Ring-widths of *Abies* at tree-line ecotone reveal three centuries of early winter season temperature changes in Yunnan, China

Yingfeng Bi^{1,2,3} · Whitney Cory⁴ · Zenxin Fan⁵ · Jingchao Yang⁶ · Zhikun Wu⁷ · Wei Huang^{1,2,3} · Jianwen Li^{1,2,3,8} · Xuefei Yang^{1,2,3,8}

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Abstract

Variation in winter temperature is less well understood than in annual and summer temperatures over long timescales, particularly in low-latitude (0–30° N) montane regions with large spatial and topographic heterogeneity. Understanding these variations could be critical for forest management in these important tree growing regions. We collected tree-ring cores from Yunnan Province in montane southwest China, to determine how winter temperature has varied in the past and to explore its possible drivers in this region. Six highly correlated site chronologies were combined into a long and well-replicated regional composite (RC) chronology. The RC chronology correlated strongest with mean early winter season temperature (EWST) from November to January during 1959–2015. We applied RC chronology to reconstruct EWST from 1653 to 2015. The reconstruction shows five relatively warm historic intervals (1658–1718, 1743–1755, 1771–1791, 1929–1959, and 1995–2015) and cold intervals (1720–1742, 1792–1852, 1860–1883, 1905–1928, and 1960–1994). The 1950s with six anomalous *warm* seasons and 1980s with five anomalous *cold* seasons were the warmest and coldest decades, respectively. The EWST reconstruction generally agreed with other winter temperature reconstructions from nearby areas and with historical documents. Atlantic Multidecadal Oscillation (AMO) might be a key forcing of multidecadal winter temperature variations in montane southwest China over the past three and half centuries. Both warm and cold temperature periods coincide with respective positive and negative phases of AMO. Besides, the cold intervals during the nineteenth and twentieth centuries may also have been influenced by large volcano eruptions in low-latitude regions of south and east Asia. Our EWST reconstruction not only improves our understanding of the trends and variations of winter climate history, but also supports planning for resilience in conservation, agriculture, and forestry management in montane southwest China into the future.

Keywords Winter temperature · Climate change · Tree-ring · Montane southwest China

1 Introduction

Winter temperatures are one of the most critical factors for plant growth and forest dynamics (Körner 2003). They can determine the post winter performance of plant species (Luedeling et al. 2011), shape the geographic range of many forest species (Crumpacker et al. 2001), and influence

species composition and production in most regions (Kreyling 2010). Winter temperatures also indirectly regulate the quality and health of forests by controlling the outbreak of forest pests and pathogens (Marcais and Desprez-Loustau 2014; Marcais et al. 2017). Temperature increase due to global warming is more pronounced in winter than that in summer in high-latitude regions and in low-latitude regions with high elevation (IPCC 2014). However, compared with summer temperatures, the variations and changing trends of winter temperatures have received little attention and are poorly understood (George 2014; Wu et al. 2013).

Large sets of long-term climate data form the basis for climate change research (Wu 1990). These can be used to understand the basic characteristics of long-term changes (Yang et al. 2014), explore the physical mechanisms of climate change (Shi et al. 2017), and verify a climate model for

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✉ Whitney Cory
cory.whitney@uni-bonn.de

✉ Xuefei Yang
xuefei@mail.kib.ac.cn

Extended author information available on the last page of the article

a region (Wu 1990). Although the meteorological network has contributed to our knowledge about variations in the weather and climate, it is still lacking in its spatial and temporal coverage. For example, China just has 699 continually monitored meteorological stations. Most only started recording in 1950 (National-Meteorological-Information-Center 2019). To understand the climate variation before instrumental records, climate proxies such as ice cores, lake or loess sediments, and tree-rings can provide records spanning of thousands of years and extending to sites beyond where samples were taken (Yang et al. 2014). Annual resolved tree-rings have been widely used to reconstruct past climate variations (Briffa et al. 1990; Shao et al. 2010). However, winter temperature reconstructions are rarely available in comparison with that of summer (George 2014).

Winter temperatures vary geographically (Rozas et al. 2009). Site-specific reconstructions with geographic representation are therefore critical for global and regional climate change research, especially in regions with complicated topography (Szeicz and MacDonald 1995). Tree-ring based winter temperature reconstructions were started in the 1990s in the northern hemisphere (Shao and Fan 1999), and more than 80% of them were conducted around the mid-latitudes (30–60° N) (Appendix Table S1). In contrast, winter temperature reconstruction at low-latitudes (0–30° N) are limited, although significant winter warming has been widely observed in this region (IPCC 2014), which requires more scientific attention.

Montane southwest China, comprises Sichuan Province, southeast Tibet Plateau, and northwest Yunnan Province. It is globally significant both in topography and biodiversity. As denoted by its name, the high lands above 5000 m above sea level (m a.s.l.) account for 25% of the total area with 28 summits above 6000 m a.s.l. (Li et al. 2011). Montane southwest China is designated as a world's biodiversity hotspot (Mountains of Southwest China) (Myers et al. 2000), and ranks the second largest forested area in China (Dai et al. 2016; National-Bureau-of-Statistics-of-China 2015). However, this region is also known as one of the most environmentally sensitive regions to global climate change (Xu et al. 2009). For example, in the past four decades, the annual temperature increases in Yunnan (~ 0.30 °C/decade) (Fan et al. 2011) exceeded those globally (~ 0.14 °C/decade) (Brohan et al. 2006), and in the rest of China (~ 0.26 °C/decade) (Huang et al. 2005). Interestingly, these increases are also more prominent in winter (0.33 °C/decade) than in summer (0.26 °C/decade) (Fan et al. 2011).

Several dendrochronological studies have been carried out in this region and some results show that tree radial growth is limited by winter temperature at high elevations (Fan et al. 2009a; Panthi et al. 2018; Zhang et al. 2015). Yet not much work has been done to generate winter temperature reconstructions (Appendix Table S2). Our knowledge on

past winter variation in montane southwest China is based on a small number of studies, including research carried out at Qamdo on the Tibetan Plateau (Appendix Table S2), and at Bama Snow Mountain (Bi et al. 2017). The studies reveal some inconsistent patterns in some periods including 1800s to 1820s and 1930s to 1950s (Bi et al. 2017; Huang et al. 2012, 2018; Shao and Fan 1999; Song et al. 2007; Zhang et al. 2015). Therefore, we conducted this dendrochronological study to gain an enhanced understanding about trends in winter temperatures over the past three and half centuries in montane southwest China.

Abies georgei is a dominant conifer species distributed in the montane southwest China. This species is distributed along the elevation range from 3200 m a.s.l. to 4400 m a.s.l. The radial growth of *A. georgei* in many parts of montane southwest China is limited by winter temperature (Fan et al. 2009a; Panthi et al. 2018; Zhang et al. 2015), making it a potentially useful species for winter temperature reconstruction. In this study, we collected tree-ring cores of *A. georgei* from northwest Yunnan Province to test the suitability of this species for winter temperature reconstructions and to answer two major research questions: (1) how has the winter temperature varied in the past three and half centuries? and (2) what are the possible drivers of winter temperature variation in this region? We present reconstruction and compare it with other reconstructions to enhance the understanding of the winter climate history of this region at long time-scales. These long-term reconstructions can be used to validate regional climate models and enhance our ability to predict future climate change. All these findings will have important implications for the management of forests in montane southwest China.

2 Materials and methods

2.1 Study areas

The sampling sites (red circles in Fig. 1) are in northwest Yunnan Province, montane southwest China. Due to the rich biodiversity and complex landscape, this area is part of the UNESCO (United Nations Educational, Scientific and Cultural Organization) World Heritage *Three Parallel Rivers* covering the drainage basins of the middle upper Yangtze, upper Mekong and upper Salween rivers (Fig. 1). The elevation ranges from 1550 m a.s.l. at the lowest point (near the Yangtze River) to 7764 m a.s.l. at the summit of Meili Snow Mountain. In summer, the climate of this area is influenced by the southwest summer monsoon (Bohner 2006) and in winter by the south branch of westerly (Duan 1997). According to Shangri-la meteorological records (National-Meteorological-Information-Center 2019), the mean annual temperature from 1958 to 2019 was 6 °C, with the lowest temperature (-2.9 °C) occurring in January, and the highest

Fig. 1 Locations of tree-ring sampling sites and the meteorological stations in in montane southwest China

Fig. 2 Variations of monthly air temperature (°C) and precipitation (mm) recorded at the Shangri-La Meteorological Station during 1959–2019, Mountains of southwest China

temperature (13.7 °C) occurring in July. The observed annual total precipitation from 1958 to 2019 was 626 mm, which fell mainly in summer (June–August, 64.3%). Less precipitation fell in autumn (September–November, 20.7%) and spring (March–May, 14.6%), and the least in winter (December–February, 0.4%). Winter weather at all these locations is characterized as cold and dry (Fig. 2).

2.2 Climate data

Long-term climate observations are lacking in the study sites. We partially solved for this using data from the Shangri-La Meteorological Station (27.83° N, 99.70° E, 3276.70 m a.s.l.) as a proxy to evaluate growth–climate relationship. The station records temperatures at 02:00, 08:00, 14:00 and 20:00 Beijing Time each day, from which the daily average, daily minimum, and daily maximum temperatures are calculated. Monthly mean temperature is calculated as the averaged results of daily records for given months. For our study, monthly temperature and total precipitation data from 1958 to 2015 were downloaded from the China Meteorological Data Sharing Service System (National-Meteorological-Information-Center 2019).

2.3 Tree-ring sampling and chronology development

We collected tree-ring increment cores from six (MM, BM, JD, JJ1, JJ2 and HB) sites on north-west Yunnan Province at elevations ranging from 3900 to 4200 m a.s.l. (red circle in the Fig. 1 and Table 1) during several field campaigns from 2014 to 2016. Site MM and BM, and site JJ1, JJ2 and HB are close to each other with around 20 km from each other. Site JD has around 70 km away from site MM and HB. The average spatial distance is around 80 km. More than 25 trees were selected near the upper tree-lines with open canopy at each site. One to three cores per tree were collected at breast

Table 1 Site information and tree-ring chronologies statistics

Sites	MM	BM	JD	HB	JJ2	JJ1	RC
Location (lat./lon.)	28.37 °N/98.78 °E	28.38 °N/98.99 °E	28.10 °N/99.63 °E	27.35 °N/100.07 °E	27.17 °N/99.98 °E	27.20 °N/99.96 °E	–
Elev. (m a.s.l.)	4000	4100	3900	4200	4000	4188	–
Trees	40	25	35	31	30	33	194
Time span	1707–2015	1651–1999	1504–2013	1725–2014	1747–2014	1710–2015	1504–2015
Duration (trees)	309	349	509	290	268	306	511
Mean length of series	160	237	250	160	168	178	190
MS	0.18	0.15	0.17	0.15	0.16	0.16	0.17
Rbar	0.33	0.44	0.38	0.45	0.38	0.42	0.37
EPS > 0.90 (trees)	1820 (14)	1734 (8)	1664 (9)	1806 (5)	1803 (8)	1784 (2)	1653 (10)

Lat. latitude, Lon. longitude, Elev elevation, MS mean sensitivity, Rbar mean inter-series correlation, EPS expressed population signal

height with increment borers (Haglof, Sweden). All tree-ring samples were collected from trees from natural forest stands without evidence of fire, insect damage or human disturbance.

Samples were processed following dendrochronological practices (Stokes and Smiley 1996). After air-drying, the cores were mounted on sample holders. The wood surface was prepared with razor blades, followed by sanding with progressively finer sandpaper (from grade 300 to 1200), in order to increase the visibility of individual tree-ring boundaries. Ring-width was measured with a LINTAB II measuring system (Rinntech Inc., Heidelberg, Germany) with a precision level of 0.01 mm. Dates for all the cores were cross-validated by pattern matching and statistical tests were conducted in the software package TSAP (Rinn 2005). The quality of the cross-dating was checked using COFECHA to ensure exact dating for annual ring-width series and Pearson correlation for comparison of series of ring-widths from individual trees with those of a master series (Holmes 1983).

Each raw series of measurement was detrended using the program RCSigfree (Tree-ring-Lab-of-Columbia-University 2014b). We applied a data-adaptive power transformation to all raw ring-width data to reduce the potential heteroscedasticity found in the raw ring-width measurements (Cook and Peters 1997). We applied conservative detrending with negative exponential curves or linear regression curves of any slope to detrend all series at each site. To assess the ability of tree-rings to track recent temperature trends (Shi et al. 2017), we further applied age-dependent spline to the detrended series that resulted from the previous step. Ring-width values were calculated as the residuals between actual ring-width values and estimated values. We used the robust biweight mean method (Cook and Kairiūkštis 1990) to average all the series into a chronology. To reduce the potential trend distortion in the chronology, we applied the Melvin signal-free method (Melvin and Briffa 2008). We, finally, used the Keith Briffa Rbar-weighted approach to stabilize variance of the chronologies (Osborn et al. 1997). At last, we employed mean inter-series correlation (Rbar) and Expressed population signal (EPS) (Wigley et al. 1984) to assess the strength of the chronology. Both Rbar and EPS were calculated in 31-year running windows. We used the threshold EPS value of 0.90 to determine the most reliable time span of each chronology (Shi et al. 2017).

2.4 Development of regional composite chronology

Although, the lengths of most chronologies are more than 300 years, their reliable lengths are generally less (Table 1). In order to develop a long-term and well replicated regional composite chronology (RC chronology), we compared similarities among six site chronologies by calculating the Pearson correlation coefficient (Tang 2008) among them over

their common period of 1820–2015, and then compared the growth–climate relationships. We found that six site chronologies, all within around 80 km from each other, were significantly correlated (Appendix Table S3), and they consistently and significantly correlated with temperatures from November–January (Appendix Fig. S2). Consequently, we pooled all the raw ring-width measurements from the six sites to build a final RC chronology for more than 363 years. The development of the RC chronology followed the same procedure and method of chronology development described above.

2.5 Growth–climate response

We used Pearson correlation analysis to study the growth–climate relationship. To test whether the growth–climate relationship is caused by climate change trends such as warming trends, we calculated the Pearson correlation coefficients (Tang 2008) for two datasets: raw data of chronology and monthly climate variables, and first difference data (the value of the current year minus the value of the previous year) of chronology and monthly climate variables. Where those two datasets resulted in similar correlation coefficients, the growth–climate relationship was not considered to be caused by climate change trends (Shi et al. 2017).

2.6 Temperature reconstruction and validation

Based on the significant temperature–growth relationships, we developed a linear reconstruction model to reconstruct temperature variations over the past three and half centuries using the RC chronology as a predictor. Calibration and verification tests were conducted to evaluate the goodness of reconstruction model between reconstructed and the observed temperatures (Huang et al. 2018). The full period 1959–2015 was split into 1959–1986 and 1987–2015, respectively as the calibration and verification periods. Then the calibration/verification periods were changed to 1986–2015 and 1959–1987. We used the Pearson correlation coefficient (r), variance explained (R^2), sign test (ST), and reduction of error (RE) to evaluate the performance of the linear reconstruction model (Fritts 1976). We follow the standard that a positive value for RE indicates a valid reconstruction model (Bi et al. 2017). We also applied the Durbin–Watson (DW) test to see if the residuals of the reconstruction model were normally distributed, if the value of DW was close to two the residuals of the reconstruction model were considered to have a normal distribution (Draper and Smith 1998). To ensure the most reliable temperature reconstructions, we chose those year with EPS (Expressed Population Signal) > 0.90, and a number of sampled trees > 10, as the starting point for the final temperature reconstructions.

To validate the reliability of our reconstruction, we compared our reconstruction with three temperature related records. Firstly, for the periods with instrumental records such as the years after 1959 in our case, we compared the spatial similarity between instrumental and reconstructed data by calculating the spatial correlations between instrumental, reconstructed temperature and temperature data from $0.5^\circ \times 0.5^\circ$ CRU TS4.03 grids (Climatic-Research-Unit 2017) from 1959 to 2015. This analysis was performed using the Royal Netherlands Meteorological Institute (KNMI) climate explorer (KNMI 1999). Secondly, For the periods without instrumental records, we graphically compared our reconstruction with four other tree-ring-based winter temperature reconstructions (Appendix Table S5), and calculated the Pearson correlation coefficient (Tang 2008) between them. Two reconstructions come from high-elevation sites in southeast Tibet Autonomous Region in middle-latitude regions (Huang et al. 2018; Zhang et al. 2015) (purple and orange triangles in Fig. 1), one reconstruction came from a low-elevation site in west Sichuan Province in middle-latitude regions (Song et al. 2007) (blue triangle in Fig. 1), and one reconstruction came from a low-elevation site in northwest Yunnan Province in low-latitude regions (Shi et al. 2017) (green triangle in Fig. 1). In addition, we used the historical documented evidence from Tibet Autonomous Region, Sichuan Province, and Yunnan Province to validate the signal of cold and warm extremes embedded in our temperature reconstruction.

2.7 Temperature variation and its possible drivers

To assess the variation of temperature during the past three and half centuries, we first defined a period as a *cold* period or *warm* period if the average value for this period was more than e 0.5 standard deviations below or above the average temperature during past 363 years. We then defined a year as the *cold* (*coldest*) or *warm* (*warmest*) if the reconstructed temperature values exceeded 1.5 standard deviations (or two standard deviations) below or above the average temperature during past 363 years. Finally, we counted the number of *cold* and *warm* seasons for each decade for comparison.

The Atlantic Multidecadal Oscillation (AMO) offers a good representation of the alternation of warm and cold phases in North Atlantic sea surface temperatures (Kerr 2000). When the AMO is in the cold phase, it induces high surface pressure over the North Atlantic, which extends to the Eurasian Continent, strengthens the Mongolian Cold High and East Asian cold air activity, and results in a colder winter in East Asia (Kerr 2000). Previous studies have shown a significant positive relationship between the AMO and land surface air temperature in China, and the influence of the AMO on winter temperature is most significant in southwest China (Shi et al. 2017; Wang et al.

2009). So, winter temperature variation in Yunnan may relate to AMO. In order to test whether the large-scale climate driver AMO influences tree radial growth through its effects on local climate, we performed Pearson correlation analysis between the reconstructed AMO series and winter temperature reconstructions including present reconstruction and other four reconstructions in montane southwest China (Appendix Table S8). In addition, large volcanic eruptions lead to a massive release of sulphur into the stratosphere, cooling the land surface by reflecting sunlight. It may also influence temperature variation in the surrounding regions. Here, we also graphically compared the relationship between the cold periods in our reconstruction and volcano eruption events in east Asia (8°S – 30°N / 90°E – 129°E) with a volcanic eruption index (VEI) greater than four (Global Volcanism Program 2013). We also used wavelet transform and multi-taper spectrums to evaluate the periodicity of the reconstructed temperature. We conducted wavelet analysis with the package of “dplR” (Andrew 2008) in the R programming language software (Core-R-Team 2019), and multi-taper analysis with the MtM program (Tree-ring-Lab-of-Columbia-University 2014a).

3 Results

3.1 Composite chronology statistics

The final length of the RC chronology was 363 years (1653–2015) (Table 1). The Rbar of the RC chronology was 0.37, indicating that the RC chronology is suitable for the study of growth–climate relationships. Although our RC chronology extended back to the year 1550, the EPS of the RC chronology was greater than 0.90 only after the year 1653 with 10 trees, suggesting that measurements of variations may be more reliable after the year 1653 (Appendix Fig S1). We used the year 1653 as the starting point for the final climate reconstructions.

3.2 Response to climatic variables

The RC chronology had significant correlations ($p < 0.01$) with mean temperatures in November and December in the preceding year and January, March, April, June, and August in current year (Fig. 3). The only significant relationship for radial growth and precipitation was detected in June. The radial growth of *A. georgei* was significantly correlated with different winter temperature windows from November to April for raw data (Fig. 3a) and first difference data (Fig. 3b) of the chronology and temperature, and with different summer temperature windows from Jun to September for raw data of the chronology and temperature (Fig. 3a). Among them, early winter season temperature

Fig. 3 Correlation coefficients between regional composited (RC) chronology of *Abies georgei* and monthly mean temperature and precipitation from the Shangri-La Meteorological Station for raw data (a) and first difference data (b). Horizontal dashed lines denote the 99% confidence intervals. “p” on the horizontal axis refers to month in the preceding year

(EWST) from November to January had the highest correlation coefficient ($r=0.71$, $p<0.01$) with the growth of *A. georgei* both for raw data ($r=0.71$, $p<0.01$) and first difference data ($r=0.68$, $p<0.01$) (Fig. 3). We, therefore, used the standard RC chronology as a predictor to reconstruct the EWST in this region.

3.3 Temperature reconstruction and validation

We developed a linear reconstruction model (Model = $4.72 \times \text{RC} + 5.70$, $r=0.71$, $p<0.01$) to reconstruct EWSTs for northwest Yunnan Province back to the year 1653. The reconstruction model explained about 50% of the total variance over the common period from 1959 to 2015 (Table 2). The RE value (RE=0.45) was positive, and the result of the sign-test was also statistically significant, further suggesting the robustness of the reconstruction model. The value of statistic DW of Durbin–Watson test was approximately two, suggesting that the residuals of the reconstruction model were normally distributed (Table 2).

The actual and reconstructed EWST were consistent both at low frequency data (Fig. 4a) and high frequency data (Fig. 4b).

We found a general agreement between the spatial correlations of instrumental records (Fig. 5a) and the reconstructed EWST (Fig. 5b) from 1958 to 2015. This result also further suggests that the EWST of montane southwest China is spatially coherent with the temperatures in the Himalaya region, Indian peninsula, southwest China, Myanmar, Thailand, Laos, Cambodia and south Vietnam (Fig. 5). Most of the *cold* and *warm* periods of our reconstruction matched with those of other winter temperature reconstructions in montane southwest China (Fig. 7). And the Pearson correlation coefficients between those winter temperature reconstructions were all significant at the 0.01 level (Appendix Table S5). Furthermore, we found around 91% of the *coldest* years corresponded to cold waves or chilling damage events, and 57% of the *warmest* years were also recorded as a warm year in montane southwest China (Appendix Table S6). All these results support the reliability of our reconstruction.

Table 2 Calibration and verification statistics for the tree-ring based early winter season temperature (EWST) reconstruction

Calibration					Verification					
Period	r	R^2	ST	RE	Period	r	R^2	ST	RE	DW
1959–1986	0.68	0.46	20+/9–**	–	1987–2015	0.68	0.46	21+/7–**	0.43**	1.72
1986–2015	0.64	0.41	22+/7–**	–	1959–1987	0.64	0.41	21+/9–**	0.46**	1.83
1959–2015	0.71	0.50	42+/15–**	–	–	–	–	–	–	–

r correlation coefficient, R^2 R square, ST sign test which counts the number of agreement and disagreement between the reconstructed and instrumental series, RE the reduction of error, DW Durbin–Watson test

The 0.01 significant levels are indicated by **

Fig. 4 Comparisons between the instrumental and reconstructed mean early winter season temperature (EWST) for their common period of 1959–2015 for raw data (a) and first difference data (b); EWST reconstruction in northwest Yunnan Province since the year 1653 along with 11-year smoothing average (c), horizontal dashed lines represent the long-term mean and 1.5 standard deviations (σ) from the mean

3.4 Temperature variation

The mean and standard deviation of EWST reconstructions were -0.94 °C and 0.58 °C, respectively, during past 363 years. There were five warm periods (temperature value $> \text{mean} + 0.5\sigma$) during 1658–1719, 1743–1755, 1771–1791, 1929–1959, and 1995–2015, and five cold periods (temperature value $< \text{mean} - 0.5\sigma$) during 1720–1742, 1792–1852, 1860–1883, 1905–1928, and 1960–1994 (Fig. 4c). The periods 1929–1959 and 1792–1852 represented the longest warm and the longest cold periods respectively (Fig. 4c). While the warmest period was 1929–1959, and 1960–1994 was coldest.

The EWST reconstructions revealed 25 anomalous warm seasons (temperature value $> \text{mean} + 1.5\sigma$) and 25 anomalous cold seasons (temperature value $< \text{mean} - 1.5\sigma$) over the past 363 years (Table 3). These seasons accounted for 7% of the total years (Table 3). Three coldest decades occurred in the 1980s, 1810s and 1960s, while three warmest decades occurred in 1950s, 1940s and 2000s. In addition, the frequency of cold early winter seasons has significantly reduced since 1990s (Table 3). Multi-taper spectrum and wavelet analysis together revealed high frequency peaks mainly in 2-year periods, and significant low frequency peaks ranging from 50 to 80-year periods (Fig. 6).

4 Discussion

4.1 Growth response of *A. georgei* to climate at the alpine tree-line

We investigated the growth–climate relationships for *A. georgei* on northwest Yunnan, montane southwest China. We found that winter and early spring (November to April) temperature has significant positive effects on the radial growth of *A. georgei* around the tree-line (Appendix Fig. 4 and Fig. S2). This phenomenon was frequently observed in alpine tree-line species in montane southwest China (Appendix Table S4). For example, trees from tree-lines of Mt. Baima (Fan et al. 2009a), Mt. Shika in northwest Yunnan (Panthi et al. 2018), and Mt. Sygera in the southeastern Tibetan plateau (Liu et al. 2016) have shown similar response patterns. Similar growth–response patterns were also documented in other mountainous regions (Appendix Table S4). These findings suggest that tree radial growth is likely to be linked to winter temperatures in high-elevation zones, where warmer winter temperatures might benefit tree growth in the subsequent season, and then result in wider rings.

In montane southwest China, the cambium activity of *A. georgei* at tree-line ecotone (4360 m a.s.l.) starts in late May and ends in early October (Li et al. 2017). So, winter

Fig. 5 Spatial correlations between observed early winter season temperature (EWST) from the Shangri-La Meteorological Station (**a**), the EWST reconstruction in present study (**b**) and EWST from CRU TS 4.03 during the period 1959–2015. Only correlations at 95% significant level are shown. Black points are the location of sampling sites

temperatures may indirectly affect the growth of *A. georgei* through regulating the onset date and duration of cambium activity. Lower winter temperatures generally lead to thicker frozen soil layers and delay of snowmelt and cambium onset date. This may shorten the length of growing season and limits tree growth (Hollesen et al. 2015; Shi et al. 2017). This explanation is further supported by the monitoring study of cambium activity. For example, Li et al. (2017) monitored the cambium activity of *A. georgei* in the Sygera Mountains (around 500 km northwest of our sites) from 2007 to 2010. They found that the latest onset and the shortest duration of cambium activity occurred in 2008 with the coldest winter, the latest snowmelt and soil thawing, while the earliest onset of cambium and the longest duration of cambium activity was observed in 2010 with highest winter temperature during four studied years (Li et al. 2017). Additionally, winter temperatures may also indirectly affect the growth of *A.*

georgei by affecting photosynthetic potential in the following season. *A. georgei* is an evergreen coniferous species that can maintain photosynthesis in winter and spring (An et al. 2014). Colder winters can trigger bud damage, winter frost desiccation, reduce root activity (Körner 2012) and decrease the tree's photosynthetic potential. Warmer conditions and less physiological injuries, on the other hand, can favor synthesis and storage of carbohydrates that can be used in cambium growth in the growing season.

We also found a positive relationship between summer (June and August) temperatures and tree-ring width of *A. georgei* (Fig. 3). Similar response pattern was reported in northwest Yunnan on Mt. Baima and Mt. Shika (Fan et al. 2009a; Panthi et al. 2018), and on Mt. Sygera on the southeastern Tibetan plateau (Liang et al. 2009; Liu et al. 2016). High summer temperatures could directly promote tree root growth and cambial activity, which can further enhance tree growth. Furthermore, past studies have demonstrated that warm winters may limit the radial growth of some species at high elevations (Bi et al. 2017; Lu et al. 2015). Tree-ring widths of broadleaf species in the same region such as *Rhododendron* have been shown to be negatively correlated with winter temperatures (Bi et al. 2017; Lu et al. 2015). Such discrepancies in plant growth responses to winter temperatures at high-elevations are likely due to the differences in species and life-form, and require further research.

4.2 Comparison of reconstructed EWST with other winter temperatures in montane southwest China

Most of the *cold* and *warm* periods of our EWST reconstruction generally matched well with other four winter temperature reconstructions in montane southwest China (Fig. 7). This offers further support for the reliability of our EWST reconstruction. However, spatiotemporal discrepancies still exist in those winter temperature reconstructions (Fig. 7). This difference may be due to differences in site and geographic conditions such as latitude and elevation. Our EWST reconstructions are highly correlated with the winter temperatures reconstructed by Huang et al. (2018) in Tibet Autonomous Region and share low correlation coefficients with the winter temperatures reconstructed by Song et al. (2007) in Sichuan Province. This is probably due to the differences in site elevation. The sites of present study and of Huang et al. (2018) were both located in high-elevation regions (> 4000 m a.s.l.), while the site of Song et al. (2007) was located at low-elevation region (around 3000 m a.s.l.). Spatial distance between sites may also play a role as the site of the Song et al. (2007) was around 800 km further away from our study area (Appendix Table S5).

The recent warming phenomenon in our study site seems to have started around 1985, later than the years that found

Table 3 Early winter season temperature (EWST) and the number of anomalously *cold* and *warm* seasons at decadal scale

Period	Mean (°C)	TD (°C)	No. C	No. W	Period	Mean (°C)	TD (°C)	No. C	No. W
1650–1659	−1.25	−0.31	0	0	1840–1849	−1.17	−0.23	0	0
1660–1669	−0.88	0.06	0	0	1850–1859	−0.83	0.11	0	0
1670–1679	−0.73	0.21	0	0	1860–1869	−1.47	−0.53	2	0
1680–1689	−0.56	0.38	0	0	1870–1879	−1.36	−0.42	2	0
1690–1699	−0.29	0.65	0	3	1880–1889	−0.88	0.06	0	0
1700–1709	−0.46	0.48	0	2	1890–1899	−1.05	−0.11	0	0
1710–1719	−0.73	0.21	0	0	1900–1909	−1.14	−0.2	2	0
1720–1729	−1.16	−0.22	1	0	1910–1919	−1.45	−0.51	1	0
1730–1739	−1.41	−0.47	1	0	1920–1929	−1.09	−0.15	0	0
1740–1749	−0.67	0.27	0	0	1930–1939	−0.32	0.62	0	4
1750–1759	−0.72	0.22	0	0	1940–1949	−0.17	0.77	0	4
1760–1769	−0.96	−0.02	0	0	1950–1959	−0.08	0.86	0	6
1770–1779	−0.63	0.31	0	0	1960–1969	−1.48	−0.54	3	0
1780–1789	−0.52	0.42	1	0	1970–1979	−1.39	−0.45	1	0
1790–1799	−1.06	−0.12	0	0	1980–1989	−1.64	−0.70	5	0
1800–1809	−1.09	−0.15	0	0	1990–1999	−1.07	−0.13	1	0
1810–1819	−1.57	−0.63	2	0	2000–2009	−0.20	0.74	0	4
1820–1829	−1.42	−0.48	3	0	2010–2015	−0.46	0.48	0	2
1830–1839	−1.11	−0.17	0	0	1653–2015	—	0	25	25

TD temperature departure, No. C number of *cold* early winter seasons, No. W number of *warm* early winter season

Fig. 6 Multi-taper spectrum of the reconstructed early winter season temperature (EWST), with 95% and 99% confidence level inferred from red noise spectra, and significant period at 95% level are marked (a); wavelet transform of our EWST reconstruction, with significant periods ($p < 0.05$) highlighted by solid black lines (b)

in middle-latitude regions in southeast Tibet Autonomous Region (Huang et al. 2018; Zhang et al. 2015) and Sichuan Province (Song et al. 2007) (Fig. 7). But this recent warming phenomenon in our study site was earlier than that found in the nearest series of Shi et al. (2017) at a middle-elevation site in same area. The magnitude of warming in our study area was also more prominent as indicated by higher standardized coefficient of temperature in the linear model (Fig. 7). This suggests that the recent warming phenomenon over montane southwest China may have started earlier at middle-latitude region, and with greater magnitude of warming in higher-elevation regions. These findings support the general speculation that climate warming started earlier in high-latitude and high-elevation regions and more prominent at higher elevations (Fan et al. 2011; IPCC 2014). More evidence is still required to validate this conclusion.

4.3 Relationship between winter and summer temperatures across long-term scale in Yunnan

Meteorological station records over the past several decades show that variations in winter temperature are more prominent than those in summer (Fan et al. 2011; IPCC 2014). However, few studies have explored these seasonal temperature variations over long-time scales (Cai et al. 2016; Liu et al. 2009). Here, we explored the relationship between our EWST reconstruction and tree-ring based summer temperature reconstructions from Yunnan

Fig. 7 Comparison between reconstructed early winter season temperature (EWST) in the present study at low-latitude ($0\text{--}30^\circ\text{N}$) regions with high elevation ($4000\text{ m a.s.l.} \leq \text{elevation}$) sites in Yunnan Province (**b**) and other tree-ring based winter temperature reconstructions in montane southwest China, Southeastern Asia. **a** Winter mean temperature reconstruction based on tree-ring width of *Picea likiangensis* and *Sabina tibetica* at middle-latitude ($30\text{--}60^\circ\text{N}$) regions with higher elevation ($4500\text{ m a.s.l.} \leq \text{elevation}$) sites in southeastern Tibetan Plateau (Huang et al. 2018); **c** winter mean temperature reconstruction based on the tree-ring width of *Pinus yunnanensis* at low-latitude regions with middle elevation ($3000\text{ m a.s.l.} \leq \text{elevation} \leq 3500\text{ m a.s.l.}$) sites in Yunnan Province; **d** winter minimum temperature reconstruction based on the tree-ring of *Picea likiangensis*

at middle-latitude regions with high elevation sites in southeastern Tibetan Plateau (Zhang et al. 2015); **e** winter minimum temperature reconstruction based on tree-ring width of *Abies chensiensis* at middle-latitude regions with low elevation ($\text{elevation} \leq 3000\text{ m a.s.l.}$) site in Sichuan Provinces (Song et al. 2007) (more detail information in Table S5 in Appendix and Fig. 1). All temperature series have been standardized using the Z-score method during the common period 1653–2015. The thin and thick lines represent the annual average temperature and an 11-year smoothing average temperature, respectively. Shaded areas show the trend consistency of reconstructed temperature from different studies. Dark-grey and grey shaded areas denote cold and warm period, respectively

Province (Fan et al. 2009b, 2010; Li et al. 2012). Our EWST reconstruction generally showed good agreement with summer temperature variations at annual and decadal scale in Yunnan province, and the Pearson correlation coefficients were all significant at the 0.01 level (Appendix Fig. S3). Similar relationships were also found in south central China (Cai et al. 2016). The 1930–1950s was likely the warmest period, and the 1960–1980s was the coldest period in the last few centuries both in summer and winter seasons (Appendix Fig. S3).

However, the variations in winter temperatures are higher than those in summer as indicated by higher standard deviation (Appendix Table S7). This variation seems to have increased in recent decades. For example, the summer temperatures consistently display a decreasing trend from 1960 to 1980 (dark-grey shaded areas in Fig. S3), and increasing trend since 1980 (light-grey shaded areas in Fig. S3). However, the winter temperature displays a

short decreasing trend followed by a short increasing and then by a decreasing trend from 1960 to 1985, and displays a continuous increasing trend since 1985. In addition, the recent warming amplitudes in winter are much bigger than that in summer as indicated by higher standardized coefficient of temperature in linear model (Appendix Fig. S3). According to Shangri-La Meteorological Station, summer temperatures also show a decreasing trend before 1975, followed by an increasing trend (Appendix Fig. S3c), while winter temperatures display a continually increasing trend but with high fluctuation especially before 1980s (Appendix Fig. S3d). The increasing amplitude of winter temperatures is higher than that of summer temperatures as indicated by higher standardized coefficient of temperature in linear model (Appendix Fig. S3c and Fig. S3d).

Additionally, we also compared the winter temperature reconstructions developed by Shi et al. (2017) in Yunnan and tree-ring based summer temperature reconstructions

in Yunnan (Fan et al. 2009b, 2010; Li et al. 2012). We found that winter temperature reconstructions from (Shi et al. 2017) also has significant positive relationships with summer temperature reconstructions (Appendix Fig. S4). But the winter temperature reconstructions from (Shi et al. 2017) has higher standard deviation ($SD = 0.73$) than summer temperature reconstructions (SD ranges from 0.24 to 0.56) (Appendix Table S7), and higher warming rates in recent decades as indicated by higher standardized coefficient of temperature in linear model (Appendix Fig. S4). These results provide further evidence that variation in winter temperature is more prominent in winter season than that in summer in Yunnan.

Dendroclimatological studies in south China (Cai et al. 2016; Liu et al. 2009) and the Tibetan Plateau (Gou et al. 2008) also suggest that more pronounced warming in winter than that in summer during recent decades. All the above cases likely imply that variation in winter temperatures may be stronger than that in summer temperatures in montane southwest China and south China, but more evidence is needed to support this conclusion. The temporal discrepancies among winter and summer temperature reconstructions may reflect the asymmetry of temperature variation. Therefore, further efforts should be taken to develop more tree-ring networks to reconstruct temperatures for different seasons in different representative regions.

4.4 Possible drivers of EWST variation

The main low frequency peaks of our EWST reconstruction range from 50 to 80 years (Fig. 6), which fall within the variability of large-scale oscillations of AMO with a periodicity of roughly 40 to 80 years (Kerr 2000), suggesting possible teleconnection between the EWST in montane

southwest China and AMO. Our EWST reconstructions match well with two reconstructed AMO series (Gray et al. 2004; Mann et al. 2009) in the past three and half centuries, and the Pearson correlation coefficients between them were all significant at the 0.01 level (Appendix Fig. S5). In addition, the winter temperature reconstructions developed by Huang et al. (2018), Shi et al. (2017), Zhang et al. (2015) and Song et al. (2007) also have consistent significant positive relationships with AMO series (Appendix Table S8). All these results provide further evidence of the teleconnection between temperatures in montane southwest China and AMO. In autumn and winter, when the AMO is in the warm phase, it induces low surface pressure over the North Atlantic, which extends to the Eurasian Continent, weakens the Mongolian Cold High and East Asian cold air activity, and results in warmer winters in East Asia (Wang et al. 2009; Shi et al. 2017).

The major cold intervals during the nineteenth and twentieth centuries in our EWST reconstruction are also likely due to large volcano eruption events in low-latitude regions of south and east Asia. Because, there were 14 large eruption events ($VEI \geq 4$) during the nineteenth century and 10 eruption events ($VEI \geq 4$) during the twentieth century (Fig. 8), including two of the largest volcano eruptions in the past 500 years (Global Volcanism Program 2013). Previous studies also suggest that the cold periods in 1810s and 1980s in Nepal and China may relate to large volcano eruption events in low-latitude regions of south and east Asia (Cook et al. 2003; Huang et al. 2018; Shi et al. 2017). Large volcanic eruptions lead to massive releases of sulphur into the stratosphere, cooling the land surface by reflecting sunlight. For example, the solar radiation decreased around 25–30% (Ellsworth and John 1992), mean solar radiation was lower than the 600 years mean during 1790–1830 (Appendix Fig. S6) after volcano

Fig. 8 Comparison of our early winter season temperature (EWST) reconstructions and volcanic eruption times and volcanic eruption index in low-latitude regions of southeast Asia (8°S – 30°N / 90° – 129°E). Volcanos name with an eruption index greater than four are shown by name above the x-axis

eruption events in this region, which provides indirect support for the cooling effect of volcanic eruptions.

Our findings suggest that AMO might be the most important driver of the multidecadal winter temperature variation in montane southwest China, although large volcano eruption events also contributed to the cold winters. Given that winter temperatures are one of the most critical factors for plant growth and forest dynamics (Körner 2003). Future research should consider AMO in generating management plans for montane forests in southwest China. This has broad implications for how forests should be managed, maintained and what can be expected for future forest distribution.

5 Conclusion

Given the importance of winter temperatures for montane forest dynamics, we should seek to better understand the variations of winter temperature across long time-scales as well as the possible drivers of the variations. This is particularly important in low-latitude (0–30° N) montane regions where much of the tree production takes place. Based on a network of six tree-ring width chronologies in northwest Yunnan, montane southwest China, we extended the early winter season (November–January) temperature reconstruction of Yunnan back to the year 1653. This is by far the longest existing winter temperature reconstruction for Yunnan Province, and the third longest winter temperature reconstruction in montane southwest China. This winter temperature reconstruction is generally consistent with other winter temperature related records. The 1980s and 1950s are the coldest and warmest decades over the past three and half centuries. AMO may be the most important factor for winter temperature variation in montane southwest China over the past three and half centuries, although large volcano eruption events also contributed to cold winters during the nineteenth and twentieth centuries. The present study adds a new and relatively long winter temperature reconstruction for montane southwest China. However, there are still few winter temperature reconstructions in this region and those existing reconstructions are mainly concentrated in upper and middle regions of the Yangtze River. More tree-ring based winter temperature reconstructions are still needed in order to improve our understanding and provide a comprehensive picture of climate history and its influence on forests in the region.

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Affiliations

Yingfeng Bi^{1,2,3} · Whitney Cory⁴ · Zenxin Fan⁵ · Jingchao Yang⁶ · Zhikun Wu⁷ · Wei Huang^{1,2,3} · Jianwen Li^{1,2,3,8} · Xuefei Yang^{1,2,3,8}

¹ Key Laboratory for Plant Diversity and Biogeography of East Asia, Kunming Institute of Botany, Chinese Academy of Sciences, Kunming 650201, China

² Key Laboratory of Economic Plants and Biotechnology, Kunming Institute of Botany, Chinese Academy of Sciences, Kunming 650201, China

³ Yunnan Key Laboratory for Wild Plant Resources, Kunming 650201, China

⁴ Institute of Crop Science and Resource Conservation (INRES), Department Horticultural Sciences/ Center for Development Research (ZEF, University of Bonn, Auf dem Hügel 6, 53121 Bonn, Germany

- ⁵ Key Laboratory of Tropical Forest Ecology, Xishuangbanna Tropical Botanical Garden, Chinese Academy of Sciences, Menglun 666303, Mengla, China
- ⁶ State Forestry and Grassland Administration, Kunming Institute of Survey and Design, Kunming 650216, China

- ⁷ Guizhou University of Traditional Chinese Medicine, Guiyang 550025, China
- ⁸ Southeast Asia Biodiversity Research Institute, Chinese Academy of Sciences, Menglun Yunnan 666303, Mengla, China