A 400-year tree-ring δ18O chronology for the southeastern Tibetan Plateau: Implications for inferring variations of the regional hydroclimate

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Abstract

We developed a new tree-ring cellulose δ18O chronology for the southeastern Tibetan Plateau from Balfour spruce (Picea likiangensis var. balfouriana [Rehd. et Wils.]) that covered the period from 1600 to 2008, and compared the results with a previous study to explore climatic variations in the Nyingchi-Bomi area. Our tree-ring δ18O chronology correlated significantly with the previous study (Shi et al., 2012; Climate of the Past 8, 205–213) during the common period from 1781 to 2005, and provided new insights into long-term regional hydroclimatic variations. Besides the significant positive correlations between tree-ring δ18O and the temperature and sunshine duration during the growing season, tree-ring δ18O was strongly negatively correlated with regional cloud cover, relative humidity, and precipitation in July and August. The correlations with cloud cover data were stronger than in previous research, but the correlations with precipitation and relative humidity in July and August were weaker. When Indian summer monsoon conditions prevail, regional hydroclimatic variations (and especially cloud cover) have the dominant influence on tree-ring δ18O in the study area. Based on the regional data, δ18O in tree rings can be an effective proxy to infer the temporal variations in regional hydroclimatic conditions and the strength of the Indian Summer Monsoon. Our results reveal that the Indian Summer Monsoon weakened from 1600 to 1650, followed by continuous strengthening until 1740 and a slight weakening from 1740 to present. The temporal variations in the cellulose δ18O chronology generally corresponded well to the δ18O and glacier snow accumulation records found in ice cores from the middle Himalaya.

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1. Introduction

The Tibetan Plateau plays a key role in the atmospheric circulation patterns and climate variability of eastern and southern Asia due to its elevation, size, and geographical position (Wu and Zhang, 1998; Ye and Wu, 1998). The climate of the southeastern Tibetan Plateau is influenced by the Indian summer monsoon during the summer and by the prevailing westerly airflow during the winter, and is therefore seasonally affected by air masses from very different regions (Zhou and Yu, 2005; Liu et al., 2008; Zhao et al., 2012). Meteorological data from 1960 to 2008 shows no significant trend in annual precipitation in the southeastern Tibetan Plateau, but summer precipitation has decreased at a rate of about 1.5 mm/decade (Li et al., 2011). Unfortunately we cannot fully understand the long-term variations in climate based on the available meteorological datasets, which are sparse and limited to less than a century. Therefore, high-resolution climatic proxies are needed to better understand the long-term climate trends, including variations in the strength of the Indian monsoon, in the southeastern Tibetan Plateau.

Precipitation δ18O is often strongly controlled by geographic parameters, such as altitude and distance from the ocean, and by meteorological parameters, such as air temperature, relative humidity, and the amount of precipitation (Dansgaard, 1964; Araguás-Araguás et al., 2000; Lee and Fung, 2008; Zhao et al., 2012). In addition, they are controlled by the sources of the moisture and by transport processes (Araguás-Araguás et al., 1998; Tian et al., 2007; Liu et al., 2008; Breitenbach et al., 2010), and differences in any of these parameters can result in pronounced differences in the annual and seasonal δ18O of the regional precipitation (Gao et al., 2011). A complete 1-year precipitation δ18O record from the Bomi area, in the southeastern part of
the plateau, was significantly positively correlated with temperatures from November to May, and was significantly negatively correlated with precipitation from June to October (Gao et al., 2011). This result has crucial implications for understanding the climatic information recorded in tree-ring δ18O in this region.

A number of proxies for precipitation δ18O obtained from mid to high-latitude continental environmental archives, such as ice cores (Duan et al., 2006; Yao et al., 2008) and tree-ring cellulose (Treydte et al., 2006; Liu et al., 2009, 2012), have been developed and used successfully to reconstruct regional paleohydrology and paleoclimate. Cellulose δ18O values from tree rings offer the potential of long, high-resolution climate reconstruction in temperate regions (McCarron and Loader, 2004; Treydte et al., 2007) and tropical regions (Poussart et al., 2004). Based on modeling of tree-ring δ18O (Rodent et al., 2000), the δ18O of plant tissues reflects the variations in three factors: δ18O values of the source water (Danis et al., 2006), evaporative enrichment of 18O in leaf water due to transpiration (Barnard et al., 2007), and biochemical fractionation during the synthesis of organic matter (Gessler et al., 2006). Thus, the cellulose records a signal that integrates both the conditions at the leaf interface (≈60% of the isotopic signal in tree rings) and the isotopic ratio in the source water (≈40% of the isotopic signal) (Rodent et al., 2000). The δ18O signal in tree rings therefore varies as a function of temperature, relative humidity (which affects the vapor-pressure deficit), and the amount of precipitation, and depends on the water sources and climate conditions in the region (Miller et al., 2006; Treydte et al., 2006; Liu et al., 2009; Bale et al., 2010). These results suggest that changes in tree-ring δ18O can be used as an effective proxy for local or regional climate change.

Several investigations have focused on the climatic signals inferred from tree-ring δ18O on the northeastern Tibetan Plateau. Griesinger et al. (2011) believed that the cellulose δ18O mainly reflected the amount of summer precipitation. Similar results have recently been reported for regions near this part of the plateau (Sano et al., 2012; Xu et al., 2012), where tree-ring δ18O was used to assess long-term trends for the Indian Summer Monsoon. In addition, other studies found that tree-ring δ18O was linked with seasonal variations in regional cloud cover in southwestern China, where there is large amount of cloud during the summer monsoon season (Shi et al., 2011, 2012; Liu et al., 2012). Cloud cover is an important factor in climate feedback because it affects the surface radiation budget (Stephens, 2005), and changes in cloud cover can therefore create large variations in the predictions of climate models (Nardino and Georgiadis, 2003). A previous study in the Bomi area (Shi et al., 2011, 2012) found that tree ring δ18O was significantly correlated with cloud cover inferred from records at a local meteorological station, but the strength of the correlation was relatively weak and the overall chronology only covered ca. 200 years. Therefore, a longer tree-ring δ18O chronology would be helpful to evaluate low-frequency climatic cycles and centennial-scale climatic variations.

Based on previous research in the Nyinjchi-Bomi area, we designed the present study (1) to develop a new, long-term, and well-replicated tree-ring δ18O chronology for the Nyinjchi-Bomi area using Balfour spruce (Picea likiangensis var. balfouriana) at the timberline, (2) to identify the dominant climatic factors that affect tree-ring δ18O in this region, and (3) to characterize the climate-change signals recorded in tree-ring δ18O values. Our new results provide evidence of the hydroclimatic information recorded in the regional tree-ring δ18O chronologies and increase the length of the chronology available for reconstructing paleohydroclimatic variations in the southeastern Tibetan Plateau.

2. Materials and methods

2.1. Study area

Our study area is located in the southeastern part of the Tibetan Plateau (Fig. 1a). The Indian summer monsoon enters the study area from the southwest along the Yarlung Tsangpo River valley, delivering abundant moisture obtained from the Indian Ocean and Arabian Sea during the summer (Breitenbach et al., 2010). According to meteorological records from the Bomi station (29°52'N, 94°46'E, 2736 m a.s.l.), the mean annual precipitation in the study area from 1961 to 2008 was 837 mm (Fig. 2). Precipitation during the monsoon season (June to September) totaled 427 mm, which amounts to 51.0% of the annual total. July (mean temperature of 16.6 °C) and January (0.2 °C) are the warmest and the coldest months, respectively. Records from the meteorological station at Nyinjchi (29°34'N, 94°28'E, 3000 m a.s.l.), located in a river valley on the western side of the Sygera Mountains, show a mean annual precipitation from 1960 to 2008 of 676 mm, of which 71.9% occurs from June to September (Fig. 2). July (mean temperature of 15.8 °C) and January (0.5 °C) are the warmest and the coldest months, respectively. Additionally, in situ measurements at our study site show that the mean relative air humidity in the Nyinjchi-Bomi region is above 80% at the timberline (Liu et al., 2011; Fig. 1b).

Balfour spruce is the dominant tree species at elevations of 3200 to 4300 m a.s.l. in the study area. We selected an open-canopy stand around Ranwu Lake (Fig. 1a; 29°26'N, 96°29'E, 4150 m a.s.l.) and sampled spruce trees at that site in May 2009. The sampling site is located about 130 km and 400 km east of the Bomi and Nyinjchi weather stations, respectively, and 120 km from a previous research site (Shi et al., 2012; Fig. 1a). The local growing season extends from June to August. The cover of Balfour spruce at this site ranges from 20 to 40%, and the cover of understory rhododendron shrubs is always greater than 40%. Soils in the study area are primarily subalpine meadow soils on a moderate slope, with a thickness of 20 to 50 cm, but they are thin and can even be absent on steeper and eroded slopes.

2.2. Tree-ring dating

We obtained 36 cores at breast height (about 1.3 m) from a total of 25 dominant living trees using increment borers with a 12-mm diameter. The samples were crossdated and measured in the laboratory using standard dendrochronology procedures. We used the COFECHA software (Holmes, 1983) to check the quality of the dating and measurements. Finally, we checked our dating results against the regional tree-ring width chronology of Zhu et al. (2011), whose study site is close (~100 km) to ours. During the common period (1300 to 2002), the two chronologies revealed the same extreme growth events (data not shown).

2.3. Cellulose extraction and oxygen isotope measurements

Shi et al. (2011) concluded that pooling at least four trees with one core per tree was sufficient to meet the criteria (an expressed population signal >0.85) for a robust tree-ring oxygen chronology in the Bomi area. Therefore, we selected eight cores from eight trees with homogeneous growth patterns and pooled wood from their annual rings prior to cellulose extraction (Leavitt, 2008; Liu et al., 2009). We first milled the pooled annual wood samples (to <80 μm), and then extracted α-cellulose using a method based on those of Green (1963) and Loader et al. (1997). To obtain better homogenization of the cellulose, we used a JY92-2D ultrasound machine (Ningbo Scientz Biotechnology Co. Ltd., Ningbo, China) to break the cellulose fibers, according to the method of Laumer et al. (2009).

The δ18O values were determined using a High Temperature Conversion Elemental Analyzer (TC/EA) coupled to a Finnigan MAT-253 mass spectrometer (Thermo Electron Corporation, Bremen, Germany) at the State Key Laboratory of Cryospheric Sciences, Chinese Academy of Sciences. To improve precision, the δ18O analysis of each sample was repeated four times, and after excluding outliers (values more than three standard deviations from the mean), we calculated the mean values. We measured the ratio for a benzoic acid working standard (repeated four times) with a known δ18O value (IAEA-601,
\( \delta^{18}O = 23.3\% \) every seven measurements to monitor the analytical precision and to calibrate the samples for analytical accuracy (Liu et al., 2009, 2012). The IAEA-C\textsubscript{3} cellulose standard (\( \delta^{18}O = 32.2\% \)) was also used to calibrate the tree-ring oxygen measurements for each sequence of measurements. The resulting standard deviation of the replicates was less than 0.3\% based on three or four measurements.

### 2.4. Tree-ring \( \delta^{18}O \) and climate analysis

To determine the concurrent and time-lagged relationships between tree-ring \( \delta^{18}O \) and selected climate variables, we calculated Pearson’s correlation coefficients (Blasing et al., 1984) over the period from 1960 to 2008 between tree-ring \( \delta^{18}O \) and each of the monthly-resolution climate variables.
climate variables: mean temperature, total precipitation, mean relative humidity, low cloud cover, total cloud cover, sunshine duration, and diurnal temperature range (DTR). All data were obtained from the Bomi and Nyingchi meteorological stations, which were the stations closest to our site (Fig. 1a). Significance levels were tested using a bootstrap procedure (Guiot, 1991) and statistical significance was defined at $p < 0.05$. We compared the values of the climatic parameters at the Nyingchi and Bomi meteorological stations for the whole year and for the spring (March and April) and summer (July and August) periods, and found strong similarity (and significant correlations) between the climatic records at the two stations (Table S1). In addition, the temporal variations in the temperature, precipitation, relative humidity, sunshine duration, cloud cover, and DTR in the spring (March and April) and summer (July and August) periods were significantly correlated between the two weather stations (Table S2). On this basis, we combined the two datasets to represent the regional mean climatic values that we compared with the $\delta^{18}O$ series. Furthermore, we computed the spatial correlations between tree-ring $\delta^{18}O$ and a gridded CRU TS 3.0 data set (Mitchell and Jones, 2005) to test the spatial coherence between the tree-ring $\delta^{18}O$ values and local to regional climatic variations. Only the data from 1960 to 2008 were used in this part of our study.

3. Results and discussion

3.1. The tree-ring $\delta^{18}O$ chronology

The tree-ring $\delta^{18}O$ values ranged from 18.2 to 27.8‰ (Fig. 3a) over the period from 1600 to 2008, with an overall mean of ca. 23.1‰, which is similar to the mean tree-ring $\delta^{18}O$ values (24.8‰) and range (7.8‰) for the period from 1781 to 2005 that were reported by Shi et al. (2012) and Giese et al. (2011) for samples obtained near our study site. Tree-ring $\delta^{18}O$ was significantly ($p < 0.05$) lower from 1660 to 1850 than from 1851 to 2008, with averages of ca. 22.6‰ (below the mean) and 23.7‰ (above the mean), respectively (Fig. 3a). The periods with the highest tree-ring $\delta^{18}O$ values were at ca. 1630 to 1650 and 1880 to 1990, whereas tree-ring $\delta^{18}O$ was less than the long-term average from 1700 to 1750. Compared to ring width, which had a first-order autocorrelation of 0.52, the first-order autocorrelation for the tree-ring $\delta^{18}O$ series was 0.36 (Fig. 3a), indicating that some influence of the previous year on the current year’s wood cellulose could be detected. Our tree-ring $\delta^{18}O$ chronology compared well with yearly values from a previous study in the Bomi area ($r = 0.55$, $p < 0.001$; Shi et al., 2012) during the common period from 1781 to 2005 (Fig. 3a). When we correlated the two chronologies at different time scales (3
or 5 years), the coefficients were similar ($r = 0.55$ at a 3-year scale; $r = 0.56$ at a 5-year scale); however, because the number of degrees of freedom decreased greatly at these longer scales, we have retained the annual time scale in the rest of our analysis. This observed coherence of the two $\delta^{18}O$ chronologies demonstrates that the $\delta^{18}O$ values in tree rings may contain similar climatic signals that reflect regional climatic variations. However, from 1850 to 1870, the correlations between the two time series were lower and not statistically significant (Fig. 3b). The cause of this difference from the other periods is not known.

3.2. Climatic response of tree-ring $\delta^{18}O$

To identify the influence of climatic factors on tree-ring $\delta^{18}O$ fractionation, we correlated the cellulose $\delta^{18}O$ values with instrumental climatic parameters from the Bomi and Nyingchi meteorological stations for the period from 1960 to 2008 (Fig. 4). The correlations between the $\delta^{18}O$ series and the climate variables for the two stations generally showed similar patterns, but different signal strengths. Tree-ring $\delta^{18}O$ was negatively correlated with low cloud cover, total cloud cover, precipitation, and relative humidity from July to August at one or both stations, and significantly positively correlated with temperature, sunshine duration, and DTR at one or both stations. These patterns are similar to the moisture-related responses reported by Shi et al. (2011). The March–April period was also significant. Tree-ring $\delta^{18}O$ values were significantly positively correlated with cloudiness and relative humidity at both stations, but significantly negatively correlated with sunshine duration and DTR at both stations (Fig. 4). Overall, the correlations between the tree ring $\delta^{18}O$ values and the climatic parameters in the March–April and July–August periods showed inverse patterns.

Considering the coherence of the climatic variations at the Nyingchi and Bomi meteorological stations, especially during the summer (Table...
S1), we computed the climatic responses of tree-ring δ¹⁸O to regional climatic parameters (i.e., based on the average values for these two stations) to reveal the regional climatic signals contained in the tree-ring δ¹⁸O. The tree-ring δ¹⁸O values appeared to be most strongly (significantly negatively) correlated with the July–August degree of cloudiness in the Nyingchi-Bomi area (Table 1). From March to April, tree-ring δ¹⁸O was significantly positively correlated with RH and significantly negatively correlated with sunshine duration and DTR based on records from the two stations and regional data. However, for the July–August season, the energy-related parameters (temperature, sunshine duration and DTR) were significantly positively correlated with tree-ring δ¹⁸O, whereas the hydroclimatic parameters (precipitation, relative humidity, and cloud cover) were significantly negatively correlated with tree-ring δ¹⁸O, and the correlations were strongest for cloud cover (Table 1). The opposite directions of the responses of tree-ring δ¹⁸O to DTR and to cloudiness can be explained by the substantial inverse covariance among precipitation, cloud cover, and DTR (Dai and Trenberth, 1999). At the Bomi weather station, DTR was significantly negatively correlated with precipitation and cloud cover, which agrees with the global pattern (Dai and Trenberth, 1999). Clouds reduce DTR

### Table 1

Correlations (Pearson’s r) between the tree-ring δ¹⁸O and climatic parameters based on data from the Bomi and Nyingchi weather stations. Significance: ns, not significant. Note: Temp = mean temperature; Prec = precipitation; RH, relative humidity; Low C = low cloud cover; Total cloud = total cloud cover; Sunshine = sunshine duration; DTR = diurnal temperature range.

<table>
<thead>
<tr>
<th></th>
<th>Temp</th>
<th>Prec</th>
<th>RH</th>
<th>Low C</th>
<th>Total C</th>
<th>Sunshine</th>
<th>DTR</th>
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<tr>
<td></td>
<td>(a)</td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Bomi</td>
<td>– 0.048</td>
<td>0.292*</td>
<td>0.325*</td>
<td>0.115</td>
<td>0.114</td>
<td>– 0.317*</td>
<td>– 0.403**</td>
</tr>
<tr>
<td>Nyingchi</td>
<td>0.006</td>
<td>0.074</td>
<td>0.360**</td>
<td>0.459***</td>
<td>0.483**</td>
<td>– 0.369**</td>
<td>– 0.401**</td>
</tr>
<tr>
<td>Mean</td>
<td>– 0.022</td>
<td>0.264*</td>
<td>0.376**</td>
<td>0.281</td>
<td>0.281</td>
<td>– 0.375**</td>
<td>– 0.426**</td>
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<tr>
<td></td>
<td>(b)</td>
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<tr>
<td>Bomi</td>
<td>0.395**</td>
<td>– 0.272</td>
<td>– 0.274</td>
<td>– 0.515**</td>
<td>– 0.473**</td>
<td>0.406**</td>
<td>0.406**</td>
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<tr>
<td>Nyingchi</td>
<td>0.444**</td>
<td>– 0.274**</td>
<td>– 0.388**</td>
<td>– 0.325</td>
<td>– 0.406</td>
<td>0.270</td>
<td>0.240**</td>
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<tr>
<td>Mean</td>
<td>0.427**</td>
<td>– 0.296*</td>
<td>– 0.352**</td>
<td>– 0.486**</td>
<td>– 0.500**</td>
<td>0.353*</td>
<td>0.343**</td>
</tr>
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* p < 0.05.
** p < 0.01.
mainly by decreasing the daytime maximum temperature, which leads to little enrichment of the δ18O in leaf water, thereby decreasing cellulose δ18O.

Cloud cover is a moisture-related variable, and was significantly positively correlated with relative humidity, precipitation, and sunshine duration during July and August (Table S2, Fig. S1). Increased cloudiness that is linked to changes in regional convective activity will increase the probability of rainfall with low δ18O values (Gao et al., 2011) and will decrease the atmospheric vapor-pressure deficit (Reinhardt and Smith, 2008), resulting in low cellulose δ18O. Thus, the strong correlation between cellulose δ18O and cloud cover could be explained by the influence of cloud cover on precipitation, relative humidity, and solar irradiance, all of which directly affect cellulose δ18O discrimination (Shi et al., 2011). Thus, we consider cloud cover to be a potentially effective and comprehensive surrogate for the ensemble of parameters that influence tree-ring δ18O fractionation in our study region. The slope of the relationship between tree-ring δ18O and the average summer RH in the Nyingchi-Bomi area was −0.112‰/% (R² = 0.13, p < 0.05, n = 48, July–August), which is comparable to a previous observation (−0.12‰/%, Shi et al., 2011). In addition, the slope of the relationship between the tree-ring δ18O and the average summer cloud cover in the Nyingchi-Bomi region was −0.085‰/% (R² = 0.25, p < 0.05, n = 48, July–August).

In contrast to the relationships during the wet July–August season, we found that tree-ring δ18O was significantly positively correlated with March–April cloudiness and relative humidity, but significantly negatively correlated with sunshine duration and DTR. The influence of the March–April conditions likely resulted from our inclusion of both earlywood and latewood in our cellulose samples. The earlywood forms during the late spring and early summer, when much of the water that is taken up by trees is provided by melting snow that was deposited during the previous winter. The climatic response pattern (Fig. 4) may explain why our chronology showed weaker correlations between tree-ring δ18O and cloud cover records than in the results of Shi et al. (2012).

To determine whether the March–April and July–August relationships between tree-ring δ18O and regional moisture availability existed over a large scale, we calculated the spatial correlations between the tree-ring δ18O chronology and the CRU TS 3.0 datasets for significant climatic parameters from 1960 to 2008 (Fig. 5). Though the CRU cloud cover dataset is not derived from direct cloud observations, we nonetheless found a close relationship between the observed cloud cover at the meteorological stations and the CRU data in July and August, but not in March and April (Fig. S2). The tree-ring δ18O was significantly negatively correlated with cloud cover during the Indian summer monsoon season (July and August) in the region surrounding the sampling site, particularly to the west and south of our study site, which lie in the direction of the source of the monsoon precipitation in the Bay of Bengal (Fig. 5a). There was no significant correlation between tree-ring δ18O and cloud cover from the CRU data in March and April (not shown) due to large differences between the observed data and the CRU data (Fig. S2a). Because of the close linkage between cloud cover and DTR, we also found significant relationships in both the March–April period (a negative correlation; Fig. 5b) and the July–August period (a positive correlation; Fig. 5c), which indicates strong coherence between the local observations and the CRU dataset in terms of the DTR variations. The spatial map of the correlation between tree-ring δ18O and July–August precipitation (Fig. 5d), which was dominated by negative correlations, indicates the presence of a regional moisture signal. The positive response of tree-ring δ18O to regional July–August temperature (Fig. 5e) reflects the stronger enrichment of leaf water δ18O caused by increasing temperatures in the summer (Fig. 2a).

These results demonstrate that our tree-ring δ18O is a good proxy for variations in the regional hydroclimate, especially in terms of regional cloudiness in July and August. However, the strengths of the correlations between tree-ring δ18O and climatic factors are not yet sufficiently strong to allow climate reconstructions based on the observed data. We believe that separate measurements of tree-ring δ18O in the latewood may be an effective way to obtain stronger and more reliable climatic information from only the summer climatic data (An et al., 2012).

3.3. Tree-ring δ18O comparisons and common climatic trend

An increasing number of tree-ring δ18O studies have recently been conducted in the southeastern Tibetan Plateau (Fig. 1a). Shi et al. (2012) reconstructed the regional cloud cover from June to August based on correlations between tree-ring δ18O and the CRU dataset. Based on the correlations among tree-ring δ18O, the observed regional cloud cover, and the CRU cloud cover in July–August (Fig. S3), we reconstructed the regional cloudiness covering the period from 1600 to 2008 (Fig. S4). It should be noted that some bias in this reconstruction was caused by using the CRU data for calibration. Apparently, our chronology showed higher mean cloudiness than in the results of Shi et al. (2012). However, during the common period, the two series revealed similar fluctuations based on the 21-year moving average (Fig. S4). Shi et al.’s (2012) results suggest that the period from 1807 to 1817 was characterized by the lowest cloud values, perhaps attributable to two consecutive volcanic eruptions. In addition, temperatures during this period were lower in the study region (Zhu et al., 2011). However, based on our reconstruction (Fig. S4), the cloud cover during this period was not the lowest value since 1600. Thus, our chronology can provide additional paleoclimatic information, over a longer time scale, than in the study by Shi et al. (2012).

Considering the possible biases between the interpolated CRU and the observed data and the absence of a high-frequency signal in the CRU data (Fig. S2) in certain seasons caused by limited observed meteorological data from the Tibetan Plateau, we have only compared the common hydroclimatic trends revealed by the tree-ring δ18O chronologies with other series. We found common hydroclimatic signals in the tree-ring δ18O values (Fig. 6). However, the signal’s strength was different, largely due to the site-dependence. In the Nepal, Reting, and Hongyuan areas, tree-ring δ18O was significantly negatively correlated with monsoon rainfall, and was sensitive to different moisture-related climatic parameters (PDSI, Sano et al., 2012; precipitation, Griesinger et al., 2011; relative humidity, Xu et al., 2012), with variable signal strength. These three sites all demonstrated a common “tree-ring δ18O–monsoon intensity” relationship. In the Bomi area, Shi et al. (2011) found that tree-ring δ18O was significantly correlated with the observed total precipitation and mean relative humidity and with the CRU estimate of cloudiness from June to August, but not with the observed cloud index. In contrast, our tree-ring δ18O series showed significant correlations with both the observed data and the CRU data in July and August. This may be related to site-specific differences, since our sampling site is often covered with thick clouds (Fig. 1b). Thus, the tree-ring δ18O values at our study site were more sensitive to local cloudiness, which is closely correlated with variations in the hydroclimatic conditions. Considering all the above mentioned evidence, tree-ring δ18O values in the study area are closely related to variations in hydroclimatic conditions (Griesinger et al., 2011; Sano et al., 2012; Shi et al., 2012), and are thus strongly related to the intensity of the Indian Summer Monsoon (Xu et al., 2012).

The annual variations in the various tree-ring δ18O chronologies were also coherent in the southeastern Tibetan Plateau (Fig. 6). Our tree-ring δ18O series was significantly positively correlated with the other series; the correlation coefficient reached 0.55 (n = 225, p < 0.001) for the Bomi tree-ring δ18O series (Shi et al., 2012), 0.39 (n = 406, p < 0.001) for the Reting series (Griesinger et al., 2011), and 0.17 (n = 223, p < 0.01) for the Nepal series (Sano et al., 2012), suggesting a common climatic forcing that was independent of potential differences among species, elevations, and local ecological
conditions. All of the series exhibit generally increasing trends in δ¹⁸O values since 1800, though at different rates. However, site-dependent discrepancies were also revealed that appear to be related to the location and the associated distance from the Indian Ocean and Bay of Bengal. For instance, the apparent dry conditions since 1950 recorded in the Reting and Nepal chronologies were not revealed in the other two records (Fig. 6). The wetter climate from 1700 to 1750 in our chronology was not obvious in the precipitation reconstruction from...
Our tree-ring \( \delta^{18}O \) series extends back to 1600 and captures climate conditions during the late period of the Little Ice Age climate pattern.

3.4. Regional hydroclimatic variation compared with ice core proxies

A more complete interpretation of the past evolution of the Indian Summer Monsoon since 1600 can be obtained by comparing our tree-ring \( \delta^{18}O \) chronology with the high-resolution ice-core \( \delta^{18}O \) values from Dasuopu in Tibet (Thompson et al., 2000) and Indian summer monsoon precipitation, as indicated by snow accumulation in the Himalaya (Duan et al., 2006). Tree-ring width in our study region reflects the temperature variability in a broad region (Zhu et al., 2011), and indicates a warming trend from 1700 to the present (Fig. 7a). The increase in sea surface temperatures over the equatorial Indian Ocean (Wilson et al., 2006) and low-amplitude warming on the Tibet Plateau (Zhu et al., 2011) would lead to a decrease in the thermal contrast between land and sea, therefore decreasing water vapor transport from the Indian Ocean to the Tibet Plateau, resulting in decreasing intensity of the Indian Summer Monsoon (Duan et al., 2006). These results support our interpretation of the long-term increasing trend in tree-ring \( \delta^{18}O \) values since 1600 (Fig. 7b).

Comparison between our tree-ring \( \delta^{18}O \) series and the ice-core \( \delta^{18}O \) series from Dasuopu (Fig. 7c) reveals common long-term decreasing trends since 1600. The ice-core \( \delta^{18}O \) series from Dasuopu responded to fluctuations in the intensity of the Indian summer monsoon, which is reflected in the decreasing trend that began in the late 18th century and two major monsoon failures, from 1790 to 1796 and from 1876 to 1877 (Thompson et al., 2000). The tree-ring and ice-core \( \delta^{18}O \) chronologies suggest similar fluctuations before 1900, but then an offset occurs during the 1900s. Both series have a slight increasing trend since 1870 and since 1940 for the tree-ring \( \delta^{18}O \) and ice-core \( \delta^{18}O \), respectively. We hypothesize that the difference in the starting date for the increasing \( \delta^{18}O \) in different proxies was caused by increased leaf-water \( \delta^{18}O \) enrichment that resulted

![Fig. 6. Comparison of the regional tree-ring \( \delta^{18}O \) chronologies in the southeastern Tibetan Plateau: (a) the present study; (b) a study in the Bomi area (Shi et al., 2012); (c) a study in Nepal (Sano et al., 2012); and (d) the precipitation anomaly based on tree-ring \( \delta^{18}O \) from Reting (Grieginger et al., 2011). All data were smoothed using 21-year fast Fourier transform filters to emphasize decadal-scale oscillations. Horizontal lines represent the long-term mean for each parameter. Because of the negative correlation between tree-ring \( \delta^{18}O \) and the hydroclimatic parameters, the y-axis for \( \delta^{18}O \) has been reversed to facilitate comparison with the monsoon precipitation reconstruction.](image)
from the continuous warming (Fig. 7a). In contrast to the higher snow accumulation from 1820 to 1920 shown in the Dasuopu ice core (Fig. 7d), no significant peaks occur in the tree-ring $\delta^{18}O$. The difference between these records may be related to the different locations and to different environmental responses peculiar to the proxy material being analyzed.

4. Conclusions

In this study, we developed a well-replicated 400-year tree-ring $\delta^{18}O$ series with annual resolution for the southeastern Tibetan Plateau using Balfour spruce, and used this chronology to infer past hydroclimate variability (from 1600 to 2008). Our analysis revealed that regional cloud cover, which is negatively correlated with sunshine duration and the diurnal temperature range, was the dominant influence on the tree-ring $\delta^{18}O$ values. The 400-year tree-ring $\delta^{18}O$ chronology displays an overall increase in $\delta^{18}O$ from 1700 to the present, with climate warming occurring since 1850. Comparison of our study results with previous tree-ring $\delta^{18}O$ results revealed coherence in the hydroclimatic signals recorded in $\delta^{18}O$ series on the southeastern Tibetan Plateau. The regional tree-ring $\delta^{18}O$ chronologies revealed a weakening Indian Summer Monsoon intensity since around 1900. This phenomenon is also recorded in other natural climate archives, such as ice-core $\delta^{18}O$ and glacier accumulation recorded in the middle Himalaya, on the Tibet Plateau, and in neighboring regions.

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Appendix A. Supplementary data

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References

