



Recent decades are the warmest in over 400 years in Northeast Asia: evidence from tree-ring anatomy

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Abstract

Recent rapid climate warming often exceeds the rate and magnitude of pre-industrial variability over the Common Era. While contemporary warming exhibits a high degree of global synchrony, pre-industrial warm and cold periods were often spatially heterogeneous and asynchronous. Consequently, reliable regional paleoclimate records remain essential for understanding past climate variability. Here, we present an annually resolved, 411-year-long reconstruction of June 9–July 30 minimum temperature based on cell wall thickness of *Picea glehnii* trees from Hokkaido, Japan. Our reconstruction, developed from non-detrended chronologies and daily climate data ($r = -0.68$, $p < 0.05$), provides a spectrally unbiased record of temperature variability across Northeast Asia. It captures pronounced cold and warm periods that align with historical records and regional climate reconstructions. For instance, it reveals cold periods in the 17th century corresponding to the Maunder Minimum and a distinct cold period in the 1830s, which coincided with one of the severest famines in Japanese history. Crucially, the period from 1994 to 2016 was the longest warm period in the entire reconstruction. The reconstruction also shows significant, persistent multidecadal variability that likely reflects the Pacific Decadal Oscillation. These findings highlight the strong potential of quantitative wood anatomy to improve our understanding of long-term climate dynamics and, thus, reduce uncertainties in future climate projections.

Keywords Dendroanatomy · Daily climate data · Lumen area · Sakhalin spruce · Peatland

1 Introduction

Understanding the patterns and drivers of past climate variability is essential for placing contemporary climate change in a long-term context and improving the accuracy of future projections (Bradley 2015; Kageyama et al. 2024; Tierney et al. 2020). Although instrumental climate records are invaluable, their relatively short length limits the assessment of long-term trends. To overcome this limitation, high-resolution proxy records are required to

reconstruct climate beyond the instrumental period (Bradley 2015). Furthermore, continued refinement of methods for extracting and interpreting proxy data is essential for improving the accuracy and reliability of climate reconstructions (Björklund et al. 2023; Jones et al. 2009).

Tree rings are exceptional proxies for reconstructing the climate of the past few millennia, offering high-resolution, absolutely dated records (Bradley 2015; Fritts 1976; Hughes 2002). To date, most reconstructions are based on tree-ring width (TRW) and maximum latewood density (MXD). However, these parameters require detrending to remove age- or size-related growth trends. Although detrending is necessary to isolate climatic signals, it can inadvertently distort and eliminate part of the climatic variability (Hughes 2002; Melvin and Briffa 2008).

To address this limitation, the emerging field of quantitative wood anatomy (QWA) offers a promising alternative. Quantitative measurements of wood-cell anatomical traits, such as lumen area and cell wall thickness, can provide stronger or additional environmental information than other tree-ring parameters (Edwards et al. 2022, 2025; Fonti and García-González 2008; Fonti et al. 2010; Martin-Benito et al. 2013). Crucially, certain QWA parameters may not require detrending and therefore better preserve low-frequency climatic variability (Björklund et al. 2020; Liang et al. 2013; Lopez-Saez et al. 2023; Pritzkow et al. 2016). Although relatively few QWA-based climate reconstructions currently exist (Björklund et al. 2023; Edwards et al. 2025; Lopez-Saez et al. 2024; Panyushkina et al. 2003; Zharkov et al. 2024; Ziaco et al. 2016), they have demonstrated considerable potential for improving the fidelity of climate reconstructions. For instance, Björklund et al. (2023) have shown that a Fennoscandian temperature reconstruction based on QWA captured a significantly stronger climate signal and consistently outperformed an MXD-based reconstruction at both interannual and decadal timescales.

Despite the growing potential of QWA, its application remains geographically limited and it has rarely been explored in Northeast Asia. The climate of this region is shaped by a complex interplay between the East Asian monsoon and large-scale ocean–atmosphere circulations, including the El Niño–Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO). Interactions among these phenomena generate spatially heterogeneous seasonal and interannual climate variability (Huang et al. 2012; Choi and Kim 2019; Kosaka et al. 2013; Nguyen–Le and Ngo-Duc 2025; Wang et al. 2024) and influence the occurrence of extreme climatic events (Wang et al. 2025). Given the region’s high population density and its strong dependence on climate-sensitive sectors such as agriculture, forestry, and water resources, even modest changes in climate variability can have substantial ecological and socio-economic impacts.

In Japan, climate changes have already manifested in shifting species distributions and phenology (Ogawa-Onishi and Berry 2013) as well as in altered forest growth, productivity, and composition (Altman et al. 2016; Altmanová et al. 2025; Fibich et al. 2021; Hiura et al. 2019; Tsunoda et al. 2024). Although warming might benefit rice and other crops cultivated at higher latitudes, such as in Hokkaido (Kunimitsu 2015), recent climate changes have generally been associated with reduced rice quality (Ishigooka et al. 2021; Lee et al. 2013; Okada et al. 2011). Moreover, substantial reductions in rice and potato yields have been linked to extreme heat events and climate-induced shifts in phenology (Nemoto et al. 2011; Shimono 2011; Sugawara et al. 2021). Climate projections indicate that temperature and precipitation in East Asia are likely to increase at rates exceeding the global means during

the 21st century, especially at higher latitudes (Min et al. 2006; Xin et al. 2013). In addition, higher latitudes may experience an increase in the frequency and intensity of tropical cyclones (Altman and Korznikov 2025; Altman et al. 2018; Liu et al. 2024). Improving our understanding of regional climate dynamics is therefore essential for developing effective adaptation strategies and reducing future socio-economic risks in Northeast Asia.

While the eastern side of the North Pacific has been extensively studied with high-density proxy networks, the western side remains comparatively underrepresented (PAGES2k Consortium 2017). Furthermore, most of the existing tree-ring-based reconstructions in this region rely on detrended chronologies. Additionally, these reconstructions are based on calibration against monthly climate data, which may not align with the specific seasonal window during which tree growth is most climatically sensitive (Jevšenak and Levanič 2018).

In this study, we developed TRW- and QWA-based chronologies from *Picea glehnii* trees from Hokkaido, Japan. Using daily climate data, we explored the climatic signals of these chronologies and reconstructed the climatic variable with the strongest signal. This approach provides new insights into climate-growth relationships and past climate variability in Northeast Asia.

2 Materials and methods

2.1 Study site

The study was conducted in the Orochi-ga-hara montane peatland on Mount Muine (Muine-yama volcano), Hokkaido, Japan (42.94 N, 141.06 E; Fig. 1), at an elevation of 980 m a.s.l. The hydrological regime is strongly influenced by heavy winter snow accumulation and

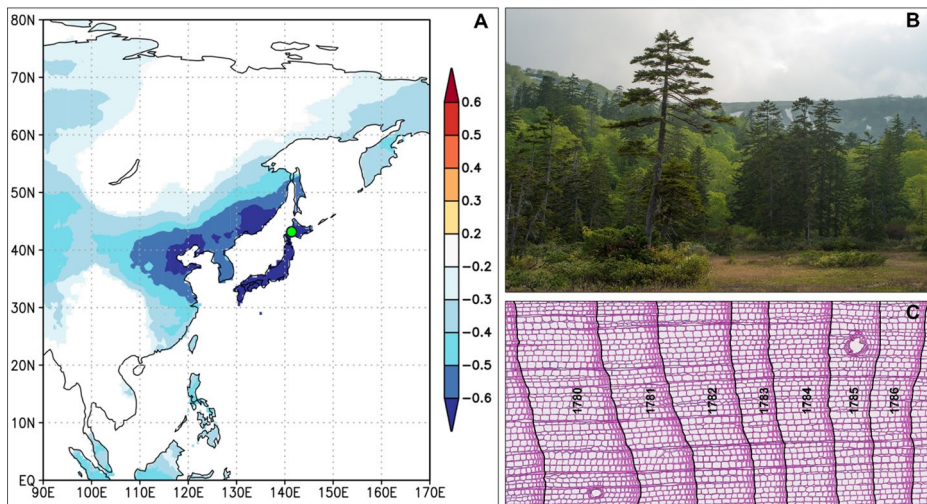


Fig. 1 (A) Map of Northeast Asia showing the study site in Hokkaido, Japan (green dot), and the spatial correlation between the cell-wall-thickness chronology and average June–July minimum temperature for 1901–2016 (CRU TS4.07; Harris et al. (2020)). The colour scale indicates the strength and direction of correlations. (B) View of the study site where *Picea glehnii* samples were collected. (C) Representative microsection of *Picea glehnii* analysed in ROXAS, showing tree rings from 1780 to 1786

subsequent snowmelt in late spring, and the water table remains relatively high throughout the year. The vegetation is characterized by a monodominant canopy of *Picea glehnii*. The understory comprises *Pinus pumila* and *Sasa kurilensis*, with a *Sphagnum*-dominated ground layer and a diverse assemblage of herbaceous species, including *Eriophorum vaginatum*, *Lysichiton camtschatcensis*, and *Nephrophyllidium crista-galli*.

Picea glehnii is a native spruce species in Hokkaido adapted to grow in challenging environments, such as infertile serpentine soils, peatlands, or coastal sand dunes (Haraguchi et al. 2003; Kayama 2006; Nakata and Kojima 1987; Nishijima and Nakata 2004). It is a slow-growing, shallow-rooted species, characterized by low nutrient requirements, low photosynthetic rates, low nitrogen concentration in needles, and extended needle longevity (Kayama et al. 2007; Nishijima and Nakata 2004). The climate of Hokkaido is characterized by pronounced seasonal changes (Fig. S1), driven primarily by the East Asian monsoon (Ding and Chan 2005; Chang 2004). During winter, the East Asian winter monsoon advects cold and dry air from Siberia, leading to cold temperatures, low humidity, and heavy snowfall. In summer, the East Asian summer monsoon brings warm and moist air from the Pacific Ocean, resulting in mild temperatures, high humidity, and rainfall (Chang 2004). These monsoonal dynamics are substantially modulated by large-scale ocean–atmosphere circulations, including the ENSO (Ohba and Sugimoto 2022; Wang et al. 2000; Zhang et al. 2024), the PDO (Nguyen–Le and Ngo-Duc 2025; Wang et al. 2024), and the Arctic Oscillation (Gong et al. 2011; He et al. 2017). Furthermore, Hokkaido is affected by tropical cyclones, although they occur much less frequently than in lower-latitude regions (Altman et al. 2016; Hoshino et al. 2020).

2.2 Tree-ring width analysis

In the summer of 2017, one increment core per tree was collected from 42 living dominant *Picea glehnii* trees at breast height (i.e., 130 cm above ground) using 5-mm increment borers. Cores were air-dried, surfaced with a core microtome (Gärtner and Nievergelt 2010), and chalked to enhance the visibility of tree-ring boundaries. Tree-ring widths were measured from pith to bark to the nearest 0.001 mm using the Lintab 6 measuring device and TSAP-Win™ time series analysis software. Cross-dating was performed visually (Yamaguchi 1991) and then statistically verified using the program Past 5 (SCIEM Inc.) based on the percentage of parallel variation (Gleichläufigkeit; Eckstein and Bauch (1969) and the similarity of growth patterns between series (Baillie–Pilcher’s t-value; Baillie and Pilcher (1973)).

2.3 QWA analysis

For QWA analysis, we selected 12 trees that showed the highest correlation with the mean TRW chronology. Sample preparation followed the approach of Fonti et al. (2025). Increment cores were cut into ~2.5-cm-long segments and embedded in paraffin using a Leica TP1020 embedding station combined with a Leica HistoCore Arcadia. Microsections of 10-μm thickness were produced from the segments using a Leica RM2245 rotary microtome, double-stained with 0.8% Safranin and 0.5% Astra Blue, and permanently fixed with Euparal. Microsection images were captured using a Zeiss Axioscan.Z1 slide scanner at a resolution of 2.27 pixels per μm. Tree-ring borders and tracheid cells were automatically

detected in ROXAS software (Prendin et al. 2017; von Arx and Carrer 2014) and manually corrected. Tree rings containing compression wood were excluded to avoid potential bias in the climatic signal. In total, 4,313,882 tracheid cells were measured. Finally, the output for all cell anatomical parameters provided by ROXAS was calculated.

After screening a wide range of anatomical parameters, including those used in previous studies, we selected lumen area (LA) and cell wall thickness (CWT) for further dendroclimatological analysis. These parameters were chosen due to their high year-to-year variability, as indicated by the mean Gleichläufigkeit (Table S1). For LA, we used the 90th percentile value, representing the widest earlywood cells. For CWT, the median value for each ring was calculated. In the latewood, CWT was determined as the average thickness of both tangential and radial walls, whereas in the earlywood, only tangential wall thickness was considered to avoid artefacts caused by the wall widening near pits, which is commonly present in radial walls. The transition from earlywood to latewood was defined using Mork's index with a threshold value of 1 (Denne 1989).

2.4 Development of chronologies

The subsample signal strength (SSS) was used as a confidence measure of predictive power of the TRW, LA, and CWT chronologies back in time. A threshold of 0.85 was applied, above which a chronology is considered to provide a reliable palaeoclimatological proxy record (Buras 2017; Wigley et al. 1984). In addition, we calculated the mean interseries correlation (R_{bar}), which quantifies the strength of the common signal among individual series by averaging all pairwise correlations (Cook and Kairiukstis 1990; Wigley et al. 1984).

TRW series were detrended and prewhitened to remove non-climatic growth trends. Specifically, a 50-year cubic smoothing spline was fitted to each measured TRW series, and ratios from these fitted growth curves were calculated. In contrast, LA and CWT series were not detrended, as our data did not exhibit any age- or size-related trend over the period with $\text{SSS} > 0.85$ (Figs. S2–S7). Subsequently, individual detrended TRW series, and raw LA and CWT series were averaged into their respective chronologies using Tukey's biweight robust mean.

SSS calculations, detrending of TRW series, and chronology construction were performed using the 'dplR' package (Bunn 2008) in R (R Core Team 2025).

2.5 Dendroclimatological relationship

To assess the influence of climate on the chronologies, mean, maximum, and minimum temperatures, and precipitation were obtained for the grid cell nearest to our study site from a forcings dataset of gridded ($0.5^\circ \times 0.5^\circ$) land surface blend of Climatic Research Unit (CRU) and Japanese reanalysis (JRA) data, hereafter CRU JRA v2.4 (University of East Anglia Climatic Research Unit and and Harris 2024) for the period 1901–2016. The original 6-hourly data were aggregated to daily resolution for our analysis. The validation and justification for using this dataset are provided in the Supplementary methods and associated figures (Figs. S8–S9).

We quantified the climatic signal by correlating daily climate data with each chronology using the 'dendrotools' package (Jevšenak and Levanič 2018) in R (R Core Team 2025). This daily-resolution approach offers a higher temporal resolution compared to the tradi-

tional approach using monthly climate data, allowing us to detect climatic signals more precisely without the constraints of artificially defined monthly boundaries. To find the most relevant period influencing our chronologies, we calculated the correlations for possible window lengths from 30 to 270 days moving by one day from January 1 of the year preceding ring formation to December 31 of the ring-formation year.

We performed spatial correlations to evaluate the spatial extent of the climatic signal for the strongest relationships found between individual chronologies and climate parameters. The spatial correlation and resulting map were created using KNMI Climate Explorer (climexp.knmi.nl; Trouet and Van Oldenborgh (2013) with the gridded $0.5^\circ \times 0.5^\circ$ CRU TS4.07 data (Harris et al. 2020) for the period 1901–2016.

2.6 Climate reconstruction

The climate reconstruction was based on the strongest dendroclimatological relationship. Reconstruction skill was assessed using split-sample calibration–verification regression models to examine the temporal stability of this relationship (Touchan et al. 2008). Specifically, this approach divides the period covered by CRU JRA v2.4 climate data in half (1902–1959 and 1960–2016) to perform cross-validation of calibration and verification regression models. To evaluate the robustness of the dendroclimatological relationship for climate reconstruction, we utilized the following statistical indicators: the reduction of error (RE), the coefficient of efficiency (CE), and the Durbin–Watson test (DW) with its associated p -value (p). Positive RE and CE values, together with a Durbin–Watson statistic close to the optimum value of 2 and a non-significant p -value ($p > 0.05$), indicate good reconstruction skill and confirm that the regression residuals are free of autocorrelation and long-term trends.

In the reconstruction, warm and cold periods were defined as intervals of three or more consecutive years when the minimum temperature deviated by more than ± 0.5 times the standard deviation from the mean (following the approach of Lyu et al. (2016). Periods separated by only one year were merged. To place our findings into a broader context, these identified periods were compared with previously published regional tree-ring and Japanese documentary-based temperature reconstructions. In particular, the Japanese documentary-based records used for comparison included reconstructions of summer temperatures originally derived by other authors from long and continuous weather descriptions recorded in personal diaries or official administrative documents. For example, Mikami (1996) reconstructed July mean temperature using the number of rainy days recorded in daily diaries kept by the Ishikawa family over multiple generations since 1721. Hirano et al. (2013) estimated July maximum temperature based on the frequency of “fine weather” days described in a personal diary. Additionally, Maejima and Tagami (1983) reconstructed climatic variability using the official administrative records of the feudal Tsugaru Clan, which contain detailed daily descriptions of weather conditions dating back to the 16th century.

Furthermore, we compared our entire reconstruction graphically and statistically with four regional tree-ring-based temperature reconstructions (Cook et al. 2013; Lyu et al. 2016; Wiles et al. 2015; Zhu et al. 2017). Reconstructions that were not originally standardized were transformed into z -scores to allow direct comparison of variability, as absolute values may differ among study sites and reconstructed temporal windows. In addition to interannual variability, we also examined low-frequency variability by calculating running means

for each reconstruction. Similarities among the reconstructions were evaluated visually and quantified using pairwise Pearson correlations calculated for both annual values and the smoothed time series.

Furthermore, we assessed the impact of large volcanic eruptions on our reconstruction using double-bootstrap Superposed Epoch Analysis (Rao et al. 2019). Data on large volcanic eruptions occurring during the reconstruction period were extracted from Wang et al. (2022), who compiled a list of volcanic eruptions over 30–90° N with stratospheric aerosol optical depth at 550 nm ≥ 0.03 , based on the eVol2k dataset (Toohey and Sigl 2017) and the Coupled Model Intercomparison Project Phase 6 volcanic forcing dataset. In total, 24 eruption years were included, with a subset of 12 years corresponding to tropical eruptions only. Statistical significance was assessed using 10,000 Monte Carlo simulations and response uncertainty was estimated using 1000 bootstrap resamples of event years.

To assess periodic variations and their temporal variability within the reconstruction, we conducted a wavelet analysis, using the ‘morlet’ function in the ‘dplR’ package (Bunn 2008) in R (R Core Team 2025). Default parameter settings were used, with a 95% confidence interval to define the cone of influence. Potential drivers of the periodicities identified in the wavelet analysis were further evaluated based on wavelet coherence (Grinsted et al. 2004) using the ‘wtc’ function from the ‘biwavelet’ R package. Subsequently, these drivers were compared with the reconstructed and instrumental June 9–July 30 minimum temperature.

3 Results

3.1 TRW and QWA chronologies

The TRW chronology spanned from 1378 to 2016, with a reliable segment (SSS > 0.85) covering the period from 1524 to 2016 (Figs. S5 and S10). The LA and CWT chronologies were based on nine trees as three trees were excluded from the final analysis due to frequent cell damage. For the LA chronology, the period with SSS > 0.85 extended from 1769 to 2016 (Figs. S6 and S10). The reliable part of the CWT chronology with high predictive power (SSS > 0.85) covered the period from 1606 to 2016 (Fig. S7 and S10). All three chronologies exhibited strong common high-resolution variability among their series (Table S1) and high interseries correlations (Fig. S11).

3.2 Climatic signal of individual chronologies

All three chronologies exhibited stronger correlations with temperature than with precipitation (Table 1). The strongest significant signal was observed between CWT and minimum temperature from June 9 to July 30 ($r = -0.681$, $p < 0.05$; Fig. 2). LA also showed the strongest significant relationship with minimum temperature, spanning from June 13 to July 14 ($r = 0.470$, $p < 0.05$; Fig. 2). TRW was mainly related to maximum temperature from May 26 to June 30 ($r = 0.444$, $p < 0.05$; Fig. 2). Precipitation had a significant negative effect on TRW from May 11 to June 30 ($r = -0.310$, $p < 0.05$; Table 1). In the subsequent period, from early July to mid-October, precipitation showed a significant positive correlation with LA ($r = 0.369$, $p < 0.05$; Table 1) and a significant negative correlation with CWT ($r = -0.306$, $p < 0.05$; Table 1). The spatial correlation between the CWT chronology and June–July min-

Table 1 Periods of the strongest correlations between *Picea glehnii* tree-ring-width (TRW), lumen-area (LA), and cell-wall-thickness (CWT) chronologies and climatic parameters. Dates indicate the start and end of the optimal window within the ring-formation year. All correlation coefficients (r) are significant at $p < 0.05$

Chronologies	Mean temperature		Maximum temperature		Minimum temperature		Precipitation	
	Period	r	Period	r	Period	r	Period	r
TRW	May 27– July 17	0.429	May 26– June 30	0.444	May 26– July 30	0.341	May 11– June 30	-0.310
LA	June 13– July 14	0.448	June 09– July 14	0.369	June 13– July 14	0.470	July 06– October 10	-0.369
CWT	June 06– August 03	-0.616	June 06– August 05	-0.512	June 09– July 30	-0.681	July 12– October 11	-0.306

imum temperature demonstrates the extensive geographical coherence of this relationship across Northeast Asia (Fig. 1A).

3.3 Temperature reconstruction

The CWT chronology was the only chronology showing robust reconstruction skills, particularly for minimum temperature (Table S1–S2; Figs. 3 and S12). Hence, we reconstructed minimum temperatures from June 9 to July 30 for the period 1606–2016 based on the CWT chronology (Fig. 4).

The average reconstructed June 9–July 30 minimum temperature over the past 411 years was 13.11 °C, with a standard deviation of ± 0.63 °C. Ten cold and nine warm periods were identified (Fig. 4, Table S3). The cold periods occurred in 1621–1625, 1646–1653, 1656–1669 (the longest cold period), 1672–1674, 1686–1689, 1831–1838, 1895–1897, 1900–1903, 1907–1909, and 1930–1934. The warm periods were in 1634–1636, 1714–1716, 1723–1732, 1775–1780, 1816–1818, 1823–1827, 1878–1880, 1957–1959, and 1994–2016. The 1994–2016 interval represents the longest continuous warm period and exhibits the highest 20-year average minimum temperatures within the entire record (Fig. S13). The comparison of our temperature reconstruction with other published tree-ring-based reconstructions revealed the strongest agreement with the April–July minimum temperature reconstruction from northeastern China by Lyu et al. (2016) (Figs. 5 and S14–S15).

Superposed Epoch Analysis did not identify any statistically significant effects ($p > 0.05$) of large volcanic eruptions on the reconstructed June 9–July 30 minimum temperatures, although there was a tendency towards negative anomalies in the first year after the eruption year ($p < 0.1$; Figs. S16–S17).

The wavelet analysis of the reconstruction revealed significant periodicities across various time scales (Fig. 6). Specifically, significant multidecadal variability was present at ~ 32 –64 years ($p < 0.05$) for a substantial portion of the record. Interannual variability, occurring at ~ 3 –4, 6–8, and 14–20 years, showed intermittent power and statistical significance throughout the reconstruction. The decadal variability may be associated with the PDO, as further supported by the results of wavelet coherence (Fig. S18) and also by the comparison between the PDO index and the June 9–July 30 minimum temperatures reconstructed in this study ($r = -0.15$, $p < 0.1$ for annual values) and obtained from CRU JRA v2.4 ($r = -0.28$, $p < 0.01$ for annual values; Fig. 7, Table S4).

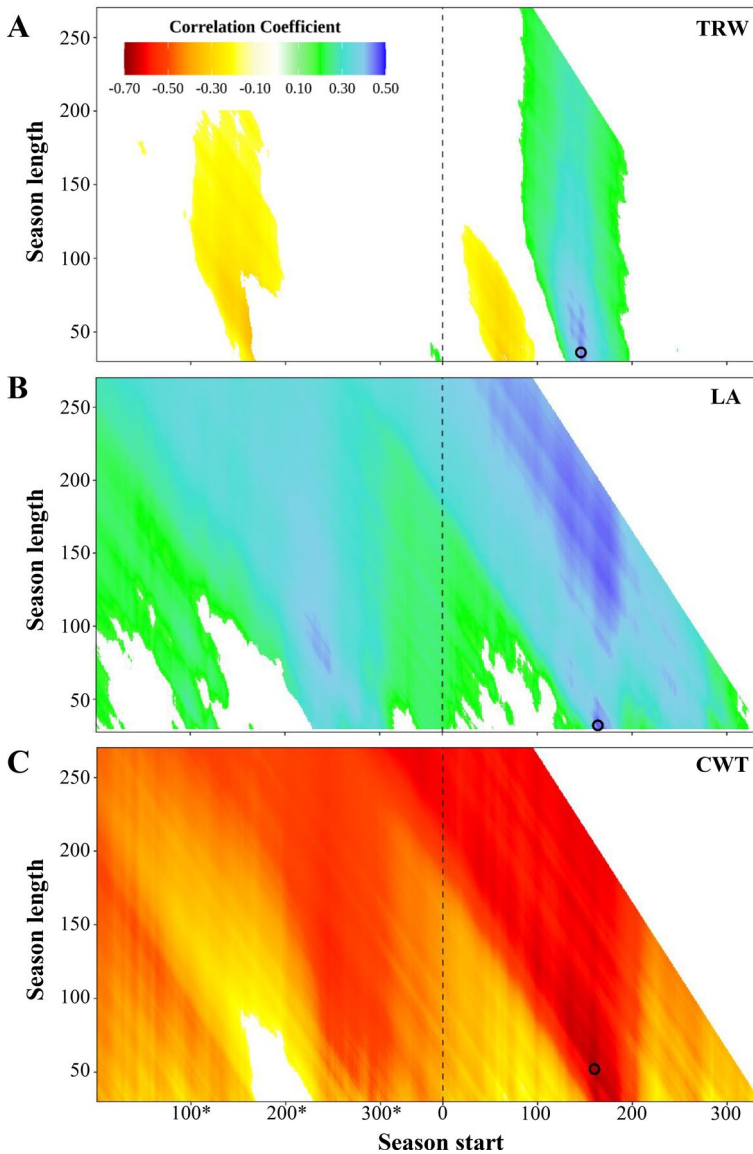


Fig. 2 Heatmaps of correlations between tree-ring chronologies and temperature variables: (A) tree-ring width (TRW) and maximum temperature, (B) lumen area (LA) and minimum temperature, and (C) cell wall thickness (CWT) and minimum temperature. Tile colour indicates the strength and direction of the correlation across different season start days (x-axis) and season lengths (y-axis). The dashed vertical line separates the year preceding ring formation (marked with an asterisk) from the ring-formation year. Each value in the correlation matrix represents the correlation calculated for a moving window aligned to its starting day. Black circles indicate the strongest significant correlation. Only correlations significant at $p < 0.05$ are shown

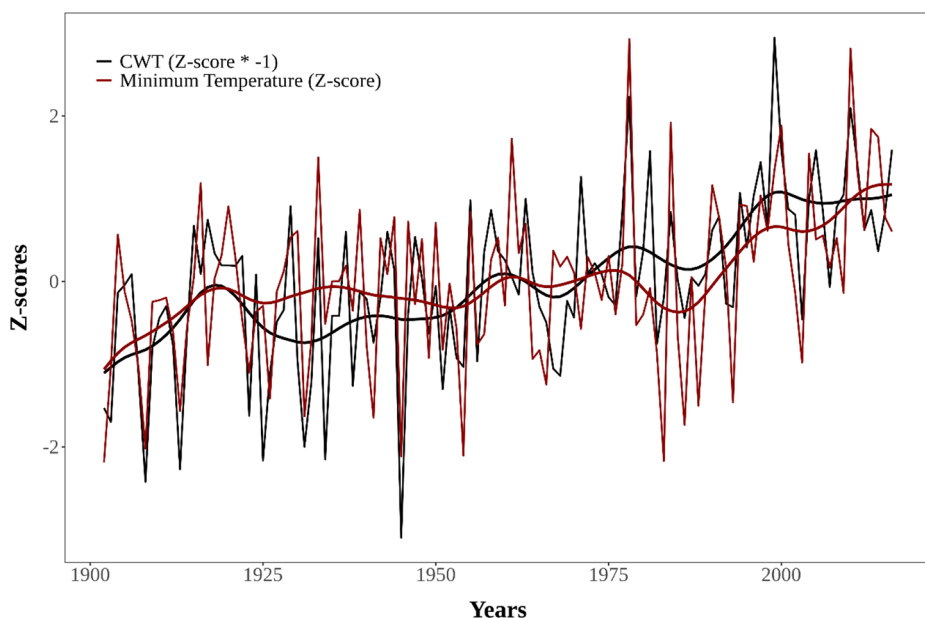


Fig. 3 The Z-scored cell-wall-thickness (CWT) chronology and June 9–July 30 minimum temperatures. The figure demonstrates a strong correspondence between the two time series in both high-frequency (interannual signal; thin lines) and low-frequency variability (20-year cubic smoothing spline; bold lines)

4 Discussion

4.1 Pronounced impact of temperature on *Picea glehnii* growth

Our analysis revealed a clear dominance of temperature over precipitation in influencing the growth of *Picea glehnii* in the studied montane peatland. The most pronounced relationship was a significant negative correlation between CWT and minimum temperature from June 9 to July 30. Conversely, the correlation between LA and minimum temperature for a nearly identical period was significantly positive (Fig. S20). The strong sensitivity of these anatomical traits to minimum (i.e., nighttime) temperature aligns with the fact that trees grow predominantly at night (Han et al. 2026; Steppe et al. 2015; Zweifel et al. 2021). The contrasting effects of minimum temperature on LA and CWT are consistent with experimental evidence showing that elevated nighttime temperatures can stimulate cell enlargement while constraining secondary wall deposition (Han et al. 2026). Higher night temperature can lead to larger LA because cell division and enlargement require turgor pressure to exceed a certain threshold, which is typically reached during nighttime (Cabon et al. 2020; Peters et al. 2021; Steppe et al. 2015). However, enhanced cell division and enlargement may increase carbon demand and shift allocation among competing sinks, potentially limiting carbohydrate availability for subsequent cell wall thickening (Han et al. 2026).

Our findings are consistent with previous research on climate-growth relationships in peatlands. Similarly, Dinella et al. (2021) found a negative relationship between June mean temperature and mean CWT of *Pinus sylvestris* growing on a bog in the Italian Alps. Furthermore, the positive effect of May–July maximum and mean temperatures on TRW observed

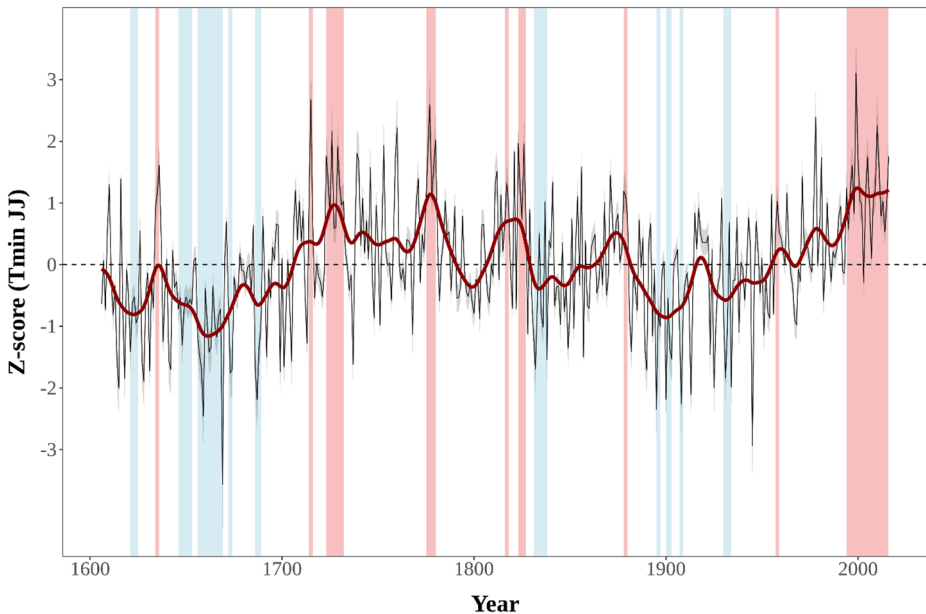


Fig. 4 Reconstructed June 9–July 30 minimum temperatures (Tmin JJ) from 1606 to 2016, expressed as z-scores (reconstructed values in °C are shown in Fig. S19). The black line represents the annual values, while the bold red line indicates a smoothed version with a 20-year cubic spline. Blue and red vertical bands highlight cold and warm periods, respectively, defined as intervals of ≥ 3 consecutive years with Tmin JJ deviating by more than ± 0.5 times the standard deviation from the mean. The grey shading shows the 95% bootstrap confidence interval of reconstructed Tmin JJ based on 1000 resampled calibrations of the regression model

in our study is supported by several earlier investigations of peatland forests (Blanchet et al. 2017; Edvardsson and Hansson 2015; Linderholm et al. 2002; Moir et al. 2011).

It is worth noting that our results contrasts with commonly observed positive relationship between CWT and temperature in high-latitude conifer sites (e.g., Björklund et al. 2020). There are two main factors explaining this discrepancy. First, a montane peatland, where our study was conducted, presents fundamentally different conditions for tree growth compared to mineral soils (Dinella et al. 2021). Second, most previous QWA studies have focused on maximum CWT, whereas we used median CWT. These metrics represent different aspects of wood anatomy and may therefore exhibit varying climate responses.

4.2 Comparison of reconstructed temperature dynamics with other records

All cold and warm periods identified in our reconstruction largely coincide with those reported in other independent proxy records, including regional tree-ring-based reconstructions and Japanese documentary-based reconstructions (Table S3). This provides external validation and a broader spatiotemporal context for our findings, although the periods often vary in duration. The four cold periods in 17th century (1646–1653, 1656–1669, 1672–1674, 1686–1689) coincide with the Maunder Minimum, a period of reduced solar activity spanning approximately from 1645 to 1715 (Eddy 1976; Shindell et al. 2001). These cold periods also overlap with previous reconstructions and documentary evidence (Choi 1994;

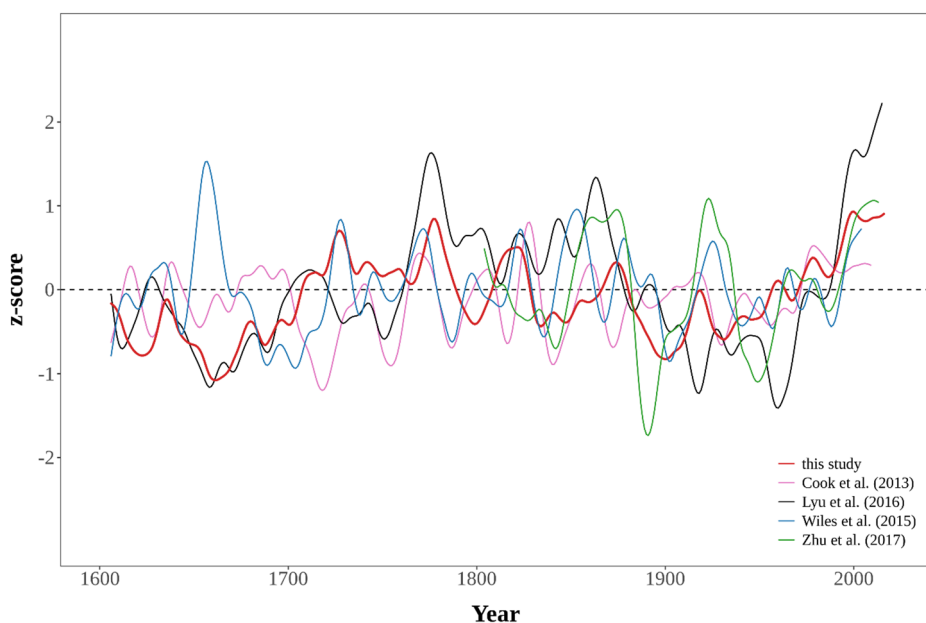


Fig. 5 Comparison of our temperature reconstruction (red) with other regional tree-ring-based reconstructions, including: the East Asian June–August temperature reconstruction (Cook et al. 2013; pink); the April–July minimum temperature reconstruction from northeastern China (Lyu et al. 2016; black); the May–July temperature reconstruction from Sakhalin Island (Wiles et al. 2015; blue); and the May–July temperature reconstruction from northeastern China (Zhu et al. 2017; green). The lines represent 20-year smoothed trends (annual values are shown in Fig. S15)

Kitagawa and Matsumoto 1995; Lyu et al. 2016; Maejima and Tagami 1983, 1986; Tagami and Fukaishi, 1992; Ukhvatkina et al. 2018; Wiles et al. 2015; Table S3).

The cold period from 1831 to 1838 aligns with cold periods identified in all examined Japanese documentary sources (Hirano et al. 2013; Maejima and Tagami 1983, 1986; Mikami 1996; Tagami and Fukaishi, 1992) and several regional tree-ring-based reconstructions (Davi et al. 2002; Doležal et al. 2014; Choi 1994; Ohyama et al. 2013; Park et al. 2001; Ukhvatkina et al. 2018; Yonenobu and Eckstein 2006). Notably, this period coincides with one of the severest famines in Japanese history, known as the Tenpō famine, which occurred in the 1830s (Arakawa 1957; Hirano et al. 2013; Jannetta 1992; Mikami 1988, 1996). Major famines in Japan have often been attributed to poor rice harvests resulting from cold and rainy summer weather conditions (Arakawa 1957; Mikami 1996). However, it is noteworthy that although cold conditions have often led to famines, when viewed from a broader perspective, it seems that cold periods have supported rapid advancements in Japanese society, such as the origin of pottery, transition between hunter-gatherer lifestyle and rice agriculture, or the collapse of feudalism and the start of the modern state (Batten 2009; Kawahata et al. 2017).

The longest warm period identified in our reconstruction, spanning from 1994 to 2016, is consistent with other regional tree-ring-based reconstructions (Davi et al. 2002; Doležal et al. 2014; Li et al. 2020; Lyu et al. 2016; Ukhvatkina et al. 2018; Yu et al. 2018; Zhu et al.

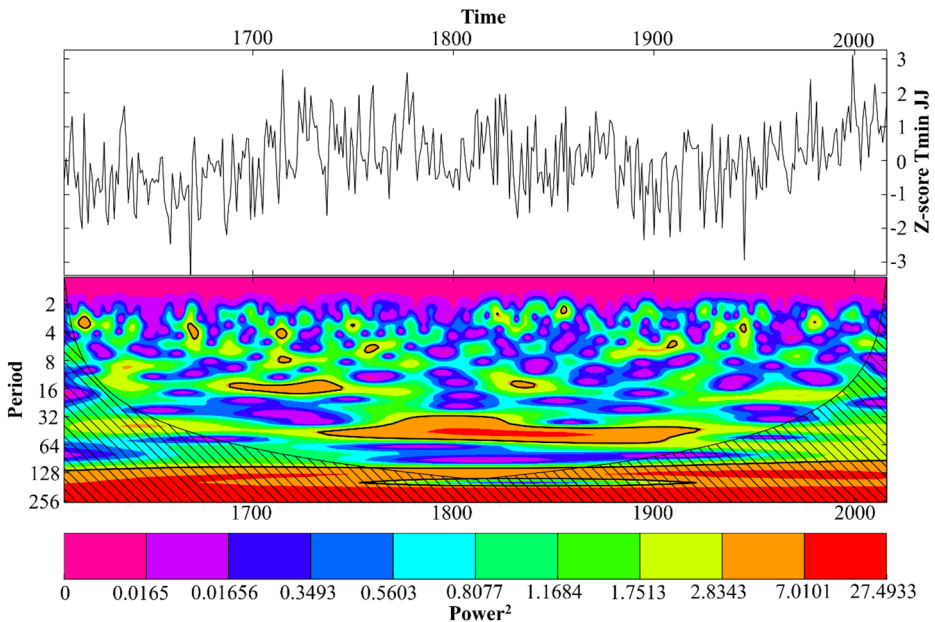


Fig. 6 The wavelet analysis of the June 9–July 30 minimum temperature reconstruction from 1606 to 2016. The upper panel shows the Z-scored June 9–July 30 minimum temperature reconstruction (Tmin JJ). The lower panel presents the wavelet power spectrum, where the x-axis represents time and the y-axis represents period. The colour scale indicates spectral power, with warmer colours (red, orange, yellow) representing higher power and cooler colours (blue, purple, pink) representing lower power. Black contours enclose areas of statistically significant power ($p < 0.05$). Hatched areas at the bottom and sides of the wavelet spectrum indicate the cone of influence, where edge effects reduce confidence

2017). This sustained warming trend since 1990 is also evident from instrumental records, which show a rapid global temperature increase (Samset et al. 2023).

Despite broad agreement with other proxy records, some discrepancies remain (Table S3). These likely arise from differences in 1) the target time period (for instance, our daily data approach extends beyond traditional monthly intervals, 2) the temperature parameters reconstructed (minimum, mean or maximum), 3) the proxy type, 4) the detrending and standardization methods applied to tree-ring series, 5) the criteria used to define cold and warm periods, and 6) the geographic distance between study sites.

4.3 Drivers of temperature variability

Our reconstruction did not show any abrupt cooling following large-scale volcanic eruptions, including the eruptions of Mt. Asama (Japan) and the Laki fissure (Iceland) in 1783, as well as the eruption of Tambora (Indonesia) in 1815. However, this absence is consistent with previous studies from Japan. Davi et al. (2002) and Ohyama et al. (2013) also did not detect significant cooling events associated with these volcanic eruptions in their Japanese tree-ring records. Similarly, Kawakubo et al. (2017) found no cooling after the Tambora eruption in their coral-based sea surface temperature reconstruction from the northern Ryukyus, Japan. Additionally, Wilson et al. (2023) reported that the Japanese main islands

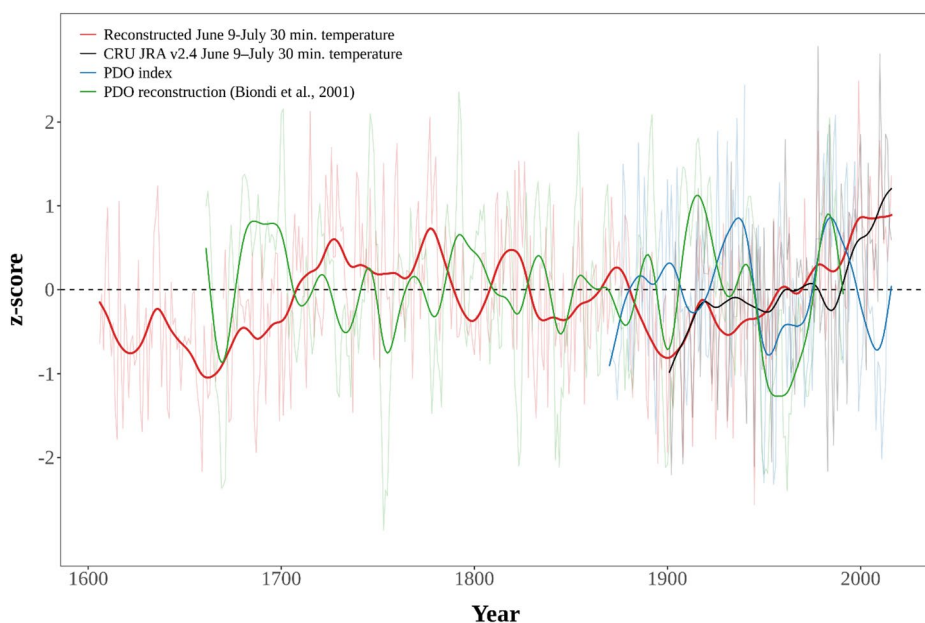


Fig. 7 Comparison between our reconstructed June 9–July 30 minimum temperature ($T_{\min}$; red), the CRU JRA v2.4 June 9–July 30 minimum temperature (black), the annual Pacific Decadal Oscillation (PDO) index based on HadISST1.1 (blue), and a PDO reconstruction developed by Biondi et al. (2001; green). All series are expressed as z-scores. Thin lines represent raw annual values, while bold lines denote 25-year smoothed trends

Hokkaido, Honshu, and Kyushu were the only islands (out of 31 investigated globally) where no impact on food production was recorded after the Tambora eruption in 1815.

Our findings indicate that the multidecadal (~ 32 – 64 years) variability in the reconstructed June 9–July 30 minimum temperatures may be driven by the PDO. The negative relationship between the PDO and temperatures in Hokkaido identified in our study is consistent with recent findings of spatial relationships between PDO and surface air temperature across East Asia (Guo et al. 2024). In addition, other tree-ring studies have identified similar periodicities and attributed them to the PDO (D’Arrigo et al. 2015; Jacoby et al. 2004; Ukhvatkina et al. 2018).

The interannual variability in the 3–8-year range might reflect the influence of the ENSO, which would be consistent with findings from previous studies across East Asia (D’Arrigo et al. 2015; Jacoby et al. 2004; Ukhvatkina et al. 2018; Zhu et al. 2017). However, we did not find any significant statistical relationship between an ENSO index and either the reconstructed or the instrumental June 9–July 30 minimum temperatures (results not shown). The absence of a clear linear relationship might be related to the intermittent occurrence of this periodicity in the wavelet analysis. Additionally, this fluctuation may be modulated by the Yamase wind, a cold northeasterly wind that affects northern Japan’s summer weather with ~ 5 – 6 -year periodicity (Kanno 2004; Kurihara 2003).

5 Conclusion

We provide novel insights into the climate-growth relationships of trees growing in peatlands where studies on intra-annual wood anatomical features are still scarce. The observed relationships between climate variables and tree-ring proxies reflect the delicate balance that trees must employ to optimize growth in peatlands, where substantial fluctuations in both moisture and temperature occur throughout the year.

Our 411-year-long reconstruction of June 9–July 30 minimum temperature revealed significant long-term variability, with cold and warm periods that align with historical records and existing regional reconstructions. Notably, our temperature reconstruction captures well-documented events, such as the Maunder Minimum, the Tenpō famine, and highlights the exceptional warming since 1994. Furthermore, the reconstruction exhibits persistent multidecadal variability, likely driven by the PDO.

Overall, this study highlights the potential of combining non-detrended QWA records with daily climate data as a promising avenue for future climate reconstructions. Future research should further explore QWA's potential across various tree species and environments, which could lead to enhancing our understanding of long-term climate dynamics and responses of forest ecosystems to environmental changes.

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Data availability The dataset used during the current study is available from the corresponding author on reasonable request.

Declarations

Ethics approval and consent to participate Not applicable.

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