Late Cenozoic Climate Changes in China's Western Interior: A Review of Research on Lake Qinghai and Comparison with Other Records

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Late Cenozoic climate changes in China's western interior: a review of research on Lake Qinghai and comparison with other records

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Abstract

We review Late Cenozoic climate and environment changes in the western interior of China with an emphasis on lacustrine records from Lake Qinghai. Widespread deposition of red clay in the marginal basins of the Tibetan Plateau indicates that the Asian monsoon system was initially established by ~8 Ma, when the plateau reached a threshold altitude. Subsequent strengthening of the winter monsoon, along with the establishment of the Northern Hemisphere ice sheets, reflects a long-term trend of global cooling. The few cores from the Tibetan Plateau that reach back a million years suggest that they record the mid-Pleistocene transition from glacial cycles dominated by 41 ka cycles to those dominated by 100 ka cycles.

During Terminations I and II, strengthening of the summer monsoon in China's interior was delayed compared with sea level and insolation records, and it did not reach the western Tibetan Plateau and the Tarim Basin. Lacustrine carbonate $\delta^{18}O$ records reveal no climatic anomaly during MIS3, so that high terraces interpreted as evidence for extremely high lake levels during MIS3 remain an enigma. Following the Last Glacial Maximum (LSM), several lines of evidence from Lake Qinghai and elsewhere point to an initial warming of regional climate about 14 500 cal yr BP, which was followed by a brief cold reversal, possibly corresponding to the Younger Dryas event in the North Atlantic region. Maximum warming occurred about 10 000 cal yr BP, accompanied by increased monsoon precipitation in the eastern Tibetan Plateau. Superimposed on this general pattern are small-amplitude, centennial-scale oscillations during the Holocene. Warmer than present climate conditions terminated about 4000 cal yr BP. Progressive lowering of the water level in Lake Qinghai during the last half century is mainly a result of negative precipitation–evaporation balance within the context of global warming.

1. Introduction

The dryland ecosystem on the northern Tibetan Plateau (Fig. 1A) is vulnerable to climate changes and human impacts. Degradation of grasslands, contraction of lakes, and desertification have become major environmental problems in recent years, affecting socio-economic development in the region. Restoring these lands or reversing these processes requires knowledge about the baseline conditions of the drylands and the dynamics of landscape change, particularly during past interglacials when climate boundary conditions were similar to the present. Studying Late Cenozoic climate changes on the Tibetan Plateau also is crucial for understanding the complex interaction among the atmosphere, lithosphere, hydrosphere, cryosphere, and biosphere of the Earth system on a longer time scale. Long, undisturbed lacustrine sediment sequences are important for addressing these questions. Previous studies reveal that sediment cores from Lake Qinghai contained abundant information about regional environmental history at various time scales. Because most of these results appear

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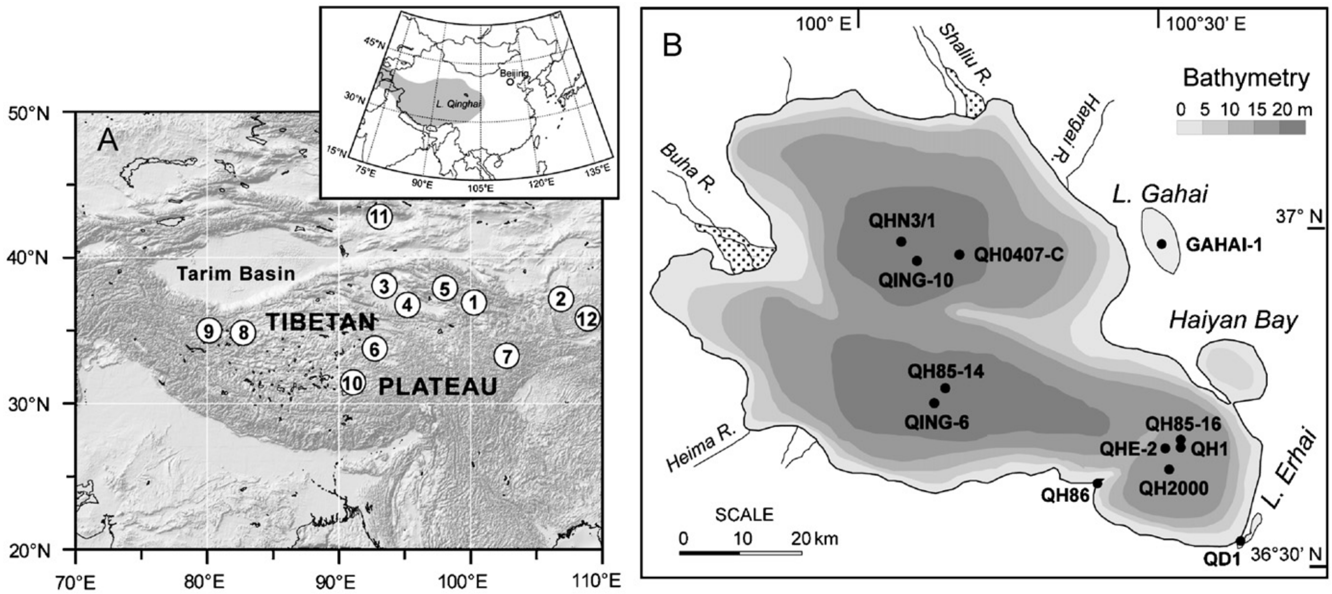

Fig. 1. (A) Relief of the western interior of China; numbered circles indicate sites discussed in the text. 1. Lake Qinghai; 2. Xifeng loess site; 3. Qarhan Playa; 4. Kunteyi Playa; 5. Dunde Ice Core; 6. Lake Co Ngoin; 7. Zoige Basin; 8. Guliya Ice Core; 9. Lake Tianshuihai; 10. Lake Man Co; 11. Lake Balikun; 12. Luochuan loess site. (B) Bathymetry of Lake Qinghai. Solid dots show the location of cores discussed in the text.

in Chinese-language journals, they are not accessible to the international community. In addition, future studies of drill cores from the Lake Qinghai Drilling Project, jointly funded by the Chinese Academy of Sciences (CAS) and the International Continental Drilling Program (ICDP), require background information about previous studies of climate and environment changes in this area. Hence, this review of paleoenvironmental studies was undertaken before the Lake Qinghai Drilling Project retrieved its cores from Lake Qinghai in late 2005.

The importance of Lake Qinghai sediment records for understanding past global changes has been increasingly recognized since the 1870s (cf. Chen et al., 1990), due to its unique geographical location (Fig. 1A). Preliminary geological and geomorphologic mapping in the lake area were performed during the first half of the 20th century (Shi et al., 1958; Chen et al., 1964). However, limnological studies using modern techniques did not start until the 1960s, when fundamental data on lake biology, water chemistry, and hydrology were first obtained during a multi-disciplinary expedition (Lanzhou Institute of Geology and Chinese Academy of Sciences (LZIG-CAS), 1979). Yang and Jiang (1965) examined Quaternary vegetation history by analyzing pollen assemblages of a 210-m-long drill core (QH5) on the Erlangjian terrace (Fig. 1B). Interest continued to increase in the 1980s. Approximately 80 cores were drilled around the lake by the Qinghai Geological and Mineral Resource Administration. Two of these cores (DH-54 and DH-64) reached depths of 500 and 300 m below the sediment surface, respectively, and thus provide long potential records of Quaternary climate changes. Coordinated by the Institute for Salt Lake Studies, CAS, a Sino-Swiss-Australian cooperative project was implemented in 1984, aiming to elucidate the recent climate history of the area. In 1987, a 155-m-long drill core (QH86) from the Erlangjian terrace (Fig. 1B), along with three short piston cores, numbed QH85-14, QH85-15, and QH85-16, from the southern sub-basin of the lake, were recovered. ^{210}Pb and ^{137}Cs dating (Huang and Sun, 1989) of these piston cores, along with a variety of other analyses, including pollen (Du et al., 1989; Kong et al., 1990; Shan et al., 1993), carbon (Kelts et al., 1989; Huang and Meng, 1991), elemental and isotope geochemistry (Zhang et al., 1989a, b, 1994; Lister et al., 1991; Sun et al., 1991) have been carried out. These proxies reveal Lateglacial to Holocene changes in lake hydrology and catchment environment associated with the advance and retreat of the Asian summer monsoon front (Wang and Shi, 1992; Wei and Gasse, 1999; Yu and Kelts, 2002).

The foci of ongoing paleolimnological studies are on high-resolution changes in lake hydrology and regional climate, based on multiple stratigraphic analyses of short cores, i.e. QING-6, QHE-2, QING-10, QHN3/1, QH0407-C, and GAHAI-1 (Fig. 1B). A group from the University of Lanzhou and University College London (Guo et al., 2002a; Henderson et al., 2003; Shi et al., 2003; Zhang et al., 2003; Henderson, 2004), and one from the Institute of Geography and Limnology, CAS, (Shen et al., 2001; Liu et al., 2002; Zhang et al., 2002a, b, 2004; Liu et al., 2003a, b, c; Shen et al., 2005) have played leading roles in this wave of investigation. However, systematic studies of modern conditions and limnological processes are rare (Chinese Academy of Science Lanzhou (CAS-LZ), Research Center for Resource and Environment of Western China, Chinese Academy of Sciences (RCREWC-CAS), 1994). Numerical models have been used to try to understand the hydrological and chemical evolution of the lake under various scenarios of climate forcing (Qin and Huang, 1998a, b; Yan et al., 2002), although few data exist to constrain such models.

Here we compile and synthesize previously published results of paleolimnological studies of Lake Qinghai. We then compare paleoclimate records from the Tibetan Plateau and nearby areas for four time intervals (Late Cenozoic, the Last Interglacial, Marine Isotope Stage (MIS) 3, and the Postglacial), three of which are currently represented by data from Lake Qinghai, in an effort to synthesize paleoclimate information from the Tibetan Plateau.

2. Environmental setting

2.1. Neogene climate history

The inception and subsequent intensification of the Asian monsoon have long been ascribed to the upward/outward growth of the Tibetan Plateau (Ruddiman et al., 1989; Molnar et al., 1993; An et al., 2001; Li et al., 2001; Liu and Yin, 2002), within the context of global cooling during the Neogene (Maslin et al., 1998; Hay et al., 2002; Gupta et al., 2004). Although magneto-stratigraphy of loess-soil sequences and other sediment records from surrounding areas provide an *a priori* constraint on the uplift history of the Tibetan Plateau (Quade et al., 1989; Rea et al., 1998; An et al., 2001; Dettman et al., 2003), the timing and mechanisms have been the subject of debate (Harrison et al., 1992; Coleman and Hodges, 1995; Chung et al., 1998). Independent geological evidence reveals that a modest uplift in the eastern Tibetan Plateau began at ~40 Ma (Harrison et al., 1992; Chung et al., 1998), as a result of the Indio-Asian collision. However, persistent dry conditions following the Paleocene–Eocene thermal maximum in Qaidam Basin (Wang et al., 1999) suggest that the plateau did not reach an elevation high enough to modify the prevailing atmospheric circulation pattern.

Extensive uplift about 22 Ma is indicated by widespread failures of the Paleogene planation surfaces and increased sedimentation rates in the neighboring basins (Zhang et al., 2001). This large-scale surface uplift, along with substantial retreat of the Paratethys Sea (Ramstein et al., 1997), is believed to have initiated desertification of the Asian interior (Guo et al., 2002b). A weak Asian summer monsoon circulation may have been established when the plateau reached a threshold elevation of ~1500 m (Ruddiman and Kutzbach, 1989). Palynological evidence from Qaidam Basin (Wang et al., 1999) and Linxia Basin (Ma et al., 1998) indicate that dry conditions were not fully established in the western interior of China until ~15 Ma. This signals the initial cooling of the Late Cenozoic (Zhou and Zhu, 2001). Superimposed upon this general trend were two drying episodes that occurred at about 14 and 8 Ma (Guo et al., 2002b), the first of which can be correlated to the initial growth of the Antarctic Ice Sheet and probably associated with internal feedbacks in the climate system. A wealth of evidence suggest that the latter can be attributed to the Tibetan uplift (An, 2000), which raised the plateau surface to its present altitude and thus gave rise to the modern Asian summer monsoon. Widespread red clay deposition beginning ~8 Ma in the marginal basins of the Tibetan Plateau (An, 2000) and increased dust transport to the North Pacific (Rea et al., 1998) indicate that the Asian winter monsoon was also established at that time. Superimposed upon a global cooling trend was the gradual strengthening of the Asian winter monsoon from 3.6 Ma onward, thereby leading to thick loess deposition along the margin of deserts and gobis (An, 2000) and dust deposition in the North Pacific (Rea et al., 1998). In this context, Lake Qinghai lies at the transition from the arid to the semi-arid zones, where the climate is controlled by the Asian summer monsoon, the Asian winter monsoon, and the westerlies. It is thus well suited to monitor changes in Late Cenozoic climate.

2.2. Geomorphologic and tectonic backgrounds

Lake Qinghai (36°15′–38°20′N, 97°50′–101°20′E) is situated on the northeastern Tibetan Plateau, and is the largest inland water body of China by surface area. The lake is developed within a basin surrounded by three mountain ranges (Bian et al., 2000): Datong Mts to the north, Riyue Mts to the east, and Qinghai Nanshan Mts to the south. These mountains, with general elevations above 4000 m, account for ca 70% of the drainage area. The lake basin is open to the west, from where it receives its major runoff from the Buha River, which, along with other major rivers, have created fluvial plains and deltas in the western and northern shores. Dunes and beach ridges are common along the eastern shore, reflecting a prevailing westerly wind pattern. Fault escarpments and terraces are extensively developed along the southern shore, where faulting and block tilting are still active. Glacial and periglacial landforms can be found on Qinghai Nanshan Mts (Porter et al., 2001), whereas small modern glaciers only occur on mountains in the upper Buha River drainage basin.

The Late Cenozoic tectonic evolution of the region represents the history of growth of northeastern margin of the Tibetan Plateau (Molnar et al., 1993). A northwest-dipping thrust fault is present along southern range-front of the Qinghai Nanshan Mts The mountain range is thus a tectonic ramp that thrusts southward over the Gonghe basin. The Riyue Shan fault zone consists of a high-angle, right-lateral strike–slip fault in the middle of the range and a west-dipping, low-angle thrust fault along the eastern range-front of the mountain. Therefore, Lake Qinghai is basically a piggyback basin behind the thrust ramps to the south and west.

The onset of Cenozoic tectonics in the Lake Qinghai region is believed to be either Late Miocene or early Pliocene along the Qinghai Nanshan fault, which caused uplift of the Qinghai Nanshan Mts (Metivier et al., 1998), thereby separating Lake Qinghai from the Gonghe basin. The uplift of the Riyue Shan Mts intersects the NWWtrending Qinghai Nanshan Mts near Lake Qinghai during the late Pleistocene (Yuan et al., 1990). This uplift eventually blocked the river system that used to drain Lake Qinghai to the east, and isolated Lake Qinghai to form the closed intra-mountain piggyback basin that we see today.

2.3. Climate and vegetation

Lake Qinghai lies in the transition from semi-arid to arid zones, where mean annual precipitation is ~360 mm (CAS-LZ and RCREWC-CAS, 1994), decreasing along a gradient from surrounding mountains (~400 mm) to the lake area (e.g. ~220 mm on Mt Haixin). The annual precipitation also shows temporal variability and most (60%) falls in summer months (June–August). With much sunshine (3640 h/a) and high insolation ($\sim 6.5\times 10^{15}$ J/m^2 a), annual mean evaporation is ~800 mm. It decreases from the lake area (~1000 mm) to surrounding mountains (~310 mm). More than 60% of the evaporation occurs in summer. Annual mean temperature is ca −0.7 °C and exhibits remarkably high seasonality, varying from ca −11 °C in winter to ca 12 °C in summer. Winds blow onshore at daytime and offshore at night with an average speed of 4–6 m/s.

The lake has a drainage area of 29 660 km^2, which is mainly vegetated by montane shrubs, alpine steppes, and high-alpine meadows. The vegetation shows a distinct vertical zonation (Chen and Peng, 1993). Forests are rare and scattered. Major tree species are *Sabina przewalskii*, *Sabina vulgaris*, *Salix oritrepha*, *Picea crassifolia*, *Populus cathayna*, and *Populus simonii*. Shrubs, dominated by *Caragana jubta*, *Potentilia fruticosa*, *Potentilia glabra*, *Myricaria sgamosa*, *Hippophae neurocarpa*, and *Cotoneaster acutifolius*, appear locally and mainly colonize on the base of the southern slopes. Alpine steppes, composed of *Achnatherum splandens*, *Kobresia pygmaea*, *Kobresia humulis*, *Kobresia bellardii*, *Kobresia capillifolia*, *Leontopodium nanum*, *Androsacatapete*, *Thylacospermum caespitosum*, *Stipa purea*, *S. breviflora*, *Artemisia ordosis*, *Agropyron desertorum*, *Carex stenophylla*, *Oxytropis falcate*, and *Poa sinoglauce*, frequently occur between 3200 and 4500 m. Desert steppes, including mainly *Artemisia sphaerocephala*, *Cxytropis aciphalla*, *Agropyrom cristatum*, *Ephedra intermedia*, and *Kobresia robusta*, occur rarely. Wetland communities have low species diversity, and the major species are *Triglochin palustre*, *Triglochin maritimum*, *Blysmus sinocompressus*, *Carex* spp., and *Potentilia acaulis*. Aquatic plants are rare: only two species, *Potamogeton pectinatus* and *Ruppia maritime*, as well as a few emergent sedge species are found in quiet waters near the shore (Chen, 1987a, b).

Human impact on modern vegetation of the area, at least in relation to preservation of pollen spectra, are thought to be minimal. Archeological sites in the area are rare, and no late Neolithic sites occur around the margins of Lake Qinghai (Madsen et al., 2006). Forested areas are uncommon and scattered, so human deforestation is not prevalent. Even today, the Lake Qinghai area is remote and sparsely populated.

2.4. Lake hydrology and water chemistry

The lake basin is located at 3194 m above sea level (asl), with a surface water area of 4400 km^2 and volume of 7.16×10^{10} m^3. Five large rivers seasonally discharge to the lake basin with annul runoff of $\sim 1.34\times 10^{9}$ m^3 (Wang, 2003). Annual sand discharge to the lake is 4.98×10^{5} T. The Buha River, with a watershed of 14 337 km^2, is the largest river both by runoff (50% of the total) and sand discharge (70% of the total). Meltwater from surrounding mountain glaciers accounts for only 0.3% of the total runoff. The hydrological residence time of the lake was estimated as 33 years (Lister et al., 1991). The lakebed is generally flat, with an average water depth of 21 m. The lake is divided into two nearly equally sub-basins by a NNW-trending horst, from which an island (Mt Haixin) emerges. Maximum depth (27 m) occurs in the southern sub-basin (12 km south of Mt Haixin). Several minor fault scarps also can be found in the lake basin.

The lake water is brackish to saline with an average salinity of 14.1 g/L and a pH of 9.2. Electrical conductivity is 20.63 ms/cm. Lake water δD is 10.0‰ (V-SMOW), and δ^{18}O is 1.97‰ (V-SMOW), values that are much higher than those of local meteoric precipitation and thus indicate that the lake is hydrologically closed, and evaporative (Zhang et al., 1994). In summer, weak thermal stratification develops, with an epilimnion of 12–15 °C, and a hypolimnion of 6 °C (Williams, 1991). The surface water is usually saturated in summer with respect to its carbonate minerals, generating a continuous rain of aragonite. The lake surface is frozen during winter months (December–-March) with a maximum ice thickness of 0.8 m.

3. Paleolimnological records of Lake Qinghai at different time scales

3.1. Potential record from long drill cores

Oxygen isotope records of marine cores reveal that the Earth's climate system has experienced significant changes during the Late Cenozoic, characterized by a gradual shift from an ice-free mode to glacial conditions after ~3.6 Ma (Zachos et al., 2001). Northern Hemisphere glaciation apparently began at about 2.7 Ma (cf. Haug et al., 2005). A leading hypothesis for this transition is the tectonically induced closure of Pacific–Atlantic seaways (Haug and Tiedemann, 1998; Cane and Molnar, 2001), which in turn caused the reorganization of ocean circulation. In contrast to oceanic changes, extensive uplift of mid-latitude mountains, particularly the Tibetan Plateau, also has been suggested as playing a vital role in triggering global climate changes through physical and weathering processes (Ruddiman and Kutzbach, 1989; Raymo and Ruddiman, 1992;

France-Lanord and Derry, 1997). Therefore, dating this change in terrestrial records, especially those on or close to the Tibetan Plateau, is important for providing evidence for the causal relationship between tectonic uplift and climate changes. Although evidence has been accumulating from areas surrounding the plateau (Quade et al., 1989; Prell and Kutzbach, 1992; Rea et al., 1998; An et al., 2001), long, high-quality records from the Tibetan Plateau itself are lacking.

Lacustrine sediments from large lakes hold one of the keys to understanding Late Cenozoic climate changes in Asian inland (e.g. Colman et al., 1995; Colman, 1996; Williams et al., 1997). Previous drilling from the Erlangjian terrace (a spit prograding into Lake Qinghai) reveals that the lake sediments may have a higher temporal resolution (Yuan et al., 1990) than the red clay-loess-soil sequences. Recent geophysical surveys (An et al., 2006) reveal that the thickness of lacustrine sediments in the lake exceeds 700 m. The age of the basal lacustrine sediments is estimated to be late Miocene (Yuan et al., 1990). Detailed chronological, sedimentological, geochemical, and biostratigraphic analyses of long drill cores from current and ongoing Lake Qinghai drilling projects have the potential to shed new light on the timing of Late Cenozoic tecto-climate events and the dynamics of the Asian monsoon system during the last eight million years.

3.2. The QH86 record

3.2.1. Stratigraphy and chronology

Core QH86, drilled from the Erlangjian terrace on the southern shore in 1987, is the longest sediment sequence of Lake Qinghai to date (Yuan et al., 1990). The upper 80 m of the core consists of continuous lacustrine silty clay interrupted rarely by coarse sand and silt. This stratigraphy reveals a generally stable sedimentary environment within the lake basin, interspersed with episodic fluctuations of lake level. The record below 80 m reveals that the lake basin was shallow, but not subaerially exposed during previous glaciations, as indicated by a thick layer of fine sandy sediments interbedded with silt and gravel between 127 and 140 m. The Brunhes/Matuyama (B/M) paleomagnetic polarity boundary (~0.8 Ma) occurs at 120 m, yielding an average sediment accumulation rate of about 0.16 mm/a. Assuming a constant accumulation rate, full lacustrine conditions (at ~80 m) were established in the lake basin after ~0.5 Ma (Yuan et al., 1990). However, Ruppia seeds were found at ~96.5 m, and freshwater algae account for 69% of the total pollen at ~120 m (Shan et al., 1993), implying that episodic lacustrine conditions occurred much earlier. Twelve uranium-series ages on authigenic carbonates were obtained by Shan et al. (1993). Although no details on the ages were given, they lead to an age–depth relationship for the upper 80 m of the core of: Age $= 4234 \times$ Depth $+ 11\,414$.

3.2.2. Pollen assemblages

Pollen analyses of 90 samples from the upper 80 m of cores from QH86 were performed by Shan et al. (1993). Pollen grains are poorly preserved below this level, and only a few grains were found. The relative abundances of major taxa in this core are presented in Fig. 2. Pollen assemblages are dominated by dwarf shrubs and herbs, which account for ca 80% of the total palynomorphs and mainly include Ephedra, Nitaria, Artemisia and Chenopodiaceae. Abundances of arboreal pollen, including Abies, Picea, Pinus, Cedrus, Tsuga, Betula, Corylus, Alnus, Ulmus, Juglans, Tilia, and Rhus, are very low, generally less than 20% of the total. Wetland and aquatic herbs, including Compositae, Umbelliferae, Polygonaceae, Cyperaceae, Typha, Potamogeton, and Ruppia, are very rare. Seven local pollen zones (Fig. 2) were defined (Shan et al.,

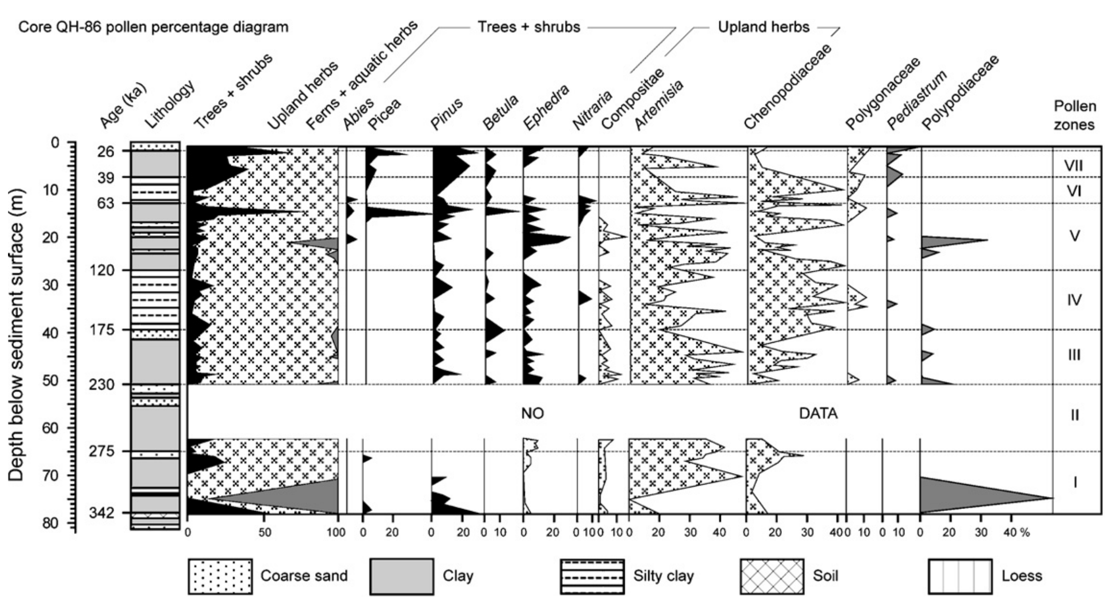

Fig. 2. Core QH86 pollen percentage diagram (after Shan et al., 1993). Only major taxa are shown.

1993), and the original interpretations are summarized below. The age of major pollen zone boundaries was calculated from the age–depth model described above.

Zone I (77.93–62.21 m; 342–275 ka): this zone is dominated by dryland herbs such as *Artemisia* and Chenopodiaceae. Tree pollen only account for 14%, with a maximum occurring at around 78 m. High values of Polypodiaceae spores occur around 75 m, indicating a period of wetland expansion possibly associated with lake-level rise.

Zone II (62.21–51.81 m; 275–230 ka): pollen concentrations are very low, and only a few *Pinus* and *Ephedra* pollen grains were found. These low pollen concentrations most likely indicate a period of extremely low lake levels and sedimentary conditions not conducive to pollen preservation.

Zone III (51.81–38.60; 230–175 ka): dryland herbs, such as *Artemisia*, dominated the alpine meadow landscape. Abundances of tree pollen decreased to 8%. Polypodiaceae and *Pediastrum* occurred episodically, indicating a period of variable lake level.

Zone IV (38.60–25.71 m; 175–120 ka): tree pollen continued to decrease. Fern spores and aquatic pollen are very rare. *Artemisia* pollen decreased substantially, concurrent with the maximum abundance of Chenopodiaceae pollen. This pollen assemblage indicates a transition from alpine meadow to an alpine steppe landscape associated with increasing dry conditions during the penultimate glaciation.

Zone V (25.71–12.25 m; 120–63 ka): pollen assemblages are still dominated by *Artemisia* and Chenopodiacea, but tree pollen, including *Picea*, *Betula*, and *Pinus*, increased gradually and reached 15%. This pollen assemblage indicates a landscape of alpine steppe with scattered mixed forests of coniferous and broad-leaved trees. Polypodiaceae and *Pediastrum* occurred again at about 25 m with an abundance reaching 39%, reflecting a substantial expansion of wetland, most likely associate with lake-level rise at about 95 ka.

Zone VI (12.25–6.52 m; 63–39 ka): this zone is marked by a gradual decrease in tree pollen. Abundances of *Artemisia* pollen increased first, followed by Chenopodiaceae. Such a succession from a landscape of alpine steppe with scattered mixed forests to an alpine steppe indicates increasingly dry conditions during MIS 4.

Zone VII (6.52–3.50 m; 39–26 ka): this zone is marked by considerable increases in tree pollen, along with a modest rise in the abundance of aquatic herbs and ferns. Abundances of *Artemisia* and Chenopodiaceae pollen decreased substantially. This pollen assemblage points to landscapes of mixed forests with coniferous and broad-leaved trees, suggesting warmer and wetter than present conditions during MIS 3 (but see Section 4.2.2).

3.3. The QH85 and QH2000 records

3.3.1. Stratigraphy, chronology, and sedimentation rates

Preliminary geophysical investigations of the lake floor topography and sediments were conducted during the Sino–Swiss–Australian project, using high-resolution (3.5 kHz) seismic-reflection techniques and coring. Seismic profiles reveal a pronouncedly layered feature of sediments and two units can be identified (Kelts et al., 1989). The *H*-series is 6–7-m-thick and thins gradually towards the shore. This unit appears to represent continuous and undisturbed lacustrine sediments (Fig. 3). A strong reflection (from the *Q*-reflector) could be correlated with the top of a yellowish silty sandy layer at Sites QH85-14B and QH85-16A. This

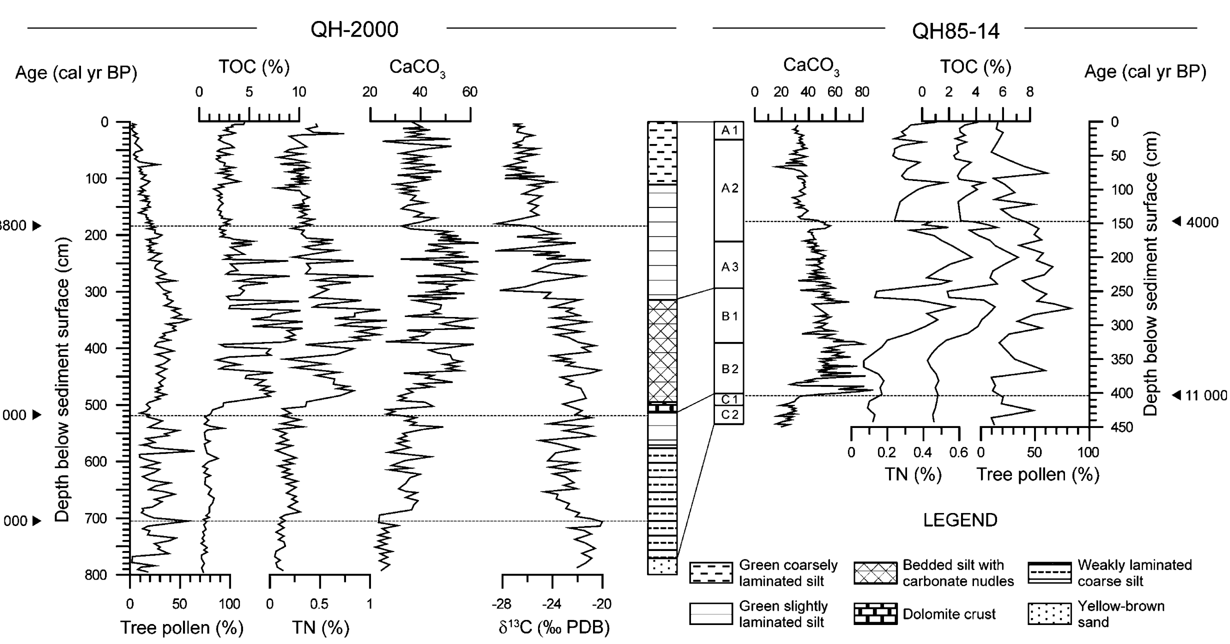

Fig. 3. Correlation of total organic carbon (TOC), total nitrogen (TN), carbonate, arboreal pollen, and organic carbon isotope records between Cores QH2000 (Shen et al., 2005) and QH85-14 (Du et al., 1989; Kelts et al., 1989; Huang and Meng, 1991). Dashed lines indicate boundaries of major changes.

layer is present in the entire basin and probably indicates a period of extremely low lake level. This inference is supported by the stratigraphy of Core QH2000 (Shen et al., 2005), where massive yellow-brownish sand occurs at ~7.7 m (Fig. 3). Extrapolations of radiocarbon chronologies for these cores suggest that the upper part of this layer was deposited at ~19 000 cal yr BP (Table 1; Fig. 3), approximately the Last Glacial Maximum (LGM). Satellite images show an underwater beach ridge in the eastern part of the lake and a 6-km-long underwater channel in the western part (Yuan et al., 1990), which were interpreted to have developed during the LGM. These findings imply that the lake level was much lowered during the LGM, presumably controlled by a dry, cold climate (Lister et al., 1991; Shen et al., 2005).

Dating sediment cores from Lake Qinghai is problematic because of potential hard water effects, which have been a great challenge to paleolimnological studies. Terrigenous plant macrofossils in the almost treeless watershed are rare, and thus bulk organic matter (OM) usually is used for radiocarbon dating. However, OM may have been reworked and delivered to the lake from various pre-aged sources by river and groundwater discharges. Linearly fitting the calibrated radiocarbon ages of OM to depth in cores (Table 1) gives a sediment–surface intercept of 1039 years for Core QH2000 (Shen et al., 2005), 439 years for QH85-16A (Zhang et al., 1994), and 1100 years for Core QH85-14B (Kelts et al., 1989). These core–top apparent ages could be ascribed to a hard-water effect or reservoir age inherited from the catchment. A radiocarbon age of 661 ± 32 yr BP for dissolved organic carbon (DOC) was reported by Henderson (2004). The finite age of modern lake water DOC again indicate the inputs of older OM by river and groundwater discharges. In contrast, the dissolved inorganic carbon (DIC) in lake water is dated to be post-1950 (Henderson, 2004), indicating that ^{14}C-rich DIC added from the post-nuclear testing atmosphere is offset by the pre-aged DIC derived from catchment weathering. Another complication is that radiocarbon ages of carbonates from some short cores are about 1000 years older than those of OM samples from the same horizon (Henderson, 2004). This may be due to the introduction of detrital carbonates from the catchment. If so, using stable isotopes of lacustrine carbonate as a climate proxy would also be problematic.

Sedimentation rates in the lake subbasins during the last 200 years have been well constrained by measuring activities of ^{210}Pb and ^{137}Cs (Huang and Sun, 1989; Zhang, 2003; Henderson, 2004; Xu et al., 2006a). Sedimentary ^{137}Cs signals are exceptionally strong in these cores, and the post-bomb maximum of 1963 is clear. A rapid increase in sedimentation rate, along with high values of magnetic susceptibility after 1952 in Core QH2000 suggests strong catchment erosion associated with overgrazing and extensive land reclamation (Zhang et al., 2002a, b). Sedimentation rates show high spatial variability (Table 2), which is consistent with preliminary sediment trap studies (CAS-LZ and RCREWC-CAS, 1994). High values occur in the

Table 1
Radiocarbon ages of Cores QH85-14B (Kelts et al., 1989), QH85-16A (Zhang et al., 1989b), and QH2000 (Shen et al., 2005), Lake Qinghai, China

Depth (cm)	Lab. ID	Materials dated	$\delta^{13}C$ (‰ PDB)	Radiocarbon age (yr BP)	2σ calibrated age (yr BP)
Core QH85-14B					
47.5–49	–	Algal threads	–	1780 ± 23	1820–1610
339.5–341	–	*Ruppia* seeds	–	8400 ± 130	9600–9000
382–383.5	–	*Ruppia* seeds	–	9730 ± 130	11 650–10 650
387–395	–	*Ruppia* seeds	–	9870 ± 170	12 050–10 750
438–440	–	*Ruppia* seeds	–	10 900 ± 250	13 350–12 150
Core QH85-16A					
14–21	ANU-6020	Organic matter	–	660 ± 140	950–400
61–70	ANU-6021	Organic matter	–	2280 ± 100	2700–2000
151.5–161.5	ANU-6022	Organic matter	–	3630 ± 100	4250–3600
236–246	ANU-6023	Organic matter	–	6370 ± 160	7600–6900
321–326	ANU-6024	Organic matter	–	7540 ± 240	9050–7850
396–406	ANU-6025	Organic matter	–	9710 ± 200	11 850–10 500
491–503	ANU-6026	Organic matter	–	11 590 ± 260	14 000–12 950
Core QH2000					
50–51	Tka-12179	Organic matter	−25.8	2700 ± 100	1820–1350
150–151	Tka-12180	Organic matter	−25.8	4240 ± 170	3850–2950
150–151	Tka-12181	Organic matter	−25.7	4010 ± 100	3380–2870
150–151	Tka-12182	Organic matter	−25.3	4000 ± 90	3360–2880
230–231	Tka-12183	Organic matter	−26.7	5060 ± 90	4850–4200
355–357	Tka-12184	Organic matter	−22.3	6760 ± 180	7000–6100
475–477	Tka-12193	Organic matter	−22.3	9660 ± 140	10 200–9300
545–547	Tka-12185	Organic matter	−23.2	14 680 ± 180	16 850–15 550
675–677	Tka-12186	Organic matter	−22.8	14 820 ± 180	17 050–15 750
745–747	Tka-12236	Organic matter	−25.0	15 610 ± 90	18 000–17 050

Table 2
Sedimentation rates (mm/a) in Lake Qinghai during the last 200 years

Site	^{210}Pb	^{137}Cs	^{14}C	Data sources
QH85-14A	0.51	–	0.32	Huang and Sun (1989)
QH85-14C	0.85	–	–	Huang and Sun (1989)
QH85-16A	0.42	–	0.44	Huang and Sun (1989)
QH2000	1.30	–	0.32	Shen et al. (2001)
QING 6	1.00	0.40	0.50	Henderson (2004)
QING 10	1.30	1.50	0.15	Henderson (2004)
QHE2/01	1.40	1.50	0.42	Henderson (2004)
QH0407-C-2	1.00	–	–	Xu et al. (2006a)

eastern sub-basin, consistent with greater shoreline erosion due to wave activity driven by the prevailing westerly winds.

3.3.2. Climate and environment changes

Detailed records from three sediment cores in the southern and southeastern part of the lake (Fig. 1B) cover the Lateglacial and Holocene. Pollen analyses and total organic carbon (TOC), total nitrogen (TN), grain size, and carbonate measurements were conducted on Cores QH85-14 (Du et al., 1989; Kelts et al., 1989; Kong et al., 1990; Huang and Meng, 1991) and QH2000 (Liu et al., 2002, 2003a–c; Shen et al., 2005) to provide complementary information on productivity, sedimentation, and climate change. Core QH1, 350-cm long, was taken just 5 m south to Core QH85-16A (Fig. 1B) by the Lanzhou Institute of Geology, CAS. This core can be regarded as a replica of the upper part of Core QH85-16A. Laboratory analyses, including grain size, organic carbon, trace elements, and carbon isotopes, were conducted on this core, covering the last 8500 years (Guo et al., 2002a; Shi et al., 2003).

Proxy records correlate well between Cores QH85-14 and QH2000 (Fig. 3). TOC, TN contents and tree pollen percentages exhibit parallel changes among the cores. Five distinct stages of regional climate changes can be defined (Fig. 3), based on the various proxies. Minor increases in TOC, TN, and carbonate contents after ca 17 000 cal yr BP indicate the termination of LGM. But the lake level was still low, and most of the OM was derived from the treeless catchment as indicated by less negative $\delta^{13}C$ compared to earlier times. A decrease in tree pollen percentages between 12 500 and 11 500 cal yr BP may correlate with the Younger Dryas stadial (Yu and Kelts, 2002), although chronologies are uncertain. Steady warming began at 11 000 cal yr BP, along with higher primary productivity as marked by significant increases in TOC, TN, and carbonate content. This trend was frequently punctuated by a number of centennial-scale cooling, possibly including the “8.2-ka event”, implying unstable climate conditions during the mid-Holocene climate optimum (Chen et al., 2001; Wunnemann et al., 2003). A significant shift in climate conditions occurred at ~4000 cal yr BP. The $\delta^{13}C$ values continued to become more negative, possibly related to an increasing input of terrigenous OM as lake level continuously lowered. However, pollen records indicate that the C_3/C_4 ratio of the vegetation did not change significantly during the Holocene (Shen et al., 2005).

3.3.3. Pollen record and vegetation history

High-resolution pollen analyses were conducted on Cores QH85-14C (Du et al., 1989) and QH2000 (Shen et al., 2005). Preservation of pollen grains in sediments below 707 cm in Core QH2000 is very poor (<100 grains per slide), which makes it difficult to infer the composition of local vegetation. Nonetheless, the presence of Ephedra, Nitraria, and Chenopodiaceae below 707 cm appears to indicate a treeless alpine desert landscape during the LGM. Occurrences of Betula pollen could be the result of redeposition of former interglacial sediments or related to long-distance transport (Herzschuh et al., 2006a).

A prominent feature of Postglacial vegetation changes on the Tibetan Plateau is the Betula expansion (in the source areas, not necessarily the Qinghai basin) around 11 000 cal yr BP (Herzschuh et al., 2006a), which then was followed by increases in Pinus around 8000 cal yr BP, indicating the onset of mid-Holocene Thermal Maximum. This pattern has been confirmed by a number of palynological studies in the neighboring areas (e.g. Van Campo et al., 1996; Tang et al., 2000), so that these pollen horizons can be used as time markers for synchronizing paleoclimatic records. Higher percentages of Betula and coniferous tree pollen have been interpreted as the presence of mixed forests in the alpine landscape of the Qinghai basin during the mid-Holocene Thermal Maximum (e.g. Tang, 2002). However, the distribution of modern vegetation on the Tibetan Plateau is governed by climate gradients—montane forests occur to the southeast and alpine steppe–desert to the western interior (Yu et al., 2001). No broad-leaved trees grow in the Lake Qinghai area today. Based on regional distributions of modern vegetation, the colonization of broad-leaved trees in this area during the middle Holocene would require more than a ca 1000 m upward shift of timberline and a climate at least 5 °C warmer than present. However, quantitative climate reconstructions from fossil pollen in the Qilian Mountains indicate that the mid-Holocene Thermal Maximum in the NE Tibetan Plateau was only 1–2 °C warmer than today (Herzschuh et al., 2006a). Therefore, we believe that the broad-leaved tree pollen are exotic and represent long-distance transport from their source areas when the early summer monsoon winds are strong. If so, the application of transfer functions to climate reconstructions from pollen records in this area would be complicated.

3.3.4. Ostracode carbonate trace elements and lake-water chemistry

In many lake environments, climate plays a major role in lake water chemistry by changing the precipitation/evaporation balance and thus the salinity of lake waters (Williams, 1966). As in marine carbonates, the Mg/Ca and Sr/Ca ratios in lacustrine carbonates change as a function of temperature and salinity, such that the Mg/Ca ratio is a reflection of both water temperature and salinity, and the Sr/Ca ratio is mainly a function of salinity (Chivas et al., 1986). In lake carbonates, these changes in trace-element chemistry can be recorded in ostracode shells, where the trace-element ratios are usually proportional to those in the ambient waters.

Several attempts to reconstruct salinity from trace elements in ostracode shells using the methods of Chivas et al. (1986) have been made for Lake Qinghai. These studies, including Zhang et al. (1989a, 1994) and Zhang et al. (2004), used partitioning coefficients determined by various calibration methods. However, several problems exist in using either Mg/Ca or Sr/Ca ratios to reconstruct water temperature or salinity in Lake Qinghai. First, because of the high Mg content of Lake Qinghai waters (Table 3), the primary precipitated carbonate is aragonite (Liu et al., 2003a). Although ostracode shells are made of low-Mg calcite, they can show evidence of aragonite overgrowths and possible diagenetic alteration in Lake Qinghai (Henderson, 1999). Such alteration obviously limits the usefulness of Mg/Ca ratios. Secondly, existing calibrations of the partitioning coefficient for Sr/Ca in relation to salinity are for low-Mg calcite, either in lacustrine (Chivas et al., 1986) or marine environments. Such calibrations show a linear relationship between the Sr/Ca partitioning coefficient and salinity. No accepted calibration function exists for Sr/Ca in ostracode shells in the presence of aragonite, as it preferentially absorbs Sr from lake water during aragonite precipitation, thus complicating the linear Sr/Ca relationships to salinity in the calcite shells of ostracodes (Engstrom and Nelson, 1991). Therefore, salinity reconstruction based on trace element geochemistry of biogenic carbonate is extremely problematic for Lake Qinghai.

Table 3
Composition and changes of water chemistry in Lake Qinghai (after CAS-LZ and RCREWC-CAS, 1994; Chen et al., 1990)

Year	1872	1880	1962	1986	1991
K^+	0.12	0.11	0.15	0.16	0.16
Na^+	3.28	3.71	3.26	3.75	3.93
Ca^{2+}	0.19	0.00	0.01	0.01	0.01
Mg^{2+}	0.31	0.73	0.82	0.79	0.79
Cl^-	1.91	5.40	5.28	5.87	5.79
SO_4^{2-}	0.87	2.11	2.03	2.38	2.35
HCO_3^-	–	–	0.53	0.69	0.68
CO_3^{2-}	–	1.73	0.42	0.52	0.52
Salinity	10.97	13.34	12.49	14.15	14.23

Note: Unit in mg/L.

3.3.5. Carbonate stable isotopes and lake hydrological and thermal conditions

Pioneering stable-isotope measurements on ostracode shells were conducted by Lister et al. (1991) on samples from Core QH85-14B (Fig. 4A). This was the longest lacustrine carbonate oxygen-isotope record from the Tibetan Plateau at that time, and provides valuable insights into the dynamics of the Asian monsoon during the Lateglacial and Holocene (Wei and Gasse, 1999). Similar changes were revealed subsequently by Zhang et al. (1989a) from Core QH85-16A (Fig. 4B) and recently by Liu et al. (2007) from QH2000 (Fig. 4C). High values of $\delta^{18}O$ prior to 14 500 cal yr BP reveal cold and dry climate conditions. Fluctuating but decreasing $\delta^{18}O$ values occurred after 14 500 cal yr BP, suggesting an unsteady strengthening of summer monsoon. The advance of monsoon front slowed

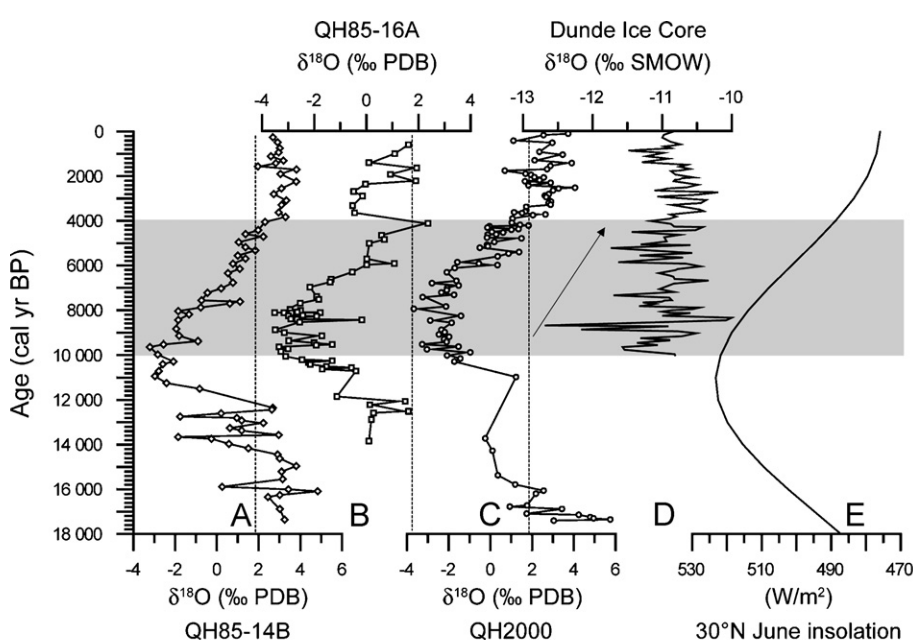

Fig. 4. Comparation of Lake Qinghai ostracode stable oxygen isotope records of Cores QH85-14B (Lister et al., 1991), QH85-16A (Zhang et al., 1989b), and QH2000 (Liu et al., 2007) with Dunde ice-core record (Thompson et al., 1989). Vertical lines indicate present-day $\delta^{18}O$ value of lake water. Summer insolation data are from Berger and Loutre (1991).

down at about 12 000 cal yr BP, possibly corresponding to the Younger Dryas stadial (Yu and Kelts, 2002). The summer monsoon culminated around 10 000 cal yr BP. A nearly 6‰ enrichment of $\delta^{18}O$ from about 11 000 to 4000 cal yr BP suggests a gradual retreat of the summer monsoon. This trend also can be observed in the Dunde ice-core $\delta^{18}O$ record (Fig. 4D). Parallel changes of the Dunde record with the Lake Qinghai records suggest that, unlike the Guliya ice core, continuous depletion of heavy oxygen isotopes in the Dunde ice core may be dominated by the "amount effect" associated with changes in monsoon intensity, modulated primarily by summer insolation to the middle latitudes (Fig. 4E). However, interpreting stable isotope records in terms of monsoon variations may be complicated due to spring water discharge to the lake, evidence of which is shown by the local presence of tufa (Zhang and Zhang, 1994).

Zhang and Zhang (1994) also reconstructed water temperature changes during the Holocene by assuming a temperature dependence of $\delta^{18}O$ fractionation between ostracode shells and ambient waters. However, in a thermally stratified lake with a relatively constant hypolimnion temperature (ca 6 °C) like Lake Qinghai, ostracode $\delta^{18}O_o$ is largely independent of temperature. Lake water $\delta^{18}O_w$ relates to the balance between the intensity of precipitation associated with the Asian Monsoon and evaporative effects in the closed basin (Wei and Gasse, 1999; Johnson and Ingram, 2004). Therefore, episodic depletions of $\delta^{18}O_o$ during the Lateglacial, e.g. the Bølling and Allerød interstadials, are likely to be manifestations of monsoon variations, rather than a signature of glacial meltwater discharge (e.g. Lister et al., 1991).

Unlike ostracode shells, the oxygen isotopes of authigenic carbonates ($\delta^{18}O_c$) is controlled by both the temperature and oxygen isotope composition of surface water ($\delta^{18}O_w$). The $\delta^{18}O_c$ is systematically higher than $\delta^{18}O_o$ in Lake Qinghai, and it has a large variability because it is subject to temperature effects in addition to the dominating evaporative effects. Therefore, comparing $\delta^{18}O_c$ with $\delta^{18}O_o$ may provide a new approach to estimating the vertical thermal structure of the lake (Kelts and Talbot, 1990), if the vital effects of the ostracodes are known. An example comes from Core QING6, where a systematic offset between $\delta^{18}O_c$ and $\delta^{18}O_o$ can be observed (Henderson et al., 2003). Assuming no isotopic stratification of the water body, this difference should be a result of oxygen-isotope fractionation induced by a temperature gradient between surface and bottom waters. This observation confirms the presence of a stable thermocline during summer, as suggested by thermodynamic hydrologic models (Qin and Huang, 1998b).

3.4. The QING-6, QING-10, QHN3/1, QH0407-C, and GAHAI-1 records

Climate changes during the last 1500 years were revealed by $\delta^{18}O$ and $\delta^{13}C$ records of fine-grained carbonate in Cores QINH-6 and GAHAI-1 (Fig. 1B). These records show parallel changes and both cores indicate that regional climate has experienced considerable variability, including three distinct phases related to the Medieval Warm Period (A.D. 800–1200), the Little Ice Age (A.D. 1200–1850), and the post-Industrialization warming (Henderson, 2004). Comparing the $\delta^{18}O$ record from Core QH0407-C with instrumental data suggests that the isotopic composition of fine-grained carbonates, which are formed in the epilimnion of the lake, primarily reflects the isotopic composition of the near-surface waters, which in turn is controlled by evaporative concentration associated with summer air temperature (Xu et al., 2006b). Consistent with these results are trends in both lake water temperature and salinity estimated in alkenone-based reconstructions on Core QHN3/1(Liu et al., 2006). The coherent changes in air temperature and precipitation are primarily related to variations of the Asian Summer Monsoon system.

3.5. Recent changes

3.5.1. Lake levels and shorelines

Past lake-level changes can be inferred from elevated terraces and beach ridges, although these features may be affected by local tectonics. Lacustrine terraces and beach ridges along the southern coast of Lake Qinghai have been described by Yuan et al. (1990) and Wang and Shi (1992). For example, three terraces can be identified at Jiangxigou. The first one above the Erlangjian terrace is very broad (3–4 km), and mainly overlain by fine sand. Its altitude is about 10 m above present lake level. Radiocarbon ages on organic clay indicate that this terrace was formed between 2000 and 1000 ^{14}C yr BP. Below this terrace, six beach ridges situated at 1.5, 2.0, 3.0, 3.5, 5.7, and 8.7 m, respectively, were found (Yuan et al., 1990; Wang and Shi, 1992), indicating stillstands of the lake level superimposed upon a long-term lowering trend induced by either climate or tectonic uplift or both. None of the beach ridges are covered by loess, implying a young age. According to Kozloff's (1909) map, a 7-m-high beach ridge now separating Lake Erhai from Lake Qinghai must have developed sometime after 1909. In Daotang valley, the highest beach ridge, dated to 1230 ± 60 ^{14}C yr BP (Wang and Shi, 1992), can be correlated to the first terrace at Jiangxigou.

The level of Lake Qinghai is very dynamic. The first lake-level measurements were carried out by O.N. Potanen during 1884–1886, and then by B.A. Obruchev and W. Filchner (cf. Chen et al., 1990). According to their results, the lake level was ca 3205 m asl in the late 1880s compared to 3193 now—it has been lowered by about 12 m during the last century or so. Regular meteorological and hydrological observations started in 1950. Instrumental data show that the lake level has continued to fall since 1958 (Fig. 5A). This is mainly caused by the 20th century warming in the area (Fig. 5B), which led to a negative precipitation–evaporation balance (Fig. 5C) and reduced river runoff

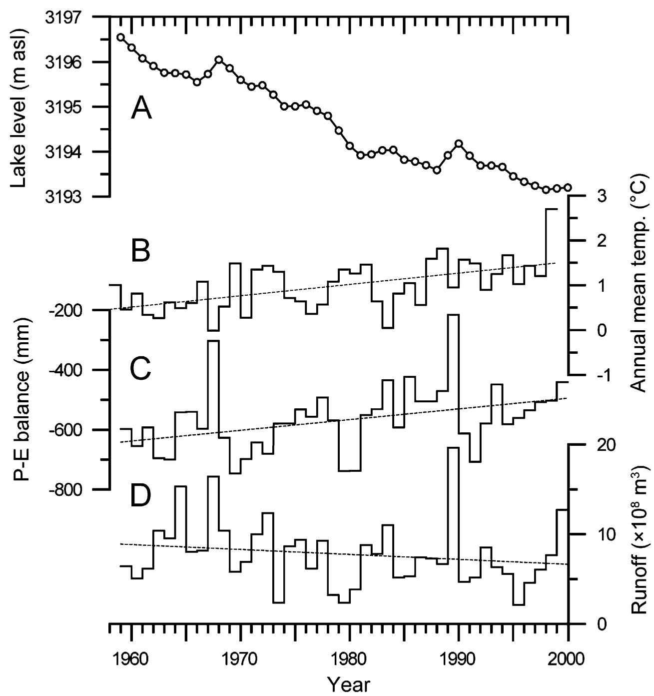

Fig. 5. Changes in lake level (A), annual mean temperature (B), precipitation–evaporation balance (C), and the Buha River runoff (D) since 1958. Dashed lines indicate the trend defined using a linear regression method.

(Fig. 5D). Agricultural and industrial water consumptions only account for 8% of this change (Qu, 1994). Four GCM experiments unanimously suggest that the local warming trend would continue under a scenario of doubling CO_2 (Qin and Huang, 1998a). A hydrological model forced by the GCMs' outputs predicts dramatic increases in both summer precipitation and river runoff (Qin and Huang, 1998b), but they are not enough to offset increased evaporation. Consequently, the historical trend of lake-level fall is likely to continue.

Lowering of lake level also led to progressive retreat of the shoreline and thus changes in the coastal landscape. According to Kozloff's (1909) map, Haiyan Bay and Lake Erhai used to be part of Lake Qinghai. Now the former has been isolated and the latter resembles a lagoon (Fig. 1B). In 1956, Bird Island was about 3800 m from the shore, and in 1972, only 2600 m (Guo, 1997). In a 1975 satellite image, this island was still separated from the mainland, but it was connected to the mainland by 1979 (Guo, 1997). Human activities, along with climate change, have also played an important role in modifying the coastal landscape. The Buha Delta prograded at a rate of 2 km/a during the period from 1956 to 1968 (the so-called "Great Leap Campaign"), primarily caused by catchment erosion associated with extensive land reclamation (Zhang, 2003). Lobes of this delta have also changed greatly during the last half century: the active lobe was located 6.2 km north of Bird Island before the 1950s (Guo, 1997), but now is 4 km southwest of the island. The morphology of the northern shore also experienced substantial changes over the last 25 years. The shoreline there advanced at a rate of 130 m/a due to increased catchment erosion (Zhang, 2003).

3.5.2. *Water chemistry*

No regular measurements of water chemistry have been made for Lake Qinghai. The earliest measurements were carried out by K. Schmidt in 1872 (cf. Chen et al., 1990). In 1962 and 1986, the Lanzhou Institute of Geology, CAS, and the Institute for Saltwater Lake Studies, CAS, performed routine measurements (CAS-LZ and RCREWC-CAS, 1994). The results indicate that the lake water is dominated by sodium and chloride. Sulfate and magnesium are also high. The chemical composition of the lake water did not change much during the last century (Table 3). The major biogeochemical processes in the lake is $CaCO_3$ precipitation and biological reduction of sulfate to sulfide (Yan et al., 2002), which governs the concentration of magnesium and heavy metals in the lake water, respectively.

4. Discussion and comparison with surrounding areas

4.1. *Late Cenozoic changes*

Loess-soil sequences and the underlying *Hipparion* red clay formation contain abundant information on past climate changes in the western interior of China. For example, the Xifeng section in the western Loess Plateau (Liu et al., 2001), containing a red clay sequence overlain by a loess-soil complex, reveals a detailed history of the Asian monsoon system during the Late Cenozoic. Multiple stratigraphic analyses reveal that the eolian red clay started to accumulate at ~8 Ma, evidently indicating the inception of the modern East Asian winter monsoon. This inception is most likely associated with the thermodynamic forcing of the rising Tibetan Plateau. Subsequent strengthening of the East Asian winter monsoon and the corresponding weakening of the summer monsoon appears to be a manifestation of global cooling after 3.6 Ma, especially when ice sheets expanded in the Northern Hemisphere at about 2.7 Ma (Maslin et al., 1998; An et al., 2001 Haug et al., 2005). Magnetic susceptibility records of the Xifeng section reveal low amplitude fluctuations of the Asian monsoon system with a 41-ka frequency prior to ~0.8 Ma (Fig. 6A), which were then replaced by large-amplitude fluctuations with a 100-ka frequency, correlative with the deep-sea oxygen isotope record (Shackleton et al., 1995). The in-phase changes in the climate of inland China with global ice volume (Fig. 6B) imply a common force thought to be the variations in the orbital parameters of the earth.

Like the Loess Plateau, climate conditions on the Tibetan Plateau are influenced by the Asian monsoon system, which brings vapor to the interior of the north-western China and feeds the lakes there. Like marine records (Prell and Kutzbach, 1992; Rea et al., 1998), continuous lacustrine records, can provide information, potentially complementary, about long-term changes in the

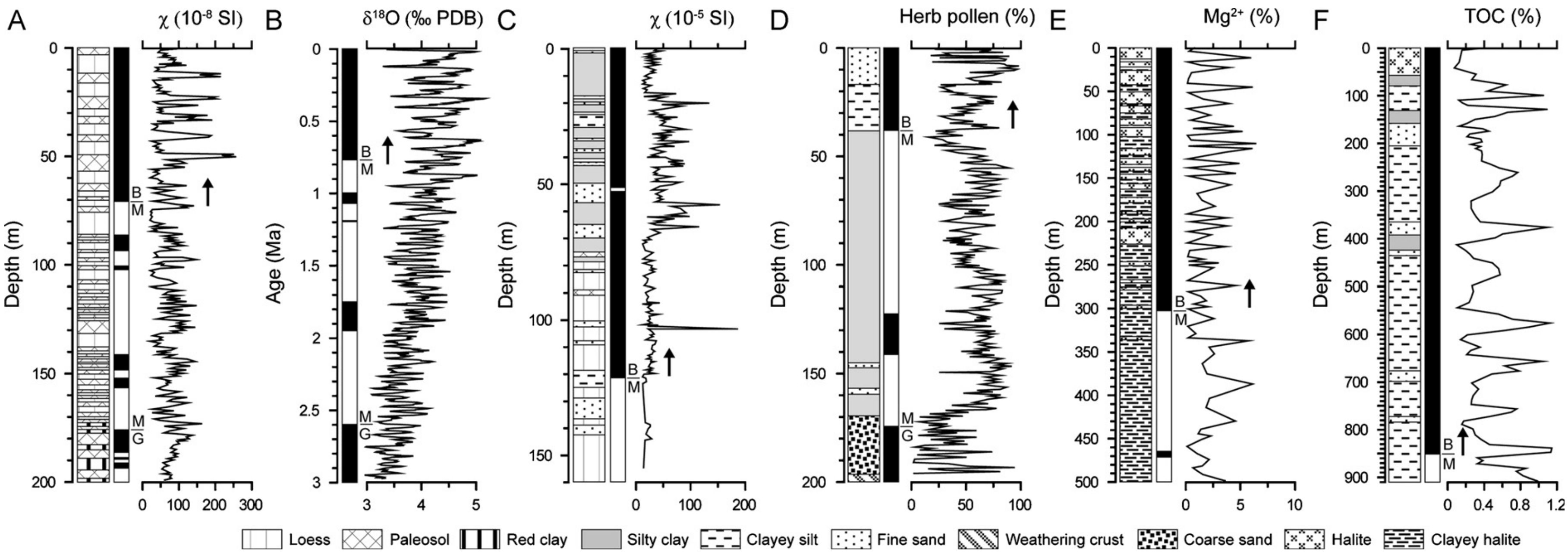

Fig. 6. Comparison of Pleistocene lacustrine records from the Tibetan Plateau with loess-soil and marine records. Arrows indicate the mid-Pleistocene transition. (A) Xifeng site, W. Loess Plateau (Liu et al., 2001); (B) Equatorial Pacific (Shackleton et al., 1995); (C) Lake Qinghai Core QH86, NE Tibetan Plateau (Yuan et al., 1990); (D) Co Ngoin, N. Tibetan Plateau (Lü et al., 2001); (E) Qunteyi Playa Core ZK3208, Qaidam Basin (Han et al., 1991); (F) Qarhan Playa Core CK-6, Qaidam Basin (Huang and Chen, 1990).

Asian monsoon. Magnetic-susceptibility record of Core QH86 (Yuan et al., 1990) suggests orbital-scale fluctuations of Quaternary climate in China's inland (Fig. 6C). Large-amplitude oscillations of lake level occurred after ~0.8 Ma, as indicated by repeated occurrences of silty clay above the B/M polarity boundary. High-resolution pollen analyses from the Lake Co Ngoin core (Lü et al., 2001) reveal dampened fluctuations of the inland vegetation (and presumably climate) prior to ~0.8 Ma (Fig. 6D), which was then replaced by longer-period, large-amplitude changes. Reciprocal changes of alpine steppe and coniferous forests at a landscape scale after ~0.8 Ma are regulated by regional climate, rather than by inter-specific competition. A similar pattern of climate fluctuations is also registered in long drill cores from two playas in Qaidam Basin (Huang and Chen, 1990; Han et al., 1991). The halite-clay sequences (Fig. 6E and F), comparable to loess-soil sequences, reveal periodical changes in dry/wet conditions associated with substantial variations of the Asian monsoon after ~0.8 Ma. Further evidence comes from a long drill core in Zoige Basin, where the organic carbon record suggests a similar pattern of inland climate fluctuations at orbital scales (Wang and Xue, 1997; Chen et al., 1999). Although mechanisms behind the shift from obliquity-dominated to eccentricity-dominated fluctuations, referred to as the "mid-Pleistocene transition," remain enigmatic, the existing long records do reveal an astronomical forcing of China's inland climate during the Quaternary.

In contrast to marine cores, lake sediments typically have higher sedimentation rates and thus can achieve finer time resolution. Lithological variations in lacustrine cores indicate different depositional environments, commonly having different sedimentation rates. Therefore, paleomagnetic stratigraphy alone is unable to provide detailed time constraints on long lacustrine records. In many of lakes on the Tibetan Plateau, carbonates dominate the authigenic minerals. Uranium-series disequilibrium dating on authigenic carbonates in the Balikun (Ma et al., 2004), Qarhan (Huang and Chen, 1990), and Qinghai (i.e. QH86; Shan et al., 1993) cores has proven promising accurate chronology for Pleistocene lacustrine records. Long and high-resolution record from Lake Qinghai with independent age controls will provide an opportunity not only to examine the dynamics of continental paleoclimate in both time and frequency domains, but also to evaluate the influence of reorganization of ocean circulation on the Asian monsoon system at millennial time scales.

4.2. The Last Interglacial–Glacial cycle

4.2.1. Delayed response of the summer monsoon during the MIS6–MIS5e transition

Periodical variations in Earth's orbital parameters have long been regarded as a pacemaker of past glacial–interglacial cycles. In some cases, however, discrepancies exist with orbital parameter, particularly for Termination II (Winograd et al., 1988). The tuning of most marine records to fit orbital time scale, precludes the opportunity to address this problem. Therefore, high-resolution, well-dated terrestrial records are important for understanding the mechanism and timing of deglaciation.

A speleothem record of $\delta^{18}O$ from Dongge Cave (not shown; Yuan et al., 2004) indicates that the Asian summer monsoon slightly lagged behind insolation (Fig. 7A) during Termination II, but was almost concurrent with the rise of global sea level as revealed by benthic foraminifera $\delta^{18}O$ record (Fig. 7B). A stacked marine record (Fig. 7C) indicates that full monsoon conditions did not occur until global sea level reached an interglacial position, which would enable the establishment of the Pacific warm pool, until about 125 ka (Clemens and Prell, 2003). No cores from the last interglaciation yet exist for Lake Qinghai, but records from Zoige Basin (Fig. 7D), the Qarhan Playa (Fig. 7E), and the Loess Plateau (Fig. 7F) reveal a much more delayed response of the Asian summer monsoon to insolation. This difference will be scrutinized in the Lake Qinghai drill core.

Comparing these records with those from the western Tibetan Plateau and Tarim Basin may also reveal spatial variability of monsoon activity in this area during the Last Interglacial–Glacial cycle. The Guliya ice-core record (Fig. 7G) indicates less negative $\delta^{18}O$ values during interstadials, suggesting that the enrichment of oxygen isotopes in regional meteoric water was dominated by the "temperature effect" rather than the "amount effect" related to monsoon intensity that otherwise would deplete the heavy ^{18}O. Further evidence comes from Lakes Tianshuihai (Fig. 7(H) and Balikun (Fig. 7I), which are located at the west side and north of the plateau, respectively. These two $\delta^{18}O$ records resemble each other but are different than other records such as Zoige in the eastern part of the Plateau, indicating that the summer monsoon did not reach the western Tibetan Plateau and the Tarim Basin even during the Last Interglacial.

Loess records (Porter and An, 1995) reveal millennial-scale variability of the Asian summer monsoon correlative with the North Atlantic Heinrich events during the Last Glaciation. And speleothem records (Wang et al., 2001) include even shorter events known as Dansgaard–Oeschger oscillations. However, existing lacustrine records from the Tibetan Plateau do not have the temporal resolution to identify these climate events. High-resolution lacustrine records from this area are needed not only to evaluate the rapidity of millennial-scale changes in summer monsoon, but also to examine the geographical extent of Heinrich and Dansgaard–Oeschger events outside the North Atlantic realm.

4.2.2. The MIS3 period of anomalously high lake levels

Lake sediments and landforms in the Qaidam basin (Chen et al., 1990) and other lakes on the Tibetan Plateau record a period of high lake levels attributed to MIS3 (Herzschuh, 2006b). In addition, the Guliya ice-core record

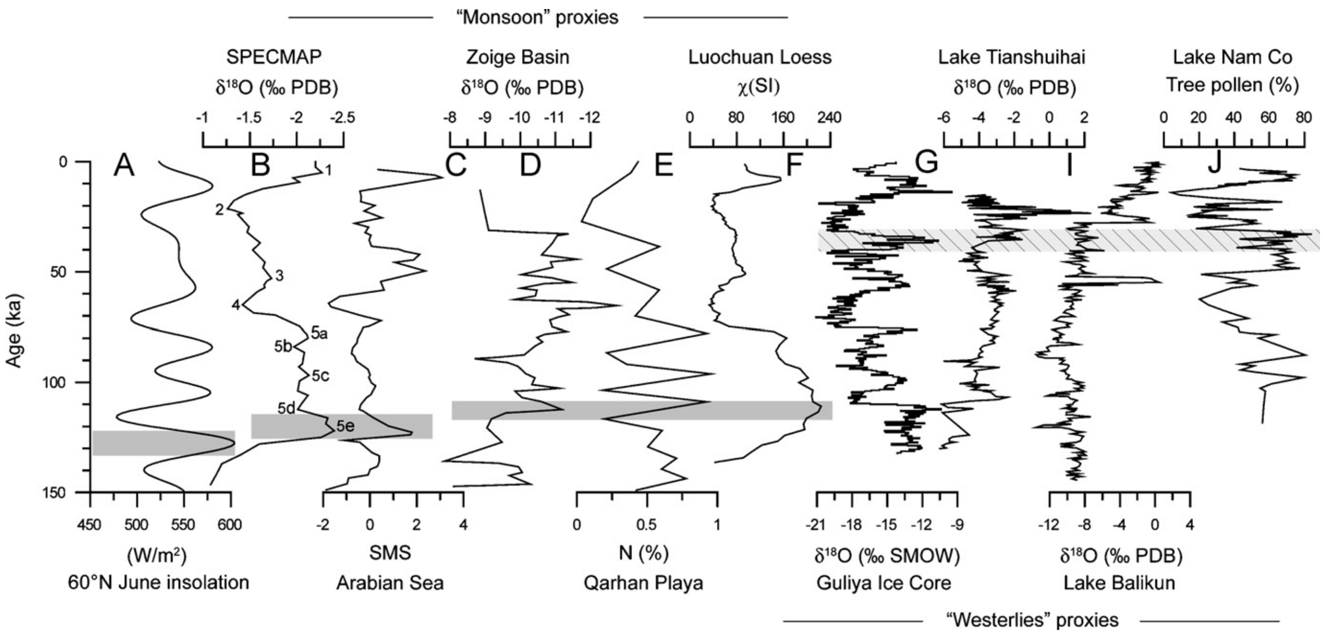

Fig. 7. Comparison of lacustrine records from the Tibetan Plateau with insolation and SPECMAP δ^{18}O record during the last 150 000 years. (A) Summer insolation (Berger and Loutre, 1991); (B) SPECMAP δ^{18}O record (Imbrie et al., 1993); (C) Arabian Sea summer monsoon stacked (SMS) record (Clemens and Prell, 2003); (D) Zoige Basin (Wu et al., 1997); (E) Qarhan Playa (Huang and Chen, 1990); (F) Luochuan section, the Loess Plateau (Xiao et al., 1999); (G) Guliya Ice Core (Thompson et al., 1997); (H) Lake Tianshuihai (Zhou and Zhu, 2001); (I) Lake Balikun (Ma et al., 2004); (J) Nam Co (Wu et al., 2004). Core chronologies are those estimated in the original studies. Arabic numerals denote Marine Isotope Stages. Shaded bands highlight the phase relationship of summer monsoon with insolation and global ice volume. The late MIS3 interval of extremely high lake levels is indicated by the hatched band.

Table 4
Reconstructed precipitation on the Lake Qinghai watershed using an energy-balance model

Age (cal yr BP)	Lake level anomaly (m)	Lake surface area (km^2)	Evaporation (mm)	Precipitation (mm)	River runoff ($\times 10^8$ m^3)	Data sources
40 000–30 000	+104	8100	1015.0	645 ± 5	–	Jia et al. (2000)
7500–5000	+45	6406	–	595 ± 15	–	Jia et al. (2000)
7000–6000	+45	6225	1096.7	528.9	35.35	Qin (1994)
7200–6000	+45	6225	1079.4	526.6	34.41	Qin (1997)
Modern	0	4448	954.5	407.2	20.56	Qin (1997)

(Fig. 7G) reveals an anomalous warming on the Tibetan Plateau at the end of MIS3 (Thompson et al., 1997). High percentages of tree pollen in the Lake Nam Co core also suggest a warm and wet phase between 40 and 30 ka (Fig. 7J). This pattern is considerably different from global ice volume indicated by the SPECMAP benthic oxygen isotope record (Fig. 7B).

Further evidence of high lake levels on the Tibetan Plateau comes from a large number of elevated terraces in several basins located at altitudes ranging from 10 to 280 m above present-day lake levels (Lehmkuhl and Haselein, 2000; Shi et al., 2002; Yang et al., 2004). For example, four terraces can be identified on the southern shore of Lake Qinghai. The highest one, situated at ~120 m above present lake level, was interpreted as having formed about 38 000 cal yr BP, and a second one at ~45 m was dated to about 7500 cal yr BP (Chen et al., 1990; Yuan et al., 1990). Lake terraces are the product of stable lake levels, when precipitation is balanced by evaporation. Therefore, an energy-balance model, originally developed by Kutzbach (1980) for a hydrologically closed watershed, can be used to estimate past precipitation using estimated watershed evaporation (Qin, 1994, 1997; Jia et al., 2000). Evaporation is catchment-dependent and can be parameterized using vegetation cover and the insolation values during specific time intervals. The reconstructed annual precipitation and lake-surface evaporation suggest that MIS3 interval of high lake levels is ~1.5 times higher than that of present (Table 4). Higher lake-surface evaporation was presumably offset by increased river runoff to maintain stable lake levels.

The volume of ice during the LGM on the mountains in the Lake Qinghai basin has been estimated at 200 ± 50 km^3 (Wang and Shi, 1992). Conservatively assuming no change in lake surface area (4400 km^2), melting of these mountain glaciers would have only raised lake level by 41 ± 10 m.

Therefore, meltwater from mountain glaciers alone cannot account for the ~104-m-high relative lake level, and tectonic deformation must be considered. On the Erlangjian terrace, a cross-bedded gravel unit that presumably was deposited near the shore suggests that the relative lake level has only dropped ~6 m since the LGM (Yuan et al., 1990).

Some pollen records, mostly from relatively large lakes, have high amounts (ca 30% of the total) of tree pollen at 30–40 ka, which has been interpreted as indicating relatively warm, moist climate, perhaps comparable to conditions in the early Holocene (e.g. Herzschuh et al., 2006a). The tree pollen in such sequences is dominated by *Abies* and *Picea*, which are derived locally (Herzschuh et al., 2006a), as confirmed by the presence of their macrofossils (Wang and Shi, 1992). Tree pollen that might be interpreted as indicating mixed forests, *Pinus* and *Betula* pollen, only account for about 5%, and thus probably represent long-distant transport rather than local mixed forests. Mixed forests of *Pinus* and *Betula* do not occur on the Tibetan Plateau today (Yu et al., 2001), although trap studies reveal the presence of their pollen (Cour et al., 1999). Thus, available pollen evidence from the Lake Qinghai area do not support conditions significantly wetter than the early Holocene during MIS3.

The altitude of the highest terraces around Lake Qinghai exhibit spatial variability, implying complex tectonic activities (Chen and Lin, 1993). For example, at Jiangxigou, the highest terrace is situated at ~104 m. A radiocarbon date of 12 100 ± 265 yr BP on charcoal (Wang and Shi, 1992) suggests that this terrace was formed during the Bølling-Allerød interstadials, rather than during MIS3. This compares with the 120 m terrace, with an estimated age of 38 000 cal yr BP, discussed earlier.

The mechanism responsible for the supposed MIS3 moist interval is open to debate. Because this period shows a perfect correlation with higher summer isolation at 30°N (Berger and Loutre, 1991), when the perihelion (i.e. precession-controlled insolation) minima occurred during summer, Shi et al. (2001) ascribed this event to the strengthening of the Asian monsoon driven by the precession cycle. However, most monsoon records do not show any strengthening between 40 and 30 ka (Figs. 7C–F). Furthermore, the Indian monsoon tends to be more sensitive to the obliquity component of orbital forcing (Clemens and Prell, 2003), which governs latent heat transport from the southern subtropical Indian Ocean and thus controls the thermal contrast between the Asian landmass and the ocean. Yang et al. (2004) argued against the monsoon hypothesis, suggesting that a moist climate during MIS3 could be better explained by the strengthened westerlies. The Tianshuihai and Balikun records (Figs. 7H and I) may provide an opportunity to test this idea, because climate conditions there are dominantly controlled by the westerlies. Carbonate $\delta^{18}O$ values in both cores progressively increase during the MIS 3, implying an aridification trend into the LGM.

4.3. Lateglacial–Holocene changes

A large number of lacustrine records cover the last 14 500 years on the Tibetan Plateau. Carbonate $\delta^{18}O$ records and their paleoclimatic implications have been thoroughly reviewed by Wei and Gasse (1999), and palynological work was recently summarized by Shen (2003). All of these records reveal an initial warming of regional climate after ~14 500 cal yr BP, accompanied by penetration of the summer monsoon into the eastern part of the plateau (He et al., 2004). Warm and wet conditions probably facilitated human colonization in this area (Madsen et al., 2006). These trends were punctuated by a cold spell between 12 500 and 11 000 cal yr BP, probably corresponding to the Younger Dryas interval (Gasse and Vancampo, 1994), although existing chronologies leave uncertainties about Younger Dryas correlatives. A gradual cooling of regional climate occurred after 10 000 cal yr BP (Herzschuh et al., 2006c), along with a progressive retreat of the summer monsoon front (Hong et al., 2003). Regional climate turned distinctly cooler and drier about 5000 cal yr BP, as indicated by a sudden transition from temperate steppe to alpine steppe (Herzschuh et al., 2006c). The deteriorations of climate conditions are thought to be the major driving force in the termination of Neolithic cultures in this area (Rhode et al., 2007). Superimposed upon this trend were centennial-scale fluctuations with small amplitude, presumably induced by variation in solar activity (Ji et al., 2005).

Because Lake Qinghai is situated near the limit of monsoonal precipitation, variations of the Asian summer monsoon may have left its fingerprint on the lake's sediments. Therefore, the Lake Qinghai record is important for examining the spatial variability of the Asian Summer Monsoon, along with other records from neighboring areas. The $\delta^{18}O$ records of lakes in the western part of the Tibetan Plateau, e.g. Sumxi-Longmu Co (Fontes et al., 1993), and Bangong Co (Fontes et al., 1996), exhibit a different pattern than the Lake Qinghai record (Lister et al., 1991), implying that the Indian Summer Monsoon did not reach the western Tibetan Plateau during the Holocene. However, several lakes, including Lakes Manas (Rhodes et al., 1996) and Issyk-Kul (Ricketts et al., 2001) show patterns of increased moisture during the mid-Holocene Thermal Maximum, similar to the record at Qinghai (Lister et al., 1991). It is not clear whether the moist intervals at these lakes were due to monsoonal precipitation or other causes.

5. Summary and conclusions

1. Comparisons of long Pleistocene lacustrine and loess-soil records with the marine oxygen isotope record show in-phase changes of China's inland climate with global ice volume, presumably driven by periodical variations in the Earth's orbital geometry. Once the plateau reached its threshold altitude for modifying the prevail-

ing wind regime, subsequent uplift had little influence on regional climate. No independent evidence supports a tectonic forcing of these glacial–interglacial cycles.

2. Lacustrine carbonate $\delta^{18}O$ and pollen records from the Tibetan Plateau suggest a delayed response of the Asian monsoon system to orbital forcing during Termination I and II. Strengthening of the summer monsoon lagged behind sea level by several thousand years, implying the importance of sea surface conditions in the development of monsoon circulation. Marine records from neighboring areas exhibit nothing anomalous in the Indian Summer Monsoon during MIS3, compared to the SPECMAP record. In addition, a lacustrine $\delta^{18}O$ record from the northern margin of Tarim Basin reveals a steady drying trend through late MIS3 into the LGM. Therefore, the extremely high lake levels and inferred warming postulated for the Tibetan Plateau remains an enigma.
3. A wealth of evidence from Lake Qinghai reveals a teleconnection of China's inland climate with that of high northern latitudes during the Lateglacial and Holocene. Initial warming began at about 14 500 cal yr BP, marking strengthening of the Asian summer monsoon after the last glaciation. The Lateglacial climate was remarkably unstable, as indicated by several cold reversals presumably related to glacial boundary conditions before the disappearance of Northern Hemisphere ice sheets. Steady warming started at 11 000 cal yr BP and terminated about 4000 cal yr BP. High summer insolation resulted in a strong Asian summer monsoon between 11 000 and 7000 cal yr BP. A striking feature of the Lake Qinghai record is the variability of the mid-Holocene Thermal Maximum at centennial time scales, probably caused by localized feedbacks, because similar changes have not been reported elsewhere.
4. Geological evidence suggests that the continued lowering of water level in Lake Qinghai within the context of global warming during the last half century is part of a long-term trajectory of aridification. Human impacts are minor in the Lake Qinghai basin. Modern studies of the limnological processes in the lake are badly needed. Such studies are important not only for interpreting paleolimnological data, but also for calibrating model parameters.

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Note added in proof

After completion of this paper, we became aware of an unpublished Ph.D. dissertation (Yu, J.-Q., 2005, Lake Qinghai, China: A multi-proxy investigation on sediment cores for the reconstructions of paleoclimate and paleoenvironment since the Marine Isotope Stage 3, University of Technology, Darmstadt, Germany, 119pp.) that deals with many aspects of the paleoclimate record of Lake Qinghai, only some of which has been published (see references). In particular, the dissertation describes a 1987 drilling effort that recovered a 26 m core (Q87) from the eastern basin of the lake. Limited data from this core suggest that it extends through sediments of MIS 3 age and that the lake was than present throughout that time.

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