Can $\delta^{18}O$ help indicate the causes of recent lake area expansion on the western Tibetan Plateau? A case study from Aweng Co

Yuzhi Zhang · Matthew Jones · Jiawu Zhang · Suzanne McGowan · Sarah Metcalfe

Abstract Glacier-fed lakes on the Tibetan Plateau (TP) have undergone rapid expansions since the late 1990s, concurrent with the changing climate. However, the dominant cause(s) of lake area increases is still debated. To identify the drivers of lake expansion, we studied Aweng Co, a glacier-fed lake in the western TP, where surface area has increased (0.74 km$^2$ year$^{-1}$) since the late 1970s and most rapidly (0.998 km$^2$ year$^{-1}$) since the late 1990s. A water balance model was used to clarify the reasons for increased lake water volume, supported by stable isotope hydrology and the $\delta^{18}O$ change recorded in recent sediments. Results showed that glacial meltwater probably had the biggest impact on changes in Aweng Co lake level in recent decades, but that precipitation was also an important contributor. Our study shows that $\delta^{18}O$ of carbonate ($\delta^{18}O_{\text{carb}}$) has great potential for indicating source changes of water supply in such lakes, but there is a need to be cautious when interpreting $\delta^{18}O_{\text{carb}}$ due to the influence of multiple hydrological factors, which can change in dominance over time.

Keywords Lake sediment · Oxygen isotope · Water-balance model · Glacial meltwater

Introduction

Lake expansion (increased lake surface area) has been identified by remote sensing across the Tibetan Plateau (TP) in recent decades (Crétaux et al. 2016; Lei et al. 2013, 2014, 2017; Song et al. 2014; Zhang et al. 2015, 2017a). New lakes (99 larger than 1 km$^2$) have appeared across the TP since 1970 and 81% of the existing lakes have expanded, with a total increase in surface area of 7240 km$^2$ between the 1970s and 2010 (Zhang et al. 2017a). Most lakes in the inner TP have undergone an apparent increase in area since...
Given the multiple potential drivers of lake level change, a lake-by-lake rather than a regional conceptual model approach is needed for down core interpretations of hydroclimatic change. Here we present a detailed study of lake area change since the late 1990s from an alpine lake, Aweng Co in the western TP. The lake is hydrologically closed, fed by direct precipitation onto the lake surface, and runoff generated by precipitation and glacier meltwater. In order to identify the dominant factors that led to lake area expansion, we analysed the components of a water balance model that could influence lake water volume change since the late 1990s, including the use of $\delta^{18}$O as a tracer. We then compared the $\delta^{18}$O$_{\text{carb}}$ values in recent sediments with the water balance model to verify the water supply variations, and to understand the relationship between the water source changes and changes in the $\delta^{18}$O$_{\text{carb}}$. This aim of the study is to improve our understanding of the factors controlling lake system change in such environments, over both recent and palaeo-timeframes.

Study site

Aweng Co (A’ong Co, 32.70° ~ 32.82° N, 81.63° ~ 81.80° E) is a closed-basin saline lake located in the western Tibetan Plateau (Fig. 1a). It lies at 4,430 m a.s.l. and is surrounded by 500 m high hills. Catchment vegetation, where present, is typical of alpine desert steppe including Stipa grasses. The catchment mainly consists of Cretaceous granite and Jurassic metamorphosed sandstone. Aweng Co is an elongated, shallow lake, which is 23.4 km long with a mean and maximum width of 2.52 km and 5.30 km respectively; the maximum water depth is 6 m. Within a large catchment area (2052.30 km²), the current lake water area is only 68.96 km² (in 2015). In the western part of the catchment, glaciers and snow at elevations higher than 5000 m a.s.l (Fig. 1b) cover an area of 125.80 km² (Li et al. 2017; Song et al. 2014; Wang and Dou 1998), and are currently approximately 50 km from the lake. Satellite imagery shows that the lake area expanded dramatically from the late 1990s (ESM 1).

In 2015 a pH of 9.2 and salinity of 29.5 g L$^{-1}$ were recorded in the lake centre, with concentrations of 1850 mg L$^{-1}$ CO$_3^{2-}$ and 2023 mg L$^{-1}$ HCO$_3^-$. Meteorological data at the Shiquanhe Station (32.50° N, 80.08° E; altitude: 4279.3 m a.s.l.), 150 km from
Aweng Co, show that mean annual temperature and total precipitation are 0.68 °C and 69.11 mm (for the period 1971–2012, https://data.cma.cn/). 87.6% of precipitation at Shiquanhe falls between May and September during the Indian Summer Monsoon (ISM) season (Fig. 1c). The mean temperatures in January and July are –12 °C and 14 °C, respectively. Monthly mean temperature is above 0 °C between May and October (Fig. 1c), and the lake surface usually freezes in October and thaws in May. The \( \delta^{18} \text{O} \) value of the central lake waters was 0.2‰ in 2015.

**Materials and methods**

**Lake volume reconstruction**

The region has been monitored by satellite imagery since the 1970s, including Landsat 4–5 Thermal Mapper (TM), Landsat 7 Enhanced Thematic Mapper (ETM), and Landsat 8 Operational Land Imager (OLI). Lake area data from the National Tibetan Plateau Data Center (https://data.tpdc.ac.cn; Zhang et al. 2014, 2019a; Zhang 2019) is averaged over 3 or 4 years, and so is of insufficient resolution to determine annual changes in lake area. Therefore, we used images with no cloud cover from the Geospatial Data Cloud (https://www.gscloud.cn/; ESM 2), sampled at a consistent time of the year (September–October) to minimize the influence of seasonal variability (Zhang et al. 2017b). This period is useful for comparing interannual changes in lake area because it records lake size at the end of the warm and wet season. Gaps in the Landsat ETM + scan line corrector-off images were filled by the neighbourhood similar pixel interpolator algorithm (Chen et al. 2011). Lake area data before 1990 was downloaded from the National Tibetan Plateau Data Center (Zhang et al. 2014, 2019a; Zhang 2019).

We calculated past changes in lake volume using a combination of the lake area measurements and a digital elevation model (DEM) of the lake, derived from a bathymetric survey by SM-5A hand-held sonar conducted in 2015. Lake volume was calculated using the VOLUME function in Surfer 11.0 sequentially lowering lake levels. Lake level altitude data was derived from ICESat Laser altimetry measurements, which were available from 2003 to 2009 (Zhang et al. 2011, 2017a). We calculated the lake level altitude for 1999–2002 and 2015 and thereby lake volume, according to the correlation between lake area and lake-level altitude from 2003 to 2009. The lake volume before 1999 was calculated from this relationship using lake area measurements from the

![Fig. 1](image-url)
National Tibetan Plateau Data Center (Zhang et al. 2014, 2019a; Zhang 2019).

Stable isotopes of water and sediments

In 2015, a 411.5-cm-long sediment core (AWC2015B) was taken from the central part of the lake (32.75° N, 81.76° E) at a water depth of 6 m using a UWITEC corer (Fig. 1b). The chronology of the core top was established by 137Cs and 210Pb using HPGe Gamma Spectrometry. 210Pb was obtained via gamma-emission at 46.5 keV and 226Ra at 351.92 keV γ-rays emitted by its daughter isotope 214Pb. The age of the top sediment was established by the Constant Rate of Supply (CRS) model (Appleby and Oldfield 1978). The top 14 cm, which covered the period with instrumental data, was used in this study, with a sampling interval of 0.5 cm.

Fine-grained carbonates (< 40 µm fraction) were collected from sediments by wet sieving and then dried at 50 °C for 6 h. The minerogenic composition was confirmed to be aragonite by X-ray diffraction analysis. Stable oxygen isotopes were analysed from the carbonates using a ThermoFisher MAT 253 mass spectrometer with an automated carbonate preparation device (Kiel IV). Four standards (NBS18, NBS19, GBW04406, GBW04405) were measured every 10 samples. Analytical precision for δ18O and δ13C was better than 0.1‰ and 0.3‰ respectively.

Hydrological model

To begin to identify the likely contributions of hydroclimate, such as evaporation, precipitation, and glacial meltwater, to the observed increase in Aweng Co lake area over recent decades, we attempted to model the lake hydrology based on Eq. 1. We made the assumption that volume changes at Aweng Co are controlled by a number of inputs and outputs to the system (Eq. 1), recognising that there is no surface outflow from the lake.

\[ \Delta V_L = P_L S_L + R_C S_C + GMW + G_I - E_L S_L - G_O \]  

where \( \Delta V_L \) is the change in lake volume (m³) in a given time, \( P_L \) is the precipitation onto the lake surface, \( S_L \) is the lake surface area; \( R_C \) is runoff from the catchment; \( S_C \) is the catchment area excluding the lake area; \( E_L \) is the total evaporation on the lake surface; \( GMW \) is the glacier meltwater and \( G_I \) and \( G_O \) are inflow and outflow groundwater components respectively. Because the lake area of Aweng Co expanded dramatically since the late 1990s, we employed the water balance model for the period of 1999–2009 in this study.

\( \Delta V_L \), the change in lake volume, is a known value, as are the lake and catchment areas from the remote sensing work. Because there is no meteorological station in the study area, precipitation in the Aweng Co basin was taken from the LZU0025 dataset (Wu et al. 2014) calculated using the Thin Plate Smoothing Spline (TPSS) method which interpolates data from all meteorological stations in China. Due to the positive correlation between precipitation and elevation in the western Tibetan Plateau (Zhang et al. 2019b), the interpolated precipitation values in the Aweng Co catchment (Table 1) are higher than those from Shiquanhe Station. The quantity of the precipitation falling on the catchment that reaches the lake is unknown, and is probably a combination of surface and groundwater, or at least subsurface flow. Here we take overland flow (\( R_C \)) to be a proportion (\( c \)) of precipitation falling on the catchment.
Glacial meltwater may reach the lake through both overland and sub-surface flow. Here GMW is taken to be the surface component, such that GMW is estimated based on the measured change in glacial volume available for the Aweng Co catchment, multiplied by a constant \( g \). To convert measured glacial area \( S_g \) to a volume \( V_g \) we used the formula from Zhu et al. (2010) based on data from 253 glaciers in the China Glaciers Catalogue, 
\[
V_g = 0.042S_g^{1.3565}
\]
Variation in glacier volume was converted to glacier meltwater volume by multiplying by 0.85 (Huss 2013), and only 42% of this volume is known to drain into the Aweng Co basin (Neckel et al. 2014). Unknown constants \( c \) and \( g \) both therefore take into account potential infiltration and evapotranspiration, i.e. factors that prevent all precipitation or meltwater flowing directly into the lake.

\( E_L \) (mm day\(^{-1}\)) is calculated using the equation of Linacre (1992) such that
\[
E_L = \left[ 0.015 + 4 \times 10^{-4}T_a + 10^{-6}z \right]
\times \left[ \frac{480(T_a + 0.006z)}{84 - A} - 40 + 2.3u(T_a - T_d) \right]
\]
where \( T_a \) is air temperature (°C), \( z \) = altitude (m), \( A \) = latitude (degrees), \( u \) = wind speed (m s\(^{-1}\)), \( T_d \) = dew point temperature = 0.527\( T_{a,min} \) + 0.607\( T_{a,max} \) - 0.009\( (T_{a,max})^2 - 2 \) °C. This has been shown to be a reasonable estimate of evaporation where full suites of meteorological data are not available (Jones et al. 2016). Because data for \( T_{a,min}, T_{a,max} \) and \( u \) are unavailable in the LZU0025 dataset, they were taken from the National Centers for Environmental Prediction (NCEP)-Department of Energy (DOE) Reanalysis 2 Gaussian Grid data (https://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis2.gaussian.html). When calculating the precipitation and evaporation on the lake surface, we used the mean lake area for the measurement year, estimated from the lake area at the beginning and end of each year. \( G_i \) and \( G_o \) are unknown.

To investigate the remaining unknowns in the lake hydrology model we firstly aimed to optimize calculated changes in lake volume, using Eq. 1, with those that have been measured. As a first order test, we aimed to optimize values of \( c \) and \( g \) such that these constants are \( C \) and the regression relationship between known and modelled \( \Delta V_L \) has a slope and \( r^2 \) of 1 and an intercept of 0.

We then took an index lake approach (Gibson et al. 2016; Jones et al. 2016) to understand whether the lake is likely to have any groundwater outflow. This approach calculates the isotopic composition of the theoretical lake \( \delta_L \) that sits at the extreme end of the local evaporation line (LEL) i.e. a fully closed hydrological system where \( P\delta_p = E\delta_E \). As \( \delta_E \) is a function of \( \delta_L \), and in the case of the index lake \( \delta_L = \delta_p, \delta_L \) can be calculated. We use the \( \delta_E \) equation based on the Craig-Gordon Evaporation model (Craig 2016).
and Gordon 1965), as used by Steinman et al. (2010a, b):

$$\delta_E = \frac{\alpha^* \delta_L - h \delta_A - \varepsilon}{1 - h + 0.001 \varepsilon_k}$$  (3)

where $\alpha^*$ is the equilibrium isotopic fractionation factor dependent on the temperature at the evaporating surface. For oxygen

$$\frac{1}{\alpha^*} = \exp(1137T_L^{-2} - 0.4256T_L^{-1} - 2.0667 \times 10^{-3})$$  (4)

and for hydrogen

$$\frac{1}{\alpha^*} = \exp(24844T_L^{-2} - 76.248T_L^{-1} - 52.61 \times 10^{-3})$$  (5)

where $T_L$ is the temperature of the lake surface water in degrees Kelvin (Majoube 1971), $h$ is the relative humidity normalized to the saturation vapour pressure at the temperature of the air water interface and $\varepsilon_k$ is the kinetic fraction factor; for $\delta^{18}O$, $\varepsilon_k$ has been shown to approximate $14.2(1 - h)$ and $12.5(1 - h)$ for $\delta^2H$ (Gonfiantini 1986). $\delta_A$ is the isotopic value of the air vapour over the lake and $\varepsilon = \varepsilon^* + \varepsilon_k$ where $\varepsilon^* = 1000(1 - \alpha^*)$.

Gibson (2002) and Gibson et al. (2016) have shown that the relationship between $\delta_P$ and $\delta_A$ varies in different environmental settings, and advocate using a measured LEL, as we have available here, to calculate the suitable regional $\delta_P$–$\delta_A$ relationship.

Finally, we attempted to balance the hydrological and isotopic components of the Aweng Co lake system, to give estimates for each of the parameters in Eq. 1. Based on optimized values of $c$ and $g$, and a constant groundwater inflow, and using average values for each modelled component from the 10 years of monitoring for which lake volumes were measured (Table 1, ESM 3) we undertook a mass balancing exercise, such that lake inputs should balance lake outputs (Lacey and Jones 2018), i.e.

$$P_L \delta_{PL} + R_i \delta_{RI} + GMW \delta_G + G_i \delta_{GI} = E \delta_E + G_o \delta_L$$  (6)

As all isotopic values are known, Eq. 6 can then be optimized, varying groundwater inputs such that the equation balances for both $\delta^{18}O$ and $\delta^2H$ values, resulting in estimates for the percentage contribution of each of these parameters to the Aweng Co hydrology.

**Results**

Changes in lake volume since the late 1990s

Between 1980 and 1999, lake area and lake volume of Aweng Co increased from 46.51 to 60.69 km$^2$ and from $57.52 \times 10^6$ m$^3$ to $83.74 \times 10^6$ m$^3$, respectively; with a rapid lake area expansion between 1996 and 1999 (Fig. 2d). Before 1999, lake area and lake

![Fig. 2](image-url)

**Fig. 2** Comparisons of lake area change and glacier area change with $\delta^{18}O_{carb}$ value, and meteorological data. (a) Interpolated precipitation in the Aweng Co catchment from 1980 to 2012 (Wu et al. 2014). (b) Interpolated mean summer temperature (June to August) in the Aweng Co catchment from 1980 to 2012 (Wu et al. 2014). (c) Calculated evaporation on the lake surface from May to September in the Aweng Co catchment. (d) Aweng Co lake area since 1979. (e) Calculated lake volume. (f) Glacier area change from 1996 to 2010. (g) Sedimentary record $\delta^{18}O_{carb}$ from Aweng Co from 1978 to 2014 (The data of $\delta^{18}O_{carb}$ are shown in ESM 5). The black dots and bars represent the chronology of the samples and the errors (full data presented in ESM 4)
volume increased slowly at 0.74 km$^2$ year$^{-1}$ and 1.38 \times 10^6$ m$^3$ year$^{-1}$, respectively. After 1999 lake area and lake volume increased reaching 69.15 km$^2$ and 125.02 \times 10^6$ m$^3$ in 2002, and then decreased until 2005; with an increase from 2006, culminating in 2008 with an area of 71.06 km$^2$ and a volume of 136.77 \times 10^6$ m$^3$, respectively. The lake reached maximum size for the study period in 2010 with an area of 71.67 km$^2$ and a volume of 153 \times 10^6$ m$^3$, and then shrank a little (Fig. 2d, e). The lake area and lake volume increased at a rate of 0.998 km$^2$ year$^{-1}$ and 6.29 \times 10^6$ m$^3$ year$^{-1}$ from 1999 to 2010.

Correlations (Fig. 3) between each known hydro-climate parameter and lake volume change show that precipitation and glacier meltwater changes both have significant and positive correlations with lake volume change. Evaporation has a negative relationship with lake volume change, but the relationship is relatively weak, and not significant.

**Aweng Co isotope hydrology**

Precipitation samples at Aweng Co (Fig. 4) lie on a local meteoric water line (MWL). Lake water samples lie to the right of the Aweng Co MWL, and with the groundwater sample describe a local evaporation line (LEL) with a gradient of 5.64 (Fig. 4).

**Model results**

The results of initial optimization showed it is difficult to optimize $r^2$, slope and intercept concurrently (Table 2), and the best combination of $c$, $g$ and groundwater input, to give variability in the model at a magnitude that matches the measured volume changes is where $c$ and $g$ are optimized to give a regression with slope of 1 (resulting in an $r^2$ of 0.64) in which case a constant amount of groundwater inflow supplying the lake is required to give the 0 intercept.

**Fig. 3** The linear correlation between each parameter (left to right: precipitation on the lake surface; evaporation on the lake surface and glacier meltwater) and lake volume change. $P_L$ represents precipitation on the lake surface. $S_L$ represents lake surface area. $E_L$ represents evaporation on the lake surface.
When calculating \( \delta_L \) for the “index lake” we used a lake system where \( \delta_E \) was equal to the intercept value of the LEL and Aweng Co MWL. In this case for \( \delta_L \), \( \delta_A \) and \( \delta_E \) to sit sensibly in \( \delta^{18}O-\delta^2H \) space (Fig. 4), an adjustment, via a constant \((k)\), is required to the standard equilibrium relationship between \( \delta_P \) and \( \delta_A \) (Gibson et al. 2016), where:

\[
\delta_A = \frac{\delta_P - k e^*}{1 + 10^{-3} \cdot ke^*} \tag{7}
\]

The value of \( k \) needed here (0.5), to fit the theoretical LEL to that measured in this study is typical for highly seasonal climates such as that at Aweng Co (Gibson et al. 2016).

The contributions of each component from the balanced \( \delta^{18}O \) and \( \delta^2H \) isotopic models (Eq. 6) are nearly the same (Table 3). Based on the \( \delta^{18}O \) balance model, the biggest supplier of water to Aweng Co is groundwater inflow, which accounts for 67% and the smallest is glacier meltwater that is 4%. Precipitation and runoff in the catchment supply 10% and 19% to the hydrological systems, respectively. Evaporation accounts for 57% of the water loss, more than the groundwater outflow, which is 43%.

### Discussion

Contributors to lake volume change: monitoring and modelling results

The combined monitoring and various modelling exercises for Aweng Co presented here have, at least on a general scale, begun to tell a coherent story for the lake system. The combined hydrological and isotope mass balance modelling (Table 3) give a similar picture to the water-isotope bi-plot (Fig. 4) in suggesting that both evaporation and groundwater are important outputs from the lake. The estimate of two thirds loss by evaporation (Table 3) is a sensible order of magnitude given the location of Aweng Co on the LEL (Fig. 4), the gradient of which is very similar to the LEL gradient (5.51) for other closed lakes that have experienced lake expansions in recent decades on the Tibetan Plateau (Yuan et al. 2011). Correlations (Fig. 3) between each known hydroclimate parameter and lake volume change indicate both glacial meltwater and changing precipitation amount could be controlling the observed lake-area change.

### Table 2

| Constant \( c \) | Constant \( g \) | \( r^2 \) | slope | intercept |
|------------------|-----------------|--------|-------|-----------|
| 0.06             | 0.13            | 0.64   | 1.00  | 31,493,754 |
| 0.09             | 0.94            | 0.51   | 0.28  | 15        |
| 0.84             | 0.74            | 0.68   | 0.17  | −38,453,042 |

### Table 3

| Hydrological component | Contribution (%) | From \( \delta^{18}O \) balance | From \( \delta^2H \) balance |
|------------------------|------------------|---------------------------------|-----------------------------|
| Water input            |                  |                                 |                             |
| \( P_L \)              | 10               | 11                              |                             |
| \( R_C \)              | 19               | 20                              |                             |
| GMW                    | 4                | 5                               |                             |
| GMW                    | 67               | 65                              |                             |
| Water output           |                  |                                 |                             |
| \( E \)                | 57               | 61                              |                             |
| \( G_O \)              | 43               | 39                              |                             |

Sediment chronology and proxies

The dating model for the top of the core showed that the sediments at 14 cm depth were deposited ca. 1898 AD (ESM 4). We used the upper 7 cm of sediment in this study, which represented the time period since the late 1970s, with a sampling resolution of 0.5 cm (2.6 years). The \( \delta^{18}O_{\text{carb}} \) values were around 1.5% between 1979 and 1984, and then decreased to 0.34% in 1989 and kept relatively stable until 1997, followed by a trough (with a lowest \( \delta^{18}O_{\text{carb}} \) value of −1.24%) around the mid-2000s and a positive trend after ~2007 (Fig. 2g).
To further refine our hydrological model we used the isotope hydrology of the site (Fig. 4). Groundwater, isotopically, lies on the Aweng Co MWL, which has a similar gradient to the MWL described by Guo et al. (2017) for Ngari, 190 km far from Aweng Co and ~ 170 m lower. If the groundwater is a mixture of both precipitation and glacial meltwater, the isotope values of these different components could help to estimate the relative contributions of the two sources. There are minimal data with which to undertake this exercise, but with that available we can make a preliminary estimate of the amount of precipitation and glacier meltwater in the groundwater entering Aweng Co. The most negative of the precipitation samples collected in the 2015 field season ($\delta^{18}O = -13.86 \%$, $\delta^2H = -115.48 \%$) was a sample of snow, and therefore probably lies towards the negative isotopic end of local precipitation. There are no isotope data from the glacier that feeds Aweng Co, but $\delta^{18}O$ values for other Tibetan glaciers are typically in the range of the catchment’s snow sample. The average $\delta^{18}O$ value from the Puruogangri Ice Cap is $-13.66\%$ in the most recent 50 years (Thompson et al. 2011), the upper meters of the Guliya Ice Cap, in the north of the plateau, and in a different climate region to Aweng Co, average $-11.2\%$ and $-13.1\%$ from the 2015 and 1992 cores respectively (Thompson et al. 2018). Given these values, and the groundwater sample ($\delta^{18}O = -12.29\%$, $\delta^2H = -99.77\%$), it appears likely that this groundwater is dominated in composition by snow and glacier meltwater ($\sim 70\%$), although distinguishing between these two would need further monitoring of the Aweng Co system. It is also possible that our precipitation values and runoff constant underestimate the amount of snowmelt that enters the lake, such that our “groundwater” value here includes all currently unmeasured inflows, including snowmelt.

For the differing inflow parameters, given the location of the groundwater sample and average precipitation in $\delta^{18}O$ $\delta^2H$ space (Fig. 4), if $\sim 70\%$ of the “groundwater” inflow comes from ice and snow melt, then approximately $50\%$ of Aweng Co inflow (surface and groundwater) comes from ice and snowmelt and $50\%$ from summer rainfall. This would suggest that for this lake system both glacier melt and rainfall changes may help to explain recent lake area expansion.

One potential way to distinguish further which component may have been more significant in recent times is to look at the potential sensitivity of the system to changes in these different parameters. Although there is a strong correlation between precipitation and lake volume change (Fig. 3), the magnitude of lake area and lake volume change through the time period of this correlation (1999–2009) is small compared to the longer term variability (Fig. 2). Over the longer period since ~ 1980 there have been larger increases in lake area, but no similar trend in increasing annual precipitation.

There are relatively few data points to observe the relationship between lake area and glacier area, as a proxy for meltwater, through the 1999–2009 window, but the significant decline in glacier area between 1997 and 1999 matches the significant period of lake area expansion which, alongside the lack of significant shifts in precipitation trends through that time period, suggests that it was glacier meltwater which drove the change in lake volume.

The analyses presented here suggest that isotope hydrology can help further the understanding of controls on changes in western Tibetan lakes, but that to fully exploit their potential a more detailed monitoring programme needs to be undertaken, ideally over a number of years.

$\delta^{18}O_{\text{carb}}$ evidence

The variation of $\delta^{18}O_{\text{carb}}$ is controlled by lake water $\delta^{18}O$ and temperature changes (Leng and Marshall 2004; Xu et al. 2006), and therefore the signals of lake hydrology variations could be preserved in the $\delta^{18}O_{\text{carb}}$ sediment record. The mean summer temperature change rise of 1.1 °C (Fig. 2b) would lead to $\delta^{18}O_{\text{carb}}$ change of $\sim 0.26 \%$ based on a temperature-dependence of carbonate fractionation of $-0.24 \%/\degree C$ (Craig 1965), which is not enough to explain the magnitude of $\delta^{18}O_{\text{carb}}$ fluctuations (1.74%) between 1997 and 2006 (Fig. 2g; ESM 5), suggesting that changes in lake water $\delta^{18}O$ have been important in driving the recorded $\delta^{18}O_{\text{carb}}$. This, alongside the importance of evaporation in the lake system (Fig. 4) suggests that the inflow to evaporation ratio (I:E) is probably the main driver of $\delta^{18}O_{\text{carb}}$ at Aweng Co. Of particular interest through recent decades is the negative excursion in $\delta^{18}O_{\text{carb}}$ between ~1999 and ~2008, which would need an increase in inflow...
or decrease in evaporation in an I:E driven system, or a significant change in the isotopic component of the inflowing water.

During the period 1999 to 2007, evaporation at the lake surface showed an overall slight increasing trend (Fig. 2c), with only a short, two year, reduction in evaporation through that time. Even within the chronological uncertainties of the core record, this is not enough to explain the trends in the δ18O_{carb} record.

Comparison of trends in precipitation (Fig. 2a) with the δ18O_{carb} record also shows no clear relationship between periods of increased amounts of precipitation and negative isotope excursions. The biggest decline in glacier area in the late 1990s does match, within the chronological errors of the core, the δ18O_{carb} shift to more negative values (~ 1.24‰ in 2006). Given I:E ratio is likely the main driver of δ18O_{carb} change, an increased amount of glacier meltwater would increase lake area/volume and lead to a negative shift in δ18O_{carb}.

Although the δ18O_{carb} returns to early 1990s values (~ 0.5‰) after the negative excursion towards the top of the core (Fig. 2), there are no similar returns for either the glacier area or the lake area. One interpretation for the difference is that the lake isotope values are returning to a steady state following the negative excursion, but these isotope values are similar to those when the lake level was lower. This could be because inflows and outflows to the system are generally the same in both the low and high lake level status. In such a system, flux, which has been considered important in controlling δ18O_{carb} in other lake systems (Jones et al. 2007), remains the same, while volume has increased due to the elevated glacial meltwater period in the late 1990s. In this scenario it is also possible that the negative excursion under discussion is a result of the particularly negative isotopic value of that glacial meltwater, rather than the amount of it, such that the impact of this input changed the δ18O_{carb} record more than the volume change. Meanwhile, the duration of the isotopic impact was limited by the relatively short residence time of the water, with negative isotopic water flushed through the system, whilst lake volume remains relatively unchanged. Overall, it is likely that a combined effect of increased inflow of particularly isotopically-negative glacial meltwater led to this negative shift in δ18O_{carb}.

This comparison exercise shows how even with instrumental data available for contrast, the interpretation of δ18O_{carb} records is complicated by the multiple potential controls that can lead to an abrupt change in a core δ18O_{carb} record. This highlights the need to have multiple proxies from which more robust interpretations of environmental changes from down-core data can be made beyond the instrumental time period.

Conclusions

The combined monitoring, modelling and palaeolimnological approach taken here shows the potential for δ18O_{carb} to be used to investigate lake area change in the western Tibetan Plateau, whilst highlighting the complexities of the system. This understanding is important for using such core records to reconstruct longer term environmental change in the region. Both the monitoring, modelling and δ18O_{carb} evidence point to the importance of glacial meltwater in influencing the lake area and isotopic record of Aweng Co, but highlight that the sensitivities of these two parts of the lake system to glacial meltwater change can be different. The flux of water through the lake system, controlled by precipitation amount and evaporation as well as glacial meltwater, is also therefore important in driving the resulting δ18O_{carb} record preserved in the sediments, and the dominant hydrological controls may change through time.

Acknowledgements

This study is supported by the National Natural Science Foundation of China (NSFC 41771212) and Fundamental Research Fund for the Central Universities (lzujbky-2017-it81). We would like to thank Juzhi Hou, Mingda Wang, Yaping Yang and Erlei Zhu for assisting the field work. We also thank Melanie Leng for her constructive suggestions in improving the quality of the manuscript, and Xian Wu for providing the interpolated meteorological data in the catchment. We thank Thomas J. Whitmore, Steffen Mischke and two anonymous reviewers for detailed comments which improved the manuscript. The authors have no conflict of interest to declare.

References

Appleby PG, Oldfield F (1978) The calculation of lead-210 dates assuming a constant rate of supply of unsupported 210Pb to the sediment. CATENA 5:1–8
Atkinson SE, Woods RA, Sivapalan M (2002) Climate and landscape controls on water balance model complexity
over changing timescales. Water Resour Res 38:50-1–50-15
Chen J, Zhu X, Vogelmann JE, Gao F, Jin S (2011) A simple and effective method for filling gaps in Landsat ETM + SLC-off images. Remote Sens Environ 115:1053–1064
Conway D (1997) A water balance model of the Upper Blue Nile in Ethiopia. Hydrol Sci J 42:165–286
Craig H (1965) The measurement of oxygen isotope palaeotemperatures. In: Tongiorgi E (ed) Stable isotopes in oceanographic studies and palaeotemperatures. Consiglio Nazionale delle Ricerche Laboratorio di Geologia Nucleare, Pisa, pp 9–130
Creux JS, Arba-del-Río R, Bergé-Nguyen M, Arsen A, Drolon V, Clos G, Maisongrande P (2016) Lake volume monitoring from space. Surv Geophys 37:269–305
Gibson J (2002) A new conceptual model for predicting isotopic enrichment of lakes in seasonal climates. Pages News 10:10–11
Gibson JJ, Birks SJ, Yi Y (2016) Stable isotope mass balance of lakes: a contemporary perspective. Quat Sci Rev 131:316–328
Gleick PH (1987) The development and testing of a water balance model for climate impact assessment: modeling the Sacramento Basin. Water Resour Res 23:1049–1061
Gonfauniti R (1986) Environmental isotopes in lake studies. In: Fritz P, Fontes JC (eds) Handbook of environmental isotope geochemistry, vol 3. Elsevier Scientific Publishing Company, Amsterdam, pp 113–168
Guo S, Wang J, Xiong L, Ying A, Li D (2002) A macro-scale and semi-distributed monthly water balance model to predict climate change impacts in China. J Hydrol 268:1–15
Guo X, Tian L, Wen R, Yu W, Qu D (2017) Controls of precipitation δ18O on the northwestern Tibetan Plateau: a case study at Ngari station. Atmos Res 189:141–151
Huss M (2013) Density assumptions for converting geodetic to water storage changes on the Tibetan Plateau 2003–2009 derived from ICESat laser altimetry measurements. Environ Res Lett 9:1–7
Phan VH, Lindenberg R, Menenti M (2012) ICESat derived elevation changes of Tibetan lakes between 2003 and 2009. Int J Appl Earth Obs Geoinf 17:12–22
Qiang M, Song L, Chen F, Li M, Liu X, Wang Q (2013) A 16-ka lake-level record inferred from macrofossils in a sediment core from Genggahai Lake, northeastern Qinghai-Tibetan Plateau (China). J Paleolimnol 49:575–590
Rouse WR (1998) A water balance model for subarctic sedge fen and its application to climatic change. Clim Change 38:207–234
Rowe HD, Guilderson TP, Dunbar RB, Southon JR, Seltzer GO, Mucciarone DA, Fritz SC, Baker PA (2003) Late Quaternary lake-level changes constrained by radiocarbon and stable isotopes on sediment cores from Lake Titicaca, South America. Glob Planet Change 38:273–290
Song C, Huang B, Ke L (2013) Modeling and analysis of lake water storage changes on the Tibetan Plateau using multi-mission satellite data. Remote Sens Environ 135:25–35
Song C, Huang B, Richards K, Ke L, Phan VH (2014) Accelerated lake expansion on the Tibetan Plateau in the 2000s: Induced by glacial melting or other processes? Water Resour Res 50:3170–3186
Steinman BA, Rosenmeier MF, Abbott MB (2010a) The isotopic and hydrologic response of small, closed-basin lakes to climate forcing from predictive models: simulations of stochastic and mean-state precipitation variations. Limnol Oceanogr 55:2246–2261
Steinman BA, Rosenmeier MF, Abbott MB, Bain DJ (2010b) The isotopic and hydrologic response of small, closed-basin lakes to climate forcing from predictive models: application to paleoclimate studies in the upper Columbia River basin. Limnol Oceanogr 55:2231–2245
Thompson LG, Mosley-Thompson E, Davis ME, Brecher HH (2011) Tropical glaciers, recorders and indicators of climate change, are disappearing globally. Ann Glaciol 52:23–34

Thompson LG, Yao T, Davis ME, Mosley-Thompson E, Wu G, Porter SE, Xu B, Lin PN, Wang N, Beauson E, Duan K, Sierra-Hernandez MR, Kenny DV (2018) Ice core records of climate variability on the Third Pole with emphasis on the Guliya ice cap, western Kunlun Mountains. Quat Sci Rev 188:1–14

Wang S, Dou H (1998) Records of lakes in China (in Chinese). Science Press, Beijing

Wu X, Huang W, Chen F (2014) Construction and application of monthly air temperature and precipitation gridded datasets with high resolution (0.025°×0.025) over China during 1951–2012 (in Chinese). J Lanzhou Univ 50:213–220

Xu H, Ai L, Tan L, An Z (2006) Stable isotopes in bulk carbonates and organic matter in recent sediments of Lake Qinghai and their climatic implications. Chem Geol 235:262–275

Xu H, Goldsmith Y, Lan J, Tan L, Wang X, Zhou X, Cheng J, Lang Y, Liu C (2020) Juxtaposition of western Pacific subtropical high on Asian summer monsoon shapes subtropical East Asian precipitation. Geophys Res Lett 47:1–10

Yang M, Nelson FE, Shiklomanov NI, Guo D, Wan G (2010) Permafrost degradation and its environmental effects on the Tibetan Plateau: a review of recent research. Earth Sci Rev 103:31–44

Yang K, Wu H, Qin J, Lin C, Tang W, Chen Y (2014) Recent climate changes over the Tibetan Plateau and their impacts on energy and water cycle: a review. Glob Planet Change 112:79–91

Yao T, Pu J, Lu A, Wang Y, Yu W (2007) Recent glacial retreat and its impact on hydrological processes on the Tibetan Plateau, China, and surrounding regions. Arctic Antarct Alp Res 39:642–650

Yao T, Thompson L, Yang W, Yu W, Gao Y, Guo X, Yang X, Duan K, Zhao H, Xu B, Pu J, Lu A, Xiang Y, Kattel DB, Joswiak D (2012) Different glacier status with atmospheric circulations in Tibetan Plateau and surroundings. Nat Clim Change 2:663–667

Yuan F, Sheng Y, Yao T, Fan C, Li J, Zhao H, Lei Y (2011) Evaporative enrichment of oxygen-18 and deuterium in lake waters on the Tibetan Plateau. J Paleolimnol 46:291–307

Zhang G (2019) The lakes larger than 1 km² in Tibetan Plateau (V2.0) (1970s-2018). Natl Tibet Plateau Data Cent

Zhang G, Yao T, Xie H, Zhang K, Zhu F (2014) Lakes’ state and abundance across the Tibetan Plateau. Chin Sci Bull 59:3010–3021

Zhang G, Yao T, Xie H, Wang W, Yang W (2015) An inventory of glacial lakes in the Third Pole region and their changes in response to global warming. Glob Planet Change 131:148–157

Zhang G, Yao T, Piao S, Bolch T, Xie H, Chen D, Gao Y, O’Reilly CM, Shum CK, Yang K, Yi S, Lei Y, Wang W, He Y, Shang K, Yang X, Zhang H (2017a) Extensive and drastically different alpine lake changes on Asia’s high plateaus during the past four decades. Geophys Res Lett 44:252–260

Zhang G, Li J, Zheng G (2017b) Lake-area mapping in the Tibetan Plateau: an evaluation of data and methods. Int J Remote Sens 38:742–772

Zhang G, Luo W, Chen W, Zheng G (2019a) A robust but variable lake expansion on the Tibetan Plateau. Sci Bull 64:1306–1309

Zhang Y, Li Y, Zhu G (2019b) The effect of altitude on temperature, precipitation and climatic zone in the Qinghai-Tibetan (in Chinese). J Glaciol Geocryol 41:505–515

Zhu L, Xie M, Wu Y (2010) Quantitative analysis of lake area variations and the influence factors from 1971 to 2004 in the Nam Co basin of the Tibetan Plateau. Chi Sci Bull 55:1294–1303

Publisher’s Note Springer Nature remains neutral with regard to jurisdictional claims in published maps and institutional affiliations.