Abstract

The decreasing trend in diurnal temperature range (DTR) since the 1950s has been widely observed in many regions of the world, including the Tibetan Plateau (TP). However, scarce instrumental records on the TP before the 1950s limit the understanding of DTR variation on a longer timescale, and its internal forcing mechanisms remain unclear. Here, we present two regional records of May–June DTR since 1753 reconstructed from tree rings on the northeastern TP (NETP) and southeastern TP (SETP), respectively. The decreasing trend in DTR in the second half of the twentieth century also occurred in the time earlier than the 1950s, indicating that the reduction in DTR is not unique to global warming. Spatially, both the instrumental and reconstructed DTR show different modes on the NETP and SETP. Composite analysis and superposed epoch analysis reveal the linkage between El Niño-Southern Oscillation (ENSO) and the DTR on the SETP. La Niña (El Niño) winters are generally followed by the stronger (weaker) early Indian summer monsoon so that may lead to the decrease (increase) in May–June DTR. The positive correlation between Niño 3.4 index and the DTR on the SETP over the past two and a half centuries is accompanied by a stable response of the decreased DTR to La Niña events. These results suggest that ENSO is a crucial driver for the DTR variation on the SETP through oceanic-atmospheric modulation.
et al., 2016). Given that ecosystem processes such as carbon assimilation and consumption of vegetation (Peng et al., 2013; Xia et al., 2018) depend not only on mean temperature \((T_{\text{mean}})\) but also on temperature extremes and their range (Lindvall and Svensson, 2015), understanding the spatiotemporal pattern of the variation in DTR and its mechanism is of importance in a broader field of research.

The Tibetan Plateau (TP), the highest and largest highland in the world, is considered as a region most sensitive to global change (Liu and Chen, 2000). Meteorological observations on the TP show that the asymmetric increase of annual and seasonal \(T_{\text{max}}\) and \(T_{\text{min}}\) is generally consistent with the global change, and the DTR has also exhibited a decreasing trend (Duan and Wu, 2006; You et al., 2016, 2017). However, numerical models often underestimate the DTR because of the bias in cloud properties, precipitation, and land surface processes (Braganza et al., 2004; Lewis and Karoly, 2013; Lindvall and Svensson, 2015). Instrumental observations, on the other hand, are usually too short to reveal the long-term variation in DTR. Besides that, the Indian summer monsoon (ISM) is one of the dominant circulation systems over the TP (Yao et al., 2012) and greatly affected by El Niño–Southern Oscillation (ENSO) (Webster and Yang, 1992; Krishnamurthy and Goswami, 2000; Roy et al., 2019). Therefore, ENSO may influence the DTR on the TP through its impacts on the ISM, but only a few studies have discussed how the DTR on the TP is associated with climate variability modes and atmospheric circulation patterns (Wu, 2010; You et al., 2016). To date, it is still unknown whether or not the decreasing trend in the DTR on the TP is a phenomenon unique to the recent global warming and how its natural variability is linked with ENSO (Duan and Wu, 2006; Wu, 2010; You et al., 2016, 2017).

Tree rings have been widely used for seasonal temperature and moisture availability reconstructions on the TP (e.g., Fan et al., 2010; Fang et al., 2010; Zhu et al., 2011). Because the DTR is affected by clouds, soil moisture, and precipitation (Dai et al., 1997, 1999), it may also have imprints in the annual growth rings of trees. Previous studies have shown a strong correlation between tree-ring widths and May–July DTR in central Spain (Büntgen et al., 2013) and presented a reliable reconstruction of DTR history using a tree-ring network in Canada (Wilson and Luckman, 2002). In this study, we focus on the TP and aim to compare the recent DTR changes with its variation during the past two and a half centuries by extending the DTR records back to AD 1753 from ring widths of juniper trees \((\text{Juniperus spp.})\). In order to understand the internal forcing mechanisms of the DTR variation, we also investigate the linkage between the DTR and ENSO in recent decades based on observations, and examine whether such linkage can be found in the reconstructed DTR and proxy-based ENSO reconstructions prior to the instrumental period.

## 2 DATA AND METHODS

### 2.1 Tree-ring data

The study area is the eastern TP (the east of 90°E). Moisture-sensitive ring-width chronologies (TRWs) of juniper trees from 23 sites in a northeast–southwest transect form the tree-ring network for analysis of DTR in this study (Figure 1). The geographic area covered by the tree-ring network is 29.45°N–37.37°N, 91.51°E–102.02°E, and the elevations span from 3,260 to 4,440 m above sea level. These site chronologies are numbered from 1 to 23 according to the descending order of latitude (Zhang et al., 2015a). At each site, increment cores were collected from at least 20 individual trees, and tree-ring widths of these samples were measured using a Lintab system with 0.001 m precision and then cross-dated. Next, each series was standardized by removing the non-climatic trends using a cubic spline of 50% frequency-response cutoff at its half-length. These series were then averaged to site chronologies and truncated to the years that have at least five replicate ring-width measurements. The common period of these 23 site chronologies is 1753–2000. More details about sample collection and chronology development were described in Zhang et al. (2015a).

### 2.2 Climate data

Instrumental observations from 85 meteorological stations over the eastern TP are provided by the China Meteorological Administration. Most stations started to operate after the 1960s and 42 of them have continuous observations of \(T_{\text{mean}}, T_{\text{max}}, T_{\text{min}}\), precipitation (Pre), total cloud cover (TC), and low cloud cover (LC) during the period of 1961–2000 (Figure 1). The previous study has found that chronologies in the tree-ring network are sensitive to May–June moisture availability (Zhang et al., 2015a), so here we also focus on May–June tree growth-climate relationship. Empirical orthogonal function (EOF) analysis for May–June DTR of the 42 stations shows that loadings of the first two spatial modes are larger on the northeastern TP (NETP) and southeastern TP (SETP), respectively (Figure 2). Therefore, two
regional DTR series (DTR\textsubscript{NETP} and DTR\textsubscript{SETP}, with the dividing line at 33°N) were calculated from 1956 to 2013 using all available data from the 85 stations because more than 10 stations have been built in each region since 1956. These two series were then used separately as dependent variable to establish the transfer functions for regional DTR reconstruction.

Gridded DTR data are obtained from Climate Research Unit (CRU) TS4.03 dataset with a horizontal resolution of 0.5° by 0.5° (Harris et al., 2020). Monthly mean geopotential height and wind field from the National Centers for Environmental Prediction/the National Center for Atmospheric Research (NCEP/NCAR) reanalysis data (Kalnay et al., 1996) and Palmer Drought Severity Index (PDSI) (Dai et al., 2004) with a horizontal resolution of 2.5° by 2.5° are provided by the National Oceanic and Atmospheric Administration (NOAA)/Oceanic and Atmospheric Research (OAR)/Earth System Research Laboratories (ESRL) Physical Sciences Laboratory (PSL) from their website at http://psl.noaa.gov/. Monthly sea surface temperature (SST) data used here is obtained from Hadley Centre Global Sea Ice and Sea Surface Temperature dataset (HadISST) with a horizontal resolution of 1° by 1° (Rayner, 2003). Monthly Niño 3.4 (5°S–5°N, 120°–170°W) index (Trenberth and Stepaniak, 2001) is obtained from https://climatedataguide.ucar.edu/climate-data/nino-sst-indices-nino-12-3-34-4-oni-and-tni. Historical El Niño/La Niña episodes since 1950 are based on the monitoring of NOAA Climate Prediction Center (CPC). Three winter Niño 3.4 index reconstructions (Cook et al., 2008; Li et al., 2013; Dätwyler et al., 2020) are obtained from https://www.ncdc.noaa.gov/paleo/study/8704, https://www.ncdc.noaa.gov/paleo/study/14632, and https://www.ncdc.noaa.gov/paleo/study/29050 (Last access: July 13, 2020).

2.3 Development of regional chronologies for DTR reconstruction

Screening processes of selecting the chronologies used for DTR reconstruction were mainly based on Pearson’s
correlation analysis during 1961–2000, and criteria for the significant correlation coefficient is at the level $p < .01$. Chronologies which are not significantly correlated with local May–June Pre or $T_{\text{mean}}$ based on the observations of the nearest station were first excluded (Figure 3a). In the second round of screening, nine chronologies that passed the first screening were correlated with May–June DTR, and eight chronologies that have significant and negative correlations with DTR were selected for further analysis (Figure 3b). Then, correlations among these chronologies were calculated on the NETP and SETP separately. One chronology (No.12) on the SETP that does not have significant correlations with all the other four chronologies on its region was excluded. Therefore, only seven chronologies were retained for final DTR reconstruction (marked with black dot in Figure 1 and * in Figure 3b). Information about these chronologies is shown in Table 1. Three chronologies on the NETP and four on the SETP were averaged separately in their common period to derive two regional chronologies ($TRW_{\text{NETP}}$ and $TRW_{\text{SETP}}$). Transfer functions for reconstructing the regional DTR from regional chronologies were established by linear regression model and verified by leave-one-out cross-validation method (Michaelsen, 1987). Spatial correlations between the DTR reconstructions and CRU TS4.03 dataset were calculated to examine their spatial representativeness. In order to compare the regional difference between $DTR_{\text{NETP}}$ and $DTR_{\text{SETP}}$, the reconstructions were truncated in 1753, which is the earliest year of the shortest chronology (Suoxian) used for reconstruction. This analysis period could ensure sufficient sample replication because time spans of the other six chronologies are much earlier than 1753.

### 2.4 DTR variation and its linkage to ENSO

Regime Shift detection (Rodionov, 2004) was performed to identify the turning points of the reconstructed DTR during the past ~250 years. Regional difference between $DTR_{\text{NETP}}$ and $DTR_{\text{SETP}}$ was compared through their 31-year moving linear trends during the instrumental and reconstructed periods. To further assess the temporal pattern of DTR reconstructions, their anomalies in the warm periods were compared with that in cold periods defined by temperature reconstructions from independent studies. Composite analysis of monthly SST evolution in the years of DTR extremes and correlation analysis between DTR and ENSO index were conducted.

#### FIGURE 3

Correlation coefficients between each TRW chronology and climatic variables (pre. $T_{\text{mean}}$, DTR) in May–June during 1961–2000. Black dashed line denotes the significance level at $p < .01$. The seven chronologies finally selected to reconstruct the DTR are marked with *.

#### TABLE 1

| Region | Site Name | Latitude (°N) | Longitude (°E) | Elevation (m) | Time Span | Number of Trees | Mean $R_{\text{sat}}$ |
|--------|-----------|---------------|----------------|---------------|-----------|-----------------|------------------|
| NETP   | Delingha  | 37.37         | 97.37          | 3,820         | 1976–2000 | 37              | 0.76             |
|        | Dulan     | 36.37         | 98.13          | 3,610         | 278B.C.E-2000.C.E | 88              | 0.68             |
|        | Xueshan   | 34.8          | 99.84          | 3,644         | 1320–2005 | 26              | 0.65             |
| SETP   | Suoxian   | 31.63         | 94.29          | 3,854         | 1753–2004 | 27              | 0.45             |
|        | Biru      | 31.12         | 93.87          | 4,350         | 1475–2005 | 31              | 0.55             |
|        | Jiali     | 30.6          | 93.46          | 4,250         | 1542–2004 | 30              | 0.61             |
|        | Linzhou   | 30.31         | 91.51          | 4,233         | 1442–2004 | 30              | 0.58             |

Note: Mean $R_{\text{sat}}$: mean inter-serial correlation.
to investigate the potential linkage between the DTR and ENSO. A modified double bootstrap method of superposed epoch analysis (SEA) developed by Rao et al. (2019) that addresses the linkage between discrete events and continuous time series was performed to identify the response of DTR to ENSO. Their teleconnection was further explained through the composite atmospheric circulation patterns. The reconstructed DTR was also compared with ENSO reconstructions to examine if ENSO had any influence on DTR even prior to the instrumental period.

3 | RESULTS AND DISCUSSION

3.1 | Tree growth-climate relationship

Correlation analysis shows that the site TRW chronologies generally have positive correlations with May–June Pre and negative correlations with $T_{\text{mean}}$. Particularly, the seven TRW chronologies finally selected for DTR reconstruction have stronger negative correlations with DTR than Pre and $T_{\text{mean}}$, indicating that the DTR is also a significant factor influencing tree growth on the eastern TP (Figure 3).

The negative correlations between tree rings and DTR revealed in this study are similar to that in central Spain, which has been well explained by tree physiology. The higher DTR in May–June corresponds to the climate conditions with less precipitation and cloud which may intensify water stress for tree growth, whereas the lower DTR corresponds to cloudy and rainy climate which may promote tree growth (Büntgen et al., 2013). The DTR is affected by the complex interplay of land surface, atmospheric boundary layer processes, clouds, aerosols, and atmospheric circulations (Lindvall and Svensson, 2015), so correlations between DTR, $T_{\text{max}}$, $T_{\text{min}}$, and other four climatic variables (Pre, TC, LC, gridded PDSI) were examined to explain the negative correlations between tree-ring widths and DTR (Figure 4). Generally, the directions of the correlations between DTR and climatic factors are the same as those of $T_{\text{max}}$ but opposite to $T_{\text{min}}$, which is consistent with the definition of DTR ($T_{\text{max}}$ minus $T_{\text{min}}$). The DTR is negatively correlated with TC and LC because clouds can reduce $T_{\text{max}}$ by reflecting

FIGURE 4 Correlation coefficients between (a–d) DTR, (e–h) $T_{\text{max}}$, (i–l) $T_{\text{min}}$ and Pre, TC, LC, PDSI (from left to right) during 1961–2000. Black dots indicate the significance level at $p < .05$. 
shortwave radiation at daytime to cool the Earth’s surface and increase $T_{\text{min}}$ by adding downward longwave radiation at nighttime to heat the Earth’s surface (Dai et al., 1997). The damping effects of precipitation and soil moisture can reduce the DTR by increasing surface evaporative cooling (Dai et al., 1997, 1999), so correlations between DTR and Pre, gridded PDSI are also negative. As an index of climate change and independent of mean temperature (Braganza et al., 2004), DTR includes not only the relative change of $T_{\text{max}}$ and $T_{\text{min}}$, but also the information of precipitation, cloud cover, and moisture availability that affect tree growth. Therefore, tree rings on the eastern TP can be regarded as a comprehensive indicator of annual change in May–June DTR.

### 3.2 Reconstruction of May–June DTR

The regional chronology TRW$_{\text{NETP}}$ is significantly and negatively correlated with regional DTR$_{\text{NETP}}$ in their common period ($r = -0.71$, 1956–2000, $p < .01$), the negative correlation between TRW$_{\text{SETP}}$ and DTR$_{\text{SETP}}$ is also significant ($r = -0.65$, 1956–2004, $p < .01$). Therefore, the transfer functions are developed as follows:

$$\text{DTR}_{\text{NETP}} = 13.50 - 0.70 \times \text{TRW}_{\text{NETP}}$$

$$\text{DTR}_{\text{SETP}} = 13.38 - 0.42 \times \text{TRW}_{\text{SETP}}.$$  

Statistics for testing the reliability of the transfer functions by leave-one-out cross-validation (Michaelsen, 1987) are shown in Table 2. The significant correlation coefficient ($r$), sign test (ST), product mean test (PMT), and reduction of error (RE) between the instrumental and leave-one-out derived series demonstrate the reliability of the regression models (Fritts, 1976). The two regional DTR reconstructions match well with the observations (Figure 5a,b), and their spatial correlations with CRU TS4.03 gridded dataset (Figure 5c–f) confirm the spatial representativeness at a regional scale. Therefore, regional DTR records on the NETP (1753–2000) and SETP (1753–2004) were reconstructed from the regional chronologies based on the transfer functions (Figure 6a,b).

### 3.3 Regional difference of the DTR variation over the eastern TP

Spatial modes of the DTR based on EOF analysis (Figure 2) indicate that the DTR variation on the NETP is not in step with that on the SETP. Climate conditions over the TP usually show different modes on its northern and southern parts due to the effects of two atmospheric circulation systems, that is, the mid-latitude westerlies and the ISM (Yao et al., 2012). In summer, the uplift ISM flows from the Arabian Sea and the Bay of Bengal bring abundant moisture to the southern TP with weakening influence northward. Due to the orographic effect of the TP, the upstream westerlies bifurcate from the west of TP and converge on the east side, leading to cyclonic southerly flows prevail the southern TP and gradually turn into prevailing westerlies northward (Liu and Yin, 2001; Yao et al., 2013). The stable oxygen isotope ratio in precipitation has revealed that the ISM dominates the moisture source over the southern TP while the westerlies dominate the northern TP (Yao et al., 2013). The interplay of these two systems in the transition season (May–June) controls the glacier mass-balance (Mölg et al., 2014) and moisture variability on the TP (Zhang et al., 2015a). Because the westerlies and the ISM greatly influence the hydrological condition over the TP, and the DTR is determined by the climatic variables associated with atmospheric moisture content (e.g., precipitation, cloud cover), the interplay of these two systems (e.g., their intensities, onset time of the ISM, and northward retreat time of the westerlies) may also cause the May–June DTR dipole on the NETP and SETP.

The 31-year low-passed DTR reconstructions reveal fluctuations during the past two and a half centuries but with different high and low periods. Regime shift detection (Rodionov, 2004) by setting cut-off length $l = 30$ years and probability level $p = .1$ shows that shifts of DTR$_{\text{NETP}}$ and DTR$_{\text{SETP}}$ are not at the same time. Although the turning point of DTR$_{\text{NETP}}$ in the 1940s is the only one with a downward shift, the mean value after this point is not lower than the past levels (Figure 6a). The shift of DTR$_{\text{SETP}}$ in the late 1980s is not the only one with a decreasing trend, and the mean value after this turning point is lower than the previous downward shift around 1850, but not yet far beyond the past average levels.

| Period  | Region | $R^2$ | $r$  | ST     | PMT    | RE  |
|---------|--------|------|-----|--------|--------|-----|
| 1956–2000 | NETP   | 0.50** | 0.67** | 34/11** | 3.47** | 0.45 |
| 1956–2004 | SETP   | 0.42** | 0.61** | 34/15** | 3.65** | 0.37 |

Note: $R^2$ is the explained variance of the regression model, $r$ is the correlation coefficient between instrumental series and leave-one-out-derived estimates, ST means sign test, PMT means product mean test, RE means reduction of error. ** represents the significance level at $p < .01$.

**TABLE 2** Statistics of leave-one-out cross-validation for the transfer functions
The results of regime shift detection indicate that the reduction in DTR on the eastern TP in the late 20th century is a notable phenomenon under global warming because of the asymmetric increase of $T_{\text{min}}$ and $T_{\text{max}}$, but not unique compared with its past change. Linear trends of the observational DTR NETP and DTR SETP from 1956 to 2013 calculated in a 31-year moving window (central year moving from 1971 to 1998) also show distinctly different modes (Figure 6c,d). Although both of them have decreasing trends in most of the windows, the DTR SETP variation is not in step with DTR NETP. DTR NETP has already declined at the beginning, and its negative trends are significant in the windows centred in 1974 and 1975, then the magnitude has become weaker (Figure 6c). On the contrary, the magnitude of the negative trends in DTR SETP has gradually become stronger, with significant decreasing trends in the windows centred in the early 1990s (Figure 6d). For the DTR reconstructions, the directions of their trends are not always opposite and show stochastic fluctuations (Figure 6e,f), indicating that DTR dipole on the TP may not stay stable during the past ~250 years.
By comparison between the reconstructed DTR and temperatures ($T_{\text{mean}}$ and $T_{\text{max}}$) over the eastern TP, we further discussed the regional difference between DTR$_{\text{NETP}}$ and DTR$_{\text{SETP}}$, and their relationships with the temperatures. Table 3 shows the number of positive and negative DTR anomalies during the warm and cold periods defined by temperature reconstructions in the references.

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DTRSETP shows regular changes during the warm and cold periods of $T_{\text{max}}$ on central and NETP, which is consistent with its definition ($T_{\text{max}}$ minus $T_{\text{min}}$). The inconsistent directions of DTRSETP anomalies further confirm the regional difference. Neither DTRSETP nor DTRSETP shows regular changes during the warm and cold periods of $T_{\text{mean}}$ because DTR is independent of $T_{\text{mean}}$ (Braganza et al., 2004). The changes of DTR and $T_{\text{min}}$ were not compared here because tree rings are rarely used for $T_{\text{min}}$ reconstruction in the growing season.

The variation in DTRNETP is consistent with the global average DTR which has reduced significantly since the 1950s and remained relatively stable after 1980 (Vose et al., 2005; Thorne et al., 2016), whereas the significant decline in DTRSETP is later than DTRNETP. Although correlations between DTR and other climatic variables are the same on the NETP and SETP (Figure 4), the regional discrepancy of DTR indicates that the driving factors of DTR are different in these two regions. Model simulations of DTR have shown that its decrease in the late twentieth century is mainly caused by anthropogenic forcing (Braganza et al., 2004; Lewis and Karoly, 2013; Liu et al., 2016; Undorf et al., 2018). Tree-ring based reconstruction of annual temperature cycle (the difference between summer temperature and winter temperature) on the TP has also exhibited a decreasing trend since 1870, and such change has been reported to be related to anthropogenic sulfate aerosol emission (Duan et al., 2017). However, the inconsistent reduction in DTRNETP and DTRSETP during the latter half of the twentieth century, and the different regime shifts during the past $\sim$250 years, are not simply caused by anthropogenic activities. Therefore, the potential internal mechanisms of the DTR variation on the eastern TP are investigated in the next section.

### 3.4 | Linkage between DTR and ENSO

The years with extreme DTRSETP and DTRSETP during 1956–2013, defined as the standardized values exceeding $\pm$1, were selected to composite the evolution of tropical Pacific (5°S–5°N, 160°E–100°W) SST anomalies from the previous July to the current June (Figure 7a–d). The patterns of SST anomalies in the years with extreme DTRNETP are not obvious, but show typical ENSO patterns in the extreme DTRSETP years. The centres of SST anomalies mainly occur in the Niño3.4 region, so the instrumental and reconstructed DTR were correlated with monthly Niño3.4 index from the previous March to the current August during their common period (1956–2000) (Figure 7e). DTRSETP is significantly and positively correlated with Niño3.4 index from the previous summer (June) to the current spring (May) ($p < .05$), and there is no significant correlation between DTRNETP and Niño3.4 index. Although it is inevitable to lose a part of amplitude when applying linear regression model for DTR reconstruction, the reconstructed DTRSETP still shows significant and positive correlations with Niño3.4 index from the previous October to the current March ($p < .05$). These results suggest that DTRSETP is associated with ENSO, and there is no significant linkage between DTRNETP and ENSO.

It is also noticeable that the SST anomalies in low DTRSETP years include the characteristics of Eastern-Pacific (EP) and Central-Pacific (CP) La Niña shown in Zhang et al. (2015b). According to the classification of La Niña winters by Zhang et al. (2015b), three low DTRSETP years (1965, 1985, 1996) occurred after EP La Niña winters (1964/65, 1984/85, 1995/96) and another 4 years (1971, 1974, 1999, and 2001) occurred after CP or Mix La Niña winters (1970/71, 1973/74, 1998/99, and 2000/01). The SST anomalies in high DTRSETP years are more like EP El Niño, and 6 years (1966, 1970, 1987, 1998, 1999, 2007) occurred after the EP El Niño winters classified by Yeh et al. (2009), whereas only 1 year (1995) is after CP El Niño. These results indicate that although EP and CP types of ENSO have different dynamic processes and climate effects (Kao and Yu, 2009), almost all types of La Niña events may cause the decrease in DTRSETP. However, the EP El Niño may have a greater impact than CP El Niño on the increase in DTRSETP.

Most of the 20 El Niño (start year: 1957, 1958, 1963, 1965, 1968, 1969, 1972, 1976, 1977, 1979, 1982, 1986, 1987, 1991, 1994, 1997, 2002, 2004, 2006, 2009) and 19 La Niña winters (start year: 1955, 1964, 1970, 1971, 1973, 1974, 1975, 1983, 1984, 1988, 1995, 1998, 1999, 2000, 2005, 2007, 2008, 2009, 2011) during 1956–2013 defined by CPC are coincident with above-average and below-average observational DTRSETP in the following May–June, respectively (Figure 8a). The 34 El Niño and 37 La Niña events during 1870–2000 identified by Yeh et al. (2009) and Ummenhofer et al. (2009) also match well with the reconstructed DTRSETP (Figure 8b). A modified double bootstrap approach to SEA (Rao et al., 2019) further confirms the response of DTRSETP to ENSO. This was conducted by randomly selecting 12 events during 1956–2013 and 30 events during 1870–2000 to make 1,000 composite matrices, then comparing with the results of 10,000 iterations of pseudo-composite matrices to obtain the significance thresholds. The SEA results in these two periods are similar, with significant higher and lower DTRSETP in the following May–June of El Niño and La Niña events, respectively (Figure 8c,d).
To further examine the possible teleconnection between ENSO and DTR$_{SETP}$, anomalies of geopotential height at 500 hPa and wind field at 850 hPa were composited in the following May–June of the El Niño and La Niña winters during 1956–2013 (Figure 9), respectively. In the early summer after El Niño winter, the geopotential height anomalies at 500 hPa are positive over the TP, while the area around Lake Baikal is with negative anomalies (Figure 9a). This circulation pattern reduces the meridional pressure gradient and also weakens the movement of cold air from the high-latitude area to the TP, and the anomalous northerly wind at 850 hPa over the Arabian Sea is not conducive to the transport of water vapour from sea to land (Figure 9c). On the contrary, in the early summer after La Niña winter, an enhanced cold high appears around Lake Baikal at 500 hPa (Figure 9b) and allows stronger cold air intrusion into TP. At the same time, the significant stronger early ISM breaks out over the Arabian Sea at 850 hPa (Figure 9d). As a result, more cold air from high-latitude area and the warm air with abundant moisture from the Arabian Sea converge on the SETP and lead to the cloudy, rainy weather, and the decrease in DTR$_{SETP}$. These results suggest that ENSO may
cause the variation in DTRSETP through the anomalous intensity of the early ISM accompanied by the anomalous geopotential height at 500 hPa.

The Indian Monsoon–ENSO relationship has been reported that during the warm phase of ENSO, descending motion over the Indian Peninsula and equatorial ascending motion form the anomalous regional Hadley circulation, which is sustained by the anomalous Walker circulation in the equatorial Indian Ocean and will lead to less ISM rainfall (Krishnamurthy and Goswami, 2000). Composite wind field at 850 hPa in Figure 9c,d is consistent with the association between ISM and ENSO, and well explains the positive correlation between DTRSETP and Niño 3.4 index (Figure 7e). Summer drought reconstruction over South Asia and the SETP in Monsoon Asia Drought Atlas (Cook et al., 2010) shows a similar correlation with the tropical Pacific SST in winter, and regional climate anomalies also occur in late spring and early summer (Li et al., 2014). To further confirm whether ENSO also had an influence on DTRSETP prior to the instrumental period, the reconstructed DTRSETP was compared with three winter (November–January and December–February, NDJ and DJF, defined based on the year of January and February) Niño3.4 index reconstructions (Cook et al., 2008; Li et al., 2013; Dätwyler et al., 2020). DTRSETP is significantly and positively correlated ($p < .05$) with these three reconstructions in their common periods, their decadal variation also showed in-phase fluctuations in general (Figure 10a–d). According to the regime shift detection of the reconstructed DTRSETP, the years with extreme DTRSETP since 1753 were defined as the standardized anomalies exceeding ±1 in each subinterval. Thirty high and low DTRSETP years were then randomly selected to perform the SEA analysis (Rao et al., 2019) with the reconstructed Niño 3.4 index, respectively (Figure 10e–j). The winter Niño 3.4 index reconstructions do not show significant response to the higher DTRSETP, but all show a significant reduction in the year with lower DTRSETP. Therefore, the correspondence between La Niña winters and DTRSETP is believed to remain relatively stable during the past two and a half centuries. Compared with the spring Niño 4 SST
reconstruction in Liu et al. (2017), fluctuations of DTRSETP were not that in phase with the spring Niño 4 SST before the twentieth century. Periods with relatively low Niño 4 SST during ~1800–1830, ~1870–1890 are characterized with moderate and high DTRSETP, respectively. However, both of them are at a low level around the 1950s. The discrepancy at the early stage may partly due to the less influence of spring Niño 4 SST on DTRSETP. Besides that, the interannual variance of Niño SSTs in Li et al. (2013) and Liu et al. (2017) have both reached their peak level during the late twentieth century, such change of ENSO activity is considered as a response to global warming (Li et al., 2013; Liu et al., 2017). The drastic variation in DTRSETP in recent decades is probably the result of the complex interplay between its internal and external forcing, especially anthropogenic activities, which not only cause the decrease in DTRSETP (Braganza et al., 2004; Lewis and Karoly, 2013; Liu et al., 2016; Undorf et al., 2018) but also affect the key driving factor of its natural variability, that is, ENSO at the same time.

Relationship between moisture-sensitive tree-rings on the SETP and ENSO shows that the positive (negative) tree growth may be caused by La Niña (El Niño) events (Cheng et al., 2019), which is coincident with the results of this study. Since the late 1970s, the structure and frequency of ENSO have changed (An and Wang, 2000) and its relationship with ISM has become weaker (Kumar et al., 1999), and ISM–ENSO teleconnection may change differently in the future (Roy et al., 2019). Meanwhile, the increased sensitivity of tree growth to ENSO has been found in southeastern China and SETP (Wang et al., 2018; Cheng et al., 2019). However, the positive correlation between DTRSETP and Niño 3.4 index is accompanied by a relatively stable response of DTRSETP to La Niña events during the past two and a half centuries in this study (Figures 8e,f and 10h–j). These findings indicate that some information of ENSO can be recorded by trees through its effect on the climate conditions associated with hydrological process, and tree-ring proxy can be used more widely in further studies to identify the long-term variation in ENSO with different phases and intensities.

4 | CONCLUSIONS

We reported the spatiotemporal pattern of regional DTR variability reconstructed from tree rings on the
eastern TP over the last two and a half centuries, and discussed its potential linkage to ENSO. We found different DTR modes on the NETP and SETP both in their observations and reconstructions. The observed reduction in DTRSETP started from the 1950s, whereas the reduction in DTRSETP started from the 1980s. The decreasing trend in DTR is not unique in the latter half of the twentieth century but also occurred differently on the NETP and SETP in the time earlier than the 1950s, indicating that the driving factors of the DTR variation in these two regions are different. The significant positive correlation between DTRSETP and Niño 3.4 index, the composite SST evolution, and atmospheric circulation all suggest that ENSO can influence DTRSETP through its oceanic-atmospheric modulation of the ISM. The positive correlation between DTRSETP and Niño 3.4 index over the past two and a half centuries is accompanied by a stable response of the decreased DTRSETP to La Niña events. However, such linkage does not exist between ENSO and DTRNETP. The internal forcing of DTRNETP variability needs further examination in the future.

Although it has been reported that the decreasing trend in DTR is mainly caused by intensifying anthropogenic emission of greenhouse gases and aerosols (Braganza et al., 2004; Lewis and Karoly, 2013; Liu et al., 2016; Undorf et al., 2018), the regional differences between the DTR on the NETP and SETP suggest that the internal driving factors for the spatial pattern of the DTR variation are quite complex and non-negligible. Knowledge of the DTR variation can provide a more comprehensive understanding of climate change, and it is particularly important for the TP because, being the source area of major rivers in Asia, climate change over here has a profound impact on both ecosystem and society. Therefore, multidisciplinary studies using different climate records and methods are still necessary to understand the mechanisms of internal and external forcing on the DTR variation.

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