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ORIGINAL PAPER

Lake evolution and hydroclimate variation at Lake Qinghai (China) over the past 32 ka inferred from ostracods and their stable isotope composition

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Abstract We used ostracod species assemblages and their $\delta^{18}O$ values in a 32-m sediment core from Lake Qinghai, China, along with information from cores collected at other sites in the lake, to infer lake evolution and hydroclimate changes since the last glacial. Dominant ostracod species *Ilyocypris bradyi* and its low $\delta^{18}O$ values showed that Lake Qinghai was small in size or even consisted of several playa lakes, and the 1F core site could have even been in a wetland setting, under cold and dry climate conditions before 15.0 ka. Presence of *Limnocythere inopinata* with low $\delta^{18}O$ values, and absence of *I. bradyi* after 15.0 ka, indicate the lake area increased or that the playas merged. The decrease or disappearance of ostracods with high $\delta^{18}O$ values showed that the lake shrunk under dry climate from 12.0 to 11.6 ka. After 11.6 ka, hydroclimate shifts inferred from ostracod species changes (*Eucypris mareotica* and *L. inopinata*) and their $\delta^{18}O$ values were as follows: (1) 11.6–7.4 ka—larger, but still small lake area with greater moisture availability under primarily dry climate conditions, (2) 7.4 to 3.2 ka—increasing lake level under a warmer and wetter climate, and (3) 3.2 ka to present—stable, large, brackish lake. The low ratio of lake water volume to runoff, and close proximity of the core site to freshwater input from the river mouth would have resulted in relatively lower ostracod $\delta^{18}O$ values when Lake Qinghai was small in area during the interval from 32.0 to 15.0 ka. Lower ostracod $\delta^{18}O$ values during interstadials and throughout the entire Last Glacial Maximum and early deglacial (ca. 24.0–16.0 ka) were caused by a greater contribution of seasonal meltwater from ice or snow and low incoming precipitation $\delta^{18}O$ values related to cold climate conditions in the region at that time.

Keywords Ostracod assemblage · Oxygen isotope · Hydroclimate · Lake Qinghai · Tibetan Plateau

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Introduction

Lake Qinghai is located on the arid, high-altitude northeastern Tibetan Plateau where the Asian Summer Monsoon (ASM), westerly jet stream, and Asian Winter Monsoon converge. The lake lies in a zone that is very sensitive to global climate change (Fig. 1a). Because of its unique geographic location, the importance of Lake Qinghai sediment records for understanding past global changes has been increasingly recognized since the 1870s (Chen et al. 1990). Previous studies revealed that sediment cores from

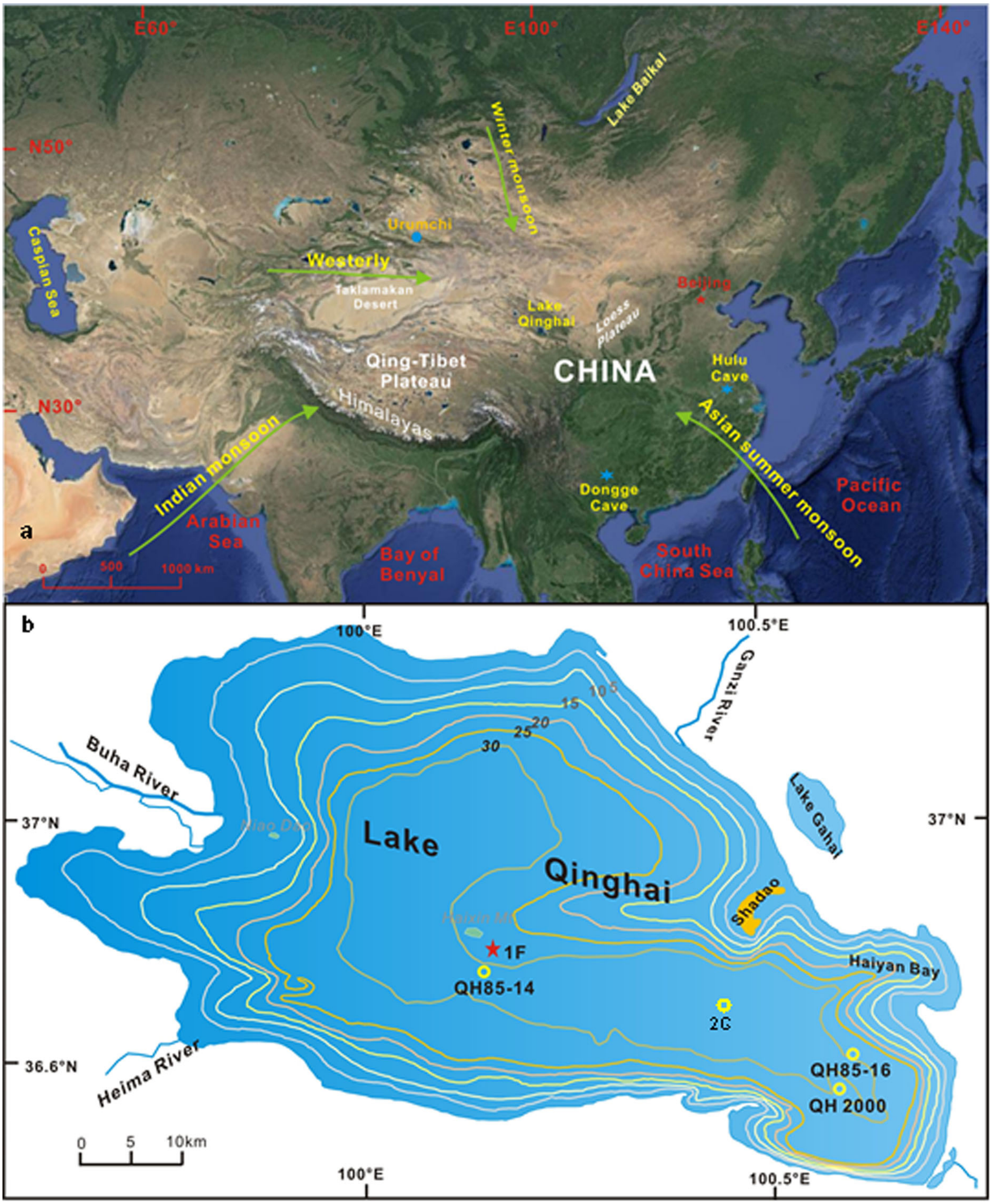

Fig. 1 **a** Location of Lake Qinghai in China and atmospheric circulation patterns of the area; **b** core sites in this study and other published reports (*red five-pointed star* shows the core 1Fs site in this study, *yellow circles* show locations of previously published cores). The 30-m *contour line* is an estimate of the maximum Lake Qinghai area before 15.0 ka. (Color figure online)

Lake Qinghai contained abundant information about regional environmental history at multiple time scales. For instance, studies using paleoenvironmental proxies including pollen assemblages (Du et al. 1989; Kong et al. 1990; Shan et al. 1993), carbonate content (Kelts et al. 1989; Huang and Meng 1991), and elemental and isotope geochemistry (Zhang et al. 1989a, b, 1994; Lister et al. 1991; Sun et al. 1991) revealed changes in lake hydrology and catchment environment during the late glacial to Holocene transition. The data suggested that environmental and hydrological conditions in Lake Qinghai were sensitive to long-term shifts in the monsoon boundary and global-scale climate changes (Wang and Shi 1992; Wei and Gasse 1999; Yu and Kelts 2002).

Paleolimnological studies focused on high-resolution reconstruction of lake hydrology and regional climate at centennial/millennial scales over the Holocene, using proxies that include sediment grain size (Zhang et al. 2003), carbonate and total organic matter content (Yu and Kelts 2002; Shen et al. 2005), stable isotope values of carbonates and total organic carbon [TOC] (Lister et al. 1991; Henderson et al. 2003, 2010; Xu et al. 2006; Liu et al. 2007, 2013; Zhang et al. 2010), rare earth elements (Shi et al. 2003), pollen (Liu et al. 2002), and biomarkers (Liu et al. 2006; Wang and Liu 2012; Liu et al. 2015a; Thomas et al. 2014, 2016). Several studies suggest the climate of this region was wettest during the early Holocene, in association with the strongest ASM (Lister et al. 1991; Liu et al. 2007; An et al. 2012; Jin et al. 2015), whereas investigations that used other paleoclimate proxies, such as OSL-based shoreline age data (Liu et al. 2015b), suggest that the early Holocene climate was relatively dry, with a weak ASM (Zhang et al. 1994; Shen et al. 2005). There are different viewpoints regarding hydroclimate in Lake Qinghai since the last deglaciation, especially with respect to hydrological conditions (i.e. high or low lake levels) in the early Holocene (Kelts et al. 1989; Chen et al. 1990; Yu and Kelts 2002; Yu 2005; Liu et al. 2013, 2014, 2015a; Wang et al. 2015a). Some studies suggested that the lake was shallow in the early Holocene, perhaps only a few meters deep, as inferred from the TOC $\delta^{13}C$ values, ostracod Sr/Ca ratios, *Ruppia* seeds, glycerol dialkyl glycerol tetraethers (reported as %thaum) and OSL-based geomorphic shoreline investigations (Zhang et al. 1989b; Yu and Kelts 2002; Yu 2005; Liu et al. 2013, 2015b). In contrast, other studies indicated that lake level was higher by as much as a dozen meters relative to the current level (z_{max} in modern Lake Qinghai = $\sim$27 m), according to geomorphic investigations, paleoshorelines and proxy records in lacustrine sediments of Lake Qinghai (Chen et al. 1990; Liu et al. 2014; Jin et al. 2015). Such different climatic and hydroclimatic inferences may be a consequence of uncertainties in the age models and different interpretations of climate proxies in Lake Qinghai.

Most previous studies concentrated on long sequences with low temporal resolution, covering the time since the last deglaciation. Alternatively, some studies focused on short sequences with high temporal resolution during the Holocene, even just the last 2000 years. There are few well-dated, high-resolution records of climate and environmental changes that occurred in the Lake Qinghai area during the last glacial period. Lake Qinghai was drilled in 2005 with the support of the International Continental Drilling Program (ICDP). The longest and highest-quality drill cores, 1F and 1A (approximately 18.6 m), were obtained from the depocenter in the southwestern sub-basin of Lake Qinghai (An et al. 2012) (Fig. 1b). Lithologic and geochemical data from cores 1F and 1A, including grain size, percent $CaCO_3$ and percent total organic carbon (TOC), and ostracod $\delta^{18}O$ values for the upper 5.0 m of core 1A were used to create a composite record that covers the last $\sim$32 ka (An et al. 2012). An et al. (2012) found that climate was affected mainly by the Westerlies during the last glacial, but ASM circulation dominated during the Holocene.

Here, we present a nearly continuous stable oxygen isotope record ($\delta^{18}O$) from ostracods, along with a record of ostracod assemblages, from core 1F in Lake Qinghai. Our ostracod $\delta^{18}O$ record is the longest yet reported and can be used to better interpret isotope values and understand paleoclimatic/paleoenvironmental changes around Lake Qinghai at the glacial-interglacial scale. We inferred past climate from ostracod $\delta^{18}O$ values by considering numerous factors, such as the lake evolution process and the distance from the core site to freshwater sources. Our approach contrasts with that of previous studies, in which ostracod $\delta^{18}O$ values were interpreted strictly in terms of effective moisture or precipitation/evaporation (P/E) ratio. Interpretation of the oxygen isotope record in the context of TOC, pollen data and other ostracod $\delta^{18}O$ records from different sites in the lake permits a detailed reconstruction of lake evolution and hydroclimate changes since the last glacial period in Lake Qinghai.

Site description

Lake Qinghai (36°32′–37°15′N, 99°36′–100°47′E) lies in a closed, intermontane basin on the Tibetan Plateau and has the largest surface area (~4260 km^2) of any inland water body in China (Fig. 1b). The lake developed within a basin that is surrounded by three mountain ranges (Bian et al. 2000), the Datong Mountains to the north, the Riyue Mountains to the east and the Qinghai Nanshan Mountains to the south. These mountains, which mostly have elevations above 4000 m, account for ~70% of the drainage area. Lake Qinghai is supplied hydrologically mainly by the Buha River and other major rivers in the western part of the basin. These rivers have created fluvial plains and deltas on the western and northern shores of the lake (Fig. 1). Dunes and beach ridges are common along the eastern shore, reflecting a prevailing westerly wind pattern (Colman et al. 2007). Fault escarpments and terraces developed extensively along the southern shore, where faulting and block tilting are still active. Glacial and periglacial landforms can be found within the Qinghai Nanshan Mountains (Porter et al. 2001), whereas small modern glaciers occur only on the mountains in the upper Buha River drainage basin. The late Cenozoic tectonic evolution of the region represents the elevation history of the northeastern margin of the Tibetan Plateau (Molnar et al. 1993). A northwest-dipping thrust fault is present along the southern range front of the Qinghai Nanshan Mountains. The mountain range is thus a tectonic ramp that is being thrust 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, ahead of the mountains. Therefore, Lake Qinghai is essentially a piggyback basin behind thrust ramps to the south and west.

Mean annual precipitation and evaporation around Lake Qinghai are 400 and 800–1000 mm, respectively (Liu et al. 2008, 2009). Precipitation shows high inter-annual variability (it has varied from 260 to 520 mm over the past 50 years), and most (60%) falls in the summer months (June–August). More than 60% of the evaporation occurs in summer (Colman et al. 2007). Within the Lake Qinghai catchment, average annual air temperature was approximately 1.2 °C (mean January and July air temperatures are −14.3 and 10.5 °C, respectively) over the last 30 years, according to meteorological data from the Haiyan, Gangcha and Tianjun weather stations. The surface of Lake Qinghai is frozen from December to March, with a maximum ice thickness of 0.8 m (Henderson et al. 2010).

Lake Qinghai has a volume of 71.6×10^9 m^3 and a catchment area of 3.0×10^4 km^2 (Cui and Li 2014, 2015). Approximately 50 rivers exceeding 5 km in length discharge seasonally into the lake, and 85% of the total discharge occurs between June and September (Cui and Li 2014). Among these rivers, the perennial Buha River has a catchment of 1.43×10^4 km^2 and the largest inflow (50% of the total runoff), with most occurring in summer (Henderson et al. 2010). Rivers that drain the surrounding area are major water sources for Lake Qinghai. Meltwater from the surrounding mountain glaciers, however, only accounts for 0.3% of the total runoff and does not appear to have made a significant contribution since the early Holocene (Lister et al. 1991; Henderson et al. 2010). These source waters, together with evaporation, represent the dominant controls on salinity in Lake Qinghai, and an excess of evaporation over precipitation has produced a brackish/saline, alkaline lake with a salinity of 16.0 g/L and a pH of 9.2 (Liu et al. 2009; Li et al. 2010). According to the bathymetric map created in the 1970s, the lake basin can be divided into two depositional centers (Fig. 1b). Surface water salinity generally increases from the shore to the center of the lake because of the frequent input of fresh water from rivers (Liu et al. 2009; Li et al. 2012). Thermal stratification is observed during the summer with surface and bottom water temperatures of approximately 13–16 and 4–6 °C, respectively (Lister et al. 1991).

The mean surface water $\delta^{18}O$ value from 18 sample sites (water depth > 11.5 m) was 3.5‰ (V-SMOW) in early June 2006 (Liu et al. 2009), much higher than values of local precipitation and river water, and indicates that the lake is hydrologically closed and evaporative (Liu et al. 2009; Henderson et al. 2010). The lake water $\delta^{18}O$ value decreased from the center of the lake to the littoral zone, along with decreasing salinity and water depth (Liu et al. 2009).

Materials and methods

Core 1Fs, which includes cores 1A and 1F (approximately 18.6 m), was drilled in 2005 using the ICDP GLAD800 drilling system. A total of 65 ^{14}C dates were originally measured to develop the core chronology. Among those samples, 52 were total organic carbon (TOC) from bulk sediments, six were Ruppiaceae seeds, and seven were plant fragments (Table 1). Most of the ^{14}C dates were run at the Xi'an AMS Laboratory, and some were measured at the University of Arizona AMS Laboratory. Eight of the AMS ^{14}C samples from 7.36 to 8.38 m had anomalous ages, ca. 17–34 ka older than expected, compared to surrounding samples thought to have reasonable ages (Table 1) (An et al. 2012). These discrepancies may have been caused by the use of improper material for dating. These materials might have been deposited in an older layer and then washed into the lake with runoff during the last deglaciation. These anomalously old ages were not included in the age model (Table 1). Therefore, a total of 57 samples were used to construct the age model (An et al. 2012) (Fig. 2). This chronological framework suggested that core 1F, with a length of 18.6 m, covered approximately 32 ka in Lake Qinghai (Fig. 2). Details of ^{14}C dating were reported by An et al. (2012) and Zhou et al. (2014).

Subsamples for analysis of ostracod species and their isotopic values were sampled at 5-cm intervals. Wet sediment samples of approximately 5–9 g were soaked in deionized water for approximately 24 h and then wet sieved with a 60-μ-mesh sieve. Material larger than 60 μ was dried at room temperature for ostracod collection. Ostracod shells were picked using a fine brush and deionized water, and shells of ostracods were cleaned with deionized water and used for isotope analysis (Li et al. 2007).

Three ostracod species (*L. inopinata*, *E. mareotica* and *I. bradyi*) were used for isotope analysis because no single ostracod species was present throughout core 1F. Only one measurement was conducted per individual sample because of the limited shell material. All ostracod samples, each composed of 4–6 shells of *Eucypris mareotica*, 10–14 shells of *Limnocythere inopinata* and ~2–3 shells of *Ilyocypris bradyi*, were analyzed for δ^{18}O at the Institute of Earth Environment, Chinese Academy of Sciences, using an isotope ratio mass spectrometer (MAT-252) with an automated carbonate preparation device (Kiel II). Results are expressed in delta (δ) notation relative to the V-PDB standard. Repeated analyses of laboratory standard carbonates (TTB1) with known δ^{13}C and δ^{18}O values were performed daily to ensure instrument accuracy. The analytical errors of the laboratory standard were approximately ±0.2‰ for the δ^{18}O values.

Results

Eight ostracod species were identified in core 1F: *L. inopinata, E. mareotica*, *I. bradyi*, *Ilyocypris echinata*, *Candona* sp., *Candoniella* sp., *Cypridopsis* sp. and *Leucocythere* sp. From 32.0 to 15.0 ka, the abundance of ostracods was low. The total number of shells was <50 in each horizon and *I. bradyi* was the dominant species, with minor proportions of other species in some levels (Fig. 3). The abundance of ostracods was extremely low between ca. 32.0 and 20.0 ka and between 18.3 and 17.4 ka (Fig. 3). Ostracods were absent during some intervals, such as from 29.8 to 29.5 ka in core 1F (Fig. 3). The abundance of ostracods increased from 15.0 to 13.7 ka, and *L. inopinata* was the dominant species (Fig. 3). Between 13.7 and 7.4 ka, the ostracod assemblage was dominated by *E. mareotica*. The abundance of ostracods, however, was still low, with the total number of shells <50 for most horizons from 13.7 to 8.8 ka, and most of the ostracod shells were broken in this interval. This period was followed by an increase in the ostracod abundance after 8.5 ka in core 1F (Fig. 3). From 7.3 ka to the present, *L. inopinata* was the dominant species, and this species was the only one present at approximately 3.5 ka in core 1F (Fig. 3).

During the interval 32.0–19.0 ka, ostracod δ^{18}O values ranged from approximately −9.4 to 3.0‰, with a mean value of −3.4‰ (Fig. 4e). From 19.0 to 15.0 ka, the δ^{18}O values varied widely and generally ranged from approximately −11.2 to 5.0‰, with a mean of −3.1‰, but the δ^{18}O values showed much lower values at ~17.9 and ~15.7 ka (Fig. 4e). Thereafter, δ^{18}O values were high, ranging between approximately 0.3 and 3.5‰ from 15.0 to 14.4 ka, (Fig. 4e). The ostracod δ^{18}O values varied between −4.5 and 5.0‰ and had an average value of approximately 1.1‰ from 14.4 ka to the present (Fig. 4e). The ostracod δ^{18}O values initially became higher (from −4.5 to 1.1‰) from 12.0 to 7.4 ka, but reached

Table 1 Age data obtained from the 1F cores from Lake Qinghai (An et al. 2012)

Lab Code	Depth (m)	Material	Measured ^{14}C age (yr BP)	Calibrated age (Cal yr BP) (Median probability)	Lab Code	Depth (m)	Material	Measured ^{14}C age (yr BP)	Calibrated age (Cal yr BP) (median probability)
AA78739	0.01	TOC	30 ± 30	60	XA1751	4.75[a]	Seed	9790 ± 40	11,215
AA78740	0.05	TOC	200 ± 35	180	AA77708	4.75[a]	Seed	9700 ± 60	11,120
XA2718	0.08	TOC	1210 ± 40	1140	AA78019	4.77	TOC	9755 ± 60	11,190
XA1748	0.09	TOC	1430 ± 30	1330	XA2701	4.80	TOC	9900 ± 45	11,290
XA2730	0.21	TOC	1550 ± 20	1470	AA77761	4.81	TOC	9890 ± 50	11,290
XA2772	0.44	TOC	1220 ± 40	1150	XA3195	4.85	TOC	10,290 ± 30	12,070
XA2770	0.62	TOC	1780 ± 40	1700	XA3194	4.90	TOC	10,310 ± 30	12,100
XA2769	0.83	TOC	2000 ± 40	1950	XA1752	4.97	seed	11,020 ± 35	12,940
XA2721	0.86	TOC	2310 ± 30	2340	XA2754	5.14	TOC	11,280 ± 60	13,170
XA2711	1.04	TOC	2180 ± 20	2240	XA3210	5.21	TOC	11,510 ± 30	13,350
XA2768	1.28	TOC	2470 ± 40	2560	XA3226	5.72	TOC	12,630 ± 40	14,910
XA2766	1.71	TOC	3565 ± 40	3865	XA2752	5.77	TOC	12,230 ± 50	14,090
XA2713	1.94	TOC	3740 ± 40	4110	AA77765	6.05	TOC	12,010 ± 60	13,870
XA2765	2.15	TOC	4340 ± 40	4920	XA3187	6.70	Plant residue	14,000 ± 40	16,680
XA2764	2.38	TOC	4380 ± 40	4940	*XA3186*	*7.36*	*Plant residue*	*31,350 ± 110*	*35,260*
XA2763	2.57	TOC	4990 ± 50	5720	*XA2748*	*7.99*	*TOC*	*34,470 ± 200*	*39,810*
XA2700	2.82	TOC	5570 ± 20	6350	*XA3228*	*8.17*	*Plant residue*	*35,170 ± 205*	*40,250*
AA77753	2.84[a]	TOC	5680 ± 40	6465	*XA2747*	*8.18*	*TOC*	*47,905 ± 450*	*51,530*
XA1755	2.84[a]	TOC	5735 ± 40	6530	*XA3185*	*8.23*	*Plant residue*	*31,490 ± 110*	*35,370*
XA2762	3.16	TOC	6270 ± 40	7210	*XA3240*	*8.23*	*Plant residue*	*31,690 ± 105*	*35,550*
XA2759	3.44	TOC	6780 ± 55	7630	*XA3184*	*8.38*	*Plant residue*	*33,620 ± 140*	*38,730*
XA2758	3.68	TOC	7510 ± 40	8335	*XA3966*	*8.38*	*Plant residue*	*32,230 ± 130*	*36,440*
XA2716	3.91	TOC	8450 ± 30	9485	AA78823	9.01	TOC	18,820 ± 130	22,360
XA2757	4.12	TOC	8600 ± 50	9560	XA3229	11.62	TOC	21,680 ± 70	25,880
XA1756	4.13[a]	TOC	8600 ± 40	9550	XA3963	11.89	TOC	21,190 ± 70	25,570
AA77758	4.13[a]	TOC	8560 ± 50	9530	XA2777	13.23	TOC	22,620 ± 100	27,370
AA77711	4.16	seed	8540 ± 50	9520	XA2705	13.63	TOC	23,900 ± 90	28,730
XA2756	4.34	TOC	9430 ± 50	10,660	XA2776	14.25	TOC	23,350 ± 160	28,130
AA77712	4.54	seed	9280 ± 50	10,460	XA2775	14.73	TOC	22,810 ± 100	27,540
XA2755	4.55	TOC	9370 ± 50	10,590	AA77785	15.50	TOC	27,250 ± 335	31,940
AA77760	4.56	TOC	9410 ± 60	10,640	XA2774	15.74	TOC	26,330 ± 150	31,210
XA3200	4.63	TOC	10,235 ± 30	11,990	AA77790	17.73	TOC	28,670 ± 300	33,150
XA1586	4.75[a]	seed	9670 ± 40	11,110					

Italicised samples were omitted in creating the age model

[a] Average age of the same depth was used in creating the age model

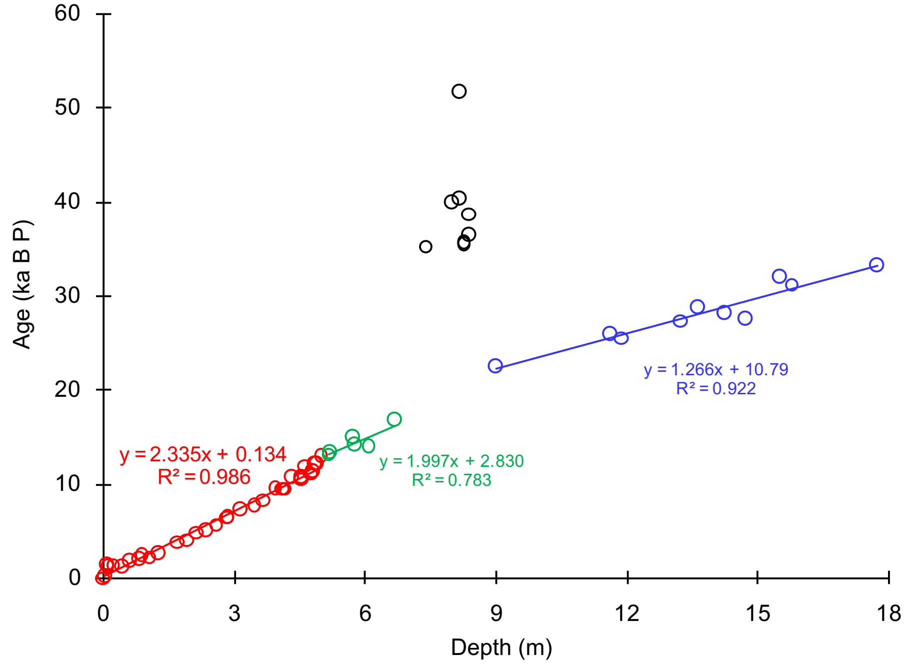

Fig. 2 Age model for core 1Fs using linear regression in segments, after An et al. (2012). The AMS ^{14}C data were divided into three segments, according to their lithology and sedimentation rates

minimum values at approximately 9.3 ka (Fig. 4e). The $\delta^{18}O$ values then showed a sustained increasing trend from 7.4 ka to the present (Fig. 4e).

Discussion

Hydroclimate implications of ostracod assemblages

Ostracod species are sensitive to changes in environmental factors, especially water temperature and salinity (Holmes et al. 1998). Also, the presence of ostracods in lakes is related to water depth, dissolved oxygen content and the nature of the substrate (De Deckker 1981; Benzie 1989; Holmes 2001; Schwalb et al. 2002; Frenzel et al. 2010; Zhai et al. 2010; Yan and Wünnemann 2014). Therefore, changes in ostracod assemblages have the potential to indicate hydroclimatic changes in the lake environment. The ostracods *L. inopinata, E. mareotica* and *I. bradyi* were the dominant species at different times in core 1F. Therefore, a summary of the controlling factors on those species is necessary before discussion of past hydroclimatic variations at Lake Qinghai.

In the Lake Qinghai area, abundant *I. bradyi* were found in rivers, springs, reservoirs and swampy areas with salinities ranging from 0.32 to 2.19 g/L (Li et al. 2010). This poor swimmer was also found in ponds and small lakes on the Qinghai–Tibet Plateau (Kotlia et al. 1998; Zhang et al. 2013). Therefore, presence of *I. bradyi* probably indicates shallow and unstable conditions (or lakeside environments) in Lake Qinghai (Curry 1999; Wu et al. 2001; Li et al. 2010).

The ostracod *E. mareotica* is strongly associated with sodium chloride-containing waters and can tolerate a wide range of salinity, from 0 to 325 g/L (Baltanás et al. 1990; Santamaria et al. 1992; Sun et al. 1993; Li et al. 1997; Wu et al. 2001). Exclusive occurrence of this species may indicate high salinity (Li and Liu 2014).

Limnocythere inopinata can live in lakes with salinities up to 46 g/L, but the maximum abundance of this species was reported for a salinity range of 3–9 g/L (Forester 1983). Maximum abundance of this species in the Lake Qinghai region was recently found in a salinity range from 5 to 15 g/L (Li et al. 2010), and the monospecific assemblage consisting of *L. inopinata* indicated that the salinity was <16 g/L in the lake (Li and Jin 2013; Li and Liu 2014). Therefore, species assemblages dominated by *I. bradyi*, *L. inopinata* and *E. mareotica* generally indicate conditions of temporary lakes or lakes undergoing frequent changes in

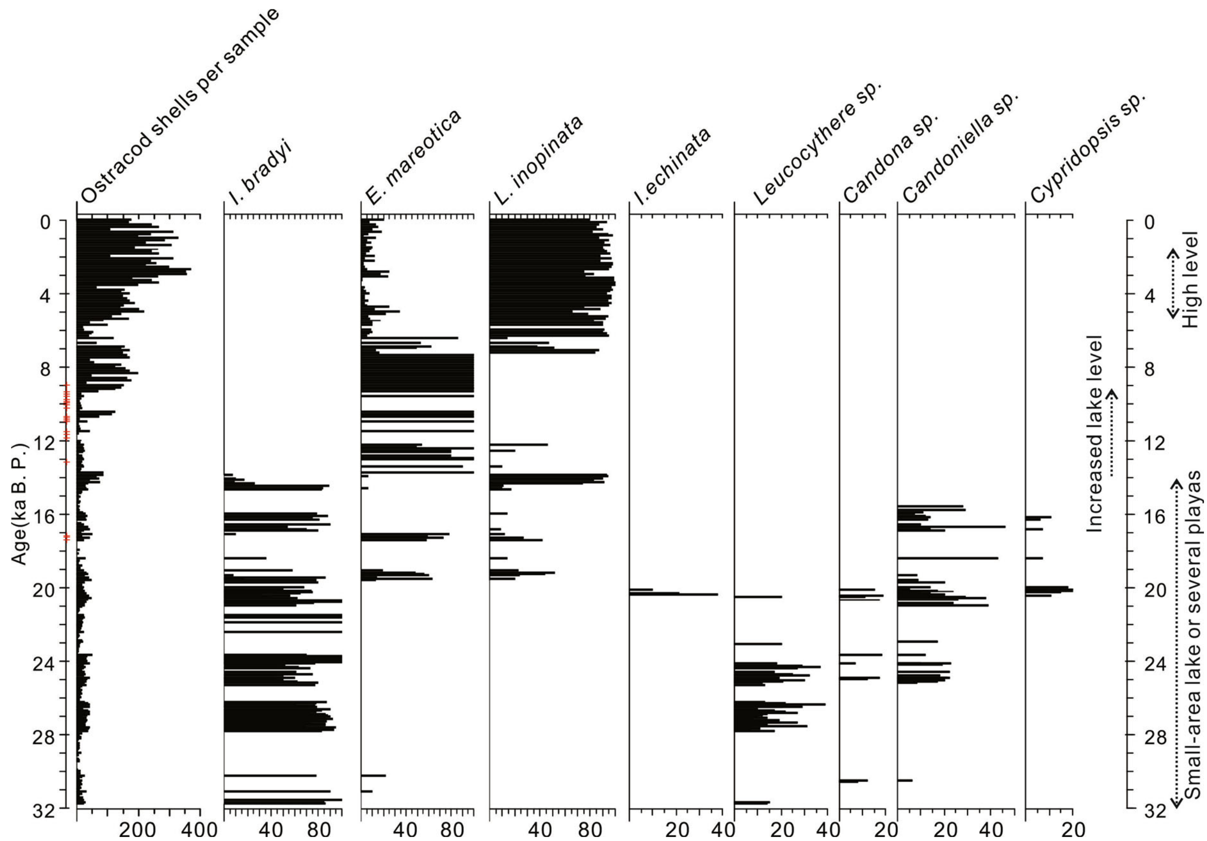

Fig. 3 Abundance of total ostracods per sample (*red plus sign on left age axis* shows periods containing *Ruppia* seeds) and percentages of ostracods in each assemblage in the 32-ka core from site 1F in Lake Qinghai. (Color figure online)

water level, moderate salinity and high salinity, respectively.

Factors that affected $\delta^{18}O$ values in ostracod shells from core 1F during the last glacial

The $\delta^{18}O$ of ostracod shells has been used to investigate changes in the P/E ratio (precipitation/evaporation, also known as effective precipitation) or I/E ratio (input water/evaporation), related to paleoclimate changes near Lake Qinghai (Lister et al. 1991; Henderson et al. 2003; Liu et al. 2007; An et al. 2012). The long-term low mean $\delta^{18}O$ value of ostracod shells in core 1F from 32.0 to 15.0 ka, especially for lowest ostracod $\delta^{18}O$ values from 19.0 to 15.0 ka, however, is probably not the consequence of high effective precipitation or species-specific vital effects (Li and Liu 2010; Von Grafenstein et al. 1999; Decrouy et al. 2011).

The ostracod *Ilyocypris* is found frequently in last glacial deposits of core 1F, as more or less a monospecific assemblage, suggesting that the core site was a wetland setting, similar to wetland conditions near Qinghai Lake today. Such conditions were probably caused by cold and dry climate conditions, supported by an inference for low effective moisture based on pollen data from the Tibetan Plateau (Fig. 4a) (Herzschuh 2006). Additionally, very negative Guliya ice core $\delta^{18}O$ values also suggest very cold climate conditions during that period (Fig. 4b). The low TOC flux indicates low primary productivity in the Lake Qinghai area at that time, caused by the very cold and dry climate (Fig. 4c) (An et al. 2012). Therefore, the relative proportions of stagnant water and runoff might have been an important controlling factor on the $\delta^{18}O$ value of ostracod shells. Although the contribution to lake water from melted snow/ice is very small in modern Lake Qinghai (Jin et al. 2013),

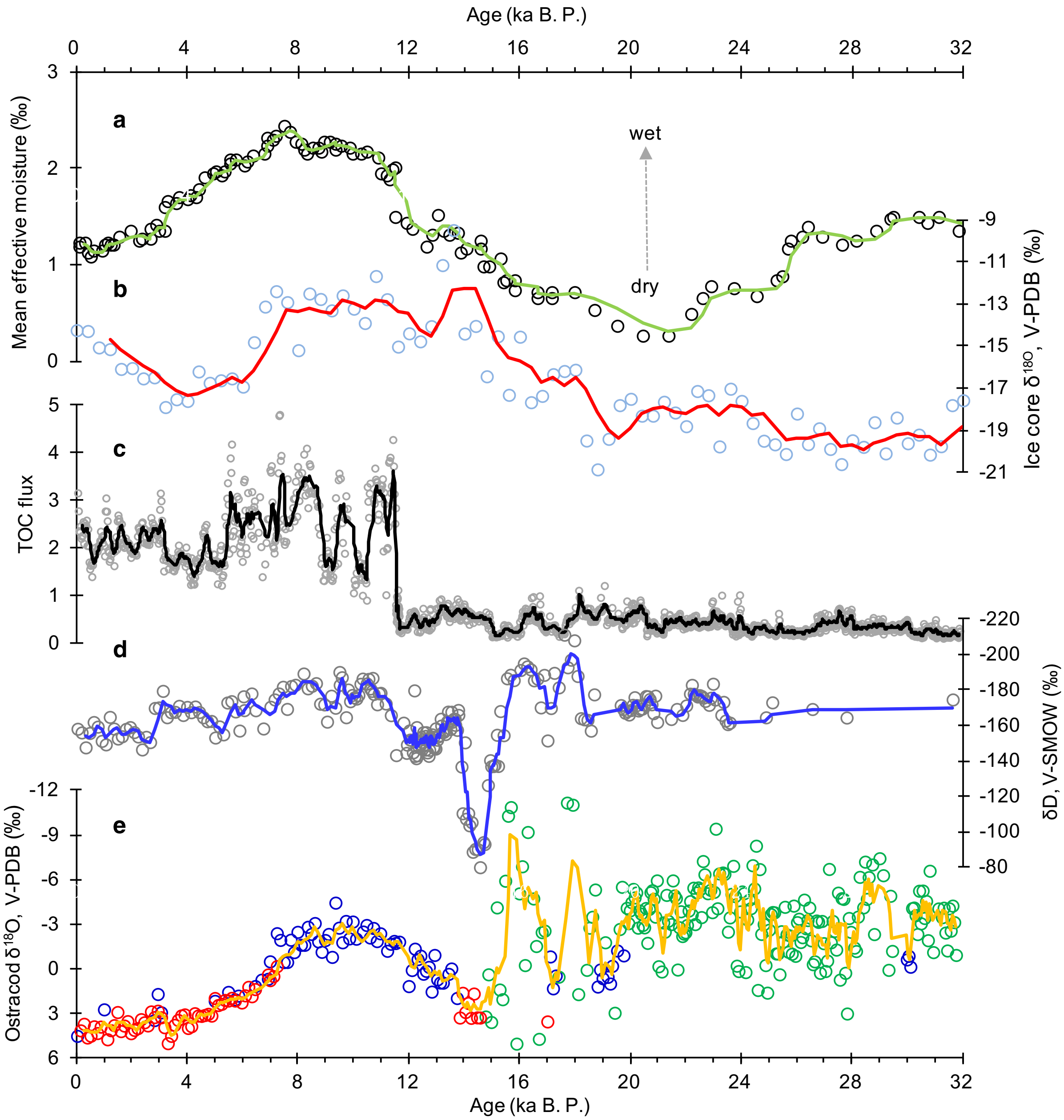

Fig. 4 The ostracod $\delta^{18}O$ record since 32 ka in core 1F compared with other records. **a** Effective moisture from pollen data in central Asia (Herzschuh 2006); **b** Guliya ice core $\delta^{18}O$ values (Thompson et al. 1997); **c** TOC flux in core 1Fs (An et al. 2012); **d** the leaf wax δD values in core 1F (Thomas et al. 2016)

the relative contribution of seasonal glacial meltwater to the lake may have been considerably larger during the last glacial because of the lake's small volume then, compared with the large lake at present.

Modern data suggest more ^{18}O-depleted values in westerly precipitation during winter than in rainfall from Asian monsoon circulation in summer (Tian et al. 2007). Precipitation $\delta^{18}O$ values should have been much lower under climate conditions during the last glacial in the Lake Qinghai region. The high contribution of seasonal meltwater and low $\delta^{18}O$ of precipitation would have caused low $\delta^{18}O$ values of lake water and ostracod shells at that time. Furthermore, reduced evaporation, caused by longer periods of ice cover under cold climate conditions of the last glacial, would have resulted in minimal lake water ^{18}O enrichment, yielding the low ostracod $\delta^{18}O$ values. In addition, past studies suggested that lake water and carbonate $\delta^{18}O$ values were related to distance of the core site from river mouths that provide fresh water. Carbonate $\delta^{18}O$ values from sites near the river mouth were lower because of frequent dilution by input water (Mischke and Wünnemann 2006; Liu et al. 2009).

Ostracod $\delta^{18}O$ values were much lower at times from 19 to 15 ka, and they also displayed large fluctuations (Fig. 4). The core 1F site was closer to the Buha River and the transition area between shallow water and deep water (Fig. 1), compared with the sites of cores QH85-14 and QH 2000, even though core 1F was drilled at a water depth of ~25 m in modern Lake Qinghai (Fig. 1b). The dominant *I. bradyi* indicates that the core 1F site was still in a wetland setting before 15.0 ka (Fig. 3). Dominance of *E. mareotica* in cores QH85-14 and QH2000 (Lister et al. 1991; Liu et al. 2007), however, indicates a stable lake environment at the sites of these two cores compared with the wetland setting inferred for the core 1F site during most of the period from 18.0 to 15.0 ka. The much lower ostracod $\delta^{18}O$ values from 16 to 15.5 ka in core 1F, compared with high ostracod $\delta^{18}O$ values in the other two cores (Fig. 5), suggest that the core 1F site was more easily diluted by river water compared with the sites of cores QH85-14 and QH 2000, before 15.0 ka.

Although effective moisture in monsoonal Central Asia has increased since 19 ka (Fig. 4a), large grain sizes and loess accumulation on the northeastern Tibetan Plateau indicate that westerly winds were very strong and that the climate was still dry ca. 16.0 ka (Wang et al. 2015b). The highly variable $\delta^{18}O$ values of ostracod shells, however, are probably not explained by increased or decreased effective moisture alone, in the interval from 19.0 to 15.0 ka.

The calculated volume of glaciers in the Lake Qinghai catchment was approximately 200 km^3 during the last glacial maximum, which is approximately 2–3 times the size of modern Lake Qinghai (Wang and Shi 1992). Our unpublished data show that $\delta^{18}O$ values of snow or ice (approximately −21.0‰) are much lower than the $\delta^{18}O$ values of precipitation and river water in the Lake Qinghai area. In the summer season, higher temperatures result in greater melting of ice or snow around Lake Qinghai. A recent study suggested that distinct meltwater events occurred at 19.0 and 14.0 ka, as indicated by distinct pulses of anomalously old organic radiocarbon dates in core 2C from Lake Qinghai (Zhou et al. 2016). In addition, the $\delta^{18}O$ values of input water may have become lower because of low $\delta^{18}O$ values in incoming precipitation, under generally cold conditions or melted ice and snow with low $\delta^{18}O$ values. So, snow/ice and incoming precipitation with low $\delta^{18}O$ values, together with high/frequent runoff to the lake relative to the lake water volume, probably accounts for the much lower ostracod $\delta^{18}O$ values at that time (Fig. 4).

High ostracod $\delta^{18}O$ values (especially for $\delta^{18}O$ values of *I. bradyi*), however, might be a consequence of high host water $\delta^{18}O$, caused by less contribution of seasonal melt water runoff and strong evaporation in the summer season. In addition, the index of water depth (%thaum) shows that lake level had risen before 15.0 ka (Wang et al. 2015a), so the relatively high contribution of stagnant water with high $\delta^{18}O$ probably also led to high ostracod $\delta^{18}O$ values (especially for *E. mareotica* and *L. inopinata*), with increased lake level in some intervals from 19 to 15 ka. In other words, high $\delta^{18}O$ values of *E. mareotica* and *L. inopinata* probably indicate that an ephemeral stable lake was present at core 1F site from 17.4 to 17.0 ka (Fig. 4).

In general, the low ostracod $\delta^{18}O$ values in the Lake Qinghai area during the last glacial might be explained by low abundance of stagnant water, a relatively high contribution of melted ice or snow, close proximity of the core site to the river mouth, and low $\delta^{18}O$ values of precipitation related to cold climatic conditions or/and the northwesterly moisture source. High $\delta^{18}O$ values of ostracods might be explained by a smaller contribution of seasonal melt water runoff, but strong evaporation in the summer season, and/or an increased contribution of stagnant water enriched in ^{18}O, with occasionally higher lake level.

Late glacial and Holocene environmental changes in Lake Qinghai

Ostracod species assemblage and oxygen isotope data from core 1F reveal the details of hydroclimate changes at Lake Qinghai since the last glacial (Figs. 3, 4). From 32.0 to 15.0 ka, *I. bradyi* shells of different

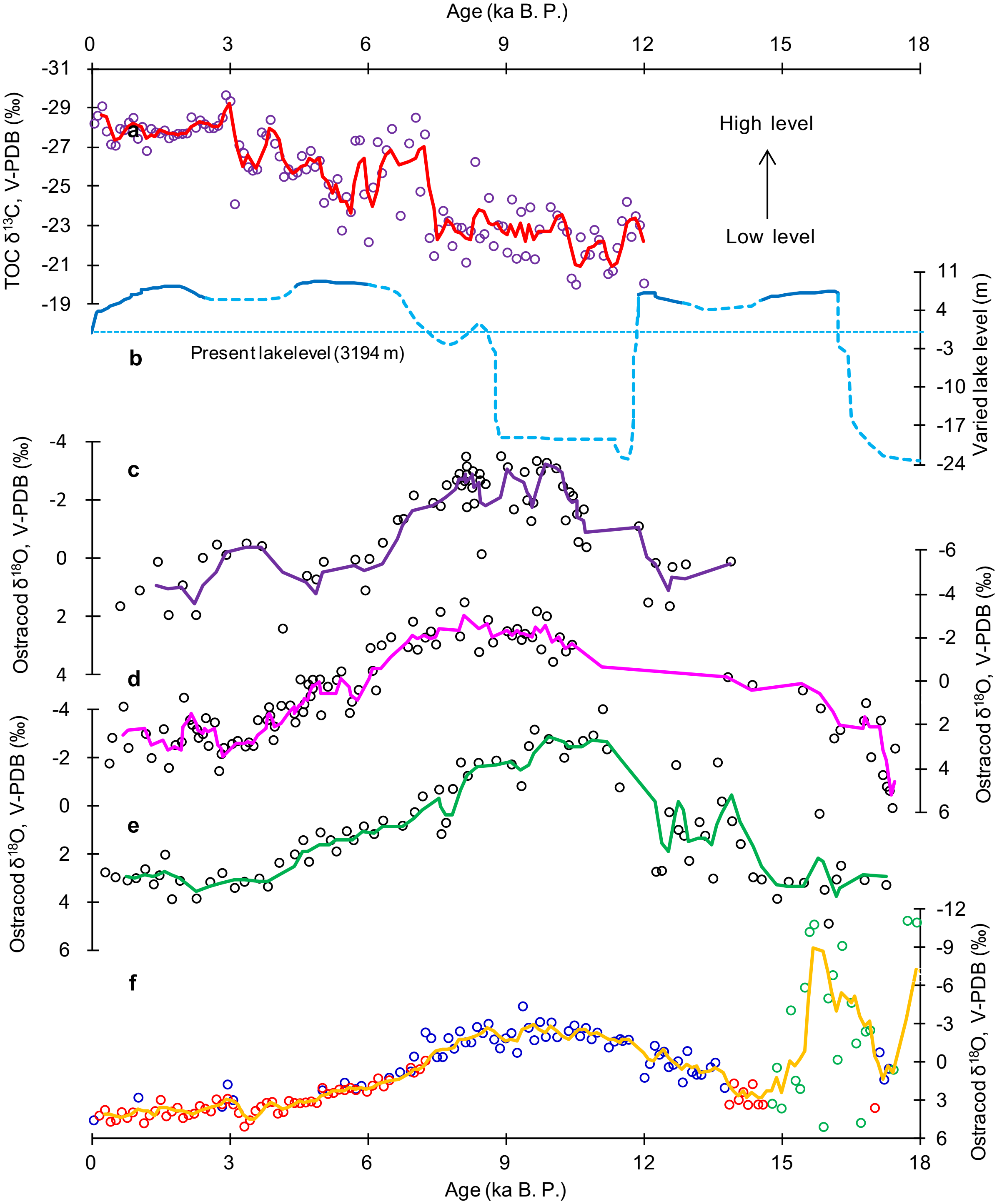

Fig. 5 The $\delta^{18}O$ variation of ostracod shells in core 1F from Lake Qinghai compared with other records since 18.0 ka. **a** Lake level variation indicated by the $\delta^{13}C$ values of total organic matter from Lake Qinghai (Liu et al. 2013); **b** reconstructed shoreline history for Lake Qinghai, plotted against age estimates for shoreline elevations from Liu et al. (2015b); **c** ostracod $\delta^{18}O$ record in core QH85-16A from Zhang et al. (1989a, b); **d** ostracod $\delta^{18}O$ record in core QH85-14B from Lister et al. (1991); **e** ostracod $\delta^{18}O$ record in core QH-2000 from Liu et al. (2007)

sizes were observed in most layers of core 1F, as a nearly monospecific assemblage, suggesting that Lake Qinghai was very shallow or consisted of several playas connected by streams or rivers most of the time (Whatley 1983, 1988; Danielopol et al. 1986; Zhai et al. 2015), and the core site was a wetland setting for some time during the last glacial, similar to some areas near Lake Qinghai today. Presence of *Leucocythere* sp. during the interval 28.0–24.5 ka (Fig. 3,), as well as the presence of *I. echinata* and *E. mareotica* from 21.0 to 19 ka (Fig. 3,), suggest more water and increased lake level during these two intervals than before (Yang et al. 2002; Mischke et al. 2003). Ostracod $\delta^{18}O$ values displayed large fluctuations from −11.1 to 5.0‰, with a long-term mean ostracod $\delta^{18}O$ value of nearly −3.3‰ at that time. Ostracod $\delta^{18}O$ values showed large fluctuations during the interval from 19.0 to 15.0 ka (Fig. 4), suggesting that the hydrological conditions at the core 1F site were very unstable during that time interval.

From 15.0 to 13.7 ka, increased abundance of ostracod shells, the presence of *L. inopinata* and decreased $\delta^{18}O$ values indicate the lake existed at the core site and water was deeper. The increased lake level probably resulted from increased moisture availability during the Bølling/Allerød warm phase. In addition, stratigraphic changes in ostracod $\delta^{18}O$ values from core 1F were strikingly similar to patterns of ostracod $\delta^{18}O$ values in cores from other sites in Lake Qinghai after 14.0 ka (Fig. 5) (Zhang et al. 1989a; Lister et al. 1991; Liu et al. 2007), which suggests that the playas probably merged after 14.0 ka.

From 13.7 to 7.4 ka, *E. mareotica* was almost the only ostracod species in core 1F (Fig. 3). Ostracod $\delta^{18}O$ values first showed a decreasing trend after 13.7 ka, and then increased after 9.3 ka (Fig. 4). Variation in ostracod $\delta^{18}O$ values suggests that lake level probably increased after 13.7 ka. Dominance of *E. mareotica* indicates that the water salinity was probably high during this period (Fig. 3), which is indicative of low lake level and supported by the presence of *Ruppia* seeds (Fig. 3). Ostracods were absent from 12.0 to 11.6 ka, which suggests that the environment or climate was not favorable for ostracod assemblages. The low TOC and pollen concentrations also suggest cold and dry climate conditions in the Lake Qinghai area (Shang 2009; An et al. 2012), which coincide with the Younger Dryas (YD), within dating uncertainty. In general, the dominance of *E. mareotica* and its decreased $\delta^{18}O$ values suggest that lake level increased from 13.7 to 7.4 ka, except for a period of reduced lake depth from 12.0 to 11.6 ka, during which the surface area of Lake Qinghai was smaller than the modern lake area.

From 7.4 to 3.2 ka, *E. mareotica* sharply decreased in relative abundance, and *L. inopinata* became the dominant species (Fig. 3). The dominance of *L. inopinata* (Fig. 3), which thrives at lower salinities, indicates that salinity decreased, related to the higher lake level between 7.4 and 3.2 ka. Greater numbers of *E. mareotica* at 6.7 and 5.8 ka, however, may indicate that salinity increased sharply and lake level declined during these two periods (Fig. 3). Higher ostracod $\delta^{18}O$ values between 7.4 and 3.2 ka, however, suggest a decrease in effective precipitation in that interval of higher lake level (Fig. 4). Increased water level in Lake Qinghai may have been caused by sustained high effective moisture, with an increasing trend from 7.4 to 3.2 ka.

From 3.2 ka to the present, *L. inopinata* was the dominant species; however, *E. mareotica* has occurred at low abundance from 3.2 ka to the present. Temporal variation in the ostracod assemblage indicates that water salinity increased again under high lake-level conditions after ca. 3.3 ka (Fig. 3). Relatively constant ostracod $\delta^{18}O$ values, with a weak increasing trend, show that effective precipitation did not change appreciably after 3.2 ka (Fig. 4).

Overall, lake state and hydroclimate changes at Lake Qinghai since the last glacial can be summarized as follows: (1) a small shallow lake or even a wetland existed for some time under cold and dry climate conditions from 32.0 to 14.0 ka; (2) increasing lake level after 15.0 ka, but a sudden drop in lake level from 12.0 to 11.6 ka; (3) rising but still shallow lake level, with increased moisture availability from 11.6 to 7.4 ka; (4) increasing lake level from 7.4 to 3.2 ka, and (5) a stable, large, deep lake with brackish water after 3.2 ka.

Comparison with other records from Lake Qinghai

Changes in dominant moisture sources have been studied using δD values of the long-chain *n*-C_{28} fatty acid from Lake Qinghai during the past 32.0 ka, which is assumed to come mainly from terrestrial plants (Thomas et al. 2016). Results of this study suggest that local and northwesterly air masses are important moisture sources to the northeastern Tibetan Plateau, and should be considered when reconstructing past hydroclimate in this region (Thomas et al. 2016). This conclusion, however, is tentative because long-chain *n*-C_{28} fatty acid may also come from aquatic plants, according to long-chain *n*-alkane $\delta^{13}C$ and δD data from Lake Qinghai (Liu et al. 2015a, 2016).

As shown in Fig. 4, both leaf wax δD and ostracod $\delta^{18}O$ values in core 1F show trends similar to those for isotopic composition of the source water, based on ice core $\delta^{18}O$ values before 14.0 ka (Fig. 4). They do not, however, agree with changes in effective moisture inferred from pollen data in this region (Fig. 4). The combined results suggest that both leaf wax δD and ostracod $\delta^{18}O$ values were more sensitive to changes in the isotopic composition of source water than to the variation in effective moisture during the last glacial. A recent review paper also suggests that ostracod $\delta^{18}O$ accurately reflects $\delta^{18}O$ of past precipitation, related to moisture source, but not the intensity of monsoon precipitation during the Holocene in the Lake Qinghai area (Chen et al. 2016).

After 14.0 ka, patterns of ostracod $\delta^{18}O$ and leaf wax δD values were still similar in core 1F, suggesting that the water sources related to the ostracod $\delta^{18}O$ and leaf wax δD were the same. Thomas et al. (2016) interpreted the low δD values of long-chain C_{28} fatty acid as reflecting strong summer monsoon precipitation (Fig. 4), so a high lake level was suggested in the early Holocene (Thomas et al. 2016), and it seems that the decreased/low ostracod $\delta^{18}O$ values in all cores from different sites in Lake Qinghai, also suggest increasing effective precipitation in the region during this period (Fig. 5). The high lake level inferred from leaf wax δD, however, contradicts the low-lake-level condition inferred from high $\delta^{13}C$ values of TOC and long-chain, C_{27} and C_{29} *n*-alkanes (Liu et al. 2013, 2015a), low %thaum (Wang et al. 2015a) and geomorphic shoreline investigations in Lake Qinghai (Liu et al. 2015b) (Fig. 5b). The results of geomorphic shoreline investigations showed that water level in Qinghai Lake was relatively low from 11.0 to 8.0 ka (Fig. 5b), with water depths of only a few meters and frequent and dramatic fluctuations (Liu et al. 2015b). Monospecific presence of *E. mareotica* indicates that water salinity was still high during the early Holocene (Fig. 3). Ostracod assemblages and their $\delta^{18}O$ values indicate that Lakes Koucha and Donggi Cona also exhibited brackish conditions caused by low moisture availability and a dry climate on the northeastern Tibetan Plateau during the early Holocene (Mischke et al. 2008, 2010). In addition, the strong aeolian activity and rapid aeolian sand accumulation also indicate dry climate in the Lake Qinghai area and neighboring regions between about 11.2 and 7.5 ka (Lu et al. 2011, 2015; Qiang et al. 2013). Recent studies in lakes on the northeastern Tibetan Plateau found that the carbon isotope value of sediment TOC and biomarkers extracted from sediments reflect mainly organic carbon contributed by aquatic plants (Aichner et al. 2010; Liu et al. 2013, 2015a, 2016). It seems that only a large contribution of aquatic plants to the long chain fatty acid (C_{28}) pool in core 1F could explain the similar trends in ostracod $\delta^{18}O$ and leaf wax δD values during the early Holocene in Lake Qinghai (Fig. 5).

Monospecific presence of the ostracod *E. mareotica* in other cores (QH85-14B and QH2000) suggests that salinity of the water was high before 14.0 ka (Lister et al. 1991; Liu et al. 2007). Thus, Lake Qinghai was probably a small, shallow lake or consisted of several playas surrounded by a salt crust, as is the case for many saline lakes of the Tibetan Plateau today. The Asian monsoon (AM) intensified (Dykoski et al. 2005), which may have resulted in greater precipitation and more river-supplied water in the early Holocene (An et al. 2012), and temperature was also high, as indicated by the $\delta^{18}O$ data from the Guliya ice core during the early Holocene on the Qinghai–Tibet Plateau (Fig. 4) (Thompson et al. 1997). Initial filling of the basin to early Holocene levels may have caused waters at the 1F core site to become saline as flooding of the basin dissolved the salt crust or contributed high-salinity lake water from other areas. Today, brackish, saline and playa lakes in the Qaidam Basin are surrounded mostly by efflorescent salt crusts. A small rise in water level would increase their salinity, despite an increase in runoff. In the early Holocene, lake water $\delta^{18}O$ values probably became lower and lake level rose as a consequence of meltwater runoff or increased river water entering the lake. This would have accounted for low $\delta^{18}O$ values in the brackish to saline water species *E. mareotica* as early as 14.0 ka. In addition, the low $\delta^{18}O$ values of ostracod shells also resulted from high temperatures during the early Holocene when the surface area of Lake Qinghai was still small (Thompson et al. 1997). Similarly, very low leaf wax δD values also can be explained by the above scenario because leaf wax was probably contributed by aquatic plants in Lake Qinghai during the early Holocene, as discussed above.

High and increasing ostracod $\delta^{18}O$ values after 7.4 ka suggest a decrease in effective precipitation at the same time that decreased abundance of *E. mareotica* and dominance of *L. inopinata* indicate that the water salinity decreased (Fig. 5). The lower $\delta^{13}C$ values of total organic carbon also indicate rising lake level after 7.4 ka (Liu et al. 2013). Lake level rise may have been caused by sustained high effective moisture, as indicated by the pollen data (Fig. 4b). Therefore, the high $\delta^{18}O$ values of ostracod shells may have been caused by the long residence time of water (Lister et al. 1991), and an increase in the proportion of resident lake water with evaporation-driven high $\delta^{18}O$ values, to input water with low $\delta^{18}O$ values, with increased lake level after 7.4 ka.

The difference between trends in ostracod $\delta^{18}O$ values and leaf wax δD values became larger after 7.4 ka (Fig. 4d, e), and the largest offset occurred in the period 4.0–3.2 ka (Fig. 4d, e). Well-developed paleosols suggest a warm and humid climate during the middle Holocene in the Lake Qinghai area (Lu et al. 2015), so the productivity of terrestrial plants would have increased in this period. Liu et al. (2016) found that δD values of long-chain *n*-alkanes of aquatic plants were generally higher than those from surrounding terrestrial plants in Lake Qinghai, and high δD values of long-chain *n*-alkanes of aquatic plants might be related to evaporation-driven, high lake water δD values (Liu et al. 2016). So the increased offset between ostracod $\delta^{18}O$ and δD values of long chain C_{28} fatty acid might be caused by a relative increase in the contribution of terrestrial plants to sediments after 7.4 ka in Lake Qinghai. In addition, ostracod $\delta^{18}O$ is known to be controlled mainly by the host water $\delta^{18}O$ and the water temperature at the time the ostracod shells formed, but the hydrogen isotopic fractionation between aquatic plants and the source water may be weakly affected by water temperature (Li et al. 2015). The ostracod $\delta^{18}O$ values may have increased because of lower water temperatures that were a consequence of greater water depth and lower air temperatures after 7.4 ka, so the offset between the ostracod $\delta^{18}O$ and leaf wax δD values became larger after that time. The greater offset and the maximum difference between ostracod $\delta^{18}O$ and leaf wax δD values may indirectly demonstrate that water level in Lake Qinghai increased considerably after 7.5 ka and reached its highest level in the interval from 4.0 to 3.2 ka. In addition, a recent study suggested that increased water depth was associated with a decrease in TOC $\delta^{13}C$ values, which reflected a shift in the types of aquatic plants in Lake Qinghai (Liu et al. 2013). Liu et al. (2013) suggested that Lake Qinghai reached its maximum level at 3.0 ka, indicated by the lowest TOC $\delta^{13}C$ value (Fig. 5a). Geomorphic shoreline investigations, however, suggested that high lake levels occurred at 5.0 and 2.0 ka (Fig. 5b) (Liu et al. 2015b), at odds with the timing shown by our results. Our age model was constructed using radiocarbon ages, whereas Liu et al. (2015b) used OSL dating. Discrepancies in the timing of high stands may result from the different age models used in the two studies.

Conclusions

A 32-ka-long record of ostracod species assemblages and their isotopic compositions, and several previously published Holocene ostracod $\delta^{18}O$ records, were used to infer lake evolution and corresponding hydroclimate changes since the last glacial period at Lake Qinghai. Dominant ostracod *I. bradyi* indicated that the lake may have featured several playas, connected by streams or rivers, during a period of colder and drier climate throughout most of the last glacial. Low ostracod $\delta^{18}O$ values during the last glacial period, compared with higher $\delta^{18}O$ values during the Holocene, may have been caused by a low ratio of stagnant lake water to runoff, low temperatures and low $\delta^{18}O$ values of incoming moisture, which was stored as snow and ice. Contributions of meltwater (ice or snow) with significantly lower $\delta^{18}O$ values probably caused lower ostracod $\delta^{18}O$ values during interstadials and the entire LGM. Presence of *L. inopinata* and absence of *I. bradyi* showed that the area of Lake Qinghai increased after 15.0 ka. The sharp decrease in abundance and even absence of ostracod shells indicates that the climate was dry and cold in the Lake Qinghai area from 12.0 to 11.6 ka, a time that coincides roughly with the Younger Dryas. After 11.6 ka, the climate and lake conditions at Qinghai can be summarized as follows: (1) increased, but still small lake area (compared with the modern lake area), caused by increased effective moisture.

Nevertheless, climate was still dry and temperatures were high during the period 11.6–7.4 ka, (2) increasing lake area and decreasing water salinity occurred under a warm and wet climate during the period 7.4–3.2 ka, and (3) a large and relatively stable brackish lake formed after 3.2 ka.

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