**Regional drought shifts (1710-2010) in East Central Asia and linkages
with atmospheric circulation recorded in tree-ring $\delta^{18}\text{O}$**

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Abstract Drought occurrence and duration in central Asia are of important socioeconomic, ecological, and geophysical significance and have received increasing research attention in recent years. Understanding long-term drought trends and their driving forces require reliable records of past drought variability with broad spatial representativeness. Here, we compiled four tree-ring $\delta^{18}\text{O}$ records from eastern central Asia (ECA) and composited them into a drought-sensitive proxy to explore regional ECA moisture variations over the past 301 years (1710-2010 CE). A robust regional standardized precipitation-evapotranspiration index (SPEI) reconstruction was established based on the tree-ring cellulose $\delta^{18}\text{O}$ fractionation mechanism and statistically significant proxy-climate relationships. We identified prominent droughts in 1710-1770, 1810-1830, and the beginning of the 21st century, and a regime shift to a persistently wet period from the 1880s to 2000. Our reconstruction reveals the impact of drought and pluvial patterns on the decline of Zhungar Empire, and on historical agricultural and socio-economical activities, including increased migration into ECA during the 1770-1800 pluvial. Our findings also suggest that wet conditions in the 20th century in ECA were related to a strengthening of the westerly circulation and thus shed light on large-scale atmospheric circulation dynamics in central Asia.

Keywords Tree rings, stable oxygen isotopes, drought reconstruction, eastern central Asia, SPEI

1 Introduction

Drought can have disastrous impacts on society, economy, and ecosystems. An enhanced understanding of drought variability is therefore crucial for sustainable development in arid regions ([Sheffield et al., 2012](#)). To improve models of the evolution of hydroclimate variability in a gradually warming world, long and reliable reconstructions of past droughts are needed. Eastern Central Asia (ECA) is such an arid region, which includes China's Tianshan and Altay Mountains and is an oasis for farming and herding cattle in northwestern China ([Qin et al. 2005](#)). Several historic

cities (e.g., Turpan, Urumqi, and Ili) are located in ECA along the Silk Road and have played an important role in China's and the world's development strategy, such as the Silk Road economic belt ([China Daily, 4 March 2017](#)). ECA thus serves as a political and geographical corridor connecting central Asia, China, and Europe. The dry climate, decreasing water resource (e.g. shrinking of glaciers), and fragile ecological and agricultural ecosystems make ECA communities strongly vulnerable to changes in moisture condition and drought ([Qin et al., 2005](#); [Shi 2006](#); [Sorg et al., 2012](#)). Thus, there is a crucial need for an improved understanding of ECA drought frequencies and intensities and their driving forces.

Moisture over the ECA region originates mainly from westerly circulation trajectories because the high mountain systems of the Himalayas and Tibetan Plateau in the South block the movement of air masses from the Indian Ocean ([Rashed et al. 2010](#); [Lee and Zhang 2011](#)). The North Atlantic Oscillation (NAO) can be considered a representation of the strength of westerly winds at different time scales (e.g. monthly, seasonal, and annual) ([Jones et al. 1997](#); [Hurrell et al. 2003](#); [Li and Wang 2003](#); [Pinto and Raible 2012](#)) and NAO variability influences ECA moisture availability in summer ([Li et al. 2008](#); [Rashed et al. 2010](#); [Lee and Zhang 2011](#)). Owing to the location of ECA in the transition zone between the Mediterranean and Asian monsoon climates, a complex interaction between the westerlies and monsoon drives the pattern and variability of moisture in the region ([Cai et al. 2017](#)). Moisture variability during the Holocene in ECA has been dominated by westerlies and has experienced synchronous and coherent changes from dry, wet to moderately moist ([Chen et al. 2008](#); [Cai et al. 2017](#)). The history of moisture over the last millennium in arid ECA is out-of-phase with that of monsoonal China ([Chen et al. 2010](#)). The ECA has experienced a dry Medieval Climate Anomaly and a wet Little Ice Age ([Chen et al. 2010](#)). A recent moisture increasing has also been reported ([Chen et al. 2010](#); [Xu et al. 2015](#)). Therefore, to explore the moisture shift and its potential driving forces, it is important to explain pre-industrial and modern regional climate. However, it is not

fully clear whether and when this modern moisture shift pattern exists in ECA at a regional scale and whether the moisture pattern differs from that of the monsoon area or not. In addition, the linkage between moisture in ECA and the westerlies and whether it is consistent through time over such a large area should be further explored.

Stable oxygen isotopes ($\delta^{18}\text{O}$) from tree rings have been used in Asia to reconstruct past precipitation (Treydte et al. 2006; Xu et al. 2013), cloud cover (Shi et al. 2012; Liu et al. 2014), drought variability (Li et al. 2011; Xu et al. 2015), and atmospheric circulation (Xu et al. 2017). One of the advantages of tree-ring $\delta^{18}\text{O}$ measurements is that – unlike tree-ring width proxies – they do not show an age trend, are less influenced by juvenile effects and can thus better retain low-frequency climate signals (Treydte et al. 2006; Esper et al. 2010; Kress et al. 2010; Young et al. 2011; Gessler et al. 2014). Tree-ring $\delta^{18}\text{O}$ can portray uniform climate signals at large spatial scales regardless of tree species and site characteristics (Treydte et al. 2007; Saurer et al. 2008, 2012; Hartl-Meier et al. 2014; Liu et al. 2014; Labuhn et al., 2016; Xu et al. 2017). In the ECA region, tree-ring $\delta^{18}\text{O}$ has previously been used to study local drought variability and moisture shift (Xu et al. 2014a, b, c; 2015), but its potential driving forces have not previously been investigated at the broad regional scale. We here combine tree-ring $\delta^{18}\text{O}$ series from four ECA sites (Xu et al. 2014a, b, c; 2015) to study regional drought variability over the last 301 years focusing on the regional common character and to explore its association with NAO variability. We also investigate why recent ECA moisture variability is out-of-phase with that of the south Asian monsoonal area.

2 Data and Methods

2.1 Regional chronology construction

We compiled four strongly coherent (Fig. S1a), previously published tree-ring $\delta^{18}\text{O}$ chronologies of spruce (*Picea schrenkiana*) and larch (*Larix sibirica* Ldb) from upper

treeline sites in ECA (Xu et al. 2014a, b, c; 2015) into a regional chronology (Table 1). Inter-series correlations were calculated between the $\delta^{18}\text{O}$ series of the four sites over the common period (1790-2008 CE; Table 1). Running mean series inter-correlation (Rbar) for these tree-ring $\delta^{18}\text{O}$ series were also calculated using two time-windows: 51 years with 25 years lag and 21 years with 10 years lag (Fig. S1b). Significant ($p < 0.05$) inter-series correlations between the four individual $\delta^{18}\text{O}$ series ($r \geq 0.34$; Table 1) and significant and stable Rbar values ($r > 0.28$; Fig. S1b) suggest common signal strength and correlations that are stable through time. Previous research showed that these tree-ring $\delta^{18}\text{O}$ series mainly record growing season moisture variability (drought or relative humidity) (Xu et al. 2014a, b, c; 2015). We emphasize the common signal in the four $\delta^{18}\text{O}$ -series by combining them into a composite $\delta^{18}\text{O}$ chronology and assess the regional climatic signal. We calculated Z-scores of each tree-ring $\delta^{18}\text{O}$ series using the mean and standard deviation over the reference and common period (1790-2008 CE) prior to averaging them (at least two series) in order to avoid possible artefacts due to different amplitudes of the individual series (Liu et al. 2014). We further applied principal component analysis (PCA; Fig. S1c) to extract the primary common signals in the four standardized tree ring $\delta^{18}\text{O}$ series as a verification series to check the strength of the common signal in the regional chronology.

2.2 Climate dataset and normalized difference vegetation index (NDVI)

We used monthly gridded temperature, precipitation, and the 4-month scale standardized precipitation-evaporation index (SPEI) from the CRU TS 3.22 dataset with grid size $0.5^\circ \times 0.5^\circ$ (Vicente-Serrano et al. 2010; Harris et al. 2014) to assess the trends of climate in our study region and then to explore climatic signals in the tree-ring $\delta^{18}\text{O}$ series. We also used the climate data (temperature and precipitation; Chinese meteorological data center (<http://www.cma.gov.cn/2011qx fw/2011qsjg x/>)) from meteorological stations in our study area to explore the inter-annual variability for temperature and precipitation. SPEI accounts for the combined effects of precipitation and potential evaporation (Vicente-Serrano et al. 2010) and has shown a better

capacity to detect drought impacts on a suite of hydrological, agricultural, and ecological variables at regional and global scales than drought indices based only on precipitation (such as the standardized precipitation index, [Vicente-Serrano et al. 2012](#)). In addition, SPEI has the advantage of different time scales (e.g., 2-month and 4-month scales) compared to PDSI and self-calibrated PDSI ([Vicente-Serrano et al. 2010](#)). We only used climate data for the period 1957-2010, because earlier data (1901-1956) are interpolated between sparse station observations, are therefore adjusted towards their climatologically “normal” value, and thus are possibly biased in our study area from undercatch of snow by raingauges or application of adjustments to compensate for these potential inhomogeneities ([Harris et al. 2014](#)). Monthly values of the NAO index from 1821 to 2010 ([Jones et al. 1997](#); data from CRU <http://www.cru.uea.ac.uk/cru/data/nao/>) and a reconstructed monthly NAO index (from 1710 to 2000 CE; [Luterbacher et al. 1999](#)) were also used for comparison. The National Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) data (including precipitation rate, water flux, and wind data) ([Kalnay et al. 1996](#); data from <https://www.esrl.noaa.gov/psd/data/gridded/reanalysis/>) were used to analyze the linkage between the climate reconstruction and potential climate driving forces.

To evaluate the influence of drought on ecosystems, we used the normalized difference vegetation index (NDVI) to represent the status of the ecosystem. Monthly mean NDVI data were derived from the Advanced Very High Resolution Radiometer (AVHRR; <https://earthexplorer.usgs.gov/>) with a 250 m spatial resolution and a 16-day temporal resolution ([Pettorelli et al. 2005](#)). The data for the period of 1981-2006 were retrieved for the study region from the KNMI climate explorer (<http://climexp.knmi.nl>) and data for the period of 2007-2010 were calculated directly from the AVHRR data set. We compared the NDVI value derived from KNMI to the calculated NDVI values to make sure they had similar mean values. NDVI values range from -1 to 1, with values below zero indicating land covered by water, snow, or

clouds, and values above zero indicating land covered by vegetation (Pettorelli et al. 2005). Higher NDVI values represent higher vegetation cover and high vegetation productivity (Pettorelli et al. 2005). We calculated monthly mean values for each pixel over the study region with NDVI >0 to reflect the status of the vegetation (Pettorelli et al. 2005).

2.3 Regional tree-ring $\delta^{18}\text{O}$ response to climate and climate reconstruction

We conducted spatial correlation analysis using the KNMI Climate Explorer (<http://www.knmi.nl/>; Trouet and Van Oldenborgh 2013) to evaluate the spatial representativeness of the climate signal in the regional tree-ring $\delta^{18}\text{O}$ series (Fig. 1). We evaluated the relationship between the regional tree-ring $\delta^{18}\text{O}$ chronology and monthly climate variables from September of the previous year to September of the current year.

For the development of our reconstruction model, we established a linear regression transfer function using the regional tree-ring $\delta^{18}\text{O}$ chronology as independent variable and March to September SPEI as dependent variable based on a reasonable relationship between tree-ring $\delta^{18}\text{O}$ and SPEI both in terms of mechanisms of stable oxygen fractionation and tree-ring $\delta^{18}\text{O}$ and climatic responses. We used a cross-validation method for calibration and verification, to test the temporal stability of the climatic signal. The reduction of error (RE), the coefficient of efficiency (CE), Pearson's correlation (r), and the sign test (Cook and Kairiukstis 1990) were used to evaluate the skill of the reconstruction.

Regression-based climate reconstructions often introduce biases in the variability of past climate (Esper et al. 2005; McCarroll et al. 2015), because of the imperfect correlation between climate proxies and the target climate parameters, leaving a proportion of the variance in the proxy unexplained (McCarroll et al. 2015). It is difficult to examine the ability of a reconstruction to replicate the measured range of climate, but Briffa et al. (2013) proposed to use a “scaled reconstruction”, where a

proxy/climate relationship can be shown to be strong and consistent through time. The advantage of this method is that the loss of variance can be removed by scaling the proxy data to the same mean and variance as the climate data over their common period, albeit at the cost of an increase in mean squared error of the reconstruction (McCarroll et al. 2015). Here, we used the equivalent variance explained (R^2_{vs}) (McCarroll et al. 2015) to evaluate the validity of our regional SPEI reconstruction.

2.4 SPEI variability and its drivers

We compared and validated our reconstruction with the Monsoon Asia Drought Atlas Palmer Drought Severity Index (MADA PDSI, Cook et al. 2010), a summer (June-August) drought reconstruction (1300-2005 CE) based on an Asia-wide tree-ring width network. We averaged the MADA PDSI grid points for the ECA (Fig. 1). We applied a spline filter (Cook and Peters 1981) to the regional MADA PDSI reconstruction and defined “30-year spline low-pass filter” as low-frequency and “30-year spline high-pass filter” as high-frequency. We also applied ensemble empirical mode decomposition (EEMD) analysis (Wu and Huang 2009), which is a method of temporal and spatial analysis based on empirical mode decomposition (EMD) analysis (Huang et al. 1999). The goal of EEMD is to sift an ensemble of signals with white noise added and with the mean treated as the final true result. The white noise is averaged out over a sufficient number of trials and the only persistent part that survives is the signal, which is then treated as the true and physically meaningful answer (Wu and Huang 2009). The added white noise provides a uniform reference frame in the time-frequency space; therefore, the added noise collates the portion of the signal of comparable scale in one intrinsic-mode function (IMF). With EEMD analysis, we can decompose nonlinear and non-stationary time series into their intrinsic modes of oscillation naturally without the *a priori* selection of any subjective criterion (Wu and Huang 2009). The IMFs decomposed by the EEMD represent the signals at different time scales (Wu and Huang 2009). The relative contribution of each IMF was calculated by means of regression analysis between IMF and the original series. We

also conducted spatial correlation analysis between our reconstruction and the March-September self-calibrated PDSI index (CRU scPDSI and UCAR PDSI) for the periods 1957-2010 and 1900-2010 to verify our reconstruction.

We used composite analysis to detect the potential dependency of the ECA SPEI reconstruction on NAO. We composited the 9 most positive and the 9 most negative summer (June-July-August) averaged NAO values (accounting about 16% of 1957-2011) and calculated precipitation and wind pattern averages for the positive minus the negative NAO years. We further analyzed the periodicity of our reconstruction and its potential climatic drivers using spectral analysis (Schulz and Mudelsee 2002) and cross-wavelet coherence analyses based on Monte Carlo tests (Grinsted et al. 2004). We compared the ECA SPEI reconstruction to monthly values of the NAO index and the reconstructed NAO index (Jones et al. 1997; Luterbacher et al. 1999). We also compared the ECA SPEI reconstruction to other moisture-sensitive reconstructions close to our study region (Shao et al. 2005; Treydte et al. 2006; Davi et al. 2013; Yang et al. 2014) using the two independent sample *t*-test and empirical distribution function to explore the moisture mean status of two periods and to detect the potential regional climate driving forces.

3 Results

3.1 Regional tree-ring $\delta^{18}\text{O}$ response to climate

A PCA analysis revealed that the four ECA $\delta^{18}\text{O}$ chronologies have 54% of variance in common – reflected by the first principal component score (PC1 with an eigenvalue of 2.14, the eigenvalue of the other PCs were lower than 1; Fig. S1c), and we thus used the regionally composited $\delta^{18}\text{O}$ series to detect regional climate signals. The March-September climate in our study region showed consistent trends: SPEI and temperature have increased and precipitation and drought showed a slight increase over the period 1957-2010 (Fig. 1a and Fig. S2). March-September temperature and precipitation also showed a coherent variability at annual scales (Fig. S3). Regional

tree-ring $\delta^{18}\text{O}$ is significantly ($p < 0.05$) positively correlated with the mean temperature from March to May, July, August, and the mean value from March to September (Fig. 2). More importantly, there are significant negative correlations between tree-ring $\delta^{18}\text{O}$ and precipitation in February, March, June, July, and March-September, and SPEI from March to September (Fig. 2). However, the regional tree-ring $\delta^{18}\text{O}$ record showed no significant correlation with the climate in the late growing season of the previous year (Fig. 2).

The individual tree-ring $\delta^{18}\text{O}$ series correlated significantly with the March to September SPEI over a large area surrounding the sampling site (Fig S4). The regional tree-ring $\delta^{18}\text{O}$ chronology was also strongly and significantly negatively correlated with regional March-September SPEI ($r = -0.70$; 1957-2010; $p < 0.01$; Fig. 2) and with a large west-east oriented SPEI region ($|r| > 0.20$, $p < 0.05$; Fig. 1b). The former result is confirmed by the significant negative correlation between the PC1 of the four tree-ring $\delta^{18}\text{O}$ series with the March-September SPEI ($r = -0.71$; 1957-2010; $p < 0.01$; Fig. S1d). The latter reflects the silk road pattern (Zhang and Zhou 2015) and is concentrated in nearby central Asia and northern Mongolia (Fig. 1).

3.2 SPEI reconstruction

We developed a 301-year long (1710-2010) SPEI reconstruction, based on the regional tree-ring $\delta^{18}\text{O}$ record, which explained 48.2% of the total variance of the March-September SPEI (1957-2010) during the calibration and verification procedure (Table 2; Fig. S5). The transfer function was statistically significant with a F value of 50.3 ($p < 0.001$). The residuals of the regression model showed a normal distribution (Shapiro-Wilk value of 0.99, $p = 0.88$) and did not vary significantly with the tree-ring $\delta^{18}\text{O}$ values, suggesting that the model had significant homogeneity of variance.

The results of cross-validation suggest that the model has good skill for reconstructing regional March-September SPEI variability (Table 2). The RE and CE values were positive and high for both verification periods (1957-1984 and 1985-2010), and the Pearson's correlation coefficients ($r = 0.80$ to 0.65) were significant ($p < 0.01$)

over the calibration and verification periods. The sign test indicated that the reconstruction was significantly ($p < 0.05$) in agreement with the CRU SPEI data during these calibration and verification periods (Table 2)

The variance scaled reconstruction showed a high value of equivalent explained variance ($R^2_{vs} = 0.40$) suggesting a limited (8.2%) loss of explained variance of the regression reconstruction for the calibration period. The Pearson's correlation coefficient between the variance scaled-reconstruction and the target is also strongly significant ($r_{vs} = 0.63$; $p < 0.001$) (Fig. S6). The variance-scaled-reconstruction, however, shows a higher increased mean squared error (MSE) of 0.23 compared to 0.18 for the regression-based reconstruction. Therefore, we used the regression-based reconstruction with minimum MSE, which captures regional SPEI variability well (Fig. S5b).

3.3 Reconstruction validation

Our reconstruction showed significant ($p < 0.05$) positive correlations with the MADA PDSI reconstruction grid points within the ECA region (Fig. S7a). At high frequency, our ECA reconstruction correlates significantly positively with the fully independent, averaged MADA PDSI time series in the region ($r = 0.34$, $p < 0.001$, 1710-2005; Fig. 3). This is confirmed by the EEMD analysis (e.g. IMF1 to IMF3) (Fig. S7b). Our reconstruction is also positively correlated with the average MADA PDSI series after applying 30-year spline low-pass filter ($r = 0.27$), but shows more variance at these low frequencies (e.g. IMF4 to IMF5) (Fig. S7b). We also detected some anti-phase between our reconstruction and the MADA reconstruction in the low frequencies (e.g., the early part of IMF 4 and IMF 5-6) but with little ($< 10\%$) contribution of the original climate signals (Fig. S7b). These divergent signals may result from the different climatic parameters or season of the reconstructions (e.g., March-September in our reconstruction versus June-August in the MADA reconstruction). Spatial correlation analysis indicated that our SPEI reconstruction was significantly correlated with the March-September CRU scPDSI and UCAR scPDSI

(Fig. S8) for the periods 1957-2010 and 1900-2010. The significant correlation pattern is similar to the tree-ring $\delta^{18}\text{O}$ spatial response (Fig. 1).

Our reconstruction indicates that the 20th century in the ECA is the wettest century over the past 301 years and that reconstructed SPEI values for the period 1880-2000 are consistently above average (Fig. 3b). Another wet period occurred in the late 18th century (1770-1800). Our reconstruction further reveals three drought periods from 1710 to 1770, 1810 to 1830, and the first decade of the 21st century, the first two of which are also recorded in the MADA PDSI reconstruction (Fig. 3).

Past drought conditions appear to have been caused by high temperatures. For example, the early part of the three drought periods (1710-1730, 1810-1830, and 2000-2010) correspond to warm periods reported from an independent Tianshan Mountains summer temperature reconstruction (Yu et al. 2013) (Fig. S9a, b, c). Over the instrumental period (1901-2010), our reconstruction was significantly negatively correlated with regional CRU TS3.22 March-September temperature and this correlation was stronger for dry years compared to wet years (Fig. S9d). For the 21st century drought, 9 out of 11 years were also years with high temperature (Fig. S9d).

ECA NDVI values were significantly positively correlated with March-September SPEI for the period of overlap (1982-2010; $r = 0.43$, $p < 0.05$; and $r = 0.39$, $p < 0.05$ after linear detrending) and showed a slightly decreasing trend since 1982 (Fig. S9e). The decreasing trend in the ECA NDVI values was stronger after 1994 when SPEI also showed a decreasing trend (Fig. S9e).

3.4 Periodicity of drought variability and potential linkages with atmospheric circulation patterns

A composite analysis indicated that ECA March-September precipitation and surface precipitation rates were higher during positive NAO years compared to negative years (Fig. 4a, b; Table S1). The water flux at 500 hPa and 850 hPa was also higher in ECA during positive NAO years compared to negative NAO years, while

the wind flux at 500 hPa and 850 hPa was more westerly and northerly (Fig. 4c, d; Fig. 10; Table S1). The large water flux area was broad in positive NAO years (Fig. S10).

The correlation between the ECA SPEI reconstruction and monthly NAO values is strongest during July (Fig. S11). The SPEI reconstruction shows a statistically significant positive correlation with the instrumental July NAO index (Jones et al. 1997) ($r = 0.29$, $n = 191$, $p < 0.001$) and with a reconstructed July NAO index (Luterbacher et al. 1999) ($r = 0.13$, $n = 301$, $p = 0.02$) (Fig. 5a; Fig. S11), which indicates that positive July NAO phases correspond to wet ECA summers. The July NAO index also shows a positive relationship with the SPEI reconstruction after 8-year low-pass filtering, which retains the 8-year cyclicity of the NAO index (Fig. 5b). The influence of NAO on ECA drought variability was primarily positive over the last 301 years, but was strongest after 1930 approximately (Fig. 5c).

A spectral analysis of our ECA SPEI reconstruction indicated quasi-2, -3, -4, -8, -11, -17 and -33-year cycles of drought (Fig. 5d). A wavelet coherence analysis (Fig. 5e; Fig. S12) indicated that the SPEI reconstruction was in phase with the July NAO index during most periods and significantly for the 2-year, 4-year, 8-year, and 11-year cycles and with the reconstructed July NAO index for the 16~33-year cycles (Fig. S12). At the 33-year cycle, SPEI and NAO variability were in phase from 1860 to 1910 and from 1950 to 2010, but this result was not statistically significant (Fig. 5e). Wavelet coherence between the two series was weak at decadal time scales from 1890 to 1930 (Fig. 5), a period during which the 31-year running correlation coefficients between the series were not significant (Fig. 5c).

4 Discussion

ECA tree-ring $\delta^{18}\text{O}$ was significantly positively correlated with temperature and significantly negatively correlated with precipitation and SPEI for the months March to September (Fig. 1; Fig. 2) and was less affected by the climate of the previous year (Fig. 2). This result is consistent with previous studies in this region (Xu et al. 2014a, b;

2015) and is also supported by a study at intra- and inter-decadal time scales (Zeng et al. 2017). The negative association between tree-ring $\delta^{18}\text{O}$ and March-September SPEI in ECA is consistent with previous studies for individual sites in this region (Xu et al. 2014a; 2015) and Europe (Labuhn et al. 2016). According to mechanistic tree-ring $\delta^{18}\text{O}$ models (Roden et al. 2000; Waterhouse et al. 2002) and empiric studies (Gessler et al. 2014; Treydte et al. 2014), tree-ring $\delta^{18}\text{O}$ is primarily affected by relative air humidity and $\delta^{18}\text{O}$ of source water (Zeng et al. 2017). Relative air humidity was negatively correlated with tree-ring $\delta^{18}\text{O}$ through strongly controlling the evaporative enrichment of $\delta^{18}\text{O}$ in leaf water (Roden et al. 2000; Waterhouse et al. 2002). Precipitation before and during the growing season affects soil water availability and relative air humidity. This holds particularly in arid areas (Xu et al. 2015) that in turn determine the degree of isotopic evaporation enrichment in leaf water (Roden et al. 2000) and therefore affect tree-ring $\delta^{18}\text{O}$. In arid and semiarid regions, temperature indirectly controls tree-ring $\delta^{18}\text{O}$ by affecting the moisture conditions (e.g., evaporation and transpiration) (Xu et al. 2015). Specifically, tree-ring ^{18}O enrichment in ECA during dry periods was partially caused by strong transpiration driven by high temperatures, which resulted in high $\delta^{18}\text{O}$ of leaf sugars and thus high $\delta^{18}\text{O}$ of tree-ring cellulose. In addition, $\delta^{18}\text{O}$ in soil water is also enriched, because of the stronger evaporative enrichment of water in shallow soil layers during dry periods (Treydte et al. 2014). As a consequence, ^{18}O enrichment in leaf sugar and soil water leads to higher tree-ring $\delta^{18}\text{O}$ values during dry periods. The reverse is the case during wet periods, when low transpiration results in low $\delta^{18}\text{O}$ of leaf sugars, soil water $\delta^{18}\text{O}$, and then tree-ring cellulose. Tree-ring $\delta^{18}\text{O}$ reflects the effects of a combination of relative air humidity, precipitation (negative), and temperature (positive) (Fig. 2), and therefore of drought (Xu et al. 2014a; 2015; Labuhn et al. 2016). This explains the significant correlation of the regional tree-ring $\delta^{18}\text{O}$ chronology with SPEI (Fig. 2).

Three decadal-scale drought periods in our reconstruction (1710-1770, 1810-1830, and 2000-2011) may have resulted from high temperatures, especially for

the earliest of the dry periods (Fig. S9), which likely led to strong evaporation and low soil moisture. The early 21st century drought has also been reported in Mongolia, where it had a disturbing effect on economic and ecological systems (Davi et al. 2013; Pederson et al., 2014) and may have been caused by anomalously high temperatures during this period in ECA (Fig. S9d). Decreased ECA NDVI values (Fig. S9e) are a further indication of drought conditions in central Asia (Yin et al. 2016). This early 21st century drought, however, was not unprecedented over the last 312 years (Fig. 3a). Future climate model projections under representative concentration pathway RCP8.5 suggest that drought risk in currently dry regions will likely further increase by the end of the 21st century because of regional scale temperature increases and soil moisture decreases (IPCC 2013).

The 1810-1830 drought was a severe drought in the ECA region and it also occurred in the northeastern Tibetan Plateau (Shao et al. 2005; Yang et al. 2014) and far-western Mongolia (Davi et al. 2009) (Fig. 6a-d). It resulted in below average streamflow in the Manasi River in northwestern China (Yuan et al. 2007), but it was relatively mild in the ECA section of the MADA PDSI reconstruction (Fig. 3). This drought was damaging to agricultural and social activities in ECA. For example, the Chinese government exempted farmers in Urumqi, Balikun, and Hami from taxation from 1816 to 1828 and loaned them funds owing to poor drought-induced harvests (Shi 2006). The 1710-1770 drought in our reconstruction corresponds with a 1720-1770 dry period in northern Pakistan (Treydte et al. 2006) (Fig. 6e) and played a crucial role in the defeat of the ethnic minority empire by the Qing dynasty, which contributed to the unification of ancient China and collapse of the Zhungar Empire (Editors of Zhungar Shilue 1985). The drought caused the decline of livestock, diminished the economy of the ethnic minority and leads to the decadence of the Zhungar Empire (Perdue 2005).

The period from 1770 to 1800 was generally wet in our reconstruction and also in a tree-ring width based ECA PDSI reconstruction (Li et al. 2006). Favorable

climatic conditions during the late 18th century were suitable for agriculture and allowed the authority of the Qing dynasty to extend their agricultural area to the northern Tianshan and the Yili River valley (Ge et al. 2005). For instance, rice was cultivated in Urumqi (Ge et al. 2005), where the land was fertile and there was sufficient water for irrigation in most years (Li et al. 2010). The suitable agricultural conditions led to an immigration increase into the ECA region during this period with three immigration peaks (Li and Zhang. 2015) (Fig. 6a).

The ECA region was anomalously wet throughout the 20th century starting in the 1880s as indicated by a two-sample *t*-test and empirical distribution function and by a shift in 1886 towards wetter conditions (Fig. 6a). Several moisture-sensitive proxies near our study region also showed that the 20th century was an anomalously wet period, such as in the north part of Qaidam Basin (Shao et al. 2005), the northeastern Tibetan Plateau (Yang et al. 2014), northern Pakistan (Treydte et al. 2006), India (Yadav 2013), and Mongolia (Davi et al. 2009, 2013) (Fig. 6b-e). A regime shift from drier to wetter conditions occurred in each of these records in the 1860s-1880s. This suggests that this shift was driven, at least in part, by a change in large-scale atmospheric circulation patterns such as the westerly circulation.

The 2-3 years cycle of our ECA reconstruction (Fig. 5d) is similar to the band of quasi-biennial oscillation (QBO) variability (Randel et al. 1999), which is found in many climatic fields in stratospheric zonal wind and temperature (Randel et al. 1999). A similar 2-3-year cyclicity was reported for reconstructed drought variability in the central Tianshan mountains (Li et al. 2006) and for precipitation oscillation in the Xinjiang Province of China (Zhu and Zhi 1991), suggesting that ECA drought variability may be linked to tropical ocean-atmosphere systems such as the QBO.

The quasi-4, -8, -11, and 33-year cycles in our ECA SPEI reconstruction coincide with the cyclicity of the NAO (Fig. 5d; Fig. S12). Similar co-occurring oscillations between NAO and drought in China have been reported earlier (Fu and Zeng 2005). ECA moisture variability over the last centuries was linked to NAO variability at

decadal to multi-decadal time-scales (Fig. 5). In the ECA region, the atmospheric moisture flux in summer is mainly transported from the west, northwest, and north (Dai et al. 2006; Liu et al. 2009; Rashed et al. 2010) by the westerly circulation (Feng et al. 2004; Li et al. 2008). The NAO represents the strength and intensity of these westerlies well (Jones et al. 1997; Hurrell et al. 2003; Pinto and Raible 2012), and reconstructed ECA SPEI has been higher during positive NAO phase since 1700 compared to negative phases (Fig. 5; 6). This has also been suggested by rainwater isotopic evidence (Liu et al. 2009) and by ECA river runoff data (Yang et al. 2010). During positive summer NAO phases, mid-latitude westerlies are northward-displaced and stronger-than average (Fig. 4; S10), resulting in ECA water flux, an enhanced latitudinal water vapor gradient, and increased precipitation transported into the region (Li et al. 2008; Lee and Zhang 2011). During positive NAO summers, anomalous convergences in lower layers of the atmosphere and divergences in the upper layers of the atmosphere occur over the ECA region (Sun and Wang 2012) producing a downward action, which is favorable for rainfall. During negative NAO phases, water vapor and precipitation decrease as a result of the divergence of water flux (Li et al. 2008; Rashed et al. 2010), which results in increased drought frequency and intensity in ECA. Documentary records also suggest that historical drought disasters in northwestern China occurred during negative NAO phases (Lee and Zhang 2011). The positive (negative) summer NAO mode thus corresponds to wetter (drier) summer moisture conditions in the ECA region (Fig. S10), a finding that has also been reported for the eastern edge of the Tibetan Plateau, Mongolia, and northwestern China (Linderholm et al. 2013; Molg et al. 2014). This summer NAO teleconnection thus even links ECA hydroclimate variability to climate variability over northwestern Europe and the Balkans (Trouet et al. 2012, 2018; Linderholm et al. 2013).

Generally, more positive summer NAO conditions since the 1880s compared to prior (Fig. 6g) – reflective of a strengthening of the westerlies - are therefore likely also linked to the 1880s wet period in the ECA SPEI record. This wet period was also

found for the north Tibetan Plateau where glaciers were characterized by less retreat and a positive mass balance (Yao et al. 2012; Molg et al. 2014). The moisture in the north Tibetan Plateau forms a dipole with that in the southeast Tibetan Plateau (Zhang et al. 2015), where glaciers and moisture showed a decrease in the 20th century (Xu et al. 2012; Yao et al. 2012; Liu et al. 2014; Fig. 6f). This dipole, as well as the 1860s regime shift, have been attributed to a weakening of the Indian summer monsoon (Xu et al. 2012; Liu et al. 2014). This proxy comparison can be helpful to explain the dipole mode of moisture conditions and glacier status in the areas dominated by the westerly and monsoon circulations (Yao et al. 2012; Molg et al. 2014). The seesaw pattern between the ECA and the north Tibetan Plateau versus the south Tibetan Plateau observed in a range of proxies can only satisfactorily be explained by an interaction of westerly circulation and monsoon circulation.

The link between summer NAO and ECA drought can further be illustrated by the results of a wavelet coherence analysis between our ECA SPEI reconstruction and the July NAO index (Fig. 5). However, we noted that the ECA SPEI-NAO relationship showed divergence during some periods (e.g., 1890-1940), which means that NAO is not the only driver of ECA drought variability. Other drivers have been shown to influence ECA moisture availability, including the Arctic Oscillation (Wang et al. 2007), the silk road teleconnection pattern (Zhang and Zhou 2015), the Asian monsoon (Zhang and Zhou 2015), as well as thermal and dynamical impacts of the Tibetan Plateau (Ye and Gao 1979; Qian and Zhu 2001). Furthermore, the increase in anthropogenic greenhouse gases (e.g., CO₂) since the 1850s also possibly resulted in a regional wetting trend through changes in the hydrological cycle, as has been suggested by climate model experiments (IPCC 2013; Tett et al. 2007). Our research is only based on a statistical and physical mechanism explanation, but it could provide validation for climate model simulations in the exploration of the relationship between the strength of westerly circulation and ECA drought.

5 Conclusion

We compiled four strongly coherent, previously published ECA tree-ring $\delta^{18}\text{O}$ chronologies into a regional chronology, and reconstructed 301 years (1710-2010 CE) of March-September SPEI variability. The reconstruction explains 48.2% of the total variance in the instrumental March to September SPEI. We found the most prominent droughts in 1710-1770, 1810-1830, and the beginning of the 21st century, and a regime shift to a persistent wet period from the 1880s to 2000. These regional drought-wet patterns reveal the impact of drought on the Zhungar Empire, historic agricultural activities, human migration, and ecosystems. Our results also indicate that wet conditions in the ECA in the 20th century were related to the strengthening of the summer westerly circulation, as represented by the summer NAO. This result encourages further explorations of the interplay between westerly and monsoon synoptics and the seesaw pattern between ECA and southeastern Tibetan Plateau that can be investigated with paleoclimate model runs. Our findings contribute to an improved understanding of moisture shifts in ECA and potential causes of out-of-phase moisture patterns observed in proxies and reconstructions from the southern Tibetan Plateau and ECA over past centuries.

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Figure captions:

Fig. 1 Linear trend (a) of the March-September SPEI for the period 1957-2010.

Spatial correlation map (1957-2010) for the composited $\delta^{18}\text{O}$ chronology and the March-September SPEI gridded dataset (b). Trends in (a) are significant ($p < 0.05$). In (b), correlation coefficients higher than 0.3 or lower than -0.3 are significant at the $p < 0.01$ significance level. The locations of the four tree-ring $\delta^{18}\text{O}$ sites used to reconstruct ECA SPEI are indicated by black triangles. The trapezoid area with the strongest correlation coefficient, homogenous trend of SPEI, and nearest to our study sites (43° - 48°N , 80.5° - 94.5°E ; red frame) was selected to calculate the regional SPEI data. The location of the comparison sites of tree-ring reconstructions (see also text) and the Monsoonal Asia Drought Atlas Palmer Drought Severity Index (MADA PDSI; [Cook et al. 2010](#)) reconstructed grid points used for averaging regional MADA PDSI are also indicated on the map.

Fig. 2 Pearson's correlation coefficients between the composite tree-ring $\delta^{18}\text{O}$ chronology and mean monthly temperature, monthly total precipitation, and SPEI from CRU TS3.22 (gridded data of the region 43° - 48°N , 80.5° - 94.5°E ; red box in Fig. 1) for the period from 1957 to 2010. Months of the year previous to the growth year are indicated by "p" prior to the abbreviation of the month and M-S represents the average of March to September. The horizontal dashed lines represent the $p < 0.05$ significance level.

Fig. 3 ECA SPEI reconstruction (a) and associated decadal mean values (b), and the MADA PDSI reconstruction ([Cook et al. 2010](#)) at annual scale and low-frequency (c). High-pass filtered ECA SPEI and MADA PDSI reconstructions are compared in (d). Grey shading in (a) represents one standard deviation. The green (yellow) squares in (b) with error bars (standard deviation of the decadal SPEI) indicate the decadal mean values above (below) the mean values of the period 1710-2010. In this study, the low- and high-frequency are defined as the result of 30-year spline low-pass and high-pass filters, respectively.

Fig. 4 Composite analyses (1950-2011) for positive minus negative summer (June-July-August) NAO years. We used the 9 most positive minus the 9 most negative summer NAO years. The negative and positive summer NAO years are listed in Table S1. March-September precipitation from the Global Precipitation Climatology Centre (GPCC) (a); March-September surface precipitation rate from the NCEP/NCAR reanalysis (b); June-August average water flux and wind pattern at 500 hPa from the NCEP/NCAR reanalysis (c); and June-August average water flux and wind pattern at 850 hPa from the NCEP/NCAR reanalysis (d). The rectangles indicate the ECA region.

Fig. 5 Comparison between our ECA SPEI reconstruction, the July NAO index ([Jones et al. 1997](#)), and the reconstructed July NAO index ([Luterbacher et al. 1999](#)) at annual scale (a) and after 8-year-low-pass spline filter (b) to retain decadal signals. All series were z-scored over 1950-2000. The 31-year running correlation

coefficients between the SPEI reconstruction and the NAO indices (c). Spectral analysis for the SPEI reconstruction (d). Wavelet coherence analysis between the SPEI reconstruction and the instrumental July NAO index (Jones et al. 1997) over their common period (1821-2010) (e). The dashed line in (c) represents the 95% confidence level. In (d), the pink line and dashed pink line indicate the 90% and 95% confidence level based on a Monte Carlo test (Grinsted et al. 2004) and significant cycles are indicated. In (e), the wavelet coherence was also based on a Monte Carlo test (Grinsted et al. 2004) and the solid black lines indicate significant ($p < 0.05$) periods.

Fig. 6 Comparison between (a) the ECA SPEI reconstruction and association with flood (green bars) and drought (red bars) events based on historical documents (Shi 2006) from 1710 to 2000 and historic immigration (Li and Zhang. 2015), annual precipitation in the Qaidam Basin (Shao et al. 2005) (b), annual precipitation on the northeastern Tibetan Plateau (Yang et al. 2014) (c), far-western Mongolia June–September drought variability (Davi et al., 2009) (d), October–September precipitation in north Pakistan (Treydte et al., 2006) (e), June–August cloud cover variability on the southeastern Tibetan Plateau (Liu et al. 2014) (f), and July NAO variability (data from CRU, Jones et al., 1997) (g). The empirical distribution patterns and mean values (vertical dashed line) of these reconstructions is indicated for the periods from 1700 (1710) to 1886 and after 1887. In the figure, the black line represents the low-frequency variability as

827 calculated with 30-year low-pass spline filter. The gray lines indicate the
828 interannual variability. The dashed lines represent the result of two-sample *t*-test
829 mean value for two periods before 1886 (1860) and after 1886 (1860 for the figure
830 e and f). The dashed vertical zones are the droughts indicated by ECA
831 reconstruction.

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Figures

Fig. 1

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Supplementary Table:

Table S1. The years of 9 most positive and the 9 most negative summer (June-July-August) averaged NAO values (accounting about 16% of 1957-2011).

Positive	Negative
1960	1962
1961	1965
1970	1966
1972	1968
1975	1973
1989	199
1993	2000
2003	2008
2010	2011

Supplementary figures:

Fig. S1 The z-scored values of four ECA tree-ring $\delta^{18}\text{O}$ series based on the 1790-2008 reference period (a). The inter-series correlations (Rbar) for the four tree-ring $\delta^{18}\text{O}$ series using time windows of 51 years (lagged by 25 years) and 21 years (lagged by 10 years). The number of series for corresponding Rbar periods (b). Comparison between the composited tree-ring $\delta^{18}\text{O}$ chronology and the first principal component (PC1) of four ECA tree-ring $\delta^{18}\text{O}$ series shows high coherence in $\delta^{18}\text{O}$ variability but a small mean offset. The PC1 series explains 54% of the total variances and the eigenvalues of the first four PC axes are indicated

in the inset (c). Regression analysis plot of PC1 and regional March-September SPEI data (1957-2010) (d).

Fig. S2 Linear trends of regional March-September climate parameters for the period from 1957 to 2010. CRU ts 4 temperature (Jones et al., 2012; Harris et al., 2014) (a); CRU ts 4 precipitation (Jones et al., 2012; Harris et al., 2014) (b). GPCP V7 + monitoring precipitation from Global Precipitation Climatology Centre (Schneider et al., 2016; Ziese et al., 2011) (c); CRU scPDSI (self-calibrated Palmer Drought Severity Index) (Wells et al., 2004; Dai et al., 2011; Harris et al., 2014; Osborn et al., 2017; van der Schrier et al., 2013) (d); UCAR scPDSI (Wells et al., 2004; Dai et al., 2011) (e). The trapezoid frame indicates our study region in the figure.

Fig. S3 Location of the sampling site and meteorological stations in our study region (trapezoid frame) (a). Inter-annual variability of March-September temperature (b) and precipitation (c) for the meteorological station for period 1960-2010. The data are z-score

transformed for the period 1960-2010 and the regional temperature and precipitation are calculated the arithmetic mean value from the stations in figure a.

Fig. S4 Spatial correlation map for the individual tree-ring $\delta^{18}\text{O}$ chronology and the March-September SPEI gridded dataset since 1957 (Vicente-Serrano et al. 2010). The spatial correlation was calculated using the KNMI web-software (Trouet and Van Oldenborgh, 2013). (a) for the FY site, (b) for the HM site, (c) for the WH site, and (d) for the WSS site. The study area was indicated by the trapezoid frame.

Fig. S5 Regression plot of the composite regional tree-ring $\delta^{18}\text{O}$ chronology and regional SPEI data for the period from 1957 to 2010 (a). Comparison of reconstructed ECA SPEI (blue line) and CRU SPEI data (black line) (b). The dashed lines in (a) represent the regression lines with their $p < 0.01$ significant intervals. The regression equations are also appended to the figure. The shaded area in (b) indicates the estimated reconstruction uncertainty with one standard deviation intervals. In addition, Pearson's r -values were calculated for the period 1902-2010, 1902-1957 and 1957-2010.

Fig. S6 Comparison of CRU SPEI data and the tree-ring $\delta^{18}\text{O}$ based SPEI reconstruction using variance scaling (a) and regional SPEI reconstructions based on linear regression and variance scaling (b).

Fig. S7 Pearson's correlation coefficients (r) between our ECA SPEI reconstruction and MADA PDSI gridded data (Cook et al., 2010) in the region for the common period 1710-2005 (a). The green dots in (a) represent statistical significance levels of $p < 0.05$. The results of the Ensemble Empirical Mode Decomposition (EEMD) analysis for the ECA SPEI reconstruction and for the averaged MADA PDSI reconstruction (b). In (b) EEMD decomposed the original signal into six intrinsic mode functions (IMFs). Note that the y-axis scales differ among the graphs. The percentage of the original signals accounted by each IMF was calculated by means of regression analysis.

Fig. S8 Spatial correlation analysis between our ECA SPEI reconstruction with CRU scPDSI (a) and UCAR scPDSI (b) for the period 1957-2010 and for the period 1900-2010 (c) and (d), respectively. The trapezoid frame indicates our study region in the figure.

Fig. S9 ECA SPEI reconstruction (a), an independent summer temperature reconstruction from the Tianshan Mountains (Yu et al., 2013) (b), and their 31-year running correlation coefficients (c). Scatter plot of the ECA SPEI reconstruction and March-September CRU TS3.22 temperature (1901-2010) (d). Negative SPEI values are indicated by red dots, positive values by blue dots. Only significant regression equations are shown. Green filled dots represent the years 2000-2010. Comparison between our ECA SPEI reconstruction (black line) and March-September normalized difference vegetation index (NDVI) for the region 40°-50°N, 80°-90°E (e). Regressions coefficients were calculated for the entire period 1982-2010 and for 1994-2010.

Fig. S10 Composite analyses for the years with positive and negative NAO. The negative and positive NAO years are listed in Table S1. NCEP/NCAR reanalysis June-July-August water flux and wind pattern at 500 hPa (a); and NCEP/NCAR reanalysis June-July-August water flux and wind pattern at 850 hPa (b). The rectangles indicate the ECA region.

Fig. S11 Pearson's correlation with p -values between ECA SPEI reconstruction and the monthly instrumental (red, 1821-2011) (Jones et al. 1997), Hurrell Station-Based (Green, 1865-2011) (Hurrell et al. 2003), and reconstructed (blue, 1700-2000) (Luterbacher et al. 1999) NAO index. Months of the year previous to the growth year are indicated by "p" prior to the abbreviation of the month. Statistically significant results ($p < 0.05$) are indicated by "x".

Fig. S12 Wavelet coherence analysis between the SPEI reconstruction and the reconstructed July NAO index (Luterbacher et al. 1999) during their common period from 1700 to 2011. The solid black lines indicate significant ($p < 0.05$) periods based on Monte Carlo test (Grinsted et al., 2004).

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