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**Abrupt Deglaciation on the Northeastern Tibetan Plateau: Evidence
from Lake Qinghai**

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Key words: Lake Qinghai, Climate history, LGM, Sediment analyses, TOC, Carbonate, XRF

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Abstract

Post glacial climate history in Central Asia was reconstructed using sediments from the core QH07 taken from the southeastern basin of Lake Qinghai, situated on the northeastern margin of the Tibetan Plateau. Results from multiple environmental indicators are internally consistent and reveal a distinct Late Pleistocene and Holocene climate record.

Carbonate content and total organic carbon (TOC) in Lake Qinghai sediments are interpreted as indicators of the strength of the Asian summer monsoon: warm and wet intervals associated with increased monsoon strength are indicated by increased carbonate and TOC content. During the glacial period (~14,600 to ~20,000 yr B.P.), the data suggest that summer monsoon intensity remained low and relatively constant at Lake Qinghai, suggesting cool, dry, and relatively stable climatic conditions. The transition between the Holocene and the late Pleistocene, about 11,500 years ago, was abrupt, more so than indicated by speleothems in eastern China. The inferred stable, cold, arid environment of the Last Glacial Maximum (LGM) seems mostly to persist through the Younger Dryas time period, and evidence of a warm interval correlative with the Bølling–Allerød oscillation is weak in the QH07 record. The Holocene (~11,500 yr B.P. to present) was a time of enhanced summer monsoon strength and greater variability, indicating relatively wetter but more unstable climatic conditions than those of the late Pleistocene. The warmest, wettest part of the Holocene (marked by maximum organic matter and carbonate contents) occurred from ~9,000 to ~11,500 yr B.P., consistent with maximum summer insolation contrast between the continent and the ocean. A gradual reduction in precipitation (weakened summer monsoon) is inferred from decreased carbonate content through the course of the Holocene.

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4 We propose that changes in the contrast of summer insolation between the continent and
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6 the ocean are the primary control on the Asian monsoon system over glacial/interglacial time
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8 scales. Secondary influences may include albedo changes due to ice-cover and vegetation shifts
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10 and sea level changes (distance from moisture source in Pacific Ocean). The abruptness of the
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12 change at the beginning of the Holocene, combined with an increase in variability, suggest a
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14 threshold for the arrival of monsoonal rainfall at the northeastern edge of the Tibetan Plateau.
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23 **Introduction**

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28 The Asian monsoon system (the Indian and East Asian monsoons) is of crucial importance not
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30 only because of its key role in global climate, but also because it provides the rainfall that
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32 sustains more than half of the world's population, as well as vast natural ecosystems. Knowledge
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34 of the climate history of the northeast Tibetan Plateau, at the inland edge of monsoon influence,
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36 will lead to a better understanding of the dynamics of the East Asian monsoon, as well as the role
37
38 of the Tibetan Plateau in the climate system.
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43 The Tibetan Plateau has long been considered as a major control on monsoonal
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45 circulation over Asia, as suggested by the contrast between climate models that include and
46
47 exclude the topography of the plateau (Kutzbach et al. 1989). The classical explanation of the
48
49 impact of enhanced heating of the atmosphere over Tibet on the Asian monsoons during spring
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51 and summer (Kutzbach et al. 1989; Clemens et al. 1991; Molnar et al. 1993; An 2000) may be an
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53 oversimplification. The dynamics of the Asian monsoon may also be affected most by the barrier
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55 posed by the plateau in the path of the subtropical jet stream (Molnar et al. 2010). Other factors
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4 may include orographic insulation, by which the height of the Himalayas on the south margin of
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6 the plateau blocks moisture exchange between the tropics and mid-latitudes (Boos and Kuang
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8 2010; Molnar et al. 2010). Whatever the exact mechanisms involved, the presence of the Tibetan
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10 Plateau clearly affects the distribution of precipitation in southern and eastern Asia.
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14 Oxygen isotopes of cave speleothems within the monsoonal region have been widely
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16 used as an indicator of the intensity of the East Asian monsoon (Wang et al. 2001; Yuan et al.
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18 2004; Wang et al. 2008; Zhang et al. 2008). These cave records generally reveal a relatively
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20 weak East Asian monsoon during the glacial period and a strong monsoon in the Holocene,
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22 following changes in the Northern Hemisphere summer insolation on the orbital timescale.
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24 Millennial-scale climate events, such as the Bølling–Allerød, the Younger Dryas, and the 8.2 ka
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26 cooling event, are clearly recorded in these records. Numerical climate simulations with an
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28 embedded oxygen-isotope model suggests that changes in oxygen isotopic compositions in the
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30 Chinese caves (including Hulu and Dongge) reflect changes in the intensity of the South Asian
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32 monsoon precipitation rather than East Asian monsoon (Pausata et al. 2011).
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38 Lake sediments properties can serve as proxies for past water temperature, biology, and
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40 chemistry, which in turn provide information about climatic variations, glacial/interglacial
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42 history, vegetation alternations, and their connections with global change (Colman et al. 1995;
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44 Johnson et al. 2002; Henderson et al. 2010; Zhao et al. 2010). For instance, sedimentary records
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46 from relatively small lakes on the northeastern Tibetan Plateau such as Hurleg Lake, Toson
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48 Lake, and Gahai Lake, derived from lithology, magnetic mineralogy, fossil pollen, carbonate
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50 isotopes, ostracode shell isotopes and trace elements, X-ray fluorescence (XRF), and gray scale
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52 data, reconstructed the climate history that extends back to the early Holocene (Zhao et al. 2009;
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54 Zhao C et al. 2010; Zhao et al. 2010). Paleoenviromental records from other lakes on the plateau
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4 have been reviewed by Herzschuh et al. (2006) and Chen et al. (2008). Sediments from larger
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6 inland lakes on the northeast Tibetan Plateau potentially provide long (extending back to LGM
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8 or even further), continuous, and high-resolution paleoclimate records, applicable for various
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10 time scales.
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14 Lake Qinghai, a closed basin, at an elevation of 3194 m, is situated on the northeastern
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16 margin of the Tibetan Plateau. The drainage area of the lake is about 30,000 km² (LZCAS 1994),
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18 from which five major rivers discharge to the lake basin. The Buha River, on the western side of
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20 the lake contributes more than 60% of the total runoff. Its drainage area is dominated by Late
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22 Paleozoic sedimentary rocks (marine limestones and sandstones). Triassic volcanics and
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24 Mesozoic granodiorites are common in catchments of smaller tributaries (Yan et al. 2002; Yu
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26 and Zhang 2008). The surrounding mountain glaciers are generally small in size, from which the
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28 melt water only accounts for 0.3% of the total inflow to the lake; thus, the influence of melt
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30 water on the water budget is relatively small. Although there is no surface outflow, groundwater
31
32 through-flow may be substantial (Jin et al. 2010).
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38 The lake has a surface area of 4400 km², and an average water depth of 21 m, with a
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40 maximum depth of 27 m. It is composed of three sub-basins based on its bathymetry: the
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42 northern basin, the southern basin and the southeastern basin. Mean annual precipitation in this
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44 region is ~360 mm, most of which falls in summer, while the potential evaporation is up to 1000
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46 mm per year at the lake surface. About 75% of the total runoff occurs in the summer time.
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48 Annual mean temperature is -0.7°C and monthly means exhibit remarkable seasonality, varying
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50 from -11°C in winter to 12°C in summer (Lister et al. 1991; Shen et al. 2005; Henderson et al.
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52 2010).
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4 The lake water is brackish (average salinity = 15.5 g/l; pH = 9.06), dominated by sodium,
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6 magnesium, potassium, chlorine, carbonate, and sulphate ions, mainly from weathering in the
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8 catchment of the Buha river, partly from atmospheric deposition (7.4-44%), and only modestly
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10 from groundwater input (3.7-4.7%) (Jin et al. 2010). The lake is supersaturated with respect to
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12 carbonates including aragonite, calcite, and dolomite (Sun et al. 1991; Yu and Zhang 2008).
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14 High Mg/Ca (61-100) promotes aragonite to precipitate directly from the lake water. The entire
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16 water column is oxic but pore water becomes suboxic a few centimeters below the
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18 water/sediment interface.
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24 Climatically, Lake Qinghai area is considered to be at the intersection of the Asian
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26 summer monsoon and the Westerlies on the Tibetan Plateau. Its unique geographical setting
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28 makes Lake Qinghai sensitive to environmental and climatic changes and thus well-suited for
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30 study of the Asian monsoon and the Westerlies at various time scales (Henderson et al. 2010; An
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32 et al. 2012).
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36 Many studies of Lake Qinghai sediments and their sources, using a variety of analytical
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38 methods, have been conducted (Colman et al. 2007; Henderson and Holmes 2009, and references
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40 therein). However, only a handful of studies (Lister et al. 1991; Shen et al. 2005) have applied
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42 multiple biogeochemical analyses of the sediments of Lake Qinghai to interpret regional climatic
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44 and environmental changes, suggesting a cold-dry glacial time, a moist early Holocene during
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46 the insolation and monsoon maximum followed by a drying trend during the mid- and late –
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48 Holocene. However, most of these studies had relatively limited chronological constraints.
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50 Radiocarbon dating of Lake Qinghai sediment is difficult because macro-organic matter is scarce
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52 and because of carbon reservoir effects in the lake (An et al. 2012). Moreover, the climate
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54 pattern reconstructed from pollen record in Hurleg Lake, 300 km west of Lake Qinghai, appears
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to be opposite to the climate pattern inferred at Lake Qinghai (Zhao et al. 2007). A few studies have also attempted to reconstruct the lake-level history of Lake Qinghai (Lister et al. 1991; Madsen et al. 2008; Liu et al. 2010b; Rhode et al. 2010; Liu et al. 2011; Liu et al. 2012). Lister et al. suggest substantial lake-level rises during the periods approximately 10,200-9800, 9500-8500, 8300-7200 yr B.P. and decreasing levels probably during the intervening intervals. But to what extent the lake level was higher or lower than today is yet not fully understood. Thus, several scientific questions remain in the Lake Qinghai area: What was the glacial-interglacial climate history in this area, in terms of changes in the Asian Summer monsoon? Was the transition from the late-glacial to the Holocene gradual or sharp? Were abrupt global climate events such as the Bølling/Allerød oscillation and the Younger Dryas, recorded in the sediments? When did the Holocene climatic optimum occur? What controlled climatic variations on different time scales, e.g., orbital or millennial? We aim to address these questions through so-called multi-proxy investigations of Lake Qinghai sediments.

Materials and methods

Sediment cores and sampling

As a complementary study to the Lake Qinghai Drilling Project (LQDP, An et al. 2006), a 3.55-m-long Uwitec piston core (QH07-1A) and an 85-cm-long mini-Mackereth core (QH07-1B-1MM) were collected from the southeastern basin of Lake Qinghai, at LQDP site 2 (36° 43' 36.7" N, 100° 29' 28.1" E; Fig. 1). The top 15 cm of QH07-1B-1MM was soupy and was extruded and sampled at 0.5 cm intervals in the field. In the laboratory, discrete samples were

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4 taken continuously at 1.5 cm intervals from the mini-Mackereth core and at 3 cm intervals from
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6 the Uwitec core.
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10 11 **Analytical methods** 12 13

14 15 16 *Magnetic susceptibility and bulk density analyses* 17

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19 Measurements of magnetic susceptibility at 1 cm intervals were performed on split cores using
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21 an automated core logger (Geotek MSCL-XYZ). The susceptibility reading was reported as
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23 volume-normalized susceptibility in units of 10^{-6} SI. In addition, the split cores were scanned
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25 using a Geotek MSCL-S to measure gamma-ray transmission, which provided an estimate of
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27 bulk density (g/cm^3), an indicator of lithology and porosity.
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34 35 *Bulk geochemical element analysis* 36

37 Bulk elemental compositions of the sediments in split cores were analyzed using an ITRAX X-
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39 ray Fluorescence (XRF) Core Scanner (Cox Analytical Instruments). The measurements were
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41 taken at 2-mm intervals with a sampling time of 90 seconds using a molybdenum X-ray source
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43 set to 30 kV and 30 mA. Initial XRF spectral data were reprocessed in a standard fitting
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45 procedure using Q-Spec spectral analysis software to refine the individual elemental peak areas
46
47 Croudace et al. 2006. Final XRF results were reported as counts per second (cps), representing
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49 relative element concentrations.
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56 57 *Carbonate analysis* 58 59 60 61 62 63 64 65

Total inorganic carbon in the sediments was measured by coulometry. Samples were freeze-dried and ground into fine, homogeneous powder and desiccated overnight. Carbon dioxide gas evolved by reaction of carbonates in the sample with acid (30 mg sediment and 5 ml 2N HCl) was recorded as micrograms of carbon. Blanks and standards were run between every 10 samples. The amount of carbonate was calculated by assuming all inorganic carbon was present as calcium carbonate.

Analysis of TOC, TN and their isotopic ratios

Samples were freeze-dried, ground into fine powder, oven-dried at 60°C overnight, and treated by acid fumigation (HCl) for 6 hrs to remove carbonate in the sediments. Fumigated samples wrapped in tin capsules were placed in a desiccator until EA-IRMS (Elemental Analysis-Isotope Ratio Mass Spectrometry) analyses. TOC and TN were reported as weight percentages. C/N weight ratios were converted to atomic ratios. Carbon and nitrogen isotopic ratios of the organic matter were determined by a Finnigan Delta Plus XP isotope-ratio mass spectrometer. The isotope ratios were reported in δ values relative to the Vienna PeeDee Belemnite standard (VPDB), using the equation $\delta = [(R_{\text{sample}} - R_{\text{standard}}) / R_{\text{standard}}] * 1000$.

Grain size analysis

About 1 cm³ wet bulk sediment was pretreated with 50 ml 30% H₂O₂ and then with 20 ml 10% HCl solution in a water bath (85°C), to remove organic matter, carbonate, and iron oxides in sediments (Konert and Vandenberghe 1997). Then, 10 ml sodium hexametaphosphate solution (5g/L) was used to disperse grains. Just before analysis, the sample tubes were placed in a shaker

bath; then analyzed on a Beckman Coulter LS 13 320 Particle Size Analyzer. Grain-size distribution was reported in 93 size classes between 0.375 and 2000 μm .

Chronology

Based on field observations, the mini-Mackereth core sampled the sediment-water interface. Comparison of magnetic susceptibility and elemental chemistry profiles among different cores at this site allows construction of a composite depth scale for the cores. The composite depth scale indicates a gap of about 2 m between the mini-Mackereth and the Uwitec core, probably due to overpenetration by the latter. The age model for the composite record (QH07) was established on the basis of 17 AMS ^{14}C ages (5 ages from QH07-1B-1MM and 12 ages from QH07-1A) (Table 1). Two of the radiocarbon ages were from samples of seeds of the aquatic plant *Ruppiaceae*. The rest were determined on bulk organic matter due to the lack of terrestrial plant macrofossils in the sediments. In this and in previous studies (Shen et al. 2005; Henderson et al. 2010; An et al. 2012), radiocarbon ages on *Ruppiaceae* seeds are indistinguishable from those on bulk organic matter at the same horizon. *Ruppiaceae* seeds occur irregularly in the core, not in any concentrated pattern that might indicate shallow water.

Radiocarbon ages of aquatic material need to be corrected for lake reservoir effects, resulting from input of old inorganic carbon, mobilized by chemical weathering of carbonate rocks in the watershed. Ages for bulk organic matter may also be affected by the residence time of terrestrial organic matter in the drainage basin before its final sequestration in the sediments. Previous studies have shown that the organic matter in the Lake Qinghai sediments is mainly aquatic (Shen et al. 2005; Yu and Zhang 2008), consistent with our measurements of radiocarbon

ages of *Ruppiaceae* seeds and bulk organic matter. This implies that terrestrial residence times will have only a negligible influence on bulk radiocarbon ages.

Previous estimates of the reservoir effect in Lake Qinghai range from 135 to 2500 years (Shen et al. 2005; Yu et al. 2007; Henderson et al. 2010; Hou et al. 2010; An et al. 2012). Henderson et al. (2010) derived values for the modern reservoir effects using two different methods. They obtained values of 661, 640, and 672 years for ^{14}C ages of lake water and multiple fraction of surface sediment, including bulk organic carbon and authigenic carbonate respectively. They also obtained an age of 737 years using a linear regression through ages at depth in a core, extrapolating that relation to the sediment surface. The consistency of these values determined by Henderson et al. (2010) lends confidence to their values. Therefore, we use a reservoir age of 700 yrs and make the assumption that this value has had no temporal variability. A constant reservoir effect is an oversimplification, but a conservative one is given definitive evidence to the contrary. Geochemical models (Yu et al., 2007) suggest that fluctuations in the magnitude of the reservoir effect as large as those suggested by An et al. (2012), 135 to 2500 years, are unlikely.

The age model (Fig. 2) was generated using CLAM software ('classical' age-depth modeling) based on probability sampling methods that take into account multi-modal and asymmetric distributions of calibrated radiocarbon dates (Blaauw 2010). Followed guidance of Bpeat software (Blaauw and Christen 2005), five out of the 17 AMS ^{14}C ages were considered outliers, either because they created apparent age reversals, were far removed from regression fits, or were incompatible with multiple age estimates from other levels close by in the core. The reason for the outliers is not known, but possible causes include contamination of samples by younger carbon, transport of carbon older than the sediments to the lake, and short-term

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4 fluctuations in sedimentation rate. Ages were weighted inversely with the error in their calibrated
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6 probabilities. The smooth-spline method in CLAM was adopted with a smoothing level of 0.4.
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8 This age modeling produced a 'best' age-model (solid line in Fig. 2) as well as uncertainty ranges
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10 (95% confidence intervals, dashed lines in Fig. 2) by finding the highest posterior probability
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12 density of the array of age-depth curves. Confidence intervals were calculated at 2 standard
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14 deviation (95%) level based on 10,000 iterations.
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19 The age model generated a reservoir corrected age of 57 yr BP for surface sediment of
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21 QH07-1B-1MM and provides a robust fit to the individual calibrated ages, particularly at the
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23 glacial/Holocene transition, and indicates that our sedimentary record reaches back close to 20 ka
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25 B.P. (Fig. 2). Sedimentation rates inferred from this age model for QH07 varied through the
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27 sequence, from about 0.37 mm/yr in the Holocene to a rate of ~0.14 mm/yr during the late
28
29 glacial period. Sedimentation rates derived from our age-depth model for QH07 are consistent
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31 with those for QH85-14B (from the south basin, Holocene sedimentation rate, 0.3-0.4 mm/yr)
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33 (Lister et al. 1991), QH-2000 (from the southeastern basin, Holocene sedimentation rate, 0.32-
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35 0.64 mm/yr) (Shen et al. 2005) and LQDP05-1F (from the south basin, Holocene sedimentation
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37 rate, 0.43 mm/yr) (An et al. 2012). Overall, the south basin cores have higher accumulation rates
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39 than those from the southeastern basin during the glacial time: ~0.67 mm/yr in LQDP05-1F and
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41 ~0.5 mm/yr in QH85-14B, versus 0.29-0.77 mm/yr in QH-2000 and ~0.14 mm/yr in QH07
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43 (Lister et al. 1991; Shen et al. 2005; Jin et al. 2010; An et al. 2012).
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53 **Results**

54 55 56 57 58 Sediment lithological properties

Water content, density and Cl/Ti

Water content shows a sharp change at the Late Pleistocene/Holocene transition, at ~11.5 ka, increasing from 20% to 50% (Fig. 3). The density data are in agreement with the water content record, exhibiting low values (~1.7 g/cm³) in the Holocene and high values (~2.3 g/cm³) in glacial times. The effects of compaction below ~15 cm (upper watery sediments) are not apparent.

Carbonate content (Ca/Ti)

The carbonate content profile is similar to the water content pattern (low in the late Pleistocene and high in the Holocene) and carbonate varies from about 20% to 60% (by weight) throughout the core (Fig. 3). During the glacial period, carbonate content is low and remains relatively stable, fluctuating in a narrow range between 20% and 28%. Early to mid-Holocene sediments have the highest concentrations and the most variability (as low as 30% and as high as 60%, average ~45%). Late Holocene sediments contain ~30% carbonate, consistent with variations in four short cores from the lake, which vary between 20-30% (Zhang et al. 2003). The carbonate minerals in the sediment of Lake Qinghai are mainly aragonite with minor calcite and dolomite (Shen et al. 2005). In addition, nodular siderite (FeCO₃) was identified in smear slides, but only at a few intervals in Holocene sediments. Therefore, the Ca/Ti calculated from XRF data should indicate the relative abundance of carbonate. The Ca/Ti profile indeed correlates well with the carbonate content determined by coulometry (Fig. 3).

Organic geochemical measurements

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7 *TOC, Inc/coh, and TN*
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9 TOC values in QH07 (Fig. 4) are broadly consistent with those in other cores from Lake
10 Qinghai, including LQDP05-1F (An et al. 2012) and QH86-14B (Yu and Zhang 2008) but are
11 much lower than those in QH2000 (Shen et al. 2005). TOC in our record is relatively low
12 (~0.6%) and relatively constant during the glacial period, suggesting low primary production in
13 the lake. An abrupt change at ~11.5 ka is marked by a change to much higher TOC (~2% on
14 average, but up to 3%), presumably resulting from increased biological productivity and greater
15 riverine input of terrestrial organic matter.
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26 Measurements of incoherent and coherent scattering were produced during XRF
27 scanning. Inc/coh can be used as a qualitative indicator of total organic matter concentrations
28 because it is dependent on the average atomic number of material in the sediment (Burnett et al.
29 2011; Liu 2011b), provided that the inorganic composition of sediments is relatively invariant.
30 Inc/coh ratios from QH07 cores appear to correlate well with TOC (Fig. 4).
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38 The TN profile closely follows the TOC profile, displaying high values (~0.2%) in the
39 Holocene, and low values (~0.07%) during glacial times. This pattern suggests higher primary
40 productivity in the lake during the Holocene relative to the low production in the Late
41 Pleistocene.
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50 *C/N*
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52 C/N ratios in QH07 are relatively constant. The late Pleistocene shows lower C/N (<10) than
53 does the Holocene (average about 12, with values of as much as 17).
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$\delta^{13}C_{OM}$

The $\delta^{13}C_{OM}$ values in QH07 range from -28‰ to -20‰, comparable with down core $\delta^{13}C_{OM}$ records in previous studies (Shen et al. 2005; Henderson and Holmes 2009). During glacial times, $\delta^{13}C_{OM}$ varies slightly and averages ~-24‰. It drastically shifts to a more positive value (as much as -20‰) in early Holocene, gradually decreases to a more depleted level of about -24‰ during middle Holocene, and decreases to a little more negative values (~-27‰) during late Holocene, probably due to a decreasing portion of terrestrial input.

$\delta^{15}N_{OM}$

The $\delta^{15}N_{OM}$ in QH07 falls in the range of +3‰ to +10‰, and it shows relatively constant values (~5‰) during glacial times, suggesting low aquatic productivity. The $\delta^{15}N_{OM}$ values in the Holocene are higher and more variable on average than that in the glacial period, indicating relatively higher lake production and more complex sources.

Terrigenous sediment properties

Si/Ti, K/Ti

The biogenic silica content of the Lake Qinghai sediments is very low. Measurements of biogenic silica were attempted by Liu et al. (2010), but preserved diatom frustules in the sediments have not been reported (Liu et al. 2010a).

Grain size

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4 In QH07, a typical Holocene-age sediment grain-size distribution has three major modes,
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6 centered at 9 μm , 28 μm , and 160 μm , and two minor modes, centered at 80 μm and 500 μm ,
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8 while a typical glacial-age sediment grain-size distribution has only the three major modes seen
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10 in the Holocene. The mean grain-size profile shows similar trends as that for the content of the
11
12 $>116 \mu\text{m}$ fraction. Coarser and more variable materials of the Holocene may reflect a more
13
14 complex and variable combination of sources, including aeolian dust, riverine load, and
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16 longshore current transport, compared to sites in the deeper south basin of the lake.
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23 *Magnetic susceptibility and Fe, and Ti content*

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25 Magnetic susceptibility, iron, and titanium profiles from QH07 exhibit parallel changes. The
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27 correlation coefficient (R) between Fe content and susceptibility is 0.76. This positive correlation
28
29 indicates that the magnetic susceptibility reflects the relative concentration of Fe-bearing
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31 magnetic minerals in Lake Qinghai. Magnetic susceptibility values, Fe content, and Ti content
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33 are higher in pre-Holocene sediments than in the Holocene. The relatively low susceptibility
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35 values were found in the Holocene.
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43 **Discussion**

44 45 46 47 48 49 Interpretation of individual paleoenvironmental indicators

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51 The Cl/Ti profile based on XRF measurements also is consistent with water content and density
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53 profiles, as expected, since most Cl is present as ions in pore water. Increased water content, high
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55 Cl/Ti, and relatively low density, all reflect the increased porosity in Holocene sediments, which
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57 in turn implies a different bulk composition.
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Carbonate precipitation in lakes (provided that CaCO_3 is the dominant carbonate) can be controlled by several climate-related factors (Talbot and Kelts 1986). First, carbonate precipitation can occur by evaporative concentration in lake water, which may cause the products of Ca^{2+} and CO_3^{2-} ion activities exceed saturation. Under these conditions, high carbonate content implies relatively dry climate conditions. Second, the solubility of calcium carbonate decreases with temperature, so that sedimentary carbonate content is enhanced by warm water. Third, photosynthesis and respiration can affect the equilibrium of the carbonate system through losses and gains of dissolved CO_2 . High primary production in lake system removes aqueous CO_2 in the process of photosynthesis, increasing pH and promoting carbonate precipitation. Fourth, if one of the ions, either the cations (Ca^{2+}) or the anions (CO_3^{2-}) are present in excess of saturation, the other becomes limiting to carbonate precipitation in the lake system. The addition of the limiting ion would result in enhanced carbonate precipitation and rapid removal of the limiting ion.

The Ca^{2+} concentrations in modern Qinghai lake water are low (~11 ppm), compared to that in major input rivers (~58.8 ppm in Buha River; ~69.6 ppm in Heima River; ~56.8 ppm in Shaliu River). In contrast, total dissolved inorganic carbon concentration in the lake is relatively high (~291 ppm) (Jin et al. 2010). Results from MINEQL+ 4.6 (Schecher and McAvoy 1992), show that the lake water is saturated with respect to calcite and near saturated with respect to aragonite. Moreover, carbonate content in recent sediments covaries with the observed water discharge of the Buha River, the largest input river for Lake Qinghai, for the last 50 yrs (An et al. 2012). Thus it appears that the carbonate chemistry of the lake is calcium limited, so that carbonate precipitation closely relates to Ca^{2+} delivered by river runoff, which in turn relates to precipitation in this region. Because Lake Qinghai apparently was always a closed basin, this

relationship applies to our entire record. Therefore, carbonate content (and/or Ca/Ti) in Lake Qinghai sediments can be used as an indicator of river runoff. In this way, Lake Qinghai behaves much the same as Mono Lake (Zimmerman et al. 2011) another closed, alkaline lake in a semi-arid environment.

Although we interpret runoff as the primary control (through calcium limitation) on carbonate sedimentation in Qinghai Lake, other factors may have secondary effects. Changes in lake level may affect carbonate flux by delivery of river-borne silicate in nearshore shallow water (Zhao C et al. 2010). Also, water temperature may have an effect, although, given the likely amplitude of glacial-interglacial temperature change, the variation in calcium carbonate solubility is small compared to the change in carbonate content in Lake Qinghai sediments. Finally, although primary productivity in Lake Qinghai increases from the Late Pleistocene to the Holocene, it is still relatively low even in the Holocene, as judged from TOC contents. Increased productivity and water temperature in the Holocene would add to the effect of increased Ca^{2+} delivery to the lake in increased runoff.

Although organic carbon may be consumed by respiration or oxidation during early diagenesis, TOC content in sediments generally reflects lake primary productivity and the input of terrestrial organic matter. Both lacustrine production and terrestrial inputs are likely linked to increased precipitation and warm temperature, probably because fresh water input decreases lake water salinity and thereby enhances aquatic productivity (Henderson and Holmes 2009). Thus, we suggest that TOC, through a link to effective moisture and terrestrial and aquatic productivity, is an indicator of monsoonal precipitation.

The C/N ratio of organic matter is extensively used as a source indicator of organic matter in terms of original proportions of aquatic and terrestrial material, despite diagenetic

alteration (Meyers 1994). Aquatic plants have low C/N, typically in the range of 4-10, whereas land plants have C/N of 20 and greater (Meyers and Ishiwatari 1993; Henderson and Holmes 2009). Higher C/N suggests increased contribution of organic matter from land plant debris and/or aquatic macrophytes in the shallow parts of the lake. The results suggest that organic matter deposited during glacial times was mainly from aquatic sources, whereas organic matter in the Holocene sediments was a mixture of aquatic and terrestrial material, but mostly aquatic.

The stable isotopic composition of bulk organic matter ($\delta^{13}\text{C}_{\text{OM}}$) in sediments is affected by the source material, fractionations during carbon assimilation, metabolism, and synthesis, and microbial reworking of organic matter during early diagenesis (Hayes 1993; Meyers and Ishiwatari 1993; Werne and Hollander 2004). $\delta^{13}\text{C}$ of land-derived organic matter depend on their distinct photosynthetic pathways: C_3 , C_4 and CAM. The carbon isotope range of C_3 plants is -22‰-35‰, with a mean of ~-28‰; that of C_4 is -9‰-16‰, with a mean of ~-14‰. CAM plants have a large range of $\delta^{13}\text{C}$ that covers the range of the C_3 and C_4 pathways (DeNiro 1987; O'Leary 1988). $\delta^{13}\text{C}$ of aquatic organic matter is mainly affected by the $\delta^{13}\text{C}$ of the dissolved inorganic carbon ($\delta^{13}\text{C}_{\text{DIC}}$) in lake water due to aquatic plants preferentially taking up ^{12}C during photosynthesis and leave the remaining dissolved inorganic carbon pool enriched in ^{13}C . Enhanced primary productivity thereby leads to a more positive $\delta^{13}\text{C}_{\text{OM}}$ (Meyers and Lallier-Verges 1999). However, carbon source, whether $\text{CO}_2(\text{aq})$ or H_2CO_3 , is also important because the two species have different fractionation factors.

Interpretation of the nitrogen isotopic composition of organic matter ($\delta^{15}\text{N}_{\text{OM}}$) is complex (Schulz and Zabel 2009). Similar to $\delta^{13}\text{C}_{\text{OM}}$, $\delta^{15}\text{N}$ of aquatic source is thought to be associated with primary productivity (Meyers and Ishiwatari 1993). Despite complications in interpreting $\delta^{15}\text{N}_{\text{OM}}$, we suggest that low-frequency variations in $\delta^{15}\text{N}_{\text{OM}}$ record reflect the aquatic

production, whereas high-frequency variations imply relative proportion of terrestrial vs. aquatic sources (Fig. 5e).

In general, all the organic geochemical measurements suggest low primary production in the lake during the glacial period, highest productivity in early Holocene, and decreasing productivity into the late Holocene. TN shows a roughly linear relationship with TOC (Fig. 5a). The C/N ratio shows linear correlation with TOC and $\delta^{13}\text{C}_{\text{OM}}$ during glacial times, but not in the Holocene (Fig. 5b, 5d). A cross plot of $\delta^{13}\text{C}_{\text{OM}}$ vs. TOC (Fig. 5c) shows that the $\delta^{13}\text{C}_{\text{OM}}$ values in glacial sediments have a narrow range, while the $\delta^{13}\text{C}_{\text{OM}}$ data in Holocene sediments are rather scattered. We conclude that during glacial times, the $\delta^{13}\text{C}_{\text{OM}}$ not only reflects the $\delta^{13}\text{C}$ values of algae (C/N <10, aquatic source), but also indicates low primary productivity. More positive values of $\delta^{13}\text{C}_{\text{OM}}$ in the early Holocene primarily indicate enhanced lake production. The variations of $\delta^{13}\text{C}_{\text{OM}}$ during the Holocene likely are caused by the introduction of a small but variable proportion terrestrial organic matter. The sources of the terrestrial organic matter during early-, mid-, and late-Holocene might have been differed in response to vegetation change.

The Si/Ti ratio, therefore, likely reflects relative quartz content, i.e., high Si/Ti suggests high quartz contents. Because K is mobilized more readily than Ti during chemical weathering, sedimentary K/Ti reflects the degree of weathering of detrital sediment, i.e., high K/Ti suggests less-altered, fresher sediment sources. In QH07, intervals of high Si/Ti and low K/Ti during the glacial period are followed by episodes of low Si/Ti and high K/Ti during the Holocene, which indicates that less quartz and fresher silicates are present in Holocene sediments compared to glacial-age sediments. The greater variability during the Holocene in Si/Ti and K/Ti may suggest multiple sediment sources.

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4 The grain size of aeolian sediment has been used to reconstruct past variations of the
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6 Asian winter monsoon on Loess Plateau (Porter and An 1995), in the North Pacific (Rea et al.
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8 1998), and in Lake Biwa (Xiao et al. 1997). However, terrigenous sediment in most lakes is
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10 made up of materials from a combination of sources including both aeolian and riverine load. By
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12 assuming multiple end members present in terrigenous sediments, the aeolian component
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14 potentially can be separated from the riverine component.
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19 An et al. (2012) suggested that the $<25\ \mu\text{m}$ fraction in Lake Qinghai sediments is derived
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21 mostly from riverine input, whereas the $>25\ \mu\text{m}$ fraction is primarily aeolian dust. This
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23 interpretation was based on observations of distinct grain-size characteristics in modern dust,
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25 loess, surface sediments, and riverine suspended particles. These results show high proportions
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27 of $>25\ \mu\text{m}$ sediments in glacial time and less coarse material during the Holocene, suggesting
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29 weakened westerlies in the Holocene (An et al. 2012).
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34 However, the grain-size record from QH07 (in the southeastern basin) appears quite
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36 different from that of LQDP05-1F in the southern basin. Compared with LQDP05-1F, the coarse
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38 fraction ($16\text{--}116\ \mu\text{m}$) in the size record from QH07 cores shows the opposite trend, i.e. higher
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40 contents of coarse fraction in the Holocene than in the glacial. Therefore, the coarse fraction in
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42 QH07 can not be simply interpreted as an aeolian signal.
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47 The southeastern basin is shallower than the southern basin and QH07 is closer to the
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49 lake shore than LQDP05-1F. QH07 is also located offshore from the Erlangjian terrace, a hook-
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51 shaped sand spit prograding into the southeastern basin of Lake Qinghai. Therefore, the coarse
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53 fraction in QH07 record may not simply reflect aeolian transport, but may relate to local
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55 processes involving erosion or progradation of the spit. The content of the $>116\ \mu\text{m}$ fraction (fine
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57 sand) increased during the Holocene compared to the late glacial. This may indicate higher lake
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4 levels during the Holocene, which would result in greater fetch. Since the fetch of the lake
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6 determines the size of waves (Wetzel 2001), larger waves would occur in the Holocene,
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8 transporting coarse grains and cause more wave-cut erosion of the spit to the southeastern basin
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10 (QH07).
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14 Medium/coarse sand (500- μ m-size) sporadically occurs in a few Holocene-age levels
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16 within the core. This seems anomalously coarse for this offshore site. The sand may have been
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18 frozen into nearshore ice in the winter. During ice break-up fragments may have rafted to
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20 offshore locations where ice melted and dropped the coarse sand. Another possibility is that the
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22 coarse sand was blown by winter winds onto the surface of the ice and was released offshore
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24 during spring melt and incorporated into the sediments, which are otherwise fine grained.
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28 The magnetic susceptibility data suggest that the relative input of detrital material was
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30 reduced in the Holocene. These relatively low susceptibility values in the Holocene result from
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32 the combined effect of higher concentrations of paramagnetic and diamagnetic materials as well
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34 as dilution with authigenic carbonate.
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38 39 40 41 Climatic implications

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43 Paleoclimate studies on the Tibetan Plateau have been focusing on various time scales (An 2000;
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45 Chen et al. 2001; Zhao et al. 2009; Chen et al. 2010; Zhao et al. 2010). Chen et al. (2008)
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47 synthesized climatic records from lake sediments at many sites in central Asia. They found a
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49 distinct difference between records from the northwest and the southeast of Tibetan Plateau in
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51 moisture history during the Holocene. The former showed a dry early Holocene, a less dry mid-
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53 Holocene, and a moderately wet late Holocene, whereas the latter, monsoonal region showed a
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55 humid early Holocene and a relatively dry late Holocene (Chen et al. 2008). Herzschuh (2006)
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4 reconstructed moisture conditions in central Asia during the last 50 ka using 75 paleoclimatic
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6 records. According to this synthesis, the middle and late marine isotope stage 3 was wet, while
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8 the LGM was dry. Following the LGM, relatively wet climate conditions returned. Among the
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10 75 records, some records from the southeastern Tibetan Plateau show that, the warmest, wettest
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12 Holocene climate conditions occurred during the early Holocene. Others (in north-western and
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14 north-central China and in Mongolia), however, show a mid-Holocene, rather than the early-
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16 Holocene temperature maximum. The majority of these paleoclimatic records suggest reduced
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18 moisture since the mid Holocene (Herzschuh 2006).
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24 In general, our lacustrine sediment record QH07 is consistent with what is known about
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26 the climate history in eastern China since the Younger Dryas cold interval, that wetter conditions
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28 in the early Holocene and drier conditions in the mid and late Holocene. The multiple results we
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30 measured in the QH07 core from Lake Qinghai are compatible with each other and capture a
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32 distinct Holocene and latest Pleistocene climate record. The major first-order change in the
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34 record occurs at the transition between the Holocene and the latest Pleistocene, about 11,500
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36 years ago, temporally correlative with the end of the Younger Dryas.
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41 Our interpretation that the carbonate content (also Ca/Ti) in Lake Qinghai sediments
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43 closely relates to river runoff allows us to use carbonate content as an indicator of the strength of
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45 the Asian summer monsoon, assuming a direct relation between precipitation and runoff.
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47 Another index of the summer monsoon strength is TOC, because TOC reflects lake primary
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49 productivity and/or contribution of land-derived organic matter, probably both. In addition to
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51 percentages, we calculated the fluxes of total organic matter and carbonate, respectively. These
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53 two indicators of the summer monsoon intensity (Fig.7) remain low and relatively constant
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55 during the glacial period (~14,600 to ~ 20,000 yr B.P.). During Holocene period (~11,500 yr
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4 B.P. to present), the indicators show higher values and greater variability in contrast to the
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6 glacial time. Our summer monsoon record (QH07) is remarkably similar to the Chinese
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8 speleothem records (Fig.7), although some differences in detail are apparent. The Dongge and
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10 Sanbao cave records show major changes in the late glacial, such as those during the time of the
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12 Bølling–Allerød oscillation, whereas the QH07 record suggests that largely glacial conditions
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14 persisted from the LGM to the end of the Younger Dryas interval (~11,600 yr B.P.). Evidence
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16 for warmer, wetter conditions at the time of the Bølling–Allerød oscillation (~12,900 to 14,600
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18 yr B.P.) is weak in the record from QH07, especially in the TOC and carbonate values,
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20 interpreted as summer monsoon indicators. Moreover, the cave records suggest a relatively
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22 gradual Younger Dryas/Holocene transition, whereas the equivalent change in QH07 is relatively
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24 abrupt. These differences between QH07 and the Chinese cave records may be related to the fact
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26 that Lake Qinghai is further inland and at higher elevation than the Dongge and Sanbao caves
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28 (Fig. 1).
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36 The Asian summer monsoon is at least partly initiated by the differential heating between
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38 the Asian landmass and the Indian and Pacific Oceans (Li and Yanai 1996), although other
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40 processes may be involved (Molnar et al., 2010). Thus, the difference in June insolation between
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42 30°N and 15°N (Huybers 2006), the approximate latitudes of Lake Qinghai and the South China
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44 Sea, respectively, is conceptually an index of this difference, so we compare it to the monsoon
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46 indicators in QH07. The pattern of relatively weak summer monsoon in the glacial time, greatly
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48 enhanced summer monsoon in the early Holocene, and weakening of the monsoon in the late
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50 Holocene broadly matches the insolation-difference record (Fig.7), and implies that orbitally-
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52 induced difference in insolation is the major control on summer monsoon variability on glacial-
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54 interglacial timescales (Weber and Tuenter 2011). Feedbacks for the weakened summer monsoon
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4 during the glacial period may include lower sea level (more distant shoreline) and decreased
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6 continental albedo from increased snow and ice cover and vegetation change over central Asia.
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9 In the early Holocene, maximum organic matter and carbonate contents in Lake Qinghai
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11 sediments signify the strongest summer monsoon, suggesting the occurrence of the warmest,
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13 wettest part of the Holocene, at time of the insolation-difference maximum (Fig. 7). Gradually
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15 reduced monsoonal precipitation (i.e., a weakened summer monsoon) since the mid-Holocene, is
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17 marked by a decrease in the carbonate content and the total organic matter content. The greater
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19 variability in the records of the Holocene implies complex, multiple sediment sources, but the
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21 overall trend follows that of insolation. The Younger Dryas interval is an anomaly, as it is in
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23 many records, compared to the insolation-difference curve. More anomalous still is the period
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25 from ~13,000 - ~15,000 yrs, which appears only slightly wetter than the LGM and the Younger
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27 Dryas, in contrast to many records that show warming (Bølling–Allerød) during this interval.
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33 The QH07 record is quite consistent with previous Qinghai core records, such as QH85-
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35 14B (Lister et al. 1991), QH2000 (Shen et al. 2005), and LQDP05 (An et al. 2012). However,
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37 our interpretation of the QH07 record differs from that of recent studies of shoreline history at
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39 Lake Qinghai that used optically stimulated luminescence (OSL) age estimates (Liu et al. 2010b;
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41 Liu et al. 2011; Liu 2011a; Liu et al. 2012). The latter infers high stands of the lake during the
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43 Bølling–Allerød and the mid-Holocene, and does not identify a high lake-level stand during the
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45 early Holocene (Liu et al. 2010b; Liu 2011a; Liu et al. 2012). The differences between the
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47 interpretation from cores and shoreline deposits might result from the inherent uncertainties in
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49 OSL and ^{14}C dating methods.
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55 There are a number of other paleoclimate records available for the northeastern Tibetan
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57 Plateau (Fig. 1); these show variable levels of consistency with conclusions based on QH07.
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4 Some agree with a warm and/or wet maximum in the early Holocene. For instance, vegetation
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6 history inferred from pollen at Hurleg Lake suggests a moist early Holocene (11.9 -9.5 cal ka
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8 B.P.) and a drying trend during the middle and late Holocene (Zhao et al. 2007). Studies of Lake
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10 Kulai and paleolake Dalianhai also imply an early-mid Holocene optimum (Mischke et al.
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12 2010b; Chen et al. 2012). In contrast, records from the Shiyang River drainage basin suggest a
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14 mid-Holocene high stand (Li et al. 2012). Studies in the northern Ulan Buh Desert document a
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16 large paleolake during 7.8-7.1 cal ka B.P., and that this paleolake began to shrink and modern
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18 desert evolved during late Holocene (Zhao et al. 2012). Some studies infer high lake levels
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20 during the late-Holocene. For example, data from Lake Donggi Cona suggest relatively low lake
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22 level (brackish lake) during the late glacial and early-mid Holocene, and a relatively high lake
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24 level (freshwater lake) after 6.8 cal ka B.P. (Mischke et al. 2010a). Lake Koucha record shows a
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26 lake level drop from 16 to 7.3 cal ka B.P. and a gradual lake rise afterwards (Mischke et al.
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28 2008). Most of these records do not show Bølling–Allerød/Younger Dryas oscillations.
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30 However, Chaka Salt Lake record from the eastern Qaidam Basin reveals Bølling–
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32 Allerød/Younger Dryas, as well as a high lake level between 17.2 and 11.4 cal ka B.P., an abrupt
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34 transition from high to low lake level at the beginning of the Holocene, and a progressively
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36 decrease into late Holocene (Liu et al. 2008). The differences among all these records suggest
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38 either problems in interpreting regional responses to climate change, or non-uniform responses
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40 of individual lakes because of local environmental factors. Larger lakes are less sensitive to local
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42 climate perturbations and thus may be more likely to record regional climate information.
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53 The QH07 record is generally consistent with the isotopic records from most Chinese
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55 caves (Wang et al. 2001; Yuan et al. 2004; Wang et al. 2008), but it also contrasts in a number of
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57 important ways. These include the relatively stable glacial interval, the lack of a significant
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change during the Bølling–Allerød interval, the extremely abrupt Late Pleistocene/Holocene (end Younger Dryas) transition, and the relatively high Holocene variability. These contrasts suggest the existence of a threshold for the arrival of monsoonal precipitation at Qinghai -a threshold that was crossed at the beginning of the Holocene. Due to the high elevation of the Tibetan Plateau, this may be related simply to the barrier that the plateau front poses to penetration of the summer monsoon into the continental interior, or it may relate to complex changes (such as the path of the westerly jet stream) within the climate system.

Conclusions

A multi-proxy approach was applied to the paleoenvironmental interpretations from the sediments in Lake Qinghai. Taken together, the results reveal climatic variations on the northeastern Tibetan Plateau since the LGM. In particular, the carbonate and TOC contents are interpreted as indicators of the strength of the Asian summer monsoon.

- Summer monsoon circulation was relatively weak and stable during the entire LGM and late glacial time, resulting in relatively cool, dry, and stable climatic conditions in the Qinghai region. In contrast, during the Holocene, the summer monsoon was stronger and showed enhanced variability, implying relatively wetter and less stable climatic conditions than during the Late Pleistocene. The strongest summer monsoon, marked by maximum organic matter and carbonate contents, occurred in the earliest Holocene and was followed by a drying trend during the mid- and late Holocene.
- All of the paleoenvironmental indicators abruptly change across the Late-

Pleistocene/Holocene transition, about 11,500 years ago. This change appears to be more abrupt than the isotopic changes at the corresponding time in the Chinese speleothems.

- Near glacial conditions persisted from the LGM to the end of the Younger Dryas interval. Evidence of significant climate change during the equivalent to the Bølling–Allerød oscillation is weak in the QH07 record.
- By comparing our Lake Qinghai record QH07 with the Northern Hemisphere insolation-difference record (between June 30°N and June 15°N), we propose that the changes in insolation contrast between the continent and the ocean is the major control on the Asian monsoon system over the glacial/interglacial time scales. Secondary influences may include albedo effects due to changing ice cover and vegetation over central Asia and sea level (distance from moisture source in Pacific Ocean).
- Differences between the Lake Qinghai record and that from most Chinese speleothems suggest that a climatic threshold existed for arrival of monsoonal rainfall on the northeastern Tibetan Plateau, and that this threshold was crossed at the beginning of the Holocene.

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Table 1 AMS ^{14}C ages of core QH07, shaded items indicate outliers of age estimate.

Fig. 1 The map of China shows the setting of Lake Qinghai on the Tibetan Plateau and the sites of other sediment records and speleothem records discussed in the text (A: Lake Qinghai; B: Dongge Cave, *Yuan et al.* 2004; C: Sanbao Cave, *Wang et al.* 2008; 1: Chaka Salt Lake, *Liu et al.* 2008; 2: Hurleg Lake, *Zhao et al.* 2007; 3: Dalianhai Lake, *Chen et al.* 2012; 4: Shiyang River drainage basin, *Li et al.* 2012; 5: Ulan Buh Desert region, *Zhao et al.* 2012; 6: Lake Donggi Cona, *Mischke et al.* 2010a; 7: Lake Kulai, *Mischke et al.* 2010b; 8: Lake Koucha, *Mischke et al.* 2008). The white square in the southeastern basin of Lake Qinghai indicates the core site location of QH07.

Fig. 2 Age-depth model for QH07 (in calibrated years BP) using a smooth spline model in CLAM (*Blaauw* 2010, smoothing factor, 0.4). Dashed lines show 95% confidence intervals around the age model. Solid line shows the “best” fit of the age estimates. A modern age was assumed for the sediment surface and a constant 700-yr reservoir effect (*Henderson et al.* 2010) was used in the age model.

Fig. 3 Results from lithological and sedimentologic measurements, including water content, Cl/Ti from XRF (water content index), bulk density, carbonate content, and Ca/Ti from XRF (reflecting carbonate content).

Fig. 4 Results from organic geochemical proxies, including total organic carbon, Inc/Coh ratio from XRF (reflecting total organic matter), total organic nitrogen, C/N, and the carbon and nitrogen isotopic composition of bulk organic matter ($\delta^{13}\text{C}_{\text{OM}}$ and $\delta^{15}\text{N}_{\text{OM}}$).

Fig. 5 Cross plots of (a): TOC vs. TN; (b): TOC vs. C/N; (c): TOC vs. $\delta^{13}\text{C}_{\text{OM}}$; (d): C/N vs. $\delta^{13}\text{C}_{\text{OM}}$; and (e): $\delta^{13}\text{C}_{\text{OM}}$ vs. $\delta^{15}\text{N}_{\text{OM}}$. Open squares indicate data for the Late Pleistocene and solid dots indicate data for the Holocene.

Fig. 6 Results for proxies derived from terrigenous sediment, including magnetic susceptibility, Fe content, Ti content, mean grain size, content of $>116\ \mu\text{m}$ fraction of lake sediment, content of $16\text{--}116\ \mu\text{m}$ fraction, Si/Ti from XRF (index of quartz content), and K/Ti from XRF (index of weathering of the sediment).

Fig. 7 Comparison of summer monsoon proxies for QH07 (dotted line: TOM flux; solid line: carbonate flux) with Dongge (solid line) and Sanbao (dotted line) cave speleothem $\delta^{18}\text{O}$ records and June insolation difference between 30°N and 15°N . Gray bars indicate the Younger Dryas (YD) and Bølling–Allerød (B/A) periods.

Lab code	Core	Depth (cm)	Composite depth (cm)	Materials	$\delta^{13}\text{C}$ (‰)	^{14}C age (yr)	Error (yr)	Calibrated age (2 σ) (yr BP)
AA88809	QH07-1B-1MM	2-3	17-18	bulk	-	2460	210	1282-2156
83368	QH07-1B-1MM	19-20	34-35	bulk	-26.59	1390	30	638-683
AA88810	QH07-1B-1MM	35-36	50-51	bulk	-	2450	210	1276-2154
83369	QH07-1B-1MM	50-51	65-66	bulk	-26.66	1720	30	905-980
AA88811	QH07-1B-1MM	67-68	82-83	bulk	-	2560	210	1360-2210
83370	QH07-1A-1U-1	6-7	206.5-207.5	bulk	-24.61	4630	35	4281-4440
AA88805	QH07-1A-1U-1	11-12	211.4-212.4	bulk	-26.2	4847	57	4527-4832
SUERC-18422	QH07-1A-1U-1	50.5-51.5	248.8-249.8	bulk	-27.7	4530	38	4146-4359
SUERC-18423	QH07-1A-1U-2	3-4	306-307	organic particles	-18.8	6939	38	7150-7257
AA88806	QH07-1A-1U-2	39-40	332.9-333.9	bulk	-21.7	8806	50	8974-9148
SUERC-18424	QH07-1A-1U-2	81-83	377-379	<i>Ruppia</i> seeds	-13.3	9475	43	9596-9926
AA88807	QH07-1A-2U-2	32.5-33.5	401.4-402.4	bulk	-22.5	9908	44	10,250-10,444
SUERC-18425	QH07-1A-2U-2	60-61	427.7-428.7	<i>Ruppia</i> seeds	-12.8	10653	46	11,245-11,505
SUERC-18417	QH07-1A-2U-3	13-14	447.5-448.5	bulk	-22.3	14270	61	16,483-16,928
83371	QH07-1A-2U-3	14-15	472-473	bulk	-24.69	4760	35	4427-4628
AA88808	QH07-1A-2U-3	58-59	491.5-492.5	bulk	-23.0	14670	230	16,637-17,684
SUERC-18418	QH07-1A-2U-3	94-95	528.5-529.5	bulk	-21.9	17062	83	19,372-19,608

Table 1

Fig. 1

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Fig. 3

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Fig. 6

Fig. 7