Different facets of dry/wet patterns in southwestern China over the past 27,000 years

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Abstract. Frequently occurring mega-droughts under current global climate change have attracted broad social attention. A paleoclimatic perspective is needed to increase our understanding of the causes and effects of droughts. Southwestern (SW) China has been threatened by severe seasonal droughts. Our current knowledge of millennial-scale dry and wet phases in this region is primarily based on the variability of the Indian Summer Monsoon. However, water availability over land does not always follow patterns of monsoonal precipitation but also depends on water loss from evaporation and transpiration. Here, we reconstructed precipitation intensity, lake hydrological balance and the soil water stress index (SWSI) for the last 27,000 yr. Grain size, geochemical and pollen records from Yilong Lake reveal the long-term relationships and inconsistencies of dry/wet patterns in meteorological, hydrological and soil systems in central Yunnan region, SW China. Our results show that the long-term trends among precipitation, hydrological balance and soil moisture varied through time. The hydrological balance and soil moisture were primarily controlled by temperature-induced evaporation change during periods of low precipitation such as the Last Glacial Maximum and Younger Dryas. During periods of high precipitation (the early to late Holocene), intensified evaporation from the lake surface offset the effects of increased precipitation on the hydrological balance. However, abundant rainfall and the dense vegetation canopy circumvented a soil moisture deficit that might have resulted from rising temperature. In conclusion, the hydrological balance in central Yunnan region was more sensitive to temperature change while soil moisture could be further regulated by vegetation changes over millennial timescales. Therefore, under future climate warming, the surface water shortage in central Yunnan region may become even more serious. Our study suggests that reforestation efforts may provide some relief to soil moisture deficits in this region.

1. Introduction

The global land area experiencing extreme-to-exceptional terrestrial water storage drought could more than double by the late twenty-first century (Pokhrel et al., 2021). In southwestern (SW) China, drought has become a climate threat which is likely to happen more frequently in the future (Qiu, 2010; Wang et al., 2016). It is generally thought that long-term dry and wet phases in SW China are primarily regulated by the intensity of monsoonal precipitation associated with the evolution of atmospheric circulation systems (Chen et al., 2014; Hillman et al., 2017; Sun et al., 2019; Wang et al., 2019). However, drought refers to the amount of water available in both the soil and hydrological systems which is dependent on precipitation and a range of other factors. These include how much water is able to infiltrate to deeper ground layers or is lost as runoff, and how much is evaporated directly from water and soil surfaces, or transpired by plants (Breshears et al., 2005; Dai et al., 2018; Feng and Liu, 2015; Mishra and Singh, 2010; Trenberth et al., 2014). Therefore, drought doesn't only happen during periods of low precipitation (Dai et al., 2018; Sun et al., 2017; Trenberth et al., 2014; Xu et al., 2019). Given this, understanding dry/wet patterns in different climate scenarios while considering physical and vegetation processes is crucial to predicting the future risk of drought.

Climate evolution in SW China since the Last Glacial Maximum (LGM) has been reconstructed using various types of paleoclimatic archives, such as speleothem oxygen isotope records (e.g., Cai et al., 2015; Dykoski et al., 2005; Zhao et al., 2015), lake sediments (e.g., Hillman et al., 2017; Hillman et al., 2020; Hodell et al., 1999; Li et al., 2018; Sun et al., 2019; Wang et al., 2020; Wu et al., 2018; Xiao et al., 2014a; Xiao et al., 2014b; Zhang et al., 2017; Zhang et al., 2019a), and peats (e.g., Huang et al., 2016; Wei et al., 2012). The major conclusion of these studies is that monsoonal precipitation and temperature was very low during the LGM, precipitation peaked in the early Holocene during a period of warmer climate, before both precipitation and temperature declined. The "cold-dry" and "warm-humid" paradigms of climate change in SW China have been widely demonstrated from a paleoclimatological perspective, despite the proposition that summer temperature and monsoon precipitation were decoupled in the early Holocene (Wu et al., 2018). Additionally, vegetation in SW China has experienced noticeable changes since the LGM (Chen et al., 2014; Cook et al., 2011; Wu et al., 2018; Xiao et al., 2014a; Xiao et al., 2014b), which may have affected evapotranspiration processes. Since water availability is a trade-off among precipitation input and water loss through evaporation, transpiration and outflow (Breshears et al., 2005; Dai et al., 2018; Watras et al., 2014), these processes may lead to different environmental signatures of the wet and dry patterns over land. However, to our knowledge few studies have tested this idea.

Definitions of ‘‘drying” in meteorology, hydrology and biology are different but connected (Mishra and Singh, 2010). In paleoclimatology, drying or moistening processes can be reconstructed by different types of proxies, but the concepts underpinning them are not necessarily the same. For example, grain-size records from several lakes in SW China reflected variations in Indian Summer Monsoon (ISM) precipitation (Ning et al., 2017; Peng et al., 2019; Sheng et al., 2015). Authigenic carbonate precipitation is strongly affected by the hydrological balance or precipitation/evaporation ratio (Leng and Marshall, 2004; Ohlendorf et al., 2013), and thus indicates changing hydrological conditions in lake-catchment systems. For terrestrial ecosystems, soil water is the main and direct source of water for most plants and a primary constraint for vegetation composition and biomass. Consequently, changes in the soil moisture can theoretically be reflected in pollen records. As yet, soil moisture status has rarely been reconstructed using pollen data, and the similarities and differences between the dry/wet patterns revealed from different paleoproxies are seldom discussed.

The Yunnan region is located in the SW of China (Fig. 1). It is primarily influenced by warm and humid airflow from the Bay of Bengal in summer. Several paleolimnological studies have shown that, since the LGM or the Holocene, precipitation, hydrological conditions and vegetation in the Yunnan region has experienced noticeable changes (Hillman et al., 2017; Hillman et al., 2020; Hodell et al., 1999; Ning et al., 2017; Sun et al., 2019; Wu et al., 2018; Xiao et al., 2014a). However, few studies have applied a multi-proxy approach at the same temporal scale to explore different aspects of the long-term dry/wet patterns. Here we have developed the first record of soil water stress index (SWSI) based on a pollen record from Yilong Lake in the Yunnan region to reveal changes in soil moisture over the past 27,000 years. By comparing the SWSI reconstruction with records of monsoonal precipitation and hydrological balance in the same core, we aim to discuss the long-term relationships of dry/wet patterns in the meteorological, hydrological and soil systems in SW China.

2. Materials and methods

2.1 Study site and modern climate

Yilong Lake (23.63–23.70° N, 102.50–102.65° E, 1414 m a.s.l.) is a faulted lake in central Yunnan region, SW China (Fig. 1a). It covers an area of 38 km^2 and has a catchment area of 303.6 km^2 (Wang and Dou, 1998). The average water depth is 2.8 m and the maximum is 6.2 m (Wang and Dou, 1998). Yilong Lake is a freshwater lake which is fed by overland runoff, lake surface precipitation and groundwater. All the inflowing rivers, except the Chenghe River at the northwest of the Lake, are seasonal (Fig. 1b). An outlet was located at the southeast end of the lake during past highstands, but it disappeared in 1978 CE due to climate- and human-induced lower lake levels (Wang and Dou, 1998).

The central Yunnan region is dominated by a subtropical monsoon climate. Observations from the closest meteorological station in Yuxi City (24.2° N, 102.338° E, 1717 m a.s.l.), 80 km away from Yilong Lake, record a mean annual temperature of 18.1 °C after lapse rate correction (0.65 °C per 100 m) and mean annual precipitation of 886 mm (1951–2017 CE; China Meteorological Data Service Centre, https://data.cma.cn/en). Annual evaporation (1035 mm) in this region is slightly higher than the annual precipitation. Seasonal climate change is characterized by high precipitation in the warm season with low precipitation in the cold season (Fig. 1c).

2.2 Sampling and dating

In May 2017, a sediment core (YLH) was retrieved from Yilong Lake at a water depth of 4.1 m (Fig. 1b) using a UWITEC sampling system. The sediment core was transported to the School of Geography, Nanjing Normal University laboratory and kept at 3.9 °C until analysis. The sediment core was split with a Geotek Core Splitter, photographed and visually described in the laboratory. The samples were subsampled at 1-cm intervals, freeze-dried, and used for further analyses.

The age of the core YLH was determined using fifteen accelerator mass spectrometry (AMS) ^{14}C dates including bulk organic matter, charcoal and plant remains (Table 1). ^{14}C dates were analysed by the Beta Analytic Testing Laboratory. The conventional ^{14}C dates were calibrated using Calib 8.20 with an IntCal20 calibration dataset (Reimer et al., 2020). Age-depth modelling was performed in R (version 3.4.4, R Core Team, 2018) using the package "rbacon" (Blaauw et al., 2020). According to the age-depth model, the basal sediment was deposited between 27.171–26.439 cal. ka BP with a median age of 26.866 cal. ka BP (Fig. 2).

2.3 Analytical methods

Samples for grain size analysis were measured at 1-cm intervals. All samples were pretreated with 30 % H_2O_2 and then 5 % HCl to remove organic matter and carbonates, rinsed with deionized water to neutralize the samples, and placed in an ultrasonic bath with 10 % $(NaPO_3)_6$ added to disperse particles. The grain size distribution was measured by a Malvern 3000E laser diffraction instrument with 100 bins ranging from 0.02 to 2000 μm.

The sediments (collected at 1-cm intervals) were pretreated with 10 % HCl to remove carbonates and then used to measure the total organic carbon (TOC) and total nitrogen (TN) using a vario EL cube Elemental Analyzer. Replicate analyses of well-mixed samples showed that the precision was ca.± <0.1 % (1 standard deviation (s.d.)). Continuous down-core X-ray fluorescence (XRF) measurements of the geochemical composition were carried out with a core scanner (MSCL-S Specifications; tube voltage 15 kV, exposure time 30 s and spatial distribution 0.5) at the laboratory of the School of Geography, Nanjing Normal University. The sample moved along a monochromatized and polarized SR beam. Core scanning started from 10 cm depth as the upper 10 cm of sediments has a high water content and are too soft to get robust measurements. Core bulk mineralogy of freeze-dried and milled samples (at 2-cm intervals) were measured by X-ray diffraction (XRD) using a PANalytical X'pert Pro (40 kV, 30 mA, from 5 to 80°, step-rate 0.0167°, Cu kα radiation) at the Qinghai Institute of Salt Lakes, Chinese Academy of Science (CAS). A total of 71 freeze-dried samples from depths where carbonate content is (relatively) high (10–176 cm and 228–280 cm) were measured for oxygen stable isotopes in carbonate ($^{18}O_{carb}$) using a Thermo-Fisher MAT 253 mass spectrometer equipped with a Kiel-IV carbonate preparation device at the Nanjing Institute of Geography and Limnology, CAS. The samples were pretreated with 100 % phosphoric acid. The $^{18}O_{carb}$ values are expressed as standard delta (δ) notation as the per mil (‰) deviation from Vienna Pee Dee Belemnite (VPDB). Four samples at different depths were imaged using a Hitachi SU8010 scanning electron microscope (SEM) to examine crystal structure and morphology of carbonate minerals.

Samples for pollen analysis were analysed at 4-cm intervals and treated standard laboratory methods (Faegri et al., 1989), including treatment with 10 % HCl and 50 % HF to remove carbonates and silicates, boiling in 10 % KOH to remove humic acid, sieving to remove the fine and coarse fractions, and mounting in silicone oil. To calculate the concentrations of pollen, tablets containing a known quantity of *Lycopodium* spores were added to each sample prior to the treatments. At least 300 terrestrial pollen grains per sample were counted. All the treatment and identification was processed in the Institute of Hydrogeology and Environmental Geology, Chinese Academy of Geological Sciences. The pollen percentages were calculated based on the total number of pollen grains from terrestrial pollen taxa and used to construct a pollen diagram and conduct numerical analyses.

2.4 Pollen-based quantitative reconstruction of SWSI

In total, 1394 surface soil pollen samples from SW China (Ni et al., 2014; Fig. S1 in the Supplement) were used in this study. Annual climate data was averaged from long-term records from 1971 to 2000 at 1814 meteorological stations across China (China Meteorological Data Service Centre, https://data.cma.cn/en). These data were interpolated into 1 km grid cells using a thin plate smoothing spline surface fitting technique (Hancock and Hutchinson, 2006) that takes the impact of elevation into account on the basis of the Shuttle Radar Topography Mission (SRTM) digital elevation model (Farr et al., 2007). The interpolation was performed in the program ANUSPLIN version 4.36 (Hutchinson, 2006). The interpolated meteorological data were used to calculate the SWSI using the SPLASH v.1.0 program (Davis et al., 2017). SWSI reflects the degree of evapotranspiration deficit and is expressed as (PET-AET)/PET, where AET and PET are the annual sums of actual and potential evapotranspiration, respectively (Prentice et al., 1993). The SWSI reconstruction was calculated using the weighted-averaging partial least square (WA-PLS) regression (ter Braak and Juggins, 1993). The pollen percentages were square-root transformed to stabilize variances and optimize the signal-to-noise ratio (Prentice, 1980). The 'leave-one-out' cross-validation was used to test the reliability and robustness of the model. The number of components to include in the transfer function was selected as those producing the lowest root mean squared error of prediction (*RMSEP*), a high coefficient of determination between observed and predicted environmental values (r^2) and low average bias (ter Braak and Juggins, 1993). These analyses were carried out using R package "rioja" (Juggins, 2017).

3. Results

3.1 Grain size and geochemical data of core YLH

Grain size in core YLH are mainly composed of clay (< 4 μm) and silt (4–63 μm), with mean contributions of 25 % and 70.2 %, respectively. Down-core variations of the clay component exhibit generally high values during ca. 16–27.5 cal. ka BP (ca. 246–456 cm) and ca. 0–3 cal. ka BP (ca. 0–40 cm), and low values with strong fluctuations during ca. 3–16 cal. ka BP (ca. 41–245 cm) (Fig. 3). The variations of the silt component are generally opposite to that of the clay component. The sand component is characterized by low percentages throughout the core (average percentage of 4.8 %), punctuated by two intervals (ca. 3–5 cal. ka BP and ca. 13–16 cal. ka BP) with a slight increase (average ca. 8–10 %) (Fig. 3). The median size was small during ca. 15–27.5 cal. ka BP (ca. 235–456 cm) and ca. 0–2 cal. ka BP (ca. 0–26 cm). In the period of 2–15 cal. ka BP, the median size was relatively big but around 11 cal. ka BP, it is as small as that in other periods (Fig. 3).

TOC within the samples varies between 0.75 % and 15.13 %, and TN varies from 0.08 % to 2.01 % (Fig. 3). TOC increases from 9 % to 15 % between 27.5–20 cal. ka BP and then decreases to 5 % around 12–9 cal. ka BP. A sudden increase in TOC to 10–14 % occurs after

9 cal. ka BP and then decreases sharply to 0.75 % between 5–3 cal. ka BP before increasing again (Fig. 3). Variations in TN and TOC are highly synchronous (Fig. 3). The C/N ratio is around 10 during 27.5–13 cal. ka BP and higher than 10 during 8–7 and 5–3 cal. ka BP, with lower than 10 during 13–8, 7–5 and 3–0 cal. ka BP (Fig. 3). The Fe/Mn ratio peaks around 13–9 cal. ka BP (Fig. 3).

XRD detects quartz, calcite, aragonite, magnetite, muscovite, gypsum, rhodochrosite, dolomite, clinochlore as major minerals in the core sediments. Carbonate in the core appears primarily in two sections, between 0–180 cm and 226–288 cm (Fig. 3). Aragonite exists only in the upper 34-cm sediments, which partially compensates for the decrease of calcite (Fig. 3). The $\delta^{18}O_{carb}$ varies from -8.435 ‰ to -2.525 ‰ during 18 to 14.5 cal. ka BP and from -10.472 ‰ to -4.371 ‰ during 8.5–0.77 cal. ka BP (Fig. 3). $^{18}O_{carb}$ enriched consistently since 8.5 cal. ka BP (Fig. 3).

3.2 Pollen assemblage variations and reconstructed SWSI over the past 27,000 years

A total of 99 pollen taxa were identified in 114 samples. Pollen from tree taxa, including *Pinus*, *Picea*, evergreen and deciduous *Quercus* (*Quercus.*Eve and *Quercus.*Dec), *Betula* and *Tsuga*, dominated the entire pollen record, with an average percentage of 85.6 % (Fig. 4). The shrub taxa only accounts for 1.08 % of the entire pollen record. The average percentage of the herbaceous taxa is 13.3 %. Among the herb taxa, only *Artemisia*, Poaceae and Asteraceae have average percentages >1 %. Herbs increased abruptly for several brief time intervals between 8−6 cal. ka BP and between 3−2 cal. ka BP (Fig. 4). The most noticeable change in the tree pollen composition occurred around 13 cal. ka BP, when *Quercus*.Dec was replaced rapidly by *Quercus*.Eve and *Betula*, and the coniferous taxa (*Abies*, *Picea*, *Pinus* and *Tsuga*) almost disappeared (Fig. 4). Another obvious change in the pollen composition occurred at 3 cal. ka BP, when *Pinus*, *Artemisia* and Poaceae increased considerably, and *Quercus*.Eve, *Quercus*.Dec, *Betula* and *Alnus* decreased. The average percentage of herbs was relatively low compared with woody plants throughout the core. But the abundance of herbaceous pollen increased noticeably between 8−6 cal. ka BP and especially after 3 cal. ka BP, with the maximum percentage more than 90 %.

A two-component WA-PLS model was selected on the basis of high r^2 (0.62), low *RMSEP* (0.159) and the smallest number of 'useful' components (Table S1 in the Supplement). High (low) SWSI values indicates large (small) water stress in soil system. The reconstructed SWSI ranges from 0 to 0.29, reflecting low to moderate water stress over the past 27,000 years. Relatively high SWSI occurred in three periods: 20−15 cal. ka BP, 8−6 cal. ka BP and 4−0 cal. ka BP (Fig. 4).

4. Discussion

4.1 Precipitation change revealed by grain size

The grain size composition of lake sediments contains information on the sources of clastic materials, lake level fluctuations and transport energy related to variations in runoff (Peng et al., 2005; Xiao et al., 2013). A previous study on the spatial characteristics of grain-size distributions in the surface sediments of Yilong Lake revealed that, as the water depth decreased, the median size increased and the grain-size distribution curve changed from "unimodal" to "bimodal" (Zhang et al., 2019b). Yilong Lake is located in the realm of the ISM, where the source of clastic materials (especially coarse particles) and the transport energy are primarily controlled by precipitation. An increase in precipitation not only enhances the erosion intensity of the basin, but also increases the runoff, which will lead to the transportation of more coarse particles into the lake. In the core YLH, most of the samples with a relatively large median grain size show a "unimodal" distribution (Fig. S2 in the Supplement), which implies a relatively high lake level and an increase in precipitation-transported coarse particles into the central part of the lake. Consequently, an increase in the sand component and median grain size coarsening should relate to intensified hydrological energy under increased precipitation.

Carbonate deposited from the lake water was assumed to preserve the $\delta^{18}O$ signal of precipitation and hence the $\delta^{18}O_{carb}$ from lake sediments is a good proxy for reflecting precipitation intensity or ISM in the Yunnan region (Hillman et al., 2017; Hillman et al., 2020; Hodell et al., 1999; Sun et al., 2019). A moderately strong negative correlation between precipitation $\delta^{18}O$ and monthly precipitation amount at Kunming, SW China, has been reported

(Hillman et al., 2017). Therefore, the assumption that the $\delta^{18}O$ of the water in Yilong Lake was controlled by the $\delta^{18}O$ of precipitation is reasonable. The changes in the median grain size in core YHL closely match that of $\delta^{18}O_{carb}$, with the small grain size corresponding to high $\delta^{18}O_{carb}$ values, and vice versa (Fig. 5). This supports the theory that the median grain size is a reliable indicator of precipitation intensity in Yilong Lake. However, it should be noted that grain-size data from the samples in the last 3,000 years cannot be simply interpreted as changes in precipitation intensity because human activities strongly affected the regional landscape and freshwater systems during this period (Wu et al., 2015; Xiao et al., 2017).

Variations of the median grain size from core YLH reflect less monsoonal precipitation during ca. 27–15 cal. ka BP and generally high precipitation between ca. 15–3 cal. ka BP (Fig. 5). This pattern is generally consistent with many lines of evidence from stalagmites (Cai et al., 2015; Cheng et al., 2016; Dykoski et al., 2005; Zhao et al., 2015) and lake sediments (Hodell et al., 1999; Peng et al., 2019; Sun et al., 2019). Additionally, variations of the median grain size also record the well-known climatic events Bølling/Allerød (B/A) and Younger Dryas (YD), shown respectively as a sharp increase and decrease in the monsoonal precipitation (Fig. 4). The occurrence time of the B/A event recorded in this study is consistent with that documented in the stalagmite $\delta^{18}O$ records from Cave Dongge (Dykoski et al., 2005), Sanbao (Cheng et al., 2016) and Xiaobailong (Cai et al., 2015) (Fig. 5). However, the YD event in this study lagged ca. 1,000 years behind that recorded in the stalagmite $\delta^{18}O$ from Cave Dongge (Dykoski et al., 2005) and Sanbao (Cheng et al., 2016). This delayed monsoon signal is recorded in the stalagmite $\delta^{18}O$ record from Xiaobailong Cave (Cai et al., 2015) which is only ca. 85 km away from Yilong Lake. Understanding the mechanisms driving this lag is beyond the scope of this study, but uneven rainfall distribution due to the complex topography of the Yunnan-Guizhou Plateau likely played an important role.

4.2 Carbonate deposition as a record of past hydrological balance

Carbonate deposition in Yilong Lake is mainly composed of autochthonous calcite and aragonite (Fig. S3 in the Supplement). A large number of mollusk shells are the main source of the aragonite that only appears in the upper 34-cm of the core. Autochthonous calcite precipitation needs a certain degree of supersaturation (Raidt and Koschel, 1988) which can be achieved by seasonal temperature increase and plankton flourishing (Robbins and Blackwelder, 1992; Stabel, 1986). ^{18}O measurements of the lake water and precipitation indicated significant evaporative effects in Yilong basin during the warm season (Whitmore et al., 1997). Increased temperature exerts a direct control on water evaporation, which will concentrate dissolved carbonates and hence facilitate carbonate precipitation. Photosynthesis of plankton influences carbonate precipitation by affecting CO_2 and pH in the epilimnetic water and/or providing nucleation sites for crystallization (Robbins and Blackwelder, 1992; Stabel, 1986). Assuming that lake primary productivity played a big role in carbonate precipitation, algae productivity should have increased during 20–14 cal. ka BP when carbonate content increased, and decreased during 14–9 cal. ka BP when carbonate content decreased. But in fact, relatively high C/N values between 20–14 cal. ka BP and low values between 14–9 cal. ka BP indicate relatively low and high algae productivity, respectively. In addition, characteristics of the crystal morphology of calcite also indicate abiotic origins (Fig. S3 in the Supplement). Consequently, lake primary productivity is not a main cause for carbonates precipitated in Yilong Lake. The carbonate content in the lake sediments should reflect the hydrological balance (i.e. P-E).

The variations of the carbonate content in the lake sediments (Fig. 3) indicate a relatively positive water balance between 27–20 cal. ka BP and 14–9 cal. ka BP, and a negative water balance between 20–14 cal. ka BP and 9–0 cal. ka BP. Carbonate records from the two nearby shallow Lakes Xingyun and Qilu (Fig.1, Hillman et al., 2017; Hillman et al., 2020; Hodell et al., 1999) show broadly similar patterns to this record, indicating that the hydrological changes reconstructed in this study reflect a regional rather than a local climate signal.

4.3 Different patterns of long-term changes in precipitation, hydrological balance and SWSI, and the potential mechanisms

Precipitation is the primary source of water for lacustrine systems and soils. It is generally assumed that dry conditions occurred when precipitation was low, and wet conditions prevailed when precipitation was high. From this point of view, we may conclude that our study region experienced “dry” conditions during the cold LGM and a “wet” phase during the warm Holocene. This pattern generally agrees with other paleoclimate reconstructions from the Yunnan region (e.g., Cheng et al., 2016; Xiao et al., 2014b). However, it does not match well with the hydrological balance and SWSI reconstructions from the same core. This study finds that during the LGM (27–20 cal. ka BP) and the YD event, precipitation was low but the water balance of Yilong Lake was positive and the soil moisture was relatively high. During the early to late Holocene (9–3 cal. ka BP), precipitation was high but the water balance was negative. The definition of dryness is the shortage of available water, which is fundamentally a trade-off between water input and output (Breshears et al., 2005; Dai et al., 2018; Mishra and Singh, 2010; Trenberth et al., 2014; Watras et al., 2014). Therefore, factors controlling water loss from terrestrial systems can also play a large role in the creation of dry conditions.

Evaporation is one of the important pathways by which water is lost from terrestrial environments. The evaporation rate is higher at higher temperatures because the average kinetic energy of the water molecules increases and more can fly off the surface. Our pollen record revealed that *Quercus*.Dec and the coniferous taxa were rapidly replaced by *Quercus*.Eve and *Betula* around 13 cal. ka BP (Fig. 3). According to the Vegetation Map of the People’s Republic of China (The Editorial Committee of Vegetation Map of China, CAS, 2007), deciduous broad-leaved forest is distributed in areas with lower temperature compared to evergreen broad-leaved forest. Therefore, the characteristics of the pollen compositional change during the transition from glacial to post-glacial period reflects rising temperature. The LGM was characterized by low temperature under which the simulated annual-mean potential evapotranspiration decreased by 10–40 % over nearly all land (Scheff et al., 2017). Decreasing evaporation is a major driver of increasing effective summer moisture availability which has a profound influence on the hydrological states of lakes (Aichner et al., 2019). These factors can explain the positive hydrological balance and relatively wet soils reconstructed during the LGM. The precipitation during 20–15 cal. ka BP was as low as the previous period while insolation was increasing (Fig. 5). Our pollen record showed an obvious decrease in *Picea*, indicating a warmer climate during 20–15 cal. ka BP than in the previous period. This period corresponds to the late glacial warming which is widely reported in the Yunnan region (Fig. 5, Wang et al., 2020; Xiao et al., 2014b; Zhang et al., 2019a). The increased temperature would intensify evaporation and lead to water deficiency in lacustrine and soil environments. Therefore, the hydrological and soil drying during 20–15 cal. ka BP should be triggered by increasing temperatures. However, in spite of the high temperature, soil moisture was still relatively high during 9–3 cal. ka BP (Fig. 5). Water loss from soil systems is more complex than that from free water surfaces because both temperature and the underlying surface conditions affect the ways and the amount of soil water loss or soil water storage capacity (Maxwell and Condon, 2016; Zhang and Schilling, 2006).

Water is distributed across the landscape during a rainfall event, with precipitation intercepted by foliage, stored in the soil profile or collected in groundwater aquifers. The amount of water stored in soil profiles and aquifers can be regulated by plant processes (Guzha et al., 2018; Mohammad and Adam, 2010). Previous studies have shown that, both at a stand level and at a global scale, plant transpiration accounts for the largest portion of the total evapotranspiration (61 % ± 15 % s.d.–64 % ± 13 % s.d., Good et al., 2015; Maxwell and Condon, 2016; Schlesinger and Jasechko, 2014). Our results show that the pollen concentration was high during 9–3 cal. ka BP. Although there isn’t an obvious linear correlation between pollen concentration and vegetation cover (Luo et al., 2009; Xu et al., 2007), a three-fold increase in the average pollen concentration from the late Pleistocene to Holocene period can be interpreted as a relatively large change in the plant biomass through time. An increase in aboveground biomass likely increased the vegetation canopy and root biomass density (Cairns et al., 1997), causing more water loss from deeper soil layers and aquifers by transpiration through leave stomata, rather than evaporative loss directly from shallow soil (Lawrence and Slingo, 2004; Markewitz et al., 2010). The abundant precipitation and denser canopy cover in the early to middle Holocene resulted in a period of higher soil moisture availability, despite increasing temperatures and plant biomass likely causing an increase in the total evapotranspiration at this time.

The results of this study show that over the recent 3,000 years the meteorological, hydrological and soil systems were all in a dry phase (Fig. 5). At the same time, human activities had intensified and changed the underlying surface conditions in the Yunnan region (Xiao et al., 2017; Xiao et al., 2018). *Pinus* sp. are typically pioneer species, which can colonize disturbed sites if competition and grazing pressures are low. Increases in Poaceae pollen abundance is commonly interpreted as reflecting increased regional aridity, but it is also influenced by early farming activities. If climate change was the only critical cause for vegetation change during the last 3,000 years, the vegetation composition should have changed to a status similar to that recorded in the previous “cool-dry” period. However, the results of the cluster analysis show that samples from the last 3,000 years appear across all three clusters identifying the late glacial period, early Holocene and mid- to late Holocene (Fig. S4 in the Supplement). Therefore, the pollen-based SWSI may have failed to reflect the real soil moisture condition for the recent 3,000 years. At present, we cannot estimate the impacts and effects of human activities on lake hydrologic regimes and watershed landscapes in this period. Thus, producing an accurate reconstruction of the monsoonal precipitation, hydrological balance and soil moisture for the past 3,000 years is limited.

5. Conclusions

Sedimentological and pollen data from Yilong Lake provide a 27,000-yr perspective of local and regional variations in the ISM precipitation, hydrological balance and soil moisture condition in SW China. The results show that the reconstructed precipitation is generally consistent with the regional pattern, with low precipitation during the LGM and the YD event and high precipitation during the B/A event and the early to middle Holocene. But since the LGM, the long-term changes in precipitation, hydrological balance and soil moisture are not completely consistent. On a millennial scale, the hydrological balance was more sensitive to temperature change which directly controls the lake surface evaporation rate. In addition to precipitation and temperature, plant processes may also play a large role in regulating soil moisture. Plant biomass in the Yilong area increased during the early to middle Holocene as recorded in the pollen records. This likely increased the vegetation canopy, causing less water loss from shallow soil via evaporation. Human activities intensified during the last 3,000 years. It is difficult to estimate the impact of human activities on the regional landscape and on the watershed hydrology in SW China in the late Holocene. Therefore, the reconstructed deficit in the hydrological and soil systems during this period cannot be interpreted simply as climate change.

Finally, our study highlights that "wetness" and "dryness" should be precisely defined when interpreting different paleoproxies.

Data availability

The reconstructed data presented in the paper can be accessed by contacting Mengna Liao or Jian Ni.

Author contributions

ML, KL and JN developed the research questions. ML, KL and WS designed and conducted the experiment. KL conducted the fieldwork. ML, KL, WS analysed samples and processed data. ML and KL wrote the manuscript with contributions from all co-authors.

Competing interests

The authors declare that they have no conflict of interest.

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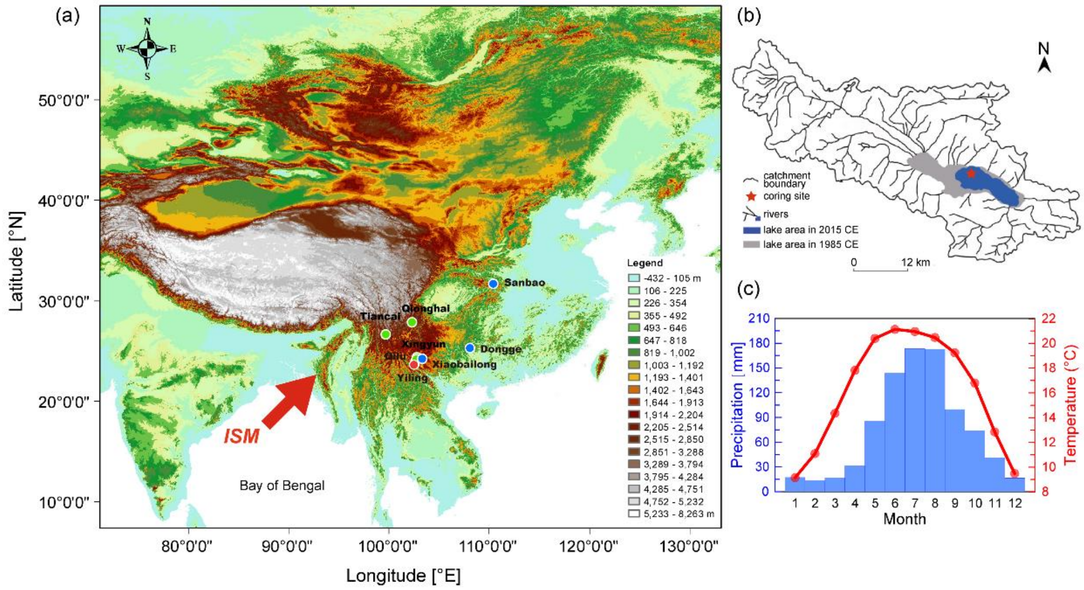

Figure 1 (a) Map showing the location of Yilong Lake and other paleoclimatic records; (b) catchment of the lake and the location of the core YLH; (c) average monthly temperature and precipitation from 1951 to 2017 CE (China Meteorological Data Service Centre, https://data.cma.cn/en).

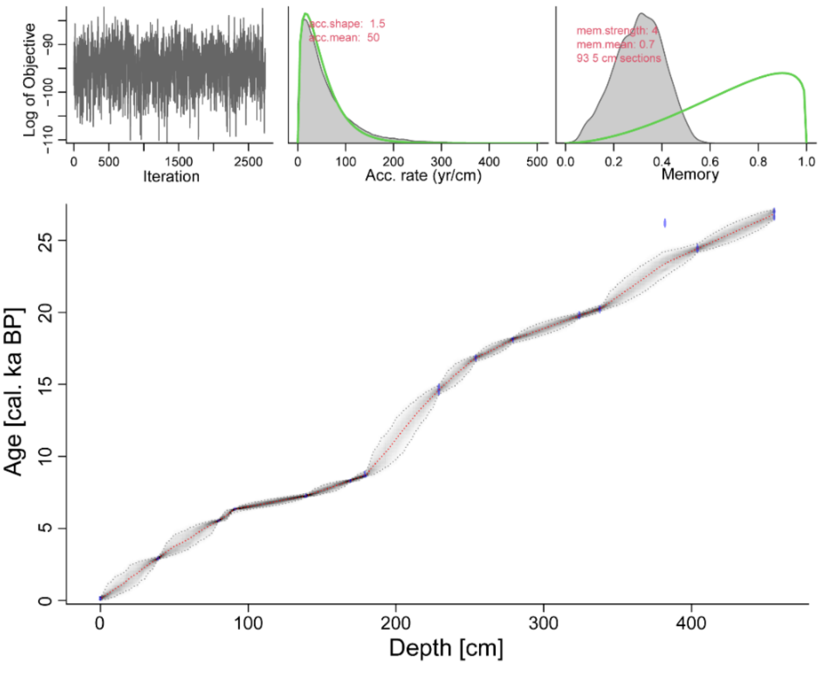

Figure 2 Age-depth model of the sediment core YLH.

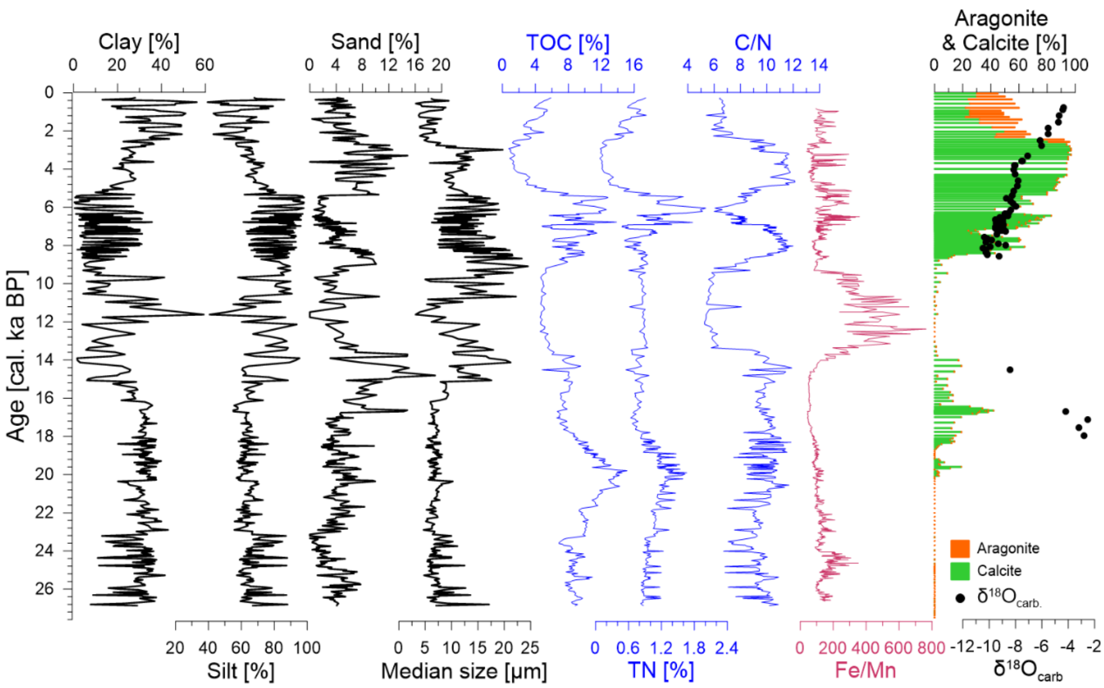

Figure 3 Down-core variations in the grain-size components (clay, silt and sand), median grain size, TOC and TN content, C/N and Fe/Mn ratios, calcite and aragonite content, and $\delta^{18}O_{carb}$.

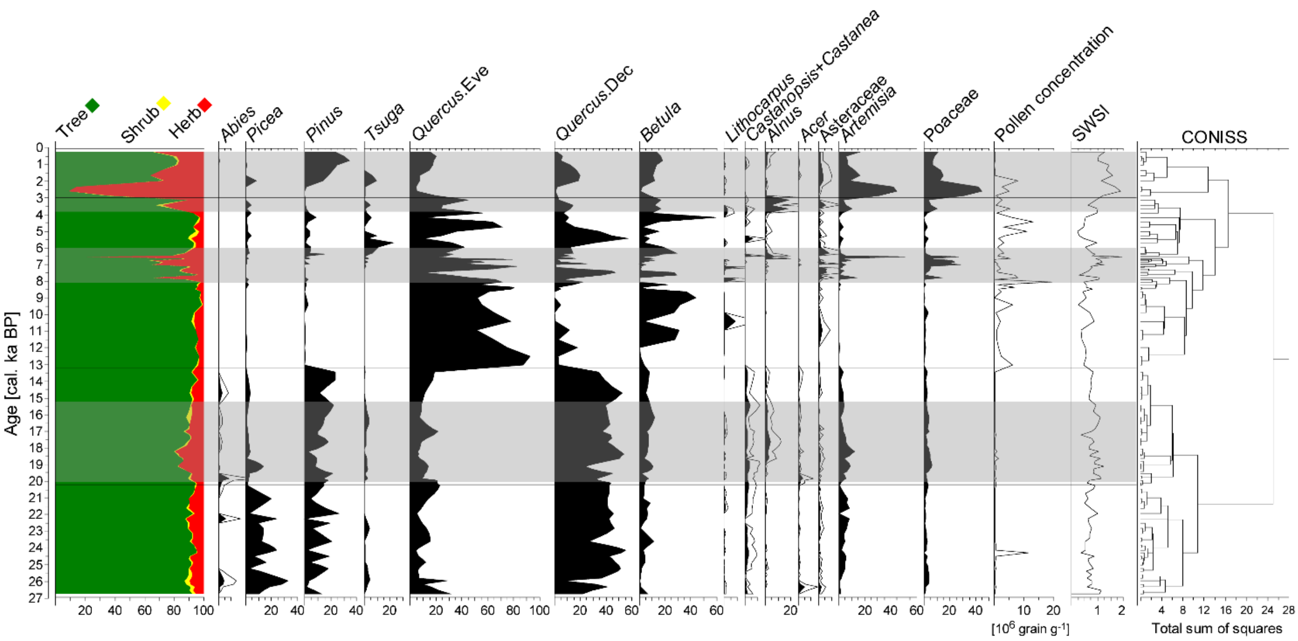

Figure 4 Pollen percentages diagram of the main taxa (average percentages ⩾ 5 %), the pollen concentration and the reconstructed soil water stress index (SWSI). Pollen types with relatively low percentages have been magnified 3 times. Gray shadows mark periods of relatively high SWSI.

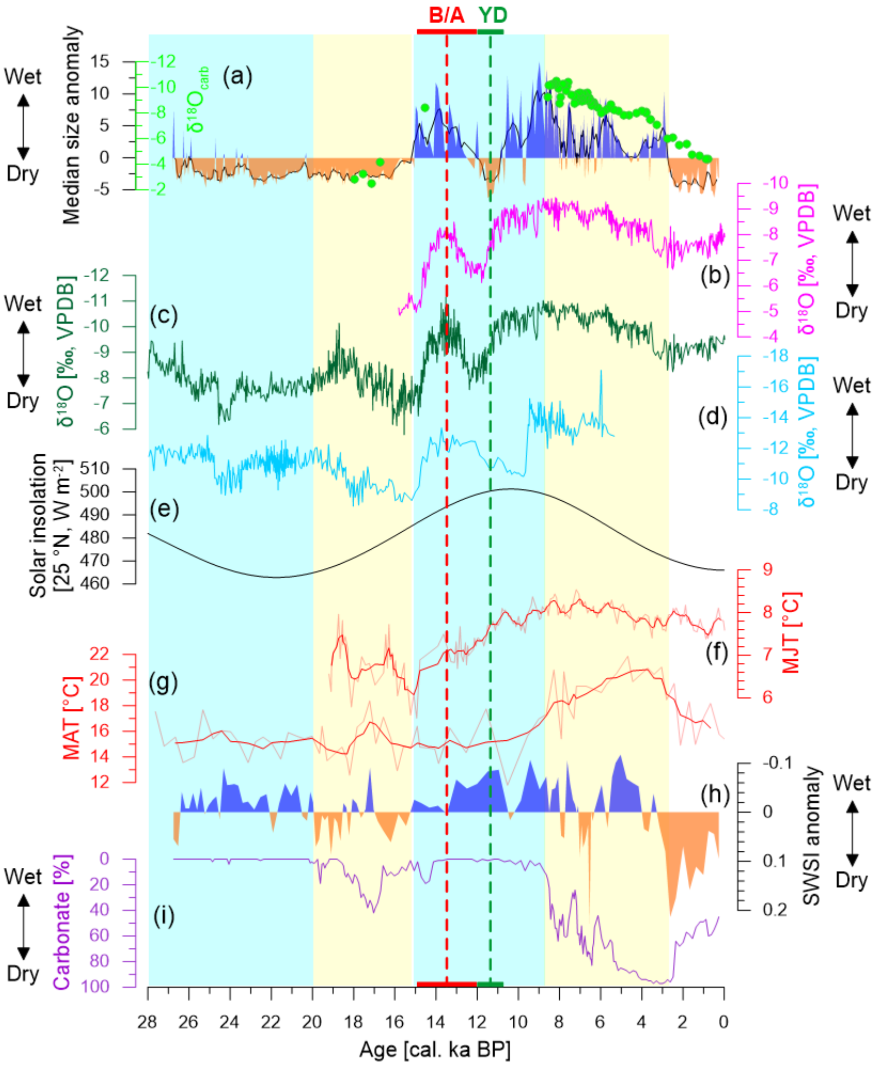

Figure 5 Median size and $\delta^{18}O_{carb}$ values from Yilong Lake (a); Stalagmite $\delta^{18}O$ records from Sanbao Cave (Cheng et al., 2016) (b), Dongge Cave (Dykoski et al., 2005) (c) and Xiaobailong Cave (Cai et al., 2015) (d); summer insolation curve for 25° N (e); chironomid-mean July temperature (MJT) from Tiancai Lake (Zhang et al., 2019a) (f); brGDGT-derived mean annual temperature (MAT) from Qionghai Lake (Wang et al., 2020) (g); pollen-based reconstructed soil water stress index (SWSI) for Yilong region (h); carbonate content of sediments from Yilong Lake (i). Light yellow (blue) shadows denote periods of inconsistency (consistency) of dry/wet condition among precipitation, SWSI and hydrological balance.

Table 1 AMS ^{14}C results from the YLH sediment core.

Lab ID	Depth [cm]	Materials	IRMS $\delta^{13}C$ [‰]	Conventional Radiocarbon Age [yr BP]	Calibrated Age [2σ, cal. yr BP]
Beta-468347	1	Sediment-TOC	24.6	-30 +/- 30	100.37 +/- 0.37 pMC
Beta-492284	40	Charcoal and plant remains	-21.5	2820 +/- 30	2848–3004
Beta-492285	80	Charcoal and plant remains	-20	4820 +/- 30	5477–5539
Beta-468348	91	Sediment-TOC	-27.3	5560 +/- 30	6297–6398
Beta-492286	140	Charcoal and plant remains	-18.7	6330 +/- 30	7237–7318
Beta-468349	170	Charcoal and plant remains	-21.5	7510 +/- 30	8289–8386
Beta-492287	180	Charcoal and plant remains	-28.7	7860 +/- 30	8585–8771
Beta-492288	230	Charcoal and plant remains	-13.1	12460 +/- 40	14322–14745
Beta-468350	255	Sediment-TOC	-17.6	13880 +/- 40	16678–17025
Beta-492289	280	Sediment-TOC	-18.8	14800 +/- 40	17986–18242
Beta-468351	325	Sediment-TOC	-19.9	16400 +/- 50	19582–19923
Beta-541605	338	Plant remains	-27	16730 +/- 50	20058–20408
Beta-468352	383	Sediment-TOC	-19.3	21980 +/- 70	25973–26393
Beta-537522	404	Charcoal	-25.1	20320 +/- 70	24183–24631
Beta-468353	457	Plant remains	-17.8	22560 +/- 80	26829–27171