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Climatic changes documented by stable isotopes of sedimentary carbonate in Lake Sugan, northeastern Tibetan Plateau of China, since 2 kaBP

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Abstract **Lake Sugan at the northern edge of the Qaidam Basin was selected as the research object. The temporal sequence of sedimentary cores retrieved from Lake Sugan since 2 kaBP was reconstructed using the ^{210}Pb, AMS ^{14}C and conventional ^{14}C dating methods. Carbon and oxygen isotopes of carbonate in the fine-grained lake sediments were analysed. Combined with the changes of δ^{18}O values of surface water and air temperature observation data in the study area, it might be thought that the δ^{18}O value of the carbonate indicates effective moisture, and the changes in δ^{13}C values are related to annual freeze-up duration of the lake and indirectly indicate air temperature changes in winter half year. From the above, the sequence of climatic changes in the region since 2 kaBP was established. The climatic changes experienced five stages: Warm-dry climate during 0—190 AD; cold-dry climate during 190—580 AD; warm-dry climate during 580—1200 AD (MWP); cold-wet climate during 1200—1880 AD (LIA); cold-dry climate during 1880—1950 AD; and climate warming since 1950s. The air temperature changes in winter half year reflected by carbon isotope since 2 kaBP are in good agreement with the historical literature records and other geologic records, which shows that the climate changes recorded by the stable isotopes from Lake Sugan since 2 kaBP are of universal significance.**

Keywords: Lake Sugan, carbonate, stable isotope, climatic changes.

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The time interval since 2 kaBP is an important "time window" to link up geologic records, instrumental data and literature data, and further to predict the future changes of human living environment. Therefore, it has long been one of the important time intervals in the study of the past global changes (PAGES). Over the past many years, the high-resolution climatic changes sequence was reconstructed by using the climatic information recorded by ice cores[1—3], tree rings[4], lake sediments[5,6] and historical literatures[7—11] since 2 kaBP. Moreover, characteristic climatic stages such as the Medieval Warm Period (MWP), the Little Ice Age (LIA) and the warming in the 20th century were identified. The Qaidam Basin, located in the northeastern part of the Tibetan Plateau, with an area of 121000 km^2, is one of the typical arid regions in the interior of China or even in Asia. Owing to the limitation of sedimentary strata, the previous researches of palaeoclimatic changes in the Qaidam Basin have two features: one is selecting saline lakes as the research objects and the other is concentrating on the low-resolution climatic reconstruction on long timescale during the Quaternary[12—17]. In addition, no reports of Holocene palaeoclimatic research were available, especially high-resolution climatic changes since 2 kaBP in the region. In view of this, the sediments of Lake Sugan at the northern edge of the Qaidam Basin were employed to reconstruct the climatic changes sequence since 2 kaBP in the region with a resolution of 50a by means of stable isotopic geochemistry. The climatic changes history was discussed and compared with other relevant records. This paper provides some basic data for further understanding the aridification history and the climatic changes in the characteristic periods since 2 kaBP in the Qaidam Basin and northwestern China, and it will contribute to a profound understanding of the palaeoclimatic implications of carbon and oxygen isotopes of lacustrine carbonate.

1 Materials and methods

The Sugan Lake basin (38°51′N, 93°54′E) is a closed depression mosaic in the Qaidam Basin (Fig. 1). Its relief is high in the east and low in the west, with an elevation varying between 2800 m and 3200 m a.s.l. Mean annual temperature in the western part of the basin is 2.75℃, annual precipitation 18.7 mm and annual potential evaporation 2967.2 mm, indicating the hyperarid zone. Vegetation in the basin is sparse and there are grasslands to the east of the lake, where groundwater seeps out. Other regions mostly exhibit a desert landscape, such as gobi, sandy gravel land and sandy land. A large part of the region is sparsely populated and only Kazak people herd their livestock there. Lake Sugan as a closed lake is the confluence center of surface water and groundwater and also the base level of groundwater cycle. The nourishment water of the basin mainly comes from the two rivers, the Big Harteng River and the Small Harteng River in the eastern part of the basin. After reaching the gobi zone of the piedmont plain, the river water entirely infiltrates into the 8—15 km wide subsurface layer, then it rises to the surface as spring water in the fine earthy plain to the east of the lake and forms rivers to recharge the lake. Lake

Sugan has an area of 103.68 km^2, its mean water depth is 2.5 m, and the water at its sedimentary center is about 5 m. Mean salinity of modern lake water is 31.83 g/l, belonging to Cl-SO$_4$-Na-Mg chemical type, with a pH value of 8.5 and an electrical conductivity of 2.18 m/s (25℃). The lake water has a higher HCO^{3-} content and Mg^{2+}/Ca^{2+} ratio, and they are 683.5 mg/l and 21, respectively. The data of boreholes drilled from the Mahai Lake to the south of the Sugan Lake basin reveal that Lake Sugan had been closed since the late Pleistocene[18]. Accordingly, Lake Sugan is an ideal site for the research of Holocene climatic changes in the region.

In December 2000, five cores were drilled in the sedimentary center of Lake Sugan with the Piston corer. The boreholes SG001227-C, SG001227-D and SG001227-E were parallel holes (38°51.19′N, 93°54.09′E), with a water depth of 4.6 m. The SG001227-C core has a drilling footage of 8.52 m and the core length is 8.34 m. The SG001227-D and SG001227-E cores are short-hole drilling and their depths are 3.0 m and 1.2 m, respectively. After lifting up the core, both ends of the sampling tube were sealed using adhesive tape. From bottom to top, the SG001227-C core consists of grayish white silty clay at the depth of 852—473 cm, alternately gray-white laminated clay at the depth of 473—178 cm, and dark gray silty clay at the depth of 178—0 cm. The core shows a good sedimentary continuity and no depositional breaks are found. For the sample division, the length of the SG001227-C core was calibrated to the actual drilling depth and its surface 1—2 mm polluted layer was scraped off. Then it was cut into 785 pieces of samples at 1—2 cm interval. They were oven-dried to constant weights at 60℃ and weighed. The SG001227-E core was cut at 1 cm interval and oven-dried at 110℃. Their dry bulk densities were calculated and then used for the analyses of ^{210}Pb and ^{137}Cs. Owing to lack of terrigenous organic remains in the sediments of Lake Sugan, the seeds of submerged plants *Ruppia maritima* in the lake were used for the analysis of AMS ^{14}C age. A total of 36 surface samples of the lake sediments were collected using the GRAB driller, with a drilling depth ranging from 3 cm to 5 cm. In January 2002, 16 surface water samples were collected in the basin, including one snowfall sample, one river water sample, one spring sample and 13 lake water samples. In addition, two snowfall samples were collected at Liushaping in the central part of the Qaidam Basin and the Dachaidan Town. These snow samples represent the same snowfall process.

The modern sedimentation rate of the lake was determined by the ^{210}Pb and ^{137}Cs dating. 5 g oven-dried sample was weighed, picked out organic remains and then ground to pass 100-mesh sieve. The ground samples were placed into a 5ml cylindrical plastic tube and the determination was conducted using the γ spectral analytical system consisting of 16k multichannel analyzer, including the high-purity germanium well-type detector, Ortec 919 spectral controller and IBM microcomputer made by EG & Ortec Co. USA. The ages of surface samples of Lake Sugan were determined by the conventional ^{14}C dating method. Since it requires a large number of surface samples, five surface samples were mixed to form a dating sample and the ages of organic carbon and inorganic carbon components of the sample were determined using LKB-1220 ultra-low background liquid scintillator. Surface water samples were analysed by a MAT-252 gas mass spectrometer using CO$_2$-H$_2$O equilibrium method, with a precision of ±0.05‰, and results were expressed by SMOW standard.

Carbonate content in the SG001227-C core varies be-

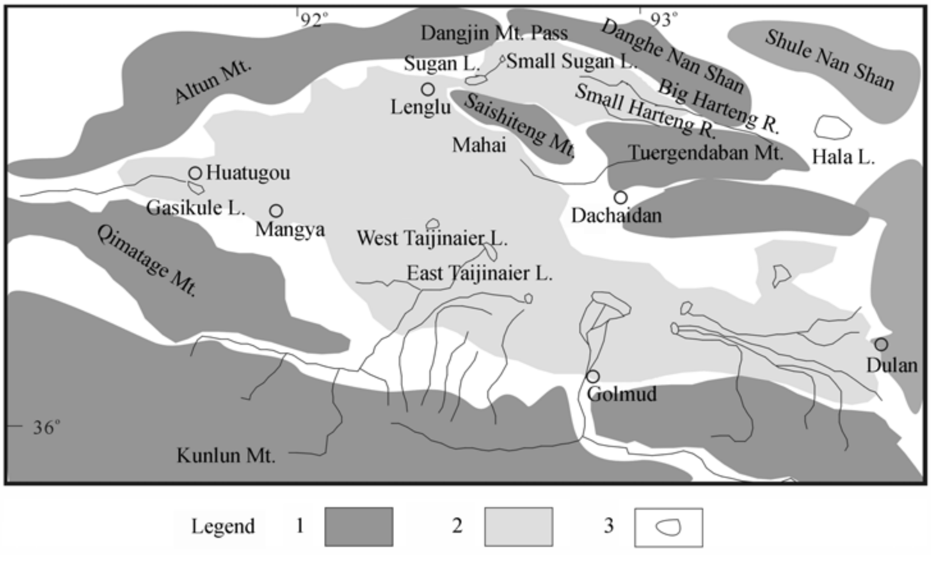

Fig. 1. Physiographic situation in the study area. 1 Mountain; 2 basin; 3 lake.

tween 29.7% and 89.1% and averages 49% above the depth of 250 cm. The stable isotopes of carbonate in Lake Sugan were analysed using fine-grained sediments. The core samples were washed with two layers of sieves. The upper one has a sieve mesh of 125 μm and the lower one has a sieve mesh of 38.5 μm. The 125 μm sieve was used to separate ostracod shells and thereby to eliminate the effect of biogenic carbonate on the analytical results, for example, the "vital effect", etc. The samples passing the 38.5 μm sieve were collected and oven-dried for the analyses of carbon and oxygen isotopes. The carbonate component in fine-grained samples was determined by X-ray diffraction method (XRD). The results (Table 1) show that the carbonate minerals are dominated by aragonite and calcite, accounting for about 50%, and the dolomite is relatively low (<8.2%). Some fine-grained samples were examined under the scanning electronic microscope and found that there are rhombohedral calcite and acicular aragonite with perfect crystal form, which demonstrates that the samples contain large amount of authigenic carbonate precipitating from surface lake water. High salinity, high alkalinity and high Mg/Ca ratio of a lake are favorable to the deposition of dolomite. In the high-evaporation environment, dolomite is considered to be an evaporite mineral, for example, the modern dolomite is found in the lakes in Victoria region of Australia[19]. In some cases, authigentic dolomite minerals also exhibit a rhombohedral crystal form. Taking the sedimentary environment in Lake Sugan into account, authigenic dolomite minerals are very likely to occur in the lake. The nourishment water of Lake Sugan mainly comes from glacier meltwater. The main nourishment source is groundwater in most part of the basin and the rivers slowly flowing into the lake are also derived from spring. Therefore, there may be no chance for the rivers to bring clastic grains to the lake. Moreover, precipitation is sparse in the study region and it is insufficient to cause large runoff. Wind erosion in the study area is severe, and especially fine particles are subject to constant deflation. There is no large area of limestone exposed in the region. There is a large area of playa to the west of Lake Sugan, which seldom produces clastic carbonate. Therefore, the wind-deposited clastic carbonate particles in the lake sediments cannot affect the analysis of stable isotopes. To sum up, we think that the fine-grained carbonate in the sediments of Lake Sugan basically represents the authigenic carbonate and can be used for the isotopic analysis.

Stable isotopes of core samples were determined by phosphoric acid method[20] using Delta Plus gas mass spectrometer. 20—30 mg of fine-grained sample was weighed, and then 1 ml 100% H_3PO_4 was added into the sample from side neck to react for 20 min at 90℃. The resultant CO_2 gas was purified and collected for isotopic analysis. The precisions of $\delta^{13}C$ and $\delta^{18}O$ were ±0.054‰ and ±0.153‰, respectively. The isotopic results are expressed by $R = (R_{\text{sample}} - R_{\text{standard}})/R_{\text{standard}} \times 1000‰$ (PDB, $R = \delta^{13}C/\delta^{18}O$). The upper 29 cm of SG001227-C core was used for the isotopic analysis at about 1 cm interval and below the depth of 29 cm it was analysed at about 4 cm interval.

2 Results

2.1 Chronology

The determined results of ^{210}Pb and ^{137}Cs of the SG001227-E core are shown in Fig. 2. The age of ^{210}Pb was calculated using the constant rate of supply model (CRS). The model can quantitatively reflect the variations of depositional flux with time and thereby obtain rational ages[21,22]. From the sedimentation flux, it has been found that the modern depositional rate of Lake Sugan is 0.31 cm/a through the fitting of least squares method. The age of modern sediments at the depth of 26.6 cm in Lake Sugan is 1845. A 155-year sequence of the sediments was obtained. The results of ^{137}Cs show that it had a marked peak value in 1963 and was older than that of ^{210}Pb in the same horizon. The same situation has been found in the South Hongshan Lake[23] and it may be related to the migration of ^{137}Cs nuclide[24]. Here, we used the ^{210}Pb dating to establish the time sequence of the core.

The seed amount of *Ruppia maritime* in the deposits is limited and they only occur in a few layers of the SG001227 core. Therefore, only four samples can be used for AMS ^{14}C dating. *Ruppia maritime* is perennial or annual submerged plant of *Potamogetonaceae* with rigid subterraneous stem and multiramose aerial stem. It is cespitose and occurs in the tropic oceans and inland saline lakes in the world[25]. The KIA18572 sample is a piece of plant remains with $\delta^{13}C$ value of −18.29‰ (Table 2). Other analyses were not conducted due to the limited amount. Judged from the modern environmental features of Lake Sugan basin and the value of its carbon isotope, the plant remains should come from endophyte of the lake. Submerged plants and endophytes can take up soluble inorganic carbon in the lake water for photosynthesis, and hence the results of ^{14}C dating may be affected by reservoir effect. The results of AMS ^{14}C dating (Table 2) show

Table 1 XRD analytical results of carbonate minerals in fine fraction (<38.5 μm) of the sediments in Lake Sugan

Sample No.	Depth/cm	Carbonate/%	Aragonite/%	Calcite/%	Dolomite/%
SG-C-000	0.5	56.4	0	53.4	3.0
SG-C-051	56.1	52.1	0	48.3	3.8
SG-C-107	117.2	56.6	23.7	24.7	8.2
SG-C-207	226.3	53.8	27.2	18.6	8.0

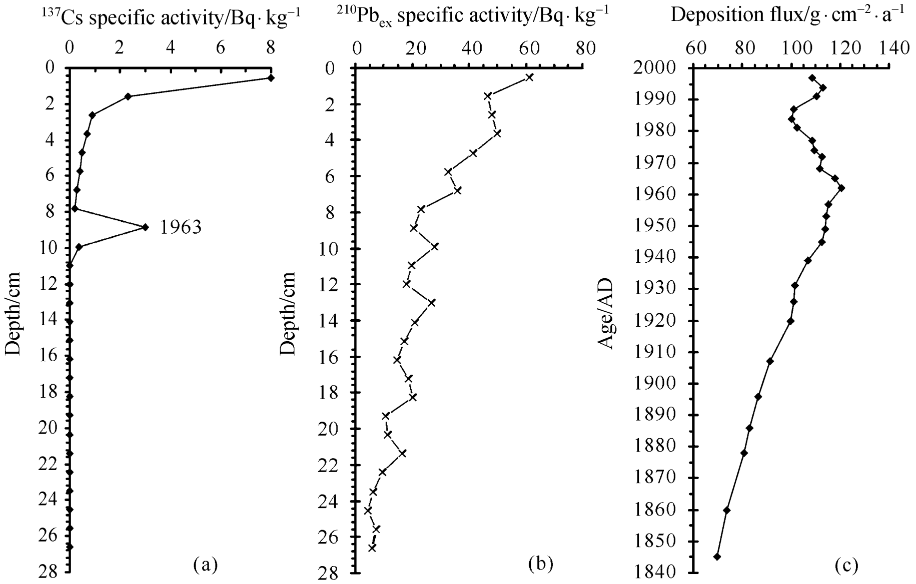

Fig. 2. Determined results of ^{210}Pb and ^{137}Cs of the modern sediments of Lake Sugan.

Table 2 Results of AMS ^{14}C dating of the SG001227-C core in Lake Sugan

Lab No.	Sample No.	Material	Seed/remains number	Depth /cm	δ^{13}C /‰	^{14}C age /a BP	Calibrated age /Cal a BC
KIA18567	SG-C-090	Seed	3	98.6	−10.15	2950±35	1209; 1200; 1191; 1177; 1163; 1140; 1131
KIA18568	SG-C-117		5	128.1	−9.47	3050±25	1369; 1360; 1347; 1344; 1316
KIA18569	SG-C-144		5	157.5	−12.80	3264±49	1521
KIA18571	SG-C-378		3	422.4	−15.43	5612±43	4455; 4415; 4414
KIA18572	SG-C-529	Plant remains	1	580.7	−18.29	7004±32	5873; 5855; 5843

that the ^{14}C age at the depth of 98.6 cm is (2950±35) a BP. Assuming that the sedimentation rate is approximate to 0.31 cm/a and constant above the depth of 98.6 cm, the ^{14}C ages must be significantly affected by reservoir effect. Combined with the ^{210}Pb ages, the reservoir effect was estimated to be 2200a after the ^{14}C ages were converted into the calendar ages[26] (Table 2).

In order to further check the influence of reservoir effect, we determined the ^{14}C ages of the surface sediment samples of the lake. PM value of inorganic carbon of the surface sample is 0.9951±0.0077, and the age of inorganic carbon is about 2594a, being slightly older. Organic carbon PM value is 1.0147±0.0075, with a δ^{13}C value of –16.2‰. The age of organic carbon is about 2433a, being slightly older. Considering the errors of the AMS ^{14}C dating and the conventional ^{14}C dating and burial depth difference of the surface samples, it is reasonable to deduct the reservoir effect (2200a) from the ^{14}C ages. Through the field investigations and the sampling of surface sediments, it has been found that there is a large amount of *Ruppia maritime* on the floor of the lake. They are the main source of organic matter in the surface sediments. Hence, the reservoir effect on the ages obtained from the *Ruppia maritime* seeds is relatively approximate to the age from the organic carbon in the surface sediments and this also has been demonstrated by the δ^{13}C value of the organic matter in the surface sediments. Based on the linear regression of the ages deducted the reservoir effect, the temporal sequence of the lake sediments since 2 kaBP was established and the age at the depth of 212 cm is about 2000 Cal aBP. The resolution of each sample above the depth of 29 cm in the SG001227-C core is 9—11a, while the resolution below the depth is 42a.

2.2 Stable isotopes

The δ^{18}O values of snow samples vary between –20.79‰ and –29.36‰. The δ^{18}O values of the Small Sugan Lake water and the Sugan River water are essentially the same within the analytical accuracy, and they are –4.33‰ and –4.30‰, respectively. The δ^{18}O value of spring water to the east of the lake is –2.98‰. The δ^{18}O values of Lake Sugan water vary between 2.40‰ and 5.85‰ with a mean value of 4.00‰.

The stable isotopic results of carbonate from Lake Sugan are shown in Fig. 3. The δ^{18}O values vary between

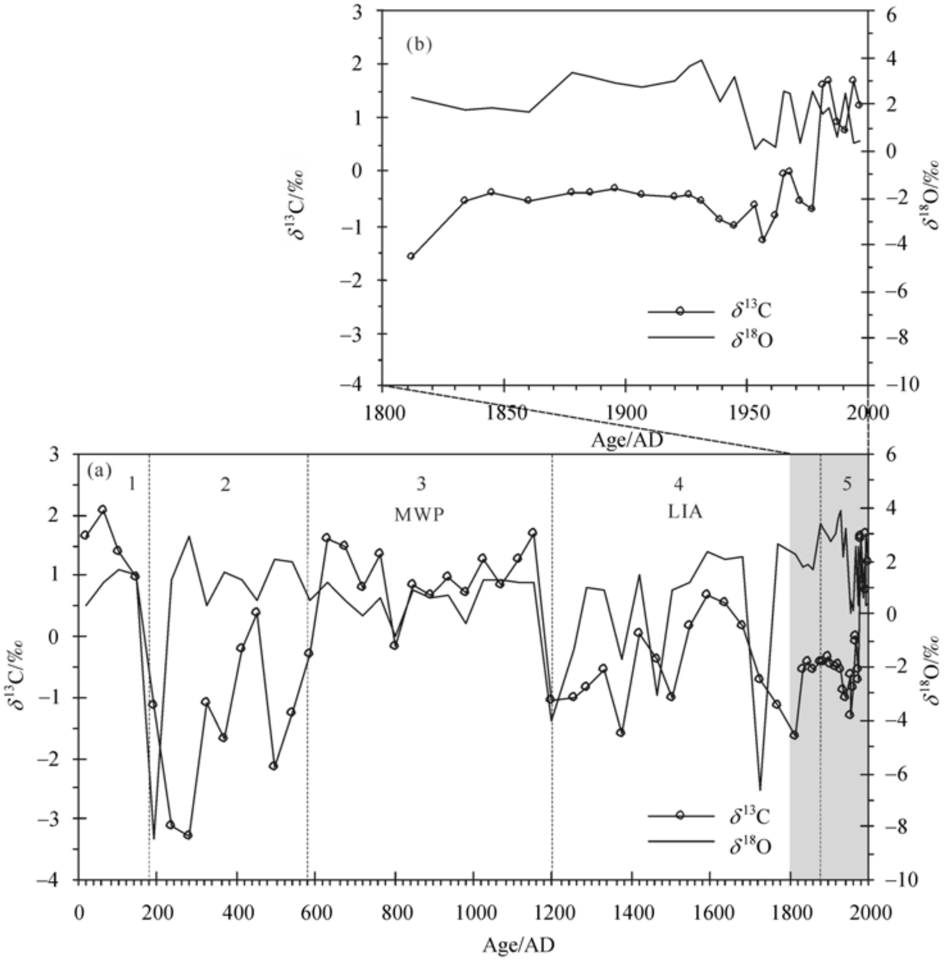

Fig. 3. Variations of stable isotopic values of carbonate in Lake Sugan since 2 kaBP (a) and during the period of 1800—2000 AD with a resolution of 9—11a (b). The numbers of 1—5 represent climatic change stages, and vertical dotted line indicates the boundary of climatic change stage.

–8.47‰ and 3.90‰, with a mean value of 0.99‰. On the whole, the oxygen isotopic values fluctuated around its mean value, and five remarkable negative values occurred in 190 AD, 1200 AD, 1370 AD, 1460 AD and 1720 AD. The carbon isotopic values vary between –3.27‰ and 2.06‰, with a mean value of –0.08‰. The variations of δ^{13}C values can be divided into five stages, and its high values occurred in 0—190 AD, 580—1200 AD and 1960—2000 AD and its low values occurred in 190—580 AD and 1200—1960 AD. There are some secondary fluctuations in the two low value stages. The higher carbon isotopic values occurred in 320 AD and 450 AD during the period of 190—580 AD, and they are –1.10‰ and 0.37‰, respectively. The higher values occurring in 1330 AD, 1420 AD and 1590 AD were –0.56‰, 0.06‰, and 0.69‰, respectively. The carbon isotopic values covary with that of oxygen isotope with a small variation amplitude during 580—1200 AD stage. Both of them became negative simultaneously around 800 AD, and became negative in phase during 580—800 AD and positive in phase during 800—1200 AD.

3 Discussion

3.1 Climatic implication of carbon and oxygen isotopes

Variations of oxygen isotope of authigenic carbonate in lakes mainly depend on δ^{18}O composition of the lake water and settling temperature of carbonate[27,28]. For the closed lakes in arid regions, δ^{18}O variations of authigenic carbonate mainly depend on the oxygen isotope composition of the lake water, and the variations of δ^{18}O values mainly reflect the ratio of precipitation to evaporation (P/E) in catchment[29−31]. Factors affecting the oxygen isotope of lake water include evaporation processes, water source of a lake (meltwater or precipitation changes due to the alteration of circulation pattern) and changes of precipitation season and lake level[32]. Lake Sugan basin is located in the hyperarid zone of China, hence, the precipitation amount or seasonal changes of precipitation in the basin have little influence on the isotopic composition of the lake. Lake Sugan is mainly fed by meltwater and the oxygen isotopic changes of the lake water are controlled by the ratio of inflow to evaporation (I/E). The δ^{18}O values of snowfall samples, spring samples and lake water samples exhibit a gradual increase tendency, suggesting the effect of evaporative enrichment in the basin on the oxygen isotopic fractionation. It can be seen that the changes in δ^{18}O values of carbonate in Lake Sugan mainly reflect the I/E ratio or effective moisture in the basin[33].

Temperature effect is little (0.0355℃$^{-1}$) in the sedimentation process of authigenic carbonate in lakes, and the fractionation effect of carbon isotope between the total dissolved inorganic carbon (TDIC) and the solid-phase carbonate is also small[34], hence the δ^{13}C changes of carbonate mainly depend on the carbon isotopic composition of TDIC[35]. The carbon isotopic composition of TDIC is mainly controlled by two factors. One is CO_2 production rate caused by local biological action; the other is the degree of exchange between lake water and atmospheric CO_2, with the latter as a leading factor[29,36]. As for the Manas region, Rhodes et al. suggested that when the δ^{13}C value is <-2‰, it perhaps records environmental processes, such as short residence time of lake water, longer annual freeze-up time and intense organism activity; when δ^{13}C value is >2‰, it reflects the equilibrium state with atmospheric CO_2, i.e. a longer residence time and/or well mixed lake water[29]. From the research of carbon isotope of ostracod shells in Lake Aibi since the late Pleistocene, Li Guosheng found two noticeable negative δ^{13}C values and thought of them as two events of sharp decrease in air temperature in the region[36]. In fact, the increase in freezing time of a lake would reduce the carbon isotope exchange between the lake and atmospheric CO_2, and then result in the negative δ^{13}C value of carbonate in the cold region[37]. The biological productivity is low in the study area, and the organic carbon transported to the lake seems unlikely to have an important role in changing the carbon isotopic composition of TDIC. The δ^{18}O variations shows that the oxygen isotope has been in a relatively stable state since 2 kaBP except for the several short negative intervals (Fig. 3), which means the lake inflow (meltwater transforming into groundwater) is relatively stable. Also, the Mg/Ca ratio of ostracod shell in the SG001227-C core has been relatively stable since 2 kaBP. That is to say, the salinity of the lake is relatively stable, and the lake productivity has little changes. The biomass in the lake might have little influence on the carbon isotopic composition of TDIC. Accordingly, the δ^{13}C changes of TDIC are related to the exchange degree with atmospheric CO_2. Lake Sugan is located in the northeastern part of the Tibetan Plateau with high elevation, its freeze-up period lasts from November to next March and the annual freeze-up duration can directly affect the exchange between the lake water and atmospheric CO_2. Therefore, the δ^{13}C variations of carbonate may indirectly indicate the air temperature changes in winter half year in the basin[33].

Mean monthly air temperature variations from November to March of next year since 1957, observed at the Lenghu Meteorology Station (60 km from Lake Sugan), can be divided into two stages: 1957—1980 and 1980—2000 (Fig. 4). The temperature in the latter was 0.7℃ higher than that of the former. Similarly, the variations of δ^{13}C can also be divided into the two stages (Fig. 4). Despite a low resolution of carbon isotopic record, it has a corresponding relation with the mean monthly temperature in winter half year in the variation tendency. Considering that each carbon isotope sample represents the climatic information of 3—5 years, the mean air temperature corresponding to the δ^{13}C value in the air temperature variation sequence was extrapolated forward and backward for two years, respectively, i.e. five-year mean value as the temperature of the interval, and then a correlation analysis with the carbon isotope sequence was made. They have a better positive correlation (r=0.64), which supports the explanation on the climatic implication of carbon isotope.

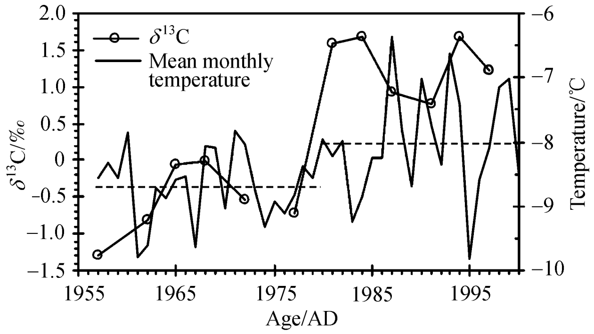

Fig. 4. Comparison of the δ^{13}C values of Lake Sugan with the mean monthly air temperature from November to next March observed at the Lenghu Meteorology Station since 1957. Dotted line represents average value of the mean monthly air temperature in winter half year during 1957—1980 and 1980—2000 periods[33].

The carbon and oxygen isotopes exhibit a noticeable covariance during the period of 580—1200 AD (Fig. 3). The covariance of δ^{13}C and δ^{18}O in a closed lake mainly reflects the changes of lake hydrologic process, exchange of CO_2 gas, lake productivity and total CO_2 concentration[38,39]. In the case of intense evaporation and longer residence time for the lake water, ^{18}O will be enriched in lake water. Meanwhile, the condensation of lake water due to evaporation and lake water warming results in release of CO_2 from the lake water and increase in vertical mixing degree of the lake water, and leads to the improvement of lake productivity, which causes ^{13}C enrichment of TDIC.

The covariance of carbon and oxygen isotopes in 580—1200 AD interval reflects intense evaporation in the basin.

As described above, the stable isotopes of carbonate in Lake Sugan can reveal the climatic variations of the region. The changes of oxygen isotope indicate the effective moisture changes of the basin, while the carbon isotope reflects the freeze-up time of the lake and indirectly reflects the air temperature variation in the winter half year or even variation of air temperature on greater timescale in the region.

3.2 Climatic changes since 2 kaBP

The climatic changes recorded by the stable isotopes of sedimentary carbonate in Lake Sugan can be divided into five stages (Fig. 3):

Stage 1 (0—190 AD): Carbon and oxygen isotopes had high values. A dry and warm climate prevailed in the study area. The effective moisture in the basin was low and air temperature was relatively high in winter half year.

Stage 2 (190—580 AD): Oxygen isotope had a significant negative value in 190 AD, but it lasted a short time. The δ^{18}O value was high in other time. This shows that the effective moisture increased around 190 AD while the climate was dry in other time intervals. A noticeable negative δ^{13}C value also occurred around 190 AD, but the δ^{13}C reached the lowest value around 280 AD. They increased around 320 AD and 450 AD but were lower than those of stage 1. A negative δ^{13}C value occurred around 500 AD again. On the whole, the δ^{13}C indicated the cold climate but the air temperature rose during 320—450 AD period. In this stage, a dry and cold climate prevailed.

Stage 3 (580—1200 AD): Carbon and oxygen isotopes have a covariance in their changes, which indicates a dry and warm climate. The values of them became negative around 800 AD, indicating a weak cold-wet climate events. The carbon and oxygen isotopes variation trend reflects that the climate was warmer and drier during 800—1200 AD than during 580—800 AD. The stage corresponds to the Medieval Warm Period (MWP).

Stage 4 (1200—1880 AD): Four negative values of oxygen isotope suggest the discontinuity increase in effective moisture in the region. On the whole, carbon isotope in this stage generally showed the negative values and indicated cold winter. The stage with cold-wet climate corresponds to the Little Ice Age (LIA), during which there were three climatic fluctuations in the stage and the coldest climate occurred in 1200 AD, 1370 AD, 1500 AD and 1810 AD.

Stage 5 (1880—2000 AD): The δ^{18}O had a high value but the δ^{13}C had a low value and changed with a small amplitude, which indicates a cold and dry climate during 1880—1950 AD period. In the 1950s, carbon and oxygen isotopes had low values and then the δ^{18}O frequently fluctuated, suggesting the fluctuations of the effective moisture in the region. After the 1960s, the δ^{13}C value rose dramatically, reflecting the fact of winter warming in the 20th century[9].

The climatic change pattern in the study area roughly exhibited a warm-dry and cold-wet assemblage, which may be related to the fact that the Sugan Lake basin is located in the hyperarid zone of China, the annual potential evaporation is much larger than the precipitation, and only under the temperature drop condition can the effective moisture increase. The wet events recorded by oxygen isotope of carbonate in Lake Sugan in 190 AD and LIA are all associated with the sharp decreases in temperature. The warm-dry event during the period of 0—190 AD should be included in the warm period during the West and East Han Dynasties of China (206 BC—220 AD). Through the combined research of ice cores, tree rings, fluviolacustrine sediments, lakeshore terraces, paleosols and literature records, Yang et al.[40] suggested that the climate in the arid northwest China was warm and wet at that time, which is different from the dry climate reflected by the δ^{18}O of Lake Sugan. The Medieval Warm Period recorded in Lake Huguangyan[41] includes two stages: 670—760 AD and 880—1260 AD, when the climate was dry. This is basically in agreement with the climatic changes recorded in Lake Sugan. In addition, the records of Guliya ice core in China[42], Quelccaya ice core in Peru[42] and Lake Moon[43] in the Great Plain of North America also indicate a dry climate in the Medieval Warm Period. The wet climate in the Little Ice Age has been confirmed by the records of Guliya ice core[1], Lake Qinghai[6] and Lake Huguangyan[41]. The warming of the 20th century indicated by the carbon isotope of Lake Sugan has also been found in Lake Qinghai[6]. The warming since 1980 not only has been demonstrated by the observation data at the local meteorology station but also coincides with the records in the whole northwest China[44].

Yang et al.[45] synthesized the palaeoclimatic records such as lake sediments, tree rings, ice cores and literatures, and obtained the regional area-weighted temperature changes in China since 2 kaBP (Fig. 5). The major cold periods occurred in 240—600 AD and 1400—1920 AD, while the warm periods occurred in 0—240 AD, 600—1400 AD and 1920 AD—present. The climate experienced secondary fluctuations on decade to century timescale. Basing on the records of historical literatures, Ge et al.[9] reconstructed the temperature anomaly variation sequence in winter half year since 2 kaBP in East China, with a resolution of 10—30a (Fig. 5). The warm periods occurred in 0s—200s AD, 570s—770s AD, 930s—1310s AD, 1920s AD—present, while the cold periods occurred in 210s—560s AD, 780s—920s AD, 1320s—1910s AD. The δ^{13}C of Lake Sugan clearly recorded the 0—190 AD warm period, 190—580 AD cold period, MWP, LIA and warming in the 20th century, and has a better comparability with the two records or even in detail (Fig. 5). A climatic change pattern of two troughs and one peak occurred during the 3rd—6th centuries for all of the three records. The temperature in the weighted sequence and the sequence of winter half year sharply decreased during 1230—1250 AD and around 1180 AD, respectively, which is consistent with the temperature decrease at the beginning of LIA as reflected by the δ^{13}C of Lake Sugan. The average value of temperature anomaly in the mean temperature variation sequence of winter half year in East China during 210s—560s AD was lower than that of LIA[9]. It is consistent with the fact that the δ^{13}C value during 190—580 AD was lower than that of the period of 1200—1880 AD. In addition, the three records all display the climatic warming trend in the 20th century (Fig. 5). However, the climatic change in MWP recorded by carbon isotope of Lake Sugan was more stable, and there was difference in the start and termination time of climatic stages recorded by stable isotopes and other two records. For example, the initial time of LIA in the stable isotope record was 1200 AD with 100—200 years earlier than that in East China and the whole China. The presence of the difference may be related to resolution of the isotopic record of Lake Sugan and responses of different climatic proxies or different regions to climatic changes.

The characteristics of the climate evolution phase in China during the past 2000 years obtained from the analysis of climatic information recorded in historical literatures demonstrate that the abrupt climatic changes occurring in the 280s AD and 1230s AD were characterized by temperature decreases[10], which is in agreement with the temperature decrease reflected by the δ^{13}C of Lake Sugan around 280 AD and 1200 AD. Such phasic characteristics also show that the climate became warm around 360s AD and became cold after 490s AD, and it gradually turned warm during 560—580 AD and lasted until 1230s AD. It is similar to the climate changes reflected by δ^{13}C. It can be seen that the phasic characteristics of the climate in China since 2 kaBP also support the history of climatic changes inferred from the carbon isotope as described in this paper.

In a word, the stable isotopes of Lake Sugan can record the climatic changes in the characteristic periods since 2 kaBP, such as 3rd—6th centuries, MWP, LIA and warming in the 20th century. Furthermore, the climatic changes and its abrupt events also coincide with other records, especially air temperature variations reflected by the carbon isotope.

4 Conlcusion

Stable isotopes of carbonate in Lake Sugan better recorded the climatic changes. The oxygen isotope indicates the effective moisture in the region, and the carbon isotope indicates the freeze-up duration of the lake and further reveals the air temperature variations in winter half year. The climatic changes in the study area have experienced five stages since 2 kaBP: 0—190 AD, warm-dry climate; 190—580 AD, cold-dry climate; 580—1200 AD,

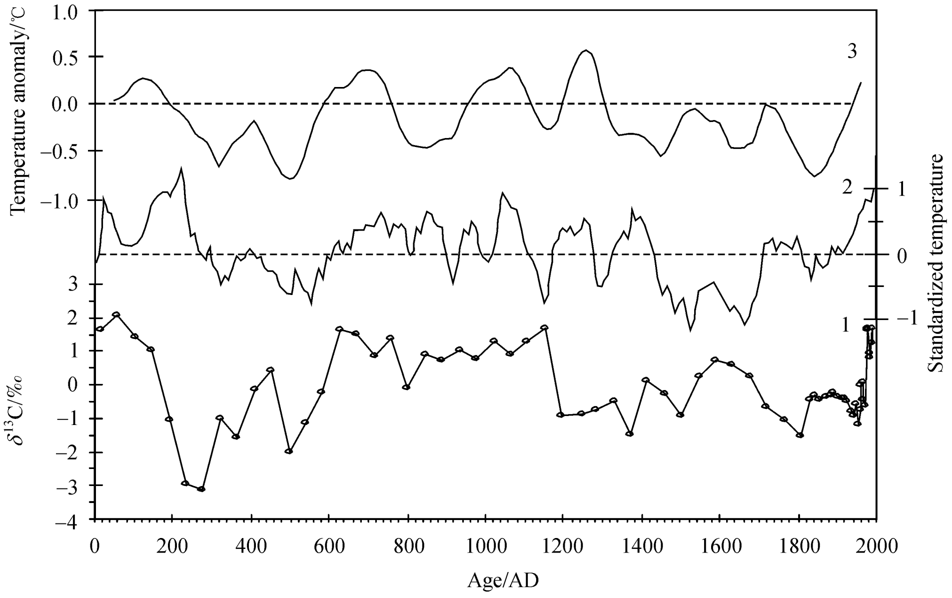

Fig. 5. Comparison of the δ^{13}C value variations of Lake Sugan since 2 kaBP (curve 1) with regional area-weighted temperature curve in China[45] (curve 2) and mean temperature variation sequence in winter half year in East China[9] (curve 3). Curve 3 is a 3-point moving average curve.

warm-dry climate, corresponding to the Medieval Warm Period; 1200—1880 AD, cold-wet climate, corresponding to the Little Ice Age; 1880—1950 AD, cold-dry climate; and from 1950s onwards, climate warming, which is consistent with the observation data. The climatic changes history recorded by stable isotopes of Lake Sugan since 2 kaBP, especially the climatic cooling and warming variations revealed by carbon isotope, not only has better comparability with other records, but also coincides with the temperature variations recorded by historical literatures of China. It can be seen that the climatic changes recorded by stable isotopes since 2 kaBP are of universal significance, and it means that the research of palaeoclimatic changes in the lake still has a great potential. However, further research is needed. The accurate ages of the lake sediments need to be further determined and an attempt is being made to determine the ages of authigenic carbonate in the sediment using the thermal ionization mass spectrometric technique (TIMS) so as to check and correct the reservoir effect of the lake. In addition, the carbonate mineral composition of the lake and the fractionation mechanism of carbon isotope in the atmospheric CO_2-lake water-carbonate system also need to be further understood.

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References

1. Yao, T. D., Climatic and environmental changes documented by Guliya ice cores over the past ~2000 years, Quaternary Sciences (in Chinese), 1997, 17(1): 52—59.
2. Yao, T. D., Qin, D. H., Tian, L. D. et al., Variations in temperature and precipitation in the past 2000 a on the Xizang (Tibet) Plateau——Guliya ice core record, Science in China, Ser. D, 1996, 39(4): 425—433.
3. Yao, T. D., Thompson, L. G., Dunde ice core records and temperature variation over the past 5000 years, Science in China, Ser. B (in Chinese), 1992, 22(10): 1089—1093.
4. Kang, X. C., Graumlich, L. J., Sheppard, P., The last years climate changes inferred from tree ring records in Dulan region, Qinghai, China, Quaternary Sciences (in Chinese), 1997, 17(1): 70—75.
5. Wang, S. M., Xue, B., Xia, W. L., Climatic changes documented by Ximen Co sediments since 2 kaBP, Quaternary Sciences (in Chinese), 1997, 17(1): 62—67.
6. Shen, J., Zhang, Z. L., Xia, W. L., Records from lakes sediments of the Qinghai lake to mirror climatic and environmental changes of the past about 1000 years, Quaternary Sciences (in Chinese), 2001, 21(6): 508—513.
7. Wang, S. W., Reconstruction of temperature series of North China from 1380s to 1980s, Science in China, Ser. B, 1991, 34(6): 751—759.
8. Wang, S. W., Climate of the Little Ice Age, Quaternary Sciences (in Chinese), 1995, 15(3): 202—211.
9. Ge, Q. S., Zheng, J. Y., Fang, X. Q., Winter temperature changes in East China since 2 kaBP, Quaternary Sciences (in Chinese), 2002, 22(2): 166—173.
10. Zhang, P. Y., Wang, Z., Liu, X. L. et al., Phasic characteristics of climatic changes since 2 kaBP in China, Science in China, Ser. B (in Chinese), 1994, 24(9): 998—1006.
11. Zhang, D. E., Winter temperature changes during the last 500 years in South China, Chinese Science Bulletin, 1980, 25(6): 497—500.
12. Zhang, P. X., Zhang, B. Z., A preliminary study on evolutions of climate and environment in Qaidam Basin during the last 3Ma years, Acta Geographica Sinica (in Chinese), 1991, 46(3): 327—335.
13. Shen, Z. Q., Tong, G. B., Zhang, J. P. et al., Geological environments since Pliocene and accumulation process of saline deposit in west Qaidam Basin, Qinghai, China, Marine Geology & Quaternary Geology (in Chinese), 1990, 10(4): 89—99.
14. Huang, Q., Sun, N. J., Meng, Z. Q. et al., The study on features of palaeoclimatic variation of dry and cold areas (Ⅱ)——preliminary study on fluctuation model of paleoclimate of Qarhan salt region in Qaidam Basin, Science in China, Ser. B, 1991, 34(3): 338—351.
15. Chen, K. Z., Bowler, J. M., Preliminary study on sedimentation characteristics and evolution of palaeoclimate of Qarhan Salt Lake in Qaidam Basin, Science in China, Ser. B, 1985, 28(11): 1218—1231.
16. Yang, W. B., Spencer, R. G., Krouse, H. R. et al., Stable isotopes of lake and fluid inclusion brines, Dabusun Lake, Qaidam Basin, western China: Hydrology and paleoclimatology in arid environments, Palaeogeography, Palaeoclimatology, Paleoecology, 1995, 117: 279—290.
17. Phillips, F. M., Zreda, M. G., Ku, T. L. et al., ^{230}Th/^{234}U and ^{36}Cl dating of evaporite deposits from western Qaidam Basin, China: Implications for glacial-period dust export from Central Asia, Geological Society of America Bulletin, 1993, 105: 1606—1616.
18. Liu, S. Q., Zhang, F. S., Sedimentation Environment and Salination in Mahai Sub-basin in Qaidam Basin (in Chinese), Beijing: Geology Press, 1992, 9—12.
19. Deckker, P. D., Last, W. M., Modern dolomite deposition in continental, saline lakes, western Victoria, Australia, Geology, 1988, 16:

29－32.

20. McCrea, J. M., On the isotope chemistry of carbonates and palaeotemperature scale, Journal of Chemical Physics, 1950, 18: 849－857.
21. Wan, G. J., ^{210}Pb Chronology of modern sediments, Quaternary Sciences (in Chinese), 1997, 17(3): 230－239.
22. Cheng, Z. Y., Liang, Z. C., Lin, R. F. et al., ^{210}Pb age of modern sediments calculated by the CF model for Lake Dianchi, Yunnan Province, Chinese Journal of Geochemistry (in Chinese), 1990, 4: 327－332.
23. Zhu, L. P., Chen, L., Li, B. Y. et al., Environmental changes reflected by the lake sediments of the South Hongshan Lake in Northwest Tibet, Science in China, Ser. D, 2002, 45(5): 430－439.
24. Xiang, L., Limitations in application of ^{137}Cs chronology to lacustrine sediments: A case study on Lake Crawford, Journal of Lake Sciences (in Chinese), 1995, 7(4): 307－313.
25. Delectis Florae Reipublicae Popularis Sinicae, Agendae Academiae Sinicae Edita, Flora Reipublicae Popularis Sinicae (Tomus 8) (in Chinese), Beijing: Science Press, 1992, 83－85.
26. Stuiver, M., Reimer, P. J., Braziunas, T. F., High precision radiocarbon age calibration for terrestrial and marine samples, Radiocarbon, 1998, 40: 1041－1083.
27. Craig, H., The measurement of oxygen isotope palaeotemperature, Stable Isotopes in Oceanographic Studies and Palaeotemperature (ed. Tongiorgi, E.), Pisa: Cons. Naz. Rich. Lab. Geo. Nucl., 1965, 161－182.
28. Stuiver, M., Oxygen and carbon isotope ratios of freshwater carbonate as climatic indicators, Journal of Geophysics Research, 1970, 75: 5247－5257.
29. Rhodes, T. E., Gasse, F., Lin, R. F. et al., A late Pleistocene－Holocene lacustrine record from Lake Manas, Zunggar (northern Xinjang, Western China), Palaeogeography, Palaeoclimate, Palaeoecology, 1996, 120: 105－121.
30. Lister, G. S., Kelts, K., Chen, K. Z. et al., Lake Qinghai, China: closed-basin lake levels and the oxygen isotope record for ostracoda since the latest Pleistocene, Palaeogeography, Palaeoclimatology, Paleoecology, 1991, 84: 141－162.
31. Leng, M. J., Marshall, J. D., Palaeoclimate interpretation of stable data from lake sediment archives, Quaternary Science Reviews, 2004, 23: 811－831.[DOI]
32. Fontes, J. C., Gasse, F., Gibert, E., Holocene environmental changes in Lake Bangong basin (western Tibet), Part 1: Chronology and stable isotopes of carbonates of Holocene lacustrine core, Palaeogeography, Palaeoclimateology, Palaeoecology, 1996, 120: 25－47.
33. Qiang, M. R., Chen, F. H., Gao, S. Y., A preliminary study on the climatic implications of carbonate stable isotopes from Sugan lake sediments, Quaternary Sciences (in Chinese), 2004, 24(5): 598－599.
34. Romanek, C. S., Grossman, E. L., Morse, J. W., Carbon isotopes fractionation in synthetic aragonite and calcite: Effects of temperature and precipitation rate, Geochimica et Cosmochima Acta, 1992, 56: 419－430. [DOI]
35. Gasse, F., Fontes, J. C., Plaziat, J. C. et al., Biological remains, geochemistry and stable isotopes for the reconstruction of environmental and hydrological changes in the Holocene lakes from North Sahara, Palaeogeography, Palaeoclimatology, Palaeoecology, 1987, 60: 1－46.
36. Li, G. S., On the δ^{13}C record and abrupt climate events since the last Deglaciation in Lake Aibi, Chinese Science Bulletin, 1994, 39(4): 302－306.
37. Gasse, F., Arnold, M., Fontes, J. C. et al., A 13,000-year climate record from western Tibet, Nature, 1991, 353: 742－745. [DOI]
38. Li, H. C., Ku, T. L., δ^{13}C－δ^{18}O covariance as a paleohydrological indicator for closed-basin lakes, Palaeogeography, Palaeoclimatology, Palaeoecology, 1997, 133: 69－80.
39. Li, H. C., Bischoff, J. L., Ku, T. L. et al., Climate variability in eastern-central California during the past 1000 years reflected by high-resolution geochemical and isotopic records from Owens Lake sediments, Quaternary Research, 2000, 54: 189－197. [DOI]
40. Yang, B., Braeuning, A., Shi, Y. F. et al., Evidences for a late Holocene warm and humid climate period and environmental characteristics in the arid zones of northwest China during 2.2－1.8 kyr B.P., Journal of Geophysical Research, 2004, 109(D2), doi: 10.1029/2003JD003787.
41. Chu, G. Q., Liu, J. Q., Qing, S. et al., The ‘Mediaeval Warm Period’ drought recorded in Lake Huguangyan, tropical South China, The Holocene, 2002, 12: 511－516. [DOI]
42. Thompson, L. G., Climatic changes for the Last 2000 years inferred from ice-core evidence in tropical ice cores, Climatic Variations and Forcing Mechanisms of the Last 2000 Years (eds. Jones, P. D., Bradley, R. S., Jouzel, J.), Berlin: Springer-Verlag, 1996, 281－296.
43. Laird, K. R., Frits, S. C., Maasch, K. A. et al., Greater drought intensity and frequency before AD 1200 in the Northern Great Plain, USA, Nature, 1996, 384: 552－554. [DOI]
44. Shi, Y. F., An Assessment of the Issues of Climatic Shift from Warm-dry to Warm-wet in Northwest China (in Chinese), Beijing: Meteorology Press, 2003. 18－20.
45. Yang, B., Braeuning, A., Johnson, K. R. et al., General characteristics of temperature variation in China during the last two millennia, Geophysical Research Letters, 2002, 29 (9), doi: 10.1029/2001 GL014485.

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