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RESEARCH ARTICLE

Lacustrine record of 800 yr hydrological variations on the central Tibetan Plateau

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Abstract Zige Tangco is a meromictic saline lake located on the central Tibetan Plateau. Two parallel cores (ZGTC A-1 and ZGTC A-2) were collected from the lake at a water depth of 25 m during summer 2006. The chronology of core A-1 was reconstructed based on the Constant Initial Concentration (CIC) model of ^{210}Pb and three accelerator mass spectrometry (AMS) ages from the chitin fragments. The hard water effect calibration of the sediment ^{14}C age showed that the reservoir effect ranged from 1655 yr at 1950 AD to 1540 yr at 1610 AD. The hydrological variation in Zige Tangco during the past 800 yr was reconstructed using multi-proxies, including organic and carbonate content, stable isotopes of fine-grained carbonate minerals (< 38.5 μm) and grain-size distribution of the lake sediments. Our results show that there were strong fluctuations in the lake level between 1200 and 1820 AD, and at least three dry periods were recorded between 1235 and 1315 AD, 1410 and 1580 AD, and 1660 and 1720 AD characterized by high carbonate content, abrupt positive shifts of stable isotopes, and high sand content. The low-lake-level periods during the Little Ice Age (LIA) in Zige Tangco correspond to the lower δ^{18}O values in the Guliya ice core and the lower precipitation reconstructed from tree rings in Delingha. This demonstrated that the summer monsoon on the central Tibetan Plateau weakened during the dry and cold periods, whereas the winter monsoon strengthened. Relatively wetter periods or higher lake levels in Zige Tangco occurred at 1580–1650 AD and 1820–1900 AD. Negative shifts in stable isotopes were related to increased lake levels between 1800 and 1820 AD. Our results also showed that the summer monsoon precipitation on the central Tibetan Plateau was mainly controlled by solar activity during the past 800 yr.

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1 Introduction

Numerous studies on changes in the climate, vegetation, and landscape of the Tibetan Plateau during the Holocene have been published in recent decades (e.g., Chen et al., 2020). The Medieval Warm Period (MWP) and Little Ice Age (LIA) were two key climatic fluctuations in the past millennium (Xu and Yi, 2014). Environmental changes in the Tibetan Plateau over the past 1000 yr were studied through various archives, including tree rings (Sheppard et al., 2004; Shao et al., 2005), ice cores (Yao et al., 1997; Hou et al., 2018a), and lake and peat sediments (Zhang et al., 2003; Hou et al., 2008, 2012, 2016, 2017a, 2017b, 2018b, 2019; Zhang et al., 2014; Jin et al., 2016; Li et al., 2017; Pu et al., 2020; Chen et al., 2022; Pu and Meyers, 2022; Yuan et al., 2022). Early studies show that cool winters during the past millennium occurred during 1470–1520 AD, 1620–1720 AD, and 1840–1890 AD in China (Zhu, 1973). The δ^{18}O value from ice cores of the Tibetan Plateau show three cold periods during the LIA (Yao et al., 1997). The LIA on the Tibetan Plateau occurred between 1350 and 1900 AD based on multiple paleoclimate proxies (Yang et al., 2002). Asian monsoon history recorded from precisely dated stalagmites from Wanxiang Cave indicated that the LIA began at 1340–1360 AD and ended between 1850 and 1880 AD (Zhang et al., 2008). In addition, ice accumulation and precipitation reconstructed from tree rings suggest that the LIA is not simply a period of prolonged drought (Yao et al., 1996; Thompson et al., 1997, 2000; Shao et al., 2005; Pu et al., 2020; Pu and Meyers, 2022). More work is needed to understand the nature and temporal-spatial variability of environmental changes, (such as hydrological variation, humidity, and vegetation) during the LIA on the Tibetan Plateau using different archives (Holmes et al., 2007; Xu and Yi, 2014; Li et al., 2022).

Nevertheless, climate variations during the past millennia are poorly illustrated owing to low sampling resolution or the loss of the upper part of sediment sequences since most previous studies on lacustrine sediments cover post-glacial and Holocene periods (Gasse et al., 1991; Lister et al., 1991; Fontes et al., 1993, 1996; Gu et al., 1993; Shen et al., 2005; Herzschuh et al., 2006; Morrill et al., 2006; Wu et al., 2006; Mischke et al., 2008; Hou et al., 2012, 2016, 2017a, 2017b, 2018b, 2019; Wang et al., 2017, 2018). Few studies have focused on the past millennium on the Tibetan Plateau, such as those on the Qinghai Lake (Zhang et al., 2003, 2004; Henderson et al., 2003), Sugan Lake (Qiang et al., 2005; Holmes et al., 2007), Nam Co (Wrozyna et al., 2010), Chen Co (Zhu et al., 2001), Lake Ngoring (Pu et al., 2020), and Ximencuo (Pu and Meyers 2022). In this study, we used stable isotopes of fine carbonates (< 38.5 μm) and other independent proxies to reconstruct paleohydrological changes in the Zige Tangco region since 1200 AD.

2 Study area

Zige Tangco is located in the central Tibetan Plateau (Fig. 1). It is a meromictic alkaline/saline lake with a pH of 10 and salinity of 41 g/L (Li et al., 1998; Li, 2001; Li et al., 2001a) and the major ions in the lake water are Na^+ and CO_3^{2-}– (HCO_3^-). It is one of the 18 lakes that evolved along the Banggong-Gerze-Nujiang Rift (Guan et al., 1984). The closed hydrological system drains a watershed of 3430 km^2 with gentle slopes. It is mainly supplied by seven rivers from all directions (Fig. 1). The lake area is 187.0 km^2 and the maximum water depth is 38.9 m (measured in 1999). Meteorological records from nearby stations at Naqu, Anduo, and Bange Counties indicate that the mean annual air temperature varies between −3.4°C and −0.4°C and the annual precipitation ranges from 240 to 500 mm (Li et al., 2001a). Up to 80% of rainfall occurs between June and September. Annual evaporation varies between 800 and 1110 mm (Li et al., 2001b).

Two parallel short cores [ZGTC A-1 (96 cm) and ZGTC A-2 (80.5 cm)] were taken from Zige Tangco at a water depth of 25 m in September 2006 (Fig. 1) using a gravity corer and sampled at an interval of 0.5 cm for the top 20 cm and 1 cm below 20 cm. Wet samples were stored at 4°C until they were analyzed. The dating results for ^{137}Cs and ^{210}Pb in core A-2 were published by Yao et al. (2008). A proxy analysis of the upper 60.75 cm of core A-1 is presented in this study.

3 Material and methods

Water content analysis (WC) was determined by drying wet samples at 100°C for 5 h. The difference between the wet and dry samples was expressed as a percentage of the wet sample weight. Organic matter content of the sediments was measured by determining the loss on

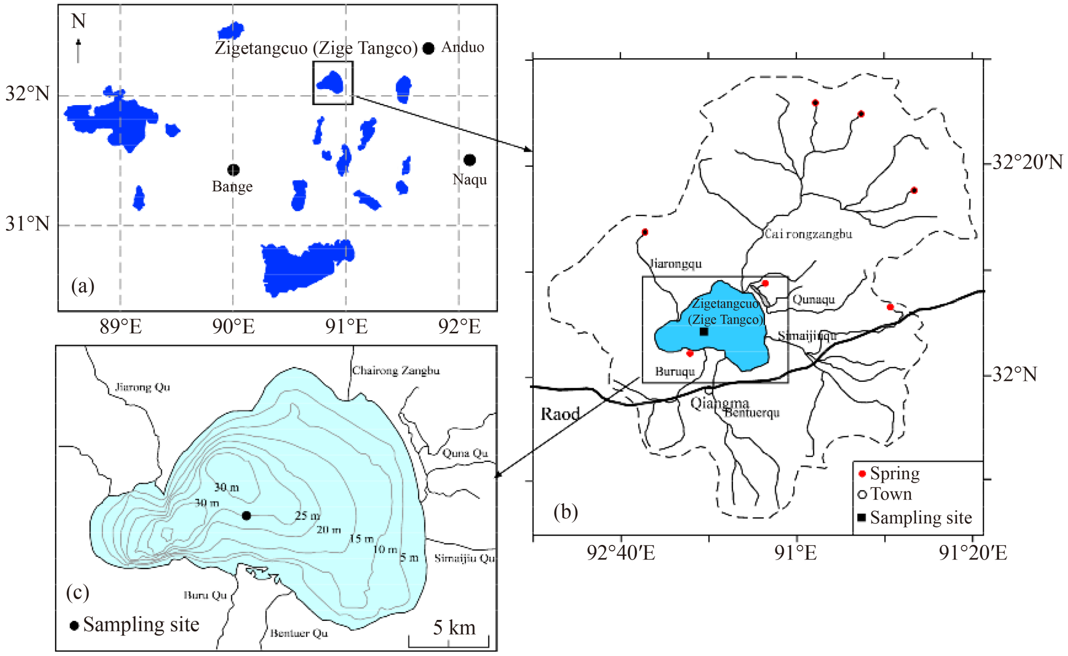

Fig. 1 The general description of the study area. (a) The location of Zige Tangco on the central Xizang Plateau. (b) The general description of Zige Tangco and its catchment. (c) The isobath of Zige Tangco (Li, 2001) and the core site in summer 2006.

ignition (LOI) at 550°C for 4 h, considering that only an undefined part that represents organic carbon content and certain mineral components such as pyrite can influence the LOI values (Zhang et al., 2003). Carbonate content was determined by heating the ashed sample to 950°C for 4 h after LOI estimation. The difference between the ash weight and the weight lost at 950°C is multiplied by a factor of 1.36 (the difference between the molecular weights of $CaCO_3$ and CO_2) and is expressed as a percentage of the dry weight.

The grain size was analyzed using a Malvern Mastersizer 2000 laser analyzer at the State Key Laboratory of Lake Science and Environment, Institute of Geography and Limnology, CAS. This machine possesses a measurement range between 0.02 and 2000 μm with a repeated error of below 2%. Carbonates and organic matter in the samples were removed using HCl and H_2O_2 during the pre-treatment process (Peng et al., 2005).

X-ray diffraction (XRD) analysis of the surface lake sediments was performed at the Modern Analysis Center, Nanjing University. Most of the carbonates were authigenic calcite and aragonite. Fine-grained carbonates were selected for stable isotopes to minimize the influence of carbonate particles from ostracods and molluscs (Qiang et al., 2006). The fine-grained sediment (< 38.5 μm) was washed in distilled water to remove interstitial pore water, oven-dried at 40°C, and ground to a fine powder using a mortar and pestle (Herczeg et al., 2003). The stable oxygen and carbon isotopes of the fine-grained carbonates from core A-1 were analyzed using a Gasbench II mass spectrometer at the State Key Laboratory of Lake Science and Environment, Institute of Geography and Limnology, CAS. The results are expressed in delta notation relative to the Vienna Peedee Belemnite (VPDB) standard, with a mean deviation of less than 0.2‰.

4 Results

4.1 Core A-1 chronology

Unsupported ^{210}Pb activity in the upper 10 cm of core A-2 displayed an exponential form (Fig. 2), and the average sedimentation rate was 0.0806 cm/yr since 1890 based on the CIC model (Yao et al., 2008). The CRS age beyond the 1920s deviates from the CIC age, and is regarded to be older than the 'true' age (Fig. 2). ^{137}Cs peaks at 3.25 cm depth in core A-2 and corresponds to the maximum fallout in 1962−1964. The sedimentation rate was 0.076 cm/yr using this peak as a time marker for 1963, and this result was consistent with that of ^{210}Pb (Table 1).

Three chitin samples were collected from different depths in core A-1 for AMS ^{14}C dating at Beijing University. Two ages at 2.25 cm and 8.25 cm depth (1295 ± 65 and 1695 ± 35 ^{14}C a B.P.) correspond to 1980 AD and 1910 AD, respectively according to ^{210}Pb age (Table 2). The difference between ^{14}C age at a certain depth and its known-age before 1950 AD can be considered as the reservoir effect considering the bomb effect on ^{14}C dating (Wu et al., 2007). The difference between the AMS and ^{210}Pb age at 8.25 cm points to a reservoir effect of approximately 1655 yr. The effect at 2.25 cm depth appears to be 330 yr younger than that at 8.25 cm owing to the influence of bomb experiments.

The hard water effect calibration of the sediment ^{14}C age of core A-1 was carried out using the relationship between ^{14}C levels in lake water and the atmosphere considering the different reservoir effects at different times. This method can eliminate the effect of atmospheric ^{14}C variation on the dissolved inorganic carbon (DIC) ^{14}C concentration in lake water (Wang et al., 2007). The hard water effect at 8.25 cm and 30.75 cm depths ranges between 1540 and 1655 yr during the past

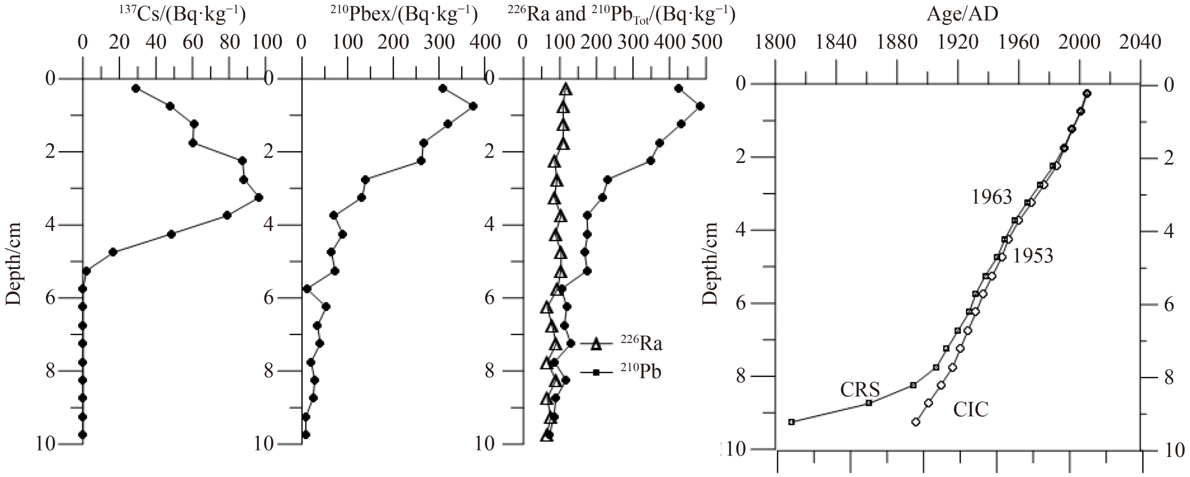

Fig. 2 The left: the depth profiles of ^{137}Cs, ^{210}Pb, and ^{226}Ra in core A-2. The right: age–depth profile of the upper 10 cm based on CRS and CIC model (according to Yao et al., 2008).

Table 1 ^{137}Cs, ^{226}Ra, and ^{210}Pb results for core ZGTC-A at Zige Tangco (Yao et al., 2008)

Depth/cm	Mass depth/(g·cm^{-2})	^{137}Cs/(Bq·kg^{-1})	Unsupported^{210}Pb/(Bq·kg^{-1})	^{226}Ra/(Bq·kg^{-1})	^{210}Pbt/(Bq·kg^{-1})
0.25	0.0871	29.12	308.64	115.35	424.00
0.75	0.233	47.88	376.17	107.88	484.05
1.25	0.408	60.64	320.43	110.16	430.59
1.75	0.520	60.23	265.62	107.58	373.20
2.25	0.725	87.39	262.01	85.72	347.73
2.75	0.980	88.39	139.72	91.72	231.44
3.25	1.214	96.45	131.50	84.80	216.31
3.75	1.479	78.88	70.73	103.11	173.84
4.25	1.612	48.19	89.34	87.00	176.34
4.75	1.740	16.86	64.10	103.55	167.64
5.25	1.962	1.96	72.27	102.42	174.70
5.75	2.093		10.47	93.64	104.11
6.25	2.238		53.35	65.20	118.55
6.75	2.407		33.76	78.47	112.24
7.25	2.532		39.98	90.09	130.07
7.75	2.696		19.54	65.86	85.40
8.25	2.992		28.07	88.79	116.85
8.75	3.184		23.77	65.04	88.81
9.25	3.443		8.40	75.14	83.54
9.75	3.628		7.70	63.57	71.27

Table 2 ^{14}C dating results of Zige Tangco

Sample number	Depth/cm	^{14}C age/(a B.P)	Material	Hard water effect/yr	Sample age/AD	Sedimentation rate/(cm·yr^{-1})
ZGT-5	2.25	1295 ± 65	Cladocera remains		1980	
ZGT-17	8.25	1695 ± 35	Cladocera remains	1655	1910	0.08594
ZGT-51	30.75	1880 ± 45	Cladocera remains	1540	1610	0.075

400 yr. Hence, the sedimentation rate was calculated to be 0.075 cm/yr between 8.25 cm and 30.75 cm depth (Table 1). The chronology of the core between 30.75 cm and 60.75 cm was extrapolated according to the sedimentation rate of 0.075 cm/yr, and an age of 1208 AD was obtained for the sediment at 60.75 cm depth. Our chronology has a resolution of 5–7 yr from 1755 AD to the present and 13–14 yr between 1200 and 1755 AD based on the age model and sampling intervals.

4.2 Oxygen and carbon isotopes

The δ^{18}Ocarb values vary between −8‰ and −2‰, and the δ^{13}C fluctuates between + 2‰ and + 7‰ during the past 800 yr. Several periods of abrupt positive shifts of δ^{18}Ocarb with fluctuations exceeding 3‰ were observed between 1200 and 1820 AD (Fig. 3). The δ^{18}Ocarb values fluctuate with minor amplitudes from −8‰ to −6‰ after approximately 1820 AD. The $\delta^{13}C_{carb}$ values range between + 2‰ and + 7‰, and display higher amplitudes between 1200 and 1820 AD. than those after this date range. The $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ values correlate owing to the hydrologically closed lake system (Fig. 3). The correlation coefficient was 0.76 ($R^2 = 0.58$) for the entire period under consideration.

4.3 Water content, organic and carbonate content

There were strong fluctuations in water, organic matter, and carbonate content over the past 800 yr (Fig. 3). Water content ranged from 50%–70%, and the lowest value occurred during 1360–1860 AD. Organic and carbonate contents fluctuated between 10%–28% and between 5%–30% respectively, and had good anti-correlation with each other. The organic and carbonate contents showed large fluctuations with amplitudes of > 10% from 1200 to 1820 AD. The lowest organic content and highest carbonate content occurred after 1820 AD. The organic content has decreased from 16% to 8%, and the carbonate content has increased from 15% to 30% since the 1940s.

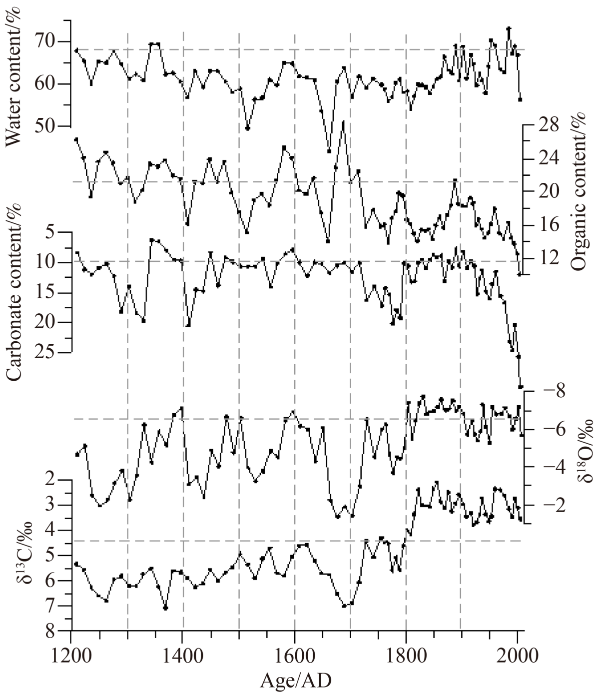

Fig. 3 Variations of water content, organic and carbonate content, $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$(VPDB) in fine carbonates from core A-1 since 1200 AD

4.4 Grain size

The median size of the lake sediment in Zige Tangco varies between 3.5 and 22 μm (Fig. 5), with an average of 6.6 μm. Clay (< 4 μm) and fine silt (4–20 μm) content account for over 80% of the lake sediments. Coarse silt (20–63 μm) accounts for 10%–15% of the lake sediments. Under 10% of the lake sediment was composed of sand (> 63 μm). The several abrupt increases in sand content between 1400 and 1820 AD corresponded to the positive shift of $\delta^{18}O_{carb}$. The clay content increased to a relatively higher value since 1820, whereas the sand content decreased to a lower value with smaller fluctuations.

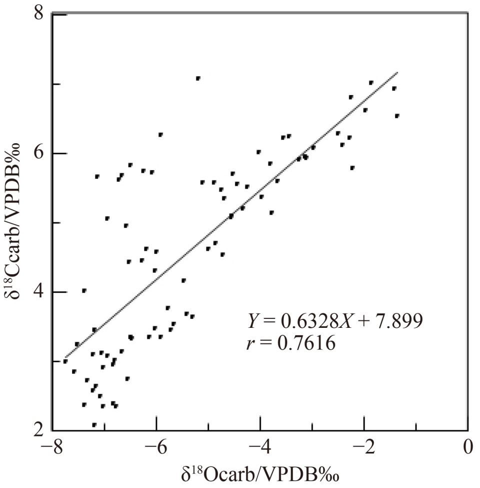

Fig. 4 Isotopic composition of fine-grained carbonate in core A-1 and the analysis of correlation coefficient between $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$.

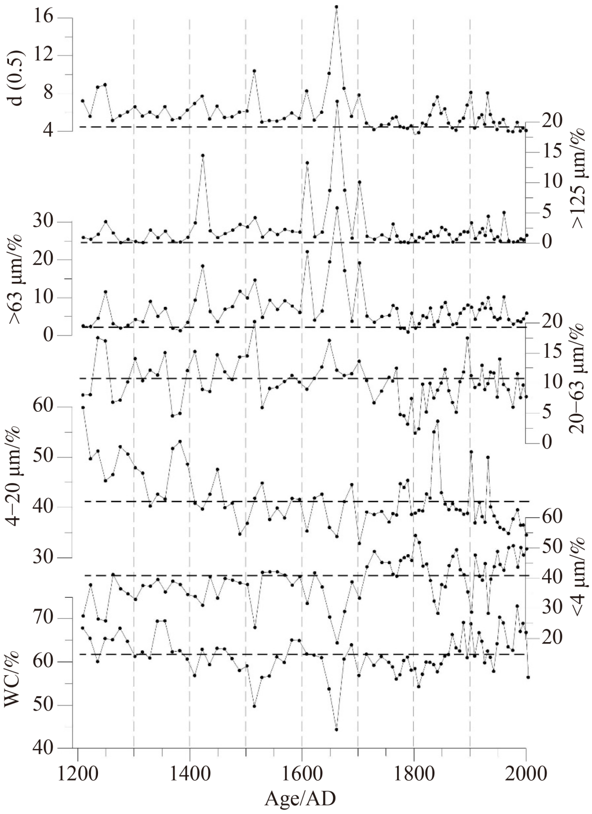

Fig. 5 Time series of water content and grain-size distribution record of core A-1 since 1200 AD.

5 Discussion

5.1 Interpretation of the stable isotopes in fine-grained carbonates

Stable isotopes in authigenic carbonates have become useful tools for the paleohydrological reconstruction of closed-basin lakes (Talbot 1990; Li and Ku, 1997; Schwalb et al., 1995; Schwalb, 2003; Leng and Marshall, 2004; Liu et al., 2008; Zhang et al., 2014). Oxygen isotopes in fine-grained carbonates reflect water temperature and host water chemistry (Henderson et al., 2003). An increase in water temperature can cause a decrease of $\delta^{18}O_{carb}$ in carbonate at a rate of 0.24‰ per 1°C (Wei and Gasse, 1999). The range of > 5% in $\delta^{18}O_{carb}$ in core A-1 would correspond to a temperature change of over 20°C during LIA, assuming that the observed $\delta^{18}O_{carb}$ fluctuations are mainly caused by water temperature. Such large temperature variations are more extreme than those during the glacial–Holocene transition (Thompson et al., 1997), and may not have occurred during the LIA. Hence we assume that water temperature changes cannot be used to explain the fluctuations of

$\delta^{18}O_{carb}$ in Zige Tangco.

The $\delta^{18}O_{carb}$ value of the host water mainly reflects the regional effective moisture, which is related to variations in precipitation/evaporation (P/E) ratios and the isotopic composition of the water inflow to the lake (Wei and Gasse, 1999; Leng and Marshall, 2004). The $\delta^{18}O_{carb}$ of water inflow includes the isotopic composition of meteoric waters linked to the origin and type of condensation of air moisture and the average temperature during the rainy (snowy) season (Wei and Gasse, 1999). Yao et al. (1996) suggested that the $\delta^{18}O_{carb}$ of precipitation on the Tibetan Plateau correlates with a temperature at a rate of + 0.75‰ per °C. The temperature during LIA would be 6°C–7°C higher than today if $\delta^{18}O_{carb}$ value fluctuations in core A-1 are mainly caused by the variation of $\delta^{18}O_{carb}$ in water inflow, which is impossible. Therefore, the $\delta^{18}O_{carb}$ fluctuation in core A-1 cannot be interpreted as the variation of $\delta^{18}O_{carb}$ in inflow water.

The $\delta^{18}O_{carb}$ value in Zige Tangco and its rivers were previously investigated (Li, 2001). The $\delta^{18}O_{carb}$ value of precipitation collected around Zige Tangco on Aug 10 and 12, 1998 was −17.35‰ and −13.26‰, respectively. The $\delta^{18}O_I$ value varied between −15.35‰ and −12.97‰ from river samples collected on August 6th−13th, 1998. The $\delta^{18}O_I$ value is close to $\delta^{18}O_p$ but fluctuated at a lower amplitude than $\delta^{18}O_p$ owing to the buffering effect. The δ^{18}Ocarb value of lake water collected from the lake surface on August 4th−12th ranged from −5.43‰ to −5.12‰ with an average value of −5.12‰. The big difference in $\delta^{18}O_{carb}$ between lake water and inflows indicates that the oxygen isotope composition of Zige Tangco is greatly concentrated through evaporation and that the annual precipitation can dramatically affect the $\delta^{18}O_{carb}$ of lake water. The annual evaporation in Zige Tangco is approximately 800–1110 mm (Li et al., 2001b); this is much higher than the annual precipitation in this region (240–500 mm) leading to a dry climate. We speculate that evaporation could have had a significant effect on the isotopic composition of lake water under these climatic conditions. Evaporation preferentially removes water containing ^{16}O from the water body owing to its high vapor pressure; this increases the $\delta^{18}O_{carb}$ composition of the residual lake water. Therefore, the variation of $\delta^{18}O_{carb}$ in Zige Tangco during the past 800 yr can be interpreted as the variation of P/E ratios and/or lake volume fluctuation. The abrupt positive shift of $\delta^{18}O_{carb}$ between 1200 and 1820 AD indicates low P/E ratios or a decrease in lake level, and the negative shift of $\delta^{18}O_{carb}$ value since 1820 AD can be interpreted as an increase in lake level.

The $\delta^{13}C_{carb}$ value of authigenic carbonates mainly reflects the $\delta^{13}C_{carb}$ value of dissolved inorganic carbon (DIC) in lake water (Leng and Marshall, 2004). The $\delta^{13}C_{carb}$ value of dissolved inorganic carbon within a lake is mainly controlled by three factors: the photosynthesis/respiration cycle, CO_2 exchange between the atmosphere and lake water, and the isotope composition of the inflow water (Henderson et al., 2003; Leng and Marshall, 2004). The productivity at Zige Tangco was not high enough to significantly change the $\delta^{13}C_{DIC}$ owing to the high DIC concentration (> 10 g/L). The $\delta^{13}C_{DIC}$ in Zige Tangco in 2006 is + 2‰− + 4‰ (Lei et al., 2011) and is significantly higher than that in rivers (approximately −4‰). This indicates that DIC in the lake experiences long-term elevated concentrations. The variation of $\delta^{13}C_{carb}$ in Zige Tangco is mainly interpreted in terms of CO_2 exchange between the atmosphere and lake water. The DIC in the lake is concentrated to a larger extent in the dry period, and CO_2 escape into the atmosphere is larger; this results in a positive shift of $\delta^{13}C_{DIC}$ (Li and Ku, 1997). The δ^{13}Ccarb in Zige Tangco can also be influenced by $\delta^{13}C_{DIC}$ in rivers. Organic matter decomposition in warm and wetter climates accounts for a larger portion of the DIC in rivers, which leads to more negative values of $\delta^{13}C_{DIC}$. Hence, the high value of $\delta^{13}C_{carb}$ between 1200 and 1820 AD indicates a low lake level and dry climate. The negative shift of $\delta^{13}C_{carb}$ from 5.5 ‰ to 2.5 ‰ indicates increasing lake levels or climate transition from dry to wet conditions between 1800 and 1820 AD.

The correlation between $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ variation in closed lakes is high ($r \geqslant 0.70$) and covariance can be treated as a test for open/closed history of lake hydrology (Talbot, 1990). The high correlation coefficient (r = 0.7616) between $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ in Zige Tangco indicates that their covariance can be used as an indicator of lake paleohydrology (Talbot, 1990). $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ evolve toward heavier and lighter values in a closed lake, when the lake volume decreases and increases, respectively. Positive values of $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ between 1200 and 1800 AD indicate a relatively lower lake level, whereas negative values after 1820 AD indicate a higher lake level, with an abrupt rise between 1800 and 1820 AD. $\delta^{13}C_{carb}$ fluctuates with much less amplitude than $\delta^{18}O_{carb}$ between 1200 and 1820 AD. The cause of $\delta^{13}C_{carb}$ insensitivity to lake level fluctuation is owing to the high alkalinity in Zige Tangco dampening the variation of $\delta^{13}C_{DIC}$ when the lake level is relatively low. A high correlation is suitable for less alkaline lakes or in the stage of relatively rapid increases and decreases in lake volume; furthermore, a covariant trend may not be found in hyperalkaline lakes owing to the insensitivity of $\delta^{13}C_{carb}$ to lake volume changes (Li and Ku, 1997). Therefore, fluctuation of the $\delta^{18}O_{carb}$ value was more sensitive to lake level changes than $\delta^{13}C_{carb}$ in Zige Tangco during the past 800 yr.

5.2 Interpretation of other proxies

The organic content is affected by many factors, and is generally an indicator of productivity in lakes and catchments. The organic matter in lake sediments on the

Tibetan Plateau is mainly controlled by terrestrial organic inputs and internal lake productivity (Wang et al., 1997; Zhang et al., 2003). Therefore, the sharp decrease in organic content between 1200 and 1820 AD can be interpreted as a low organic input and lake productivity.

Zige Tangco is a carbonate-type lake with a high DIC concentration (Li, 2001) that facilitates the precipitation of carbonate minerals. Carbonate minerals in Zige Tangco mainly precipitate in summer caused by relatively high-water temperature and intensive evaporation (Lei et al., 2011); this is partly controlled by calcium inflow with high carbonate content. The high carbonate content in the sediment may reflect aridity, which is linked to evaporation. Therefore, dry climate conditions occurring between 1200 and 1820 AD correspond to the records of stable isotopes.

The organic and carbonate content in core A-1 differs from the natural process of lake evolution since the 1950s. The organic content and carbonate content decreased and increased to their lowest and highest values during the past 800 yr, respectively. The decrease in organic content from 16% to 8% since the 1950s was probably owing to a drop in lake productivity. Conversely, the abrupt increase in the carbonate content from 15% to 30% was probably owing to global warming over the past 50 yr.

The dry periods indicated by carbonate content and stable isotopes were not synchronous during the LIA. The occurrence of carbonate content peak is 40–50 yr later than the abrupt positive shift of $\delta^{18}O_{carb}$ (Fig. 3). However, the minimum organic matter is coincident with the abrupt positive shift of $\delta^{18}O_{carb}$, but jumps to a higher value more quickly than the negative shift of $\delta^{18}O_{carb}$ (Fig. 3). A possible reason for this asynchrony is that the calcium concentration in the lake is very low owing to its high alkalinity when the calcium input by inflow is high and not during the driest period.

The grain size of lake sediments in Zige Tangco is mainly controlled by lake level variations and vegetation cover in the surrounding area. Coarse material transported by rivers reaches the core site more easily during summer season when the lake level is low (Peng et al., 2005; Shen et al., 2007). More importantly, coarse materials from the surroundings can be transported onto the frozen surface by strong winds in winter and spring and deposited on the lake bottom after ice melts in late spring (Qiang et al., 2006). This may account for a large amount of sand content in cold and dry regions, such as Zige Tangco. Hence, we interpreted the high sand content as a strong winter monsoon impact under sparse vegetation during 1400–1780 yr.

The hydrological variation over the past 800 yr can be divided into two stages based on the analysis of all proxies. The dry and unstable climate in Zige Tangco with some abrupt drops in lake level recorded between 1200 and 1820 AD corresponded to the Little Ice Age. Subsequently, a wet climate occurred after 1820 with relatively small fluctuations in the lake level. However, human activity has influenced the natural process of lake evolution in Zige Tangco according to the organic and carbonate content since 1950s. Similarly, the Ximencuo sediments show that human activity increased during natural processes (Pu and Meyers, 2022).

5.3 Comparison with other records

The hydrological variations of Zige Tangco from 1200 AD were compared with well-dated and high-resolution records on the Tibetan Plateau, including $\delta^{18}O_{carb}$ in the Guliya ice core (Yao et al., 1996; Thompson et al., 1997) and precipitation reconstructed from tree rings at Delingha (Shao et al., 2005). The low lake level or dry periods recorded in core A-1 during the LIA are synchronous with the cold and weak summer monsoon periods reconstructed from the Guliya ice core and tree rings within the age uncertainty (Fig. 6). The low lake level at Zige Tangco was recorded using independent proxies in core A-1 between 1235 and 1315 AD. The $\delta^{18}O_{carb}$ value in the Guliya ice core shows negative shifts of over 3‰ between 1250 and 1300 AD. Precipitation in Delingha experiences low values from 1250 to 1310 AD. The dry period at Zige Tangco from 1410 to 1580 AD. is also recorded in these archives. Low temperatures during 1430–1530 AD. were indicated by low $\delta^{18}O_{carb}$ values in the Guliya ice core, with some strong fluctuations. Low precipitation occurred in the Delingha tree rings between 1420 and 1510 AD. The oscillation between the dry and wet periods in Zige Tangco is similar to the fluctuation of the lake level in Nam Co reconstructed from ostracod transfer functions (Wrozyna et al., 2010; Zhang et al., 2014).

Relatively high lake water levels or wetter periods in Zige Tangco during 1580–1650 AD and 1820–1900 AD are also found in other studies. Precipitation in Delingha around the 1600s was at its highest value during the past millennia. The temperature during the 1600s recorded by $\delta^{18}O_{carb}$ in the Guliya and Malan ice cores was relatively high (Yao et al., 1997; Wang et al., 2006a). An obvious desalination reconstruction from the diatom-based conductivity transfer function in Chen Co and Nam Co during 1845–1885 AD indicated a cold and moist climate (Yang et al., 2004; Wang et al., 2006b). The relatively dry period of 1900–1950 AD reconstructed from core A-1 is also reflected by lake level fluctuations in Nam Co (Wrozyna et al., 2010) and low precipitation in Delingha tree rings (Shao et al., 2005). Cold periods are not always synchronous with low precipitation in Zige Tangco; cold and wet periods with relatively high lake water levels occurred during 1820–1900 AD.

The upwelling of the Arabian Sea indicated weakening of the Indian summer monsoon during the LIA (Gupta et al., 2003). However, the fluctuation of the Indian

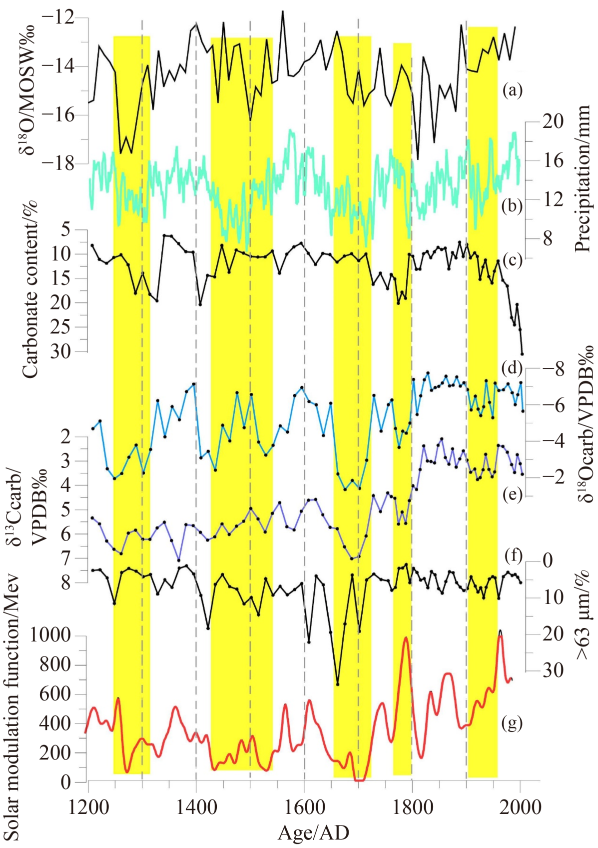

Fig. 6 Comparisons of proxies in core A-1 with other archives on Tibetan Plateau since 1200 AD. Shadow area means dry periods or low lake level. (a) δ^{18}O of Guliya ice core (Thompson et al., 1997) (SMOW), (b) precipitation reconstructed from tree rings at Delingha (Shao et al., 2005), (c) carbonate content in core A-1, (d) δ^{18}O$_{\text{carb}}$ of fine-grained carbonates in core A-1(VPDB), (e) δ^{13}C$_{\text{carb}}$ of fine-grained carbonates in core A-1(VPDB), (f) sand content (> 63 μm) in core A-1, (g) solar modulation function (Muscheler et al., 2007).

summer monsoon during the LIA is unclear owing to the low resolution of marine sediments. Over 85% of the annual precipitation around Zige Tangco occurs during the summer monsoon season; therefore, lake level variation is an indicator of the intensity of the summer monsoon on the central Tibetan Plateau. Three dry periods or low lake levels were recorded in Zige Tangco during the LIA by independent proxies, corresponding to low periods of solar irradiance (Fig. 6) (Muscheler et al., 2007).

Published records of total solar irradiance (TSI), δ^{18}O variations in the Greenland ice core, biologically sensitive indicators from Lake Qinghai, and variations in the δ^{13}C$_{\text{org}}$ values in Lake Ngoring sediments can reflect lake level fluctuations. High-lake-level phases correspond to higher North Atlantic Oscillation Index (NAOI) values and lower Southern Oscillation Index (SOI pr) (Pu et al., 2020). Earlier studies show that the intensity of the Indian Monsoon is strongly correlated with interdecadal variations in various indices of the El Niño–Southern Oscillation (ENSO) (Torrence and Webster, 1999; Krishnamurthy and Goswami, 2000). A decrease in Indian monsoon rainfall is generally associated with the warm phases of ENSO and vice versa (Pu et al., 2020). It seems that stronger solar radiation will amplify the evaporation of water from the ocean and increase the mean meridional temperature gradient between land and sea in middle-low latitudes, leading to an intensified summer monsoon that further increases rainfall and corresponds to a negative ENSO event (Krishnamurthy and Goswami, 2000). Meanwhile, a warmer North Atlantic caused by a higher TSI would lead to higher evaporation from the surface seawater and an increased pressure difference between the Azores High and the Icelandic Low. The Indian Monsoon intensifies during summer and controls the pattern of atmospheric circulation in the source area of the Yellow River (SAYR), whereas the westerly circulation moves southward in winter and plays a major role in local precipitation (Pu et al., 2020). This model indicates that both atmospheric systems would deliver more moisture to the SAYR during periods of intense solar radiation.

A similar phenomenon was observed in the Yellow River during 1950–2000 by comparing monthly SST anomalies, SOI, and annual precipitation (Wang et al., 2006c). The SOI may have a significant impact on precipitation variability in the entire Yellow River, including its source area on the TP. Endorheic lakes in the interior of the Tibetan Plateau have expanded and dramatically deepened in response to considerable increases in precipitation since the late 1990s (Lei et al., 2013, 2014, 2019; Yang et al., 2018; Wang et al., 2023). Thus, we assumed that solar activity may have played a key role in controlling summer monsoon variability on the central Tibetan Plateau and influenced the Zige Tangco hydrological variation during the LIA.

6 Conclusions

The hydrological variations in Zige Tangco were reconstructed from 1200 AD using multiple proxies, and the following conclusions were drawn.

Multi-proxies (including organic and carbonate content, stable isotopes of fine-grained carbonate minerals (< 38.5 μm) and grain-size distribution) were used to reconstruct hydrological variation in Zige Tangco since 1200 AD. Zige Tangco experienced at least three dry periods or abrupt drops in lake level at 1235–1315 AD, 1410–1580 AD, and 1660–1720 AD. The dry periods in Zige Tangco correlated with low δ^{18}O values in the Guliya ice core and low precipitation reconstructed by tree rings in Delingha. Hence, we assume that the Indian summer monsoon weakened and the winter monsoon on the central Tibetan Plateau strengthened during the LIA. Our data also indicate the control of lake hydrology in

Zige Tangco by solar forcing.

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