Absence of late-summer warming trend over the past two and half centuries on the eastern Tibetan Plateau

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\section*{1. Introduction}

Past temperature variability has attracted intense research attention in recent decades (Jones and Briffa, 1992; Jones and Moberg, 2003; Bekryaev et al., 2010; Leclercq and Oerlemans, 2012). The Fifth Assessment Report of the IPCC (Intergovernmental Panel on Climate Change) indicates an increase in global mean surface air temperature of approximately 0.89 °C during the past century (1901–2010) over the last three decades (1979–2009). Instrumental observations since the mid-1950s indicated a strong warming trend in winter temperature. Using the tree-ring data, we reconstructed the late-summer temperature for the period A.D. 1765–2009. The reconstruction accounted for 71.4% of the variance in instrumental temperature in the period 1954–2009. The reconstructed temperature did not show a warming trend in the past 250 years, suggesting a unique characteristic of late-summer temperature variability in the study region of the Tibetan Plateau. Our tree-ring data are useful for further understanding of the plateau’s thermal condition in late summer and its role in regulating southward retreat of the Southwest Asian monsoon.

Climate on the Tibetan Plateau is often considered sensitive to global changes, yet the degree of sensitivity in different spaces and seasons prior to instrumental records is not clear. In this paper, we studied tree-ring maximum latewood density (MXD) to examine the late-summer (August–September) temperature variation over the past two and half centuries on the eastern Tibetan Plateau. The MXD regional chronology was established from 109 tree-ring samples of Balfour spruce \textit{[Picea likiangensis var. balfouriana (Rehd. et Wils.)]} at four high elevation sites. The chronology exhibited a high correlation ($r = 0.82$, $n = 56$, $p < 0.001$) with the mean August–September temperature. Using the tree-ring data, we reconstructed the late-summer temperature for the period A.D. 1765–2009. The reconstruction accounted for 71.4% of the variance in instrumental temperature in the period 1954–2009. The reconstructed temperature did not show a warming trend in the past 250 years, suggesting a unique characteristic of late-summer temperature variability in the study region of the Tibetan Plateau. Our tree-ring data are useful for further understanding of the plateau’s thermal condition in late summer and its role in regulating southward retreat of the Southwest Asian monsoon.

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Having an average elevation of more than 4000 m above sea level (a.s.l.) and an area of about 2.3 million km\(^2\), the Tibetan Plateau is the highest and largest highland in the world and exerts a great influence on regional and global climate through thermal and mechanic forcing (Yanai et al., 1992; Duan and Wu, 2005). The Tibetan Plateau has been widely considered as one of the ideal areas for climate change studies (Yao et al., 1996; Liu and Chen, 2000; Duan and Wu, 2006; Wang et al., 2008). Instrumental observations since the mid-1950s indicated that the Tibetan Plateau exhibited a significant warming trend in winter (0.32 °C/decade) and moderate warming trend in summer (0.09 °C/decade) and autumn (0.17 °C/decade) (Liu and Chen, 2000). Li et al. (2011) found a rapid increase rate in temperature during spring and winter (0.59 and 0.35 °C/decade, respectively), but a reduced rate during summer and autumn (0.15 and 0.17 °C/decade, respectively) at the Hengduan Mountains, southeastern Tibetan Plateau. However, a strong warming trend was observed in summer and autumn over the northeastern Tibetan Plateau (Li et al., 2006). You et al. (2010) reported that a dramatic warming (about 0.26 °C/decade) occurred during autumn on the whole Tibetan Plateau, especially for the northeastern subregion, whereas summer exhibited a much lower increase.
rate (0.14 °C/decade) than other seasons in the southwestern and southeastern Tibetan Plateau. Due to the shortage of instrumental data availability (only approximately 50 years), it is not clear whether the characteristics of summer and autumn temperature variability remain the same in periods prior to instrumental records on the Tibetan Plateau.

Pre-instrumental data of temperature are mainly obtained from proxy records. Tree-ring maximum latewood density (MXD) is such a proxy in paleoclimate studies. High temperature in summer tends to increase MXD by promoting photosynthesis and enzyme activity for latewood cell wall thickening and lignification (Yasue et al., 2000; Kirdyanov et al., 2003). The sensitivity of MXD to summer temperature has been well verified by numerous studies in North America and Europe (Briffa et al., 2002; Wilson and Luckman, 2003; D’Arrigo et al., 2004; Luckman and Wilson, 2005; Battipaglia et al., 2010). Till now, several temperature reconstructions of warm season have been developed based on MXD in northwest and southwest China (Bräuning and Mantwill, 2004; Fan et al., 2009; Duan et al., 2010; Wang et al., 2010; Chen et al., 2012). The reconstructed temperature records of warm season showed an evident warming trend in northeastern Tibetan Plateau, but decreasing trend in southern Tibetan Plateau (Bräuning and Mantwill, 2004), whereas the record from Hengduan Mountains is much stable (Fan et al., 2009). Compared with the numerous dendroclimatological researches on tree-ring widths, tree-ring density and its temperature signals on the Tibetan Plateau are, however, much less studied largely due to its dependence on expensive measurement instruments. Expansion of temperature records both in space and in time will help us understand the complex pattern in seasonal temperature variations.

Herein, we report a dataset of tree-ring MXD for the past two and half centuries in the eastern Tibetan Plateau. The objectives of the study were 1) to examine if the tree-ring MXD is related with summer temperature and, if so, to reconstruct summer temperature for the past two and half centuries, and 2) to investigate if the reconstructed summer temperature shows a warming trend since the Industrial Revolution on the eastern Tibetan Plateau.

2. Materials and methods

2.1. Climate in the study area

Our study area is situated in Changdu Prefecture of the eastern Tibetan Plateau (Fig. 1). Meteorological data from two stations in the study area, Dingqing (N31°25′, E95°36′) and Changdu (N31°09′, E97°10′), showed significant correlation for monthly mean temperature and for total monthly precipitation since the observation in 1954 ($r$ ranges from 0.795 to 0.947 for temperature, and from 0.315 to 0.823 for precipitation, $n = 56$, $p < 0.05$). Therefore, we averaged the monthly mean temperature records of the two stations to represent regional temperature variations. The mean monthly temperature ranges from $-4.3 \, ^\circ\text{C}$ in January to $14.3 \, ^\circ\text{C}$ in July. Most precipitation occurs from June to September, which accounts for approximately 75% of the annual precipitation, reflecting typical monsoon climate (Fig. 2).

2.2. Tree-ring sampling and density data

The tree species under study is Balfour spruce [Picea likiangensis var. balfouriana (Rehd. et Wils.)], which is the dominant species at the four

![Fig. 1. Location of the four tree-ring density sampling sites and the two meteorological stations on the eastern Tibetan Plateau.](image-url)


study sites (Changniang, Suonangka, Sangduo and Shangka) (Fig. 1, Table 1). Other species in the area mainly include Sabina tibetica and shrubs of Rhododendron. The elevation of tree-ring sampling at these four sites was between 4119–4345 m a.s.l. Balfour spruce at three sites, Changniang, Suonangka and Sangduo, were growing close to the tree lines, and the other site Shangka was in the middle slope. Using 12-mm-diameter increment borers, core samples were collected at breast height from trees that had no sign of obvious rot and damage in the stem. In the laboratory, the MXD of these tree-ring samples was measured using Itrax MultiScanner, an X-ray microdensitometry apparatus developed by Cox Analytical Systems (http://www.coxsys.se). Before measurement, these core samples were soaked in 80 °C water for 48 h to extract resins and other movable compounds in the wood that are not related to the annual production of wood tissue (Schweingruber et al., 1978). The samples were then cut into 1.0 mm thin laths using a twin-blade circular saw with an adjustment to reach an angle of 90° to the axis of the fiber direction (longitudinal tracheids). Thereafter, the sample laths were scanned using a focused high-energy X-ray beam in steps of 20 μm with the following settings for the X-ray tube: voltage of 30 kV, current value of 50 mA and an exposure time of 120 ms. The output from this system was a 16 bit digital radiograph with a resolution of 0.01 mm (1270 dpi). The digital gray-scale image was subsequently analyzed using the software WinDENDRO (http://www.regentinstruments.com) to obtain the density data of each series.

The tree-ring samples were cross-dated visually under microscope and the measured series were quality checked using the program COFECHA (Holmes, 1983). Standard MXD chronology was developed from the cross-dated-ring density series using the program ARSTAN (Cook, 1985). Because most MXD series show a slight linear decreasing trend with age, the cross-dated MXD series were fitted with linear lines with negative slope to remove the age-related growth trends while preserving as much low-frequency variation as possible. Expressed population signal (EPS) was calculated for 30-year moving windows with 15-year overlaps, and the threshold of 0.85 in EPS was considered to denote sufficient strength of signal for use in paleoclimate reconstruction (Wigley et al., 1984).

2.3. Reconstruction of past climate

Climate-tree growth relationship was identified by examining the Pearson correlation coefficient between the MXD regional chronology and the instrumental climate records over the period 1954–2009. These analyses were conducted over an 18-month climate window extending from May of the previous year to October of the current growth year. Based on the climate–tree growth relationship, a transfer function was established in which the MXD was the predictor variable and the corresponding climatic factor was the predictand variable. Split validation was performed to verify the reliability of the transfer function. The following statistics were examined in the validation: Pearson’s correlation coefficient (R), reduction of error (RE), coefficient of efficiency (CE), sign test and product mean test (Pmt). Both RE and CE are measures of shared variance between the actual and estimated series, with a positive value suggesting that the transfer function is stable and valid (Cook et al., 1994).

To illustrate the spatial representativeness of the reconstructed temperature, the KNMI climate explorer (Royal Netherlands Meteorological Institute; http://climexp.knmi.nl) was used to calculate the spatial correlations between the reconstructed temperature and the gridded temperature data (CRU TS 3.1, Climatic Research Unit, University of East Anglia) for the period 1954–2009.

3. Results

3.1. Chronology statistics

MXD chronologies were established for each of the four sites. The four independent MXD chronologies were significantly correlated with each other (r ranges from 0.562 to 0.796 with the mean value of 0.673) for the common period 1806–2009, suggesting that climatic factors common at the four sites are influencing the tree growth in the region. Therefore, we pooled the MXD series from the four sites to develop a regional chronology (RC) (Fig. 3a). The statistics of the MXD regional chronology were shown in Table 2. The length of the regional chronology was from AD 1765 to 2009 during which the EPS values were greater than 0.85. The variations of the sample depth and running EPS for the regional MXD chronology were shown in Fig. 3b.

3.2. Relationships between MXD and climate

The regional MXD chronology depicted a significantly positive correlation with mean temperature in August, September and previous December (Fig. 4). There were similar correlation patterns between the regional MXD chronology and mean maximum temperature. Significant negative correlation was also found with precipitation from August to September at the p < 0.01 significance level. This is mainly a result of association between variations in late summer temperature and variations in precipitation (r = −0.663, p = 0.001). The relationship between precipitation and MXD disappeared in partial correlation analysis (r = −0.081, p = 0.554). The highest correlation (r = 0.82, p = 0.001) was found between the regional MXD chronology and mean August–September temperature, indicating that temperature in late summer was the most important factor controlling tracheid cell wall thickening. However, the fitting of linear model underestimated temperatures with high values (approximately above 13.4 °C) (Fig. 5). Further examination of the MXD chronology and mean August–September temperature showed that a non-linear relationship existed between the two series (Fig. 5).

![Fig. 2. Annual precipitation (gray bars, left axis) and temperature (black line, right axis) averaged from the two nearest meteorological stations (Dingqing, Changdu) on the eastern Tibetan Plateau.](image)

Table 1

<table>
<thead>
<tr>
<th>Site</th>
<th>Location</th>
<th>Elev. (m)</th>
<th>Cores</th>
<th>Time span (A.D.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Changniang (CN)</td>
<td>N31°18.3', E95°13.5'</td>
<td>4268–4345</td>
<td>25</td>
<td>1748–2009</td>
</tr>
<tr>
<td>Suonangka (SNK)</td>
<td>N31°15.9', E95°13.6'</td>
<td>4212–4280</td>
<td>32</td>
<td>1733–2009</td>
</tr>
<tr>
<td>Sangduo (SD)</td>
<td>N31°09.8', E95°46.4'</td>
<td>4136–4232</td>
<td>36</td>
<td>1733–2009</td>
</tr>
<tr>
<td>Shangka (SK)</td>
<td>N31°26.1', E96°41.6'</td>
<td>4119–4195</td>
<td>16</td>
<td>1625–2009</td>
</tr>
<tr>
<td>Dingqing</td>
<td>N31°22', E95°36'</td>
<td>3874</td>
<td>–</td>
<td>1954–2009</td>
</tr>
</tbody>
</table>
3.3. Reconstruction of past climate

Given the non-linear relationship between the MXD chronology and mean August–September temperature, a transfer function was developed using a logistic model for regression to particularly high temperatures as close as possible. The range of MXD indices in the period 1954–2009 covers most of the data distribution during the past two and half centuries (Fig. 5). The transfer function accounted for 71.4% of the variance in instrumental temperature over the calibration period 1954–2009 (Table 3). In order to evaluate the reliability of the logistic model, the split validation was conducted in four different ways (Table 3). The results of each validation showed positive values in RE and CE (greater than 0.6), indicating fairly reasonable skill of the logistic model (Fritts, 1976; Cook et al., 1994). The significance \( p < 0.01 \) of sign test and product mean test additionally validated the reliability of the transfer function. Thus, the reconstruction model was considered to be stable and valid over time.

Applying the transfer function to tree-ring data, the late-summer (August–September) temperature was reconstructed for the period A.D. 1765–2009. The variation of the reconstructed late-summer temperature agreed well with that of the variation of the observed temperature (Fig. 6a). They matched particularly well in extreme high temperature years (e.g. 1971, 1992 and 1994) and extreme low temperature years (e.g. 1962, 1965 and 1987). Spatial correlation analyses using KNMI climate explorer showed that the reconstructed late-summer temperature correlated significantly with regional surface temperature over grids approximately 26°–35° north latitude and 90°–102° east longitude \( r > 0.5, p < 10\% \) (Fig. 7).

3.4. Patterns of reconstructed climate

The reconstructed late-summer temperature for the past 245 years showed annual to decadal fluctuations punctuated with cool and warm periods (Fig. 6b). The warmest year was 1892 with a temperature of 14.2 °C. The temperature difference during the warmest year and the coldest year (1835) was 2.9 °C. This reconstruction also captured the cold event after the Tambora volcanic eruption in 1815 (Siebert and Simkin, 1994). From 1997 to the present, the late-summer temperature increased significantly and was much higher than the average rate over the past 245 years. Nevertheless, the recent decade was not the warmest period in the context of the past two and half centuries. The 20-year low-pass filter of the reconstruction indicated no evident warming trend in the reconstructed late-summer temperature during the past two and half centuries.

We compared our reconstruction with three summer temperature reconstructions developed from surrounding areas (Liang et al., 2008; Fan et al., 2009; Wang et al., 2010) (Fig. 8). The cold periods in 1810s–1820s, 1850s–1870s, 1890s–1920s and 1970s–1980s as shown in our study were consistent with other three reconstructions. Particularly, the cold periods in 1810s–1820s and 1890s–1920s were also widely recorded in tree rings, ice core and fluctuation of glacier from other areas on the Tibetan Plateau (Liang et al., 2009). Based on the comparison, our late-summer temperature reconstruction was in good agreement with previous reconstructions of summer temperature variation inferred from tree ring width/MXD in surrounding areas on the Tibetan Plateau.

4. Discussion

4.1. Response of tree-ring MXD to late-summer temperature

In this study, the Balfour spruce tree-ring MXD responded strongly to August–September temperature \( r = 0.82, n = 56 \), suggesting that the MXD is a reliable indicator of late-summer temperature on eastern Tibetan Plateau. Compared with tree-ring width of the same samples in this study (data not shown), the MXD data have stronger ability to
track variation in a single climate variable (Esper et al., 2012; Schrope, 2012). Several studies reported that, for some conifer species, the time of vigorous secondary cell wall formation seemed to be in August (Yamamoto et al., 1978), and the wall thickening and lignification ended in the late September or early October (Thibeault-Martel et al., 2008; Li et al., 2013). Air temperature (especially minimum temperature) constrains the ending of xylem formation for trees in high latitudes and altitudes (Rossi et al., 2008; Moser et al., 2010; Li et al., 2013). In this study, the regional climate record shows that the mean (minimum) monthly air temperature in August–September is 12.4 °C (7.1 °C) which allows the above physiological activities to proceed. The mean (minimum) air temperature in October dropped to 6.2 °C (0.5 °C) which probably ceases the xylem formation. The association between tree-ring MXD and summer temperature was also reported in other studies on the eastern Tibetan Plateau (Bräuning and Mantwill, 2004; Duan et al., 2010; Wang et al., 2010). By contrast, the MXD indicates warm season temperature (approximately April–September, i.e., the whole growing season) at high latitudes and altitudes in Europe (Frank and Esper, 2005; Grudz, 2008; Battipaglia et al., 2010) and North America (Briëta et al., 1994; Luckman and Wilson, 2005). The discrepancy in seasonal responses may be related to the timing of cell wall thickening and lignification in different climate systems.

It seems that the late-summer temperature is such a single climate variable that has dominant and non-linear effects on latewood cell-wall thickening and lignification. Such non-linearity might be a particular feature in latewood cell wall development. The MXD will reach its wood-anatomical upper threshold when the latewood cell contains almost no lumina but only cell walls (Battipaglia et al., 2010). Two types of models (linear and logistic) were tested to describe the relationship between the regional chronology and the actual August–September temperature. The logistic model exhibited the highest $R^2$ and lowest Root-MSE (Table 4). When the temperatures are close to the high ($T_N > 13 °C$) or low ($T_b < 12 °C$) values, the logistic model displayed much lower residual sum of squares (RSS) than the linear model. Validation test using leave-one-out method also showed that the logistic model exhibits much higher RE than the linear model. Therefore, we chose the logistic model to reconstruct the mean August–September temperature.

4.2. Absence in late-summer warming trend

The tree-ring density data in our study showed no evident warming trend in August–September temperature over the past 245 years on the eastern Tibetan Plateau. Although the age-related trend in tree-ring MXD was removed, the absence of warming trend is not a result of the detrending process. This is because the linear lines with negative slope were chosen to detrend the tree-ring MXD series. If the trend we removed is of climate, it would be cooling trend not warming trend. A general warming trend since the Little Ice Age was reported from studies of tree rings (Liu et al., 2005; Zhu et al., 2011), ice cores (Yao et al., 1991; Thompson et al., 2006; Yao et al., 2006) and lake sediment (Liu et al., 2006) on the Tibetan Plateau. Additionally, a speleothem record (Tan et al., 2003) from North China and tree-ring records (Briëta et al., 2013) in high latitude regions of Asia both indicated unusual warming in the 20th century for the warm season. Zhu et al. (2011) studied tree-ring widths on the southeastern Tibetan Plateau and
detected a pronounced warming trend in summer (August) minimum temperature in the recent two centuries. It should be noted that the maximum latewood density in our study mainly responded to late-summer mean temperature and maximum temperature, not the minimum temperature (Fig. 4). However, warm-season minimum temperature played strong roles on the process of MXD formation in high latitude regions (Mervi, 2005; Frank et al., 2007; Kirdyanov et al., 2008). As indicated by observed meteorological data, the minimum temperature was rising much more rapidly than maximum temperature during the recent warming on the Tibetan Plateau (Gou et al., 2008; Fan et al., 2009), Asia (Tabari and Talaei, 2011), North America (Zhang et al., 2000; Wilson and Luckman, 2003) and the globe (Easterling et al., 1997; Stone and Weaver, 2002). Tree-ring density studies in northern high latitudes showed long-term warming trend in summer temperature since Industrial Revolution (Luckman and Wilson, 2005; Büntgen et al., 2006; Grudd, 2008; Briffa et al., 2013). Terrestrial changes in summer albedo due to land use and melting of ice and snow might contribute substantially to the summer warming trend in high-latitude regions (Chapin et al., 2005; Serreze and Barry, 2011).

The absence of warming trend in August–September temperature over the past 245 years in our study might be related to the topography and vegetation cover of the region under study. Firstly, Changdu is located in a confluence area of Lancangjiang (Mekong) and several rivers, and the elevations of our sampling sites are high and close to treelines. It has been demonstrated that treelines on the southeastern Tibetan

### Table 3


<table>
<thead>
<tr>
<th>Period</th>
<th>$R^2$</th>
<th>RMSE</th>
<th>F</th>
<th>Period</th>
<th>$R^2$</th>
<th>RMSE</th>
<th>F</th>
<th>Period</th>
<th>$R^2$</th>
<th>RMSE</th>
<th>F</th>
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<tbody>
<tr>
<td>$y_{3i}$ ± 1</td>
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<td>0.346</td>
<td>88.1</td>
<td>$y_{3i}$ ± 1</td>
<td>0.888</td>
<td>0.776</td>
<td>0.734</td>
<td>$y_{3i}$ ± 1</td>
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<td>1954–1991</td>
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<td>60.4</td>
<td>1992–2009</td>
<td>0.883</td>
<td>0.766</td>
<td>0.811</td>
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<td>111.7</td>
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<td>0.613</td>
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<tr>
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<td>102.8</td>
<td>1954–2001–09</td>
<td>0.834</td>
<td>0.676</td>
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<tr>
<td>1954–2009</td>
<td>0.714</td>
<td>0.339</td>
<td>145.9</td>
<td>Logistic model: $\text{Tem}_{8–9} = 15.589 - 4.306 / (1 + (x / 1.066)^{20.689})$</td>
<td>15.589</td>
<td>4.306</td>
<td>15.589</td>
<td>15.589</td>
<td>4.306</td>
<td>15.589</td>
<td></td>
</tr>
</tbody>
</table>

$y_{3i} ± 1$, the 1st, 2nd, 4th, 5th, 7th, … year; $y_{3i}$, the 3rd, 6th, 9th year; $\text{Tem}_{8–9}$, reconstructed August–September temperature; RC, MXD regional chronology; R, correlation coefficient; $R^2$ and $R^2_{\text{adj}}$, coefficients of determination and adjusted coefficients of determination of regression analysis; RMSE, root mean square error; F, F statistic for regression model significance; RE, the reduction of error; CE, coefficient of efficiency; sign test, sign of paired observed and estimated departures from their mean on the basis of the number of agreement/disagreements; Pmt, the product mean test.

* $p < 0.05$.
** $p < 0.01$.

Fig. 6. Comparison between observed (black line) and reconstructed (red line) mean August–September temperatures for their common period 1954–2009 (a) and the reconstruction of late summer temperature from tree-ring maximum latewood density in the eastern Tibet (b). Superimposed on the reconstruction is a 20-year low-pass filter. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
Plateau hold much moisture during the rainy season (Liang et al., 2011). This abundant moisture is favorable to mitigate the potential warming by evapotranspiration (Körner, 2003; Holtmeier and Broll, 2005). Secondly, our study region is well-forested relative to central and western Tibetan Plateau (Zhuo et al., 2010). Particularly in the growing season, these forests could reduce surface heating by discharging excess

Fig. 7. Spatial correlations of instrumental (a) and reconstructed late summer (August–September) temperatures (b) with regional gridded August–September temperatures for the period 1954–2009 (p < 10%). The gridded climate dataset is from the University of East Anglia Climate Research Unit (CRU TS 3.1).

Fig. 8. Comparisons between the late-summer (August–September) temperature in this study (a) and other temperature reconstructions in the nearby regions. Reconstruction of August–September mean temperature in Changdu (Wang et al., 2010) (b); reconstruction of April–September mean temperature in Hengduan mountains (Fan et al., 2009) (c); reconstruction of June–August mean minimum temperature in the source region of the Yangtze river (Liang et al., 2008) (d). All chronologies were smoothed by 15-year moving average.
heat through evapotranspiration of water, which further leads to atmospheric cooling by increasing cloud cover (Betts, 2000). The absence of warming trend in summer or increasing trend in tree-ring growth was also reported in other studies from Changdu region (Brauning and Mantwill, 2004; Linderholm and Brauning, 2006) and eastern Tibetan Plateau (Collins et al., 2002; Gou et al., 2008; Liang et al., 2009; Fan et al., 2010).

Absence of warming trend in summer does not mean the same in other seasons. In the current study region, meteorological data in the period 1954–2009 showed that pronounced warming trend (0.32 °C/decade) occurs in winter (December–February). Long-term warming trend in winter was reported from tree-ring and ice core studies on the northeastern Tibetan Plateau (Yao et al., 1991; Liu et al., 2005; Zhu et al., 2008). The pronounced seasonality of long-term warming trend, characterized by slight increase in summer and marked warming in winter, was observed not only on the Tibetan Plateau but also in other regions. Cook et al. (2003) established a tree-ring chronology network in Nepal and found that winter temperature exhibited a strong and constant increase over the past 400 years, whereas spring–summer temperature was much stable for centuries and declined after 1960. Ice-core 618O data from Law Dome in East Antarctica over the last 700 years showed that summer temperature had relatively little change, whereas winter temperature had significant fluctuations through time (Morgan and Ommen, 1997). The recent warming trend in winter might be associated with increased concentration of greenhouse gas in atmosphere due to residential heating (Cao et al., 2007; Zhang and Tao, 2008) and decreased surface albedo due to deposition of black soot aerosols (Xu et al., 2009).

5. Conclusion

Our study reconstructed the mean August–September temperature on the eastern Tibetan Plateau for the period AD 1765–2009. The reconstruction represented 71.4% of the variance in the observed August–September temperature during the period 1954–2009. We found that there is no warming trend in the late-summer temperature over the past 245 years. Because late summer is the season that southwest Asian Monsoon retreats southward, our temperature reconstruction for this season will allow further understanding of the regional thermal condition and its role in regulating monsoon retreat. Particularly, the absence of warming trend in late summer suggests that climate change is not spatially universal on the Tibetan Plateau, and the climate features unique to the study region are worth greater attention in large spatial scale modeling of climate change.

Acknowledgments

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References
