Climate Reconstruction from Tree-Rings in the Tatra Mountains

Ryszard J. Kaczka, Barbara Spyt, Karolina Janecka, Tadeusz Niedźwiedź and Zdzisław Bednarz

Abstract This chapter examines the long-term variability of summer (June–July) air temperature and summer humidity (precipitation and Standardised Precipitation Evapotranspiration Index, SPEI) in the region of the Tatra Mountains, which represents natural climate conditions, free of strong anthropogenic influences. The reconstruction of temperature is available for the period since the beginning of the 17th century and reconstruction of humidity related parameters since the beginning of the 18th century by means of the methods based on the tree-ring chronologies. The main proxies utilized for temperature reconstruction were tree-ring widths of Norway spruce (*Picea abies* (L.) H. Karst) and Stone pine (*Pinus cembra* L.) growing in the timberline ecotone. The precipitation and SPEI were reconstructed based on Scots pine (*Pinus sylvestris* L.) tree-ring widths of trees growing at $\sim$1000 m a.s.l. The reconstruction of summer temperature from tree-rings pointed to a relatively cold interval as a part of the Little Ice Age (from the mid 16th to late 19th centuries). In the 20th and at the beginning of the 21st centuries, general increase of air temperature was observed. However, in this recent warm period and during earlier main climatic periods, temperature conditions were not uniform. Analysing series of summer temperature (the 17th–21st centuries) several shorter warm and cool fluctuations were observed. The reconstructed humidity variables exhibited less variability. This is the first attempt of precipitation reconstruction in mountains regions based on the tree-ring chronologies. But the correlation between flood events and humid periods is poor due to the predominant character of the flood caused by short term intensive precipitation of short duration.

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1 Introduction

Different annually resolved tree-ring proxies such as tree-ring width (TRW), maximum wood density, and stable isotopes are widely used to reconstruct past variations in local- to hemispheric-scale temperature means (e.g. Briffa et al. 2001; Jones and Mann 2004; Wilson et al. 2016).

Such natural proxy evidence generally derives from sites located in the high northern latitudes or high-elevations (Fritts 1976; Schweingruber 1996). For Europe, multi-centennial to millennial tree-ring records (Büntgen et al. 2005, 2006, 2013) and network analyses (Frank and Esper 2005a, b) predominantly derive from the Alps, whereas only little evidence exists for the Carpathian arc (Kaczka and Büntgen 2006; Büntgen et al. 2007, 2013; Ponocná et al. 2016) and the lower part of the Upper Vistula Basin (Szychowska-Krąpiec 2010). These examples show clearly that knowledge about regions and processes can be improved much more by studying the local network than singular sites. Even though few TRW based studies introducing the growth/climate response analyses were conducted at local scale (Feliksik 1972, 1992; Bednarz 1976, 1983, 1984, 1996, 2015; Bednarz et al. 1998–1999; Szychowska-Krąpiec 1998; Kaczka 2004; Niedźwiedź 2004; Bednarz and Niedźwiedź 2006; Kaczka and Büntgen 2006; Savva et al. 2006; Büntgen et al. 2007; Popa and Kern 2009; Kaczka and Czajka 2014; Kaczka et al. 2015), studies considering the entire mountain system and climate reconstructions are broadly missing. The initial analyses of growth/climate response for a wider area (Kaczka and Büntgen 2006; Sidor et al. 2015) as a function of geographical settings provide the evidence that multi-site approach allows building the reliable tree ring based reconstructions of summer temperature (Büntgen et al. 2007).

The potential of TRW as precipitation proxy was also probed in the Carpathians (Büntgen et al. 2012; Kaczka et al. 2012; Brzęk et al. 2014) but such reconstructions are still missing. The detailed information about the short- and long-term trends of humidity-related variables would play an important role in understanding the temporal dynamics of natural hazards since summer precipitation has caused most of the floods in the Tatra Mountains region (Kundzewicz et al. 2014; Niedźwiedź et al. 2015). The flood risk at the northern foothills of the Tatra Mountains is considered as one of the main natural hazards. The regular climatic and hydrological records are scattered and lack of data poses the challenge to build applicable models explaining the local and regional flood processes (Parajka et al. 2010; Marchi et al. 2010; Ballesteros-Cánovas et al. 2015a, b; Ruiz-Villanueva et al. 2016). That gap can be bridged using historical records (Figs. 1 and 2) and different proxies including tree rings.

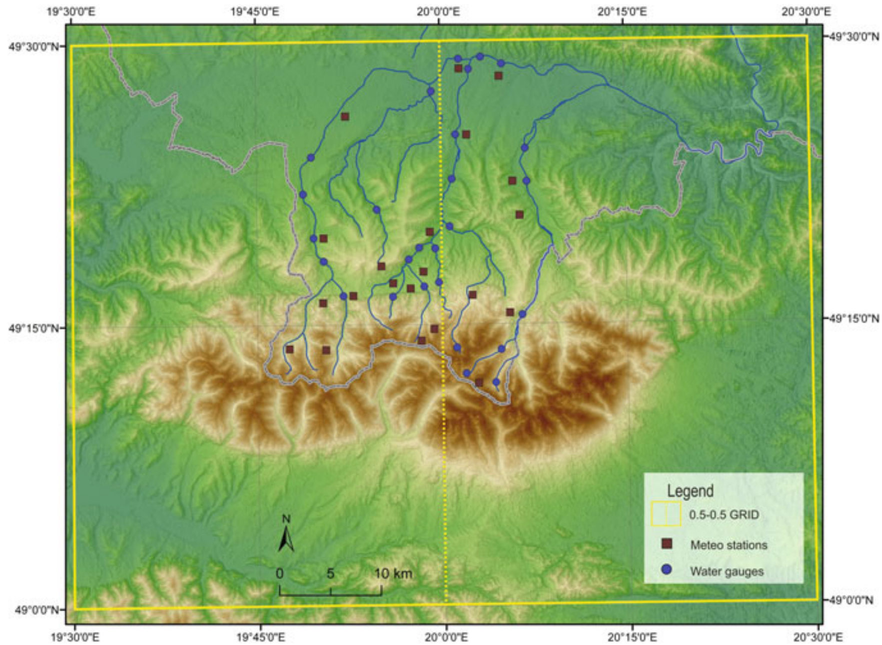

Fig. 1 Location of meteorological stations and flow gauges on the northern slopes and at the foothills of the Tatra Mountains (compare with Ballesteros-Cánovas et al. 2015). The spatial extent of the gridded data employed for dendroclimatic analyses and reconstructions (CRU TS3.23 (Harris et al. 2014))

2 Geographical Setting

The northern part of the studied region can be described as the Tatra Mountains and the foothills traditionally known as Podhale (Kondracki 1994). The Tatras are the highest mountain system (with Gerlach being the highest peak, 2655 m a.s.l., located in Slovakia) in the Carpathian arc. The Tatra Mountains are bio-geographically a part of the Western Carpathians (Warszyńska 1995).

Widespread forest stands that reach from the montane to sub-alpine zones are typical landscape elements with thermally induced treeline between 1500 and 1700 m a.s.l. along the north-south gradient. The dominant coniferous species of the sub-alpine zone is Norway spruce (*Picea abies* (L.) H. Karst: PCAB), occasionally Stone pine (*Pinus cembra* L.: PICE). The high-located forest of the Tatras has been under long-term human impact since at least the 15th century. The human activity was mainly related to logging and shepherding, affecting the structure of the mountain forest and lowering treeline of sub-alpine zone. Long-lasting selective logging resulted in decreasing the numbers of the Stone pine stands. The Scots pine (*Pinus sylvestris* L.: PISY) sites are constituted by small isolated groups of trees

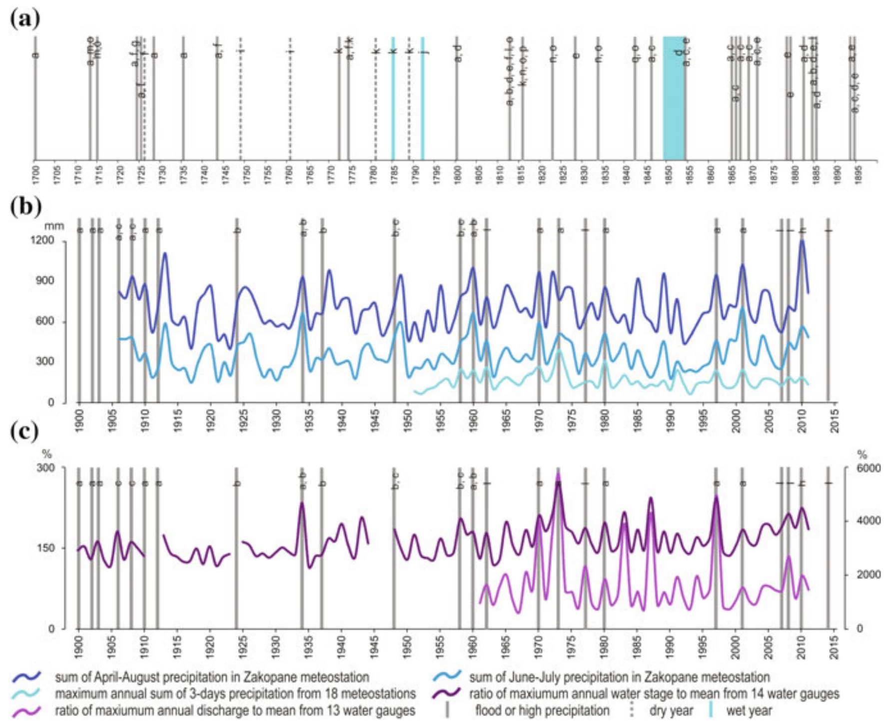

Fig. 2 The comparison of existing precipitation, water stage and discharge data (based on the network shown in Fig. 1) and historical records (*a*—Kotarba (2004), *b*—Wierchy (1923–2015), *c*—Bielański (1984), *d*—Stolarczyk (1915), *e*—Pamiętnik Towarzystwa Tatrzańskiego (1876–1920), *f*—Siemionow (1992), *g*—Generisch (1807), *h*—Woźniak (2013), *i*—Niedźwiedź (2004), *j*—Gustawicz (1883), *k*—Szewczuk (1939), *l*—Rajwa (2014), *m*—Komoniecki (1704), *n*—Adamczyk (1991a), *o*—Bednarz (2015), *p*—Adamczyk (1991b), *q*—Więckowski (1834))

located at the altitude of appr. 1000 m a.s.l. and related to combination of carbonate substrate soil and southern exposure (*Erico-Pinion*, Perzanowska 2010).

The Tatra Mountains represent natural climate conditions free of strong anthropogenic influences. The winter conditions are affected by polar-continental air masses arriving from the east and northeast while in the other seasons weather is predominated by oceanic polar-maritime air masses from the west. Foothills in the Western Carpathians are characterized by mean July temperature of ~19 °C compared to ~22 °C in the southern part, and 4.1 °C compared to 5.4 °C respectively for the highest locations (~2500 m a.s.l.). According to altitude, the mean annual temperature drops from about 6 °C at the level 600–650 m a.s.l. (Carpathian foothills, Orawa-Nowy Targ Basin) to −4 °C on the highest peak of the Tatra Mountains (Hess 1974; Niedźwiedź 1992). After Hess (1974), six vertical climatic belts of 2 °C width can be distinguished in the Western Carpathians and

Tatra Mountains. They are strictly connected with types of vegetation zones. For dendroclimatic investigations, the forest zone below the treeline (1550 m a.s.l.; mean annual temperature 2 °C, in July 10 °C) is the most important. A cool climatic belt (annual temperature between 2 and 4 °C) is located below this line and above the level of ~1150 m a.s.l. It is covered by the coniferous forest with *Picea abies* (L.) H. Karst. A moderately cool belt (from 4 to 6 °C) with mixed forest is located at the level 650–1150 m a.s.l.

The annual precipitation is lower than in the other mountains of Central Europe, such as the Alps and Czech Massive, due to the effect of the distance of around 1000 km to the Atlantic Ocean and the influence of those mountains on the mild air, thus increasing the continental character of upper-lands. The highest rates of annual precipitation (>2000 mm) are reported from the summits of the Tatra Mts and their northern slopes. In the Podhale region, the annual precipitation varies from below 800 mm in Orawa-Nowy Targ Basin to ~1100–1200 mm in Zakopane (Niedźwiedź 1992).

3 Materials and Methods

3.1 Tree-Ring Data

The tree-ring network spans across a northern part of the Tatra Mountains of ~55 km covering the Western, High and Belianske Tatras (Fig. 3). The network was established based on 10 PCAB and 5 PICE sites from high-elevation, near local timberline and six sites of PISY located in montane forest (Fig. 4). The data sets include tree-ring width series that were originally developed and downloaded from the International Tree-Ring Data Bank (ITRDB). The qualities of TRW

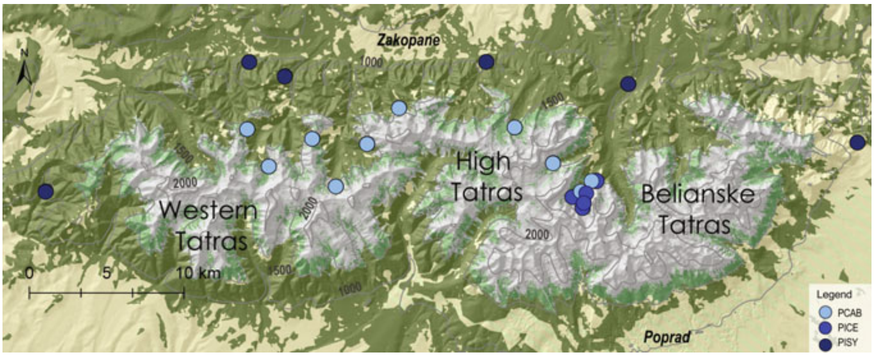

Fig. 3 Location of tree-ring sampling sites in the Tatra Mountains: *PCAB*—Norway spruce (*Picea abies* (L.) H. Karst), *PICE*—Stone pine (*Pinus cembra* L.), *PISY*—Scots pine (*Pinus sylvestris* L.)

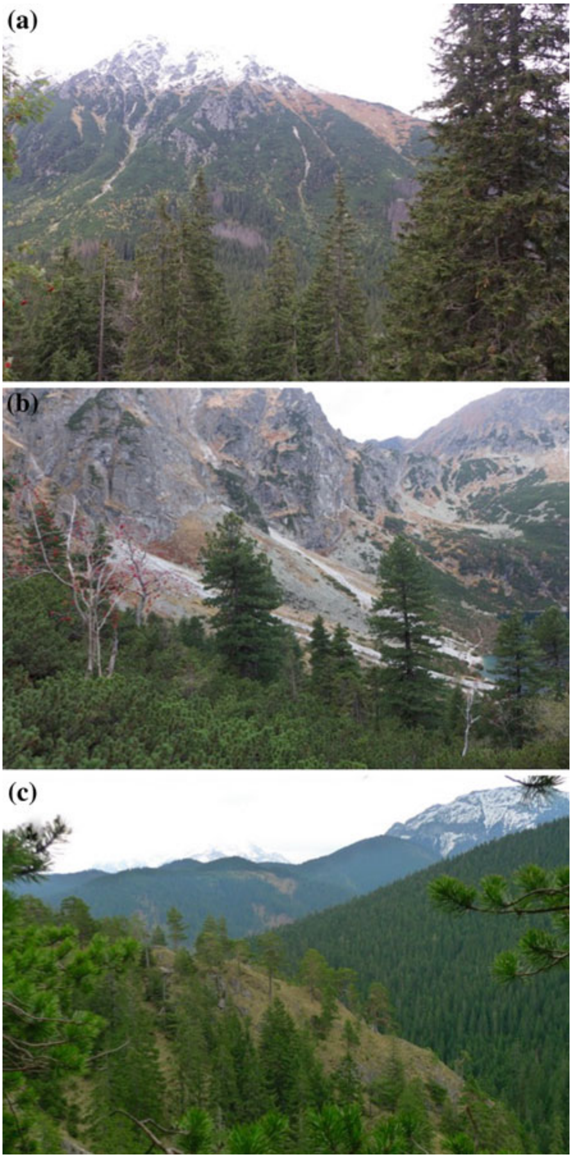

Fig. 4 Three types of the sampled forest: **a** Subalpine spruce forest, *Plagiothecio-Piceetum tatricum*, **b** Stone pine stands, *Pino cembrae-Piceetum*, **c** semi-dry Scots pine forest, *Erico-Pinion*

measurements were visually checked using CDendro program (Larsson 2003) and screened for possible mistakes and missing rings using statistical analyses of COFECHA program (Holmes 1983). To remove non-climatic, age-related growth trends from raw measurement series (Fritts 1976), spline-detrending standardization was applied to individual series using ARSTAN software (Cook 1985). Indices were calculated as ratios from 200-year cubic smoothing splines (Cook and Peters 1981). Signal strength of the chronologies was assessed using a 'moving window' approach of the inter-series correlation (*Rbar*) (Fritts 1976), and the expressed population signal (*EPS*; Wigley et al. 1984). For chronology development, series were averaged using the bi-weight robust mean (Cook 1985), and then truncated at a minimum sample size of five series and commonly accepted threshold of EPS = 0.85 (Wigley et al. 1984) (Fig. 4).

3.2 Climatic Data and Growth/Climate Response

The records of monthly resolved gridded (0.5° × 0.5°) temperature, precipitation and Standardised Precipitation Evapotranspiration Index (SPEI, Vicente-Serrano et al. 2010; Beguería et al. 2014) means from dataset CRU TS3.23 (Harris et al. 2014) were used. Two sets of boxes (19.5–20.5°E, 49.0–49.5°N) of gridded data covering the region were established (Fig. 1). The growth/climate response and reconstructions were computed employing the temperature and precipitation anomalies calculated against the commonly used 1961–1990 climatological normal period (New et al. 2000).

Growth/climate response analyses using an 18-month window from May of the year prior to the tree growth, up to October of the growing season, and various seasonal means (April–September = A–S, May–September = M–S, June–September = J–S, April–August = A–A, May–August = M–A, June–August = J–A, June–July = J–J) emphasized the dendroclimatic 'response' (Fritts 1976) of each chronology. Monthly and seasonal growth response of TRW chronologies to temperature and precipitation was computed using Pearson's correlation coefficient.

3.3 The Climate Reconstruction

The direct use of regression model for climate reconstruction caused a variance reduction effect and loss of variance therefor the scaling technique was employed. The means and standard deviations of a proxy (TRW series) were brought to the same level as the corresponding values of climate records (temperature and precipitation) over an established common period (Esper et al. 2005). The overlap period was used as the calibration period spanning over 1902–1956 and was the same for all three reconstructions (PCAB, PICE and PISY). The model was checked against the climate data over the verification interval, 1957–2013. The adequacy of the calculations was tested using the reduction of error (RE) and coefficient of efficiency (CE) (Fritts 1976). The Durbin-Watson (WD) statistics were computed to detect the autocorrelation in the residuals (Durbin and Watson 1951).

4 The Results and Discussion

4.1 The Tree-Ring Width Chronologies

The three coniferous chronologies were developed covering the last three centuries. The PCAB chronology was established based on 575 TRW series from 10 sites located at or near the timberline in the Polish part of the Tatra Mountains (Fig. 5). The spruce chronology is the longest, spanning the interval 1637–2013, after truncation according to the series replication and EPS. The PICE chronology represents smaller and more concentrated sites (341 series from five sites) and covers a shorter period (1735–2013). Both species represent the subalpine environment but Stone pine grows often above the spruce-dominated timberline and often creates the treeline. Therefore PICE is more exposed to severe climate conditions (Carrer et al. 2007; Janecka and Kaczka 2015). Nevertheless, the growth of both conifers is similar and for common period of 1735–2013 standard chronologies correlate at 0.45 (p = 0.01). The PISY chronology was established utilizing 135 series derived from 6 sites of relatively low elevation and characterized by carbonated bedrock of cliff-like locations. The signal strength of all chronologies was assessed by Rbar and ranged from 0.28 to 0.60, whereas the EPS values varied amongst 0.84–0.99, indicating the internal consistency in the common variance of chronologies.

The inter-annual variability allows the identification of pointer years, caused by singular events, as summer cooling effects due to the radiative forcing of volcanic eruptions (e.g. 1816 and 1912/13) (Briffa et al. 2001, 2004). Two timberline chronologies showed this kind of short-term changes whereas PISY chronology did not demonstrate both cooling periods. The only one pronounced, well synchronised

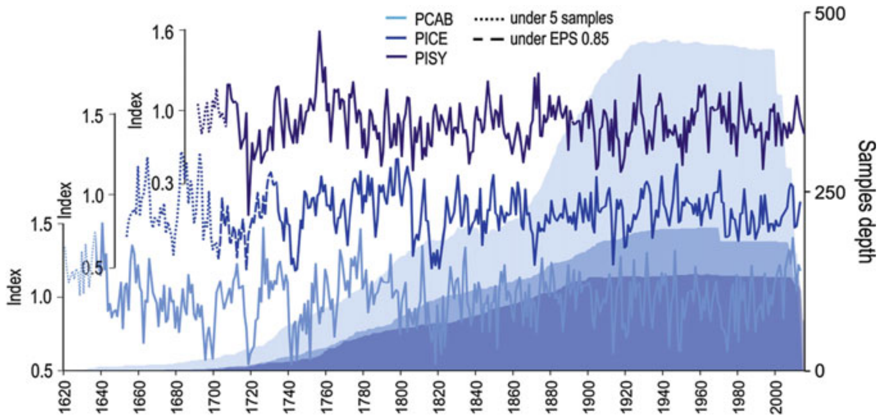

Fig. 5 Three (*PCAB*, *PICE*, *PISY*) standard chronologies and sample depth. The *solid line* represented the period exhibiting EPS value over threshold of 0.85 (Wigley et al. 1984) and/or trucked at <5 series

growth reduction for all three chronologies occurred in 1718, which could also be linked to volcanic activity (Crowley 2000; Robock 2000). The lowest growth rates are observed during the Dalton Minimum and coincide with an increased volcanic activity in the early 19th century and followed by improved radial growth momentarily interrupted by light decreases in the 1910–20s, and 70–80s. Interestingly, only PCAB chronology indicated the highest productivity within the last two decades. The previously constructed TRW chronologies emphasized the growth reduction in the 1970s and 1980s as a result of air pollution (Büntgen et al. 2007; Bednarz 2015; Treml et al. 2015). In this work, highly replicated chronologies do not exhibit this issue.

4.2 *Growth/Climate Response*

The response of tree-ring width differs between two subalpine and low elevation species. The growth of PCAB and PICE is driven by summer temperature whereas the growth of PISY mainly depends on humidity (precipitation and SPEI). The highest correlations between TRW and temperatures were found for June-July of the current year for PCAB chronology (r = 0.69, $p < 0.01$) (Fig. 6). The PICE standard chronology correlates with temperature of the same season (June–July) but with lower value (r = 0.49, $p < 0.01$). The growth of both species responds to several factors, but temperatures of June and July are the most important. The impact of temperature of longer seasons, such as April–September, May–October and April–October is also significant. The correlation with previous year autumn and current year late winter differentiates PCAB and PICE chronologies. Only Norway spruce chronology correlates positively with October in previous year ($p < 0.01$).

Similar results were reported in literature from several high elevation sites (Oberhuber 2004; Frank and Esper 2005b; Büntgen et al. 2007). The growth of the recent season could be linked with the amount of the reserves inherited from previous year warm autumn. The negative correlation with March temperature is individual feature of PICE chronology. The warm March decreases the possibility to produce wide ring in PICE stems (r = 0.28, $p < 0.01$). Vaganov et al. (1999) reported similar phenomena from the Ural Mountains and pointed out the danger posed by early onset of vegetation provoked by a warm end of winter and the beginning of spring. Precipitation has little influence on the growth of both subalpine species. The PCAB chronology correlates positively with January, March and January–March precipitation ($p < 0.01$) but the relationship is much lower than for summer temperature. The clear influence of June–July temperature signal allowed to select it as a reconstructed factor. The PISY chronology exhibits different relation with climate. The correlations with temperature are rather low while both humidity parameters (precipitation and SPEI) drive the growth of Scots pines. The PISY chronology exhibits statistically significant correlation values for precipitation and availability of water for the entire growing season, including spring

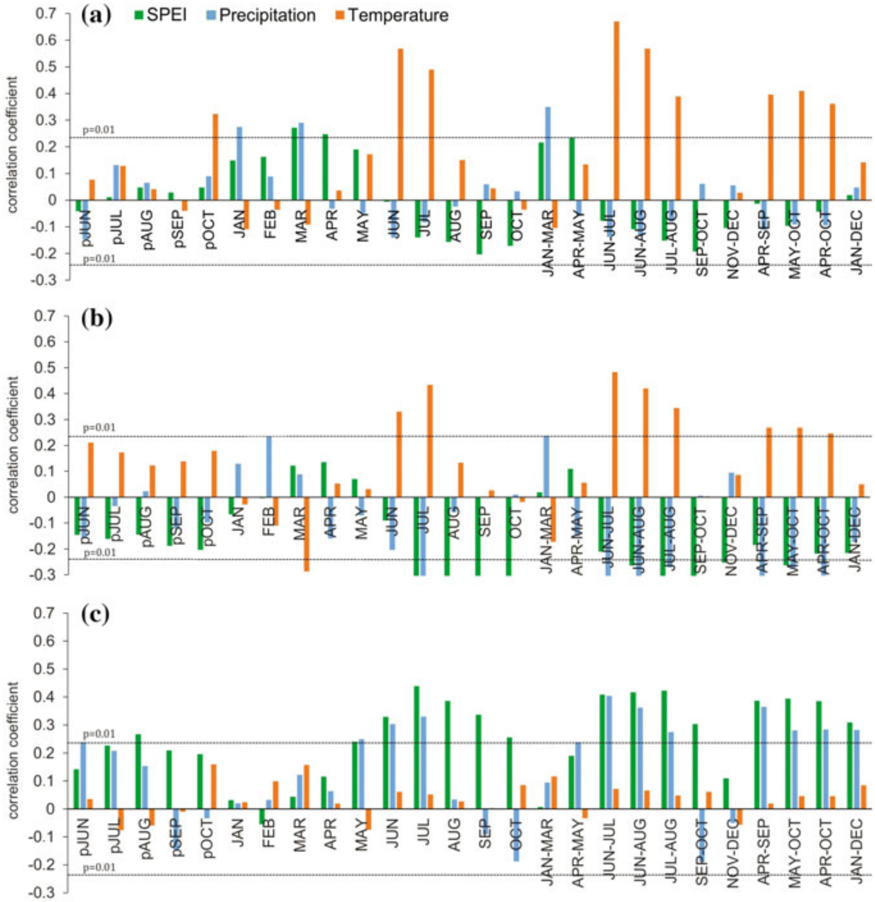

Fig. 6 The growth/climate response of three chronologies: **a** PCAB, **b** PICE, **c** PISY

and autumn and even annual means. The strongest relationship in the PISY chronology is present for summer precipitation, especially June–July (r = 0.41, $p < 0.01$) and July SPEI (r = 0.44, $p < 0.01$). The June–July precipitation constitutes an appropriate climatic factor for the TRW reconstruction.

One of the most important features of growth/climate response is the temporal stability of such correlation. The results of the 21-years running correlation between June–July temperature in case of PCAB and PICE and June–July precipitation reveal the changes over 1902–2013 period. The temperature signal registered in TRW of spruce is stable for most of the testing period (Fig. 7a). There are some fluctuations and except for the 1960, the values of Pearson correlation remain significant at $p = 0.05$. Although the stationary correlation between PICE chronology and June–July temperature is comparable to PCAB chronology, the

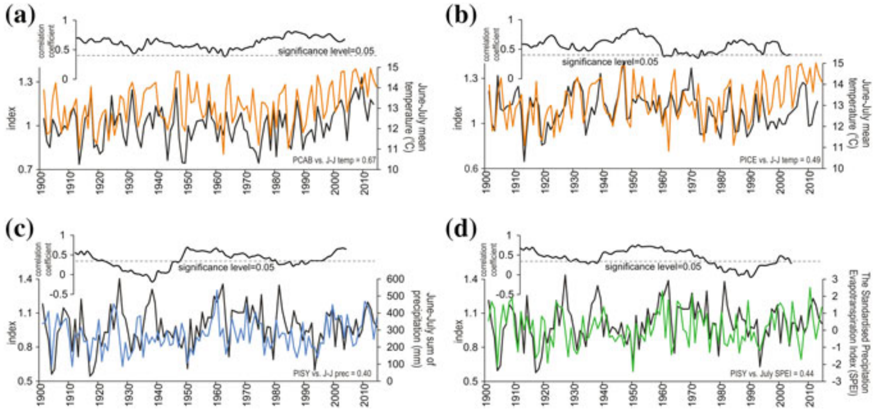

Fig. 7 Comparison of chronologies and selected climate factors over common period of 1901–2013: **a** PCAB and June–July temperature, **b** PICE and June–July temperature, **c** PISY and June–July precipitation, **d** PISY and June–July SPEI. The temporal stability of TRW/climate response was tested using 21-year running correlation

signal is much less stable over time. The results are not significant for longer period between 1960 and 1980 and also in the beginning of the 21st century (Fig. 7b). The lack of temporal stability of climate signal carried by TRW series suggests also poor performance of temperature reconstruction. The PISY chronology also exhibits low temporal stability of climatic signal for both precipitation and SPEI (Fig. 7c, d, respectively).

4.3 The Climate Reconstruction

The climate variables revealing the highest correlations with TRW chronologies were selected to develop climate reconstruction. In case of PCAB, June–July temperature was chosen based on the fact that this variable explains 69 % of variance and is stable over the tested period of $\sim$100 years. Although slightly lower variance is explained (49 %) and temporal stability is weaker, the mean temperature of the same season was selected for reconstruction based on the PICE chronology. Since the PISY chronology reveals completely different relationship with climate, the July SPEI and June–July precipitation (explaining 44 and 41 % of variance, respectively) were applied for climate reconstruction. The temporal stability of both is not fully satisfactory but they are also somehow complementary. The period of lacking significant correlation for precipitation showed high values for SPEI. The calibration and verification statistics (Table 1) exhibit positive values for the RE and CE indicating the predictive skills of the model. The value of DW suggests low autocorrelation of residuals. The results of the same procedure performed for PISY chronology and humidity time series (July SPEI and June–July

Table 1 Statistics of the calibration (1902–1956) and verification (1957–2013) for PCAB and PICE June–July temperature, and PISY June–July precipitation reconstructions

Chronology	Period	Calibration		Verification		
		R^2	DW	R^2	RE	CE
PCAB	1902–1956	0.59	1.1	0.58	0.58	0.55
	1957–2013	0.75	1.8	0.58	0.37	0.33
PICE	1902–1956	0.57	1.3	0.56	0.52	0.45
	1957–2013	0.59	1.5	0.58	0.51	0.55
PISY	1902–1956	0.54	1.7	0.52	0.55	0.53
	1957–2013	0.51	1.6	0.54	0.57	0.49

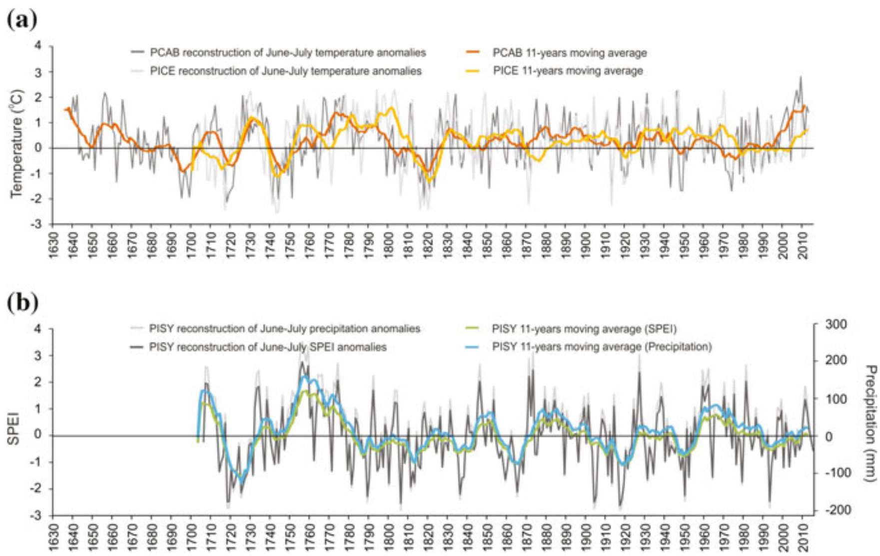

Fig. 8 Climate reconstruction of the Tatras Region, over the period 1705–2013 AD: **a** Reconstructed June–July mean temperature anomalies based on PCAB standard chronology and PICE standard chronology, **b** reconstructions based on PISY standard chronology June–July precipitation and July SPEI anomalies

precipitation) show higher agreement for SPEI than for precipitation but the calibration and verification statistics present the predictive skills of both models (Table 1).

Reconstructions of summer temperature, based on the Norway spruce and Stone pine, reveal high similarity (long-term trends) and also several differences (in short-term trends and performance in the most recent interval) (Fig. 8). The PCAB reconstruction is longer and covers the cold period, ~1640–1700, induced by the Late Maunder Minimum (Eddy et al. 1976; Niedźwiedź 2010) followed by a short interval of warmer summers in the three first decades of 1700s. During the cold phase 1643–1651, two major flood occurred (1650 and 1651) (Siemionow 1992).

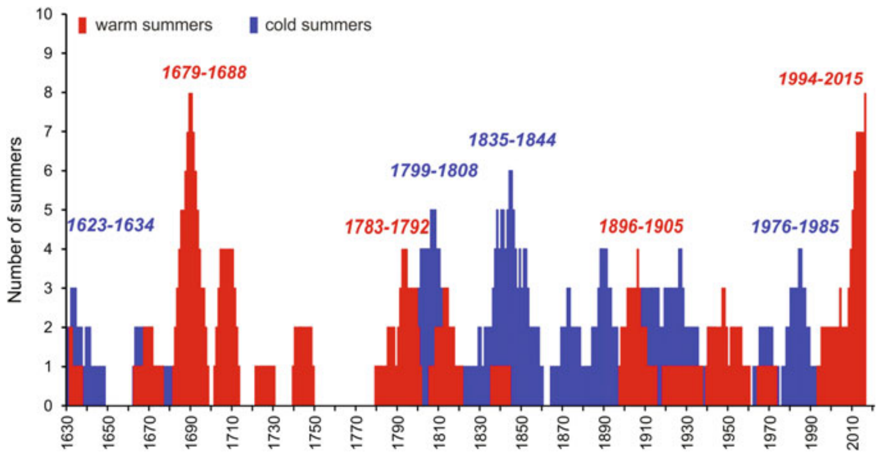

Fig. 9 Variability of the number of extreme warm and cold summers (June–August) in the Tatra Mountains (Hala Gąsienicowa 1520 m a.s.l.) during the consecutive 10 year periods

This was succeeded by warmer period of the 1660s, which was also noted in historical record (Bokwa et al. 2001) followed by the next cooling and warming (1670s and 1680s, respectively) (Fig. 9). The next cold period lasted also one decade but the temperature decrease was more severe, by almost 2 °C comparing with 1 °C for earlier variations. The year 1695 was very cold, due the eruption of Hekla, Serua and Aboino volcanoes (Grove 1988). This agrees with reports from other locations in Europe.

The next similar cooling happened in the 1710s, with the minimum in 1718 when the temperature dropped by 2 °C and reached the minimum of the reconstructed period. That cooling interval, with duration reaching almost a decade, could be associated with a series of volcanic activities in 1715–1718. The historical records inform about heavy snowfalls in July and August 1724 and again August 1725 (Siemionow 1992). The following years were warmer but can be characterized by oscillations of temperature (from +2 °C in 1726 to 0.0 °C in 1729). The common period of both temperature reconstructions started from 1737. The 1740 marked the beginning of the next cool phase, which has lasted till 1753. The volcano activity increased in 1739 and 1740 and could have resulted in decrease of temperature in the following years. The 1750s and 1770s were the next period of temperature changes around 0.7 °C with exception of 1771. The PICE reconstruction registered pronounced cooling whereas PCAB showed this year as 0.3 °C above the average. The remaining part of the 18th century was relatively warm. The eruption of Laki volcano (1783) could be linked with short but severe temperature drop (up to 1.0 °C in 1886) captured by the PICE chronology.

The next significant cooling at the beginning of 1800s was related to the Dalton solar minimum (Eddy et al. 1976) and major volcanic activity (Crowley 2000; Robock 2000). The Tambora eruption on April 1815 (VEI = 7; Newhall and Self 1982) played the most crucial role in inducing the coldest period with most severe and longest in duration temperature drop. Depending on the reconstruction, the temperature dropped by 2.4 °C (PICE) or 1.9 °C (PCAB) in comparison with the average. The Stone pine chronology registered the minimum temperature in 1816 while Norway spruce in 1818. The delay in the response of severe volcanic eruption-related cooling was discussed by Esper et al. (2015) and D'Arrigo et al. (2013). The cooling was probably less pronounced in Central Europe than in Western Europe but the term "year without summer" (1816–1818) undoubtedly applies also to the Tatra Region (Grove 1988; Harrington 1992; Bednarz and Trepińska 1992). The following decades showed sine-like cycle of dropping and rising temperature but with slightly positive long-term trend. The noticeable cooler periods occurred in 1843–44, 1851–54 and 1864–1871 mainly registered by PICE chronology. They were also mentioned by the historical sources reported the crop yield failures (Stolarczyk 1915). The cool summers mentioned above belong to the second phase of millennial climate variability named the Little Ice Age (1575–1850), cf Grove (1988). In reconstructed temperature series this cooling is prolonged to 1870 (Fig. 8a). The similar picture is observed in the tree-ring chronology derived from Norway spruce growing on Babia Góra Mountain (Bednarz 2015). After temperature reconstruction for Hala Gąsienicowa (1520 m a.s.l.) near the treeline in the Tatra Mountains, the end of the Little Ice Age can be placed around 1895 (Niedźwiedź 2004).

The beginning of the 1900s marked the end of the Little Ice Age in the Tatras Region, which lasted longer than in the other parts of Europe (Niedźwiedź 2004). Well documented cooling caused by the Katmai eruption in 1912 interrupted the warm period. The reaction to this event was delayed by one year and the summer of 1913 was colder by more than 2 °C in the Tatras. The summer of 1913 is the coldest one also in the other temperature reconstructions (Bednarz 1976, 1983, 1984, 2015; Bednarz et al. 1998–1998; Niedźwiedź 2004). This drop of temperature was registered by both species but PICE reacted stronger also creating strong signal in the form of pale rings (Janecka and Kaczka 2015).

The following decades again showed since-like cycle of dropping and rising temperature but with higher amplitude which could be attributed to the increase of climate continentality (Niedźwiedź 2004). The 1960s and 1970s were the most recent colder phases but less pronounced in comparison to the Little Ice Age period. The PICE chronology revealed problems to perform well during this period. The recent warming has been present in the Tatras since the end of 1980s, which is documented by instrumental data (Żmudzka 2009, 2010) and behaviour of the PCAB reconstruction. Interestingly, the TRW of PICE failed to capture this change. Stone pine often grows above the timberline in harsher conditions where locally and globally induced changes of temperature influence the growth stronger.

In the instrumental data on the treeline on Hala Gąsienicowa (1520 m a.s.l.) the last cool summer (8.3 °C) was indicated in 1984. After 1987 11-year average temperature increased from the level of 9.9 to 11.4 °C during the last 11 years (2005–2015). The summer of 2015, with temperature 12.9 °C was the warmest in the whole temperature series (where the long-term average is 10.0 °C). In Fig. 9, the variability of numbers of extreme warm (>11 °C) and cool (<9 °C) summer seasons (above and below the 10 and 90 ‰) on the treeline has been presented in consecutive decades since the beginning of the 18th century (Niedźwiedź 2004, data updated to 2015). In the last decade (2005–2015), eight summers were extremely warm. The four warmest summers are also noted in the decades 1896–1905 and 1783–1792. The six coolest summers are located in the decade 1835–1844 and five in the years 1799–1808 during the final phase of the Little Ice Age. The last decade with four coolest summers spans over the years 1976–1985.

The summer precipitation and Standardized Precipitation Evapotranspiration Index, SPEI, were reconstructed based on the PISY chronology. The changes of humidity-related climate variables are less pronounced over the reconstructed period of 1710–2013 (Fig. 8b). The cold period of the Late Maunder Solar Minimum (Eddy et al. 1976) is also the time of increase of summer humidity with high SPEI index (3.02) in 1707. In the Western Carpathians, after historical chronicles, three consecutive summers 1713–1715 were rainy and cold (Bednarz 2015). Relatively short, but deep dry period 1717–1726 preceded the long wetter summer conditions in 1733–1779 with the highest SPEI index (5.54) in 1756. The most humid conditions (SPEI > 2.00) in 1751–1775 coincide with the large increase of temperature after the cool period 1740–1752 (compare Fig. 8a). The mega drought in Western Europe, reconstructed by Cook et al. (2015), is also visible by decreasing the humidity after ~1770 (Fig. 8b) to the SPEI below −1.00 in 1789. The local but more severe drought happened in amount of summer precipitation and SPEI in 1806 (−1.34) and 1836 (−1.01). Extreme cooling during the Dalton Minimum of solar activity at the beginning of the 19th century (Fig. 8a and Bednarz 2015) is not visible in the SPEI variability. But in the historical sources, damaging floods are indicated in Nowy Targ in 1813, 1816, 1823 and 1834 (Bednarz 2015). Significant increases of SPEI index (3.08 in 1846) are strongly marked during 1840–1853 in the final cool phase of the Little Ice Age. This wet period was also reported in historical documents and tree-ring reconstructions from Moravia, Czech Republic (Büntgen et al. 2011). The last 30 years of the 19th century (1871–1900) can be distinguished as a very humid interval, with maximum SPEI in 1871 (3.26) and 1873 (3.48). In the 20th century, three wet periods (1924–1942, 1954–1987 and 2006–2010) and three drier intervals concentrated around the years 1917, 1950 and 1993 were observed.

The correlation between flood events and reconstructed or registered humid periods (Fig. 10) are difficult due to the predominant character of the flood caused by short-term intensive precipitation. Lack of the synchronicity between the

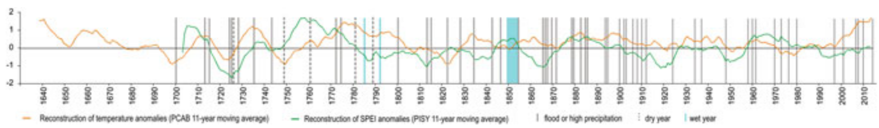

Fig. 10 Compression between the historical and instrumental flood records and reconstruction of summer temperature and precipitation

long-term and even medium-term changes of humidity and flood occurrences poses an important challenge for predicting the occurrence of floods and managing of this natural risk. The comparison of instrumental data enriched by historical records of floods with climate reconstructions demonstrates that even during drier period major flood can happen (as in intervals: 1721–1730 or 1865–1872).

5 Conclusion

The reconstruction of climate factors different than temperature in mountain regions of temperate climate is challenging but allows us to pose more complete image of paleoclimate. The coupling of regional reconstruction of humidity and temperature for the Tatra Mountains spanning an interval in excess of 300 years proved to be a promising approach to reconstruction of climate data. The temperature reconstruction based on two subalpine species demonstrates benefits of combining the records of Norway spruce and Stone pine proxies. The former better register long-term changes and latter are more susceptible to short-term cooling, clearly indicating events of more extreme nature.

The compilation of all performed climate reconstructions for the last three centuries allows us to conclude:

- The 17th–18th period of the Little Ice Age was characterized by complex changes of climate. The cold phases were followed by warm periods and single warm and cold years were frequent. The beginning of the 19th century was associated with a more pronounced cooling caused by the Dalton Solar Minimum and volcanic activity (Tambora eruption in 1815). The second part of the 19th century was also relatively cold and the end of the Little Ice Age in the Tatras Region could be placed at the beginning of the 20th century.
- Considering the entire 20th century, the recent warming is less pronounced. The significant increase of summer temperature has started since the 1980s and the 1990s and was registered by Norway spruce TRW whereas PICE failed to capture it.
- After 1985, cool summers (below 90 ‰) did not occur and in the decade 2006–2015 the number of the warmest summers (above 10 ‰) increased to eight.
- The temperature reconstructed from PCAB and PICE TRW reveals largely individual character of the Tatras in comparison with the Alps and Europe. This problem was also reported by Büntgen et al. (2007) and D'Arrigo et al. (2008).
- The short-term global events are better synchronised. The short (one or few years) decreases of summer temperature, mainly caused by volcanic activity are clearly present in the climate of the studied region. The 1718 cooling is the first and severe event, followed by 1816–20, 1912–13 and 1974–1980 periods.
- The correlation between flood events and reconstructed humid periods is poor due to the predominant character of the flood caused by short and intensive precipitation.

- Lack of the synchronicity between both long- and short-term changes of humidity and flood occurrences poses an important challenge for direct implementation to predict the floods and manage that natural risk. The comparison of instrumental data enriched by historical records of floods with climate reconstructions proves that even during drier period major floods can occur.

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