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Abstract

Knowledge of Asian monsoon variability remains limited because of sparse instrumental data available only for short time series. Here, an updated tree-ring width record covering the period AD 1037–2009 was developed for the south-central Tibetan Plateau (TP). Correlation analysis revealed a significant relationship ($r = 0.71$) between the tree-ring index and annual (previous July to current June) precipitation series for the instrumental period 1963–2008, which accounts for 50.41% of the rainfall variability. Based on a linear regression model, the longest available regional precipitation history was reconstructed. Spatial correlation between tree ring width and annual precipitation data from previous July to current June indicates that the reconstruction is representative of precipitation changes on the south-central TP. Regional wet conditions occurred during AD 1095–1161, 1376–1403, 1414–1446, 1518–1537, 1549–1572, 1702–1757, 1848–1878 and 1891–1913, while dry periods were identified during AD 1189–1242, 1256–1314, 1329–1357, 1470–1491, 1573–1623, 1636–1686, 1761–1821, 1823–1847, 1879–1890 and 1931–1985. The negative correlation between our reconstructed precipitation and India monsoon rainfall series indicates the seesaw pattern over northern and southern monsoon Asia. It is suggested that solar radiation-induced sea surface temperature (SST) anomalies over the tropical Pacific influence regional rainfall patterns. The degree of this influence has been stable at the multidecadal scale during the past 1000 years.

Keywords

Asian summer monsoon, past millennium, precipitation variation, solar activity, south-central Tibetan Plateau, tree-ring records

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Introduction

Possible shifts in moisture patterns in monsoon Asia are a serious concern because of their significant effects on regional water resource management (Shi et al., 2007), since more than half of the world's population depends on directly or indirectly on Asia monsoon rainfall (Buckley et al., 2007). Changes in monsoon strength, timing and distribution may result in droughts or floods, and crop failure or famine, and can thus inflict economical losses on local populations. For example, between November 2008 and February 2009, central China was hit by the area's worst drought within half a century, which affected 1.0×10^{11} m² of crops and adversely impacted residents in 12 provinces of China (Tao et al., 2009). According to Lei and Duan (2011), the durations of prolonged dry events during the dry season have significantly increased over southern China and the Yellow River valley during the past 51 years. The significant drought in Indochina in the 1980s (Sano et al., 2009) affected local environment and ecosystems seriously. Therefore, understanding the causes of drought/flooding and forecasting future monsoon variability is of great socio-economic benefit. However, such knowledge is limited at present by the short instrumental records available in the Asian monsoon region; for example, records available for the Tibetan Plateau (TP) seldom extend back beyond the 1950s, providing only a limited spatiotemporal perspective on natural climate variability. The existing instrumental climate records are also too short to investigate multidecadal or longer timescale climate change. This situation can be mitigated by adopting a paleoclimatic approach, employing annually resolved, well replicated and exactly dated proxy records, such as tree rings (Fritts, 1976).

During the past few years, many researchers have investigated climate change in the Asian monsoon region based on tree-ring networks (Bräuning and Gießinger, 2006; Esper, 2000; Fan et al., 2008; Li et al., 2006; Sheppard et al., 2004; Tian et al., 2007; Yang et al., 2009). The resulting longer-term paleoclimate records may allow evaluating the climate variability at a decadal to multidecadal scale. Recently, Cook et al. (2010) used tree rings from 327 tree-ring sites across the Asian monsoon region and reconstructed the seasonal Palmer Drought Severity Index (PDSI) of the summer (June–July–August) season for the past 700 years, thereby providing a new perspective on Asian drought variability. Such long-term records are urgently needed for climate modeling (Barlow et al., 2002; Hoerling and Kumar, 2003; Mariotti, 2007), and for better understanding of large-scale atmosphere–land–sea interactions (Gergis and Fowler, 2009; Robock et al., 2003; Trouet et al., 2009; Wang et al., 2001). In addition, the responses to the natural driving forces of these modes also need further investigation over long timescales for the Asian monsoon region, especially over the TP.

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The TP acts as an extensive elevated (mean elevation >4000 m above sea level) heat source in summer and greatly influences the Asian monsoon circulation through its thermal and dynamic effects (Liu and Chen, 2000; Rangwala et al., 2009; Webster et al., 1998; Yanai et al., 1992). Many researchers have recently reconstructed the climate history for the area (Fang et al., 2010; Gou et al., 2007; Li et al., 2011, 2012; Liang et al., 2009; Yang et al., 2010a, 2010b, 2010c; Zhang et al., 2003, 2011; Zhu et al., 2008); however, very few studies have focused on the southern TP (Bräuning, 2001; Bräuning and Mantwill, 2004; Fan et al., 2008; Griesinger et al., 2011) because of its remote location. Therefore, an improved tree-ring data set with a dense spatial coverage and long time span is needed for the southern and south-central TP region. Bräuning and Griesinger (2006) reconstructed the summer monsoon intensity of the late Holocene in the Tibetan-Himalayan region from tree rings, but the reconstruction is not well validated with instrumental data and the chronology is only credible from AD 1200 to 1998 because of the low replication in earlier years. In this study, we updated a 937 year long tree-ring chronology for this area, and investigated both the low- and high-frequency climate variability over the last millennium.

Our objectives in this study were: (1) To identify the relationships between tree-ring growth and climate parameters in the study area. (2) To reconstruct regional precipitation history during the past 1000 years for the south-central TP. (3) To analyze regional rainfall patterns and their associated physical mechanisms. This presented reconstruction is a new contribution to the scarce existing knowledge about long-term moisture variability on the south-central TP.

Materials and methods

The study area is located in Linzhou County (91.5°E, 30.3°N), north of Lhasa, on the south-central TP. Figure 1 shows the locations of the sample site and nearby available meteorological stations (Lhasa, Jiali, Naqu, Suoxian, Dangxiong, Bangge and Zedang). Although some individual climate records started in 1954, the common period of all seven instrumental records in the wider area only covers the period 1963–2008. The two other nearby meteorological stations (Nimu and Mozugongka) only have records commencing in 1974 and 1991, respectively, and were excluded from our analysis because of statistical considerations. Principal component analysis (PCA) was used to assess the common signal between the seven precipitation records over the 1963–2008 periods. The PC#1 accounted for 65.48% of the

total variance, which is relatively high considering the different elevations and the microenvironment of the station's locations. The detailed PC#1 loadings (varying between 0.66 and 0.89) of the seven stations indicate that the year-to-year variability is spatially homogeneous within the region. Furthermore, all stations have the same climatic features characterized by a main wet summer season when temperatures are highest (Figure 2). Our tree-ring width series showed higher correlations with the annual precipitation records with all stations than precipitation during individual months. The same results are obtained in the case of leaving one or two observations from the seven stations. Since temperature is a continuous variable, monthly mean temperature will tend to be normally distributed, thus, we calculated the regional monthly mean temperature series from the corresponding seven instrumental stations with a weighted average (Jones and Hulme, 1996). The lower bound of zero means that most precipitation time series are not normally distributed. Furthermore, in regions of high precipitation gradients there will also be high gradients in precipitation means and variances, we calculated the regional monthly precipitation series with the weighted standardized anomalies for each of the seven stations; the monthly standardized anomalies were then converted to 'absolute' precipitation values using the average of the means and standard deviations over the region of interest for the common reference period that has little missing data. Hence, for the common period 1963–2008, regional mean annual temperature is 2.52°C, with a minimum mean monthly temperature of −7.86°C in January and a maximum mean monthly temperature of 11.47°C in July. Mean annual precipitation is 515.72 mm and the wettest month is July (121.06 mm) (Figure 2). Our studied forest stands on a south-facing slope characterized by well-drained soils, indicating that there is no influence of groundwater on tree growth.

In the fall of 2009, we extracted 2–3 cores at breast height from each juniper (*Juniperus tibetica* Kom.) tree using increment borers. The forest was located in a semi-arid region with an elevation range of 4200–4575 m. In the laboratory, all cores were mounted on sample holders with their transverse surfaces exposed. After being air dried, the wood surface was prepared with razor blades and then enhanced with chalk to ensure that the narrower rings could be recognized. The ring-width of each core was measured with a LINTAB 6 measuring system at a resolution of 0.01 mm. Cross-dating between the individual growth curves was accomplished by visual inspection, and the statistical *t*-test and sign test were calculated by TSAP software (Rinn, 2003). Cores that were difficult to be cross-dated or cores containing many false or missing rings were excluded from the master series. In total, 161 cores from 63 trees were selected for further analysis. The quality of the cross-dating was checked with the COFECHA program (Holmes, 1983).

Before calibrating the tree-ring chronology against climate data, the biological age trend inherent in the raw data series had to be removed. To remove bias caused by so-called heteroscedasticity, a data-adaptive power transformation was first applied to the raw data (Cook and Peters, 1997). Next, we employed two standardizing procedures to remove the biological growth trend inherent in the series: first, a negative exponential or linear regression curve of any *k* was applied to each series. If the method turned out to be inappropriate for some growth curves, the tree-ring sequences were detrended using a cubic spline with a 50% frequency-response cutoff equal to 67% of the series length. The final tree-ring chronology was calculated as the residuals between the raw measurements and fitted trend curves, resulting in a dimensionless index series, which was averaged to a standard site chronology by a bi-weight robust estimate of the mean to minimize the influence of outliers (Cook and Kairiukstis, 1990). To reduce the potential influence of the variable sample size, we stabilized the variance of the chronology using the method described by Osborn et al.

Figure 1. Locations of the tree-ring sample site and meteorological stations.

Figure 2. Mean monthly precipitation and temperature from the meteorological stations Lhasa, Jiali, Naqu, Suoxian, Dangxiong, Bange, Zedang, and the regional mean for the common period AD 1963 to 2008.

(1997). All calculations were carried out with the ARSTAN program (Cook and Kairiukstis, 1990). To eliminate ambiguous parts of the chronology due to decreasing sample replication in the early part of the chronology, the threshold of the Expressed Population Signal (EPS) of 0.85 was applied (Wigley et al., 1984) by using a 30-year moving window with 15-year overlaps.

Correlation coefficients and response functions (Fritts, 1976) of the tree-ring width chronology with regional temperature and precipitation were calculated for the common period 1963–2008 using the software Dendro-Clim 2002 (Biondi and Waikul, 2004). In addition, the relationship between ring width series and local PDSI (31.25°N, 91.25°E) grid data developed by Dai et al. (2004) was investigated. To test the validity of our climate-growth model, we used split sample calibration-verification tests (Meko and Graybill, 1995). The sign test and product mean test were also conducted. Since the common period of the climate data is short (46 years), we also performed a leave-one-out verification (Michaelsen, 1987) for the derived climate-growth relationship (Table 1).

We examined the spatial representativity of our ring width chronology by correlating it to monthly precipitation data extracted from the Climatic Research Unit (CRU) Time Series

(TS) 3.1 0.5°×0.5° gridded data set to check that our sampled trees was spatially representative. Since instrumental observations did not begin until 1954, we only used the years from 1954 onwards in the CRU TS3.1 data set. We also compared the spatial correlations of the calculated regional precipitation data from the seven instrumental stations during the common period 1963–2008. All correlation analyses were carried out using the KNMI climate explorer (Royal Netherlands Meteorological Institute; <http://climexp.knmi.nl>).

Results

Characteristics of the tree-ring width data

The master series spans 973 years from AD 1037 to 2009. The expressed population signal (EPS = 0.94), signal-to-noise ratio (SNR = 19.288), mean sensitivity (MS = 0.284) and all series Rbar (0.425) indicate that the chronology is sensitive to climate variability and that the sampled trees respond to a common climate forcing. Based on the EPS cutoff value of 0.85, the final reliable period of the standard chronology is AD 1085–2009, including the period AD 1115–1130 with a slightly lower EPS value of 0.76. Specifically, the mean segment length (MSL) of our tree-ring chronology is 377 years, indicating that decadal- and multidecadal-scale variability may probably be preserved (Cook et al., 1995).

Response of tree growth to climate parameters

The results of the correlation and response function analysis are shown in Figure 3. Significant positive correlations (at the 0.05 level) with temperature (Figure 3a) occur in the prior August (0.322), November (0.390), December (0.432) and current January (0.377), February (0.309), August (0.480) and September (0.402). The influence of temperature during the winter season is a common feature in high-altitude tree-ring growth sites (Bräuning and Griesinger, 2006; Liu et al., 2010a); this is because warmer conditions during the cold season allow higher levels of hydro-carbonate to be stored in the trees, thus favoring a wide ring in the following growth season. In addition, warm conditions before the growing season could stimulate root system activity and be beneficial to carbon absorption and transportation. Significant negative correlation with temperature was found for the current May (−0.461). High temperatures during May might strengthen the tree's transpiration, leading to drought stress and thus decrease tree-ring width. The response function also yields significant value with temperature in current May (−0.412).

Table 1. Regression model statistics for different calibration and verification periods.

Calibration	Verification	R_{cal}	R^2_{cal}	R^2_{adj}	R_{ver}	R^2_{ver}	RE	CE
1963–1994	1995–2008	0.59	0.35	0.32	0.79	0.62	0.59	0.54
1978–2008	1963–1977	0.72	0.52	0.51	0.60	0.36	0.58	0.48
Full period	R	R^2	R^2_{adj}	ST1	Pmt 1	ST2	Pmt 2	
1963–2008	0.71	0.50	0.49	33/12**	5.00**	27/17	4.11**	0.48L

R_{cal} : correlation coefficient for the calibration period; R^2_{cal} : explained variance of the regression model for the calibration period; R^2_{adj} : adjusted explained variance of the regression model; R_{ver} : correlation coefficient for the verification period; R^2_{ver} : explained variance of the regression model for the verification period; RE: reduction of error statistic for the calibration period; CE: coefficient of efficiency for the verification period; R, R^2 and R^2_{adj} are the correlation, explained variance and adjusted explained variance of regression model for the full period AD 1963–2008. ST1: general sign test between observation and reconstruction; ST2: the sign test between observation and reconstruction departures from mean of their first order differences; Pmt 1: product mean test between observation and reconstruction; Pmt 2: product mean test between observation and reconstruction departures from mean of the first order differences; ** significant at the 0.01 level. ^LThe value is calculated by the leave-one-out method for the full period.

Figure 3. Correlations (solid bars) and response functions (straight lines with triangles) of Linzhou tree-ring series with regional (a) monthly temperature, (b) precipitation and (c) Palmer Drought Severity Index (PDSI). Correlations with temperature and precipitation were calculated for the common period 1963–2008, while that for the PDSI is 1950–2005. The horizontal dashed lines indicate the 95% confidence level for the correlation function, respectively. Response functions significant at 0.05 levels are marked with an asterisk.

Correlations between the tree-ring index and regional temperature are weak compared with those of precipitation (Figure 3b). Significant positive correlations with precipitation were found for the prior July (0.418) and August (0.321), and the current April (0.345), May (0.736) and June (0.437). Precipitation during spring and early summer months provides available moisture and favors formation of the early-wood cells, which account for the majority of the total juniper ring width, as confirmed by studies of the cell anatomy of the juniper species (Vaganov et al., 2005). The significant correlations with precipitation in the non-growing season suggest that previous year's moisture input favors tree growth in the subsequent growing season.

Based on the above analysis, we considered that both temperature and precipitation affect tree growth in the study area. Therefore, we calculated the correlations between tree-ring series and the extracted PDSI over the period 1950–2005, since PDSI integrates the combined effects of temperature and precipitation (Dai et al., 2004; Palmer, 1965). All correlations (Figure 3c) are significant at the 0.05 level, with maximum values occurring in the period from current April (0.560), May (0.685) and June (0.689). Response function coefficients are slightly lower, with significant values occurring in the previous July (0.209) and current May (0.250) and June (0.310). All positive correlations between PDSI and the tree-ring series indicate that the soil moisture availability significantly affects tree growth on the south-central TP.

The climate conditions in the individual months are less important compared with those of the seasonal or annual averages, so correlations with combined months were also computed. The correlation between ring width series and current March to July PDSI is 0.709, while that for the previous July to current June is 0.634, indicating that the tree-ring width chronology can also be used as an indicator of PDSI variability. However, considering that the regional annual precipitation has a higher correlation ($r = 0.71$) with the tree-ring width series, we reconstructed the precipitation history based on the time window from previous July to current June for the past 1000 years.

Regional precipitation reconstructions

The spatial correlations (Figure 4) of the tree-ring width chronology (Figure 4a) and the calculated precipitation data (Figure 4b) with the annual (previous July to current June) gridded precipitation for the past five decades indicate that our research site can be used as an indicator of the precipitation variability on the south-central TP. Furthermore, we found that precipitation variability at our study region had a positive correlation with precipitation in the Aral area and a negative correlation with that in the Far East territory. Such teleconnections indicate the interactions between the Asia monsoon system and the Westerlies over Eurasia region, emphasizing the importance of our study region for the large-scale atmosphere circulation studies in the future.

The following linear regression model was developed for the precipitation reconstruction:

$$P = 168.4 \text{STD} + 342.7 \quad (1)$$

where STD represents the index of the standard tree-ring width chronology and P is the annual precipitation from previous July to current June.

The reconstructed precipitation series derived from this model shows similar patterns to the observed instrumental data (Figure 5a). The correlation coefficients between the actual and reconstructed precipitation in the calibration period 1963–1994 (1978–2008) and verification intervals 1995–2008 (1963–1977) are 0.587 (0.723) and 0.785 (0.600), respectively. The reduction error (RE) and the coefficient of efficiency (CE) based on the respective tests ($p < 0.001$) are positive and high (Table 1), providing further evidence that the model has significant skill (Fritts, 1976). For the full calibration period (1963–2008), the correlation between tree-ring width series and instrumental rainfall records is 0.71, which accounts for 50.41% of the precipitation variability from prior July to current June. Moreover, the product mean test and leave-one-out results are all significant. Based on the above statistical characteristics, we are confident that the regression model can be used to examine the rainfall variability for the south-central TP over the past millennium.

Figure 4. Spatial correlations between the gridded (40°–130°E, 0°–60°N) annual precipitation from previous July to current June (PJ–CJ) data of the Climatic Research Unit (CRU) Time Series (TS) 3.1 networks and (a) the Linzhou tree-ring width series from 1954 to 2009, and (b) regional annual (PJ–CJ) precipitation averaged from the seven instrumental records (Lhasa, Jiali, Naqu, Suoxian, Dangxiong, Bange, Zedang) for the common period 1963–2008.

Figure 5. (a) Comparison between instrumental (straight black line) and reconstructed (dashed blue line) precipitation from 1963 to 2008; (b) reconstructed precipitation for the period AD 1085–2009 (blue line; see colour figure online) and the 11-year running mean series (black thick line); (c) expressed population signal (EPS) and (d) sample depth. The horizontal dashed line is the mean precipitation over the full reconstructed period and the horizontal dotted gray line denotes the 0.85 EPS criterion for signal strength acceptance (Wigley et al., 1984).

By applying the above model, the annual precipitation (previous July–current June) was recovered for the period AD 1085–2009 (Figure 5b). The reconstructed precipitation ranges from 400.63 mm to 653.06 mm and the mean annual precipitation was 509.18 mm, with a standard deviation ( ) of 30.04 mm. Wet (dry) periods were defined as having an 11-year running mean that was continuously higher (lower) than the long-term average from AD 1085 to 2009 for more than 10 years. According to this definition, regional wet conditions occurred during AD 1095–1161, 1376–1403, 1414–1446, 1518–1537, 1549–1572, 1702–1757, 1848–1878 and 1891–1913; dry periods were identified during AD 1189–1242, 1256–1314, 1329–1357, 1470–1491, 1573–1623, 1636–1686, 1761–1821, 1823–1847, 1879–1890 and 1931–1985. During the past 1000 years, the most extended wet period (66

years) occurred in the first half of the 12th century, while the most prolonged dry phase started in the second half of the 16th century and lasted until the end of the 18th century. Furthermore, based on the increasing precipitation trend from the second half of the 20th century to the present day, the reconstruction indicates unprecedented pluvial conditions for the past decade.

Discussion

Comparison of the reconstructed precipitation with other associated results

In order to identify whether our tree-ring width based precipitation reconstruction represents large-scale spatiotemporal climate features, we compared our results with other reconstructions from

Figure 6. Tree-ring width based annual precipitation anomalies (b) compared to summer precipitation anomalies inferred from tree-ring $\delta^{18}\text{O}$ series (a) by Griesinger et al. (2011) and seasonal PDSI (c) recovered by Cook et al. (2010) during the past 1000 years. The thick black lines denote the smoothed (with a 5-year window) series and the horizontal black lines indicate the mean values over the full period for each series.

nearby regions. Griesinger et al. (2011) established a ^{18}O tree-ring series from Linzhou (Reting; $30^{\circ}18.5'\text{N}$, $91^{\circ}31'\text{E}$) to reflect the Asian summer monsoon variability during the past 800 years. Figure 6a shows the detailed information of the tree-ring ^{18}O series. We found that the correlation with our annual precipitation series was low ($r = 0.08$) over the common period AD 1163–2005. On the centennial timescale, however, common periods of reduced precipitation are found during much of the 14th century and the 20th century. The 16–19th centuries were generally wet in both series, with the exception of the second half of the 17th century and around AD 1800 when some dry events are identified in our precipitation reconstruction. Compared with precipitation reconstructions for south and south-central Tibet (Liu et al., 2010a, 2010b), we found common dry years during the 1570s–1620s, 1800s–1850s and 1960s–1980s, while wet periods occurred in the 1510s–1550s, 1700s–1750s, 1850s–1870s, 1890s–1910s and after 1995. According to the reconstructed seasonal (May–June) PDSI for north Tibet (Wang et al., 2008), dry conditions occurred during the 1600s–1690s and wet years during the 1700s–1750s, again illustrating synchronous variability with our precipitation reconstruction. However, contrary to the wet conditions during AD 1931–1943 revealed by Wang et al. (2008), we found dry conditions on the south-central TP during that time. The discrepancies between the start and end years of the above- or below-average precipitation in the different archives may be partly attributed to the different seasonality of the climate signal represented by the proxies or different regional climatic conditions.

Specifically, based on the tree-ring width series, the increasing trend in annual precipitation during the past 50 years was also found on the southeastern (Bräuning and Mantwill, 2004) and northeastern TP (Qin et al., 2010, 2011; Yang et al., 2010b), and the unprecedented pluvial condition since the mid-1980s is consistent with the results from Liu et al. (2010a). This rapid growth increase can not be attributed to the age structure of the studied trees used in developing the chronology since a homogenous age distribution of young/old trees is indicated in Figure 5d. This growth trend is also not a statistical artifact of the detrending procedure because we followed a detrending procedure (applying an

adaptive power transformation to the raw data and calculating residuals between the raw measurements and fitted trend curves to derive as the final dimensionless index series) suggested to be free of the ‘end effect’ problem (Cook and Peters, 1997; Melvin and Briffa, 2008). Rather, as shown in Table 1, the results of verification statistics are more robust for the calibration period 1963–1994 than for the period 1978–2008, which may indicate that the dramatic increase in rainfall during the last decade is a real climatic effect. Figure 5a illustrates the increasing trend in the instrumental climate data for the past decade. Furthermore, 2007, which is one of the three significant drought years with 389 mm of annual precipitation also showed relatively higher rainfall compared to the other two drought events occurring before (1972 with 359 mm, and 1995 with 333 mm annual precipitation, respectively). Since the available regional instrumental record are rather short, there is not an unambiguous trend for the past five decades. In conclusion, we consider that the significant increase of tree growth during the past decade is probably attributed to the current composited climate change, while the other factors mentioned above may just strengthen this trend. However, we still need some work to verify the results in the future.

We also compared the recovered seasonal (June–July–August) PDSI-based Asian monsoon variability (Cook et al., 2010) with the annual rainfall series during the past 700 years. The significant correlations between the two series ($r = 0.371$, $n = 706$, $p = 0.01$; $r = 0.422$, after applying 5-year mean) indicate that our tree-ring width chronology is able to represent regional precipitation variability. As inferred from Figure 6 (b, c), common wet periods in both series occurred during AD 1320–1332, 1364–1395, 1524–1538, 1551–1576, 1713–1730, 1856–1877 and 1899–1914; while dry years prevailed during AD 1451–1473, 1597–1624, 1674–1698, 1792–1830 and 1957–1976. During the past 700 years, significantly different periods occurred in the 15th century and in the first half of the 20th century. In the remaining periods, both series showed common increasing or decreasing trends. Differences between the two series may be due to the fact that the tree-ring index used for the reconstruction of the PDSI is derived from sparsely distributed data over the southern TP (Bräuning, 2001;

Bräuning and Mantwill, 2004; Fang et al., 2010) and therefore lacks details of rainfall variability. The different seasonal representations of the two series can also lead to the biased trends. Third, the 'point-by-point' regression model used for finding the best correlation (Cook et al., 2010) may also decrease the sensitivity of the tree-ring index to the moisture variability. Hence, we recommend that more high-resolution proxy records are needed to better understand regional climate variability over the last millennium.

Concerning the southern monsoon Asia region, we found a negative correlation ($r = -0.312$, after applying 5-year running mean, $n = 134$) between the recovered precipitation series and all India monsoon rainfall for the common period AD 1871–2009, while correlations reach -0.418 with the averaged monsoon rainfall only from west, northwest and core India. Previous studies have reported the so called 'seesaw rainfall pattern' over monsoon Asia, where precipitation in the monsoon shadow zone of the western Himalaya was negatively correlated to summer monsoon rainfall over central northeast India (Yadav, 2011a). Kaspari et al. (2007) found that the reduced northward incursion of the south Asian monsoon inferred from East Rongbuk glacier, located on the northeast ridge of the Mt Everest, was contrary to the strengthened monsoon south of the Himalayas. Such 'seesaw patterns' have been verified by observations (Chandrasekar and Kitoh, 1998; Goes et al., 2005) over the past decades.

Possible physical mechanism of precipitation variability over the south-central TP

Given the successful verification of our long-term reconstruction, we applied the multitaper spectral analysis method (MTM) (Mann and Lees, 1996) to examine the characteristics of local rainfall variability in the frequency domain. The analysis over the full reconstruction period (AD 1085–2009) revealed both low- and high-frequency variability. As shown in Figure 7, the significant cycles are restricted to 104, 125, 156, 208, 312 years and also to periods of less than 10 years, with particular concentrations around the interannual cycles (2–3 years). The cycles of 104–312 year periods may be linked to longer-term changes in the large-scale modes of solar activity (Agnihotri et al., 2002; Wang et al., 2008). Similar periodicities were detected in climatic reconstructions from other regions (Bhutiyan et al., 2010; Hong et al., 2001; Qin et al., 2010). The interannual cycles fall within the range of the El Niño–Southern Oscillation (ENSO) variability (Allan et al.,

1996, 2003; Mann et al., 2005) and are in good agreement with studies from elsewhere in the Himalaya–TP region (Fan et al., 2008; Li et al., 2011, 2012; Wang et al., 2008; Yadav, 2011b), suggesting teleconnections between ENSO variability and the wet/dry episodes in the studied area and nearby regions.

To provide deeper insights into the dynamics of monsoon variability, Cook et al. (2010) used Distinct Empirical Orthogonal Functions (DEOFs) to analyze the spatiotemporal precipitation variability over monsoon Asia. The leading patterns of DEOF1 are characterized by dominant same-sign loadings over India and Southeast Asia (hereafter as southern monsoon Asia), with opposite-sign loadings over the TP, northern Pakistan, the Pamir and Tien-shan Mountains (hereafter as northern monsoon Asia). Furthermore, sea surface temperature (SST) anomalies over the eastern tropical Pacific are positively correlated with the northern monsoon Asia series and negatively correlated with the southern monsoon Asia series for the March–August season during the available common data coverage period AD 1856–2004. DEOF4 (that explains 6.48% of the total variance) showed significant positive loadings over the eastern TP, and the time series positively correlated with the eastern tropical equatorial SST anomalies of the previous winter (December–February) season. That is to say, wet periods over the northern monsoon region are associated with warm SSTs and a dry event is related to cold SSTs over the eastern tropical Pacific. This association is reversed to the southern monsoon Asia, which was also indicated by other studies (Hoerling and Kumar, 2003; Lau and Wu, 2000; Sano et al., 2011; van Loon et al., 2007). Considering that the summer Asian monsoon is driven by components of the climate system from both the northerly higher elevation and the southerly lower elevation regions (Kaspari et al., 2007), we assume that local topography modifies the monsoon system to some extent (Huang and Zhang, 2007; Liu and Chen, 2000; Wu and Zhang, 1998; Wu et al., 2007). However, we need other models integrating topographic parameters to further verify these results (e.g. Thomas, 2008).

Based on the above analysis, we infer that SST anomalies over the eastern tropical Pacific have influenced local precipitation variability during the period of instrumental records. To investigate whether the influence is stable at multidecadal timescales over the past 1000 years, we compared our reconstructions with the SST anomalies over tropical Pacific inferred from Mann et al. (2005). This analysis was based on applying the Zeblak–Cane model to analyze the ENSO response to solar radiative forcing, volcanic forcing and the composite-natural forcing (solar + volcanic). We smoothed the SST anomalies (forced by solar radiation) and the recovered precipitation series with a 41-year window to highlight their multidecadal variability. The higher positive correlation ($r = 0.471$) between the two series (Figure 8) indicates that dry periods at our study site are associated with cold SST conditions in the eastern tropical Pacific, and vice versa. Therefore, we are confident that local precipitation is influenced by SST anomalies at the multidecadal scale during the past millennium. In Figure 8 we also show the indirect correlations between solar radiation (rescaled by Mann et al., 2005) and local precipitation variability during the past 1000 years because the associated SST anomalies were induced by solar forcing. How solar activity directly influences local precipitation is still under investigation (Kaspari et al., 2007; Tiwari et al., 2005). Herein, it is suggested that solar radiation-induced SST anomalies over the tropical Pacific interact with different components of the large-scale atmosphere–land–ocean system and thus influence local climate through different pathways. The significant correlations between modeled SST anomalies and the reconstructed precipitation also suggest that model simulations are likely to direct future predictions of temperature and precipitation patterns (Gergis and Fowler, 2009) under the complicated climatic change in the future.

Figure 7. Multitaper method (MTM) spectral density of the Linzhou tree-ring width based precipitation for the south-central Tibetan Plateau (TP). The red dashed, blue dotted, and green dash-dotted lines indicate the 99%, 95% and 90% confidence levels, respectively (colour figures available online).

Figure 8. The recovered precipitation (blue line) variability compared with solar radiative forcing (orange line) and sea surface temperature (SST) anomalies over the eastern tropical Pacific (red line) taken from Mann et al. (2005). The three thick lines denote the smoothed (with a 41-year window) series and the horizontal dashed lines indicate the mean values for each series (colour figures available online).

Conclusions

We have presented a 925-year juniper (*Juniperus tibetica* Kom.) tree-ring width chronology that yields high correlation ($r = 0.71$) with the regional annual precipitation from previous July to current June on the south-central Tibetan Plateau. Based on statistical tests, we recovered a millennial-scale precipitation history, which is consistent with other climatic records around the region. The reconstructed precipitation series revealed both low- and high-frequency variability, which falls within solar activity and ENSO periodicities, respectively. The significant negative correlation between our recovered precipitation variability and Indian monsoon rainfall series is consistent with previous results that have revealed the seesaw rainfall patterns over northern and southern monsoon Asia. It is suggested that SST anomalies over the tropical Pacific affected the rainfall patterns over monsoon Asia during the instrumental period. Furthermore, the influence of solar radiation-induced SST anomalies over tropical Pacific on local rainfall variability was stable on the multidecadal scale during the past 1000 years. Anyhow, regional circulation models are needed to verify the physical mechanisms underlying the complicated regional rainfall patterns.

Considering that our reconstruction is based on one sample site and that precipitation is a spatially complex climate variable, more work is needed in the near future to develop large-scale and longer-term climatic reconstructions over this and surrounding areas to yield a clearer picture of regional climate variability. More effort should also be made to develop a tree-ring network including elevation gradients to clarify the influence of topography on the spatiotemporal variability of local climate. This would be the foundation for sound regional climate modeling and for more reliable forecasts of trends in future monsoon rainfall.

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