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

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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 $km^2\ yr^{-1}$) since the late 1970s and most rapidly (0.998 $km^2\ yr^{-1}$) 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 melt water 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 ($\delta^{18}O_{carb}$) has great potential for indicating source changes of water supply in such lakes, but there is a need to be cautious when interpreting $\delta^{18}O_{carb}$ due to the influence of multiple hydrological factors, which can change in dominance over time.

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 1998 (Zhang et al. 2017a), verified both by satellite images and by ICESat altimetry measurements between 2003 and 2009 (Phan et al. 2012; Song et al. 2013). Total water storage in 312 lakes (>10 km^2) across the whole TP increased by an estimated 4.3 Gt from the early 1970s to 2000 and by 88.1 Gt between 2000 and 2011 (Song et al. 2014).

The major factors leading to lake area increases have been identified as increases in runoff generated by more precipitation (Lei et al. 2014; Yang et al. 2014), glacier melt water (Yao et al. 2007, 2012) and permafrost thaw caused by higher temperatures (Yang et al. 2010). However, the impacts of climatic drivers on lake water balance are complex, often interconnected, and variable across the interior TP. Lakes in different regions of the TP are recharged by different sources of water including rainfall, snowfall, glacier melt water and groundwater. In the central, northern, and northeastern TP, lake levels increase during the warm season and decline in the cold season, related to annual changes in monsoonal precipitation and evaporation (Lei et al. 2017). In the northwestern TP, however, lake levels increase both in spring (March to May) and in summer (June to August); this is closely linked to increased snowfall in spring and glacier melt in summer, which currently accounts for 30-40 % of total annual precipitation (Lei et al. 2017).

Water balance models based on meteorological data, are an efficient way to investigate the effects of climate change on hydrological processes for water resource planning (Gleick 1987; Guo et al. 2002; Rouse 1998; Song et al. 2014). These models have been widely used for estimating the relative importance of hydrological sources and sinks (Conway 1997; Xu et al. 2020; Zhu et al. 2010) and predicting conseque nces of future changes in streamflow (Atkinson et al. 2002). Such models have, however, only previously been applied to six lakes located in the central TP (Lei et al. 2013; Zhu et al. 2010) where hydrological controls are largely monsoonal-driven. Less is known of hydrological processes in the western TP. Lake sediment records are widely used for studying regional hydrological variations, such as lake level changes (Magny 2004; Qiang et al. 2013; Rowe et al. 2003). Combining water balance models based on instrumental datasets with palaeolimnology is a potentially powerful validation approach for understanding the drivers of lake change.

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 melt water. 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 $\delta^{18}O_{carb}$ 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 $\delta^{18}O_{carb}$. 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 km^2), the current lake water area is only 68.96

km^2 (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 km^2 (Li et al. 2017; Song et al. 2014; Wang et al. 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: 4279.3m 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 $\delta^{18}O$ 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 (http://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 (http://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 inter-annual 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 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 ^{137}Cs and ^{210}Pb using HPGe Gamma Spectrometry. ^{210}Pb was obtained via gamma-emission at 46.5 keV and ^{226}Ra at 351.92 keV γ-rays emitted by its daughter isotope ^{214}Pb. 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 hours. 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 $\delta^{18}O$ and $\delta^{13}C$ was better than 0.1 ‰. Values were reported relative to the Vienna Pee Dee Belemnite (VPDB) standard. All the measurements were carried out in the Key Laboratory of Western China's Environmental Systems, Lanzhou University.

During the field season (in July 2015) a number of lake water and precipitation samples were taken from the lake and catchment to better understand the isotope hydrology of the lake system. Lake water and a groundwater sample from a catchment spring were filtered using a syringe filter with 0.45-μm membranes and then hermetically stored in a 5 mL polyethylene bottle. Rainfall samples were also collected and sealed in a 5 mL polyethylene bottle. Falling snow was collected in a clean stainless steel bowl, which melted quickly as it was at the end of June, and the resulting water was then transferred into a 5 mL polyethylene bottle. All the water samples were stored at 4 °C before analysis. Stable isotopes of all the samples were measured by an Isotopic Liquid Water Analyzer (Picarro L1102-i) at Lanzhou University. The values are reported relative to the Vienna Standard Mean Ocean Water (VSMOW) standard. Analytical precision for $\delta^{18}O$ and $\delta^{2}H$ 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 melt water, to the observed increase in Aweng Co lake area over recent decades, we attempted to model the lake hydrology based on equation 1. We made the assumption that volume changes at Aweng Co are controlled by a number of inputs and outputs to the system (Equation 1), recognising that there is no surface outflow from the lake.

$$\Delta V_L = P_L S_L + R_C S_C + GMW + G_I - E_L S_L - G_O \qquad (1)$$

where ΔV_L is the change in lake volume (m^3) 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 melt water 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.

ΔV_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 sub-surface flow. Here we take overland flow (R_c) to be a proportion (c) of precipitation falling on the catchment.

Glacial melt water 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, $V_g = 0.042 S_g^{1.3565}$. Variation in glacier volume was converted to glacier melt water 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 melt water flowing directly into the lake.

E_L (mm day^{-1}) is calculated using the equation of Linacre (1992) such that

$$E_L = [0.015 + 4 \times 10^{-4} T_a + 10^{-6} z] \times [480 \frac{(T_a + 0.006z)}{84 - A} - 40 + 2.3u\,(T_a - T_d)] \qquad (2)$$

where T_a is air temperature (°C), z = altitude (m), A = latitude (degrees), u = wind speed (m s^{-1}), T_d = dew point temperature = $0.52T_{a\,min} + 0.60T_{a\,max} - 0.009(T_{a\,max})^2 - 2$ °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 equation 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 ≥ 0 and the regression relationship between known and modelled ΔV_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 (δ_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 δ_E is a function of δ_L, and in the case of the index lake $\delta_E = \delta_P$, δ_L can be calculated. We use the δ_E equation based on the Craig-Gordon Evaporation model (Craig and Gordon 1965), as used by Steinman et al. (2010a, b):

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

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

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

and for hydrogen

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

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 ε_k is the kinetic fraction factor; for $\delta^{18}O$, ε_k has been shown to approximate $14.2(1-h)$ and $12.5(1-h)$ for δ^2H (Gonfiantini 1986). δ_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 δ_P and δ_A varies in different environmental settings, and advocate using a measured LEL, as we have available here, to calculate the suitable regional δ_P - δ_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 equation 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.

$$P_L \delta_{PL} + R_i \delta_{\mathrm{R}i} + GMW \delta_G + G_i \delta_{Gi} = E \delta_E + G_o \delta_L \qquad (6)$$

As all isotopic values are known, equation 6 can then be optimized, varying groundwater inputs such that the equation balances for both δ^{18}O and δ^{2}H 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 km^2 to 60.69 km^2 and from 57.52×10^6 m^3 to 83.74×10^6 m^3, respectively; with a rapid lake area expansion between 1996 ~ 1999 (Fig. 2d). Before 1999, lake area and lake volume increased slowly at 0.74 km^2 yr^{-1} and 1.38×10^6 m^3 yr^{-1}, respectively. After 1999 lake area and lake volume increased reaching 69.15 km^2 and 125.02×10^6 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 136.77×10^6 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 153×10^6 m^3, and then shrank a little (Fig. 2d, 2e). The lake area and lake volume increased at a rate of 0.998 km^2 yr^{-1} and 6.29×10^6 m^3 yr^{-1} from 1999 to 2010.

Correlations (Fig. 3) between each known hydroclimate parameter and lake volume change show that precipitation and glacier melt water 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.

When calculating δ_L for the "index lake" we used a lake system where δ_E was equal to the intercept value of the LEL and Aweng Co MWL. In this case for δ_L, δ_A and δ_E to sit sensibly in $\delta^{18}O$ - δD space (Fig. 4), an adjustment, via a constant (*k*), is required to the standard equilibrium relationship between δ_P and δ_A (Gibson et al. 2016), where:

$$\delta_A = \frac{\delta_P - k\varepsilon^*}{1 + 10^{-3} \cdot k\varepsilon^*} \qquad (7)$$

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 δ^2H isotopic models (equation 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 melt water 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 %.

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 $\delta^{18}O_{carb}$ 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 $\delta^{18}O_{carb}$ value of −1.24 ‰) around the mid-2000s and a positive trend after ~2007 (Fig. 2g).

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 melt water and changing precipitation amount could be controlling the observed lake area change.

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 melt water, 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 melt water in the groundwater entering Aweng Co. The most negative of the precipitation samples collected in the 2015 field season ($\delta^{18}O$ = −13.86 ‰; $\delta^{2}H$ = −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 ‰, $\delta^{2}H$ = −99.77 ‰), it appears likely that this groundwater is dominated in composition by snow and glacier melt water (~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 snow melt.

For the differing inflow parameters, given the location of the groundwater sample and average precipitation in $\delta^{18}O$ $\delta^{2}H$ space (Fig. 4), if ~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 melt water, 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 melt water 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.

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

The variation of $\delta^{18}O_{carb}$ 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 $\delta^{18}O_{carb}$ sediment record. The mean summer temperature change rise of 1.1°C (Fig. 2b) would lead to $\delta^{18}O_{carb}$ change of ~0.26 ‰ based on a temperature-dependence of carbonate fractionation of −0.24 ‰/°C (Craig 1965), which is not enough to explain the magnitude of $\delta^{18}O_{carb}$ 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 $\delta^{18}O_{carb}$. 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 $\delta^{18}O_{carb}$ at Aweng Co. Of particular interest through recent decades is the negative excursion in $\delta^{18}O_{carb}$ 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 $\delta^{18}O_{carb}$ record.

Comparison of trends in precipitation (Fig. 2a) with the $\delta^{18}O_{carb}$ 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 $\delta^{18}O_{carb}$ shift to more negative values (−1.24 ‰ in 2006). Given I:E ratio is likely the main driver of $\delta^{18}O_{carb}$ change, an increased amount of glacier melt water would increase lake area/volume and lead to a negative shift in $\delta^{18}O_{carb}$.

Although the $\delta^{18}O_{carb}$ 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 $\delta^{18}O_{carb}$ in other lake systems (Jones et al. 2007), remains the same, while volume has increased due to the elevated glacial melt water 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 melt water, rather than the amount of it, such that the impact of this input changed the $\delta^{18}O_{carb}$ 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 melt water led to this negative shift in $\delta^{18}O_{carb}$.

This comparison exercise shows how even with instrumental data available for contrast, the interpretation of $\delta^{18}O_{carb}$ records is complicated by the multiple potential controls that can lead to an abrupt change in a core $\delta^{18}O_{carb}$ 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.

δD

Conclusions

The combined monitoring, modelling and palaeolimnological approach taken here shows the potential for $\delta^{18}O_{carb}$ 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 $\delta^{18}O_{carb}$ evidence point to the importance of glacial melt water 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 melt water change can be different. The flux of water through the lake system, controlled by precipitation amount and evaporation as well as glacial melt water, is also therefore important in driving the resulting $\delta^{18}O_{carb}$ record preserved in the sediments, and the dominant hydrological controls may change through time.

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Tables

Table 1 Annual values for the parameters of the water balance model from 10/1999 to 10/2009. P_L is the precipitation on the lake surface (Wu et al. 2014). S_L is the lake surface area (average of the area at the start and end of the time period). S_C is the area of catchment excluding the lake area. E_L is the evaporation from the lake surface. LVC is the lake volume change in each period

start	end	P_L (mm)	S_L (m^2)	S_C(m^2)	E_L (mm)	LVC (m^3)
10/1999	10/2000	199	63897313	1988402687	916	27017800
10/2000	10/2001	165	67175062	1985124938	979	1161000
10/2001	10/2002	205	68197610	1984102390	914	13097000
10/2002	10/2003	149	68967180	1983332820	886	−10094000
10/2003	10/2004	130	67938056	1984361944	997	−5669000
10/2004	10/2005	159	66809354	1985490647	942	−2652000
10/2005	10/2006	166	67833813	1984466187	962	21009000
10/2006	10/2007	157	69168096	1983131905	984	384000
10/2007	10/2008	205	70129135	1982170865	975	8777000
10/2008	10/2009	125	70121542	1982178458	1006	−14563000

Table 2 The results of changing model constants *c* and *g* to optimize r^2, slope and intercept of the relationship between known and modelled ΔV_L

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 Estimates of contributions of different hydrological components of the Aweng Co system from the isotope mass balance calculation (Equation 6)

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

Figure captions

Fig.1 The location of Aweng Co (a), the topography of Aweng Co catchment (b) and the monthly mean temperature and total monthly precipitation at Shiquanhe Station (c). The boundary of the inner TP (a) was defined according to Zhang et al. (2015).

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).

Fig. 3 The linear correlation between each parameter (left to right: precipitation on the lake surface; evaporation on the lake surface and glacier melt water) 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.

Fig. 4 $\delta^{18}O$ vs. $\delta^{2}H$ of different waters from Aweng Co. Blue dots are lake water from Aweng Co. Green Rhombus is the groundwater from the Aweng Co basin. Light blue crosses are lake water isotopes from the western Tibetan Plateau (Yuan et al. 2011). Small and large dark blue triangles are sampled precipitation and mean precipitation in the Aweng Co region respectively. Yellow triangle is the estimated isotopic value of the air vapour over the lake, black cross is the isotope value of the calculated evaporation from the lake surface, orange square is the calculated regional index lake (see text for details). MWL is the meteoric water line from Ngari station (Guo et al. 2017). Aweng Co MWL is the local meteoric water line. LEL is the local evaporation line. The isotope data used in this study are shown in ESM 5.

Figures

Fig. 1

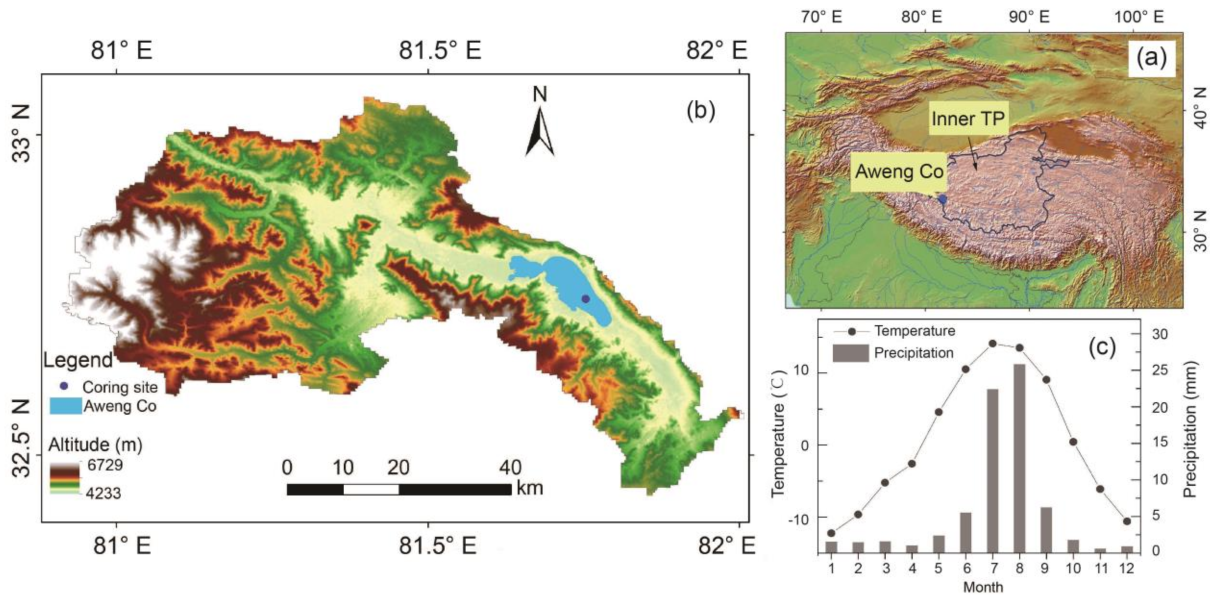
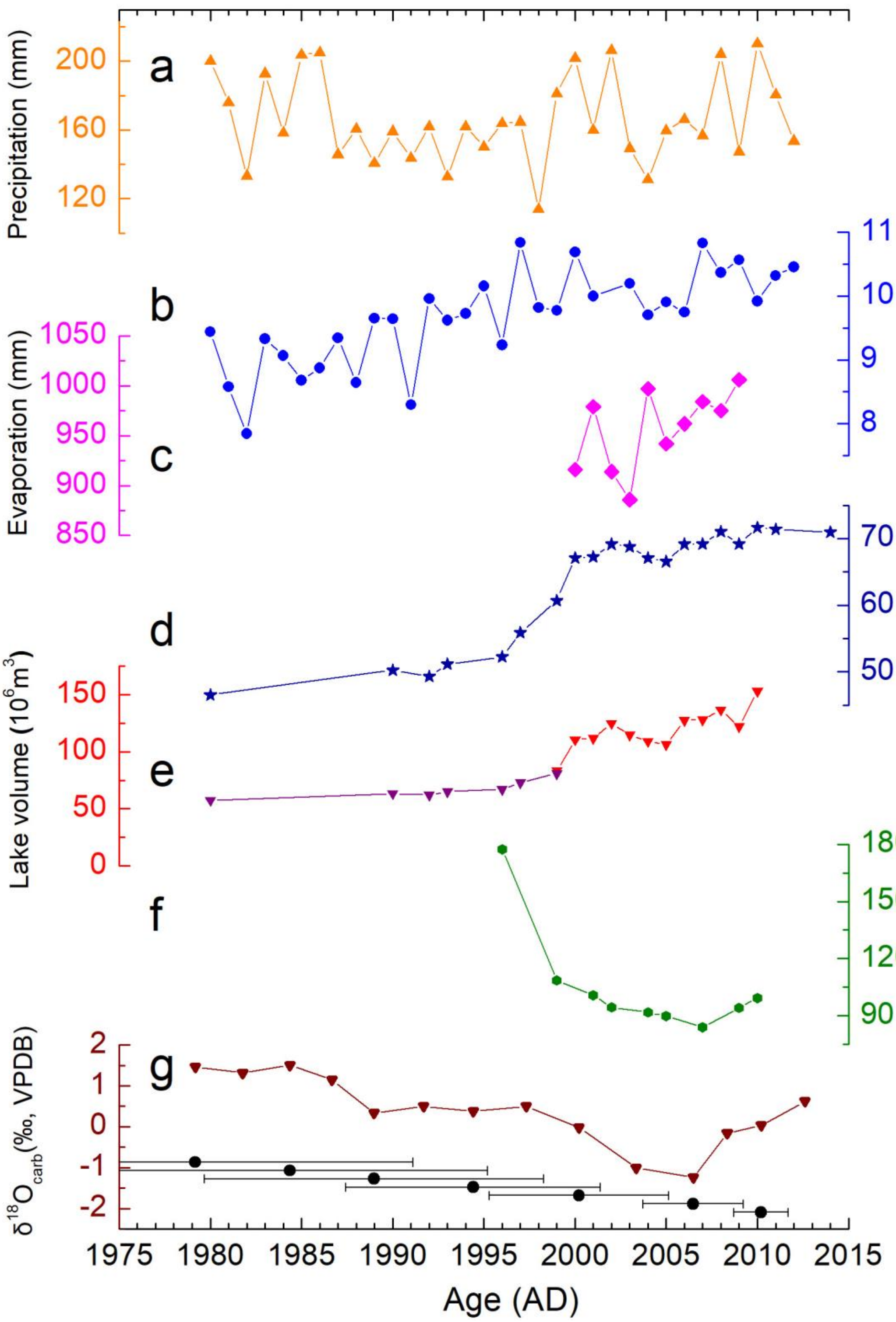
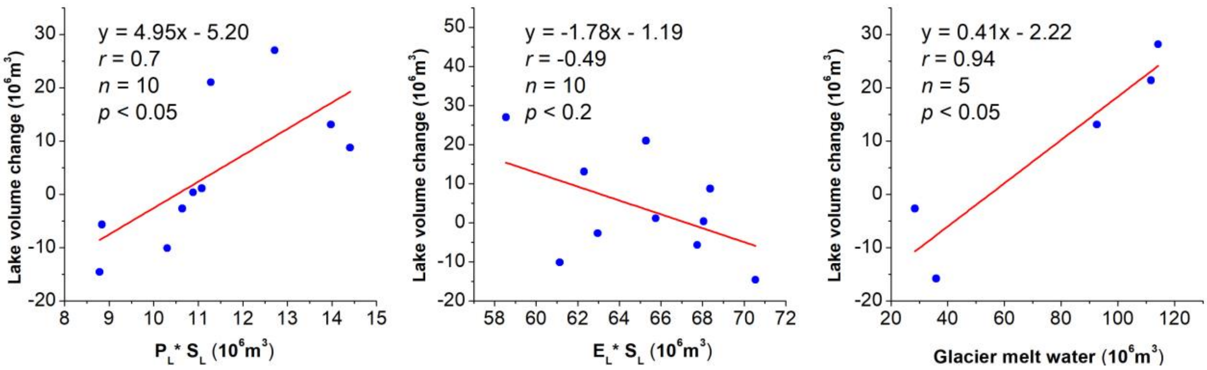


Fig. 2



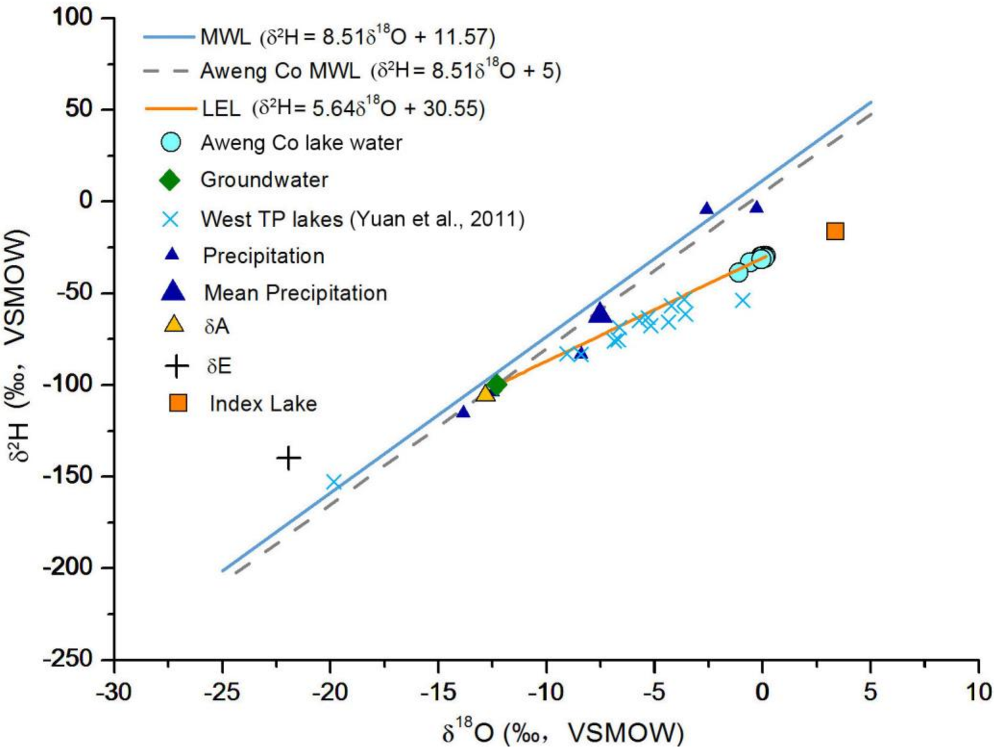
a
b
c
d
e
f
g
Precipitation (mm)
200
160
120
Temperature (°C)
11
10
9
8
Evaporation (mm)
1050
1000
950
900
850
Lake area (km^2)
70
60
50
Lake volume (10^{6}m^3)
150
100
50
0
Glacier area (km^2)
180
150
120
90
$\delta^{18}O_{carb}$(‰, VPDB)
2
1
0
-1
-2
1975
1980
1985
1990
1995
2000
2005
2010
2015
Age (AD)

Fig. 3



y = 4.95x - 5.20
r = 0.7
n = 10
p < 0.05
Lake volume change (10^6m^3)
P_L* S_L (10^6m^3)
y = -1.78x - 1.19
r = -0.49
n = 10
p < 0.2
Lake volume change (10^6m^3)
E_L* S_L (10^6m^3)
y = 0.41x - 2.22
r = 0.94
n = 5
p < 0.05
Lake volume change (10^6m^3)
Glacier melt water (10^6m^3)

Fig. 4


MWL (δ²H = 8.51δ¹⁸O + 11.57)
Aweng Co MWL (δ²H = 8.51δ¹⁸O + 5)
LEL (δ²H= 5.64δ¹⁸O + 30.55)
Aweng Co lake water
Groundwater
West TP lakes (Yuan et al., 2011)
Precipitation
Mean Precipitation
δA
δE
Index Lake
δ²H (‰, VSMOW)
δ¹⁸O (‰, VSMOW)
100
50
0
-50
-100
-150
-200
-250
-30
-25
-20
-15
-10
-5
0
5
10