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Short communication

Growth response of alpine treeline forests to a warmer and drier climate on the southeastern Tibetan Plateau

Chunming Shi[a,*], Miaogen Shen[b], Xiuchen Wu[c,*], Xiao Cheng[a], Xiaoyan Li[c], Tianyi Fan[a], Zongshan Li[d], Yuandong Zhang[e], Zexin Fan[f], Fangzhong Shi[c], Guocan Wu[a]

[a] Joint Center for Global Change Studies (JCGCS), College of Global Change and Earth System Science, Beijing Normal University, Beijing 100875, China
[b] Institute of Tibetan Plateau Research, Chinese Academy of Sciences, Beijing 100101, China
[c] State Key Laboratory of Earth Surface Processes and Resource Ecology, Faculty of Geographical Science, Beijing Normal University, Beijing 100875, China
[d] State Key Laboratory of Urban and Regional Ecology, Research Center for Eco-Environmental Sciences, Chinese Academy of Sciences, Beijing 100085, China
[e] Key Laboratory of Forest Ecology and Environment, State Forestry Administration, Institute of Forest Ecology, Environment and Protection, Chinese Academy of Forestry, Beijing 100091, China
[f] Key Laboratory of Tropical Forest Ecology, Xishuangbanna Tropical Botanical Garden, Chinese Academy of Sciences, Menglun, Mengla, Yunnan 666303, China

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ABSTRACT

Forest growth at high altitudes and latitudes is sensitive to climate warming. However, warming-induced drought stress has decreased forest growth and survival rates, and constitutes a key uncertainty in projections of forest ecosystem dynamics. A fast warming rate has occurred over the Tibetan Plateau (TP), and the response pattern of alpine forest growth on the TP to a warmer and possibly drier climate is still unknown. By compiling tree-ring width records from ten alpine treeline ecotones (ATEs), we developed an index of regional tree growth in ATEs (RTGA) on the southeastern TP, which is a major forested region of the TP. Our results showed a stable and clear coherence between RTGA and the regional summer (June-August) minimum temperature during the studied period (1950–2012, $R^2 = 0.59$, $P < 0.001$), despite a prominent drying trend since the 1990s. We conclude that warming-induced drought stress has not limited ATE forest growth on the moist southeastern TP.

1. Introduction

Warming rates that are faster than the global mean have been observed at high elevations and latitudes (Brohan et al., 2006; Pepin et al., 2015). Forest growth at the upper altitudinal and high latitudinal limits is generally limited by low temperatures; therefore, the growth of these forests has been considered to be particularly sensitive to climate warming (Korner and Paulsen, 2004; Rossi et al., 2007). However, growing evidence of decreased tree growth rates and increased drought sensitivity in the northern high latitudes has been reported (Briffa et al., 1998; Driscoll et al., 2005; Hellmann et al., 2016; Porter and Pisaric, 2011; Wilmking et al., 2004). This phenomenon, known as the divergence problem (DP), has been mainly attributed to drought stress aggravated by warming and has caused a high degree of uncertainty in tree-ring-based climate reconstructions and forest growth projections (D'Arrigo et al., 2008; Hellmann et al., 2016). The DP has mainly been observed in northern high latitudes; however, it has seldom been tested in regions with high elevations, such as the Tibetan Plateau (TP).

In recent decades, the rate of warming on the TP has been faster than that recorded across the Northern Hemisphere and that in other regions at the same latitude (Liu and Chen, 2000; Kang et al., 2010). In association with a fast rate of warming, precipitation has been reported to limit forest growth at the alpine treeline ecotone (ATE) in the dry regions of the TP, such as in the northeast (Liang et al., 2016a; Liu et al., 2006; Yang et al., 2013) and on the northern slope of the Himalayas (Liang et al., 2014), since the 1950s. Drought stress induced by fast warming is responsible for ATE forest growth limitation in both regions (Liang et al., 2016a; Schwab et al., 2018).

Nevertheless, forest growth has showed substantial spatial incoherence in different regions of the TP due to diverse hydrothermal conditions (Brauning and Mantwill, 2004). A comprehensive understanding of the climate response patterns of ATE tree growth is essential for predicting alpine forest growth on the TP. On the southeastern TP, where the highest alpine treeline in the world is located (Miehe et al., 2007), increased ATE forest growth has been reported in site-specific studies. However, whether the growth stimulation is due to increased CO_2 concentration or increased growing season temperature is still unclear (Huang et al., 2017; Li et al., 2017). Moreover, attribution of

* Corresponding authors.
E-mail addresses: chunming.shi@bnu.edu.cn (C. Shi), xiuchen.wu@bnu.edu.cn (X. Wu).

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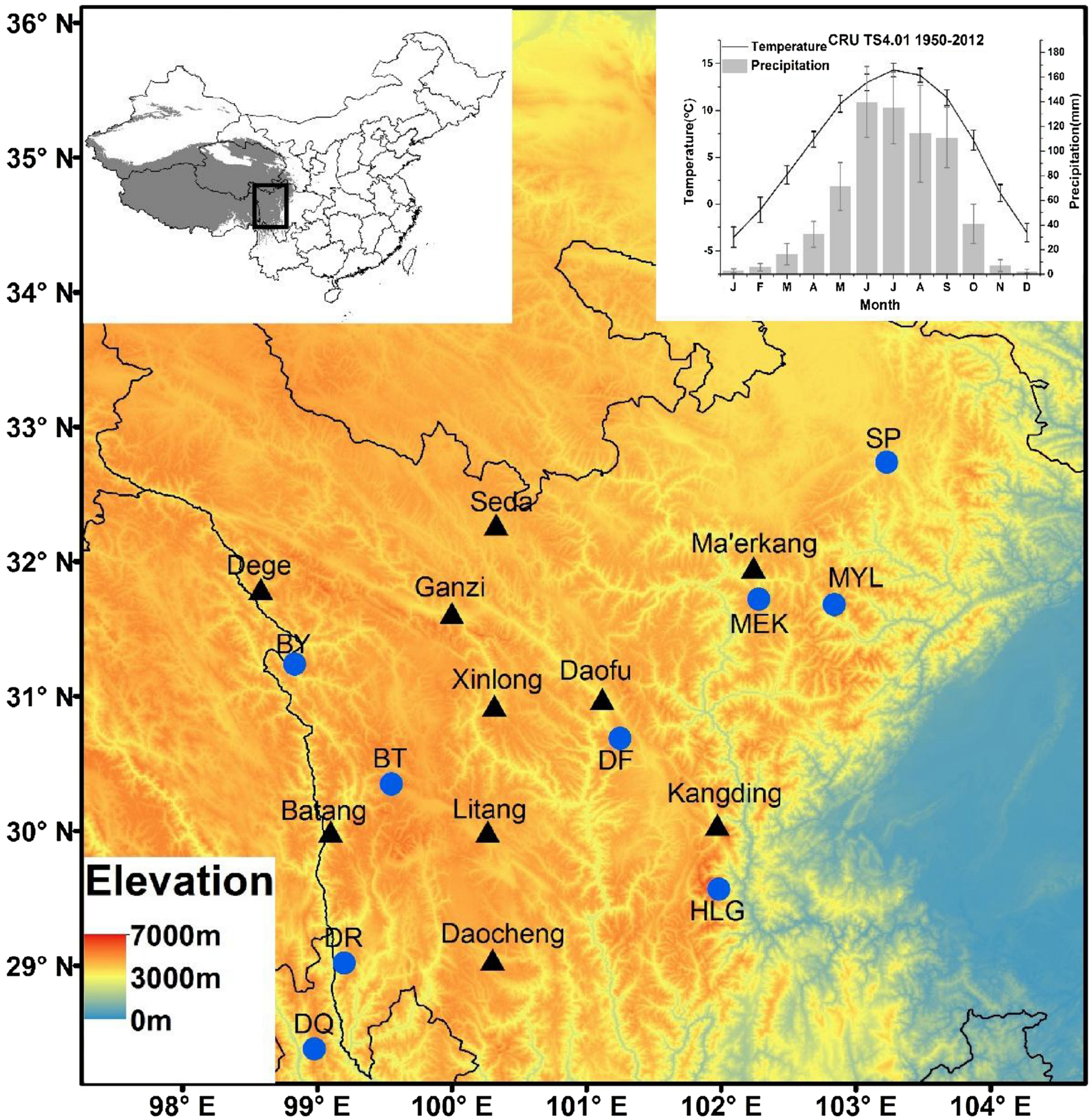

Fig. 1. Sampling sites (blue circles) and meteorological stations (black triangles). The top left insert shows the TP (grey shading depicting elevation 3000 m a.s.l.), with a black rectangle delimiting the region shown in the main figure. The top right insert is the 1950–2012 CRU TS4.01 dataset monthly mean temperature (black curve) and precipitation (grey bars) in the study region (28–33 °N, 98.5–103.5 °E). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Table 1
Latitude (Lat.), longitude (Long.), mean elevation (Elev.) of trees, tree species, and number of trees (No. tree) sampled at each site and the time span of each TRW chronology with an expressed population signal (EPS) value greater than 0.85.

ID	Lat.	Long.	Elev.	Species	No. tree	Chrono. span
BT	30.35	99.55	4160	*Abies squamata*	36	1675-2011
DR	29.02	99.20	4221	*Abies squamata*	37	1752-2011
MEK	31.72	102.28	3967	*Abies faxoniana*	34	1774-2013
DF	30.67	101.25	4113	*Abies squamata*	38	1795-2012
BY	31.24	98.83	4012	*Abies squamata*	38	1796-2012
MYL	31.68	102.84	4150	*Abies faxoniana*	57	1826-2009
HLG	29.57	101.99	3750	*Abies squamata*	26	1869-2012
SP	32.74	103.23	3611	*Picea asperata*	32	1645-2014
DQ	28.38	98.98	4200	*Larix potaninii*	46	1696-2003
MYL	31.68	102.84	3750	*Sabina saltuaria*	37	1796-2009

alpine forest growth variability could be biased by non-climatic factors at the local scale, including inter- and intra-species competition (Liang et al., 2016b; Qi et al., 2015; Wang et al., 2016), soil nutrient availability (McNown and Sullivan, 2013; Sullivan et al., 2015), and topography (Liu et al., 2016; Salzer et al., 2014; Wang et al., 2017). Therefore, a better understanding of the climate response of ATE tree growth at the regional scale could improve our ability to predict regional alpine forest growth.

In this study, we addressed the question of whether the DP exists for forest growth in ATEs on the southeastern TP through a regional tree-ring chronology network and gridded climate data. Specifically, we aimed to understand the following: 1) the relative importance of different climate variables in controlling local and regional ATE forest growth and 2) the temporal stability of the climate response of regional ATE forest growth.

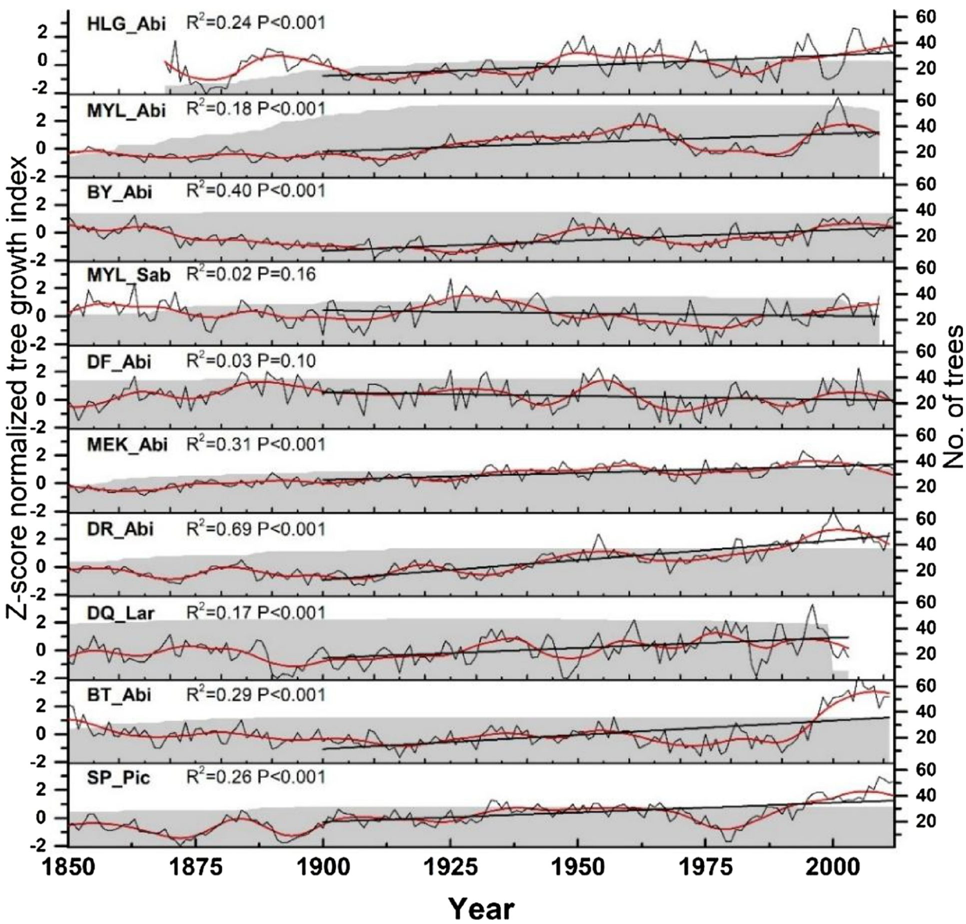

Fig. 2. Tree-ring width chronologies standardized by Z-score (thin black lines) and the corresponding sampling replications (number of trees, grey shading). The thick red lines show the 30-year LOESS smoothing of each chronology. The thick black lines are linear regressions of each chronology. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

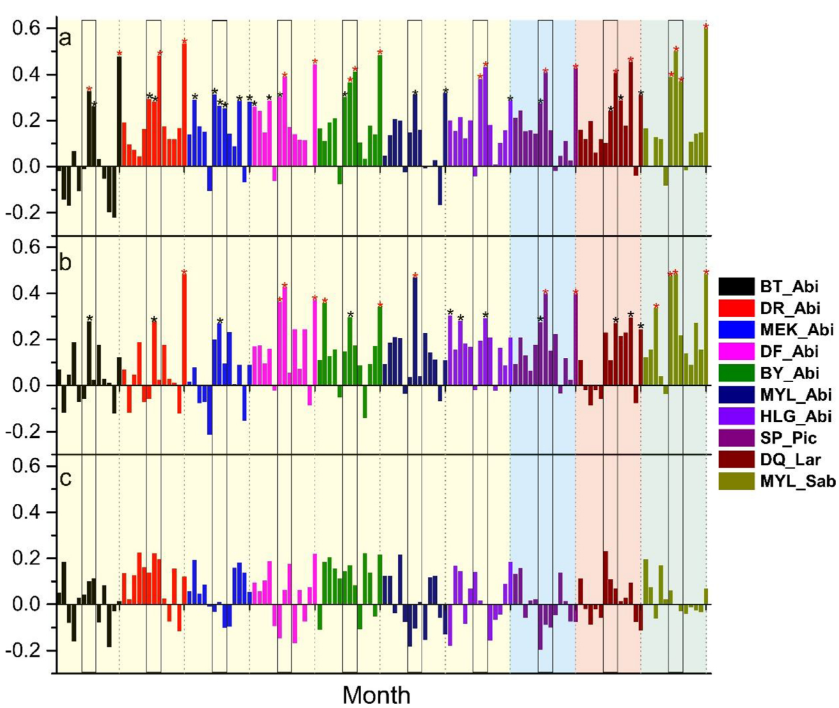

Fig. 3. Correlation coefficients between TRW chronologies and the 1950–2012 monthly CRU TS4.01 climate data of the corresponding grid. a) Minimum temperature; b) mean temperature; and c) precipitation. This analysis was conducted for each month from January to December of the current year and for the current summer (June-August) mean (indicated by grey dashed lines). The results for each chronology are shown in blocks separated by grey dashed lines. The yellow, blue, red and green shading represents species from the genera *Abies, Picea, Larix* and *Sabina*, respectively. Black and red stars indicate coefficients significant at the 95% and 99% levels, respectively. The summer (JJA) season was highlighted with black rectangle. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

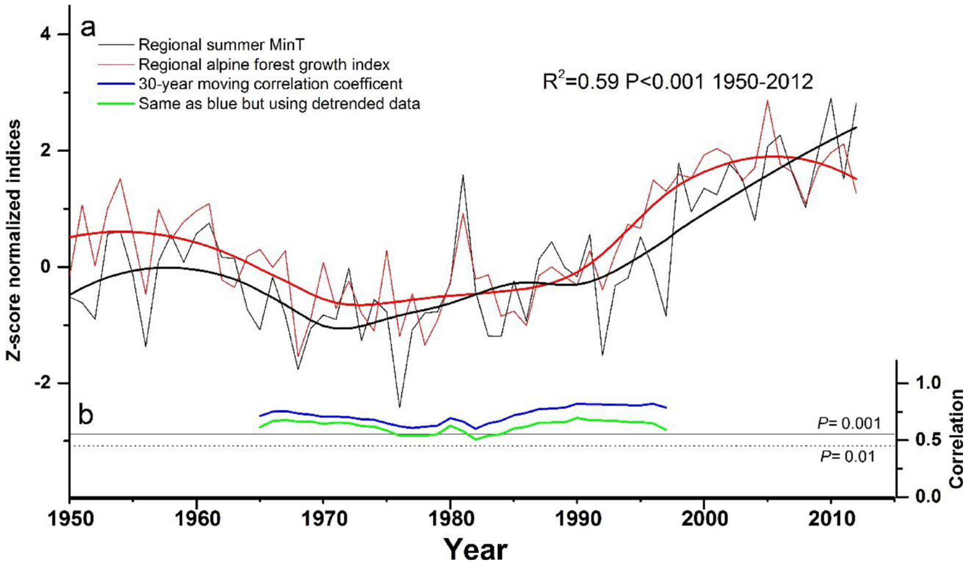

Fig. 4. Index of regional tree growth in alpine treeline ecotones (RTGA, black line) and regional 1950–2012 CRU summer minimum temperature (MinT, red line). The thick lines represent 30-year LOESS smoothing (a); the blue line is the 30-year moving correlation coefficient of the regional RTGA and the summer MinT, and the green line is same as the blue line but using the RTGA and summer MinT detrended by their 30-year LOESS smoothing (b). Horizontal black solid and dashed lines in panel b are 0.001 and 0.01 significance levels for the moving correlation analysis using both raw and detrended data (blue and green lines, respectively). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

2. Materials and methods

2.1. Study area and climate data

Our study region included the western Sichuan and northwestern Yunnan Provinces, which are located on the southeastern TP. Monthly temperature, precipitation, and self-calibrating Palmer Drought Severity Index (scPDSI) data spanning the period from 1901 to 2012 were extracted from the Climate Research Unit (CRU) TS 4.01 dataset with a spatial resolution of 0.5°. Monthly relative humidity data were obtained from ten meteorological stations of the China Meteorology Administration (black triangles in Fig. 1). The Monsoon Asia Drought Atlas (MADA) reconstruction was downloaded fromhttps://www.ncdc.noaa.gov/paleo-search/study/10435 (Cook et al., 2010). The CRU dataset is an interpolation of nearby station data. As extensive meteorological observations did not start until the 1950s over the TP, the CRU data for the TP before the 1950s are less reliable. The annual mean temperature and annual total precipitation for the study region (defined as 28–32 °N, 98.5–103.5 °E) provided by the 1950–2012 CRU dataset were 5.58 °C and 753 mm, respectively, according to our own calculations. Approximately 57% of the precipitation falls in the summer (June-August), which is the main growing season for ATE trees on the southeastern TP.

2.2. Tree ring sampling, processing and chronology construction

Tree cores were extracted from trees in ten ATEs at nine sites on the southeastern TP (Fig. 1). For site MYL, two ATEs with different tree species were sampled (Table 1**)**. In general, 26–57 trees of the following species were sampled in each ATE: *Abies squamata*, *Abies faxoniana*, *Sabina saltuaria*, *Picea asperata* and *Larix potaninii* (Table 1).

The selected trees had open canopies, straight stems and no signs of damage or internal rot. Two cores per tree were bored at breast height with an increment borer (inner diameter of 5.14 mm). The cores were polished, and tree-ring width (TRW) was measured and cross-dated according to standard dendrochronology procedures. The cross-dated TRW series were quality-checked using COFECHA 2002 software (Holmes, 1998). Then, the TRWs of the cores from the same tree were averaged. To preserve the climate signals and avoid the effects of trend distortion, the TRWs of the same species within each ATE were detrended and combined into a standard chronology with Climatic Research Unit Standardisation of Tree-ring data (CRUST) software using signal-free regional curve standardization (sf-RCS) (Melvin and Briffa, 2014a,b).

The portions of the chronology at both ends with an expressed population signal (EPS) value less than 0.85 were truncated. All the standard chronologies were normalized by Z-score for further analysis. The leading principal component (PC1) of the normalized chronologies was calculated to test the common signals among these chronologies.

2.3. Calibration with climate data and response stability test

To determine the key climate factors dominating ATE forest growth, correlations and response functions were calculated with DendroClim 2002 software between the chronologies and the 1950–2012 CRU data of the corresponding grid point, including minimum temperature, mean temperature and precipitation (Biondi and Waikul, 2004). This analysis was conducted for each month from January to December of the current year and for June-August (JJA). Then, the ten chronologies were averaged into one chronology to represent the regional tree growth in ATEs (RTGA), which was calibrated to the 1950–2012 regional climate index exhibiting the highest correlation and response coefficients with the chronologies. The stability of the forest growth response to the climate was tested using a 30-year moving correlation analysis, with 14 and 15 years before and after the year of interest. To eliminate a possible trend effect on the moving correlation result, the same analysis was conducted using the time series detrended by the 30-year locally weighted scatterplot smoothing (LOESS) trends.

3. Results

All TRW chronologies normalized by Z-score are shown in Fig. 2, with a 30-year LOESS smoothing. In the post-1900 period, eight out of ten chronologies showed a statically increasing trend ($P < 0.001$), while the other two chronologies depicted insignificant ($P > 0.05$) decreasing trends (MYL_Sab and DF_Abi). PC1 explained 41% of the total variance, suggesting a strong common signal embedded in these chronologies.

The monthly correlation analysis of the TRW chronology and local climate data showed the following: (1) all chronologies were significantly ($P < 0.05$) and positively correlated with the summer (JJA) minimum temperature (Fig. 3a); (2) six out of ten chronologies were significantly ($P < 0.05$) and positively correlated with the summer mean temperature, with lower correlation coefficients than those for the minimum temperature (Fig. 3b); and (3) correlations between the chronologies and precipitation were insignificant for all months (Fig. 3c).

The response function analysis demonstrated that eight and five

TRW chronologies exhibited significant ($P < 0.05$) positive responses to the local summer minimum and mean temperatures, respectively (Appendix Fig. A1a and b). No significant response was found to summer precipitation (Appendix Fig. A1c).

Calibration showed that 59% of the variance in the 1950–2012 regional CRU summer minimum temperature can be explained by the RTGA index (the averaged chronology) (Fig. 4a). The coherence of the RTGA index and the CRU summer minimum temperature is highly stable over the period from 1950 to 2012, with significant ($P < 0.001$) moving correlation coefficients for the raw data (blue line in Fig. 4b) and the detrended time series ($P < 0.01$, green line in Fig. 4b).

4. Discussion and conclusion

Similar to the DP observed at the northern high latitudes, increased temperature insensitivity and drought-limited forest growth have been frequently reported in ATEs in subtropical and mid-latitude mountains (Gonzalez De Andres et al., 2015; Morales et al., 2004), especially in dry ATEs, such as the Mediterranean Basin (Diego Galvan et al., 2015), the northeast TP (Liang et al., 2016a; Zhang and Wilmking, 2010), and the north slope of the Himalayas, which has a significant rain shadow effect (Dawadi et al., 2013; Liang et al., 2014). For ATEs with sufficient moisture availability, the DP was not detected, and forest growth was mainly controlled by the growing season temperature (Buentgen et al., 2008; Salzer et al., 2009).

Alpine trees start or cease growing when the air temperature is above or below a certain limit (Moser et al., 2010; Rossi et al., 2007; Ziaco et al., 2016). The summer minimum temperature was shown to be a good approximation of the growing season length and was well calibrated with the annual radial growth of ATE forests on the southeastern TP (Li et al., 2017; Liang et al., 2010). Our study region, with an annual precipitation ranging between 500 and 800 mm, is among the wettest regions north of the Himalayas on the TP (Appendix Fig. A2).

The summer warming of the study region has been most prominent since the 1980s (Appendix Fig. A3a) and has been accompanied by a drying trend with decreased PDSI, precipitation and relative humidity since the 1990s (Appendix Fig. A3b–d). However, this drying trend has not yet reached a threshold beyond which ATE tree growth has started to be limited by drought because the RTGA index showed a close and stable coherence with the summer minimum temperature in the studied period. Therefore, we conclude that the DP does not yet exist for the ATE forests on the southeastern TP.

However, little is known about when ATE forest growth will reach the point of moisture limitation. The TP is expected to continue to warm, according to all future projections, except in the most optimistic scenario (i.e., RCP 2.6, and even under this scenario temperatures will continue to rise until the mid-century) (Chapter 14 in IPCC AR5, WG1, 2013). Predicting alpine treeline growth responses to future climate change is inherently complex and depends on many variables including climate, soils and local adaptation (Cavin and Jump, 2017; McCullough et al., 2017; Monson and Grant, 1989), but our study demonstrates that warming summer temperature, not precipitation, is the dominant driver of tree growth in the TP. Thus, the divergence problem has not yet been observed.

Acknowledgements

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Appendix A

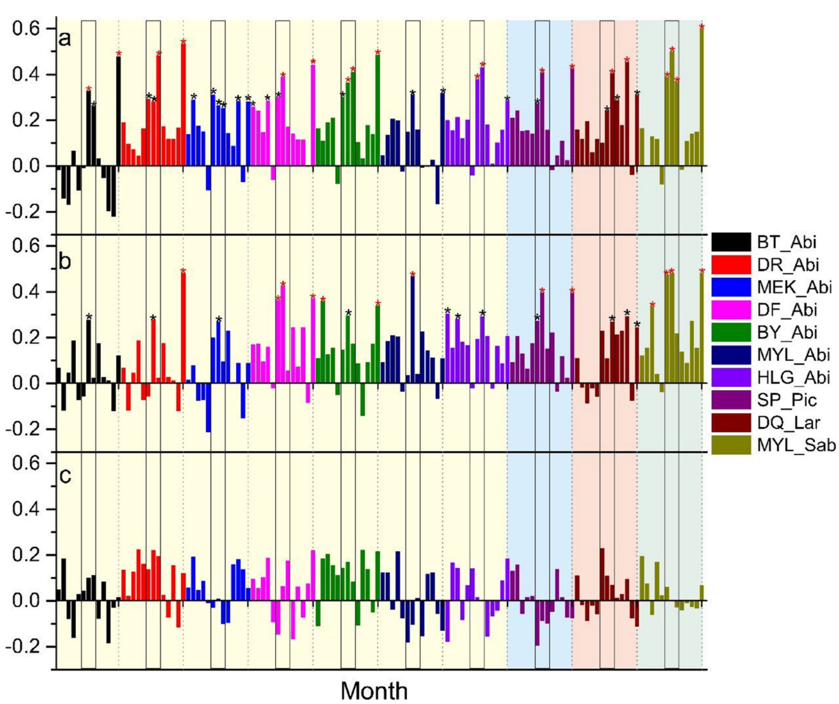

Fig. A1. Response coefficients between TRW chronologies and the 1950–2012 monthly CRU TS4.01 climate data of the corresponding grid. a) Minimum temperature; b) mean temperature; and c) precipitation. This analysis was conducted for each month from January to December of the current year and for the current summer (June-August) mean (indicated by grey dashed lines). The results for each chronology are shown in blocks separated by grey dashed lines. The yellow, blue, red and green shading represent species of the genera *Abies, Picea, Larix* and *Sabina*, respectively. Black and red stars indicate coefficients significant at the 95% and 99% levels, respectively. The summer (JJA) season was highlighted with black rectangle.

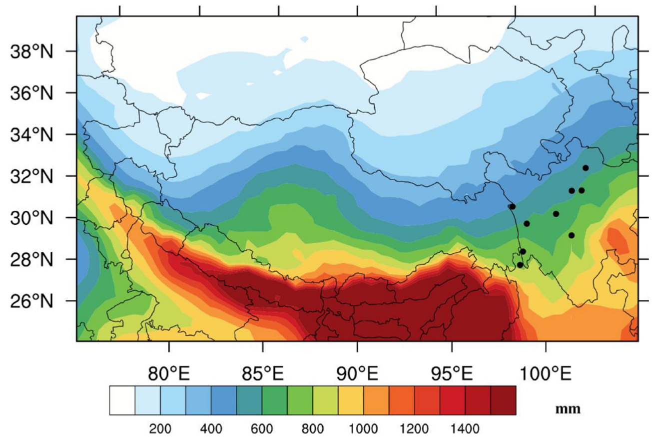

Fig. A2. 1950–2012 annual total precipitation (CRU TS4.01 dataset) for the Tibetan Plateau and the surrounding region. Filled black circles indicate the sampling sites.

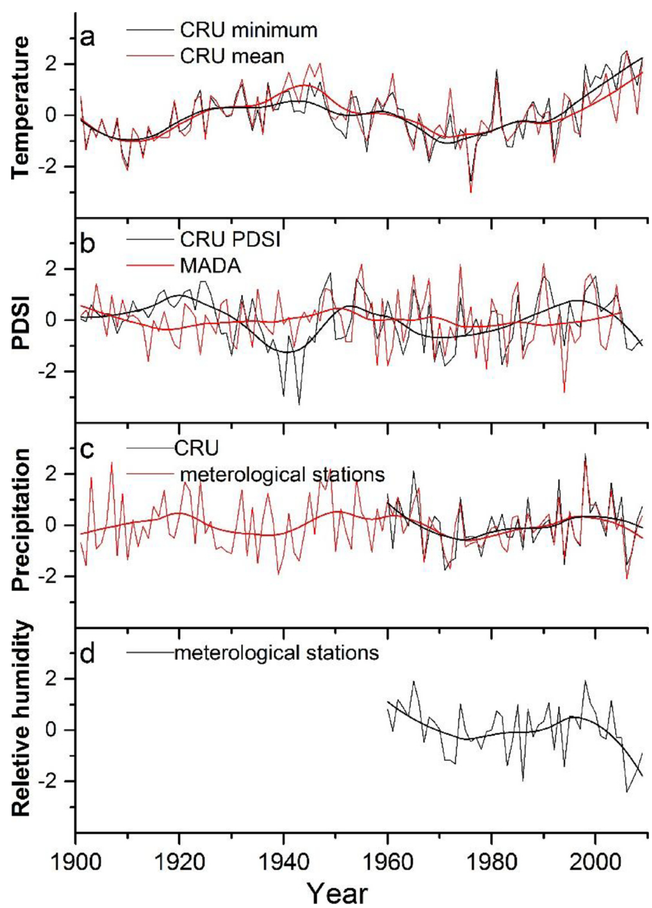

Fig. A3. Summer (JJA) climate variability over the study region, including the CRU mean and minimum temperature (a); the CRU self-calibrating PDSI and the Monsoon Asia Drought Atlas (MADA) reconstruction by Cook et al., 2010 (b); total precipitation documented by the CRU datasets and meteorological stations (c); and relative humidity (d). The thick lines are the 30-year LOESS smoothing.

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