1   **Oxygen isotopic composition** of bulk carbonates in recent sediments from Lake Kuhai (**NW China**)
2   and **implications for** hydroclimatic **changes** in headwater areas of the Yellow River on the
3   northeastern Tibetan Plateau

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Abstract

The water resource issue is one of the most significant problems in the Yellow River basin, and has received much attention from the public and government because of the dramatically decreased Yellow River streamflow in the last several years. In this study, oxygen isotopic compositions of bulk carbonates from Lake Kuhai in the headwaters of the Yellow River were evaluated as an indirect proxy of past Yellow River streamflow in order to place the recent flow reduction in a long-term context. The results indicated that $\delta^{18}\text{O}$ values of bulk carbonates from core KHC14-1 generally vary with changes in the Yellow River streamflows related to the precipitation/evaporation (temperature) ratio above the Tangnaihai hydrological station over the past 50 years. In general, the Yellow River streamflow in the headwaters area continued to decrease from the early 1980s to the late 1990s because of decreased precipitation and increased temperature. Then, the streamflow increased with enhancing precipitation over the last two decades. In addition, $\delta^{18}\text{O}$ values of bulk carbonates in the Lake Kuhai core roughly correlate with the streamflows of the upper reaches of the Yellow River recorded by tree ring width over the past 800 years. The enriched $\delta^{18}\text{O}_{\text{BC}}$ values showed that the precipitation/evaporation (P/E) ratio or streamflow was very low and the climate might be very dry in the middle of the 1400s AD in the headwaters of the Yellow River over the past 800 years. Our results suggested that changes in the P/E ratio or streamflow of the upper reaches of the Yellow River were dominated by variations in Indian Summer Monsoon precipitation over a long time scale, consistent with other records from the northeastern Tibetan Plateau.

Keywords: oxygen isotope, bulk carbonates, streamflows, Yellow River, Tibetan Plateau

1. Introduction

As one of the major water resources for a large part of China, the Yellow River runoff supplies 15% of the water for agricultural production and furnishes 12% of the water used by the country's population (Zhu and Li, 2003). Nearly 40% of the runoff of the Yellow River comes from areas of the arid and semi-arid upper reaches above the Tangnaihai hydrological station on the northeastern Tibetan Plateau (Mu et al., 2003). Therefore, variations in the runoff from this drainage area have great impact on the downstream discharge and water supplies (Kang et al., 2005; Fu et al., 2007). Drying up episodes of the Yellow River would lead to severe ecological crises such as widely degraded grassland and frequent sandstorms in the headwater area of the Yellow River. In this case, it is very important to explore the mechanism of dynamic variations of streamflow in the region, for the purposes of management and the ecological sustainable development in the source region of the Yellow River.

Studies based on the data from hydrological stations and weather stations have shown that variations in the streamflow from the upper reaches of the Yellow River in the current period are mainly influenced by climate change (precipitation and/or temperature) instead of human activities (Li et al., 1998; Huang et al., 2006; Wang and Li, 2013; Li et al., 2014). However, the natural stream gauge records from hydrological stations only exert limited ability to predict long-term streamflow history (Gou et al., 2007, 2010; Pan et al., 2013). Alternatively, information on the history, regularity, and mechanisms of hydrologic variability over centennial/millennial time scales would help estimate the recently observed changes from a long-term perspective.

Climatic or hydroclimatic changes of the upper reaches of the Yellow River are still poorly investigated on both short and long time scales, since sporadic studies on tree rings and lacustrine sedimentary materials around the region (e.g., Ma et al., 2004; Gou et al., 2007, 2010; Mischke et al., 2008, 2010; Herzsuh et al., 2009; Zhang and Mischke, 2009; Aichner et al., 2012; Pu et al., 2013). For the Holocene, a study from Lake Donggi Cona on the upper reaches of the Yellow River showed that the higher lake level occur at the Holocene Climate Optimum between 11.5 and 6.5 cal kyr BP (Opitz et al., 2012). However, the moisture availability recorded by ostracods and their geochemical compositions from Lake Koucha in the upper reaches of Yellow River since the late glacial period is different with many paleoclimate records available on the Tibetan Plateau, probably suggesting its regional characteristics (Mischke et al., 2008). Inconsistency between the

reduced early Holocene effective moisture in the Lake Koucha region and the higher monsoonal precipitation suggested by other records from the Tibetan Plateau simply implies an enhanced evaporation caused by warm temperature over the rise of monsoonal precipitation in the period (He et al., 2014). As for the last millennial, much less studies have been conducted so far. For example, Gou et al. (2007, 2010) reconstructed the streamflow history of the Yellow River over the past 593 and 1234 years using principal component analysis on six tree ring width profiles from the headwater areas of the Yellow River. Therefore, more works, especially for the lacustrine investigations, should be carried out to better evaluate the recently observed reduction in the Yellow river discharge in response to present global warming and the future changes in the discharge in response to future warmer climatic conditions from a centennial-scale perspective.

In this study, we present oxygen isotopic compositions of bulk carbonates predominantly laminated lake sediments covering the past several hundred years from the saline closed-basin Lake Kuhai in the headwaters of the Yellow River. This proxy was evaluated as an indirect proxy of the past Yellow River streamflow changes. Our objective is to investigate the historical natural variability of the river discharge in centennial scales, and probe the main controller on this.

2. Materials and methods

2.1 Study area

Lake Kuhai (35.32°N, 99.21°E, 4150 m asl) is located in the eastern Tibetan Plateau, with an area of approximately 45 km² and a maximum water depth of 21 m (Fig. 1) (Mischke et al., 2010). The lake is aligned NW-SE and NE-SW, corresponding to the large Kunlun Fault system which moves in the NW-SE direction toward a location approximately 20 km to the southwest of the lake (Van der Woerd et al., 2002). So, the lake is mainly supplied by the river water, meltwater and groundwater around the foothills of the A'nyêmaqên Mountains, which are also the main runoff-generating areas of the upper reaches of the Yellow River (Gou et al., 2007; Mischke et al., 2010). Lake Kuhai is now brackish/saline alkaline lake, with a salinity of 20.4 g/L, a high Mg/Ca ratio of 31.1, and a pH of 8.9-9.1, due to the intensive evaporation over precipitation in the area (Mischke et al., 2010; Zhou et al., 2010).

The climate in the Lake Kuhai region has been suggested to be dominated by the Asian monsoon system (Mischke et al., 2010). Regional instrumental data collected by the Maduo

Weather Station (98.22°E, 34.91°N, 4273 m asl) over the last 50 years reveals strong seasonal contrasts with moderately warm and wet summers and cold and dry winters ($T_{\text{July}}=7.8^{\circ}\text{C}$, $T_{\text{ann}}=-4.0^{\circ}\text{C}$, $P_{\text{ann}}=321$ mm). The lake remains covered by ice from late October to early May because of the low temperature in this region (Mischke et al., 2010). The mean annual potential evaporation is probably of the order of about 1000 mm. The lake itself is densely populated by aquatic macrophytes, primarily *Potamogeton pectinatus*, to a depth of 7 m according to our observations in the field. Alpine meadows and steppes with *Kobresia*, *Poaceae* and *Artemisia* are dominant in the catchment of Lake Kuhai (Mischke et al., 2010).

2.2 Sediment sampling

A 56-cm-long core (KHC14-1) was drilled in the northern part of the lake (34.0108°N and 97.2422°E, water depth of 7.8 m) using Livingstone coring equipment in the late September of 2014. The core consisted of sandy marl with some submerged plant fragments belonging to the *Potamogeton*. The core did not show finely laminated sediments as described by Mischke et al. (2010). Obvious lithostratigraphic difference was observed between the upper parts of all the cores (up 54 cm) and the deeper layers (bottom 2 cm). In order to avoid unnecessary sample contamination, the bottom 2 cm of the core was discarded and the top 54 cm of the core was sub-sampled at continuous 0.5 cm intervals for further analysis.

2.3 Experimental analysis

Radioactivity was measured by a germanium detector manufactured by AMETEK Company at the Institute of Polar Environment, USTC, Hefei, China. The samples were dried to a constant weight at a temperature of 50°C, homogenized using a mortar and pestle, and passed through a 120-mesh sieve. Samples were then packed into standard counting geometries for gamma spectrometry and sealed and stored for about one week to allow radioactive equilibration between ^{226}Ra and its daughter product ^{214}Pb . Spectra were continuously measured for 24 h to obtain enough counts. The resulting spectrum files showed ^{210}Pb activity with a peak at 46.5 keV. In addition, after the standard acid-alkali-acid procedure based on the laboratory's analytic standard pretreatment protocols, two macrophyte remains (*Potamogeton*), selected at approximately 11 cm and 54 cm depths of core KHC14-1, were measured for radiocarbon dating by the acceleration mass spectrometry (AMS) at the Xi'an AMS Center, China (Table 1).

Approximately 0.5 g of sediments from the core were pretreated according to the method of

Lu and An (1997) and were further analyzed by a Malvern Mastersizer 2000 for grain size property. Basically, the measurement range is 0.02-2,000 μm and the measurement error based on the repeated samples is less than 2%. Approximately 1.0 g of each sediment sample was leached with deionized water to isolate calcium carbonate from those of the pore water and water-soluble carbonates. After freeze-dried, the samples were then ground in an agate mortar, sieved through a 200 mesh screen, and homogenized for bulk carbonates content and their isotopic measurements ($\delta^{13}\text{C}_{\text{BC}}$ and $\delta^{18}\text{O}_{\text{BC}}$). The bulk carbonates content was determined by using the neutralization-titration method (Hesse, 1971). Approximately 0.3 g of a sediment sample was reacted with 30 ml HCl (0.2 mol/L) for 24 hours and then centrifuged. Ten milliliters of supernatant with 2 drops of phenolphthalein solution were titrated with NaOH (0.1 mol/L), resulting in an analytical precision better than 0.5%.

All bulk carbonate samples were analyzed for the $^{13}\text{C}/^{12}\text{C}$ and $^{18}\text{O}/^{16}\text{O}$ content 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). The results are expressed in delta (δ) notation relative to the V-PDB standard. Repeated analyses of laboratory standard carbonates (TTB1) with known $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values were performed daily to ensure instrumental accuracy. The analytical errors of the laboratory standard were approximately better than $\pm 0.1\%$ for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$.

3. Results

3.1. Chronology

The excess ^{210}Pb activity showed a simple exponential relationship with depth (Fig. 2). Accordingly, the chronology of the upper 17 cm of the core was first determined by the ^{210}Pb - ^{137}Cs dating method (Fig. 2), using a Constant Initial Concentration (CIC) computer model (Appleby, 2001). From the ^{210}Pb profile, the age at 8.5 cm with peaked ^{137}Cs was determined to be AD 1970, which is very close to the presumed ^{137}Cs peak age of AD 1963. The average sedimentation rate was then calculated to be 0.15 cm/year for the upper 17 cm of the core, which is much higher than the earlier result (Mischke et al., 2010).

AMS ^{14}C ages are 2018 ± 28 BP and 2991 ± 32 BP at depths of 11 cm and 54 cm, respectively (Table 1), older than the calculated age based on the ^{210}Pb and ^{137}Cs measurements in the same core, suggesting a apparent reservoir effect in the lake. By calculating the difference

between the AMS ^{14}C age and ^{210}Pb age at 11 cm, the reservoir effect in the lake is determined as 2018 years. The reservoir-corrected radiocarbon ages of the samples were converted to calendar ages (Table 1) using the software Calib702 (Reimer et al., 2009). The age-depth profile below 17 cm was established using a linear interpolation between the depths of 54 cm and 11 cm. In general, core KHC14-1 dates to approximately AD 1170 based on these two age models in Lake Kuhai.

3.2. Grain size, bulk carbonates content, $\delta^{13}\text{C}_{\text{BC}}$, and $\delta^{18}\text{O}_{\text{BC}}$

$\delta^{13}\text{C}_{\text{BC}}$ varies between -2.24‰ and 3.50‰, with an average of 0.71‰ (Table 2, Fig. 3). $\delta^{18}\text{O}_{\text{BC}}$ varies between -8.32‰ and 1.60‰, with an average of -2.94‰ (Table 2, Fig. 3). $\delta^{18}\text{O}_{\text{BC}}$ shows no linear changing trend but varies with fluctuation (Fig. 3). In general, $\delta^{13}\text{C}_{\text{BC}}$ co-varies with $\delta^{18}\text{O}_{\text{BC}}$ well ($r=0.96$, $p<0.01$), indicating a general feature of the hydrological-closed lakes (Talbot, 1990; Li and Ku, 1997). Bulk carbonates content is significantly correlated with $\delta^{13}\text{C}_{\text{BC}}$ and $\delta^{18}\text{O}_{\text{BC}}$. The correlation coefficients for bulk carbonates content vs. $\delta^{13}\text{C}_{\text{BC}}$ and bulk carbonates content vs. $\delta^{18}\text{O}_{\text{BC}}$ are 0.85 ($p<0.01$) and 0.88 ($p<0.01$), respectively (Fig. 3).

The percentage of $>63\ \mu\text{m}$ grain size varies from 0 to 47%, with an average of 22% (Fig. 3). In general, the coarse grain ($> 63\ \mu\text{m}$) content shows large variations. A large content of the coarse grains coincides well with higher oxygen isotope values, except that there is no obvious decrease in the grain size curve during the period of AD 1310-1450 (Fig. 3).

4. Discussion

4.1 Climatic significance of carbonate content and grain size

In general, lacustrine bulk carbonates mainly come from authigenic carbonate, detrital carbonate derived from the local bedrock as well as aeolian dust. The contribution of detrital carbonate from bedrock and aeolian dust may be very limited because the rocks are mainly identified as sillimanite-rich metamorphic quartz wacke around the Lake Kuhai area (Mischke et al., 2010). Therefore, the lacustrine bulk carbonates is mainly contributed by authigenic carbonate in Lake Kuhai, and the environmental change within the lake is the main controller on the $\delta^{18}\text{O}_{\text{BC}}$ and $\delta^{13}\text{C}_{\text{BC}}$ records. In addition, the precipitation rate of the chemically deposited carbonates in Lake Kuhai should be predominantly controlled by the salinity of the lake water as reported in other brackish lakes, with high salinity condition resulting in a high bulk carbonates content in lake sediments. In arid and semi-arid areas, the salinity of the lake water is primarily influenced by regional precipitation/evaporation (P/E) ratios (Liu et al., 2009), thereby an indirect relationship

between the P/E ratio and the bulk carbonate content may be expected. Coarse grained fractions in the sedimentary core from Lake Kuhai are composed of subangular grains of quartz and other minerals with sizes ranging from 50 to 200 μm , which is considered as a detrital component with an aeolian origin in Lake Kuhai (Mischke et al., 2010). Thus, content of the coarse grain ($>63 \mu\text{m}$) particles can be used as an indicator of dust storm events related to dry climatic condition (Qiang et al., 2007; Mischke et al., 2010).

4.2 The hydroclimatic implications of $\delta^{18}\text{O}_{\text{BC}}$ in the upper reaches of Yellow River

The oxygen isotope values of lacustrine carbonates are influenced by the temperature and, more importantly in closed lakes, the isotopic composition of the water when the carbonates precipitated (Leng and Marshall, 2004; Liu et al., 2009; Mischke et al., 2010). $\delta^{18}\text{O}$ of lake water is governed by $\delta^{18}\text{O}$ of the rainfall and by the evaporation in the arid area. In general, higher evaporation correlates with higher $\delta^{18}\text{O}$ of the lake water, and vice versa. The effect of water temperature on $\delta^{18}\text{O}$ values of carbonates is usually considered minimal with 0.24‰ decrease in the $\delta^{18}\text{O}$ value of carbonates for every 1°C increase (O'Neil et al., 1969). Therefore, the oxygen isotopic values of the sedimentary carbonates are assumed to be the indicator of changes in the regional effective moisture in arid and semi-arid northwestern China (Lister et al., 1991; Leng and Marshall, 2004; Qiang et al., 2005; Henderson et al., 2010). Thus, the $\delta^{18}\text{O}_{\text{BC}}$ was used to indicate variation in the effective precipitation (P/E) ratio in the Lake Kuhai area.

The water loss caused by evaporation related to high temperature mainly occurs in the summer season (June - August) because Lake Kuhai was covered by ice from late October to early May. In addition, the precipitation in the nearby foothills of the A'nyêmaqên Mountains mainly supplies Lake Kuhai. Therefore, we assume that the effective precipitation (P/E) ratio was mainly controlled by the balance between annual precipitation and the water loss through evaporation in summer seasons in Lake Kuhai. Past studies revealed that the streamflow is mainly controlled by changes in precipitation in the upper reaches of the Yellow River (Lan et al., 2004; Qiu, 2011; Wang and Li, 2013) but has also been affected by the evaporation related to increased temperature over the last 50 years in this region (Zhang et al., 2004; Kang et al., 2005; Wang and Li, 2013). Briefly, streamflows of the upper reaches of the Yellow River are mainly controlled by the balance between precipitation and evaporation related to temperature in this region. The results of a correlation analysis show that the mean annual runoff of the upper reaches of the Yellow River

above the Tangnaihai hydrological station is positively correlated with the precipitation/summer temperature ratio (representative of the P/E ratio) according to modern instrumental data over the last 50 years ($r=0.6$, $p<0.01$). In addition, the precipitation in the A'nyêmaqên Mountains mainly supplies both Lake Kuhai and the upper reaches of the Yellow River. Thus, we would like to determine whether the $\delta^{18}\text{O}_{\text{BC}}$ from Lake Kuhai can be indirectly used to trace variation in the P/E ratio or Yellow River streamflows above the Tangnaihai hydrological station in the long term in the upper reaches of the Yellow River.

To test this assumption, we compare the record from $\delta^{18}\text{O}$ values of the bulk carbonates from the core KHL14-1 with instrumental record of streamflows from the Yellow River above the Tangnaihai hydrological station over the last 50 years. The low $\delta^{18}\text{O}_{\text{BC}}$ values in core KHL14-1 correspond to a high streamflow during the period from AD 1960 to 1990. Then, the $\delta^{18}\text{O}_{\text{BC}}$ values became high and were accompanied by low runoff after AD 1990 (Fig. 5). However, the changes in $\delta^{18}\text{O}_{\text{BC}}$ values appear to lag behind the variation in runoff (Fig. 5), probably due to the lagging formation time of carbonate and the lagging sedimentary times of carbonate from surface water to sediments in Lake Kuhai. Therefore, $\delta^{18}\text{O}_{\text{BC}}$ values in Lake Kuhai may have recorded the changes in the P/E ratio or runoff of Yellow River above the Tangnaihai hydrological station over the past 50 years.

We next evaluate whether the $\delta^{18}\text{O}_{\text{BC}}$ value can be effectively used throughout core KHL14-1 by comparing the $\delta^{18}\text{O}_{\text{BC}}$ values in core KHL14-1 with tree-ring evidence for the streamflow change over the past ~800 years (Fig. 6) (Gou et al., 2007). In general, an increased effective precipitation would be expected to accompany a shift to less positive $\delta^{18}\text{O}_{\text{BC}}$ values in the lake carbonates, whereas a decreased effective precipitation coincides with the opposite in Lake Kuhai. Meanwhile, the increased effective precipitation also resulted in increased Yellow River streamflow above the Tangnaihai hydrological station. Therefore, the $\delta^{18}\text{O}_{\text{BC}}$ values would become high with decreased river runoff if the $\delta^{18}\text{O}_{\text{BC}}$ value has an indirect relationship with the Yellow River streamflow above the Tangnaihai hydrological station.

As shown in Figure 6, fluctuations in river runoff from AD 1170 to 2000 could potentially explain the changes in the KHL14-1 $\delta^{18}\text{O}_{\text{BC}}$ values during that period considering the uncertainty in our age model (Fig. 6). In general, the enriched $\delta^{18}\text{O}_{\text{BC}}$ values in core KHL14-1 correspond to the low Yellow River streamflow inferred from the tree ring data during the following periods:

1290s-1300s, 1470s-1500s, 1580s-1600s, 1640s, 1790s, 1820s-1840s, 1860s-1890s (Gou et al., 2007). Meanwhile, the increased and high content of bulk carbonates with a coarse grain size ($>63\ \mu\text{m}$) also indicated the low and decreased effective precipitation/river runoff related to dry climates during these periods (Fig. 3). In addition, the lowest runoff generally corresponds to the highest $\delta^{18}\text{O}_{\text{BC}}$ value in core KHC14-1 in the 1470s (Fig. 6). On the contrary, high river runoff periods were identified from tree rings that correspond to low $\delta^{18}\text{O}_{\text{BC}}$ values in core KHC14-1 for periods such as the 1250s, 1350s, 1510s-1570s, 1720s-1730s, 1750s-1760s, 1940s and 1970s (Fig. 6). Similarly, the bulk carbonate content and percentage of grain size $>63\ \mu\text{m}$ also decreased to low values, which may be related to humid climate conditions in this area (Fig. 3). Despite the age uncertainty, the comparison of the $\delta^{18}\text{O}_{\text{BC}}$ and the tree-ring reconstructed stream flow during the period of AD 1400~1500 is not as obvious as in other periods (Fig. 6), indicating these variables could change independently and either the tree-ring based stream flow is not reliable or the $\delta^{18}\text{O}_{\text{BC}}$ of carbonates was influenced by other factors. However, our results suggest that variation in the P/E ratio indicated by the $\delta^{18}\text{O}_{\text{BC}}$ values from Lake Kuhai was very similar with the change in the tree-ring reconstructed streamflows in the upper reaches of the Yellow River.

4.3. Climate-forcing mechanism for varied streamflow/effective precipitation of the upper reaches of the Yellow River

As shown in Figures 4 and 5, both annual precipitation and summer temperature increased, but the streamflow of the Yellow River above the Tangnaihai hydrological station only fluctuated in a relatively stable way over the last 50 years. The streamflow of the upper reaches of the Yellow River is significantly correlated with annual precipitation ($r=0.48$, $p<0.01$) but not correlated with summer temperature over the past 50 years in this region, further suggesting that the streamflow of the upper reaches of the Yellow River mainly is affected by precipitation.

The climate in the studied region is influenced by several atmospheric systems, including the Indian Summer Monsoon (ISM), the East Asian Summer Monsoon (EASM) and the Westerlies in the upper reaches of the Yellow River on the northeastern Tibetan Plateau, which create a more complex climate system compared to other parts of the Tibetan Plateau (Pu et al., 2013; Li et al., 2014). Modern meteorological data shows that the moisture is primarily brought by the southwesterly wind from the Indian Ocean (Cao and Pan, 2014). The monthly changes in the stable isotopic compositions in precipitation at Maduo show that the summer precipitation is

significantly influenced by the summer monsoon circulation in the upper reaches of the Yellow River (Ren et al., 2013). It has been suggested that most of the precipitation comes from the Indian Ocean via the summer monsoons since the last glacial maximum (LGM) (Mischke et al., 2008, 2010; Zhang et al., 2011; Cai et al., 2012). To understand the climate-forcing mechanism for varied streamflow in the upper reaches of the Yellow River on the northeastern Tibetan Plateau over the last 800 years, we compared the varied pattern of streamflow/effective precipitation of the upper reaches of the Yellow River recorded by $\delta^{18}\text{C}_{\text{BC}}$ values in core KHC14-1 with a record of Indian Summer Monsoon intensity or ISM-related precipitation deduced from stalagmite $\delta^{18}\text{O}$ record (Berkelhammer et al., 2010; Sinha et al., 2011).

As showed in Figure 6, the varied pattern of effective precipitation/streamflow recorded by $\delta^{18}\text{C}_{\text{BC}}$ values in core KHC14-1 from Lake Kuhai area generally corresponded to the ISM intensity since AD 1170 (Sinha et al., 2011). However, the effective precipitation/streamflow recorded by $\delta^{18}\text{C}_{\text{BC}}$ values in Lake Kuhai does not accurately parallel the ISM intensity (Fig. 6). The difference between these two records over the past 800 years can be explained by relatively larger dating uncertainties in our record, since the annual-resolution stalagmite $\delta^{18}\text{O}$ record contains a more reliable chronology. In general, the $\delta^{18}\text{O}$ record of stalagmite showed that the ISM precipitation was high before AD 1310, which corresponded to the high effective precipitation/streamflow indicated by low $\delta^{18}\text{C}_{\text{BC}}$ values in the upper reaches of the Yellow River (Fig. 6). Then, the ISM intensity weakened at approximately AD 1310-1640, and the effective precipitation/streamflow also declined during this time stage in the upper reaches of the Yellow River (Fig. 6). The period of AD 1310-1640 approximately corresponds to the Little Ice Age (LIA) period, which was suggested to be cool and dry on the northeastern Tibetan Plateau (Liu et al., 2009; Henderson et al., 2010; Chen et al., 2010; He et al., 2013a, 2013b; Pu et al., 2013). However, the low $\delta^{18}\text{C}_{\text{BC}}$ values showed that effective precipitation/streamflow in the upper reaches of the Yellow River suddenly increased in the early 1500s (Fig. 6). Although the increased effective precipitation/streamflow in the upper reaches of the Yellow River can be explained by strengthened ISM precipitation, the stalagmite $\delta^{18}\text{O}$ record showed that the ISM intensity was not as strong during this period (Fig. 6). However, the maximum air temperature of the headwater area of the Yellow River is very high in the late 1400s and the early 1500s, and the summer temperature was also very high around the headwater area of the Yellow River in the early and

middle 1500s on the Tibetan Plateau (Gou et al., 2008; Zhu et al., 2011). Increased temperature may cause a large amount of melting snows of the high mountains around the headwater area of the Yellow River. Significant contribution of melted snows may result in the low $\delta^{18}\text{C}_{\text{BC}}$ values observed in the Lake Kuhai core. Thus, the reconstructed streamflow of the headwaters of the Yellow River based on the $\delta^{18}\text{O}_{\text{BC}}$ value may be higher than real values. But, the low P/E ratio indicated as the high $\delta^{18}\text{O}_{\text{BC}}$ values in the Lake Kuhai area was different with the high streamflows in the upper reaches of the Yellow River and the strengthened ISM in period of AD 1400-1500 (Fig. 6). The difference cannot be explained using the uncertainty of age model, but the carbonate $\delta^{18}\text{O}$ may have changed independently because it was influenced by many other factors, such as changes in the moisture source, besides the P/E ratio. The result of past studies suggested that the moisture in this area is a mixture of ISM and convective precipitation, which results from the evaporation of local water bodies on the northeastern Tibetan Plateau (Tian et al., 2008; Ren et al., 2013). Usually, strong evaporation of water from the lakes on the Tibetan Plateau creates precipitation with higher $\delta^{18}\text{O}$ values than the ISM precipitation (Tian et al., 2008; Ren et al., 2013). Therefore, the increased convective precipitation with high $\delta^{18}\text{O}$ values and lake water loss in high temperature summer season may be one of the reasons for high $\delta^{18}\text{O}_{\text{BC}}$ values in Lake Kuhai during the period of AD 1400 to 1500. After 1640 AD, the ISM intensity gradually became strong as supported by the stalagmite $\delta^{18}\text{O}$ record, which corresponds to an increasing effective precipitation/streamflow since AD 1640 in the upper reaches of the Yellow River (Fig. 6).

Therefore, we suggest that ISM intensity/precipitation would be an important factor influencing the changes of effective precipitation/streamflow in the upper reaches of the Yellow River. Usually, the intensified (weakened) ISM can cause more (less) water vapor to be transported from the sea to the inland, resulting in more (less) continental precipitation. In fact, changes in the precipitation or streamflow in the Tibetan Plateau are well associated with the ISM intensity as shown by other instrumental records or model results (Zhang et al., 2007). Thus, it could be reliable that most changes in effective precipitation/streamflow of the upper reaches of the Yellow River in response to ISM intensity have been recorded by the $\delta^{18}\text{O}_{\text{BC}}$ values in core KHC14-1 from Lake Kuhai over the past 800 years but uncertainties should be still considered.

Both the tree-ring data and the highest $\delta^{18}\text{C}_{\text{BC}}$ values reveal that the streamflow of the upper reaches of the Yellow River was very low and the climate was dry at approximately AD 1470s

(Fig. 6). The streamflow in the period from AD 1980 to 2000 is close to the values in period of AD 1470s in the upper reaches of the Yellow River (Fig. 6). Nevertheless, it is still within the range of earlier streamflow fluctuations reconstructed for the past several centuries (Fig. 6). In recent decade, decreased $\delta^{18}\text{C}_{\text{BC}}$ values indicated an increasing streamflow again (Fig. 6). Although the Tangnaihai station only records streamflow from the headwater area, the high correlation between this station and the Sanmenxia station, which may be frequently affected by human activities over the past 60 years, ($r=0.61$, $p<0.01$) suggests that reconstructed streamflow in the upper reaches of the Yellow River using $\delta^{18}\text{C}_{\text{BC}}$ values from Lake Kuhai may be representative for a much larger region. In recent decades, the issue of freshwater resources in terms of quality and quantity is already a large problem in the Yellow River region. Hydrologic characteristics of arid inland drainage basin in western China are low precipitation, high evaporation and distinct drought periods. Both climatic variation and human activities can lead to extremely fragile water resources in this region, so water shortage is a major environmental and social concern in this area. Therefore, the development of further streamflow reconstructions from longer lake records or other records is highly valuable. These records should provide much needed and improved long-term estimates of water supply variability that are not available from the short instrumental records of streamflow.

5. Conclusions

This study presents high-resolution multi-proxy records from Lake Kuhai over the last 800 years in the upper reaches of the Yellow River on the northeastern Tibetan Plateau. By comparing our records with the instrumental record and the record derived from tree-ring data back to the AD 1170s, we can closely examine changes in the oxygen isotope of the bulk carbonates in response to changes in effective precipitation/streamflow of the upper reaches of the Yellow River. The $\delta^{18}\text{O}_{\text{BC}}$ values in Lake Kuhai mainly respond to variation in the P/E ratio, which is indirectly associated with changes of streamflows in the upper reaches of the Yellow River. Lake Kuhai $\delta^{18}\text{O}_{\text{BC}}$ records parallel the changes of the Yellow River streamflow recorded by instrumental stations and the 800-year tree ring records in the headwater areas of the Yellow River on the northeastern Tibetan Plateau. In general, the highest $\delta^{18}\text{O}_{\text{BC}}$ values showed that the streamflow was very low and the climate might be very dry in the middle of the AD 1400s in the upper reaches of the Yellow River over the past 800 years. In addition, the streamflow of the upper

reaches of the Yellow River was also very low because of decreased precipitation and increased temperature from the early 1980s to the late 1990s, and then the streamflow increased with increasing precipitation. In general, similar changing trends between $\delta^{18}\text{O}_{\text{BC}}$ values and ISM intensity recorded by stalagmite $\delta^{18}\text{O}$ values suggested that changes in the effective precipitation/streamflow of the upper reaches of the Yellow River were dominated by variations in ISM precipitation as observed in other records from the northeastern Tibetan Plateau.

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Captions

Table 1 The AMS radiocarbon dates and calendar ages for the Lake Kuhai sedimentary core (KHC14-1).

Table 2 Carbon and oxygen isotopic compositions of bulk carbonates from core KHC14-1 in Lake Kuhai.

Figure 1. Map of Lake Kuhai and its surrounding environment, the solid red five-pointed star in the map shows the location where core ELC14-1 was collected.

Figure 2. Activity profiles of ^{210}Pb (a) and ^{137}Cs (b) for the sediment core KHC14-1.

Figure 3. The variation in bulk carbonates content (yellow), content of grain size $>63\ \mu\text{m}$ (blue) (a), and the stable isotopic variations of bulk carbonates (green: $\delta^{13}\text{C}$ values, black: $\delta^{18}\text{O}$ values) that were recorded by the core KHC14-1 covering approximately 800 years

Figure 4. Mean annual precipitation (red) and mean annual summer temperature (black) over the last 50 years from Maduo Weather Station, which is ca. 70 km southwest of Lake Kuhai.

Figure 5. Comparison of varied bulk carbonates $\delta^{18}\text{O}$ values from core KHC14-1 (green circle) with instrumental records of the Yellow River streamflow above Tangnaihui hydrological station over the last 50 years (purple) in the headwater areas of the Yellow River.

Figure 6. Comparison of bulk carbonates $\delta^{18}\text{O}$ values (green line shows three-point moving average $\delta^{18}\text{O}$ values) from core KHC14-1 with a) streamflow of the upper reaches of the Yellow River deduced from tree ring data (Gou et al., 2007) and c) Indian Summer Monsoon intensity/precipitation recorded by stalagmite $\delta^{18}\text{O}$ (blue line shows thirty-point moving average $\delta^{18}\text{O}$ data) (Sinha et al., 2011) over the last 800 years.

Figure

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Graphic abstract

Table 1

Lab. Code	Sample No.	Depth (cm)	Material	Measured ¹⁴ C age	Corrected by 2018 yr age (cal. yr BP) (Median Probability)
XA13184	KHC14-3-22	11	macrophyte remains	2018 ± 28 BP	0
XA12814	KHC14-1-108	54	macrophyte remains	2991± 32 BP	863

Table 2

Depth (cm)	Year (AD)	$\delta^{13}\text{C}$ (V-PDB, ‰)	$\delta^{18}\text{O}$ (V-PDB, ‰)	Depth (cm)	Year (AD)	$\delta^{13}\text{C}$ (V-PDB, ‰)	$\delta^{18}\text{O}$ (V-PDB, ‰)
0.5	2010	2.06	1.26	27.5	1689	-0.16	-4.84
1.0	2007	2.22	1.49	28.0	1679	-0.71	-5.74
1.5	2005	2.00	1.10	28.5	1670	0.50	-4.09
2.0	2003	2.06	0.85	29.0	1660	0.75	-3.56
2.5	2001	2.03	0.73	29.5	1650	1.40	-1.89
3.0	1999	2.06	0.41	30.0	1640	2.66	1.31
3.5	1997	1.43	-1.19	30.5	1630	1.62	-1.71
4.0	1995	-0.17	-4.64	31.0	1621	1.17	-2.13
4.5	1992	-0.11	-4.81	31.5	1611	2.14	0.76
5.0	1989	-0.34	-5.02	32.0	1601	1.30	-1.71
5.5	1987	-0.06	-4.45	32.5	1591	0.65	-2.93
6.0	1984	-0.59	-5.16	33.0	1581	-0.32	-4.42
6.5	1981	0.09	-4.53	33.5	1572	-0.75	-5.56
7.0	1978	0.20	-4.02	34.0	1562	1.55	-0.45
7.5	1976	1.31	-2.01	34.5	1552	0.50	-2.84
8.0	1973	1.34	-2.25	35.0	1542	-0.93	-5.64
8.5	1970	1.93	-0.76	35.5	1532	-2.24	-8.32
9.0	1966	1.71	-1.45	36.0	1523	-0.33	-4.35
9.5	1963	0.65	-3.51	36.5	1513	0.72	-1.83
10.0	1959	0.06	-4.61	37.0	1503	1.79	0.48
10.5	1955	-0.30	-5.52	37.5	1493	1.40	-0.47
11.0	1950	-0.49	-5.66	38.0	1483	1.17	-0.60
11.5	1946	0.09	-4.73	38.5	1474	2.25	1.60
12.0	1941	-0.19	-5.37	39.0	1464	1.02	-0.71
12.5	1936	-0.83	-6.11	39.5	1454	1.18	-0.63
13.0	1932	-0.88	-6.37	40.0	1444	1.42	-1.15
13.5	1927	-0.36	-5.69	40.5	1434	2.12	-0.64
14.0	1923	-0.13	-4.90	41.0	1424	1.78	-0.89
14.5	1919	0.08	-4.64	41.5	1415	1.99	-0.20
15.0	1916	1.30	-2.10	42.0	1405	1.84	-0.46
15.5	1912	0.89	-2.38	42.5	1395	1.97	-0.39
16.0	1909	0.68	-2.60	43.0	1385	1.89	-1.15
16.5	1905	0.70	-2.70	43.5	1375	2.18	-0.63
17.0	1895	0.51	-2.63	44.0	1366	2.16	-0.79
17.5	1885	0.50	-3.02	44.5	1356	1.82	-1.90
18.0	1876	0.65	-2.86	45.0	1346	-0.52	-5.19
18.5	1866	1.91	-1.30	45.5	1336	0.06	-4.58
19.0	1856	1.65	-1.86	46.0	1326	-1.74	-7.00
19.5	1846	3.06	0.29	46.5	1317	-1.55	-7.16
20.0	1836	2.49	-0.33	47.0	1307	-1.67	-7.31
20.5	1827	2.65	-0.13	47.5	1297	-1.21	-6.65
21.0	1817	3.50	0.53	48.0	1287	1.10	-2.65
21.5	1807	3.30	0.63	48.5	1277	0.04	-4.54
22.0	1797	1.54	-2.03	49.0	1268	-0.34	-5.17
22.5	1787	2.35	-0.93	49.5	1258	-1.37	-6.93
23.0	1778	0.57	-3.34	50.0	1248	-1.37	-7.04
23.5	1768	-0.33	-5.03	50.5	1238	0.53	-4.71
24.0	1758	0.43	-3.76	51.0	1228	0.13	-4.22
24.5	1748	0.78	-3.51	51.5	1219	-0.23	-4.28
25.0	1738	0.12	-4.52	52.0	1209	0.87	-2.88
25.5	1728	-0.08	-4.89	52.5	1199	0.66	-3.33
26.0	1719	-0.39	-5.08	53.0	1189	-0.78	-5.09
26.5	1709	-0.19	-4.64	53.5	1179	1.04	-1.94
27.0	1699	-0.40	-4.69	54.0	1169	0.54	-3.40