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Are Karakoram temperatures out of phase compared to hemispheric trends?

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Abstract In contrast to a global retreating trend, glaciers in the Karakoram showed stability and/or mass gaining during the past decades. This "Karakoram Anomaly" has been assumed to result from an out-of-phase temperature trend compared to hemispheric scales. However, the short instrumental observations from the Karakoram valley bottoms do not support a quantitative assessment of long-term temperature trends in this high mountain area. Here, we presented a new April–July temperature reconstruction from the Karakoram region in northern Pakistan based on a high elevation (~3600 m a.s.l.) tree-ring chronology covering the past 438 years (AD 1575–2012). The reconstruction passes all statistical calibration and validation tests and represents 49 % of the temperature variance recorded over the 1955–2012 instrumental period. It shows a substantial warming accounting to about 1.12 °C since the mid-twentieth century, and 1.94 °C since the mid-nineteenth century, and agrees well with the Northern Hemisphere temperature reconstructions. These findings provide evidence that the Karakoram temperatures are in-phase, rather than out-of-phase, compared to hemispheric scales since the AD 1575. The synchronous temperature trends imply that the anomalous glacier behavior reported from the Karakoram may need further explanations beyond basic regional thermal anomaly.

Keywords Paleoclimate · Climate variability · Tree rings · Dendrochronology · Karakoram Anomaly

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1 Introduction

With 7820 glaciers covering an area of 17,946 km^2, the Karakoram region in northern Pakistan comprises the most glaciated area outside the polar regions. Unlike the retreating status of most glaciers elsewhere, the Karakoram glaciers are stable or re-advance over the past 20 years (Bolch et al. 2012), which has been coined as the "Karakoram Anomaly" (Hewitt 2005; Kumar et al. 2015). Anomalous decreasing summer temperatures, compared to globally warming trends, as well as increasing winter precipitation were assumed to be the main climatic drivers of the Karakoram Anomaly (Hewitt 2005; Kapnick et al. 2014). However, most assessments were based on the relatively short meteorological observations (typically <50 years), recording conditions in the deep and semi-arid Karakoram valleys

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rather than in the high, glaciated regions (Hewitt 2011, 2014), limiting our understanding of mountainous climate variability on longer timescales.

Tree rings, as an annually resolved and quantitative climate proxy, have been used to investigate past temperature variability over the Tibetan Plateau, Himalaya, and neighboring regions (Cook et al. 2003; Deng et al. 2014; Liang et al. 2010; Thapa et al. 2014; Lv and Zhang 2013; Wang et al. 2014; Yadav et al. 2011; Zhang et al. 2015). In the Karakoram area, several studies revealed tree rings to be sensitive to changing temperatures (Ahmed et al. 2011; Esper 2000; Esper et al. 2002b, 2007; Zafar et al. 2015). Recent work by Zafar et al. (2015) presented a rigorously calibrated summer (June–August, JJA) temperature reconstruction extending over the past 500 years based on *Picea smithiana* and *Pinus gerardiana* tree-ring data. This reconstruction indicated that temperature variations in the Karakoram were out-of-phase compared to the hemispheric trends. In contrast, the Northern Hemisphere shows general warming trends during the twentieth century (Esper et al. 2002a; IPCC 2007; Wilson et al. 2016). However, the Zafar et al. (2015) reconstruction is based on a negative association between tree-ring data and summer temperatures, indicating that moisture stress is the key driver of tree growth at the study sites.

The objective of this study is to present a new high-elevation temperature reconstruction for the Karakoram region based on a positive response of tree growth to temperature, and to compare the retained temperature trends with previous attempts based on inverse growth/temperature correlations. We expect that the Karakoram temperature variability is synchronous with the trends at larger spatial scales and assess this by comparison with reconstructions of Northern Hemisphere and Northern Hemisphere extra-tropical temperature variability.

2 Materials and methods

2.1 Study area

Our study area is situated in the Bagrot valley in the Karakoram region of northern Pakistan (Fig. 1). It covers an area of 452 km^2, characterized by extreme relief from 1500 to 7788 m a.s.l. at the Rakaposhi summit (Mayer et al. 2010). Precipitation in this region changes gradually with altitude, semi-arid in lower elevations (<1500 m) and wet in higher elevations (>4000 m) (Mayer et al. 2010; Winiger et al. 2005). The precipitation in the study region is estimated to increase to 1000+ mm at the upper treeline in 4000 m a.s.l. (Cramer 2000). In summer, the monsoon is contributing to the precipitation in this region, while westerlies dominate during the cold season (Hewitt 2005; Kapnick et al. 2014;

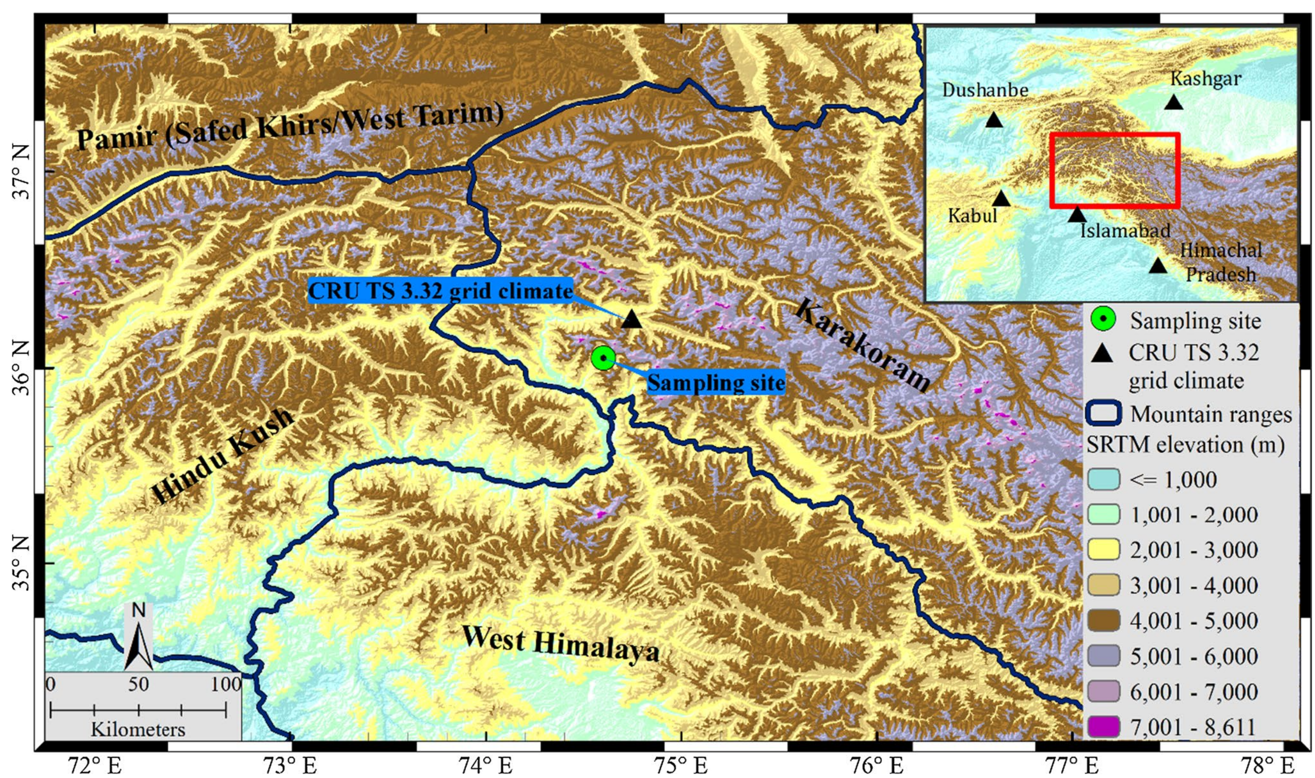

Fig. 1 Location of the tree-ring sampling site and the CRU TS 3.32 grid point in the Karakoram, northern Pakistan

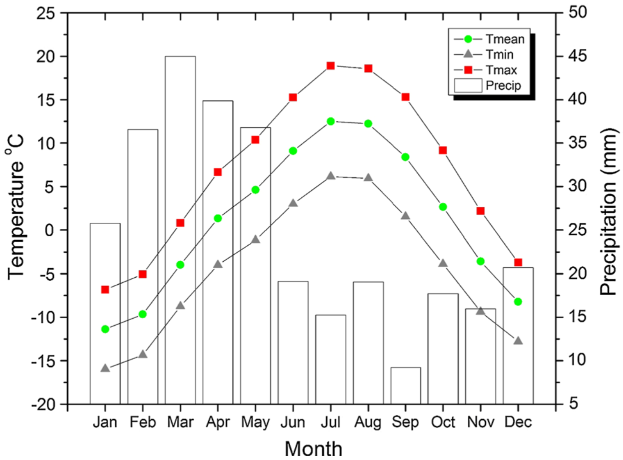

Fig. 2 Climate diagram of the gridded monthly climate data sets of CRU TS 3.23, covering the period of 1955–2013

Table 1 Site information and statistics for the tree-ring chronology

Study site	Bagrot (Karakoram)
Species of tree	*Pinus wallichiana*
Latitude	36° 01
Longitude	74° 39
Elevation range (m)	3550–3710
Cores/trees	60/33
Time span	1194–2013
Percent of missing rings (%)	0.17
Mean sensitivity	0.13
Standard deviation	0.22
Mean correlation between all series	0.45
Mean correlation between trees	0.44
Mean correlation within a tree	0.65
First order autocorrelation AC (1)	0.68
Signal to noise ratio (SNR)	42.5
Expressed population signal (EPS)	0.97

Kumar et al. 2015). The steep precipitation gradient, from the semi-arid valley bottoms to the humid summits, results from an inner-mountainous circulation system generating clouds and downpour at the slopes and summits, and persisting clear sky conditions above the valley centers (Weiers 1998). According to the gridded monthly climate data sets (36° 25′N, 74° 75′E from 1955 to 2013) of CRU TS 3.23 (Mitchell and Jones 2005), the mean annual precipitation, typically measured by stations situated in the valley bottoms, is 301 mm. Annual mean temperature is 1.2 °C, ranging from −11.4 °C in January to 12.5 °C in July (Fig. 2).

2.2 Tree-ring sampling and chronology development

Pinus wallichiana is a naturally distributed evergreen species in Pakistan, Bhutan, Afghanistan, China, Nepal and India (Orwa et al. 2015). It usually grows at an altitudinal range from 1800 to 3900 meters a.s.l (Singh and Yadav 2007). *P. wallichiana* is a dominant species at the timberline in the Karakoram reaching ages up to about 700 years (e.g. Astore-Rama; Ahmed et al. 2011). We here selected healthy *P. wallichiana* trees between 3550 and 3710 m a.s.l. near the upper treeline in the Bagrot valley, with no evidence of human disturbance or fire injuries (Fig. 1). A total of 60 increment cores were extracted from 33 living trees in May 2014 (Table 1). Tree-rings samples were air dried, mounted, and sanded using progressively finer sandpaper following established dendrochronological techniques (Fritts 1976; Stokes and Smiley 1968). The tree-rings were counted and tree-ring width (TRW) measured at 0.01 mm resolution using a LINTAB 5 measuring system. The samples were cross-dated by comparing the TRW visually using TSAP software and validated using the COFECHA computer program (Holmes 1983).

To preserve low-frequency climate signals in the final TRW chronology, we used ARSTAN program (Cook 1985) to calculate ratios between the raw measurement series and negative exponential functions (Cook and Kairiukstis 1990). For 18 cores we used 90-year cubic smoothing splines for standardization to avoid artificially increasing trends in the index series. We averaged the detrended series to produce a standard mean chronology (STD; Fig. 3). The statistical characteristics of this chronology over the period common to all trees (1800–2013) are shown in Table 1. To investigate whether there is a trend bias in the STD chronology calculated using ratios, we also produced a chronology (using the same negative exponential and spline fits) by calculating residuals subsequent to the application of a data-adaptive power transformation (Cook and Peters 1997).

The mean inter-series correlation is used to estimate coherence inherent to the chronology, and the Expressed Population Signal (EPS) is used to evaluate temporal changes of the reconstruction signal strength (Wigley et al. 1984). Both metrics were calculated over 50-year intervals lagged by 25 years. We considered an EPS > 0.85 threshold to define the period over which the chronology likely reflects the signal expressed by a theoretically infinite population. According to the 50-year running EPS values, the period after AD 1575 was considered for climate reconstruction (Fig. 3).

2.3 Climate signal assessment

In order to determine the climate signal inherent to the tree-ring data, we correlated the STD chronology with climate variables over the period 1955–2013. Since most of the weather stations in this area are situated on the valley floors, we used gridded data centered at 36°25′N and 74°75′E

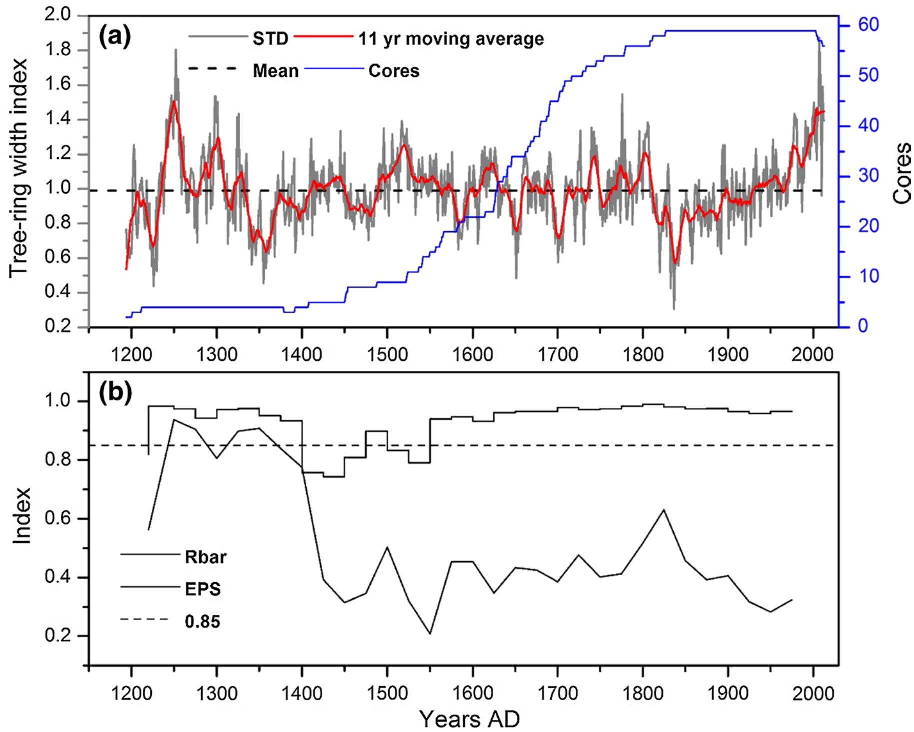

Fig. 3 **a** Standard chronology from 1194 to 2013, its long mean (*dotted line*), 11-year moving average (*bold red line*), and numbers of tree cores (*blue line*). **b** Expressed population signal (EPS), and running inter-series correlation (*Rbar*). Critical EPS level of 0.85 is highlighted with *dotted line*

Table 2 Calibration-Validation results of the Bagrot (Gilgit) April-July temperature reconstruction

	Calibration 1955–1983	Validation 1984–2012	Calibration 1984–2012	Validation 1955–1983	Calibration 1955–2012
r	0.60	0.74	0.78	0.58	0.70
R^2	0.36	0.55	0.61	0.40	0.49
R^2_{adj}	0.31	0.53	0.58	0.32	0.48
DW	2.30	1.56	1.34	2.31	1.90
F value	7.22*	33.14*	19.96*	13.85*	26.78*
ST		21 (21, 22)		25 (21, 22)	
FST		19 (20, 22)		16 (20, 22)	
RE		0.58		0.26	
CE		0.49		0.05	

r: Pearson correlation coefficient; R^2: variance explained; R^2_{adj}: adjusted variance explained; DW = Durbin–Watson statistic; ST: sign test with numbers of significant levels of 0.05 and 0.01; FST: the first-difference sign test; RE: reduction of error; CE: coefficient of efficiency

* Significant at $p = 0.01$

from the CRU TS 3.23 dataset (Fig. 2; Mitchell and Jones 2005) to capture the regional climate signals. Meteorological data from the nearby station in Gilgit are available since AD 1955, thus the CRU data are most consistent since 1955 for this region. The climate variable considered here include monthly mean (T_{mean}), maximum (T_{max}), and minimum temperatures (T_{min}), as well as monthly and annual precipitation sums. Considering that the climate in the year prior to tree-ring formation may have an effect on tree growth (Fritts 1976), Pearson correlations were calculated with the climate data from prior-year September to current-year October.

2.4 Calibration and validation

To reconstruct past temperature variations, we transferred the STD chronology into estimates of seasonal temperature using multiple linear regression. The highest temperature correlations were selected to develop a reconstructions model with minimum residuals. Current year growth (at lag t) and 1 year lag (t + 1) were utilized as independent variables, acknowledging lagging effects of climate on tree growth (Fritts 1976). This method, however, may reduce high frequency variability in the final reconstruction, particularly where there is opposite direction of change in growth between lag t and lag t + 1. Independent calibration and validation statistics were conducted to evaluate the predictive skills of the regression model in the periods 1955–1983 and 1984–2012, respectively. The following statistics were calculated considering the two, equally long sub-period (Table 2): explained variance (R^2), adjusted variance explained (R^2_{adj}), Durbin-Watson test (DW), F value, reduction of error (RE), coefficient of efficiency (CE), Sign test (ST), and first-difference sign test (FST) (Cook and Kairiukstis 1990; Fritts 1976). Sign-test is a nonparametric procedure quantifying the number of sign agreements and disagreements between the proxy and observation data (Fritts, 1976). We used the original (STD and observational) data as well as first-difference data to calculate ST. We also calculated spatial correlations between our reconstruction and the CRU TS 3.23 April–July minimum temperature field from 1955 to 2012 utilizing the KNMI climate explorer (http://climexp.knmi.nl), to evaluate its spatial representativity.

3 Results and discussion

3.1 TRW chronology characteristics

The long-lived *Pinus wallichiana* trees allowed the development of a chronology reaching back to the twelfth century (Fig. 3a). The inter-series correlation is r = 0.40 back in 1425, and enhance before likely due to the dominance of within-tree correlations (Fig. 3b). The mean correlation (0.44) between trees during the period 1800–2013 (Table 1) reveals significant common variance among the TRW series from these high elevation pines sampled in >3500 m a.s.l. The moving EPS values indicate that the chronology is representative back to 1575.

The chronology fluctuates around the mean from AD 1400 to 1800, followed by a substantial decrease in the first-half of the nineteenth century, culminating in the 1840s, and a subsequent long-term increasing trend towards present. This increasing trend is also present in the raw TRW chronology as well as the STD chronology produced using residuals (and power transformation) (Supplementary Fig. S1). It appears that potentially biasing effects of ratio calculation are of minor importance, as the other two STD chronologies show similar variance and trends. Consequently, we used the ratio STD chronology for further analysis, as it does not require extra transformation of the raw data.

3.2 TRW climate signals

The high-elevation *P. wallichiana* tree-ring chronology correlates positively with temperatures of each month from prior-year September to current-year October (Fig. 4). Significant ($p < 0.05$) and relatively high correlations are found from late spring to early summer of the current year (April–July) and from late autumn to early winter (October–November) of the previous year. Monthly minimum temperatures correlate better than mean and maximum

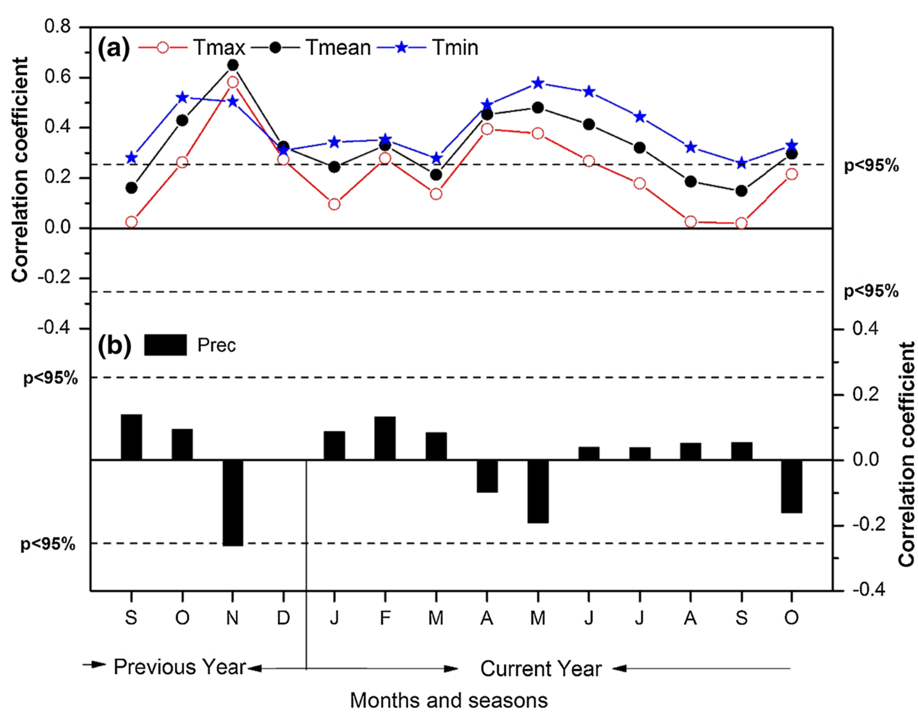

Fig. 4 Pearson correlations between the standardized TRW chronology and monthly maximum (T_{max}), mean (T_{mean}) and minimum (T_{min}) temperatures (**a**), and precipitation (**b**) from the previous-year September to current-year October. The *horizontal dotted line* at ±0.254 represents the significant level at $p < 0.05$

temperatures, particularly over the growing season from April to July. Weaker correlations are found between the STD chronology and precipitation, with only previous-year November precipitation showing a significantly ($p < 0.05$) negative relationship with tree growth. These results demonstrate that low temperature rather than precipitation is the fundamental constraining climate factor for tree growth towards the upper treeline in the Karakoram.

These findings are in contrast to an earlier study by Zafar et al. (2015) showing high elevation *Picea smithiana* and *Pinus gerardiana* growth in the Karakoram to be negatively related to summer temperature. Zafar et al. (2015) found significant negative correlations between seven TRW chronologies from the Gilgit and Hunza valleys with summer temperatures of both prior year and current year during the calibration period from 1955 to 2005. While our *Pinus Wallichiana* chronology reveals a significant positive correlations with late-spring and early-summer (growing season; April–July) temperatures, the difference with Zafar et al. (2015) might arise from the particular sampling site elevations. Our site is located about 300–700 m above the elevational range of the Zafar et al. (2015) sites. Their samples originate from 2850 to 3296 m a.s.l., far below the natural timberline. This also means that our site is 1.8–4.2 °C cooler if considering a temperature lapse rate of 0.6 °C/100 meters. Lower site temperatures typically reduce the evapotranspiration from trees and soil, while precipitation increase with elevation as a result of the inner-mountainous wind circulation system, thus moisture may not limit tree growth. On the other hand, limited warmth reduces tree growth through limiting tracheid division and enlargement (Rossi et al. 2008). In addition, low temperature in pre-growth season may damage buds and reduced root activity affecting tree metabolism (Körner 1999). At lower elevations, higher temperature may cause moisture stress to tree growth through enhanced evapotranspiration (Fritts 1976). Therefore, the cold conditions near the upper timberline in the Karakoram provides a physiological background for the temperature reconstructions. Selecting trees from sites near the upper treeline ecotone is critical for retrieving a temperature rather than a drought/precipitation signal.

3.3 Calibration, validation and temperature reconstruction

According to the tree-growth/climate signal assessment, we selected April-July minimum temperatures as the reconstruction target. The reconstruction explains 49 % of the variance of the climate records during the full 1955–2012 calibration period, and 48 % after adjusting for the loss of the degrees of freedom due to two predictors (Table 2). Positive RE and CE statistics, over the early (1955–1983) and late periods (1984–2012), reveal predictive skill of

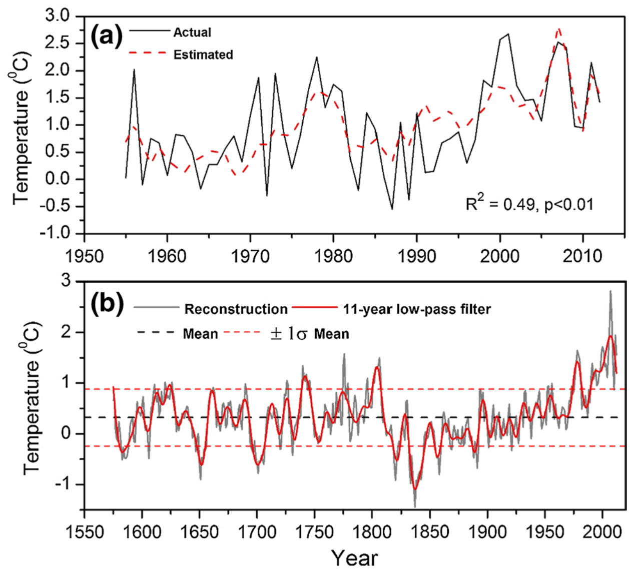

Fig. 5 **a** Comparison of instrumental and reconstructed April–July minimum temperatures (AMJJT) from 1955 to 2012. **b** The reconstructed temperature (T_{min}) shown together with $\pm 1\sigma$ (see "Methods" for detail), and 11-year low-pass filter

Table 3 Cold/warm periods during AD 1575–2012

No.	Cold periods	Warm periods
1	1578–1594	1595–1601
2	1602–1608	1609–1629
3	1630–1635	1636–1640
4	1641–1657	1658–1665
5	1666–1670	1671–1678
6	1679–1684	1685–1693
7	1694–1710	1711–1716
8	1717–1722	1723–1729
9	1730–1736	1737–1749
10	1750–1758	1759–1763
11	1764–1768	1769–1783
12	1784–1788	1789–1809
13	1810–1826	1894–1899
14	1830–1893	1929–1935
15	1900–1928	1939–2012

the reconstruction model (Cook and Kairiukstis 1990). The split calibration/validation approach also supports the stability and reliability of the reconstruction model over the full period. Both early and late period calibration equations are significant at $p < 0.01$, in spite of relatively lower R^2 (36 %) of the early compared to the late period (61 %). Application of the sign test supported this conclusion reaching $p < 0.05$ over the 1984–2012 and 1955–1983 periods. However, neither first-difference sign test passed the significant level of 0.05 (Table 2). Accordingly, the

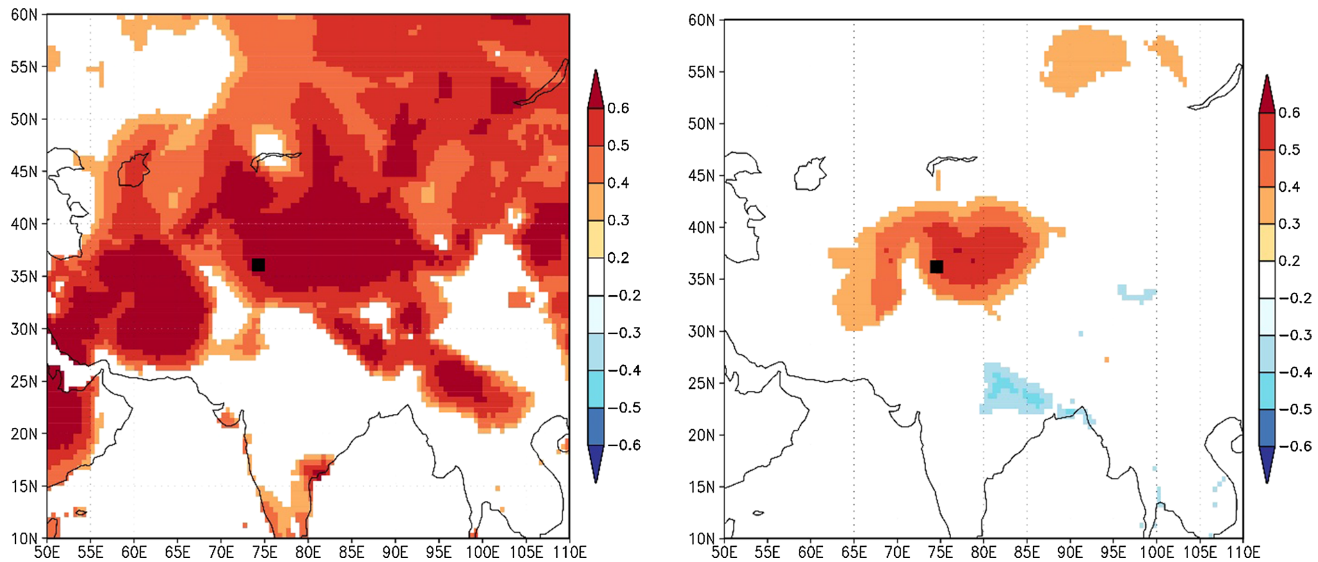

Fig. 6 Spatial correlation field of reconstructed April–July mean minimum temperatures (T_{min}) with gridded T_{min} from the CRU TS 3.23 dataset 1955–2012. The *left panel* show the results for the original data, the *right panel* for first-differenced data. Correlation analysis was performed using the KNMI climate explorer (http://climexp.knmi.nl/). *Black square* is the sampling site in the Bagrot valley. Correlations are not shown if significant level $p > 0.01$

good calibration results are supported by the coherent low frequency trends, rather than the year-to-year variations, similar to temperature calibration models detailed in Nepal (Cook et al. 2003) and the northeast Tibetan Plateau (Zhu et al. 2008). April–July temperatures were reconstructed back to AD 1575 using the transfer function of the full period: Temp $= -2.63 + 2.01\ TR_t + 0.983\ TR_{t+1}$. The reconstruction captures decadal scale fluctuations from the cool 1960s to the warm 1970s, as well as the cool 1980s to the warm 2000–2010s quite well (Fig. 5a).

The most prominent feature of the reconstruction is the warming trend from the early nineteenth century until present (Fig. 5b). It exhibits a long cold period from 1830 to 1893 and subsequent warming from 1939 to 2012. Application of a 11-year low pass-filter emphasizes cold and warm periods over the past 400+ years (Table 3). Over the past 438 years, 59 extremely warm and 61 extremely cold years are identified based on a 1σ criterion (0.56 °C) deviating from the long-term mean (0.32 °C). Thirty-two of the warmest years were identified in the twentieth and twenty-first century. The difference between the overall warmest (2007, 2.82 °C) and coldest years (1837, −1.45 °C) is 1.37 °C.

The comparison of the Bagrot reconstruction with gridded April-July minimum temperatures from CRU TS 3.23 shows significant ($p < 0.01$) correlations over the Karakoram, West Tian Shan, Tarimu Basin, Kunlun Mountain, Pamir and central-east Himalaya (Fig. 6). For the first-differenced data, the correlations are weighted eastward into the Karakoram, Pamir, and western Kunlun Mountains. The correlations decrease dramatically along the Karakoram and Himalaya towards the lower Indus Basin into central-southern Pakistan and the Indian sub-continent. These results suggest that the Bagrot reconstruction is representative for high central Asia, distinguished from temperature variations of the monsoonally dominated lowlands in the south.

3.4 Comparison with High Asian and Northern Hemispheric reconstructions

To assess regional versus large-scale trends, we compare the new Bagrot reconstruction with the mean summer (JJA) temperature reconstruction for High Asia and reconstructions representing the Northern Hemisphere extratropics (Fig. 7). There is a significant ($p < 0.001$) correlation ($r = 0.45$) between our local and the regional High Asian records (Cook et al. 2013; PAGES 2k 2013) at inter-annual scale. This correlation increases to $r = 0.58$ after 11-year low-pass filtering on the series (Supplementary Table 1). Both records showed relatively stable conditions from 1600 to 1800, followed by marked cooling in the 1840s, and subsequent warming. Prominent cold decades in the 1578–1594, 1642–1657, 1695–1710, 1784–1788, 1811–1826, 1832–1847, and 1909–1918, and 1960–1969 are consistent between the High Asian reconstructions (shaded periods in Fig. 7). The substantial cold period from 1811 to 1826 and 1832 to 1847 also corresponds with other temperature reconstructions from Western Himalaya (Yadav et al. 2004), Tibetan Plateau (Liang et al. 2008; Lv and Zhang

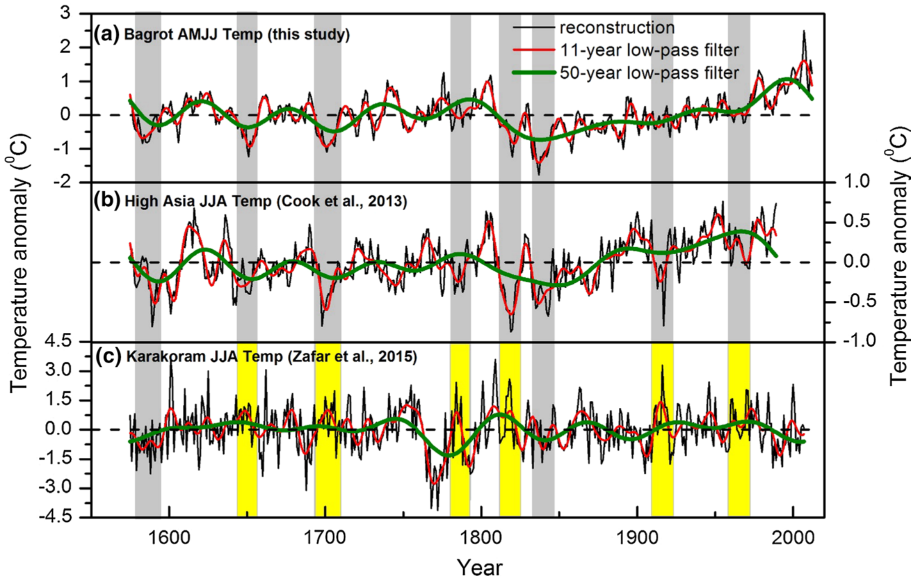

Fig. 7 Comparison of temperature reconstructions of the Bagrot–Karakoram region (April–July, T_{min}; this study) (**a**), High Asia (JJA; Cook et al. 2013; PAGES 2k 2013) (**b**) and Karakorum (JJA; Zafar et al. 2015) (**c**)

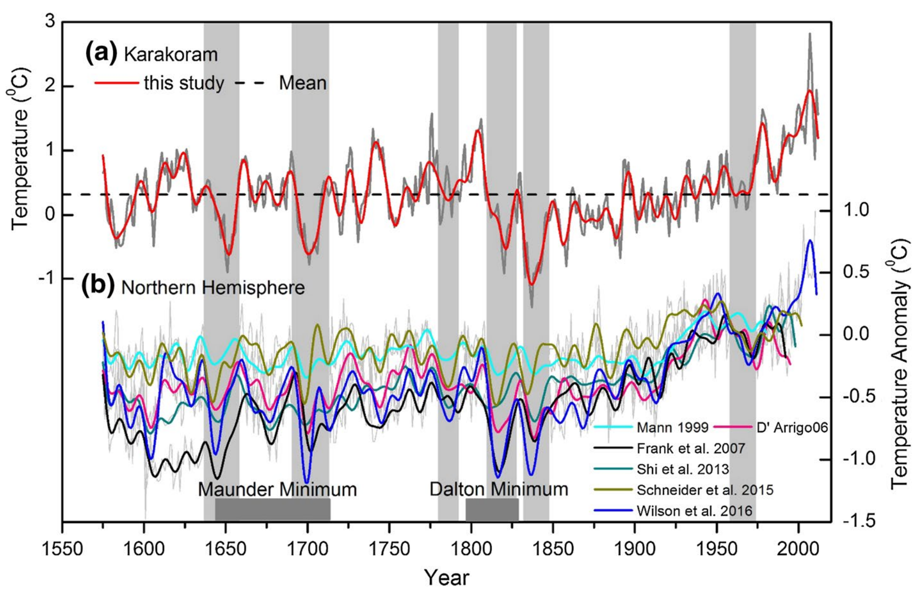

Fig. 8 Comparison between the Karakoram temperature reconstruction of this study and Northern Hemisphere reconstructions (D'Arrigo et al. 2006; Frank et al. 2007; Mann et al. 1999; Schneider et al. 2015; Shi et al. 2013; Wilson et al. 2016)

2013; Zhu et al. 2011), and central Asia (Davi et al. 2015), despite differences in magnitude. The early nineteenth century cold conditions are likely driven by solar variability (the Dalton Minimum; Wagner and Zorita 2005) and volcanic forcing (Tambora eruption; Esper et al. 2013b). These shared warm/cool variations re-validate the reliability of our reconstruction in Karakoram at decadal to multi-decadal scales.

However, most of these reconstructed cold and warm intervals are out of phase with the Karakoram JJA temperature reconstruction by Zafar et al. (2015). The latter record appears flat, containing relatively minor low frequency variance over the past 400+ years. There is no correlation (r = 0.01) between the series of Zafar et al. (2015) and our reconstruction. Possible explanations include the earlier reconstruction to be sensitive to drought stress, which is the combination of high temperature and low precipitation (Cook et al. 2010), and cloud cover changes. The latter has been suggested by Esper et al. (2007) as a possible influence of growth of old junipers of this area.

Our temperature reconstruction from the Karakoram region is also in good agreement with large-scale reconstructions integrating variability at the Northern Hemispheric extratropical scale (D'Arrigo et al. 2006; Frank et al. 2007; Mann et al. 1999; Schneider et al. 2015; Shi et al. 2013; Wilson et al. 2016) (Fig. 8). All of these reconstructions show a cool Little Ice Age and subsequent warming into twentieth century. There are consistent cold periods at decadal to multi-decadal scales including the cold episodes 1641–1657 and 1694–1710 (Maunder minimum), and 1812–1826 (Dalton Minimum). Another cold event in the first half of nineteenth century from 1830 to 1847 is also seen in Northern Hemisphere records. However, temperatures in the Karakoram were not cold in 1816 subsequent to the Tambora eruption, despite cool deviations in Northern Hemisphere, the east Himalaya (Krusic et al. 2015), and the southeast Tibetan Plateau (Duan and Zhang 2014; Liang et al. 2009; Wang et al. 2010; Zhu et al. 2011). Similar differences in the severity of the 1816 year without a summer are, however, also reported from Europe (Esper et al. 2013a, b). Our reconstruction correlates significantly ($p < 0.01$) with the Northern Hemispheric records. The strongest correlation ($r = 0.55$; Supplementary Table 2) is found with the Wilson et al. (2016) tree-ring based temperature reconstruction. The correlation increases to 0.68 after 11-year low-pass filtering on both records, revealing increased similarity at decadal to inter-decadal timescales.

4 Conclusions

We developed an April–July temperature reconstruction extending back to AD 1575 based on *Pinus Wallichiana* TRW data from the upper timberline in the Karakoram, northern Pakistan. The reconstruction is characterized by decadal to multi-decadal variations during the Little Ice Age, and a subsequent warming trend since the mid-nineteenth century, in line with reconstructions from High Asia and the Northern Hemisphere. Our reconstruction demonstrates that Karakoram temperatures are in-phase, rather than out-of phase with large-scale continental and hemispheric trends. However, it should be mentioned that there is need to collect more samples and develop additional high-elevation records to support this argument. Since little is known about longer-term glacier fluctuations in the region, additional dendroglaciological studies are needed to evaluate the Karakoram Anomaly and explain the long-term relationships between Karakoram climate variation and glacier fluctuations.

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References

Ahmed M, Palmer J, Khan N, Wahab M, Fenwick P, Esper J, Cook E (2011) The dendroclimatic potential of conifers from northern Pakistan. Dendrochronologia 29:77–88

Bolch T, Kulkarni A, Kääb A, Huggel C, Paul F, Cogley JG et al (2012) The state and fate of Himalayan glaciers. Science 336:310–314

Cook ER (1985) A time-series analysis approach to tree-ring standardization. PhD dissertation. Department of Geosciences, The University of Arizona, Tucson

Cook ER, Kairiukstis LA (1990) Methods of dendrochronology: applications in the environmental sciences. Springer Science & Business Media, Berlin

Cook ER, Peters K (1997) Calculating unbiased tree-ring indices for the study of climatic and environmental change. Holocene 7:361–370

Cook ER, Krusic PJ, Jones PD (2003) Dendroclimatic signals in long tree-ring chronologies from the Himalayas of Nepal. Int J Climatol 23:707–732

Cook ER, Anchukaitis KJ, Buckley BM, D'Arrigo RD, Jacoby GC, Wright WE (2010) Asian monsoon failure and megadrought during the last millennium. Science 328:486–489

Cook ER, Krusic PJ, Anchukaitis KJ, Buckley BM, Nakatsuka T, Sano M (2013) Tree-ring reconstructed summer temperature anomalies for temperate East Asia since 800 CE. Clim Dyn 41:2957–2972

Cramer T (2000) Geländeklimatologische Studien im Bagrottal:(Karakorumgebirge, Pakistan). Terrimago Pachnicke, Göttingen

D'Arrigo R, Wilson R, Jacoby G (2006) On the long-term context for late twentieth century warming. J Geophys Res Atmos 111:1–12

Davi NK et al (2015) A long-term context (931–2005 CE) for rapid warming over Central Asia. Quat Sci Rev 121:89–97. doi:10.1016/j.quascirev.2015.05.020

Deng Y, Gou X, Gao L, Yang T, Yang M (2014) Early-summer temperature variations over the past 563 yr inferred from tree rings in the Shaluli Mountains, southeastern Tibet Plateau. Quat Res 81:513–519. doi:10.1016/j.yqres.2013.08.002

Duan J, Zhang QB (2014) A 449 year warm season temperature reconstruction in the southeastern Tibetan Plateau and its relation to solar activity. J Geophys Res Atmos 119:11578–11592. doi:10.1002/2014jd022422

Esper J (2000) Long-term tree-ring variations in Juniperus at the upper timber-line in the Karakorum (Pakistan). Holocene 10:253–260

Esper J, Cook ER, Schweingruber FH (2002a) Low-frequency signals in long tree-ring chronologies for reconstructing past temperature variability. Science 295:2250–2253

Esper J, Schweingruber FH, Winiger M (2002b) 1300 years of climatic history for Western Central Asia inferred from tree-rings. Holocene 12:267–277

Esper J, Frank DC, Wilson RJ, Büntgen U, Treydte K (2007) Uniform growth trends among central Asian low-and high-elevation juniper tree sites. Trees 21:141–150

Esper J, Büntgen U, Luterbacher J, Krusic PJ (2013a) Testing the hypothesis of post-volcanic missing rings in temperature sensitive dendrochronological data. Dendrochronologia 31:216–222

Esper J, Schneider L, Krusic PJ, Luterbacher J, Büntgen U, Timonen M, Sirocko F, Zorita E (2013b) European summer temperature response to annually dated volcanic eruptions over the past nine centuries. Bull Volcanol 75:1–14

Frank D, Esper J, Cook ER (2007) Adjustment for proxy number and coherence in a large-scale temperature reconstruction. Geophys Res Lett 34:1–5. doi:10.1029/2007GL030571

Fritts H (1976) Tree rings and climate. Academic, New York

Hewitt K (2005) The Karakoram anomaly? Glacier expansion and the 'elevation effect', Karakoram Himalaya. Mt Res Dev 25:332–340

Hewitt K (2011) Glacier change, concentration, and elevation effects in the Karakoram Himalaya, Upper Indus Basin. Mt Res Dev 25:332–340

Hewitt K (2014) Glaciers of the Karakoram Himalaya: glacial environments, processes, hazards and resources. Springer, Berlin
Holmes RL (1983) Computer-assisted quality control in tree-ring dating and measurement. Tree-ring bulletin 43:69–78
IPCC (2007) Climate change 2007-synthesis report. Contribution of Working groups I II and III to the fourth assessment report of the Anpassungsstrategien und Klimaschutzaktivitäten
Kapnick SB, Delworth TL, Ashfaq M, Malyshev S, Milly P (2014) Snowfall less sensitive to warming in Karakoram than in Himalayas due to a unique seasonal cycle. Nat Geosci 7:834–840
Körner C (1999) Alpine plant life: functional plant ecology of high mountain ecosystems; with 47 tables. Springer & Business Media, Berlin
Krusic P, Cook E, Dukpa D, Putnam A, Rupper S, Schaefer J (2015) Six hundred thirty-eight years of summer temperature variability over the Bhutanese Himalaya. Geophys Res Lett 42:2988–2994. doi:10.1002/2015GL063566
Kumar P, Kotlarski S, Moseley C, Sieck K, Frey H, Stoffel M, Jacob D (2015) Response of Karakoram-Himalayan glaciers to climate variability and climatic change: a regional climate model assessment. J Geophys Res Lett 42:1818–1825
Liang E, Shao X, Qin N (2008) Tree-ring based summer temperature reconstruction for the source region of the Yangtze River on the Tibetan Plateau. Glob Planet Change 61:313–320
Liang E, Shao X, Xu Y (2009) Tree-ring evidence of recent abnormal warming on the southeast Tibetan Plateau. Theor Appl Climatol 98:9–18. doi:10.1007/s00704-008-0085-6
Liang E, Shao X, Eckstein D, Liu X (2010) Spatial variability of tree growth along a latitudinal transect in the Qilian Mountains, northeastern Tibetan Plateau. Can J For Res 40:200–211. doi:10.1139/x09-186
Lv L-X, Zhang Q-B (2013) Tree-ring based summer minimum temperature reconstruction for the southern edge of the Qinghai–Tibetan Plateau, China. Clim Res 56:91–101
Mann ME, Bradley RS, Hughes MK (1999) Northern Hemisphere temperatures during the past millennium; Inferences, uncertainties, and limitations. Geophys Res Lett 26:759–762
Mayer C, Lambrecht A, Mihalcea C, Belò M, Diolaiuti G, Smiraglia C, Bashir F (2010) Analysis of glacial meltwater in Bagrot Valley, Karakoram: based on short-term ablation and debris cover observations on Hinarche Glacier. Mt Res Dev 30:169–177. doi:10.1659/MRD-JOURNAL-D-09-00043.1
Mitchell TD, Jones PD (2005) An improved method of constructing a database of monthly climate observations and associated high-resolution grids. Int J Climatol 25:693–712. doi:10.1002/joc.1181
Orwa C, Mutua A, Kindt R, Jamnadass R, Simons A (2015) Agroforestree database: a tree reference and selection guide version 4.0. 2009. http://www.worldagroforestry.org/af/treedb/. Accessed on 15 Feb 2011
PAGES 2 k C (2013) Continental-Scale temperature variability during the Past Two Millennia. Nat Geosci 6:339–346
Rossi S, Deslauriers A, Griçar J et al (2008) Critical temperatures for xylogenesis in conifers of cold climates. Glob Ecol Biogeogr 17:696–707. doi:10.1111/j.1466-8238.2008.00417.x
Schneider L, Smerdon JE, Büntgen U, Wilson RJ, Myglan VS, Kirdyanov AV, Esper J (2015) Revising midlatitude summer temperatures back to AD 600 based on a wood density network. Geophys Res Lett 42:4556–4562
Shi F, Yang B, Mairesse A, Gunten LV, Li JP, Bräuning A, Yang F, Xia X (2013) Northern Hemisphere temperature reconstruction during the last millennium using multiple annual proxies. Clim Res 56:231–244. doi:10.3354/cr01156
Singh J, Yadav RR (2007) Dendroclimatic potential of millennium-long ring-width chronology of Pinus gerardiana from Himachal Pradesh, India. Curr Sci 00113891:93
Stokes M, Smiley T (1968) An introduction to tree-ring dating. University of Chicago Press, Chicago, IL
Thapa UK, Shah SK, Gaire NP, Bhuju DR (2014) Spring temperatures in the far-western Nepal Himalaya since AD 1640 reconstructed from Picea smithiana tree-ring widths. Clim Dyn. doi:10.1007/s00382-014-2457-1
Wagner S, Zorita E (2005) The influence of volcanic, solar and CO2 forcing on the temperatures in the Dalton Minimum (1790–1830): a model study. Clim Dyn 25:205–218
Wang L, Duan J, Chen J, Huang L, Shao X (2010) Temperature reconstruction from tree-ring maximum density of Balfour spruce in eastern Tibet, China. Int J Climatol 30:972–979. doi:10.1002/joc.2000
Wang J, Yang B, Qin C, Kang S, He M, Wang Z (2014) Tree-ring inferred annual mean temperature variations on the southeastern Tibetan Plateau during the last millennium and their relationships with the Atlantic Multidecadal Oscillation. Clim Dyn 43:627–640
Weiers S (1998) Wechselwirkungen zwischen sommerlicher Monsunaktivität und außertropischer Westzirkulation in den Hochgebirgsregionen Nordpakistans. PGM 142:85–104
Wigley TM, Briffa KR, Jones PD (1984) On the average value of correlated time series, with applications in dendroclimatology and hydrometeorology. J Clim Appl Meteorol 23:201–213
Wilson R, Anchukaitis K, Briffa KR, Büntgen U, Cook E, D'Arrigo R, Davi N, Esper J, Frank D, Gunnarson B (2016) Last millennium northern hemisphere summer temperatures from tree rings: Part I: The long term context. Quat Sci 134:1–18. doi:10.1016/j.quascirev.2015.12.005
Winiger M, Gumpert M, Yamout H (2005) Karakorum–Hindukush–western Himalaya: assessing high-altitude water resources. Hydrol Process 19:2329–2338. doi:10.1002/hyp.5887
Yadav RR, Park WK, Singh J, Dubey B (2004) Do the western Himalayas defy global warming? Geophys Res Lett 31:1–5. doi:10.1029/2004GL020201
Yadav RR, Braeuning A, Singh J (2011) Tree ring inferred summer temperature variations over the last millennium in western Himalaya, India. Clim Dyn 36:1545–1554
Zafar MU, Ahmed M, Rao MP, Buckley BM, Khan N, Wahab M, Palmer J (2015) Karakorum temperature out of phase with hemispheric trends for the past five centuries. Clim Dyn. doi:10.1007/s00382-015-2685-z
Zhang Q-B, Evans MN, Lyu L (2015) Moisture dipole over the Tibetan Plateau during the past five and a half centuries. Nat Commun 6:8062. doi:10.1038/ncomms9062
Zhu H, Zheng Y, Shao X, Liu X, Xu Y, Liang E (2008) Millennial temperature reconstruction based on tree-ring widths of Qilian juniper from Wulan, Qinghai Province, China. Chin Sci Bull 53:3914–3920. doi:10.1007/s11434-008-0400-8
Zhu H, Shao X, Yin Z-Y, Xu P, Xu Y, Tian H (2011) August temperature variability in the southeastern Tibetan Plateau since AD 1385 inferred from tree rings. Palaeogeogr Palaeoclimatol Palaeoecol 305:84–92. doi:10.1016/j.palaeo.2011.02.017