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Climate of the Past

Tree ring-based February–April temperature reconstruction for Changbai Mountain in Northeast China and its implication for East Asian winter monsoon

H. F. Zhu[1,2], **X. Q. Fang**[2], **X. M. Shao**[1], **and Z. Y. Yin**[3]

[1]Institute of Geographic Sciences and Natural Resources Research, Chinese Academy of Sciences, Beijing, China
[2]School of Geography, Beijing Normal University, Beijing, China
[3]Dept. of Marine Science and Environmental Studies, University of San Diego, San Diego, USA

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Abstract. Long-term climatic records are scarce in the northeast Asia for understanding the behavior of the East Asian Winter Monsoon. Here we describe a 250-year February–April temperature reconstruction (TCBM) based on tree-ring widths of Korean Pines from the Changbai Mountain area, Northeast China. The reconstruction can account for 45.7% of the temperature variance in the instrumental period (1953 to 2001). Four cold periods including 1784–1815, 1827–1851, 1878–1889 and 1911–1945, and two warm periods of 1750–1783 and 1855–1877 were identified before the instrumental period. Four shifts were also detected at 1781, 1857, 1878 and 1989. Good agreements between TCBM and other temperature records of East Asia suggest that the reconstruction is of good reliability and captures the regional cold/warm periods of East Asia. Moreover, TCBM shows negative correlations with the instrumental or proxy-based EAWM intensity records. The known weakening of the EAWM in the late 1980s is in agreement with the shift at 1989 in TCBM. These comparisons suggest that the February–April temperature reconstruction may be a good indicator of the EAWM intensity.

1 Introduction

Long temperature records are critical in studying the long-term behavior of the East Asian Winter Monsoon (EAWM), which is a key climatological feature in the northern hemisphere winter. Tree rings, as a continuously and annually resolved proxy, have been widely used to reconstruct past climate in North America and Europe (e.g., Briffa et al., 2004; Cook et al., 1999). In recent years, there are also several dendroclimatic reconstructions in East Asia (EA), such as in China, Korea and Japan (e.g., Choi et al., 1994; Shao and Wu, 1997; Yonenobu and Eckstein, 2006; Zhu et al., 2008). However, there still remains a lack of long temperature proxies for the northeast Asia, where winter climate is predominantly influenced by the intensity of the EAWM.

The purpose of this study is to reconstruct winter temperature based on tree-ring widths of Korean Pines from the Changbai (also known as Baekdu) Mountain area in Northeast China (Fig. 1). This reconstruction may also be useful for studying the long-term behavior of the EAWM.

2 Materials and methods

The study area is located in the Changbai Mountain, a volcano in Northeast China, where the climate is affected by the East Asian monsoon (Fig. 1). At the meteorological station in Dunhua (43°22′N, 128°12′E, 52 m a.s.l.), January (mean temperature of −16.9°) and July (19.9°) are the coldest and the warmest month, respectively (Fig. 2). The multi-year mean of annual precipitation amounts to 630 mm, with 88.4% of the annual precipitation falling during the growing season approximately from April to September.

Tree-ring increment cores were sampled from four sites of trees in the Changbai Mountain (Table 1). The trees sampled were all from the dominant species, Korean Pine (*Pinus koraiensis*), of closed-canopy temperate forest. The cores were mounted, crossdated and measured through traditional process (Fritts, 1976; Stokes and Smiley, 1968). Then we used the COFECHA program to evaluate the accuracy of the crossdating and measurements (Holmes, 1983).

Correspondence to: X. Q. Fang
(xfang@bnu.edu.cn)

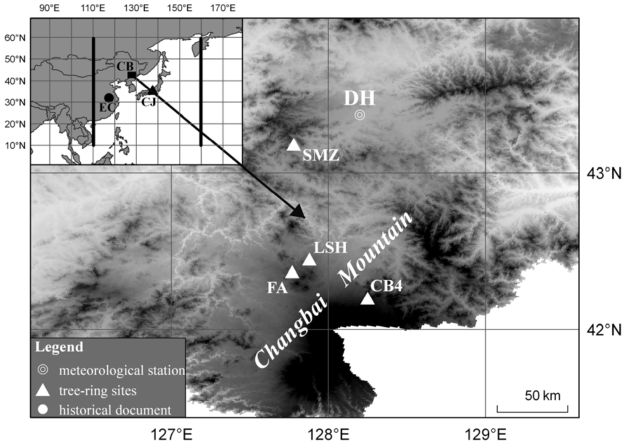

Fig. 1. Map showing the tree-ring sites, meteorological station, and other locations mentioned in text. CJ: Feburary-April temperature (Yonenobu and Eckstein, 2006); EC: Winter-half year temperature (Ge et al., 2003). The grids for the EAWMI are marked by the solid bars in the panel upper left.

Table 1. Site information and tree-ring chronology statistics[a].

Site	Lat	Lon	Elev	TS	MSL	C/T	Ac1	Rbt	$Y_{EPS>0.85}$
CB4	42°12′	128°15′	1188	1651–2002	229	27/14	0.461	0.260	1810
LSH	42°27′	127°53′	870	1743–2002	180	40/21	0.349	0.338	1825
FA	42°22′	127°46′	940	1689–2002	206	40/27	0.444	0.251	1765
SMZ	43°11′	127°47′	765	1742–2002	199	57/29	0.531	0.232	1800

[a] Lat=site latitude; Lon=site longtitude; Elev=site elevation in meters; TS=time span; MSL=median segment length; C/T=number of cores/trees; Ac1=first order autocorrelation; Rbt=correlation between trees; $Y_{EPS>0.85}$=year of EPS>0.85.

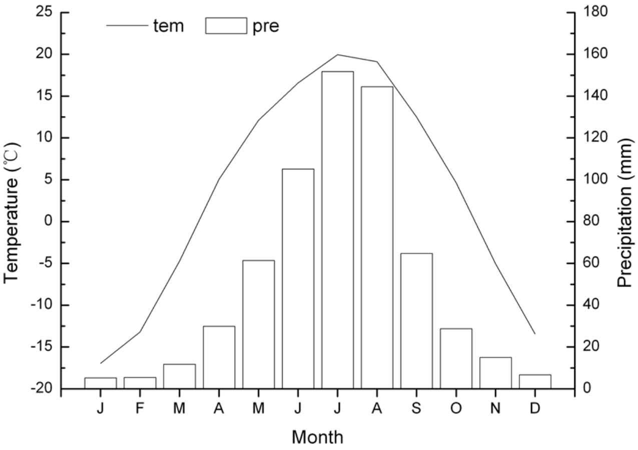

Fig. 2. Climate diagrams (1953–2002) at the Dunhua meteorological station.

Tree-ring chronologies were developed using the ARSTAN program. A 180-year cubic smoothing spline was used to remove the long-term growth trends of raw ring-width series. The resulting ratio series were then averaged to build the site chronologies. To evaluate the reliability of the chronologies, the 50-year moving Expressed Population Signal (EPS) with a 25-year lag (Wigley et al., 1984) was calculated (Table 1). Significant correlations from 0.52 to 0.63 exist among all site chronologies over the 1825–2002 common period (EPS>0.85). This allowed us to build a regional chronology (CB) by averaging all the individual ratio series of the four sites. CB is considered to be reliable after 1750 AD, when the sample depth is 20 series, although the EPS is 0.80, slightly lower than the commonly used level of 0.85.

The relationships between tree-ring indices and climate variables were analyzed using correlation analysis for the period 1953–2002. The climate variables include monthly mean temperature and total precipitation from the Dunhua meteorological station (Fig. 1). The data from previous October to current October were used for the analysis, as the climate conditions of the previous year may have effects on tree-ring growth of the current year (Fritts, 1976).

To reconstruct the past climate variations, the instrumental records were regressed against the regional chronology (CB). In consideration of the lag effects of climate on tree growth, concurrent chronologies (at t) and those lagged at $t+1$ were used as independent variables. Several seasonal mean temperature subsets were examined from previous October to current September according to results of correlation analysis (Fig. 3). The leave-one-out cross-validation method (Michaelsen, 1987) was used to choose the most successful season, since the instrumental data set from 1953–2002 was too short to be divided into two subsets for independent calibration and verification tests. The testing statistics include variance explained, adjusted variance explained, sign test of the first difference (SN1), sign test of the raw data (SN2), the reduction of error (RE) and Pearson's correlation coefficient (Fritts, 1976; Cook and Kairiukstis, 1990).

3 Results

Figure 3 illustrates the correlation coefficients between the regional chronology (CB) and the monthly climate data. Ring width shows significantly ($p<0.01$) positive correlations with the temperature in previous October, current February to April, and September. However, there were no significant correlations between CB and precipitation.

The transfer function between February-April temperature and tree-ring chronologies (t and $t+1$) acquired the best calibration and cross-validation statistics (Table 2). The final calibration model ($\mathrm{Tem}=-10.9+2.35\mathrm{CB}_t+4.12\mathrm{CB}_{t+1}$) explained 46.2% of the total variance of the instrumental records during 1953–2002 (Fig. 4), and 37.1% in the leave-one-out cross validation. The positive RE indicates good predictive skill of the regression model. The lower SN1 and highest SN2 suggest that the strength of the calibration lies more in the lower-frequency agreement between the reconstruction and the instrumental records.

Figure 5a shows the interannual to multi-decadal variations of the reconstructed February–April temperature (TCBM) for the Changbai Mountain area since 1750. Several extended warm and cold periods (>11 years) were identified before the instrumental period (1953–2002) according to the 11-year moving averages of the reconstruction and the long-term mean (1750–2002, −4.05°). Warm periods occurred around 1750–1783 (W1) and 1855–1877 (W2), while cold periods at 1784–1815 (C1), 1827–1851 (C2), 1878–1889 (C3), and 1911–1945 (C4) (Fig. 5a). Intervention analysis (Rodionov, 2004), with the mean values of 15-year periods being compared with values on either side of each year, was also applied to the TCBM to identify any significant shifts ($p<0.05$) (Fig. 5a). Four shifts were identified at 1781, 1857, 1878 and 1989.

4 Discussion

The results of correlation analysis (Fig. 3) suggest that the radial growth of Korean Pine in Changbai Mountain is mainly limited by the pre-growth season temperature. A positive effect of winter temperature on tree growth was also reported for other temperate coniferous forests, such as in central Japan and in the Hudson River Valley (Pederson et al., 2004; Yonenobu and Eckstein, 2006), and from timberline forests on the east and northeast Tibetan Plateau (Brauning, 2001; Liang et al., 2009; Liu et al., 2007; Zhu et al., 2008). Warm winters may lead to damage to the roots and positive carbon gains for conifer trees when their leaves are not frozen (Brauning, 2001; Chabot and Hicks, 1982; Havranek and Tranquillini, 1995; Pederson et al., 2004). In addition, the warmer temperature of previous October may allow an extended period of carbon accumulation for the growth of next year. However, the positive effect of current September is difficult to interpret because the cell division and enlargement have probably ceased at the end of the growing season (April–September).

Our winter temperature reconstruction captures a warming trend in the recent decades in northeast China. A warming trend in winter was also recorded by tree growth on the Tibetan Plateau (Liu et al., 2007; Zhu et al., 2008; Shao and Fan, 1999). However, a study based on tree-ring width in central Japan (Yonenobu and Eckstein, 2006) did not track such a warming trend, probably due to the consequence of anthropogenic SO_2 emissions. Increased summer temperatures were recorded by tree growth in other areas of EA, such as Mongolia (D'Arrigo et al., 2000), central Korea (Choi et al., 1994) and Tibetan Plateau (Liang et al., 2008). Unlike apparent loss of temperature sensitivity in some northern forests (D'Arrigo et al., 2008; Briffa et al., 1998), there appears to be no such shift in response at Changbai Mountain, indicating the trees' continued response to temperature.

The cold/warm periods in our reconstruction (TCBM) (Fig. 5a) are generally consistent with other temperature reconstructions in East Asia. The warm period of W1 and cold periods of C1 and C2 were all indicated by a historical documented winter-half year temperature record of East China (EC: Ge et al., 2003) (Fig. 5c) and a tree ring-based February–April temperature reconstruction of central Japan (CJ: Yonenobu and Eckstein, 2006) (Fig. 5b), although they are more than 1000 km away from our study site. Moreover, there are evidences from historical documents to suggest that cold climate occurred in the neighboring Heilongjiang Province during 1806–1810 and 1821–1850 (Gong et al., 1979), and cold winters or episodes of heavy snowfall occurred in Japan during 1807–1819 and 1826–1836 (Fukaishi

Table 2. The calibration and verification statistics of different seasonal temperature models. The leave-one-out method was used for verification during the period of 1953–2001

	Calibration		Verification			
Season	R^2	R_a^2	SN1	SN2	RE	r
P10C4	0.331	0.301	27	34*	0.226	0.475*
P10C9	0.392	0.365	24	35*	0.308	0.555*
C2C4	0.457	0.439	30	41*	0.376	0.613*
C2C3	0.374	0.360	31	36*	0.312	0.558*
C2	0.393	0.380	35*	34*	0.326	0.571*

* Significant ($p<0.01$).

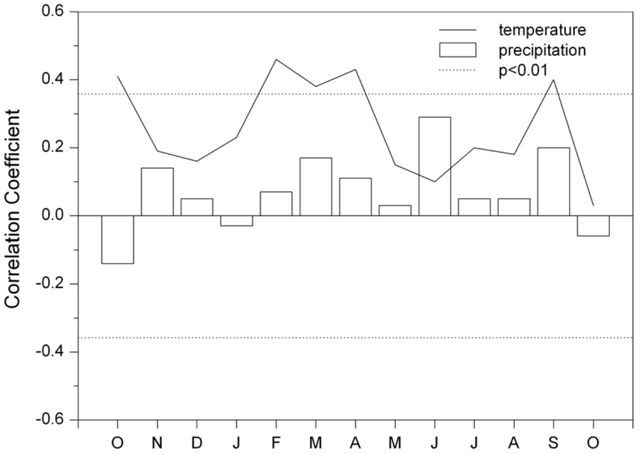

Fig. 3. Correlations between the regional chronology (CB) and monthly temperature and precipitation from previous October to current September at Dunhua station (1953–2002).

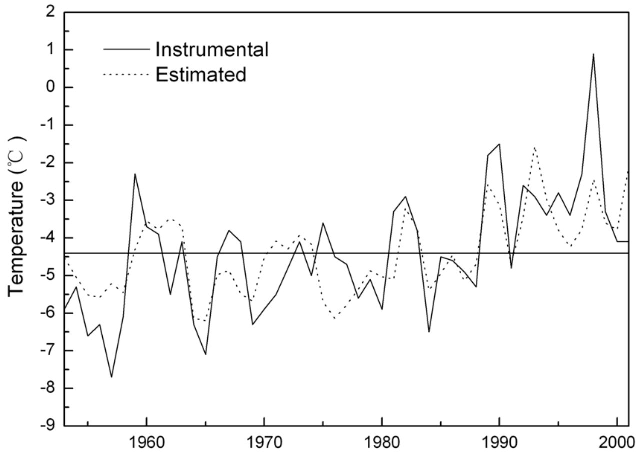

Fig. 4. Comparison between the instrumental and reconstructed February–April temperatures at Dunhua Station.

and Tagami, 1992). Despite their differences in lengths and magnitudes, the warm period W1 agrees well with the EC record. In addition, C3 is detected in both EC and CJ records, although it is not as long as in those studies. The cold period of 1910–1940s (C4) is consistent with the low February–April temperature in CJ. Overall, these agreements suggest that our reconstruction is of good reliability, and captures regional cold/warm variations in EA on the decadal scale. A former reconstruction of January–April maximum temperature for the Changbai area was based mainly on *Larix olgensis* and *Picea jezoensis* (Shao and Fan, 1999). However, due to the removing of persistence by autoregressive modeling (Cook and Kairiukstis, 1990) of their tree-ring data, little low-frequency variations were retained in the record for a comparison with the current reconstruction.

In addition, the coldest years 1837–1839 in the past 250 years in northeast China were also indicated in other proxy records of EA. In Japan, the winters were also cold in 1838–1839 (Yonenobu and Eckstein, 2006), and the occurrence frequencies of the winter monsoon weather patterns are high (Hirano and Mikami, 2008). Tree-ring records in central Korea indicate that it was cool in early summer during 1835–1844, with the lowest temperatures occurring in 1841–1842 (Choi et al., 1994). The consistency implies that there may be synchronous occurrence of extremely cold winters or even cool summers in EA.

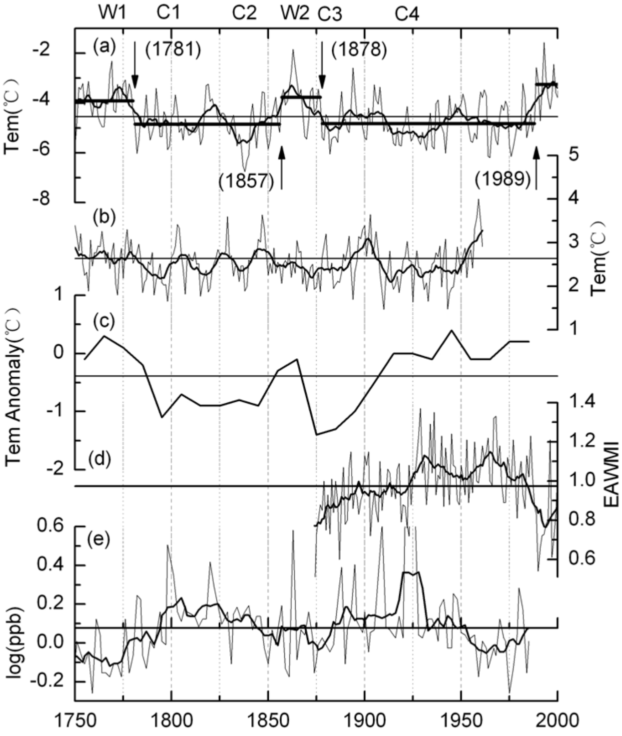

Fig. 5. Comparisons of the February–April temperature reconstruction in this study with other temperature reconstructions and East Asian Winter Monsoon (EAWM) records. **(a)** February–April temperature reconstruction of this study; **(b)** February–April temperature reconstruction in central Japan (Yonenobu and Eckstein, 2006); **(c)** winter-half year temperature reconstruction in East China (Ge et al., 2003); **(d)** East Asian Winter Monsoon (EAWM) index (Guo, 1994); **(e)** Non-seasalt potassium concentration (NSSK) in ice core from central Greenland (Meeker and Mayewski, 2002). The thick black curves are the 11-year moving averages and the horizontal lines are the long-term means. The years identified as shifts of the temperature reconstruction in this study are marked by arrows, and their long-term means are represented by the horizontal thick lines in (a).

We correlated the reconstructed temperature against the EAWM index (EAWMI), which is based on the sum of zonal SLP differences (110° E–160° E) computed at 10° intervals over 10–60° N (Guo, 1994). The reconstructed temperature correlates with the winter (December–February) EAWMI at $r=-0.342$ ($p<0.001$) over the period 1874–2000, and at -0.503 ($p<0.001$) over the period 1951–2000 (Fig. 5d). The relatively lower correlation over the whole time span may be due to the scarcity of pressure data in earlier periods, which introduced uncertainties in the EAWMI. Strong EAWM years are characterized by a frequent occurrence of cold surges that brings cold air from the Siberian/Mongolian High (SH), thus producing the below-normal temperatures in the study area (Ding, 1990; Guo, 1994; Jhun and Lee, 2004; Yin, 1999). In addition, the shift identified at 1989 in our reconstruction is also consistent with the weakening of EAWM in the late 1980s (D'Arrigo et al., 2005; Nakamura et al., 2002; Tsunoda et al., 2006; Wang and Jiang, 2006).

The relationship between the reconstructed February–April temperature in the Changbai Mountain area and the EAWM is also verified by other proxy records. The SH is the source area of EAWM, and its intensity is significantly positively correlated with the EAWM intensity, thus the intensity of SH can also be used to indicate the intensity of EAWM (Guo, 1994; Jhun and Lee, 2004; Wu and Wang, 2002). The non-seasalt potassium (NSSK) in the ice core from central Greenland was suggested to be transported from central Asia, and has been used to reconstruct the spring (March–May) SH intensity (Meeker and Mayewski, 2002). The high/low NSSK content variations are in good match with cold/warm periods in our reconstruction (Fig. 5e). Moreover, D'Arrigo et al. (2005) have developed a difference index between the normalized SH index and North Pacific index based on tree-ring records collected over broad regions of Eurasia and northwest America. Although this index provides more information on the interannual variability of EAWM, it also demonstrates stronger winter monsoon during ∼1825–1851 and 1920–1940s, which is in agreement with the cold periods (C1 and C4) of our reconstruction. In addition, C2 is consistent with a culmination of the EAWM from ∼1825 to 1841 recorded by historical documents in Japan (Hirano and Mikami, 2008).

5 Conclusions

In this study, we reconstructed a new February–April temperature record based on tree-ring widths of Korean Pines from the Changbai Mountain in Northeast China, which revealed variations at interannual to multi-decadal scales over the past 250 years. The series captures well the recent warming trend in northeast China. The cold/warm periods identified in the reconstruction are in good agreement with other reconstructions in East Asia, suggesting that these are significant periods with regional implications. In addition, our reconstruction shows negative correlations with instrumental or proxy-based EAWM intensity records. The known regime shift of EAWM in the late 1980s is captured in the reconstruction, and similar shifts have occurred over the past 250 years. On the whole, these comparisons suggest that our reconstruction is of good reliability and may also be used as an indicator of the EAWM intensity. Hence, more efforts should be paid to extend the reconstruction by collecting more old trees or excavated wood samples buried under the volcanic ashes.

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