in $\delta^{18}O_{prep}$. Thus, the use $\delta^{18}O_{carb}$ as a proxy of monsoon intensity actually faces the same problem as that faced by the use of $\delta^{18}O_{cave}$: that is, the question of whether or not changes in $\delta^{18}O_{prep}$ faithfully record the monsoon intensity or changes in precipitation amount.

Previously, the accepted interpretation of the speleothem $\delta^{18}O$ records from monsoon-dominated China was that they mainly reflected either EASM intensity (Wang et al., 2001; Cheng et al., 2006, 2009), or changes in the amount of summer rainfall associated with EASM intensity (Johnson and Ingram, 2004; Cai et al., 2010). However, the interpretation of $\delta^{18}O_{cave}$ as an EASM indicator has already been strongly challenged by observational and modelling studies (Maher, 2008; Tan, 2009, 2014; Clemens et al., 2010; Pausata et al., 2011; Caley et al., 2014; Liu et al., 2015a). A review of Holocene stalagmite $\delta^{18}O$ records from the EASM and ISM regions demonstrated that most of them exhibit a remarkably similar trend of variation (Yang et al., 2014b). On short-time scales, nearly all the $\delta^{18}O_{cave}$ records from monsoonal China exhibit an increasing trend since 1900 AD (Tan, 2014; Liu et al., 2015a). The similar trend of variations at different timescales contradicte with spatial differences of precipitation across the monsoonal China (Chen et al., 2016). Based on the concept of a “circulation effect” proposed by Tan (2009, 2014) and Maher and Thompson (2012), the isotopic

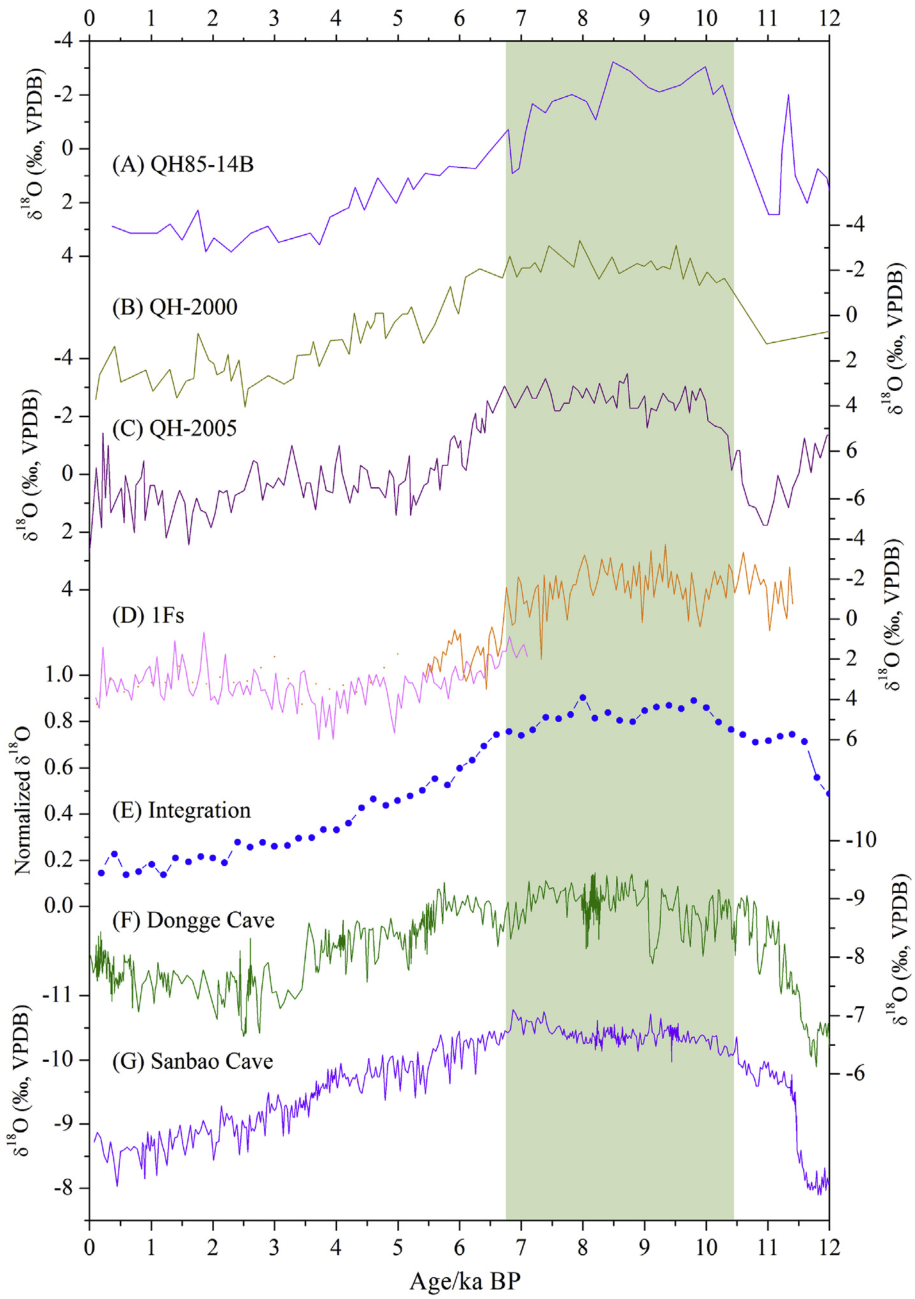

Fig. 4. Comparison of ostracod $\delta^{18}O$ records from lake sediments cores from Lake Qinghai (A- Core QH85-14B (Lister et al., 1991); B- Core QH-2000 (Liu et al., 2007); C- Core QH-2005 (Wang et al., 2011); and D- Core 1Fs (An et al., 2012)); stacked carbonate $\delta^{18}O$ record from 10 lakes in monsoonal China (E; Zhang et al., 2011b) and Chinese stalagmites $\delta^{18}O$ records (F- Dongge Cave (Dykoski et al., 2005); G- Sanbao Cave (Dong et al., 2010)) during the Holocene. The shaded column indicates a commonly-recorded interval of depleted $\delta^{18}O$ during the early Holocene.

composition of precipitation is predominantly a signal of water source. The "amount effect" meaning that depleted $\delta^{18}O_{prep}$ values reflect increased precipitation (Fleitmann et al., 2003), may only be used in the core ISM-dominated regions to interpret speleothem $\delta^{18}O$. Based on a comparison with various independent records of Holocene EASM variability, the speleothem $\delta^{18}O$ signal in monsoon-dominated China was proposed to mainly reflect variations in the isotopic composition of precipitation (Yang et al., 2014b), and thus potentially ISM intensity - but probably not the strength of the EASM (Pausata et al., 2011; Chen et al., 2015b, 2016; Liu et al., 2015a). Therefore, if both the Chinese speleothem and lacustrine carbonate $\delta^{18}O$ records mainly reflect changes in $\delta^{18}O_{prep}$, the similar pattern of changes of the $\delta^{18}O_{carb}$ records from Lake Qinghai (Fig. 4A–D), as well as Selin Co (Kashiwaya et al., 1995), Ahung Co (Morrill, 2004), Bangong Co (Gasse et al., 1996) and Lake Tso Moriri (Mishra et al., 2015) on the Tibetan Plateau (Fig. 1B), should also document the same moisture signal. However, the status of precipitation and/or climate differs between the northern and southern Tibetan Plateau, as suggested by several researchers (e.g., Wang, 2006; Zhang et al., 2015). Therefore, the isotopic composition in Lake Qinghai cannot be regarded as an indicator of the intensity of monsoon precipitation, and it is essentially a signal of precipitation isotopic composition which is mainly dominated by moisture source.

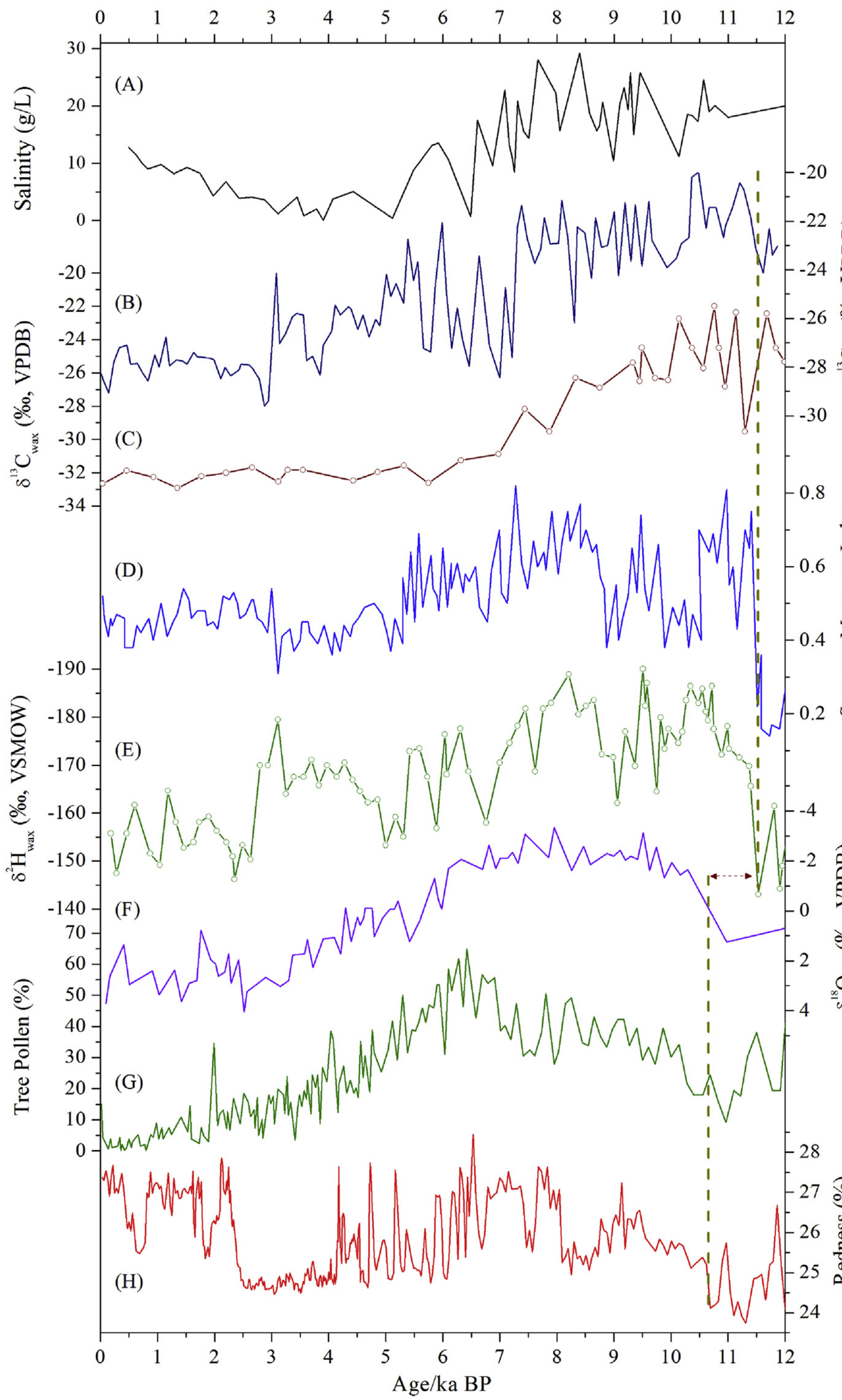

Fig. 5. Comparison of conflicting records from Lake Qinghai during the Holocene. (A) lake water salinity inferred from Sr/Ca ratios of ostracod (Zhang et al., 1994); (B) $\delta^{13}C$ of sedimentary total organic matter as a proxy of water depth (Liu et al., 2013b); (C) carbon isotopic composition of leaf waxes (Thomas et al., 2014); (D) Asian summer monsoon index based on total organic carbon and carbonate flux (An et al., 2012); (E) hydrogen isotopic composition of leaf waxes (Thomas et al., 2016); (F) ostracod $\delta^{18}O$ record (Liu et al., 2007); (G) tree pollen percentage (Shen et al., 2005b); and (H) redness record indicating river discharge (Ji et al., 2005). Curve (A) for QH85-16A; (B)–(E) for core 1F; (F)–(H) for core QH-2000. Dashed lines indicate the age difference between cores QH-2000 and 1F.

4.4. Comparison of conflicting records from Lake Qinghai during the Holocene

The conflict between proxies from different cores or even the same core is illustrated in Fig. 5. Thomas et al. (2014) converted leaf wax carbon isotope ratios (Fig. 5C) to C4 plant abundance and the results indicate that C4 plants contributed a significant proportion of local primary productivity during the early Holocene. This may be mainly caused by high growing season temperature (Rao et al., 2010, 2012) as temperature was generally high during the early Holocene in the northern hemisphere (Fig. 6I; Marcott et al., 2013). The lower stands of Lake Qinghai during the early Holocene are

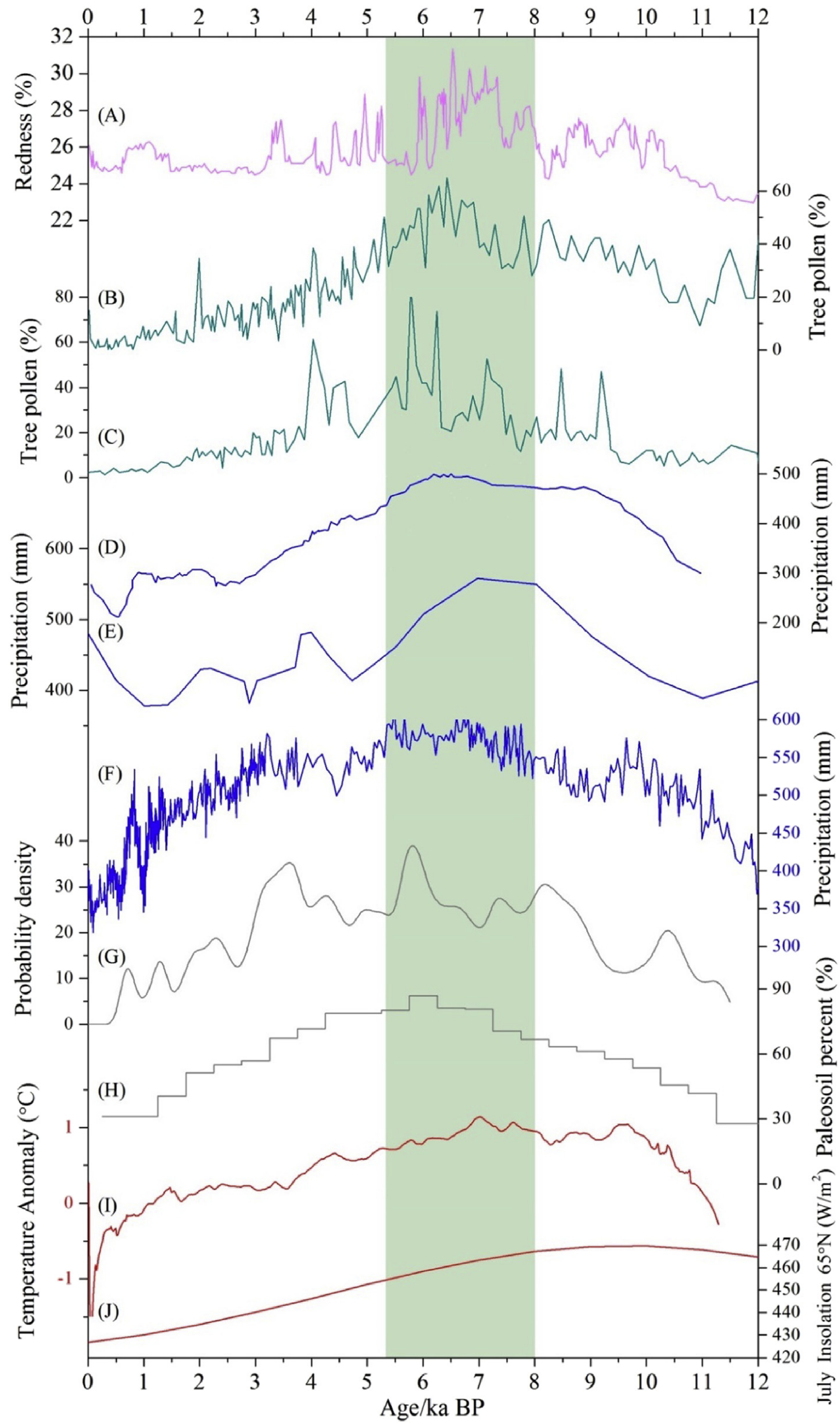

Fig. 6. Multi-proxy records of Holocene climate change from the northeastern Tibetan Plateau and comparison with other records. (A) redness record from sediment core QH-2005 from Lake Qinghai, a proxy of river discharge (Wang et al., 2011); (B) tree pollen percentages from sediment core QH-2000 from Lake Qinghai (Shen et al., 2005b); (C) tree pollen percentages from Lake Dalianhai in Gonghe Basin, south of Lake Qinghai (Cheng et al., 2013); (D) quantitative precipitation reconstruction from Lake Qinghai (Herzschuh et al., 2010); (E) quantitative precipitation reconstruction from Lake Luanhaizi in the Qilian Mountains, northeast of Lake Qinghai (Wang et al., 2014c); (F) variations in EASM precipitation recorded by Lake Gonghai, north China (Chen et al., 2015b); (G) EASM evolution indicated by the frequency distribution of palaeosol dates from the Chinese Loess Plateau (Wang et al., 2014a); (H) effective moisture changes based on palaeosol development in the deserts and Sandy Lands in northern China (Li et al., 2014); (I) Northern Hemisphere (30°–90°N) temperature record (Marcott et al., 2013); (J) July insolation at 65° N (Berger and Loutre, 1991).

demonstrated by a range of evidence, including records of $\delta^{13}C$ of sedimentary total organic matter (Liu et al., 2013b, Fig. 5B), and the lake water salinity inferred from Sr/Ca ratios from ostracods (Zhang et al., 1994, Fig. 5A). Liu et al. (2016b) used skewness and a grain-size ratio (8–50/2-8 μm) to reconstruct hydrodynamic changes in the Lake Qinghai Basin and suggested that the highest skewness during the early Holocene reflected the strongest precipitation. However, the coarser grain size during the early Holocene may have been caused by a lower lake level because of the decreasing distance from the shoreline to the core site, as well as by persisting strong westerlies during the early Holocene (An et al., 2012).

An et al. (2012) suggested that the carbonate content and total organic carbon flux in the sediments of Lake Qinghai can reflect variations in the summer monsoon and they further normalized and averaged these two proxies to create an index of the ASM (Fig. 5D). Supported by the ostracod $\delta^{18}O$ record (Fig. 4D), the authors argued that the ASM tracked northern hemisphere summer insolation during the Holocene, with strengthened monsoon intensity during the early Holocene. However, the sedimentary total organic carbon in Lake Qinghai is mainly derived from authigenic sources since C/N ratios average less than 10 during the Holocene (Shen et al., 2005b). After analyzing modern sediments and aquatic plants, Liu et al. (2013b) and Ao et al. (2014) also found that the sedimentary total organic matter in Lake Qinghai is mainly generated from internal carbon sources. Therefore, the use of TOC content as a summer monsoon indicator may not be reliable since its origin in lake sediments is complex (Meyers and Lallier-Vergès, 1999).

The TOC content or flux changed rapidly from low values during the glacial period to high values during the early Holocene (An et al., 2012; Liu et al., 2014b), followed by generally decreased values from the middle Holocene to the late Holocene with two intervals of low values at around ~9.5 ka BP and ~10.5 ka BP (An et al., 2012). The variation of TOC during the Holocene may have been mainly dominated by water temperature because it exhibits the same pattern of variation as water temperature based on the long chain alkenone record (Wang et al., 2015c). In addition, it is clear that lake productivity may depend on nutrient supplies and temperature or even solar intensity, all of which are not directly linked to the summer monsoon strength.

In addition, the TOC content of four short (~1.0 m long) cores obtained from the center of the three sub-basins of Lake Qinghai (Fig. 1C) all exhibit increasing trend over the last 1000 years (Zhang, 2001; Zhang et al., 2003). However, regional precipitation does not exhibit the same trend (Yang et al., 2014a; Gou et al., 2015) and in fact the EASM intensity decreased during the last 1000 years, with a strong summer monsoon and high precipitation in northern China during the Medieval Warm Period (Liu et al., 2011a; Chen et al., 2015c). Thus it is clear that the TOC content or TOC flux cannot be used as a reliable summer monsoon proxy, and given this fact it is unsurprising that the pattern of monsoon evolution inferred from TOC content or TOC flux records (An et al., 2012; Liu et al., 2014a, 2014b) differs from that inferred from the more reliable EASM proxies, including the pollen and redness records discussed above.

Based on the investigation of modern lake water and river water, it is suggested that the carbonate chemistry of the water from Lake Qinghai is Ca^{2+}- limited, and that carbonate precipitation is controlled by the amount of Ca^{2+} brought to the lake by rivers (Jin et al., 2010). Therefore, in turn, the carbonate content in Lake Qinghai is regarded as reflecting the regional rainfall (An et al., 2012; Liu et al., 2014b). However, there are several other possible factors that can affect the precipitation of carbonate in lake water, including the fact that sedimentary carbonate content will be enhanced by warm water as a result of the decrease in carbonate solubility under the condition of high temperature. In addition, high primary production in lake systems removes aqueous CO_2 in the process of photosynthesis, promoting carbonate precipitation and increasing carbonate content (Liu et al., 2014b). The variations of the carbonate content of both short cores and long Holocene cores from Lake Qinghai exhibit the opposite trend to that of the total organic matter or TOC content, on a centurial time scale (Zhang et al., 2003; Liu et al., 2014b). However, both exhibit the same pattern on a glacial-interglacial time scale (Shen et al., 2005b; Liu et al., 2014a), as well as the same trend as water temperature variations during the Holocene (Wang et al., 2015c). Therefore, the carbonate precipitation in Lake Qinghai during the Holocene was mainly controlled by temperature or increased lake productivity, as inferred from the variations of TOC during the Holocene.

With regard to the ostracod $\delta^{18}O$ record, it is clear that it cannot be regarded as a reliable monsoon index in the Lake Qinghai area, as discussed in section 4.3. Leaf wax is formed during the growing seasons when the precipitation in the northeastern Tibetan Plateau is mainly from monsoon sources, and it has been used to reconstruct changes in moisture sources as well as Holocene Asian monsoon evolution in the region (Rao et al., 2016b; Thomas et al., 2016). The leaf wax δ^2H record of Lake Qinghai (Fig. 5E) exhibits the same pattern of variation as the ostracod $\delta^{18}O$ record and the record of hydrogen isotopic composition from Lake Genggahai (Rao et al., 2016b), with more negative values during the early Holocene. This trend can also be observed in other lakes. For example, in a well-dated sedimentary record of compound-specific carbon ($\delta^{13}C$) and hydrogen (δ^2H) isotopes of long-chain n-alkanes from Lake Gonghai in northern China, the interval from cal. 11–5 ka BP is characterized by more negative isotopic ratios in precipitation (as indicated by the δD data); however, the interval from cal. 8–5 ka BP is characterized by enhanced humidity (as indicated by the $\delta^{13}C$ data) (Rao et al., 2016a) and the highest precipitation (Chen et al., 2015b). This highlights the uncertainties in using isotopic composition of precipitation (including both $\delta^{18}O$ and δ^2H) as indicators of EASM intensity in this region (Rao et al., 2016a). Rao et al. (2016c) reviewed Holocene lacustrine hydrogen isotopic records, authigenic carbonate and cave stalagmite oxygen isotopic records from the Asian monsoon region. They found that all of these isotopic records exhibit roughly similar long-term characteristics in both the ISM and EASM regions, i.e. they were all more negative during the early Holocene and the early to mid-Holocene (ca. 11–6 ka BP), and then became more positive towards the late Holocene. After further comparison with palaeo-humidity records from the two monsoon regions, they suggested that both $\delta^{18}O$ and δD palaeorecords could not be used directly as reliable palaeo-humidity (i.e. precipitation amount or EASM intensity) indicators in the EASM region. Therefore, the precipitation isotopic composition in EASM region is essentially a signal of the moisture source whose temperature may play an important role in determining the isotopic composition of the precipitation in monsoonal China (Rao et al., 2016a). In addition, precipitation isotopic composition in the ISM region may represent the intensity of the ISM (Fleitmann et al., 2003, 2007; Chen et al., 2014; Yang et al., 2014b; Rao et al., 2016c).

In summary, although all of the proxies discussed above exhibit an "oxygen isotope" pattern (Fig. 5F), with lowest values during the early Holocene and higher values during the middle and late Holocene and vice versa, which resembles the trend of northern hemisphere summer insolation, they are not true indices of the intensity of the EASM. The other proxies from Lake Qinghai, such as tree pollen percentages and sediment redness, which are regarded as indicators of regional precipitation and runoff respectively, all exhibit higher values during the middle Holocene and thus that the strongest precipitation occurred during that period (Shen et al., 2005b; Ji et al., 2005, Fig. 5G–H). It should be noted that *Pinus* accounts for a significant proportion of the pollen assemblages and dominates the variation of tree pollen percentages, especially during the middle Holocene. Although *Pinus* has a high pollen productivity and it is possible that *Pinus* pollen could be transported by wind from other regions (Xu et al., 2007b), *Pinus* does grown in the Lake Qinghai area at the present time. In addition, charcoal records from archaeological site 151 on the southern shore of Lake Qinghai (Fig. 1) indicate *Pinus* occupied the region to a certain degree during the mid-Holocene (Dongju Zhang, personal communication). Finally, it should be noted that the conflict between different proxies is not caused by difference in radiocarbon reservoir evaluation because this conflict occurs in one core with the same age (Fig. 5F–H).

4.5. Holocene moisture change and EASM evolution in Lake Qinghai and its environs

The availability of multiple proxies from lake sediment cores from Lake Qinghai provides a significant advantage in terms of avoiding the uncertainties involved in the use of single proxy such as the $\delta^{18}O$ of biogenic carbonate. This is especially so for reconstructing regional environmental changes and monsoon evolution during the Holocene since the region is located on the margin of the present summer monsoon-dominated areas (Fig. 1B) (Liu et al., 2008; An et al., 2012). Therefore, a review of the various proxies documented in different sediment cores from Lake Qinghai and the surrounding region can potentially provide an insight into Holocene climatic trends, especially EASM evolution, since previous studies indicated that Holocene climate change in the northeastern Tibetan Plateau was dominated by the EASM (Shen et al., 2005b; Cheng et al., 2013). Frequently-used environmental proxies include changes in palaeoshoreline elevation used to reconstruct changes in lake water depth or lake level (Fig. 2A; Liu et al., 2015b); ostracod $\delta^{18}O$ values used to reconstruct changes in the water oxygen isotopic composition of the lake (Fig. 4A–D) (Lister et al., 1991; Liu et al., 2007; Wang et al., 2011; An et al., 2012); ostracod Sr/Ca ratios used to indicate lake salinity (Zhang et al., 1994); aeolian sand deposits used to indicate aridity changes (Fig. 2) (Lu et al., 2011; Qiang et al., 2013a); TOC content used to reconstruct lake organic productivity or terrestrial vegetation cover (An et al., 2012); and pollen assemblages used to reconstruct vegetation changes (Fig. 6B) (Shen et al., 2005b). We believe that the moisture and precipitation changes, and thus EASM evolution, can only be fully understood after a comprehensive examination of all of these proxies, as well as the separation of those proxies which reflect lake status from those which reflect terrestrial environmental changes, as discussed above.

One of the most direct proxies for regional moisture, precipitation changes and EASM monsoon evolution is vegetation change on the northeastern Tibetan Plateau. The region is situated at the modern summer monsoon limit with conditions changing from semi-arid to arid from southeast to northwest (Fig. 1A). The specific topography of the Lake Qinghai, Gonghe Basin and Qaidam Basin, surrounded by high mountains, enables the vertical migration of vegetation zones in response to climate changes. The present vertical vegetation zones consist of desert grassland in the lowest areas, such as in Gonghe Basin and Qaidam Basin, and grassland, conifer forest, alpine grassland and alpine meadow with increasing altitude (Zhou et al., 1987; Cheng, 2006; Zhao et al., 2007). The most sensitive factor to vegetation development and tree growth in the region is moisture or precipitation (Gou et al., 2014; Yang et al., 2014a), which mainly depends on the EASM intensity (Liu et al., 2002; Shen et al., 2005a, 2005b; Cheng et al., 2013). Pollen assemblages have been analyzed in detail from two sediment cores, QH2005 (Liu et al., 2002; Shen et al., 2005a, 2005b) and QH85-14C (Du et al., 1989). The reconstructed vegetation changes are similar, characterized by an increase in the percentage of arboreal pollen and total pollen concentration from the last deglacial to the middle Holocene, and a decrease in the late Holocene. Based on the adjusted chronology (Shen et al., 2005b), the highest tree pollen percentages occurred during ~5–7 ka BP following a rapid increase from the last deglacial to ~10 ka, forming an early Holocene 'plateau'; this was succeeded by an increase in the middle Holocene (Fig. 6B). In addition, the highest total and tree pollen concentration also occurred during the middle Holocene (Shen et al., 2005b). It should be noted that although the later publications (e.g., Shen et al., 2005b; Liu et al., 2007) are based on chronologies corrected for the radiocarbon reservoir effect by Shen et al. (2005b), there may be age differences of a few hundred years between cores QH-2005 (Wang et al., 2011), QH-2011 (Wang et al., 2014b) and 1Fs (An et al., 2012) because of the use of different reservoir ages (Table 1). Changes in tree pollen percentages from the sediments from Lake Dalianhai (Fig. 6C) exhibit a similar trend to those from Lake Qinghai (Fig. 6B), and further support the occurrence of a relatively dense vegetation cover and strong EASM intensity during the middle Holocene (Cheng et al., 2013). Considering the chronological uncertainties between the different lake sediment cores, the quantitative precipitation reconstructions from Lake Qinghai (Fig. 6D; Herzschuh et al., 2010) and Lake Luanhaizi in the Qilian Mountains (Fig. 6E; Wang et al., 2014c) indicate a pattern of increased precipitation from the early Holocene to the middle Holocene with the strongest precipitation during the middle Holocene. This trend is quite different to that of the orbitally-forced summer insolation changes in the northern hemisphere (Fig. 6J; Berger and Loutre, 1991), which are usually assumed to be the main driver of changes in summer monsoon (Kutzbach, 1981; COHMAP Members, 1988; Wang et al., 2001; Kutzbach et al., 2008; Cheng et al., 2009); this indicates a different driving mechanism of the EASM, as suggested by Chen et al. (2015b).

The results from other proxies from sediment cores from Lake Qinghai also support the occurrence of maximum precipitation and EASM intensity during the middle Holocene (Fig. 6). Sediment redness is related primarily to the concentration of iron oxides such as hematite. In the case of Lake Qinghai, iron oxides appear to be eroded from nearby red beds or loess deposits and transported by fluvial processes to the lake (Ji et al., 2005). Therefore, redness increases at times of increased precipitation, which is associated with increasing monsoon intensity (Ji et al., 2005; Wang et al., 2011). As illustrated in Fig. 6A, redness increased during the early Holocene, indicating increasing precipitation; however, the highest redness values occurred during the middle Holocene indicating that the highest precipitation occurred during that period. It should also be noted that the redness from two cores during the late Holocene is different, as illustrated in Figs. 5H and 6A. In addition, the highest lake levels of Lake Qinghai, Lake Hurleg, Lake Toson and Lake Gahai, as indicated by palaeoshorelines, all occurred during the middle Holocene and low lake levels during the early Holocene (Fig. 2; Fan et al., 2012, 2014; Liu et al., 2015b). In addition, the water level of Lake Qinghai, as reconstructed from several forms of evidences (Fig. 2), was low during the early Holocene but high during the middle Holocene or a little later. The reconstructed salinity of Lake Qinghai was also relatively high during the early Holocene and much fresher during the middle Holocene (Zhang et al., 1994; Li and Liu, 2014), indicating higher inflow during the middle Holocene. During the early Holocene, aeolian activity was also strong (Fig. 3), partly because of the strong westerlies, as indicated by the Lake Qinghai record (An et al., 2012), and sufficient sand supply, but possibly also indicating a relatively dry early Holocene. This dry status during the early Holocene may have been caused by high evaporation; however, the lowest isotope values of the lake water during that period, indicating limited evaporation, do not support this inference. All of the above evidence indicates a dry climate and weak EASM intensity during the early Holocene, but a strong EASM during the middle Holocene.

From the foregoing it can be concluded that some of the commonly-used proxies for changes in moisture and precipitation or EASM evolution are in conflict. Proxies for vegetation such as pollen assemblages (Du et al., 1989; Liu et al., 2002; Shen et al., 2005b); for lake level such as shorelines (Liu et al., 2015b), water depth (Lister et al., 1991; Yu, 2005; Liu et al., 2013b; Wang et al., 2014a) and salinity (Zhang et al., 1994; Li and Liu, 2014); and for river recharge intensity such as lake sediment redness (Ji et al., 2005; Wang et al., 2011); all exhibit consistent changes during the Holocene and therefore can be regarded as reliable proxies for documenting the evolution of moisture and precipitation, and thus of the EASM. The results from these proxies are consistent with the record of EASM evolution recently reconstructed from a Chinese loess-palaeosol section (Lu et al., 2013) and from a stacked record of Holocene palaeosol development in the Chinese Loess Plateau (Fig. 6G; Wang et al., 2014a) and the deserts and Sandy Lands in northern China (Fig. 6H; Li et al., 2014), which suggest that the strongest EASM occurred during the middle Holocene. A recent pollen-assemblage-based precipitation record was obtained from the sediments of Lake Gonghai in the EASM region (Chen et al., 2015b). It is a reliable record of EASM intensity and reveals a gradually intensifying monsoon from 14.7 to 7.0 ka BP with a notable decrease of precipitation during the Younger Dryas cold event and increased precipitation during the warm Bolling/Allerod interstadial, maximum monsoon strength from ~7.8–5.3 ka BP, and a rapid weakening of monsoon strength after ~3.3 ka BP (Fig. 6F, Chen et al., 2015b). These results are consistent with a Holocene EASM and precipitation reconstruction based mainly on pollen data (Shi et al., 1992), but they differ from the Chinese speleothem $\delta^{18}O$ records (cf. Liu et al., 2015a; Chen et al., 2016). The results of the commonly-used biogenic carbonate (ostracod) $\delta^{18}O$ (Fig. 4A–D), leaf wax hydrogen isotope composition (Thomas et al., 2016) and TOC content or flux records (An et al., 2012; Liu et al., 2014b) are in conflict with those of the other proxies discussed above and thus they cannot be regarded as reliable indicators of the EASM intensity. However, they do exhibit a similar trend to summer insolation and temperature changes in the northern Hemisphere (Fig. 6I and J). Therefore, it can be concluded that precipitation (moisture) in the northeastern Tibetan Plateau was dominated by EASM intensity during the Holocene; and that increased EASM strength and precipitation occurred during the early Holocene, strongest EASM and highest precipitation during the middle Holocene, and decreasing EASM and precipitation during the late Holocene.

5. Conclusion

1. Reconstructed lake levels based on OSL dating of palaeoshorelines, together with evidence from sediment cores based on the distribution of certain aquatic plants and ostracod Sr/Ca data, demonstrate that the level of Lake Qinghai was not high during the early Holocene. The highest lake shoreline clearly occurred during the middle Holocene or a little later. This conclusion is also supported by lake level records from Lake Gahai, Lake Toson and Lake Hurleg in the eastern Qaidam Basin.

2. A probability density function plot of 93 OSL ages from 39 sections in aeolian sand deposits in the northeastern Tibetan Plateau demonstrates that a peak in aeolian sand accumulation occurred during the interval from ~10–7.5 ka BP, indicating enhanced aeolian deposition and strong aeolian activity during the early Holocene.

3. Ostracod $\delta^{18}O$ records from various sediment cores from Lake Qinghai demonstrate a generally similar pattern of variation with depleted $\delta^{18}O_{ostr}$ values during the early to middle Holocene and enriched $\delta^{18}O_{ostr}$ values during the late Holocene, in agreement with the Chinese speleothem $\delta^{18}O$ record. However, both records mainly reflect changes in $\delta^{18}O_{prep}$ rather than the strength of the EASM or the intensity of EASM precipitation.

4. The Holocene climatic change history of Lake Qinghai and its environs is well documented by records of pollen assemblages, river recharge intensity indicated by pollen-based precipitation reconstructions and lake sediment redness, as well as by lake level states and aeolian activity on the northeastern Tibetan Plateau. All of these proxies, which are regarded as reliable indicators of precipitation (moisture) in the northeastern Tibetan Plateau, exhibit the same pattern of Holocene evolution as the EASM during the Holocene recorded in Lake Gonghai (Chen et al., 2015b): increasing EASM intensity during the early Holocene, strongest EASM intensity during the middle Holocene and decreasing EASM intensity during the late Holocene.

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