Climate variations over the southern Altai Mountains and Dzungarian Basin region, central Asia, since 1580 CE

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Abstract
An improved knowledge of long-term climatic variations over the Altai-Dzungarian region will increase our understanding of the current climate and help to predict the effects of global warming on future water availability in this region. We sampled 77 Larix sibirica Lede. trees at upper and lower treelines in the southern Mongolian Altai mountains and reconstructed temperature and precipitation for longer periods than previous studies from this area. We reconstructed mean June–July air temperatures for the period 1402–2012 and June–December precipitation for the period 1569–2012 based on tree ring width chronologies. The temperature and precipitation reconstructions explain 39.7 and 41.3% of the respective station observation variance during the common periods. The precipitation reconstruction shows alternating wet and dry conditions during the Little Ice Age (1580–1874) followed by more stable conditions until a late 20th century wetting. The temperature reconstruction attributes the warmest period to the 20th century, which follows cooler periods related to volcanic and low solar activities during the Little Ice Age. Long-term climatic variation and change over the Altai-Dzungarian region is inferred from the analysis of the combined temperature and precipitation reconstructions for the common period 1580–2012. Accordingly, this region has become warmer since 1875 as the number of warm/moist and warm/dry years increased by 2 and 14%, respectively, while the number of cool/moist and cool/dry years both decreased by 8% compared to the Little Ice Age. Our findings also reveal a late 20th century cool and wet period, which has also been observed across other mountainous areas of China and Nepal. This period was most probably caused by volcanic-induced cooling and coincided positive phases of the Arctic Oscillation and North Atlantic Oscillation promoting an intensified subtropical westerly jet and a positive summer rainfall anomaly over the Altai-Dzungarian region.

KEYWORDS
global warming, Larix sibirica, Little Ice Age, Mongolia, precipitation, reconstruction, temperature, tree ring proxy
1 | INTRODUCTION

The Altai mountain range lies in the cross-border region of Kazakhstan, China, Mongolia, and Russia. Its vegetation zones follow moisture and temperature gradients with a decrease in moisture and an increase in temperature from north to south and from west to east (Zhang et al., 2015). The Dzungarian semi-desert basin is bounded by the Altai Mountains in the north and the Tian Shan mountains in the south. The lack of long-term instrumental observations from such outback desert and mountain areas hinders the estimation of climate dynamics and change in these regions. Schwikowski et al. (2009) suggested that climate proxies based on tree rings, relict wood, lake sediments, and glaciers, which often can be found in such remote regions in still rather undisturbed states, should be used for exploring regional climate change and variability. Accordingly, tree rings from the Altai mountain range are widely used as climate proxies to reconstruct the past variability of temperature, precipitation, and drought in Mongolia, China, and Russia (Panyushkina et al., 2005; Davi et al., 2009; Loader et al., 2010; Chen et al., 2012; 2014; Zhang et al., 2015; Buentgen et al., 2016). Buentgen et al. (2016) reconstructed the so far longest summer temperature time series starting in the sixth century AD, using tree ring chronologies from the Russian Altai. However, climate reconstructions for the southern Altai have so far only covered temperatures over the past 450 years and precipitation over the past 250 years. Moreover, up to now individual studies have usually reconstructed a single climate parameter for this area and reconstructions of different parameters have not yet been compared and combined.

Climatologically, the Altai Mountains are located at the conjunction of the North Atlantic climate system to the west and the Pacific climate system to the east (Zhang et al., 2018). Thus, the Altai-Dzungarian region is influenced by both southwesterly monsoonal airflow from the Mediterranean Sea, Black Sea, Caspian Sea, and Aral Sea (Iwao and Takahashi, 2006; Zhao et al., 2014; Zhang et al., 2018) and mid-latitude westerly and northwesterly air flow from the Atlantic Ocean during summer (Bohner, 2006; Iwao and Takahashi, 2006; Chen et al., 2015; Zhang et al., 2018). The latter brings warm air and relatively high amounts of precipitation, whereas during winter northerly and westerly winds of the Siberian thermal high-pressure system bring cold and dry airflow from Siberia and the Arctic (Chen et al., 2015; Zhang et al., 2018). Warming till the 1950s and cooling afterwards till the 1990s over the Altai Mountains was revealed in the mean June–July temperature-related tree growth series from upper-treeline sites in the Russian south-east Altai (Panyushkina et al., 2005) and in the mean July–August temperature reconstruction from tree ring oxygen and carbon isotopes of Siberian pine (Pinus sibirica Du Tour) from the central Russian Altai (Loader et al., 2010). The temperature reconstruction from Siberian larch over the Chinese Altai Mountains from Chen et al. (2012) and in east Kazakhstan from Zhang et al. (2015) show similar patterns of change in mean June–July temperatures. This regional temperature variability contrasts with the area-weighted average of the past estimated temperature of all continents from the PAGES 2k Consortium (2013), which locates the warmest period in the late 20th century between 1971 and 2000. Also, their temperature reconstruction for the Arctic suggests that the period 1941–1970 was warmer than 1971–2000; they also stated that continental-scale temperature reconstructions from multi-proxy data show no globally synchronized warm and cold episodes during the past two millennia, except for the overall relatively cold conditions between 1580 and 1880, the so-called Little Ice Age (LIA). The past climate variability thus not only differed between continents but also between eastern and western Asia due to local factors like the Tibetan Plateau and the complex monsoon systems, which affect the stability of climatic teleconnections (Shi et al., 2015). This suggests that the past temperature variability and change over the southern Altai and Dzungarian Basin region could be confined to this regional scale and differ from the mean Northern Hemisphere temperature pattern.

The Altai Mountains were identified as an important climatic boundary in previous regional moisture distribution studies. Zhang et al. (2018) revealed that precipitation increased significantly in the southern Altai between 1966 and 2015, while it decreased in the northern Altai. Also, Chen et al. (2015) found a drying trend in total annual (previous July to current June) precipitation from reconstructions for the northwestern Chinese Altai since the 1980s, while a wetting trend was found for June–July precipitation from reconstructions over the southern Chinese Altai in the 20th century by Chen et al. (2014). Davi et al. (2009) reconstructed the Palmer drought severity index (PDSI) for the growing season (June–September) and found higher moisture availability for a grid point in western Mongolia throughout the 20th and 21st centuries compared to earlier times.

In general, previous studies illustrated, that the southern Altai Mountains and the adjacent Dzungarian Basin could have distinctive temperature and moisture variabilities and trends due to their location and local factors. In this study, we reconstructed and analysed long-term (>400 years) temperature and precipitation time series over that area in order to better understand climate variability caused by the interaction among different natural factors influencing this inland area.
In section 2, we describe the sampling sites, the applied climate and tree proxy data analysis, and the statistical model development for the climate reconstruction. Section 3 shows the detected climate–growth responses of the tree ring width series, the derived temperature and precipitation reconstructions and their spatial correlations, and the inferred climate anomalies between 1580 and 2012. The paper concludes with a summary in section 4.

2 | MATERIAL AND METHODS

2.1 | Site description and tree ring data

In the southern part of the Mongolian Altai Mountains (Figure 1c), patches of Siberian larch forest are often found on north, west, and northwest facing slopes (Figure 1a, b). During our fieldwork in the area, local people asserted that the main disturbance has been intensive logging from the 1960s to 1990s and that in recent decades no widespread fires have occurred. Not much is known about the impact of wildfires on these patchy forests. In the national fire statistics for the period 1963–1997, the majority of wildfires are noted from forested areas of central and eastern Mongolia after the snowmelt in March to mid-June and in autumn (Bayartaa et al., 2007). Regional grassland fire reconstructions based on charcoal analyses of sediments from 48 lakes in northwestern Mongolia with sediment influx from the Altai revealed no recent wildfire evidence (Umbanhowar et al., 2009). The Mongolian Altai mountain forest is also too cold for insect outbreaks (Dulamsuren and Khishigjargal, 2012). In the Russian northern Altai, a Siberian silk moth outbreak was confined to southern slopes of 11–13° steepness located at approximately 400 m above sea level (a.s.l; Kharuk et al., 2016).

We took cores from 53 trees at the sites Kargait (KAR), Khets (KET), and Gurt (GUR) located at upper treelines in July 2014 in addition to 20 trees at the sites Yolt (YLT), Shiregt (SHR), and Khudagt (KUD) located at the lower treeline of forest patches in July 2013. The positions of all sites are shown in Figure 1c; see also Table S1, Supporting Information for more information on all six sites. In this study, we only used the longer and more climate sensitive KAR and KET chronologies. Information on the other chronologies and derived reconstructions is shown in Supporting Information (see, e.g., the climate sensitivity analysis for all site chronologies in Figure S1).

The two sites KAR and KET selected for this analysis had rather steep slopes (25–33°) and revealed no indications of impacts by fire (scar) or insects (loss of needles) and only minimum human disturbances according to interviews with local officers and our own site inspections. KAR is located on a north facing slope and KET on a west facing slope, which makes it more prone to westerly winds, more rainfall, and more radiation. The KAR site is characterized by large boulders covered with a thin soil layer and alpine shrubs (Figure 1a) while the KET site has a well-developed soil layer with grasses and young trees (Figure 1b). We took 32 cores from 17 trees at KAR and 46 cores from 23 trees at KET. Except for two trees, two cores were taken from each tree from opposite sides perpendicular to the slope direction.

The cores of each site were marked with calendar dates and visually cross-dated using so-called pointer years (isolated...
years with exceptionally narrow or wide rings). During cross-dating, the tree ring width series were scanned for potential missing and false rings. By matching the tree ring width patterns among the cores and by examining the ring structure (Fritts, 1976), for each site mean chronologies based on common growth sequences of the trees were derived and compared to all cores from a site to detect missing, partial, and false rings formed under severe conditions. The tree ring widths were measured to the nearest 0.001 mm with a Velmex measuring system and the MeasureJ2X software (Velmex, Inc., Bloomfield, NY). The measurement accuracy and the visual cross-dating among the individual tree chronologies at each site were checked by statistical cross-dating using COFECHAv6.06 (Grissino-Mayer, 2001), which calculates Pearson correlation coefficients between segments of individual ring width series and a master chronology consisting of all other series at the dated position and 10 positions forwards and backwards. Flagged, that is, potentially incorrectly dated segments were re-examined and corrected when missing or false rings in that ring width segment were found. Standardized indices were calculated from the individual tree ring width series by fitting a Friedman super smoother growth curve with alpha 5, and dividing each ring width for a certain year by the value of the growth curve (Figure 2a,c). This data-adaptive smoothing technique (implemented in ARSTAN for Windows, version ARS41c_xp; Cook and Krusic, 2006) preserves the low-frequency variance (Friedman, 1984) and removes effects of aging and other non-climatic trends from the series (Cook, 1985). The standardized indices from the individual cores were then averaged to obtain the site chronology (Fritts, 1976).

We wanted to treat all data consistently and therefore applied the same detrending procedure to all series. Some cores came from lower-tree line trees, which contained logging effects in the past 50 years. These effects were most effectively removed by a Friedman super smoother curve with alpha 5. Although the latter is rather flexible and potentially removes longer-term (>100 years) trends, low-frequency variance was preserved as shown through the application of more conservative detrending procedures, which had very similar outcomes (see Figure S2 for a comparison of the detrended chronologies resulting from Friedman super smoother curves with different alphas).

The climate signal strength within a site chronology and its reliability are commonly described by its mean sensitivity, expressed population signal (EPS), and mean correlation between the individual tree ring series ($\bar{r}$). The mean sensitivity is the relative difference between adjacent ring widths and indicates the range of the year-to-year variations in radial growth in response to climate (Fritts, 1976). EPS quantifies the strength of the common signal in a given set of tree ring series used for a chronology (Wigley et al., 1984; Cook and Kairiukstis, 1990), and is based on $\bar{r}$ and the sample size. Wigley et al. (1984) suggested .85 as a lower acceptable threshold for the EPS. $\bar{r}$ is a measure for the common growth signal or common variance between all tree ring width series. We calculated $\bar{r}$ and EPS with ARSTAN for all tree ring series contributing to one chronology in 50-year intervals with 25-year overlap. The actual computation of the standardized chronologies, which are later used for the climate reconstruction (Figure 2b,d) is described in section 3.1.

2.2 | Climate data

Only very few climate stations exist in this region with sufficiently long and complete records. The nearest climate

![Image](image_url)

**FIGURE 2** Mean L. sibirica ring width chronologies (dark grey) for the KAR (a) and KET (c) sites with a fitted Friedman super smoothing curve with alpha 5 (black line) and individual ring width series (light grey lines). The mean standardized KAR and KET chronologies (ring width index [RWI]) are shown in (b, d), respectively, together with the sample depth.
station Duchinjil (1951 m a.s.l.) is located in a valley approximately 36–40 km away from the sampling sites and has an instrumental record starting in 1977. In addition to this short time series, we also considered observations from more distant climate stations in the southern Altai Mountains established between 1954 and 1963 (Figure 1c and Table S2) to find appropriate time series for climate reconstruction. We also used monthly mean gridded CRU data (CRU TS4.01; Harris and Jones, 2017) averaged over the area between 46°–47°N and 91°–92°E for the period 1963–2012, which we downloaded via the Climate Explorer website of the Royal Netherlands Meteorological Institute (KNMI, http://climexp.knmi.nl). This series reflects the time series of all local climate stations well. Moreover, we examined the monthly and seasonal climate indices Arctic Oscillation (AO; NOAA/NCEP/CPC, 2019) for the period 1950–2012, North Atlantic Oscillation (NAO) based on Iceland and Gibraltar surface air pressure (Jones et al., 1997) over the period 1825–2012, and NINO4 for the period 1854–2012 (Huang et al., 2017) all obtained from the KNMI in February 2019, each of which may relate to moisture variability over the study area as suggested in previous studies (Davi et al., 2009; He et al., 2017). At the station closest to the sampling sites, Duchinjil, 55% of the total annual precipitation (129 ± 33 mm in the period 1977–2012) fell during summer (June–August), while the winter months (December–February) contributed only 6% on average to the annual sum (Figure 1d), although much larger contributions may occur occasionally. Since July is often the wettest month—typical for Mongolia and Inner Mongolia—its inter-annual variability is often used to quantify the precipitation variation in this region (Iwao and Takahashi, 2006). Mean monthly air temperatures range from −22.0°C in January to +15.8°C in July at Duchinjil (Figure 1d).

2.3 | Climate–growth response analysis

Precipitation and temperature are growth controlling climate factors, since they affect soil moisture availability, tree transpiration and respiration, and photosynthesis (Fritts, 1966). Growth of Siberian larch in the Altai Mountains mainly occurs between April and September. Although cell division and lengthening ceases at the end of the growing season, thickening of the tracheid cell walls might continue depending on the weather conditions (Chen et al., 2012). Hence, we tested correlations between the site chronologies and both monthly and seasonal mean air temperature and precipitation sums. We used DendroClim2002 (Biondi and Waikul, 2004) to estimate the relations between monthly climate variables and tree ring width for the time period of available instrumental and gridded climate data. DendroClim2002 uses 1,000 bootstrapped samples to compute Pearson’s correlation coefficients and their significance. Significances were tested at the p < .05 level.

The site chronologies (averaged standardized tree chronologies, Figure 2b,d) were correlated with the monthly mean temperatures and precipitation sums from the Duchinjil station, the monthly mean regional air temperature as estimated by the averaged longer time series of five climate stations including the Baitag, Khovd, and Ulgii in Mongolia, and the Fuyun and Altay in China, the monthly precipitation sums from the Qinghe station, monthly temperature means and precipitation sums from the CRU TS4.01 data set, and the monthly mean scPDSI, the latter two averaged over the area from 46°–47°N and 91°–92°E. Correlations were determined for all months of a 18 months window starting May of the year prior to growth and October of the current year of growth (for an overview of the correlations, also for the other four sampling sites, see Figure S1a–f). The monthly and seasonal climate data most significantly correlated with the site chronologies were selected for setting up transfer models for climate reconstructions.

2.4 | Regression model development for summer air temperature and precipitation reconstruction

Transfer functions between the chronologies (predictor) and the climate parameters (predictand) were developed for the individual site chronologies using simple linear regression. If tree growth correlated with the climate conditions of both the previous and the current year, the site chronology and its lag by 1 year relative to the climate observations were considered as predictors. In addition, the first principal component of these predictors resulting from a principal component analysis (PCA) of the complete time series was also used as predictor.

The reliability of the resulting transfer functions was assessed via calibration and verification statistics commonly used in dendrochronology, that is, Pearson’s correlation coefficient (r), coefficient of determination (R²), reduction of error (RE), coefficient of efficiency (CE), product mean (PM) test, and first difference sign test (Cook et al., 1999; Weijers et al., 2010). r quantifies the association between tree rings and climate while the percentage of explained variance by the ring widths is given by R². RE and CE range between minus infinity and one; positive values close to 1 indicate good skill and negative values lower skill (Fritts, 1976; Cook et al., 1999; Weijers et al., 2010). The PM test (Fritts, 1976) takes into account both sign and magnitude of the actual and estimated departure from the mean value, while the non-parametric first difference sign test uses only the sign of change to quantify similarities between two
series. These calibration and verification statistics were calculated over separate periods while the full calibration periods, that is, the complete overlapping periods between climate and tree ring data, were used for the construction of the transfer functions. Besides the individual site chronologies, separate regional tree network chronologies for temperature and precipitation were constructed from the first principal components of two and more individual site chronologies sensitive to the same climate condition over common periods and applied in the development of regional climate transfer functions. This was performed to verify our climate reconstructions based on single site chronologies (for an overview of the individual and regional temperature and precipitation reconstructions, see Figure S2a,b).

3 | RESULTS AND DISCUSSION

3.1 | Site chronologies and climate–growth responses

Site chronologies for KAR and KET were established for the period 1402–2012 and 1569–2012, respectively (Figure 2). Tables 1 and S3 show summary statistics of the site chronologies. After detrending, the first-order autocorrelations of the site chronologies decreased as expected. The EPS for both site chronologies remains above the threshold of .85 (Wigley et al., 1984) over their entire chronology lengths. The running-mean correlation coefficients between the individual tree ring series in a chronology (萤) for 50-year intervals with 25-year overlap ranges between .44 and .65.

Over the period 1835–2012, which is covered by all six site chronologies (i.e., KET, KAR, and the other four sites initially analysed), upper- and lower-treeline chronologies obtained from northwest and west facing slope orientation tended to be correlated significantly with each other (Table S4); for example, the KET upper-treeline chronology from a west facing slope was relatively weakly correlated with the KAR chronology (also upper treeline but north facing slope) (萤 = .26 at p < .01), whereas the KET chronology was strongly positively correlated with the GUR upper-treeline chronology from a northwest facing slope (萤 = .80 at p < .01). The KET chronology was also relatively strongly correlated with the SHR lower-treeline chronology also obtained from a west facing slope (萤 = .56 at p < .01). This finding is in line with Fritts (1976), who sees slope orientation as an even more important site factor for growth response than elevation and latitude, because it more strongly affects the water and energy balance by controlling the amount of radiation received by the site, which in turn influences the allocation of moisture. Regardless of position (upper or lower treeline) and slope orientation, significant correlations among the site chronologies suggest one or more shared external growth-driving factor, like, for example, cooling by aerosols from volcano eruptions, and water stress (Table S4).

In the correlation analyses with the climate data, the current year air temperature at KAR and the prior year precipitation at KET were identified as the main drivers of radial growth (Figures 3 and S1). The KAR chronology correlated positively with the mean monthly temperatures of June (萤 = .66, p < .01), July (萤 = .40, p < .01), and mean June–July (萤 = .63, p < .01) of the current year, and of June (萤 = .38, p < .01) and July (萤 = .27, p < .01) of the previous year of the gridded CRU TS4.01 time series over the period 1963–2012 (Figure 3a). These results corroborate findings from previous studies in the Chinese southern Altai Mountains by Chen et al. (2012), Zhang et al. (2015), and Wang et al. (2013), who showed positive correlations of tree ring widths with June–July air temperatures. They believe that the increased radial growth is caused both by higher photosynthetic rates and higher soil moisture levels due to increased snowmelt. The CRU TS4.01 temperatures for the period 1963–2012 were preferred by us over the Duchinjil station data because of the longer length of this time series. The CRU TS4.01 temperature data before 1963 were not used due to too few stations in the region and the related low reliability of interpolations for that time period.

The strongest correlations between precipitation sums (萤 = .64, p < .01) and the KET chronology were those with June through December precipitation sums of the year prior to growth as measured at the Duchinjil station for the period 1977–2012 (Figure 3b). Previous year monthly precipitation sums of July (萤 = .37), August (萤 = .36), and November (萤 = .30) were also significantly correlated (p < .05) with the KET chronology, as were those of July through November (萤 = .58, p < .01) and July through December

| TABLE 1 | Summary statistics of the raw and standardized ring width chronologies |
|---|---|---|
| Tree ring sites | KAR | KET |
| Period, years | 1402–2012, 611 | 1569–2012, 444 |
| First-order autocorrelation | | |
| Raw | .89 | .77 |
| Standardized | .47 | .67 |
| Average mean sensitivity | | |
| Raw | .212 | .233 |
| Standardized | .206 | .228 |
| EPS* | .92–.98 | .91–.98 |
| p | .44–.65 | .48–.63 |
| Period, number of years with EPS > .85 | 1450–2012, 564 | 1645–2012, 369 |

*Expressed population signal (Wigley et al., 1984).

*Mean correlation coefficient among all tree ring series used in a chronology.
Our findings are largely in line with Chen et al. (2014), who found positive correlations between the radial growth of Siberian spruce in the Chinese southern Altai Mountains with July–August and December precipitation sums of the previous year in addition to May–July precipitation of the current year. According to Fritts (1974) the growth of arid-site conifers during the current year might be enhanced by above-average precipitation in late summer and early autumn of the previous year due to its promotion of carbohydrate storage and bud formation. Winter precipitation, which mostly falls as snow from November to March in this area, increases soil moisture during the early growing season due to snowmelt and may thus also promote growth.

3.2 Summer air temperature reconstruction

We developed a temperature transfer function model based on the KAR chronology for June–July temperatures of the gridded CRU TS4.01 data set over the period of 1963–2012. The first principal component from the PCA of the predictors—the KAR ($t$) and its lag ($t + 1$) series—contained 79% of the total variance and was used in a linear regression model for the mean June–July temperature reconstruction. The statistics of the model skills were calculated for the calibration period 1963–1986 and for the verification period 1987–2012, which was repeated with both periods interchanged (see overview of the transfer function models based on all chronologies and their calibration and verification statistics in Figure S3 and Table S5). The transfer function model verification statistics for the mean June–July temperature reconstruction (Table 2) indicated—with positive values of RE, significant PM values, and significant first-difference sign tests ($p < .05$)—that our temperature reconstruction captures the high-frequency variation of the instrumental data well. Therefore, we used the transfer function model from the KAR chronology calibrated over the whole period between 1963 and 2012 for the reconstruction of mean June–July temperatures for the period 1402–2012.

**TABLE 2** Calibration and verification statistics for the transfer function model of the mean June–July temperature reconstruction from the KAR chronology and the monthly mean temperature averaged over CRU TS4.01 grids (91°–92°E; 41°–42°N)

|          | Full calibration (1963–2012) | Calibration (1963–1987) | Verification (1988–2012) | Calibration (1988–2012) | Verification (1963–1987) |
|----------|-----------------------------|-------------------------|--------------------------|-------------------------|-------------------------|
| $r$      | .631***                     | .734***                 | .603***                  | .605***                 | .733***                 |
| $R^2$    | .397**                      | .539***                 | .366**                   | .366                    | .356                    |
| Adj$R^2$ | .371**                      | .497                    | .338**                   | .338                    | .356                    |
| RE       | .347                        | .539                    | .238                     | .366                    | .356                    |
| CE       | .539                        | -.183                   | .366                     | -.198                   | .356                    |
| Sign test| 18+/7= −                   | 14+/11= −               | 22+/3= −***              | 16+/9= −                |
| Products means test | .399**                   | .597*                    | .095                     | .326**                  |

Abbreviations: Adj$R^2$, adjusted for degrees of freedom; CE, coefficient of efficiency; $r$, Pearson’s correlation coefficient; $R^2$, coefficient of determination; RE, reduction of error statistic.

$p < .05$; **$p < .01$; ***$p < .001$. 

**FIGURE 3** Significant ($p < .05$) Pearson’s correlation coefficients between the KAR chronology and the monthly and seasonal mean air temperatures of the regional temperature averaged over five stations, and the CRU TS4.01 (averaged over the grids of 91°–92°E and 41°–42°N) for the period 1963–2012, and the nearest station (Duchinjil) for the period 1977–2012 (a) and between KET and the monthly and seasonal precipitation sums as measured at the Qinghe station, the CRU TS4.01 (averaged over the grids of 91°–92°E and 41°–42°N), and the Duchinjil station over the period 1958–2007, 1958–2012, and 1977–2012, respectively (b).

c, current year; p, year prior to growth

(r = .60, $p < .01$).
The reconstructed June–July temperature time series explained 39.7% of the year-to-year variance of the gridded CRU TS4.01 data over the period 1963–2012 and contained the same general positive trend as the instrument observations (Figure 4a). This common trend between tree ring and instrumental temperature series might lead to partly spurious correlations. After the removal of these linear trends from the KAR chronology and the mean June–July temperatures by taking the difference between the value in 1 year and the previous year, a moderate correlation ($r = .49$, $p < .01$) was still found between both detrended time series. However, such common trends may be causally linked and the removal of such trends from both the tree ring and temperature series might prevent the reconstruction of similar past trends based on the tree rings and lead to a false detection of divergence (Weijers et al., 2012).

The spatial distribution of the correlation between reconstructed and gridded instrument-based CRU TS4.01 mean June–July temperatures for the period 1963–2012 (Figure 5a) demonstrates that our temperature reconstruction contains a clear regional signal, which covers the Altai Mountains and the western Sayan Mountains in northern Mongolia ($r = .5–.6$, $p < .1$), the Mongolian Plateau ($r = .3–.5$, $p < .1$), and the Dzungarian Basin ($r = .2–.5$ depending on the distance from the Altai Mountains, $p < .1$).

According to the resulting 611-year June–July temperature reconstruction, mean summer temperatures ranged from 9.3 to 16.0°C over the period 1402–2012 in the region. The observed summer temperature variability is within the uncertainty range ($\pm$ standard error of the prediction) of the reconstructed temperature over the common period (Figure S4).

In a graphical comparison with other tree ring-based reconstructed temperatures from the southern Altai from previous studies many similarities can be found (Figure 6). All reconstructions lack the 20th century warming trend observed over the Northern Hemisphere. Warm decades are suggested for the periods 1880–1910 and 1940–1975 and cold periods for 1490s, 1540s, 1680–1710, 1780s, 1810–1860, 1911–1939 and 1980s. These cold periods and the cooling in the 1930s are also revealed by a 750-year high-resolution temperature reconstruction (1250–2000) from an ice-core oxygen isotope record from the Belukha glacier in the Siberian Altai and explained as periods of low solar activity (Eichler et al., 2009; Schwikowski et al., 2009). Moreover, periods of volcanic-induced cooling (Briffa et al., 1998; Eichler et al., 2009; Buentgen et al., 2016) and periods of low solar activity (Schwikowski et al., 2009) coincide with the observed periods with low temperatures in our reconstruction. Most of the coldest summers took place during such periods of low solar activity and volcanic eruptions (Figure 6).

Chen et al. (2012) explained the late 20th century cooling as a consequence of enhanced cloudiness and rainfall over the Altai Mountains and reduced growth of Larix sibirica at the treeline. The warm period in the 1950s and the cold period between 1983 and 1998 were also observed in tree ring chronologies from mountainous areas in Nepal and China (Wang et al., 2013).

None of the temperature reconstructions from the Altai Mountains (this study and Panyushkina et al., 2005; Loader et al., 2010; Chen et al., 2012; Zhang et al., 2015, Buentgen et al., 2016) indicate a continuous 20th century warming trend as observed in the northern Mongolia reconstructions (d’Arrigo et al., 2000; 2001; Davi et al., 2015). Instead, a decrease in summer temperatures starting in the 1950s followed by a steep rise in the 1990s was observed. This recent rapid warming over Mongolia slowed down temporarily since 2002 probably following a natural global climate change pattern.
variability caused by a redistribution of heat in the ocean, volcanic eruptions, the recent minimum in the solar cycle, and the decadal cooling caused by La Niña in the Pacific Ocean (Dagvadorj et al., 2014).

3.3 Precipitation reconstruction

A transfer function model was developed based on the lagged \((t + 1)\) KET chronology and the June through December precipitation sum as observed at the Duchinjil station over the period 1977–2012 using simple linear regression. This model was first calibrated over the period 1977–1995 and verified over the period 1996–2012, which was repeated with interchanged periods. The reconstruction of June–December precipitation sums explains 41.3\% of the variance in the instrumental data over the complete period (1977–2012). The positive RE and CE values, and significant PM values suggest a high reliability of the transfer function model (Table 3). Consequently, June–December precipitation sums were reconstructed based on the KET chronology \((t + 1)\) for the period 1569–2012. The correlation between reconstructed and instrumental data over the whole common period is .64 \((p < .01)\) (Figure 4), see overview of the transfer function models based on all chronologies and their validation statistics in Figure S3b and Table S6).

The spatial distribution of the correlation between the reconstructed and the gridded instrument based June–December precipitation sums from 1977 to 2012 (Figure 5b) shows a more local signal—compared to the mean June–July temperature reconstruction—with significant positive correlations over the southwestern and southern Altai Mountains \((r = .4–.5, p < .1)\) and the Dzungarian Basin \((r = .3–.4, p < .1)\). Interestingly, the strength of the spatial correlations between the chronologies and gridded temperature and
precipitation data decreases when the longer period between 1950 and 2012 is considered (Figure 5c,d).

The reconstructed precipitation variability for the time period 1569 to 2012 shows recent wet maxima in the 1950s (1956–1962) and 1990s (1989–2005) after stable conditions between the 1830s and 1930s, which follow extreme wet and dry periods between the late 1600s to the early 1800s (Figure S3b). The observed precipitation variability is within the

**TABLE 3** Calibration and verification statistics of the transfer function model for the June–December precipitation sum reconstruction from the KET tree ring width chronology with observations from the Duchinjl station

|                      | Full calibration (1977–2012) | Calibration (1977–1995) | Verification (1996–2012) | Calibration (1996–2012) | Verification (1977–1995) |
|----------------------|-------------------------------|--------------------------|--------------------------|--------------------------|--------------------------|
| $r$                  | .642***                       | .711**                   | .520***                  | .520**                   | .711**                   |
| $R^2$                | .413***                       | .506*                    | .271**                   | .271**                   | .458                     |
| Adj$R^2$             | .395***                       | .477*                    | .222**                   |                          |                          |
| RE                   | .379                          | .506                     | .231                     | .271                     | .458                     |
| CE                   | .506                          | .193                     | .271                     | .442                     |
| Sign test            | 13+/6=                        | 10+/7=                   | 10+/7=                   | 15+/4=                   |
| Products means test  | 680**                         | 354                      | 250***                   | 392**                   |

Abbreviations: Adj$R^2$, adjusted for degrees of freedom; CE, coefficient of efficiency; $r$, Pearson's correlation coefficient; $R^2$, coefficient of determination; RE, reduction of error statistic.

*p < .05; **p < .01; ***p < .001.

precipitation data decreases when the longer period between 1950 and 2012 is considered (Figure 5c,d).

The reconstructed precipitation variability for the time period 1569 to 2012 shows recent wet maxima in the 1950s
uncertainty range (± standard error of the prediction) of the reconstructed precipitation over the common period (Figure S5).

In a graphical comparison with other precipitation reconstructions (Figure 7) from the northern and southern Altai Mountains, all reconstructions show similar wet (1915, 1956–1961, 1974–1976, and 1989–2006) and dry periods (1880–1885 and 1977–1985), although our sampling sites in the southern part of the Mongolian Altai Mountains are relatively dry due to the rain shadow probably caused by the central Mongolian Altai Mountains (Klinge et al., 2003) and the southern Chinese Altai Mountains. The differences in the high-frequency variation of the precipitation reconstructions might be explained by regional orographic effects of the Altai Mountains on the storm tracks bringing moisture to this region (Davi et al., 2009).

Zhang et al. (2018) found a significant increasing trend in precipitation in the southern Altai in the period 1966–2015 while precipitation decreased in the northern Altai. They explained these wetting conditions in the southern Altai by increased water vapour supplies from the Mediterranean Sea, Black Sea, and Aral Sea caused by an increasing westerly and southerly circulation. Also, Shi et al. (2006) noted that evaporation from the Indian Ocean was high during the 1990s—the warmest decade of the last 1,000 years—at that time—thus the northwards water vapour transport increased from the Indian Ocean through the Arabian Sea to northwestern China in the 1980s and 1990s. Moreover, Chen et al. (2014) suggested that an increased strength of westerlies related to a warming of sea surface temperatures over the North Atlantic and the Indo-West Pacific Ocean led to the wetting trend over the southern Altai in northwest China in the 1980s, whereas Zhao et al. (2014) concluded that the position of the westerly jet is more important than its strength for an increase of summer precipitation in this region as a more southern position brings warm and moist air across the Indian subcontinent. Moreover, Wei et al. (2017) found that the weak Indian summer monsoon made a southeasterwards shift of the Asian westerly jet through a southeasterwards shift of the intense and persistent South Asian anticyclone, which produces rising air and thus more rainfall over central Asia and downwelling with less rainfall over northern China. Thus, the position and strength of westerly jet defines the moisture variability over the Altai-Dzungarian region.

Our June through December precipitation sum reconstruction (Figure 7) shows a rapid increase in the 1980s but a gradual decrease since 1998. The latter decrease might be related to a re-strengthening of the Siberian High in last two decades (Jeong et al., 2011), accompanied by an increase of Eurasian winter snow cover since 2000 (Estilow et al., 2015). The strengthening of the Siberian High might reduce southerly and westerly winds over this region.

The extreme moisture variations observed here, with the wettest period 1739–1746 and the driest period 1755–1761 between the late 1600s to the early 1800s, are supported by Davi et al. (2009), who reconstructed the June–September PDSI from 1565 to 2004 in the northwestern Mongolian Altai. They found influences of solar activity, and ENSO on the moisture variation in a spectral analysis of the PDSI reconstruction.

To demonstrate the relationships between the multidecadal variability of the monthly and seasonal NAO, AO...
and NINO4 indices and our estimated June through December precipitation sum, all time series are smoothed with a running 25-year averaging window (Figure 8a). The thus filtered estimated precipitation was positively correlated with (a) the current year mean February–April NAO index over the period 1837–2000 ($r = .51, p < .001$), (b) the previous and current year mean January–March AO index for the more recent period 1962–2000 ($r = .57, p < .001$, $r = .50$, $p < .001$, respectively), and (c) the previous and current year mean June–December NINO4 for the period 1866–2000 ($r = .53, p < .001$, $r = .48, p < .001$, respectively). For the period 1962–2000, the NAO index was strongly correlated with the AO index ($r = .95, p < .001$) and the AO index with NINO4 ($r = .83, p < .001$). For the much longer period 1866–2000 the NAO index was moderately correlated with the NINO4 index ($r = .48, p < .001$). Thus, in general, all climate indices positively correlated with precipitation over the Altai-Dzungarian region on the decadal timescale.

Interdecadal relationships between the climate indices and the estimated precipitation over the Altai-Dzungarian region were analysed from their 25-year centred moving correlations (Figure 8b). Obviously, the linkages between February–April NAO and June–December NINO4 with the estimated precipitation varied during between the mid-19th century until today. The impacts of NAO and NINO4 on the precipitation were weak till the late 1800s, strengthened during the change of the century and weakened again till the late 1940s. From about the period 1964–1988 correlations changed from $-0.33$ for NAO and $-0.52$ for NINO4 to $+0.47$ for NAO and $+0.37$ for NINO4 towards the beginning of the 21st century in 2000, which coincides with the wetting trend in estimated precipitation since about 1975 (Figure 8a). The impact of January–March AO index on regional precipitation (Figure 8b) started weak in the beginning of the 1970th, strengthened till a peak in the mid-1980th ($r = .55–.57$ during 1972–2001), and slowed down again towards the change to the 21th century ($r = .21$ during 1988–2012). Thus, the enhanced rainfall over the Altai-Dzungarian region between 1988 and 2005 might have been caused by a combined effect of a positive spring NAO and periods of positive winter AO during that period. This interpretation is supported by Visbeck et al. (2001), who connect a positive NAO with stronger than average westerlies across the mid-latitudes, and by He et al. (2017), who state in their review of the AO on the East Asian climate, that a positive AO concurrent with El Niño could reinforce anomalous mid-latitude westerlies, and that a positive spring AO is usually followed by positive summer precipitation anomalies in the southern China and western Pacific.

**FIGURE 8** Twenty-five-year low-pass-filtered time series of estimated June through December sum precipitation (black line), mean June–December NINO4 index (dark grey), mean January–March AO index (dashed black line), and mean February–April NAO index (light grey) over the period 1825–2012 (a); 25-year centred moving correlation coefficients of the estimated precipitation with mean June–December NINO4 index (dark grey), mean January–March AO index (dashed black line), and mean February–April NAO index (light grey) over the same period (b).
Ju et al. (2005) found a trend in the winter AO towards a high-index polarity after the late 1970s, which might have promoted the summer rainfall anomalies changing from below normal to above normal in central China, the southern part of northeast China, and the Korean Peninsula around 1978. He et al. (2017) also pointed out that a positive AO induces a northwards shift of the polar jet and intensifies the subtropical westerly jet over the northern Tibetan Plateau. Moreover, a positive winter AO leads to warmer conditions over East Asia through its impact on the Siberian High by weakening the northerly winds from Siberia for the period from 1980 to 1992 (He et al., 2017). Gong and Ho (2002) related the warming and wetting trend across continental Asia from the late 1970s to the 1990s—especially during winter—to weakening northerly and northeasterly winds over East Asia caused by a weaker dry and cold Siberian High in parallel with a pressure reduction over the Atlantic Ocean and high- to mid-latitude Asia.

### 3.4 Long-term climatic variation and change

We inferred climate variations for a 433-year period over the Altai-Dzungarian region and combined them in 20-year smoothed reconstructed temperature and precipitation series shown as standardized anomalies in Figure 9. The inferred climate conditions were classified into cool-dry, cool-moist, warm-dry, and warm-moist periods (Table S7). This classification suggests that cool-moist and cool-dry years have both decreased by about 8% since 1875, while warm-dry and warm-moist years have increased by about 14 and 2%, respectively. Thus, the more frequent warm-dry summers have replaced the cool/warm-moist and cool-dry episodes more common during the LIA (1580–1874). We defined the periods extreme cool (1689–1691, 1705–1715, 1838–1859, 1983–1992), cool-dry (1692–1704, 1716), dry (1632–1647, 1717–1721, 1760–1768, 1820–1835), warm (1655–1664, 1889–1909, 1948–1962, 2008–2011), warm-dry (1648–1654, 2012), moist (1602–1616, 1677–1678, 1734–1749).

**FIGURE 9** Inferred climate variation from the 20-year smoothed lines of the estimated temperature (solid line) and precipitation (dashed line) time series, expressed in standardized and normalized values. The horizontal dotted and thin grey lines indicate the range of one standard deviation for the 20-year smoothed precipitation ($\sigma = \pm .29$) and temperature ($\sigma = \pm .37$), respectively.
1789–1810), and warm-moist (1665–1676) by the anomalies of the estimated temperature and precipitation above or below the ±one standard deviation range of the whole record.

The cool-moist, cool-dry, warm-moist, and warm-dry periods can be linked with different states of the general circulation discussed in sections 3.1 and 3.2. The LIA period was cooler and wetter than the present climate condition in the Altai-Dzungarian region, which is supported by Putnam et al. (2016) who inferred wetter climate conditions during the LIA (which they defined as the period between 1150 and 1845) from geomorphological, biological, and historical evidence in the Tarim Basin, which neighbour the Dzungarian Basin. During that period, the Northern Hemisphere mountain glaciers expanded as a reaction to the lower temperatures and descending snowlines. In parallel, the Tarim Basin became wetter with deeper snowpacks over the high mountains due to increasing orographic precipitation as a result of a southwards shift or a strengthening of the boreal westerlies impinging the interior Asian desert belt. Moreover, the recent cool-moist period from 1985 to 2000 (Table S7) has been related to the AO (this study, Robock, 1984, He et al., 2017). The recent cooling could have been caused by volcanic aerosols of the El Chichón eruption (VEI5, 1982) in southern Mexico, which impacted atmospheric wind patterns, including a positive phase of the AO (Robock, 1984). No large volcanic-induced cooling was observed at this time due to the simultaneous warming ocean temperature caused by El Niño (Robock, 2002). Also, the positive AO competing with El Niño could reinforce the anomalous westerlies in the mid-latitudes (He et al., 2017). During this recent cool-moist period, ice mass accumulation of the glaciers in the Russian Altai Mountains was observed and Narozhniy and Zemtsov (2011) connected this phenomenon to annual precipitation increased by 8–10% especially in winter and spring (April–May) as a result of a strengthening of the zonal circulation over the Altai Mountains.

4 | SUMMARY AND CONCLUSIONS

We reconstructed the longest time series of temperature (611 years) and precipitation variability (444 years) thus far for the Altai-Dzungarian region. On long timescales, most cool summers were observed during the LIA when solar activity was low and intense volcanic eruptions occurred (see also Buentgen et al., 2016). The observed precipitation variability seems to be related to the position and strength of westerlies linked to the Arctic and NAO. Our temperature reconstruction is representative for a region encompassing broader parts of northwestern Mongolia, while the area represented by the precipitation reconstructions is confined to the Altai-Dzungarian region due to local influences of orography on precipitation. According to the reconstructed combined variability of temperature and precipitation over the past 433 years, the Altai-Dzungarian region was cooler and wetter during the LIA and warmer throughout the 20th century. This is in line with findings for the neighbouring Tarim Basin. A continuous 20th century warming trend was, however, not observed. A short combined cool and wet period was detected in the late 20th century, which was likely a result of volcanic-induced cooling and a positive phase of the AO promoting an intensified subtropical westerly jet and a positive summer rainfall anomaly over the Altai-Dzungarian region.

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AUTHOR CONTRIBUTIONS

B.O. carried out the fieldwork with the assistance of N.S., A.B., S.G., and C.S. Laboratory analysis and interpretation of the results was carried out with the help of S.W., J.L., and S.B. The manuscript was prepared and compiled by B.O., S.W., and C.S. All authors thoroughly reviewed the manuscript and significantly contributed with comments and additions.

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**SUPPORTING INFORMATION**

Additional supporting information may be found online in the Supporting Information section at the end of this article.