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Climate of the Past Discussions

Interaction between East Asian summer monsoon and west winds as shown by tree-ring records

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Abstract:

Against the background of changes in global atmospheric circulation, local changes in the East Asian summer monsoon (EASM) and the mid-latitude westerly winds will inevitably affect the climate and ecology of the arid zone of Northwest China. Hence, it is important to study these changes. We chose to observe these changes in the Alxa Plateau using dendrochronological methods. We assembled ring-width records from Qinghai spruce trees growing in the mountain regions surrounding the Alxa Plateau: the Helan Mountains, Changling Mountain, and Dongdashan Mountain. We analyzed these records for changes on interannual and interdecadal scales. Our results show that radial growth was indeed affected by changes in the monsoons and westerlies. The heterogeneity of precipitation and climatic wet-dry changes in different regions is primarily influenced by the interactions between atmospheric circulation systems, each with its own dominant controlling factors. In the case of the Helan Mountains, both of these major atmospheric circulation systems play a significant role in shaping climate changes. Changling Mountain in the southern part of the Alxa Plateau are mainly influenced by the EASM. Dongdashan Mountain is mainly influenced by the westerlies. Understanding these local conditions will help us predict climate changes in Northwest China.

Key words: Alxa Plateau, dendroclimatology, westerly winds, EASM, interaction between winds and monsoon

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1. Introduction

1.1 Importance of climate studies in northwestern China

The alpine zone of Qinghai-Tibet, the arid zone of the northwestern interior, and the humid zone of the east constitute the three main areas of China's natural geomorphology (Chen et al., 2019b). The Northwest China inland dry zone is located in the hinterland of the Eurasian continent and is among the driest regions in the world. It exhibits typical continental climatic characteristics. This region is mainly influenced by westerly winds and the EASM. The interaction of these two factors results in high precipitation variability and hence frequent droughts. This would be the case even before global climate change began affecting the area; it is even more the case in current years. This inland dry zone is very much an ecologically fragile area (Chen et al., 2019a; Chen et al., 2019b; Zhang et al., 2023).

The semi-arid and arid regions of northern China are characterized by large areas of sand and desert. They are the second largest source of dust in the world after the Sahara. Their contribution to global climate change is large. So far inland, the influence of the EASM is often weak (Zhang et al., 2021; Liu et al., 2022). It is opposed by the westerly winds that flow from the North Atlantic climate zone toward the East Asian monsoon climate zone (Qu et al., 2004). The interaction between the westerly winds and the EASM governs precipitation, water vapor transport, and thus the climate of northwestern China (Feng et al., 2004; Wang et al., 2005; Li et al., 2008; Ma et al., 2011).

It is important to understand the history of this interaction if we are to estimate how global climate change will affect it. Global atmospheric circulation is likely to change as is the EASM. Climate change will not only affect the regional climate and regional water resources (Ding et al., 2023); it will affect East Asia (dust storms) and even the rest of the globe. Hence the study of climate in this region is of great practical and theoretical significance (Chen et al., 2019a; Chen et al., 2019b).

The westerly winds and the EASM meet at the northern boundary of the Asian summer monsoon (Huang et al., 2023). In northern China, this boundary runs from west to east, along the eastern section of the Qilian Mountains, the southern foothills of the

Helan Mountains, the Daqing Mountains, and the western section of the Daxinganling Mountains. This is not a static boundary. It fluctuates within a range of 200–700 km (Chen et al., 2018). It is important to understand the history of these fluctuations (Huang et al., 2023).

This can be done using climate records such as lacustrine, eolian, and dendrochronological (Sun et al., 2003; Liu et al., 2005; Li, 2009; Chen et al., 2010; Li et al., 2016; Chen et al., 2019b; Qin et al., 2023). Our research team specializes in dendrochronology, which is one of the best tools for studying paleoclimatic changes, due to its precise dating, high resolution, good continuity and high replication (Zhang et al., 2003; Shao et al., 2010; Yang et al., 2014; Liu et al., 2016).

1.2 Previous work

The climate history of the Baotou area, at the northern edge of the EASM, has been studied at interannual and interdecadal scales for the past 260 years, based on June–August precipitation reconstruction from tree-ring samples from the western Yinshan Mountains (Liu et al., 2001; Liu et al., 2003). Using tree-rings and historical records, Kang and Yang (2015) reconstructed the annual precipitation history of the East Asian monsoon northern fringe zone for the last 530 years. They analyzed spatial variability and possible driving mechanisms using the 400-mm isohyet.

Several May–July precipitation sequences have been reconstructed using ring-width and latewood-width data from Chinese pine (*Pinus tabulaeformis*) growing in the Helan Mountains (Ma et al., 2003; Liu et al., 2004; Chen et al., 2016). Studies of tree-ring carbon and oxygen isotopes from Chinese pine samples have shown that $\delta^{18}O$ values increase with summer precipitation, while $\delta^{13}C$ values decrease (Zhang et al., 2005a; Liu et al., 2008). Westerly winds have also been shown to affect precipitation in the Helan Mountains (Chen et al., 2010).

Principal component analysis of tree-ring chronologies constructed from data collected at several sites in Gansu suggests that trees at these sites were more influenced by EASM than by west winds (Chen et al., 2013). These researchers also found that the EASM weakened in 1970s, but recovered in the early 1990s. Tree-ring data allowed the

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reconstruction of 330 years of PDSI values for the Mount Hasi region (at the northern boundary of the summer monsoon zone) (Kang et al., 2012). This study confirmed that radial growth of Chinese pine has declined over the past three decades, due to the weakening of the EASM. Dendrochronological reconstruction of precipitation in the Mount Changling region (again using Chinese pine) suggested that precipitation in that region mainly depends on the EASM (Chen et al. 2012). Other researchers have assembled tree-ring chronologies from pines growing in the Mount Qilian region and the northern mountains of the Hexi Corridor. Here again precipitation is associated with the EASM. These chronologies have allowed scholars to compile precipitation, temperature, and drought records for the last thousand years (Gou et al., 2015a; Gou et al., 2015b; Zhang et al., 2017).

Most modern researchers studying climate change in the region are mostly carried out on single sample sites (Wang et al., 2004; Liu et al., 2005; Chen et al., 2010; Chen et al., 2016; Li et al., 2016; Liu et al., 2016; Chen et al., 2018). There is a dearth of multi-site, regional, large-scale studies on the interaction of the westerlies and the EASM. ~~Our group~~ compiled and analyzed tree-ring chronologies from Qinghai spruce (*Picea crassifolia*) growing in the Helan, Changling, and Dongdashan mountain regions that surround the Alxa Plateau, the climate response characteristics of spruce radial growth in three regions was then analyzed. Combining the relevant Westerly and East Asia monsoon circulation indices, the driving mechanism of the regional climate change by with the interaction and synergistic roles of two atmospheric circulation systems in the Alxa Plateau was explored. The results will lay a theoretical foundation for the climatic evolution of the region and the desertification control.

2. Material and methods

2.1 Study area

The Alxa Plateau is located in the western part of the Inner Mongolia Autonomous Region and is surrounded by mountains (Fig.1). It consists primarily of three deserts: Tengger, Ulan Buh, and Badan Jaran. It lies south of the Gobi desert. It is the main source of the fierce sandstorms and dust storms that blow toward eastern China and the

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Pacific. It has been much affected by climate change; sand- and dust storms have increased, much to the detriment of lands to the east. The Chinese government is doing what it can to establish an environmental defense line there. It is currently the Northern Sand Prevention Belt of the National Two Ecological Barrier and Three Belts Ecological Security Strategy Pattern (Xiao et al., 2017; Xiao et al., 2019).

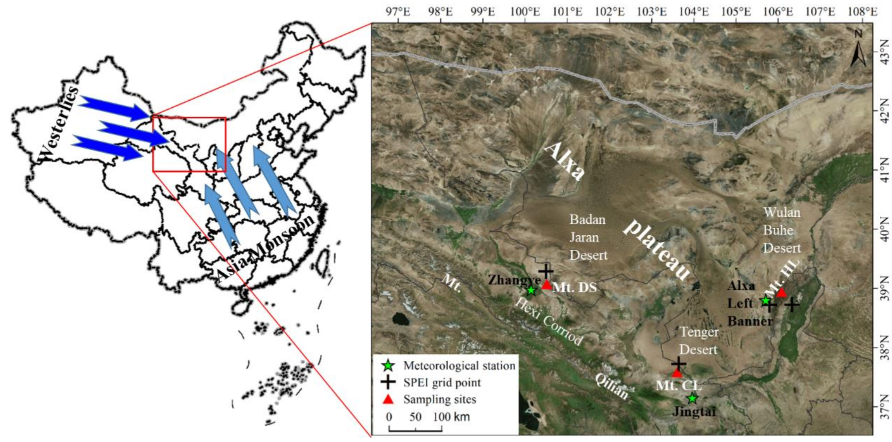

Figure 1. Location of tree-ring sampling sites and meteorological stations (the right panel is from Mapworld)

There are several mountain ranges surrounding the Alxa Desert, such as the Helan Mountains in the east, the northern mountains of the Hexi Corridor, and the outliers of the Altai Mountains in the north. These mountains not only block the eastward and southward expansion of the desert (driven by high pressure regions from Mongolia); they are also the source of mountain rivers and streams that water the oases on the plateau.

The Alxa Plateau is located in the eastern margin of the inland arid region of Central Asia. It is affected not only by the mid-latitude westerly circulation, but also by the Asian monsoon and the plateau monsoon. It is in the zone where the mid-latitude westerly circulation and the Asian monsoon interact (Xiao et al., 2017; Chen et al., 2019b). As a result, there is large interannual variability of vegetation cover in the region (Ou and Qian, 2006; Tang et al., 2006; Li et al., 2013).

The Helan Mountains (38°27′~ 39°30′N, 105°20′~106°41′E) (sampling site henceforth abbreviated as HL), are located at the eastern edge of the Tengger Desert.

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They stretch more than 200 kilometers from north to south; the main peak is ~3,556 m. The mountain forests are dominated by Qinghai spruce and Chinese pine, juniper, mountain aspen, and elm.

Mount Changling (37°12′~37°17′, 102°45′~103°48′E) (sampling site henceforth abbreviated as CL) is an independent mountain protruding northward from the remnants of the eastern Qilian Mountains, It is located at the southern edge of the Tengger Desert; its elevations range from 2100 to 2900 m. The dominant tree species are Qinghai spruce and Chinese pine.

Mount Dongdashan (39°00′~39°04′N, 100°45′~100°51′E) (sampling site henceforth abbreviated as DS) is located at the southwestern edge of the Badan Jaran Desert and the middle part of Mount Qilian. It is one of the northern mountains along the Hexi Corridor; that range consists of mountains that vary from 2200 to 2637 m in elevation. Forests are dominated by Qinghai spruce and Qilian juniper.

The temperatures of the coldest months recorded at meteorological stations in the Alxa Left Banner (a division of the Alxa League region), Jingtai (a county in Gansu), and Zhangye (a city in Gansu) all occurred in January, ranging from -9.8℃ to -6.8℃. The hottest months at those stations were in July (21.9℃ to 23.1℃). These meteorological stations are the closest stations to our three sampling sites.

Precipitation measured at those stations varied widely. The multi-year average of total precipitation from May to September was 171 mm at Alxa Left Banner station, 156 mm at Jingtai station, and 108 mm at Zhangye station. This accounted for more than 80% of the annual precipitation (Fig.2).

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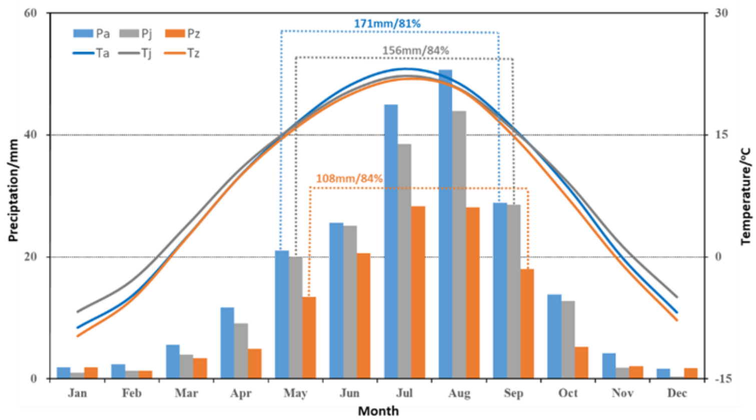

Figure 2. Climatic diagram of study area. Pa/Ta are the monthly total precipitation and monthly mean temperature at the Alxa Left Banner meteorological station (1953–2016); Pj/Tj are the precipitation and temperature figures for the Jingtai meteorological station (1957–2017); Pz/Tz are the precipitation and temperature figures for the Zhangye meteorological station (1957–2017). The dashed box and appended data indicate the total growing season precipitation in the study area and the proportion of total annual precipitation.

2.2 Sample collection, processing and data analysis method

2.2.1 Sample collection, processing and dendrochronology construction

Researchers used standard methods of tree-ring sample collection. One core was drilled from each tree in the sample site. We collected 209 cores in total, from five sampling sites at HL, 48 cores from one sampling site at CL, and 81 cores from two sampling sites at DS. Relevant information re the sampling sites is summarized in Table 1.

Chronologies were constructed using standard dendrochronological methods. We calculated the highly significant correlations ($P<0.001$) between the chronologies of different points at the HL and DS mountains; a weighting method was used to finally synthesize a chronology for each mountain.

2.2.2 Climate data, atmospheric circulation indices and the related Analyzing methods for chronological correlation

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Climate data for the study areas HL, CL, and DS were collected from the nearest meteorological stations in Alxa Left Banner, Jingtai and Zhangye, respectively (http://data.cma.cn).

We used SPEI to represent the local drought and wetness conditions. SPEI data (grid-point resolution 0.5°*0.5 °) was obtained from the grid-point datasets of the National Center for Environmental Predictions-National Center for Atmospheric Research (NCEP-NCAR). Time scales ranged from 1 month to 15 months. The mean values of data from two grid-points closest to the HL sampling site (38.75°N,105.75°E and 38.75°N,106.25°E; period 1953–2015) were chosen for subsequent analysis. Grid-point data from one site closest to our CL sampling site (37.75N, 103.75E; period 1951–2015) was used for later analysis. Grid-point data from one site closest to our DS sampling site (39.25°N, 100.75°E; period 1951–2015) was also used. As SPEI datasets are multi-scale, we preprocessed the data to identify and select 11-month scaled SPEI datasets for subsequent analysis.

We took into account the so-called lagging effect (the influence of fall and winter climate factors on the radial growth of trees shows up later in the year) and chose to use temperature, precipitation, and SPEI data from September of the previous year to September of the current year (abbreviated as P9–P12 and C1–C9), as collected at each meteorological station, for our climate response analysis.

The East Asian Summer Monsoon Index (EASMI) represents the activity strength of the EASM. Larger EASMI values indicate a stronger summer monsoon, smaller ones a weaker monsoon. In this study, the EASMI (mean values for June–August in the period 1950–2017) defined by Li and Zeng (2005) was used to study the impact of the EASM on climate change in the study area.

We used the Westerly Circulation Index (WCI annual mean) to represent the strength of the mid-latitude westerlies. The larger the WCI value, the stronger the Eurasian latitudinal circulation; the smaller the value, the weaker the Eurasian latitudinal circulation. WCI data (period 1951–2015) were derived from the Eurasian Latitudinal Circulation Index published by the National Climate Center of the China Meteorological Administration (https://cmdp.ncc-cma.net/cn/index.htm).

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Interannual and interdecadal (sliding moving average of 11a) chrono-climatic/-cyclonic index correlation and partial correlation analyses were performed using SPSS 19.0. Based on the characteristics of tree-ring series, the sequences were classified into three groups of low, average and high ring widths using mean ± 1δ (SD) as the classification criterion (with mean ± 2δ as the extreme year). Correlation statistical tests were performed with the corresponding annual circulation indices; similar treatments and analyses were performed for the two major circulation indices.

3. Results and analysis

3.1 Ring-width chronologies and their characteristics

Based on the sampling cores from five sample sites at HL, two sample sites at DS, and one sample site at CL, ring-width residual chronologies were derived for each of the three study areas (Fig. 3). Statistical parameters showed that the three chronologies meet the usual requirements for correctly done dendrochronological studies. (Table 1).

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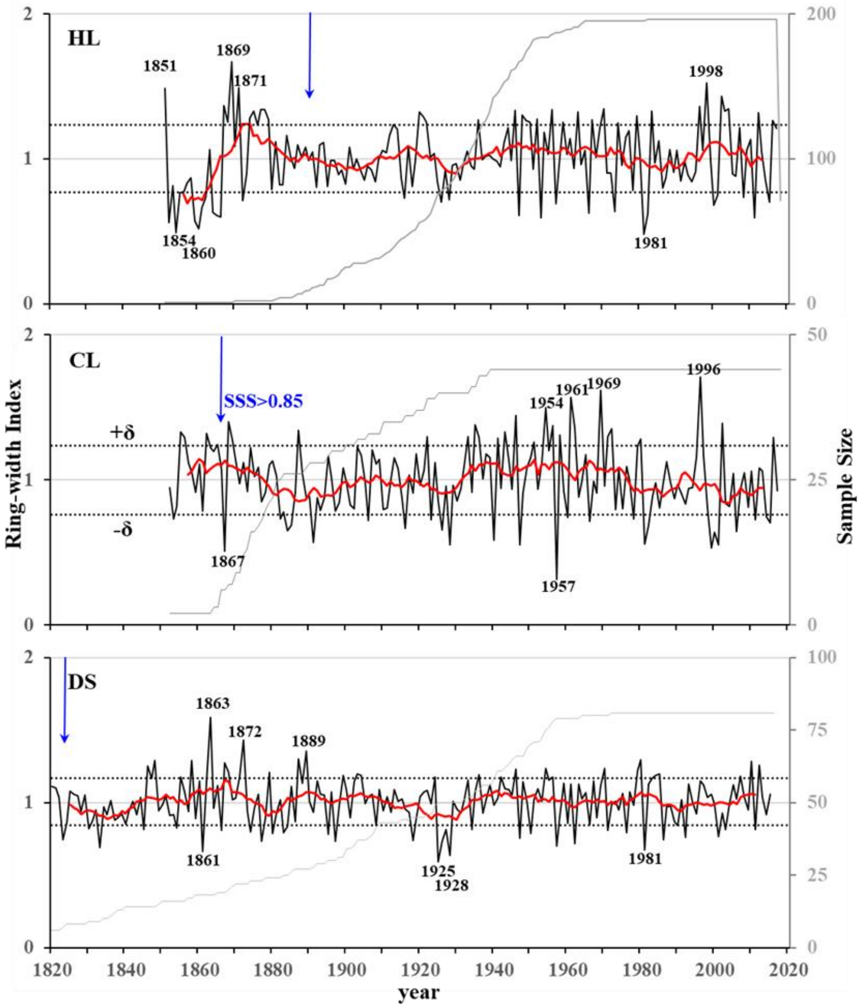

Figure 3. Residual ring-width chronologies for the three study areas. The dark lines indicate the chronology; grey lines indicate the sample depth; red lines indicate the 11-year running mean chronology; dotted horizontal lines indicate the mean value +/-1δ, years with data identified as >/< mean +/-2δ (δ: standard deviation); the blue arrows indicate the start of the reliable residual chronology (SSS>0.85).

Table 1. Statistical characteristics of the sampling sites and the tree-ring chronologies.

Sampling sites	HL(5)	CL(1)	DS(2)
Latitude (°N)	38.52–38.97	37.61	39.04
Longitude (°E)	105.83–106.02	103.71	100.78

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Elevation (m)	2200–2750	2490	2650–2700
Cores	209	48	81
Reliable period	1891–2018	1866–2017	1823–2015
MS	0.18–0.37	0.28	0.15–0.33
Rbar	0.45–0.61	0.56	0.40–0.60
SNR	22.5–56.1	38.9	25.7–42.5
EPS	0.96–0.98	0.98	0.96–0.98
PC1(%)	17.3–63.0	57.9	43.0–62.5

Reliable period, MS (mean sensitivity), Rbar (mean series intercorrelation), SNR (signal to noise ratio), EPS (expressed population signal), and PC1 (variance explained by the first principal component) refer to residual chronologies.

3.2 Climate response characteristics

Correlation analysis comparing a) monthly mean temperature and precipitation at neighboring meteorological stations and b) SPEI at the nearest grid-point showed that, overall, the three residual chronologies were correlated negatively with monthly mean air temperature, positively correlated with monthly precipitation, and positively correlated with SPEI during the growing season (Fig. 4).

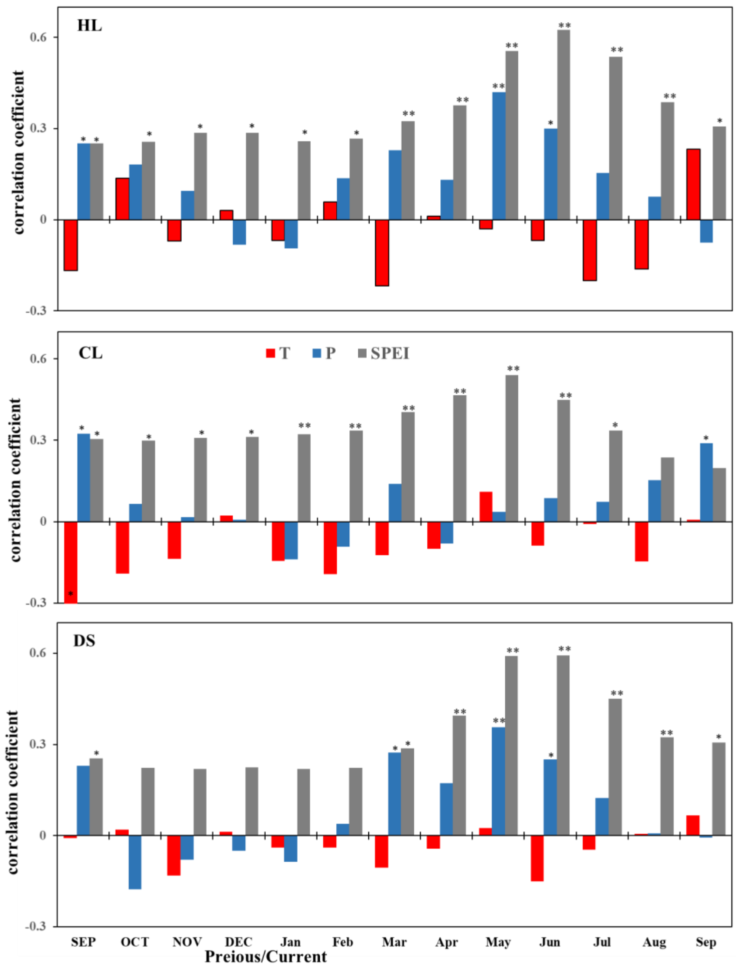

Figure 4. Correlation coefficients (Pearson's r values) between the residual ring-width chronologies of Qinghai spruce at the three study areas (HL, CL and DS) and the observed monthly temperature (T), total monthly precipitation (P), and SPEI.

* Pearson's r correlation, significant at P < 0.05.

* * Pearson's r correlation, significant at P < 0.01.

Month names of previous year are capitalized.

HL chronology was correlated negatively with mean temperature mainly in C5–C8 in the growing season, but not to the significant level. It was also positively correlated with precipitation in all months except P12, C1, and C9, reaching significant levels (P<0.05) in P9, C5, and C6. All months were positively correlated with SPEI and

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reached statistical significance (P<0.05), with C3–C8 showing highly significant correlation levels (P<0.01).

CL chronology was negatively correlated with mean temperature in all months except for P12, C5 and C9. Only P9 reached statistical significance (P<0.05). CL chronology was positively correlated with monthly precipitation, save for C1, C2, and C4. Only P9 and C9 reached statistical significance (P<0.05). All months were positively correlated with SPEI, with P9–C7 reaching significant correlation levels (P<0.05) and C1–C7 reaching highly significant correlation levels (P<0.01).

DS chronology showed weak correlations between DS chronology and monthly mean temperatures. None of the correlations reached levels of significance. DS chronology was positively correlated with P9 and C2-C8 precipitation and reached significant correlation levels for C3, C5, and C6 (P<0.05). All months were positively correlated with SPEI, with P9 and C3-C9 reaching significant correlation levels (P<0.05) and C4-C8 reaching highly significant correlation levels (P<0.01).

Overall, the radial growth of Qinghai spruce at the three study areas seems to have been limited, for the most part, by low precipitation during the growing season (April–July). The three chronologies reflect regional wet and dry variations.

3.3 Regional climate changes as recorded by tree-ring widths

3.3.1 Regional climate change viewed at interannual scales

On interannual scales, the three residual chronologies, when compare, showed highly significant correlations. HL-CL: n=166 (P<0.01); HL-DS: n=165 (P<0.01); CL-DS: n=164, P<0.01). This indicates that there was a high degree of consistency in the radial growth of Qinghai spruce in the three regions.

According to the results of the chronology-climate response analysis in the previous section, the high and low ring-width indices (mean±1~2δ) of the chronology at the three sample sites indicate wetter or drier, and extreme wet or dry years, respectively (Table 2).

Overall, the three ring-width residual chronologies (HL, CL, DS) had a total of two shared wetter years and seven shared drier years. The HL and CL chronologies shared

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four wet years and eleven dry years; the HL and DS chronologies shared five wet years and nine dry years; and the CL and DS chronologies shared five wet years and seven dry years (Table 2).

Table 2. Years of high or low growth, wet or dry climate in the three chronologies

chronology	Higher index/wetness years (≥mean+1δ)	Lower index/drought years (≤mean-1δ)
HL	**1851**, 1867, **1869**, **1871**, 1874, 1875, 1877, 1878, 1879, 1920, 1921, 1922, 1936, 1946, 1952, 1959, 1963, 1967, 1970, 1974, 1979, 1983, 1996, **1998**, 2002, 2003, 2004, 2012, 2016	1852, **1854**, 1855, 1859, **1860**, 1861, 1864, 1865, 1866, 1872, 1916, 1926, 1928, 1947, 1953, 1957, 1966, 1973, **1981**, 1982, 2000, 2001, 2011
CL	1855, 1862, 1922, 1933, 1935, 1936, 1941, 1946, 1951, **1954**, 1958, **1961**, **1969**, 1972, 1980, **1996**, 2016	**1867**, 1884, 1885, 1891, 1916, 1923, 1926, 1928, 1947, **1957**, 1960,1966, 1978, 1981, 1982, 1999, 2000, 2001, 2006, 2011, 2014, 2015
DS	1846, 1848, 1855, 1858, **1863**, 1867, 1868, **1872**, 1879, 1887, **1889**, 1899, 1903, 1904, 1924, 1936, 1942, 1946, 1954, 1956, 1959, 1970, 1979, 1980, 2007, 2010, 2012	1823, 1830, 1833, 1854, **1861**, 1874, 1877, 1880, 1883, 1884, 1895, 1897, 1908, **1925**, 1926, 1927, **1928**, 1934, 1947, 1950, 1957, 1962, 1971, 1976, **1981**, 1985, 1990, 1992, 2001, 2003, 2011

Note: Bold is used to indicate extreme years (>/<mean+/-2δ；δ：standard deviation); double underlining indicates years shared between two of the three sample sites; bold underlining shows years shared between three sample sites.

There were no extremely wet years shared by the three sample sites. However, there were two shared wetter years in 1936 and 1946 and several shared wetter years in later years among the three sample sites. For example, note the wetter years in 1922 and 2016 for HL and DS chronologies; 1959, 1979, and 2012 for HL and DS chronologies; 1855, 1954, and 1980 for CL and DS chronologies (Table 2).

The extreme drought years are consistent among the three sample sites. For instance, there was an extreme drought year in 1981 at HL and DS sample sites; it was also a drought year at CL. An extreme drought year at CL in 1957 was also a drought year for the other two chronologies. Moreover, the extreme drought year of 1928 at DS was a drought year at the other two sites. Drought years in 1926, 1947, 2001, and 2011

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were seen in all three sites and in two of the three sample sites (1916, 1966, 1982, and 2000 at HL and CL; 1854 and 1861 at HL and DS) (Table 2).

3.3.2 Characteristics of regional climate change at inter-decadal scales

On the decadal scale, the 11a running mean series indicates that at the HL site there were four wetter periods (mid-1860s to early 1880s; 1910s to 1920s; mid-1930s to mid-1970s; and late 1990s to early 2010s). Four drought periods were seen (mid-1850s to mid-1860s; early 1890s to late 1900s; circa 1930s; and mid-1970s to 1980s) (Fig.5).

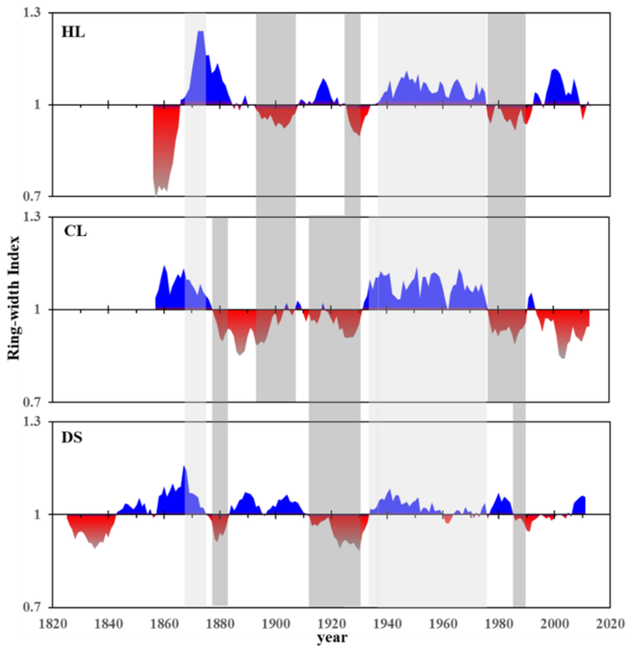

Figure 5. Three regional chronologies demonstrating alternation between dry and wet years on interdecadal scales (11 a running mean). The gray bands indicate consistent changes.

The CL regional chronology revealed two main wetter periods (mid-1850s to mid-1870s; mid-1930s to mid-1970s) and two longer drought periods (late 1870s to early 1930s; following the late 1970s) (Fig.5).

The DS regional chronology showed four main wetter periods (mid-1840s to mid-1870s; mid-1880s to late 1900s; mid-1930s to mid-1980s; and late 2000s to early

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2010s). There were four drought periods (mid-1820s to mid-1840s; mid-1870s to 1880s; early 1910s to early 1930s; and late 1980s to mid-2000s). The drought during the last drought period was less severe (Fig.5).

The three chronologies show both synchronized phases and differential changes on an interdecadal scale. The more synchronized dry phases of climate change were the drought periods of the 1930s and 1990s. When we compared the DS chronology to the HL and CL chronologies on decadal scales, we noted that DS droughts tended to last longer and that they started and ended later than CL droughts. However, HL and DS droughts tended to end at ~~close to~~ the same time (Fig.5).

There were two wet periods in 1870s and the mid-1930s to 1970s which were shared by all three sample sites. The latter period was the longest lasting wet period we saw in our study. There were also dry and wet periods that were not shared by any of our sites. There was an HL drought (mid-1850s to mid-1860s) which was not shared by the other two sites, which were wetter. HL and CL shared drought periods (1890s to 1910s; 1980s) while DS was wetter. Conversely wetter periods at HL were sometimes accompanied by drought in the other two sites. Drought at CL was sometimes accompanied by wet periods at the other two sites. DS was wet during the 2010s but the other two sites were in drought (Fig.5).

The results of the above studies show that there are diversified and complex features in the interdecadal processes of climate change in different regions around the Alxa Plateau.

3.4 Driving mechanism of the regional climate changes

3.4.1 Driving mechanism of the regional climate changes of typical years

On the interannual scales, three regional chronologies we developed showed fairly weak negative correlations between the EASM and the westerlies; none of the correlations were statistically significant. We carried out correlation analyses of the three regional ring-width chronologies and two major circulation indices. This was done in high, medium and low ring-width index groups (Fig. 6; 7).

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At HL, the results of our combined subgroup correlation analyses suggest that correlations between radial growth groups and atmospheric circulations were stable. Correlation between the higher ring-width group and atmosphere circulation indices and between the lower ring-width group and the WCI were all significant (P<0.05) (Fig. 6; 7).

At CL, correlations between the higher and middle ring-width groups to the WCI and the higher and middle WCI groups to the ring-width index were all negative. Correlations between the higher and middle ring-width groups and the EASMI, and between the higher and middle EASMI groups with the ring-width index were inconsistent (Fig. 6; 7).

At DS, correlations between the higher and lower ring-width groups and the EASMI, and between the higher and lower EASMI groups to the ring-width indices, were consistent. The correlations between the higher ring-width groups and the WCI, and between the higher WCI groups and the ring-width index were consistent. However, the correlations between the lower ring-width groups and the WCI, also between the lower WCI groups and the ring-width index, were inconsistent (Fig. 6; 7).

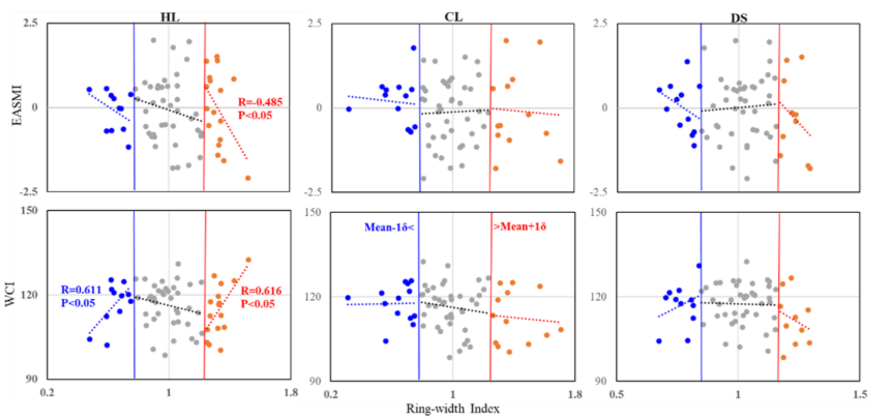

Figure 6. Grouping related charts among the ring-width index of three regions (HL, CL and DS) and the two atmospheric circulations' indices (EASMI and WCI). The noted numbers are the person correlation coefficients (two-tails test) and the corresponding significant credible level. Red dots indicate the higher ring-width index group

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(>mean+1δ), gray dots indicate the middle ring-width index group (>mean-1δ~< mean+1δ), and blue dots indicate the lower ring-width index group (>mean-1δ).

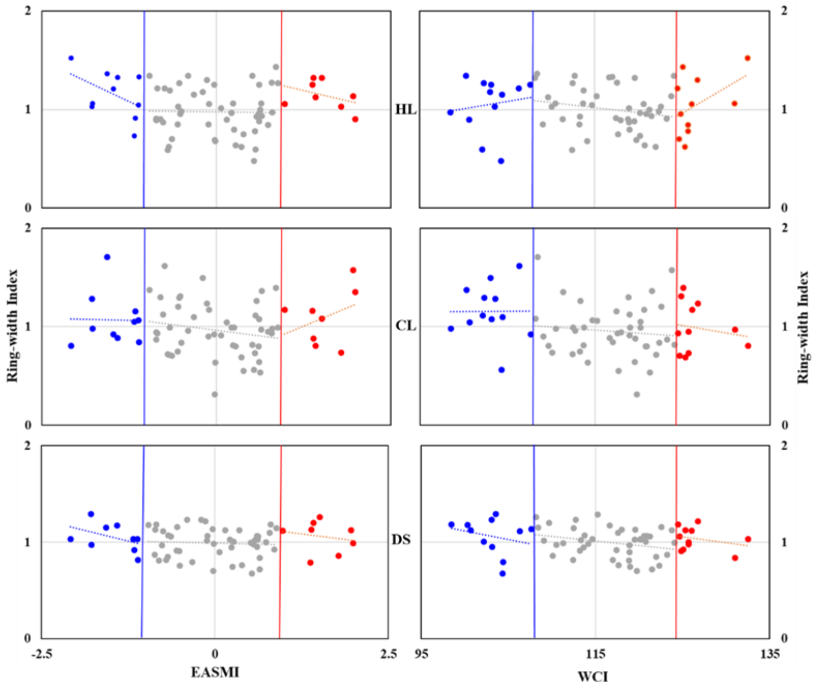

Figure 7. Grouping related charts among the two atmosphere circulations' index (EASMI and WCI) and the ring-width index of three regions (HL, CL, and DS). Red dots indicate the higher atmosphere circulations' index group (>mean+1δ), gray dots indicate the middle atmosphere circulations' index group (>mean-1δ~< mean+1δ), and blue dots indicate the lower atmosphere circulations' index group (>mean-1δ).

Except for HL, none of the ring-width groups or the atmospheric circulation index groups of the others reached a level of significance. These results suggest that HL is strongly affected by size of, and the interaction between, the EASM and the westerlies. On an interannual scale, stronger west winds and a weaker monsoon could result in variations from the ordinary climate (veering towards drier or wetter). Weaker west winds and a stronger monsoon formed the normal climate at HL. At the CL and DS sites, both atmospheric circulations were relatively weak on interannual scales. They had complex interactions.

3.4.2 Driving mechanisms of the regional climate changes on a decadal scale

At HL, both the EASM and the westerly circulation had highly significant effects on the radial growth of the Qinghai spruce. At CL, the EASM also had highly significant effects on radial growth of the Qinghai spruce. There, correlation coefficients were higher for the EASMI (EASM index) than they were for the HL index. Correlations between the WCI and radial growth were negative, but not at a significant level.

At DS, correlation between radial growth and the WCI was extremely negative (P<0.01). Correlation between radial growth and the EASM was positive (P>0.1) (Fig. 8). These results suggest that at HL, alternations between dry and wet seasons were affected both by the EASM and the westerlies. If either of the two atmospheric circulations was stronger, the climate tended to be wetter. At CL, alternations between dry and wet were affected mainly by the EASM. When the EASM was stronger, the climate was wetter. At DS, the climate was affected mainly by the westerly winds. The stronger the winds, the wetter the climate (Fig. 8).

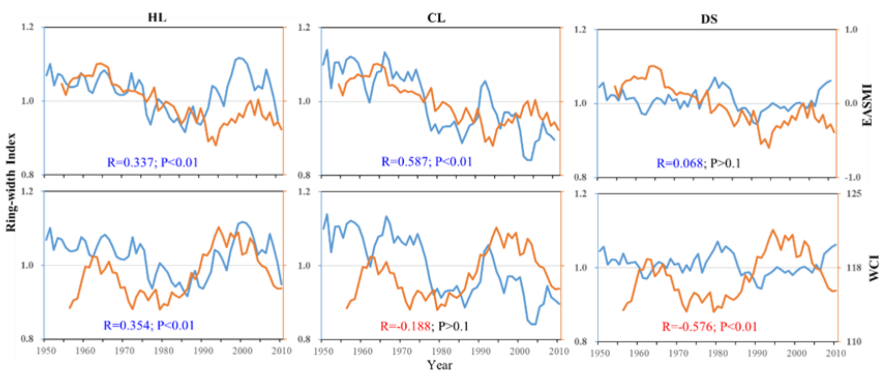

Figure 8. Interdecadal scale (11-a running average) correlations of the three residual chronologies with the EASMI and WCI.

The results of our interdecadal partial correlation analysis of the three RES-chronologies with the WCI and EASMI further illustrate the impacts of the two circulation systems on the climate of the three regions (Table 3).

At HL, if we control one factor (the WCI or EASMI) from our analysis, the other factor showed a positive correlation (P<0.0001). At CL, if we controlled the WCI, we find a positive significant correction with EASMI (P<0.0001). If we controlled the

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effect of EASMI, we saw a weak negative correction with WCI (Table 3). At DS, if we controlled EASMI, we saw a negative significant correlation with WCI (P<0.0001). If we controlled the WCI, we saw an insignificant negative correlation with EASMI (Table 3).

Table 3. Inter-decadal partial correlation analysis of the three residual-chronologies with the WCI and EASMI.

	HL	CL	DS
WCI	0.489***	0.550***	-0.172
EASMI	0.511***	-0.001	-0.591***

Correlation significance levels (two-tailed test): *** P < 0.001.

Summary: at HL (on the eastern boundary of the Alxa Plateau), both EASMI and WCI influenced the alternation between wet and dry; at CL (on the southern boundary of the Alxa Plateau), climate was mainly influenced by the EASM. At DS (on the western boundary of the Alxa Plateau and the middle part of Hexi Corridor), climate was mainly influenced by the westerly winds.

4. Discussion and conclusions

4.1 Climate changes indicated by regional chronologies

Our chronology-climate response analysis (Fig. 4) showed that the radial growth index of Qinghai spruce in the HL, CL and DS mountains were a good record of regional climate changes around the Alxa Plateau (Fig. 3). On the interannual scale, the three regional chronologies noted that the extreme drought years of 1928, 1957 and 1981 were shared by two or more locations, as were the drought years of 1854, 1861, 1916, 1926, 1947, 1966, and 2001 (Fig. 3 and Table 3).

We note that drought was also reported by other tree-ring studies for these regions (Chen et al., 2016), also for the Qilian Mountains (Zhang et al., 2011; Zhang et al., 2017). Several other drought years (1854, 1884, and 1925–1928) were also seen in the dry-wet climate history (PDSI) (recorded by tree-ring-widths) in the nearby area of

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Mount Hasi, which lies on the edge of the regions most influenced by the EASM (Kang et al., 2012).

The drought years of 1823, 1833, 1854, 1877, 1883–1885, 1895, 1908, 1971, 1992, and 2003 seen in results for the Alxa Plateau are also seen in twelve tree-ring reconstructed drought series for the Qilian Mountains (an area mainly influenced by westerly winds) (Zhang et al., 2011). We also note that wetter years seen in our three regional chronologies were also seen in results from the Hasi and Xinglong Mountains, which are also on the edge of the area influenced by the EASM) (Fang et al., 2009; Kang et al., 2012).

If we compare our results with those seen for the EASM-affected areas at Mount Guiqing, 1820–2005 (Fang et al., 2010), we noted that only three of the eight drought years in that area (1928, 2000, and 2001) were seen in our three chronologies. We also noted results from the westerly-influenced area at Mount Tianshan (Jiang et al., 2017). The wetter years of 1846, 1903 and 1942 at DS were also extreme wet years at Mount Tianshan. Two wet years, 1848 and 1959, recorded at DS are either one year earlier or one year later than extremely wet years at Mount Tianshan, which might suggest some correlation. Drier years at DS (1884, 1947 and 1951) are one or two years later than the extremely dry years at Mount Tianshan. This suggests that these phenomena could be related to broader changes in the extent and strength of the atmospheric circulation.

On a broader (interdecadal) scale, an extreme drought period in 1920s–1930s was shared by much of northern China (Liang et al., 2006; Fang et al., 2009; Fang et al., 2010). This is the same drought that we note our chronologies for HL, CL and DS (Liu et al., 2002; Chen et al., 2010; Fan et al., 2012; Liu et al., 2013; Zhang et al., 2015). A drought in 1890–1900 was noted by dendrochronological studies and regional history documents (Yuan, 1994; Ma et al., 2003; Cai and Liu, 2007).

Ma and Fu's (2006) study showed a broad shift towards a drying climate in 1977–78 (eastern area in northwestern China, also northern China). Several other dendrochronological studies showed a combination of high temperatures and low precipitation in the late 1970s to early 1990s (Zhang et al., 2005b; Cai and Liu, 2007; Cai, 2009).

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This same drought was seen at DS, if somewhat later and for a shorter time. We also noted its effects at HL and CL. This would be consistent with the increased humidity of the climate in the eastern region of Northwest China (the EASM-influenced region experiencing >400 mm precipitation). This region would include Mount Xinglong (Fang et al., 2009; Chen et al. 2015), the easternmost part of the Qilian Mountains, and Mount Guiqing (Fang et al., 2010).

The wet period that lasted from the 1940s to the early 1970s has been recorded by several tree-ring-width chronologies covering HL, CL, and DS (Liu et al., 2004; Liu et al., 2005; Gao et al., 2006; Cai, 2009; Chen et al., 2010). Regional history documents also record some severe floods disasters in this period (Yuan, 1994). We also see this wet period in tree-ring-width chronologies from Mount Xinglong (Fang et al., 2009; Chen et al. 2015) and Mount Guiqing (Fang et al., 2010).

The wet period in the 1830s–1840s evident in the chronologies in Xinglong Mountain (Fang et al., 2009)(Chen et al. 2015) and Guiqing Mountain (Fang et al., 2010) corresponds to the dry period of DS. The wet period in the 1830s–1840s corresponds to the dry period of HL and CL, and to the wet period of DS. The observed phenomena can be attributed to differences in the extent and intensity of EASM and westerly atmospheric circulations.

4.2 Influence of atmospheric circulations and their interaction on climate change in the Alxa Plateau

~~It is known that that~~ water vapor carried by the westerly winds will extend southward to the northern part of Qinghai, the Hexi region of Gansu, the northern part of Ningxia, and the northern part of Shaanxi Province, sometimes passing through the northern border of the Xinjiang region (Li et al., 2012). The area bounded by 35° and 55°N, 110°E and 140°E seems to key to fluctuations in the westerly winds. This in turn affects the distribution of rain belts in summer. Its mean WCI are weaker positively to the rainfall in the middle of Yellow River Basin and its northern regions (Yan et al., 2007).. The results showed that the middle ring-width index group of Qinghai spruce in the three sample sites, which are located in the key area for interaction between wind and monsoon, presented weaker negative correlation with WCI on the interannual scale (Fig. 6).

The EASM boundary zone has a greater influence on precipitation at higher latitudes and thus on vegetation growth. This boundary zone can fluctuate due to the interannual variability of the EASM and the westerlies. There may be lagging effects at the mid-latitudes (Ou and Qian, 2006). Again, we note that on an interannual scale, there is much variation in the strengths and interactions of the EASM and westerly circulation and thus on climate in our three study regions (Fig. 6).

Sun et al. (2019) showed that when the westerly circulation strengthens, high latitude air pressure drops across the entire Asian continent. Siberian high pressures and the EASM are weakened. The south of the cold air activity is also correspondingly weakened. That is not conducive to the north and south of the cold and warm air vapor exchange to form precipitation. When the lower of the WCI and weakened latitudinal circulation, the meridional circulation will strengthen, which favors the exchange of warm and cold air between the north and south to form precipitation.

Yang et al. (2019) proposed that in years with weak summer westerlies in the middle latitudes, the upper-level jet stream tends to shift southward. This southward displacement of the jet stream, coupled with weakened lower-level divergence, hampers the northward transport of warm air into the southwestern region. Consequently, this leads to reduced availability of water vapor sources and ultimately results in diminished summer precipitation within the transitional zone of typical monsoon activity. If the jet stream moves northward, precipitation increases.

Xu et al. (2010) wrote that in the middle Qilian Mountains the westerly winds affect precipitation directly, while the EASM only indirectly affects precipitation. When the westerly winds are stronger, the high precipitation zone moves northwestward; when they are weaker, the zone moves.

At DS, radial growth showed weak negative correlations with higher WCI and also higher, middle, and lower EASMI groups (Figs. 6;7). At HL, when high chronology indices are positive they are significantly correlated with westerly circulation; when they are negative they significantly correlate with EASM (Figs. 6; 7). At CL, which lies further to the south than HL, a higher EASMI leads to a more humid climate. Other effects are more complicated: for example, the higher and lower ring-width index groups, associated with extreme dry and wet climate years, have weak negative correlations to EASMI (Figs. 6; 7). Jiang et al. (2019) published the results of their hydrogen and oxygen isotope studies of surface water at more than 3,000 sampling sites in northern China. They showed that surface water recharge in the DS Mountains is due to the westerly winds; recharge in the CL Mountains is due to the EASM. The HL Mountains, in contrast, sit at the boundary of the EASM; water recharge there is due to both the EASM and the westerlies.

Jiang and Wang (2005) notes significant declines in the EASM in the mid-1960s and mid-1970s, which led to decline in the radial growth of Qinghai spruce in our study area. The effect of the latter declined period was much greater than that of the former, whatever the intensity or duration. The effects of these declines were stronger at CL and DS than at HL. In the mid-1970s, EASM retreat had stronger negative effects at CL and then at HL. However, decline in the EASM proved to be a facilitator of radial growth at DS (Fig. 8).

In the same period the westerly circulation also retreated. The EASM retreated again in 1990s, while the westerly winds strengthened. This resulted in a drier climate in the CL Mountains. However, it was also correlated with fluctuating wet periods at HL and a weak wet period at DS. The above results, to a certain extent, support our view on the driving mechanisms of climate change in the three study areas, especially in the DS Mountains.

When we look at this area on a geologic scale, we learn that the westerly circulation strengthened during the Ice Age. Westerly jet streams moved southward to about 35°N. When the westerly winds weakened in the Interglacial Age, the westerly jet streams moved northward to ~37°N (Sun et al., 2003). A study of Holocene lake level evolution in the ancient Zhuye lake, central Alxa Plateau, showed that lake-level change was subject to the combined effects of EASM and the arid climate of Central Asia (Li, 2009) This result further illustrates the complexity of lake evolution and climate change in the EASM marginal zone.

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The westerly circulation also interacts with the monsoon on the Tibetan Plateau, which has a profound effect on the climate of the Asian monsoon region as well as the global climate (Qu et al., 2004). There has also been much research using proxy indicator cycles indicating that our study area is also influenced by large-scale climate and ocean-atmosphere changes on interannual and interdecadal scales, such as the North Atlantic Oscillation (NAO), Pacific Decadal Oscillation (PDO), El Niño-Southern Oscillation (ENSO), and sunspot activity (Gou et al., 2015a; 2015b; Liu et al., 2016; Wang et al., 2017).

However, all of the above-mentioned large-scale climate and ocean-atmosphere changes affect the EASM and westerly circulation through different pathways (Li. 2009), which in turn have various effects on the northwestern edge zone of the EASM and the zone of interaction between the two major atmospheric circulations.

In conclusion, based on the analysis of the regional chronologies collected in the HL, CL and DS mountains that are arrayed around the Alxa Plateau, we can safely assert that the radial growth of Qinghai spruce in the study area is mainly affected by regional precipitation. This precipitation varies constantly over time and space, primarily influenced by the interactions between two atmospheric circulation systems, EASM and westerly winds. At HL, both of these atmospheric circulation systems play a significant role in shaping climate changes. At CL, the climate is mainly influenced by the EASM. At DS, climate is more heavily influenced by the westerly circulation.

In the future, it is to be hoped that more refined, smaller scale research can be done on the climate history in the deserts of the Alxa Plateau. Such research may finally to provide a theoretical basis to explain regional climate driving mechanisms and thus enable better desertification controls.

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