Tree-ring density inferred late summer temperature variability over the past three centuries in the Gaoligong Mountains, southeastern Tibetan Plateau

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ABSTRACT

Long and high-resolution proxy records are still sparse in the southeastern Tibetan Plateau (TP), hampering our understanding of past climatic variability from a long-term perspective. In this study, we developed a regional maximum latewood density (MXD) chronology of *Larix speciosa* stretching up to 523 years based on 72 tree-ring cores (44 trees) collected from three sites close to the tree line in the Gaoligong Mountains, southeastern TP. This chronology responded well to temperatures during August through September and was thus used to reconstruct late summer (August–September) temperature over the period A.D. 1690–2008. The reconstruction explains 40.9% of the total temperature variance during the calibration phase. Cold conditions prevailed during the periods 1695–1702, 1806–1821, the 1850s, 1882–1889, the 1900s and the 1960s. Warm phases occurred in 1734–1745, the 1770s, 1824–1840, the 1890s, 1927–1936, the 1940s–1950s and 2002–2008. Spatial correlation with the gridded temperature data set showed that our reconstruction captures large-scale regional temperature variations for the southeastern and southern TP. Comparison with other tree-ring inferred temperature time series in the surrounding areas, glacier fluctuations and historical documental records imply a high degree of confidence for our reconstruction.

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1. Introduction

The Tibetan Plateau (TP), with an average altitude above 4000 m a.s.l. and a total area of more than 2 million km², deeply affects large-scale atmospheric circulation pattern in its neighborhood, such as the eastern and southern Asian monsoons and even the global climate system (Manabe and Terpstra, 1974). Investigating climate variation over the TP thus provides an effective way to interpret monsoon dynamics and global climate change. However, instrumental climate records on the TP are sparse and of short length (most of the climate data are available only after the 1950s), limiting our understanding of long-term climate change in the region. Therefore, it is necessary to extend the available climate data over TP back in time by developing climate proxy records.

Tree-ring data have been increasingly used to reconstruct high-resolution and long-term climate change of the TP in the last decades. However, the dendroclimatic studies on the TP are very uneven in spatial distribution. Many works focused on precipitation/drought

reconstructions on the northeastern TP (e.g. Zhang and Wu, 1997; Zhang et al., 2003, 2013; Shao et al., 2005; Liu et al., 2006) and temperature reconstructions in the eastern and northeastern parts of the plateau (e.g. Liu et al., 2005, 2009; Gou et al., 2007, 2008; Xu et al., 2011). The southern and southeastern parts of the TP, however, have received much less attention, although they are widely covered by forests. To date, limited dendroclimatic researches have been conducted in these areas, most of which employed tree-ring width or isotope to reconstruct climate history (Fan et al., 2008, 2010; Liang et al., 2009; Zhu et al., 2011; Li et al., 2012; Sano et al., 2013; Yadav et al., 2014). Few studies used tree-ring density as a climate proxy in these regions (Bräuning and Mantwill, 2004; Bräuning, 2006; Fan et al., 2009), even though tree-ring maximum latewood density (MXD) has been demonstrated to have a great potential for temperature reconstruction (Schweingruber, 1988; Briffa et al., 1992). Dendroclimatic researches involving more sites, species and tree-ring parameters are needed in the southern and southeastern TP.

Larix speciosa is a deciduous pioneer tree species mainly distributed in the Gaoligong, Nu, and Xuelong Mountains, southeastern TP (Wang and Zhang, 1992). Fan et al. (2010) found the tree-ring width of this species from the Gaoligong Mountains responded well to temperatures during the growing season and thus reconstructed mean May–August temperature history. Here in this study, we developed a regional MXD chronology of *L. speciosa* and used this chronology to infer late summer

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(August–September) temperature variations during the period A.D. 1690–2008 for our study area.

2. Materials and methods

2.1. Study area and climate

The study area was in the Gaoligong Mountains which is the border area of northwest of Yunnan province, China and northeast of Burma (Fig. 1). The Mountains are incised by deep gorges of Nujiang River and Irrawaddy River, forming extremely steep physiognomy in the region. The Gaoligong Mountains are considered as one of the biodiversity hotspots in the world (Myers et al., 2000), where the flora is highly rich, including more than 4294 seed plant species (Wang et al., 2007). Along the elevation gradient, the vegetation from low to high altitudes is respectively tropical monsoon forest (<1000 m a.s.l.), subtropical evergreen broad-leaf forest (1000–2600 m a.s.l.), temperate deciduous broad-leaf forest (1000–3000 m a.s.l.), sub-alpine coniferous forest (2700–3500 m a.s.l.), alpine meadow and tundra (>3400 m a.s.l.). The forest limit is located about 3500 m a.s.l. (Xue, 1995; Li et al., 2000).

The climate in our study area is mainly influenced by southwest Asian monsoon from the Indian Ocean, characterized by abundant precipitation during the summer half year and much less in the winter half year. Spatially, the climatic conditions (e.g. precipitation and temperature) vary according to terrain. Records from the meteorological station in Deqin (28°29'N, 98°55'E, 3319 m a.s.l.) close to our sampling site show that the mean annual temperature and precipitation during the period 1954–2008 are respectively 5.4 °C and 648 mm (Fig. 2).

Fig. 2. Monthly mean temperature (Tmean) and total precipitation (Pre) for the meteorological station in Deqin in the southeastern Tibetan Plateau during the period 1954–2008.

2.2. Sample collection and chronology development

Three sites (SKL, 42KM and DLJ) near the tree line in the Gaoligong Mountains, southeastern TP with little evidence of fire, insect attack or human disturbance were chosen for sampling (Fig. 1). These sites locate at 3200–3276 m a.s.l. In total, 72 cores from 44 trees were extracted using an increment borer at breast height, with 16, 9 and 19 trees being obtained from SKL, 42KM and DLJ sampling sites, respectively (Table 1).

In order to obtain the MXD data of *L. speciosa*, tree-ring samples were processed following the standard densitometric analysis procedure developed by the Swiss Federal Institute of Forest, Snow and Landscape

Fig. 1. Locations of the sampling sites, grid sites and meteorological station.

Table 1
Information about the sampling sites and grid sets in the southeastern Tibetan Plateau.

Sites	Latitude	Longitude	Elevation	Time span	Years
SKL	27.78°N	98.48°E	3276 m	1706–2008	303
42KM	27.80°N	98.50°E	3217 m	1798–2008	211
DLJ	27.81°N	98.41°E	3200 m	1486–2005	520
Grid set 1	27.75°N	98.25°E	–	1950–2008	59
Grid set 2	27.75°N	98.75°E	–	1950–2008	59
Grid set 3	28.25°N	98.25°E	–	1950–2008	59
Grid set 4	28.25°N	98.75°E	–	1950–2008	59
Grid set 5	27.25°N	98.25°E	–	1950–2008	59
Grid set 6	27.25°N	98.75°E	–	1950–2008	59

Research (Lenz et al., 1976; Schweingruber, 1988; Schweingruber et al., 1991). Tree cores were cut into 1.0-mm-thin sections using a twin-blade DENDROCUT, with the angles vertical to the wood fiber, which were accomplished by a DENDROSCOPE (Wang et al., 2010). The resin and sugar in the thin wood sections were removed by soaking them in 80 °C water for 72 h. After being air-dried, the sections were put into a constant-temperature-and-humidity room for more than 2 h to ensure all of them have the same water content. Then, X-ray photography was taken using these sections in the same room. This process yielded grey-scale variations in the X-ray film, from which the MXD measurements can be acquired by the DENDRO2003 instrument. The obtained MXD measurements were cross-dated using the COFECHA program (Holmes, 1983) and were verified with wood sections when inconsistencies occurred.

The dated MXD measurements were detrended by a cubic spline with a 50% frequency-response cutoff equal to 67% of the series length to remove the biological growth trend using ARSTAN program and preserve variations related to climate (Cook, 1985). This detrending method emphasized the inter-annual to multi-decadal scale variability (Cook and Kairiukstis, 1990). All detrended series were averaged to form the MXD standard chronology for each site (Fig. 3) by calculating the biweight robust means which can decrease the effect of outliers (Cook and Kairiukstis, 1990). As the site chronologies were highly

correlated with each other during the period 1798–2005 (R : 0.589–0.651, $P < 0.01$), we decided to combine all detrended MXD series from these sampling sites to develop a longer and better replicated regional MXD standard chronology (Fig. 3). The statistical characteristics of the site and regional chronologies are shown in Table 2. According to the Expressed Population Signal (EPS) threshold of 0.85 (Wigley et al., 1984; Cook and Kairiukstis, 1990), the regional chronology met the acceptable signal strength for climate reconstruction after A.D. 1690 (> 16 cores).

2.3. Climate-growth response and reconstruction

In order to evaluate the response of tree growth to climate, we conducted bootstrapped correlation analysis between the regional MXD chronology and climate data for their common period using the DendroClim 2002 program (Biondi and Waikul, 2004). The gridded temperature and precipitation data (obtained from the Climate Research Unit (CRU TS 3.10) (Mitchell and Jones, 2005; Harris et al., 2013)), averaged from six grid sets near the sampling sites (Fig. 1) were used in our study for emphasizing the spatial representation. The climate variables include monthly mean temperature (Tmean) and total precipitation from previous October through current September.

A linear regression model was established for mean August–September temperature reconstruction by using the regional MXD chronology as the predictor. Since the available temperature data set is too short to be divided into two subsets for robust calibration and verification, we employed the leave-one-out cross-validation (Michaelsen, 1987) to verify our reconstruction. Evaluative statistics include Person's correlation coefficient (r), sign test (ST) product mean test (Pmt) and reduction of error (RE) (Fritts, 1976). The cyclic patterns of the reconstructed temperature variability were investigated using the method of multiple-taper method (MTM) spectral analysis (Mann and Lees, 1996). To investigate the spatial representativity of our reconstruction, we performed a spatial correlation analysis between the reconstructed series and the gridded temperature data set of Climate Research Unit (CRU TS 3.10) using the KNMI (<http://climexp.knmi.nl>).

Fig. 3. The MXD standard chronologies (black line) from three sites (SKL, 42KM and DLJ) and the regional MXD standard chronology with 11-year low-pass filters (thick gray line) and sample depths (gray line). The year with the Expressed Population Signal (EPS) > 0.85 is marked using an arrow.

Table 2
Statistics of the site and regional MXD standard chronologies.

Chronologies	Time span	MLS	R	AC1	MS	SD	Cores/trees	EPS > 0.85
SKL	1706–2008	182	0.425	0.437	0.045	0.052	26/16	1830/14
42KM	1798–2008	131	0.521	0.424	0.057	0.062	16/9	1880/7
DLJ	1486–2005	310	0.383	0.450	0.041	0.048	30/19	1690/16
Regional	1486–2008	224	0.369	0.418	0.038	0.045	72/44	1690/16

MLS, mean length of series; R, mean series intercorrelation; AC1, first-order autocorrelation; MS, mean sensitivity; SD, standard deviation; EPS > 0.85, year/number of cores when the Expressed Population Signal exceeds the 0.85 threshold.

3. Results

3.1. Climate responses of the MXD of *Larix speciosa*

The regional MXD chronology showed significant positive correlations ($P < 0.01$) with monthly mean temperatures from August to September over the period 1950–2008 (Fig. 4). Higher correlations were exhibited between the MXD indices and seasonal temperature. The correlation coefficient for the average August–September temperature (late summer) reached 0.640 ($P < 0.01$), which was much greater than those for individually monthly temperatures. Therefore, we employed the MXD chronology to reconstruct the mean August–September temperature history in our study area.

3.2. Calibration and reconstruction of mean August–September temperature

The statistics of the calibration and verification test are shown in Table 3. The regression model ($Y = 7.355X + 6.058$) accounts for 40.9% ($R^2_{\text{adj}} = 39.9\%$) of the actual variance of the mean August–September temperature during the calibration period 1950–2008 (Fig. 5). Also, this model passes all conventional calibration statistics ($R = 0.640$, $F = 39.503$, $P < 0.01$). The leave-one-out cross validation test produced a positive RE (0.364), revealing the predictive skill of the regression model (Fritts, 1976). Statistical significant sign test ($41 + / 18 -$, $P < 0.01$) and correlation ($r = 0.605$, $P < 0.01$) between the actual values and the leave-one-out estimates indicated validity of the reconstruction. We calibrated further the first differences of the MXD indices with those of actual temperature data, which explained 36.9% ($R^2_{\text{adj}} = 35.8\%$) of the actual August–September temperature variances over the calibration period (1951–2008) (Table 3). The leave-one-out cross validation test yielded significant ($P < 0.01$) verification statistics (r : 0.499; ST: $47 + / 11 -$) and positive RE (0.254), indicating that the developed model using the first difference of MXD index as the independent variable was able to capture the temperature variability at high-frequency domain.

Fig. 4. Correlation coefficients between the regional MXD standard chronology and the gridded mean temperature (Tmean) and precipitation from the previous October to current September, as well as seasonal mean (August–September), during the period 1950–2008. The horizontal dotted lines indicate statistical significance of 0.05.

We extended mean August–September temperature data for our study area back to A.D. 1690 using this regression model, which placed emphasis on the inter-annual to multi-decadal scale variations. According to the reconstructed temperature time series (Fig. 6), the mean is 13.387°C and the standard deviation (SD) is 0.301. Extremely warm summer temperatures occurred in 1703, 1716, 1724, 1747, 1770, 1804, 1827, 1834, 1847, 1865, 1887, 1929, 1945, 1985, 1994, 2003 and 2007; singularities for low temperature summers were 1694, 1701, 1709, 1720, 1729, 1750, 1759, 1787, 1809, 1816, 1817, 1841, 1857, 1885, 1905, 1913, 1937, 1951, 1962 and 1987. On the decadal timescale, the smoothed (11-year low-pass filter) curve of the reconstructed temperature series displays obviously warm periods during 1734–1745, the 1770s, 1824–1840, the 1890s, 1927–1936, the 1940s–1950s and 2002–2008. Cold episodes appeared during 1695–1702, 1806–1821, the 1850s, 1882–1889, the 1900s and the 1960s. MTM spectral analysis of the temperature reconstruction displayed highly significantly ($P < 0.01$) spectral peaks at 2.2, 3.5 and 12.2 years (Fig. 7). Significant ($P < 0.05$) cyclic patterns were also found at 2.0, 2.1, 2.3, 2.7, 11.0, 27.5 and 59.8 years.

Spatial correlations of our reconstruction with the gridded temperature dataset during the period 1950–2008 showed that the significant ($P < 0.1$) correlation fields covered a broad region approximately 23°N – 35°N , 83°E – 105°E , located in the southeastern and southern TP, with the highest correlations ($R > 0.6$) appearing in the north–south oriented Hengduan Mountains and the eastern part of Himalayas (27°N – 32°N , 95°E – 100°E) (Fig. 8a). Additionally, the significant correlation fields emerged in southeast China, covering a much smaller area. To investigate the spatial correlations at high-frequency domain, we further correlated the first differences of the reconstruction with the gridded temperature dataset over the 1951–2008 period and even found larger significant ($P < 0.1$) correlation fields covering the region about 20°N – 36°N , 80°E – 113°E (Fig. 8b) in comparison to the original data. The closest association fields ($R > 0.6$) for the first differences, however, only confined to a much smaller region about 27°N – 32°N , 94°E – 99°E (the Hengduan Mountains and the eastern part of Himalayas), which was very similar to those of the original data. Overall, the spatial correlation results confirm that our reconstruction captures broad-scale regional temperature variations in the southeastern and southern TP, especially in the Hengduan Mountains and the eastern part of Himalayas. Moreover, the reconstruction tracks temperature variability in some areas of south-east China to some degree.

4. Discussion

4.1. The relationship of MXD and climate

A number of previous studies involving the responses of maximum tree-ring density to climate at northern high latitudes and high altitudes have demonstrated the significant positive relationships between MXD and summer temperatures (Parker and Hensch, 1971; Briffa et al., 1990; Schweingruber et al., 1991; Wang et al., 2010; Jones et al., 2013). In the present study, we found a high positive correlation of the MXD standard chronology with August–September temperature, suggesting that *L. speciosa* forests growing at the upper tree-line of the Gaoligong Mountains, southeastern TP were likely to be influenced by late summer temperature. Temperature conditions in late summer during the

Table 3

Statistics of calibration and leave-one-out verification results over the common period 1950–2008.

	Calibration (original model: $Y = 7.355X + 6.058$)				Leave-one-out verification			
	<i>R</i>	R^2	R^2_{adj}	<i>F</i>	<i>r</i>	ST	Pmt	RE
Original (1950–2008)	0.640	0.409	0.399	39.503	0.605	41 +/18 –	3.911	0.364
First difference (1951–2008)	0.608	0.369	0.358	32.757	0.499	47 +/11 –	2.572	0.254

R is correlation coefficient; R^2 and R^2_{adj} are coefficient and adjusted coefficient of determination of regression analysis, respectively; *F* is the *F* statistic of the statistical significance of the regression analysis; *r* is the correlation coefficient between the record data and the leave-one-out-obtained estimates. ST is sign test which counts the number of agreement and disagreement between the estimated and instrumental series; Pmt is the product mean test; RE is the reduction of error; any positive value of RE denotes there is reliability in the reconstruction.

current growing season affect the number and wall thickness of latewood cells which determine the variations in MXD (Yasue et al., 2000; Hughes, 2001). Therefore, warmer temperature in the late summer can contribute in producing denser latewood. Our study confirms that there is a close relationship between the MXD of *L. speciosa* and late summer temperature, implying that the MXD of this species might have a great potential for temperature reconstruction and can even be used as an indicator of changes in climate.

4.2. Validation for our summer temperature reconstruction

4.2.1. Tree-ring inferred temperature variations in surrounding regions

Tree-ring studies regarding temperature reconstruction in surrounding regions can provide a reference for validating our reconstruction. The decadal variations (11-year low-pass filter) in August–September temperature reconstructed in the present study agrees with those for many other temperature time series inferred from tree-ring density in the southeastern, southern and eastern TP (Fig. 9). Cold periods during the 1810s, 1900s–1920s and 1960s presented in Changdu, eastern Tibet (Wang et al., 2010) and central Hengduan Mountains, southern China (Fan et al., 2009) were also indicated by the MXD of *L. speciosa* in our study area. Also, the conspicuously cold interval for the 1850s in our reconstruction was also shown in the MXD inferred August–September temperature time series in Linzhi (Bräuning and Mantwill, 2004) and Changdu (Wang et al., 2010), eastern TP. Warm phases in the 1820s, 1890s and 1930s–50s found in our study area were consistent with relatively high temperatures in Linzhi, eastern TP (Bräuning and Mantwill, 2004), the central Hengduan Mountains, southern China (Fan et al.,

2009) and Changdu, eastern TP (Wang et al., 2010). The warm periods in the 1930s–1950s were also exhibited in the reconstructed summer temperature series for the source region of Yangtze River (Liang et al., 2008). Although there are less similarities for these reconstructed temperature series before A.D. 1800, the cold period 1695–1705 and the warm period in the 1770s in our reconstruction were respectively consistent with the relatively low and high temperatures showed in the reconstructed August–September temperature series for Linzhi (Bräuning and Mantwill, 2004) and Changdu (Wang et al., 2010), eastern Tibet.

4.2.2. Glacier fluctuations

Advance and retreat of glacier is largely determined by changes in temperature, especially by summer temperature (Lowell, 2000; Liang et al., 2008). Therefore, we can use the historical glacier fluctuations in surrounding region to validate our reconstruction. The cold periods in the 1900s and 1960s corresponded to glacier advance in the Gongga Mountains of the eastern Tibetan Plateau (Li and Su, 1996) and in the Jade Dragon Snow Mountain of northwest Yunnan (He et al., 2003). Also, the cold episode during the 1960s was consistent with the accumulation period of Hailuoguo glaciers in the southeastern TP (Li et al., 2010; Duan et al., 2013). The warm phases during the 1940s–1950s, however, coincide with the rapid retreats of the glacier in the Gongga Mountains (Li and Su, 1996; Li et al., 2010; Duan et al., 2013) and the Meili Snow Mountain (Zheng et al., 1999). Conspicuous period with high temperature during the 1770s in our reconstruction was approximately concurrent with the beginning of retreat for Midui (Xu et al., 2012) and Xinluhai (Bräuning, 2006) glacier in the southeastern TP, occurred in 1767 and 1777, respectively. Thus, the reconstructed August–September temperature variability in this study may be used as an indicator of past glacier fluctuations in nearby areas to some extent.

4.2.3. Historical climatic records

Few climatic records within and around our study region are available before the 1950s, although historical records might be a useful information to validate our reconstruction. Documental records in Lhasa displayed a cold period in the early 1900s and a warm phase during most of the 1940s–50s (Liu and Chen, 2000) which were consistent with the corresponding cold and warm intervals in our reconstruction. The years 1816 and 1817 as two obviously cold years in our reconstructed temperature series were also recorded in some historical documents for Yunnan province (Yang et al., 2005; Liu et al., 2006). The extremely low temperature in these two years, with the August temperatures

Fig. 5. (a) Comparison of the actual (black line) and reconstructed (grey line) mean August–September temperature for their common period 1950–2008. (b) Comparison of the 1st differences of actual (black line) and reconstructed (grey line) mean August–September temperature during the period 1950–2008.

Fig. 6. Reconstructed mean August–September temperature (grey line) since A.D. 1690 from MXD with 11-year low-pass filter (thick black line).

Fig. 7. Multiple-taper method (MTM) spectral analysis of the reconstructed mean August–September temperature. The grey and black lines indicate 95% and 99% confidence limits, respectively.

being about 2.5–3 °C lower than those for other normal years, led to a serious famine that occurred in most areas of the province (Yang et al., 2005). In addition to Yunnan, the cold years in 1816 and 1817 were also widely reported in other areas of the world, especially for Europe and North America (Filion et al., 1986; Briffa et al., 1998), which were considered to be related to the known Tambora (Indonesia) eruption that occurred in 1815 (Sigurdsson and Carey, 1992).

4.3. Possible driving forces for the temperature variability in the Gaoligong Mountains, southeastern TP

The results of the MTM spectral analysis indicated the existence of several important cycles in our reconstruction, which may be used to explain the temperature variations in the study region. The 2–4 year cycles fall within the spectral bandwidth (2–8 years) of El Niño Southern Oscillation (ENSO) (Bradley et al., 1987). In accordance with this observation, several previous studies also indicated a possible linkage between ENSO and temperature variability of the Tibetan Plateau (Yin et al., 2000; Liang et al., 2008; Zhang et al., 2014). It has been proved that ENSO can modulate effectively temperature variability in the tropics and even across the globe (Trenberth et al., 2002; Gu and Adler, 2010), with warming up following an El Niño event (Newell

Fig. 9. Comparison of the reconstructed mean August–September temperature series in this study with other tree-ring density inferred temperature series. (a) Mean August–September temperature reconstruction for the Gaoligong Mountains, southeastern Tibetan Plateau from our study; (b) Mean August–September temperature reconstructed from maximum tree-ring density (MXD) in Changdu, eastern Tibet (Wang et al., 2010); (c) Mean August–September temperature reconstructed from MXD in Linzhi, eastern Tibet (Bräuning and Mantwill, 2004); (d) Mean April–September temperature reconstructed from MXD in the central Hengduan Mountains in southern China (Fan et al., 2009). The thick curves denote the 11-year low-pass filtered values and the horizontal lines indicate the long-term means.

and Weare, 1976). Warm ENSO event corresponded to the weakening of Indian monsoon (Kumar et al., 1999) which could decrease precipitation and hence reduce its cooling effect, resulting in above-normal temperature in our study area, a shoulder region of the Indian monsoon. The spectral peaks of 11.0 and 12.2 years were approximately equivalent to the cycle of solar activity which can affect the temperature variations on the earth as well as the TP. Tree-ring data from Sygera Mountain (Wang and Zhang, 2011) and Zuogong (Duan and Zhang, 2014), southeastern TP also displayed a cycle of about 11 years, indicating a possible effect of solar activity on tree growth through influencing temperature variations. The multi-decadal variability at spectral peaks of 27.5 and 59.8 years may correspond to the Pacific Decadal Oscillation (PDO) (Mantua et al., 1997). As shown in Fig. 10, the August–September temperature series reconstructed in this study agrees with the instrumental (Mantua et al., 1997) and reconstructed (MacDonald and Case, 2005) PDO variations during the period in the 1940s–1990s. Moreover, similar

Fig. 8. (a) Spatial correlations of August–September temperature of our reconstruction with that of Climate Research Unit (CRU TS 3.10) over the period 1950–2008. (b) Spatial correlations of the 1st differences of August–September temperature of our reconstruction with that of Climate Research Unit (CRU TS 3.10) during the 1950–2008 period.

Fig. 10. Graphical comparison of August–September temperature reconstructed in this study (b) with two PDO indexes: (a) Instrumental spring (March–April–May) PDO index during the period 1900–2000 (Mantua et al., 1997) and (c) reconstructed PDO index from MacDonald and Case (2005).

fluctuation pattern was displayed between the reconstructed temperature series and PDO index over the period in the 1770s–1810s. Therefore, PDO might be a driving force for the temperature variations in our study region. Tree-ring based temperature constructions in surrounding areas, such as the Changdu, southeastern TP (Wang et al., 2010), the source region of Yangtze River, eastern TP (Liang et al., 2008) and the Qilian Mountains, northeastern TP (Zhang et al., 2014) also suggested a possible relationship between the temperature variability and PDO. Besides the above factors, the volcanic explosion was probably an important driving force for temperature anomalies in our study area. Fine ash and gas developed in volcanic explosion can spread widely and consequently hinder the entry of sunshine to the Earth's surface, leading to the occurrence of low temperature (de Silva and Zielinski, 1998). The cold years 1809, 1810 and 1816 in our reconstruction corresponded separately to the volcanic eruption in 1808 (Chenoweth, 2001), 1809 (Dai et al., 1991) and 1815 (Sigurdsson and Carey, 1992), since a cold weather generally follows a volcanic event occurred in the previous year (Briffa et al., 1998).

5. Conclusions

- 1) The maximum tree-ring density (MXD) of *L. speciosa* growing near the upper tree line of the Gaoligong Mountains, southeastern TP is sensitive to temperature variations in late summer. The developed MXD standard chronology of this species can well explain the actual variance in the August–September temperature (40.9%) as well as its first differences (36.9%) during the calibration period 1950–2008. This study demonstrates the potential of using the tree-ring density of larch (*L. speciosa*) to infer temperature variability in the southeastern TP.
- 2) The reconstructed August–September temperature history for our study area based on the MXD index series, which highlights the inter-annual to multi-decadal scale variability, can capture broad-scale regional temperature variability in the southeastern and southern TP. Comparison of the present reconstruction with other documents in nearby regions indicates consistent cold/warm episodes at the decadal scale, i.e. cold periods during the 1810s, 1900s–1920s, 1960s and warm phases during the 1820s, 1890s and 1930s–50s. Our reconstruction provides a new temperature record for the southeastern TP, which, to some degree, improves the spatial coverage of available proxy records in this region.

- 3) The temperature variability in the Gaoligong Mountains, southeastern TP might have been related to ENSO, solar activity, PDO and volcanic explosion.

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Supplementary data

Supplementary data associated with this article can be found in the online version, at <http://dx.doi.org/10.1016/j.palaeo.2015.01.003>. These data include Google maps of the most important areas described in this article.

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