TITLE PAGE

A 7.5 ka chronology of stable carbon isotopes from tree rings with implications for their use in palaeo-cloud reconstruction

Samuli Helama[a,*], Laura Arppe[b], Mauri Timonen[a], Kari Mielikäinen[c], Markku Oinonen[b]

[a]Natural Resources Institute Finland, Eteläranta 55, Rovaniemi, Finland
[b]Laboratory of Chronology, Finnish Museum of Natural History, Gustaf Hällströmin katu 2, University of Helsinki, Helsinki, Finland
[c]Natural Resources Institute Finland, Tietotie 2, Espoo, Finland

*Corresponding author: Samuli Helama, samuli.helama@luke.fi, +358295325047

Declarations of interest: none

Keywords: isotope geochemistry; palaeoclimate reconstruction; geochronology; dendrochronology; regional curve standardization; subarctic

A 7.5 ka chronology of stable carbon isotopes from tree rings with implications for their use in palaeo-cloud reconstruction

Abstract

Tree-ring stable isotope chronologies provide very high-resolution palaeoclimatic data, and the number of records is increasing rapidly worldwide. To extend the chronologies back in time, before the period covered by the old living trees, the use of subfossil wood samples is required. Typically, the longest continuous subfossil chronologies consist of regionally collected tree-ring materials, rather than tree rings from a single site, and are likely more sensitive to data heterogeneity. Yet, the characteristics of such datasets remain hitherto unexplored. Here we produce a continuous, decadally resolved chronology of $^{13}C/^{12}C$ ratio ($\delta^{13}C$) from Finnish Lapland over the past 7.5 ka (5500 BC to AD 2010) for which there is replication of at least five *Pinus sylvestris* trees. Less negative $\delta^{13}C$ values were observed as trees age and for western sites (higher in elevation and further from the cold oceanic air flow). The age-related trends in living tree $\delta^{13}C$ data were expressed mainly over the first fifty years mimicking the “juvenile effect” whereas the subfossil data showed trend over the trees’ lifespan. These findings demonstrated the need to detrend the individual $\delta^{13}C$ series before averaging them into the mean chronology. These biases were removed from the isotopic data using the methods frequently applied for tree-ring width and density proxies, the Regional Curve Standardization (RCS) combined with signal-free approach. While the RCS procedures commonly preserve the long-term variations in the resulting chronology, not all types of them did so as demonstrated for chronologies produced using separate RCS models for the $\delta^{13}C$ series with relatively high and low isotopic level (offset from the grand mean). It was shown that these $\delta^{13}C$ levels (i.e. the relative isotopic enrichment) result both from the low-frequency climate signal and biogeographical aspects (the site longitude/altitudes). The non-climatic biases were removed from the $\delta^{13}C$ series by using separate RCS models for the subsets (western and eastern) of isotopic series. Similar to previous investigations using annually resolved $\delta^{13}C$ data from Lapland, our chronology had strong negative correlations to variations in cloud cover. Here, a bootstrapping experiment was used to verify this dendroclimatic association. The resulting palaeo-cloud reconstruction portrayed decadal to multi-millennial variations with centennial anomalies coinciding with the mid and late Holocene events of climate transitions, highlighting the value of subfossil isotope chronologies from tree rings in synthesising climate dynamics from several proxy sources over the present interglacial.

1. INTRODUCTION

Palaeontological archives provide us with the possibility of developing long isotopic chronologies and, among plant remains, the sampling of subfossil tree-ring cellulose offers the potential for high-resolution records of past climates and environments. In fact, the tree-ring stable isotope data from old living trees constitute chronologies commonly spanning the past several hundred years only. Clearly, the use of subfossil tree rings is needed to extend these isotope chronologies back in time (Becker et al., 1991; Sonninen and Jungner, 1996; Boettger et al., 2003; Mayr et al., 2003; Frumkin 2009; Edvardsson et al., 2014; Helama et al., 2018b). Apart from being annually resolved, tree-ring based records of isotope ratios are benefitted from dendrochronological dating (Stokes and Smiley 1968; Fritts, 1976), according to which the boundaries of cross-dated rings are providing a framework for cellulose samples with exact calendar year position. Combined with data from living trees, isotopic data from subfossil tree rings have so far allowed for continuous, high-resolution reconstructions of past climate variability over roughly the past millennium (Gagen et al., 2011, 2012; Young et al., 2012; Loader et al., 2013a; Naulier et al., 2015; Gennaretti et al., 2017).

Palaeoclimatic signals are stored in plant carbon as trees react to prevailing climatic conditions through their stomatal and photosynthetic responses. The $^{13}C/^{12}C$ ratio ($\delta^{13}C$) observed in tree rings can be explained using the equations of carbon isotope fractionation (Farquhar et al., 1982; Francey and Farquhar, 1982). These models predict increase in $\delta^{13}C$ when the CO_2 concentration decreases inside the leaves in relation to that outside the leaves (i.e. in the air) as trees stressed by drought close their stomata or their leaves are exposed to more sunlight and assimilate more carbon for their photosynthetic products. In terms of palaeoclimatology, this framework allows not only for interpreting the $\delta^{13}C$ variations observed in tree rings but for selecting sites potentially suitable for reconstructing either moisture or irradiance/could cover changes using the long $\delta^{13}C$ chronologies.

Their benefits notwithstanding, the isotope chronologies come with uncertainties stemming from non-climatic factors affecting the trees through their lifespans (McCarroll and Loader 2004). As for the carbon isotopes, these data are presumed to show a trend towards less negative ratios during the juvenile growth phase (Francey and Farquhar, 1982; Duquesnay et al., 1998; Gagen et al., 2008) that comprise the very first decades of tree's life. The trends may however continue beyond the "juvenile effect" and be persistently identified in tree-ring isotope data through much of the present interglacial time (Mayr et al., 2003; Helama et al., 2015). More generally, tree-ring evidence suggestive of age-related trends in D/H (Mayr et al., 2003), $^{18}O/^{16}O$ (Treydte et al., 2006; Esper et al., 2010) and $\delta^{13}C$ ratios (Esper et al., 2010, 2015; Helama et al., 2015) have been reported for several tree species and habitats.

Moreover, the longest continuous tree-ring chronologies are constructed from hundreds to more than a thousand subfossil samples and while no single site can provide such a supply, each of the chronologies actually represents a range of sites with varying characteristics e.g. altitudes and latitudes/longitudes (Pilcher et al., 1984; Eronen et al., 1999, 2002; Grudd et al., 2002; Hantemirov and Shiyatov, 2002; Kelly et al., 2002; Helama et al., 2004, 2010; Nicolussi et al., 2010). With these regards, it is known for example that carbon isotope discrimination depends on site altitude (Hultine and Marshall, 2000; Warren et al., 2001), this effect resulting in altitudinal differences in tree-ring $\delta^{13}C$ values (Leavitt and Long, 1992; Helama et al., 2018a). Averaging isotope data of (i) varying ontogenetic tree age or from (ii) differing sites into the chronology could at least theoretically create non-climatic perturbation in that chronology and potentially violate the palaeoclimate interpretations that follow. In addition, the tree-ring isotope ratios are rather commonly assessed after the pooling of the growth rings of several trees and same calendar year (Dorado Liñán et al., 2011; Liu et al., 2012). While this procedure offers the advantage of obtaining isotopic values more inexpensively and rapidly, such mixtures of wooden samples are likely to result in produced isotope data where the effects of signal (i.e. climate) and noise (i.e. inter-tree and -site variability) are not only mixed but also inseparable in any subsequent data analysis. Yet, there is a pressing need to acquire more information on non-climatic factors driving the isotope offsets between the trees and sites (Leavitt 2010).

Here we produce a 7.5 ka long tree-ring $\delta^{13}C$ chronology running continuously over the mid and late Holocene. Instead of pooling, we opted for the production of separate $\delta^{13}C$ series from individual *Pinus sylvestris* trees, the assemblage comprising a composite of nearly three hundred subfossils and living pine trees from northern Lapland (Finland). Our decadally resolved $\delta^{13}C$ chronology presents a ~ 6 ka extension to any of the existing $\delta^{13}C$ chronologies constructed from the northern timberline sites (Young et al., 2012; Loader et al., 2013a; Naulier et al., 2015; Gennaretti et al., 2017). Our $\delta^{13}C$ data originate from the region where annually resolved tree-ring $\delta^{13}C$ have shown to be sensitive to non-stomatal limitations and correlate positively with summertime changes in temperature and irradiance and negatively those in cloud cover (Young et al., 2010, 2012; Loader et al., 2013a; Helama et al., 2016, 2018a) and these are the climate variables that our $\delta^{13}C$ chronology is expected to record. Moreover, a subset of this data was previously shown to portray age-related trends contributing to a notable increase in $\delta^{13}C$ values through the trees' lifespans and thus clearly beyond the "juvenile" effect (Helama et al., 2015). We aim to identify the presence of any similar trends in this enlarged $\delta^{13}C$ dataset and detect the potential biases from not removing such non-climatic trends from the individual $\delta^{13}C$ series before calculating the mean chronology.

Our analyses of $\delta^{13}C$ data were carried out using the state-of-the-art methods developed for the purpose of removing the age-related bias from the tree-ring width and density proxy series, in particular the regional curve standardization (RCS) method (Briffa et al., 1992, 1996; Melvin and Briffa, 2008; Briffa and Melvin, 2011; Helama et al., 2017b). Importantly, this method have been shown to retain much of the climate-driven long-term (i.e. low-frequency) variability in the resulting tree-ring chronologies while still eliminating the age-related trends and reducing data heterogeneity (e.g. Briffa et al., 2013; Melvin et al., 2013; Matskovsky and Helama, 2014; Yang et al., 2014; Wilson et al., 2016; Anchukaitis et al., 2017; Helama et al., 2018b). Such RCS chronologies have previously been produced from shorter, late Holocene $\delta^{13}C$ and $\delta^{18}O$ records (Gagen et al., 2008; Esper et al., 2010; Helama et al., 2016, 2018b) but here we first time apply the RCS standardization for a multi-millennial tree-ring based isotope dataset. Further comparisons with the meteorological data were to identify the climatic signals in the $\delta^{13}C$ data, the reconstruction of climate variable best corresponding to the $\delta^{13}C$ proxy being built and validated using the instrumental and independent proxy evidence. These analyses evaluate the methods potentially suitable for combining large sets of isotopic data from plant carbon for assessing the low-frequency climate variability from tree-ring $\delta^{13}C$, in agreement with the models of carbon isotope discrimination (Farquhar et al., 1982; Francey and Farquhar, 1982), and contribute toward our understanding of inter-tree and -site variability in tree-ring isotope data over correspondingly long time-scales, essential to studies using stable isotopes to critically assess the palaeoclimate and long-term environmental changes.

2. MATERIALS ANS METHODS

2.1. Sites

Our study sites include the northernmost part of the boreal forests in northern Finnish Lapland where previous investigations have identified substantial amount of pinewood preserved as subfossils in a limited number of lakes (Eronen, 1979; Eronen et al., 1999, 2002). In addition to previously discovered material, new samples were unearthed from the lake sediments during the summer of 2012. Subfossil trunks were pulled to the shore, sawn into disks, and the trunks were returned into the lake sediments. In this study, the samples representing Scots pine (*Pinus sylvestris* L.) wood originate from 18 lake sites (Fig. 1). From south to north, the sites are characterised by the northern pine-dominated boreal forests, the subarctic timberline conditions and finally by solitary living pines. Our sites belong to this zone representing both the timberline and northern protection forest areas (Pohtila and Timonen, 1980; Veijola, 1998). Altogether 276 subfossil trees were used for isotopic analyses, among which 112 (40.6 %) originate from western and 164 (59.4 %) from eastern sites (Table S1), the boundary between the two sub-regions approximating 25°E. Living *P. sylvestris* trees were sampled at the breast height (1.3 m) with an increment borer at the same sites. Tree-ring samples from 22 living pines were used for isotopic analyses. Among them, nine (40.9 %) originate from western and 13 (59.1 %) from eastern sites (Table S1). The mean altitude the samples represent averaged 262 m above modern sea level. The sampling sites belong to the area of glacio-isostatic rebound of the earth’s crust and the modern altitudes overestimate the actual altitudes at which the trees’ have once grown. After adjustment for land uplift (Helama et al., 2004, 2010), the samples obtain a mean altitude of 234 m.

2.2. Wood samples

Tree rings were measured under the light-microscope to the nearest 0.01 mm and the ring-width time-series cross-dated against the existing master chronologies built previously in the same region

(Eronen et al., 1999, 2002; Helama et al., 2008 and unpublished data). Cross-dating was carried out both visually and statistically (Holmes, 1983). Cross-dating is a dendrochronological routine (Stokes and Smiley 1968; Fritts, 1976) where tree rings become dated to exact calendar year of wood production. Subsequent to cross-dating, the ring-boundaries provided a framework for wood material to be dissected from the sample disks under the microscope using a scalpel. Both earlywood and latewood were included in each sample, this choice being justified by observations of a very high common carbon isotope signal between the two portions of the annual ring for *P. sylvestris* (Kress et al., 2009). Tree-ring dated samples were separated as non-overlapping decadal blocks (i.e. 5500 to 5491 BC, 5490 to 5481 … AD 2001 to 2010) and samples processed to α-cellulose using a batch-approach designed by Wieloch et al. (2011), featuring multiple glass funnels connected via custom-built PTFE drainage blocks. The samples were treated with two alkaline extractions (5-7% NaOH) and a chlorination step ($NaClO_2$) in between (Wieloch et al., 2011), after which they were homogenized using an ultrasonic probe (Laumer et al., 2009). Finally, the samples were freeze-dried. The treatment resulted in dry cellulose fibres (ca. 70 µg) that were combusted for isotopic analysis on a DeltaPlus Advantage isotope ratio mass spectrometer coupled to a CN2500 elemental analyzer at the Laboratory of Chronology, Finnish Museum of Natural History (Luomus), University of Helsinki (Helama et al., 2015, 2016, 2018a, 2018b). The δ-notation as per mille (‰) expresses the deviations from the Vienna Pee Dee Belemnite (VPDB) standard. The mean reproducibility of this process, calculated by analysing sample replicates of our cellulose samples and those of internal laboratory reference (Fluka-22181 cellulose powder, Sigma-Aldrich, Lot. 442654/1), was ± 0.1 ‰ (Helama et al. 2015). These data were corrected for the 'Suess' effect i.e. changes in past $\delta^{13}C$ values as documented for the atmospheric CO_2 since AD 1000 (Leuenberger, 2007).

2.3. Analyses of $\delta^{13}C$ series

The individual isotopic series from 298 trees were analysed for their temporal trends and overall $\delta^{13}C$ levels. These analyses followed the general guidelines established for tree-ring based data (Fritts, 1976; Cook et al., 1990; Speer, 2010), especially the regional curve standardization (Briffa et al., 1992, 1996; Briffa and Melvin, 2011; Helama et al., 2017b), but with slight modifications to be compatible with our decadally sampled $\delta^{13}C$ data (Table 1).

2.3.1. Regional Curve Standardization

The $\delta^{13}C$ series were first aligned according to their biological i.e. ontogenetic ages, instead of the dated calendar years, and averaged. This alignment accounted for the number of the pith-offset i.e. the number of missing innermost rings, when the data near the pith was not available. Thereafter, the resulting mean curve was modelled using an exponential function:

$$A_t = a - bc^t \qquad (1)$$

where A_t is the value of the trend in ontogenetic year t (here, the number of decade), and a, b and c are parameters obtained from the fitting procedure and determine the level of the curve for the ontogenetically oldest measured $\delta^{13}C$ values, determine the level of the curve near the pith and control the trend curvature. The function was solved by the Levenberg–Marquardt algorithm (Moré, 1978). Aligning the $\delta^{13}C$ series first according to their ontogenetic ages, averaging the data from a wide range of calendar years, and then fitting the function to the mean curve, makes it possible to isolate any age-dependent variations in the tree-ring series (Huntington, 1912; Briffa et al., 1992, 1996). This age-dependent component of the $\delta^{13}C$ variability can be removed (i.e. detrended) from each of the $\delta^{13}C$ series by extracting the modelled $\delta^{13}C$ value from the observed $\delta^{13}C$ value. In dendrochronology, it is convention to detrend the series by dividing the observed tree-ring value by the modelled value to simultaneously eliminate the heteroscedasticity from the data (Cook and

Peters, 1997; Helama et al., 2004b). However, the $\delta^{13}C$ data of *P. sylvestris* is not heteroskedastic (McCarroll and Pawallek, 1998), meaning that the mean and the variance of $\delta^{13}C$ are not statistically related, and it is therefore recommended to remove the age-related trends from these series by subtraction (Cook and Peters, 1997). Here, we use a formula as follows:

$$D_t = I_t - A_t + \bar{I} \quad (2)$$

where D_t is the new, detrended isotopic value, I_t is the 'raw' (i.e. undetrended) isotopic value ($\delta^{13}C$) in ontogenetic year t (here, the decade), A_t is the ontogenetic trend (Eq. (1)), and $\bar{I}$ is a mean (time-invariant) of available isotopic values prior to AD 1700 (here, -24.9 ‰). Finally, the $\delta^{13}C$ series were realigned to their cross-dated, calendar years, and averaged into a mean chronology. This methodology constitutes the Regional Curve Standardization (RCS) (Briffa et al., 1992, 1996; Briffa and Melvin, 2011; Helama et al., 2017b) increasingly used in recent palaeoclimate and dendroclimatic literature for its benefits for retaining the low-frequency climate variability in the chronology (e.g. Briffa et al., 2013; Melvin et al., 2013; Matskovsky and Helama, 2014; Yang et al., 2014; Wilson et al., 2016; Anchukaitis et al., 2017; Helama et al., 2018c).

2.3.2. Signal-free approach

The RCS curves (Eq. (1)) may not represent purely age-related $\delta^{13}C$ change but contain fluctuations of climatic origin. Removing this component straightforwardly from the series may, at least in some cases, bias the mean chronology by altering its climate-dependent variations (Briffa and Melvin, 2011). Such a bias can be avoided by a signal-free implementation (Melvin and Briffa, 2008), applied together with the RCS method (Melvin et al. 2013). To do so, the undetrended $\delta^{13}C$ values were aligned by their calendar year positions and subtracted by the values of the RCS chronology, to derive a set of signal-free (SF) series. These SF series were again aligned according to their ontogenetic years and averaged to obtain a mean curve, which was then modelled using the non-linear function (Eq. (1)) now constituting an SF-RCS curve. Subtraction (Eq. (2)) was then used to remove the age-related trends from the undetrended $\delta^{13}C$ series and the resulting $\delta^{13}C$ series were averaged into a mean SF-RCS chronology. The process of signal removal was repeated (Melvin and Briffa, 2008) until no noticeable change in the chronology estimation could be attained to obtain final SF series.

In theory, the final SF series are more optimal than the raw series for assessments of the site differences because they are supposed to contain no climatic signals (Briffa et al., 2013; Melvin et al., 2013). As for the isotopic data, the SF series make it possible to address the relative isotopic enrichment (RIE) in respective tree-ring samples that may relate to site conditions or ecological status of the trees from a wide range of calendar years and climatic intervals. The overall $\delta^{13}C$ levels were obtained as relative measures (i.e. RIE), as previously suggested (Briffa and Melvin, 2011), by comparing the sum of $\delta^{13}C$ values for each SF-series to the sum of the overlapping SF-series' mean curve using the equation

$$RIE = \frac{1}{n-p}\sum_{t=p+1}^{t=n}(I_t - \bar{I}_t) \quad (3)$$

where I_t is the isotopic ($\delta^{13}C$) value in ontogenetic year t from first to last (n) ontogenetic year in that SF-series, $\bar{I}_t$ is the mean value of $\delta^{13}C$ values of all SF series in that year t, and p denotes the pith-offset. Thus, the series with RIE > 0 were found enriched in their ^{13}C relative to the $\delta^{13}C$ mean, whereas the series with RIE < 0 were those depleted in ^{13}C. Such grouping was previously used for defining multiple RCS curves from large tree-ring datasets (Briffa et al., 2013; Melvin et al., 2013; Yang et al., 2014; Matskovsky and Helama, 2014; Helama et al., 2018c) and could likewise be used for grouping the $\delta^{13}C$ series for a more elaborated RCS procedure.

2.3.3. Multiple curves

Multiple RCS curves (MRCS) were defined based on previously set guidelines to detrend tree-ring series (Briffa et al., 2013; Melvin et al., 2013; Yang et al., 2014; Helama et al., 2018c). First, two separate mean curves were calculated, for the $\delta^{13}C$ series of living trees and those of subfossil samples. The two types of $\delta^{13}C$ series were detrended by their respective models and the resulting $\delta^{13}C$ data averaged into a mean chronology. Second, two separate mean curves were calculated, for the enriched and depleted $\delta^{13}C$ series of the full dataset, their age-related trends were modelled (Eq. (1)) and removed (Eq. (2)) from the undetrended $\delta^{13}C$ series before calculation of the mean chronology. Third, two separate mean curves were calculated, for the $\delta^{13}C$ series of western and eastern origin, the respective models used for detrending the series, and the mean chronology calculated by averaging the resulting, detrended $\delta^{13}C$ series (for sample size fluctuations in the different types of $\delta^{13}C$ data, see Fig. S1).

2.3.4. Assessment of removed $\delta^{13}C$ variability

Considering that the RCS chronologies are computed as averages of detrended $\delta^{13}C$ series that, in turn, are derived as the difference between the values of the 'raw' isotopic value and that of the RCS model (Eq. 2), we also constructed the time-series of the unwanted (i.e. noise) $\delta^{13}C$ variability (U) by understanding these relationships as

$$U_T = \bar{I}_T - \bar{D}_T \tag{4}$$

where $\bar{I}_T$ and $\bar{D}_T$ express the $\delta^{13}C$ values in the mean time-series of 'raw' and detrended $\delta^{13}C$ series in cross-dated calendar year (T), respectively. The time-series of U_T were calculated separately for each type of the RCS chronologies.

2.3.5. Characterizing the records

Low-frequency comparisons were carried out between the $\delta^{13}C$ records smoothed using differently flexible spline functions (Cook and Peters, 1981), corresponding here to 500-year and 1500-year rigidities. Unsmoothed and smoothed chronologies were compared visually and using the Pearson correlations between each other. The mean chronologies and the time-series of removed $\delta^{13}C$ variability (U) (Eq. (4)) were also correlated to the mean records of RIE statuses of the $\delta^{13}C$ series, and to the time-series representing the potential sources of bias from non-climatic factors potentially affecting the 'raw' $\delta^{13}C$ data: the mean ontogenetic age of the $\delta^{13}C$ samples and the mean latitude, longitude and altitude of the sampling sites through time. To do so the time-variant mean altitude was calculated before and after the adjustment for land uplift. Conceivably, the time-series were expected to portray autocorrelation and, to account for its effect, the levels of statistical significance for the resulting Pearson correlations were assessed using the Monte Carlo simulations adopting the previously suggested method (Ebisuzaki, 1997) and published algorithms (Macias-Fauria et al., 2012). We also compared the long-term means (5500 BC to 3001 BC, 3000 BC to 501 BC and 500 BC to AD 2010) of the $\delta^{13}C$ levels using the unequal variance *t*-test (Ruxton, 2006), taking into account the autocorrelation in the data to obtain effective sample size (Franke et al., 2013) for calculations.

2.4. Meteorological data

Monthly temperature and cloud data within the study region were obtained from three stations in northern Norway providing observations on temperature and cloud cover since at least the early

20th century: Kautokeino (68.00°N, 23.03°E, 307 m above sea level), Sihccajavri (68.76°N, 23.54°E, 382 m) and Karasjok (69.47°N, 25.48°E, 155 m). These data were available from the Norwegian Meteorological Institute (http://eklima.met.no/) as the mean (TAM) and maximum temperatures (TAX), the days of clear sky (NN04), of fair weather (NN09), and of overcast (NN20), obtained as the number of days, and as the average cloud cover (NNM), minimum cloud cover (NNN), and maximum cloud cover (NNX) expressed as octas. The stations of Kautokeino, Sihccajavri and Karasjok provide temperature (could) data since 1889 (1931), 1913 (1913) and 1876 (1877), respectively. The TAX records were not available until AD 1951. The time-series of the three summer months (June, July and August) were normalised to a mean of zero and standard deviation of one calculated over the years with data from all tree stations. Next, the mean series of monthly observations were calculated over the full period by averaging the normalised time-series. The resulting composite records cover the years from AD 1891 onwards and were averaged over decades (AD 1891-1900, 1901-1901 … 2001-2010) to correspond to the resolution in our $\delta^{13}C$ data. The normalised series were then rescaled to represent the mean and standard deviation in each of the climate records (TAM, TAX, NN04, NN09, NN20, NNM, NNN, and NNX) as observed in Karasjok (the station with highest number of observations). Pearson correlations were calculated between the climate records and the $\delta^{13}C$ chronology.

2.5. Calibration and verification

A subsample replication experiment was carried out to test the skill of our $\delta^{13}C$ data to reconstruct past climate variability with reduced number of $\delta^{13}C$ samples. The full number of $\delta^{13}C$ series (n=22) over the instrumental period was randomly reduced to replication of 5 to 11 isotope series (to mimic the replication of the subfossil part of the chronology), their mean $\delta^{13}C$ chronologies computed and the calibration and verification statistics calculated for the resulting experimental chronologies, this procedure being repeated one hundred times. Similar dendroclimatic approach (i.e. bootstrapping) was recently adopted by Loader et al. (2013b) in defining the uncertainties in their *P. sylvestris* $\delta^{13}C$ and $\delta^{18}O$ datasets and further applied for analysing the calibration/verification statistic of the tree-ring width proxy dataset (Helama et al., 2017a).

First, the climate parameter with strongest association with the $\delta^{13}C$ data was used for evaluating the possible discrimination rate change to elevated CO_2 in the atmosphere. Using the previously detailed methods (Treydte et al., 2009), the Pearson correlation, the Durbin-Watson statistic and the linear slope of the regression residuals were used to assess the possible correction value between −0.05‰ and 0.10‰ per ppmv CO_2 increase, with 0.001‰ increments, to our tree-ring $\delta^{13}C$ chronology values. The history of atmospheric CO_2 concentration was obtained as a smoothed (40-yr spline) time-series as derived from Law Dome ice core (Etheridge et al., 1996, 1998; MacFarling Meure et al., 2006) and instrumental data registered at the Mauna Loa observatory (Keeling et al. 1976).

Second, the same climate parameter was used for building the palaeoclimate transfer functions. The coefficient of determination (R^2) was calculated over the instrumental period (AD 1891-2010) to express the fraction of climate variance the $\delta^{13}C$ data explained. Leave-one-out procedure (Gordon et al., 1982) was used to produce independent predictand record as long as the instrumental period. To do so, the observations (AD 1891-1900, 1901-1901 … 2001-2010) in the calibration period were withheld one at the time and the calibrations were made on the remaining values, each time predicting the withheld value. The resulting predictand record was used to verify the skill of the statistical models transforming the $\delta^{13}C$ proxy data into palaeoclimate estimates. The squared Pearson correlation coefficient (r^2) and the Reduction of error (RE) were used as a verification statistics. Positive values of the RE are commonly regarded as indication of real skill in the reconstruction (Fritts, 1976).

3. RESULTS

3.1. Tree-ring $\delta^{13}C$ variations

3.1.1. Initial data constituting the $\delta^{13}C$ chronology

The averaged, undetrended $\delta^{13}C$ values (referred hereafter as the 'raw' $\delta^{13}C$ chronology) vary roughly between -26 and -24 ‰ through mid and late Holocene time (Fig. 2a). These variations are evident at decadal to millennial scales and thus represent vigorous low-frequency $\delta^{13}C$ fluctuations superposed by isotopic excursions of shorter duration. Moreover, the mean ontogenetic age of the $\delta^{13}C$ samples exhibit temporal variations demonstrating that the chronology is constructed from isotopic samples of variable tree age in different periods (Fig. 2b). This record shows a mean of about 100 years (101.0 yrs) and standard deviation of three decades (29.4 yrs), with maximum and minimum values of 222 and 32.8 years. Interestingly, there appears a slightly positive trend in this record through the study period indicating that the longevity of the trees' has increased towards the present-day. Alternatively, the more recent subfossils may have decayed less on their external surface thus allowing a higher number of rings to have preserved in the samples.

Moreover, there seems to be a long-term change in the mean latitude of the sampling sites (Fig. 2c). Likely, this change reflects the latitudinal retreat of the northern timberline over the mid and late Holocene as commonly observed in the region and adjacent areas following the orbital forcing on summer temperatures (Eronen, 1979; Kullman, 1992; Helama et al., 2004a, 2010). This retreat is superposed by shorter-term fluctuations that may largely be more stochastic as they do not correspond with variations in the mean longitude (Fig. 2d) or altitude (Fig. 2e) of the sites which, by contrast, appear closely linked to each other. That the mean record of the longitudes correlates negatively with the corresponding record of altitudes ($r = -0.933$) mirrors the fact that the eastern sites occur at topographically lower elevations (see Table S1). The long-term descent in the site altitudes, similar to that of their latitudes, may likely be reflected in the apparent tendency of samples from eastern sites to cover the later part of the study period.

The chronology is covered by subfossil $\delta^{13}C$ series from at least five trees since 5500 BC, and by an increasing proportion of living tree $\delta^{13}C$ series from AD 1541 through 2010 (Fig. 2f). The proportion of living tree $\delta^{13}C$ series remains above 50 % since AD 1701.

3.1.2. Age-related $\delta^{13}C$ trends

The $\delta^{13}C$ series contain trends dependent upon the ontogenetic age of the sampled trees (Fig. 3a). These trends account for roughly 1.0 ‰ increase in mean $\delta^{13}C$ values over the first 35 decades of trees' lifespan and can be described using the nonlinear models (Table 2). More specifically, these models demonstrate the change towards less negative $\delta^{13}C$ values by roughly 0.2 ‰, 0.4 ‰and 0.7 ‰ over the first five, ten and twenty decades of trees' lifespan i.e. clearly beyond the "juvenile effect". These trends are evident for both the $\delta^{13}C$ series and their SF-series (i.e. signal-free), indicating that the mean trends are not influenced by almost any climatic influence. Likewise, the similarity between the RCS and SF-RCS models may result from the wide range of calendar years the samples represent as the individual variations are then suppressed in the mean curves by the law of averages (Huntington, 1912). Overall, these results show the expected benefits (Briffa et al., 1996; Briffa and Melvin, 2011) of obtaining unbiased RCS models from similarly distributed subfossil sample settings.

Yet, the age-related $\delta^{13}C$ trends are more convex in living trees and their RCS and SF-RCS models are less similar to each other (Fig. 3b). Obviously, these differences are caused by the relatively small number of living trees representing the $\delta^{13}C$ variations over the past centuries only. As a result, their mean $\delta^{13}C$ values may also exhibit short-term alterations unrelated to the ageing trend, an issue that is reflected in the relatively low R^2 values obtained for the living tree RCS models (Table 2). Similarly differentiated RCS and SF-RCS models are not found for the subfossil data (Fig. 3b) that again represents much wider range of calendar years and climatic signals of which anomalies become then cancelled out already when calculating the age-dependent mean curves. Moreover, the mean RIE values for living trees and subfossils average 0.234 ‰ and 0.006 ‰, respectively, this difference (0.228 ‰) being statistically significant (*t*-test, $p < 0.05$).

Dividing the $\delta^{13}C$ series dichotomously (RIE > 0 versus RIE < 0) results in mean curves with widely separated $\delta^{13}C$ levels throughout the trees' lifespan. In addition, the RCS and SF-RCS models for the $\delta^{13}C$ series with RIE > 0 are somewhat more convex in their shape than those with RIE < 0 (Fig. 3c). The $\delta^{13}C$ series from the western sites are represented by especially SF-RCS models with less negative $\delta^{13}C$ estimates, throughout the trees' lifespan, than those from the eastern sites (Fig. 3d). Similarly, this finding indicates that the $\delta^{13}C$ data from western sites may be characterised by higher RIE than the eastern $\delta^{13}C$ data. In fact, the western and eastern $\delta^{13}C$ series exhibit RIE values averaging 0.106 ‰ and -0.034 ‰, respectively, the difference (0.140 ‰) between these means being statistically significant (*t*-test, $p < 0.05$).

Considering that the chronology is composed of $\delta^{13}C$ samples of varying ontogenetic age (Fig. 2b), the age-related trends in $\delta^{13}C$ series (Fig. 3) indicate corresponding biases potentially introduced into the 'raw' $\delta^{13}C$ chronology (Fig. 2a) and illustrate the obvious need of removing the age-related trends from the individual $\delta^{13}C$ series prior to averaging the mean chronology. Such analyses using the RCS methods are demonstrated in the following section.

3.1.3. Constructing the $\delta^{13}C$ chronologies

Removing the modelled $\delta^{13}C$ trends (Fig. 3) from the series and characterizing the $\delta^{13}C$ variations as respective $\delta^{13}C$ chronologies demonstrates isotope excursions predominantly between 1.0 ‰ and -1.0 ‰ from the long-term $\delta^{13}C$ mean level (Fig. 4). In fact, the chronologies retain some degree of similarity yet there are also noticeable differences resulting from the means the alternative RCS methods have removed the non-climatic noise from the undetrended $\delta^{13}C$ series.

The low-frequency patterns of $\delta^{13}C$ change exhibit variations exceeding 0.2 ‰ in magnitude, with statistically significant difference between the mid and late Holocene $\delta^{13}C$ levels in most of the chronologies (Table 3). The exceptions were those chronologies constructed from the $\delta^{13}C$ series detrended by RIE-dependent models (Fig. 3c). More specifically, these chronologies appear to illustrate markedly less long-term variations in contrast to other $\delta^{13}C$ chronologies already in visual inspection (Fig. 4c) and their the low-frequency patterns of $\delta^{13}C$ change exhibit variations around 0.05 ‰ only (Table 3).

Similar patterns are found in the extreme $\delta^{13}C$ values for which the RIE-dependent RCS models resulted in $\delta^{13}C$ change (i.e. magnitude) notably less than 2.0 ‰, whereas the other chronologies portrayed magnitudes clearly above that value (Table 4). These differences are reflected in the correlations among the chronologies (Fig. 5a; for correlation matrices, see Table S2). That is, the RIE-dependent models appear to result in $\delta^{13}C$ chronologies with lowest correlations to other chronologies. These results are the same for the full period (BC 5500 to AD 2010) and for the period covered by the living tree $\delta^{13}C$ data (AD 1541 to 2010).

Notably low correlations are obtained also for the 'raw' $\delta^{13}C$ chronology, over the late period (Fig. 5). We note that this is the period over which the mean ontogenetic age of the $\delta^{13}C$ data appears unprecedentedly high (Fig. 2b). As a result, this interval exemplifies the magnitude of the bias from not removing the age-related trends in the $\delta^{13}C$ data.

3.1.4. Correlations with the RIE records

The $\delta^{13}C$ chronologies demonstrate substantial similarity to the temporal variations in the RIE data illustrated through the study period (Fig. 6). In addition to visual inspection, the relatedness of these records is confirmed by the positive correlations both with the unsmoothed and smoothed records (Fig. 5b). The $\delta^{13}C$ chronologies built using the RIE-dependent RCS models (Fig. 3c) represented the only exceptions in this regard with correlations as low as r ~ 0.1 … 0.2 against the actual RIE records over the mid and late Holocene times (Fig. 5b). Clearly, the trees' RIE statuses are not randomly distributed over time but we find $\delta^{13}C$ series with RIE > 0 more frequently roughly before 1000 BC and less so over the AD era (see Fig. 6a, b). Thus, the RIE statuses constitute low-frequency patterns and removing the age-related trends according to these statuses also removes the corresponding low-frequency variability from the resulting MRCS-RIE and SF-MRCS-RIE chronologies (see Table 3). In other words, the RIE statuses appear to contain the information vital to low-frequency climate signal in the $\delta^{13}C$ data.

Yet, the RIE data exhibits correlations to sites' geographical positions. The site-dependent RIE means associate positively with site altitudes, negatively with their latitudes and longitudes (Fig. S2). Importantly, the signs of the relationships are consistent with the evidence presenting increases in plant $\delta^{13}C$ i.e. decreased carbon isotope discrimination with altitude (Hultine and Marshall, 2000; Warren et al., 2001) and $\delta^{13}C$ decrease with latitude (Stuiver and Braziunas, 1987; van Klinken et al., 1994). The signs of these relationships are compatible with geographic patterns of smaller scale identified recently through the study region in the analyses of living tree $\delta^{13}C$ data (Helama et al., 2018a). For our $\delta^{13}C$ data, however, only those with site longitudes were statistically significant ($p < 0.05$). As previously alluded to, the western sites locate at topographically higher elevations (see Table S1). Moreover, the eastern sites may be cloudier and receive less solar radiation in summer due to the cold air flow from the Barents Sea (Venäläinen and Heikinheomo, 1997). That the $\delta^{13}C$ data originate from more northern, western and higher sites during the first half of the study period (Fig. 2) may accordingly suggest a positive non-climatic bias (i.e. less negative $\delta^{13}C$ values) in the resulting $\delta^{13}C$ chronologies over the same period, and similarly more negative $\delta^{13}C$ values in the chronology over the latter half of the period. Processing of these variations through the RCS analyses are addressed in the following section.

3.1.5. Quantifying the biases

As expected, the 'raw' $\delta^{13}C$ chronology contains a positive non-climatic influence (i.e. noise) with the mean age while no such relationship can be found for any of the RCS chronologies (Table 5a). Albeit the correlation is not remarkably high ($r = 0.231$) it is statistically significant ($p < 0.01$) thus illustrating the less negative $\delta^{13}C$ values in the resulting chronology over the periods of increasing ontogenetic age of the samples (for the 'raw' $\delta^{13}C$ chronology compared with the RCS chronologies, see Fig. S3). Moreover, the component of $\delta^{13}C$ variability being removed during the RCS process (Eq. (4)) appears highly correlative to the mean age of the samples (Table 5b) implying the strength of the detrending approaches applied here to remove the age-related bias from the $\delta^{13}C$ series. Not all methods but only those with separate RCS models for western and eastern sites are found to correct all the considered biogeographical biases in the $\delta^{13}C$ data. This is inferred from the statistically significant and negative correlations between the removed $\delta^{13}C$ component and the site latitudes and longitudes over the full period, as well as from the positive correlations between the same $\delta^{13}C$

component and the site altitudes (see correlations for SF-MRCS W/E in Table 5b). In addition, the $\delta^{13}C$ chronologies are found to show less negative $\delta^{13}C$ values over the periods with more northern samples available into the chronology, but no such correlations are statistically significant ($p < 0.05$) unless the separate RCS models for western and eastern sites were in use (Table 5a). Less negative $\delta^{13}C$ values are indeed generally expected during the periods of warmer climate (Helama et al., 2016), with similar increase in subfossil sample availability from our northernmost sites over the mid and late Holocene times (Helama et al., 2004a, 2010). Therefore, the chronology of this type (i.e. SF-MRCS W/E) appears the only form of our $\delta^{13}C$ data reliably preserving this potential palaeoclimate signal. As a result, the correspondingly detrended $\delta^{13}C$ series were used in the following sections to detect the climatic signals in the $\delta^{13}C$ data.

3.2. Climatic signals in $\delta^{13}C$ data

3.2.1. Correlations to monthly climate variables

The $\delta^{13}C$ data had highest (negative) correlations to monthly variables of average cloud cover (Fig. 7). Correlations to the number of days of clear sky and of fair weather were nearly as high (positive) but more variable in their strengths from month to month. Statistically significant associations are found also for the minimum cloud cover. Among the temperature variables, only the mean maximum temperature is somewhat related to $\delta^{13}C$ data but the correlations are not statistically significant. Clearly, the correlations to average cloud cover are markedly negative for all three summer months. These correlations are statistically significant for June and July and for the June through August (JJA) season. Using the average cloud cover (JJA) for the estimation of trees' response to CO_2 increase did show strongest climatic correlation with a slightly negative discrimination rate change whereas the Durbin-Watson statistics and the slope in regression residuals provided indications of a slightly positive discrimination rate change (Fig. S4). As a result, no correction to chronology was made. The same climatic parameter was used as dependent variable to be reconstructed from the $\delta^{13}C$ data.

3.2.2. Strength of $\delta^{13}C$ data to reconstruct cloud cover

Randomly reducing the chronology replication to $\delta^{13}C$ sample sizes (n = 5 … 11) similar to that over the subfossil period (Fig. 2f) and calibrating the resulting $\delta^{13}C$ chronologies against the average cloud cover record produced R^2 roughly between 0.4 and 0.9 (Fig. 8a). It appeared that the correlation between the proxy and climate cannot be elevated by simply increasing the sample size while the chance of low R^2 do increase with lower replication. That is, the R^2 values similar to that observed with n=10 could be obtained with n=5. Using the leave-one-out verification resulted in a similar pattern for r^2 (Fig. 8b) and RE (Fig. 8c). Here, R^2, r^2 and RE average 0.751, 0.648 and 0.712, respectively. Importantly, the RE did not reach any negative values, the majority of the RE values remaining above 0.5. The obtained statistics did not show any dependence on tree age. That is, the correlation between the R^2, r^2 and RE values and the mean age of the bootstrapped chronologies, for which the corresponding statistics were calculated, was as low as 0.054, 0.032 and 0.041, respectively, these coefficients remaining statistically non-significant. Thus, the varying ontogenetic age of the $\delta^{13}C$ samples (Fig. 2b) was not expected to influence the reliability of the reconstruction. We also note that the calibration and verification experiments with other promising climate records, those of the number of days of clear sky and of fair weather as well as the minimum cloud cover, did in fact show lower calibration/verification statistics (Fig. S5).

3.2.3. Final reconstruction model

A final statistical model was built over the instrumental period and the leave-one-out implementation used to verify the calibration. The transfer function to reconstruct the average cloud cover (C_T) in calendar year (T) was derived as follows

$$C_T = -0.662\overline{D}_T - 10.940 \qquad (5)$$

where $\overline{D}_T$ expresses the values in the mean δ^{13}C chronology (i.e. SF-MRCS W/E). Here, R^2, r^2 and RE were 0.840, 0.785 and 0.818, respectively, these values slightly exceeding the respective mean levels derived from bootstrapped δ^{13}C chronologies (see Fig. 8). Yet, even higher values could be obtained for some of the bootstrapped estimates implying that the relationship between the reconstruction skill and the chronology replication is not straightforward and that the skill may not be invariably reduced over the subfossil period.

3.2.4. Validation using proxy evidence

The full reconstruction portrays markedly variable cloudiness over the mid and late Holocene times with decadal to millennial changes in the average cloud cover (Fig. 9). We note that many of the characteristic features of the new reconstruction are at least indirectly supported by the previous palaeoclimate literature of independent proxy evidence. Long-term variations illustrate periods with reduced cloudiness since the start of the records until around 5 ka. That is, the first two millennia of our record were found to represent relatively clear skies with gradually increasing cloudiness with a culmination of this phase between 4.8 ka and 4.0 ka. Low lake-levels and correspondingly dry climate have been previously inferred from the lake sediments at various sites from central to northern Fennoscandia between 7.0 and 5.0 ka (Solovieva and Jones, 2002; Antonsson et al., 2008; Nyman et al., 2008), which should concur with our indications of reduced cloud cover over the same period and generally moister climate thereafter. Moreover, there appears a long-term decline in cloud cover with lowest reconstructed values identified especially between 3.1 ka and 2.8 ka, after which the reconstruction indicates increase in cloud cover, with highest values over the past two millennia, around 1.7 ka. The timing of this swing matches that of the climate change about 2.8 ka and the increased humidity in the North Atlantic and European regions since then (Martin-Puertas et al., 2012). It has been suggested that this change may in fact represent the biostratigraphic transition from the sub-boreal to the sub-Atlantic period (van Geel et al., 1996, 2014), similarly inferred also from the changes in the lacustrine sedimentation in Fennoscandia (Berntsson et al., 2015).

These periods were superimposed by the variations in cloud cover of shorter timescales. Two such periods of decreased cloudiness were evident between 5.2 and 4.8 ka and between 0.3 ka and 0.1 ka. Possibly, the mid-Holocene event may be linked with the climate phase observed in central Europe, with indications of hydroclimate anomalies around the globe, between 5.6 and 5.0 ka (Magny et al., 2006). The authors of that article further suggested this event to represent transition in hemispheric climate regimes similar to that observed around 2.8 ka (van Geel et al., 1996, 2014). The late Holocene event, on the other hand, overlaps with the timing of the Little Ice Age when the Northern Hemisphere temperatures where significantly below the modern (AD 1961-1991) mean level (Matthews and Briffa, 2005). Combined with our results, these findings reinforce the view of the same Little Ice Age interval as a period with cool, sunny climatic phase in the same region (Young et al., 2010, 2012; Loader et al., 2013a), and suggest potential links with our reconstructed cloud cover (Fig. 9) and large-scale, even hemispheric climate transitions. Importantly, the most recent change in the reconstruction towards the increased cloud cover agreed with the Arctic moistening observed as high-latitude precipitation increases (Zhang et al., 2007; Min et al., 2008) and thus with parallel trends in cloud cover. At local scale, the recent trend in cloudiness is known to be associated with the increase of cloudy situations, rather than amounts of clouds when they are present

(Parding et al., 2014). Based on our reconstruction, this frequency has shifted markedly since the Little Ice Age but remains far from the more cloudy conditions recorded in the context of the past 7.5 ka.

4. DISCUSSION AND CONCLUSIONS

Production of tree-ring isotope datasets is underway worldwide. Part of this development is the use of subfossil samples to construct isotope chronologies over the past millennium (Gagen et al., 2011, 2012; Young et al., 2012; Loader et al., 2013a; Naulier et al., 2015; Gennaretti et al., 2017). Yet, the extent of long, continuous subfossil tree-ring chronologies exceeds this time frame (Pilcher et al., 1984; Eronen et al., 1999, 2002; Grudd et al., 2002; Hantemirov and Shiyatov, 2002; Kelly et al., 2002; Helama et al., 2004, 2010; Nicolussi et al., 2010), these materials offering the still largely unexplored potential for high-resolution isotope records extending much over the Holocene and even the late Pleistocene (Becker et al., 1991; Mayr et al., 2003). Making use of a regionally collected sample set, our $\delta^{13}C$ chronology represents a typical subfossil assemblage, the multi-site approach with no pooling of rings from different trees demonstrating the uncertainties attributable to age-related trends as well as to site longitudes/altitudes. These factors stand for non-climatic sources of inter-tree and -site variability. Their effects were much reduced when the RCS methods similar to tree-ring width and density proxies, supplied by the signal-free approach, were used for assessing the age-related $\delta^{13}C$ trends separately for the western and eastern sites (Briffa et al., 1992, 1996; Melvin and Briffa, 2008; Briffa and Melvin, 2011; Helama et al., 2017b). Another type of methodology to produce long isotopic chronologies from tree rings was recently suggested as "offset-pool plus join-point" approach (Gagen et al., 2012; Naulier et al., 2015) in which the blocked data of consecutive cohorts from a single site were adjusted to equal isotopic level over the intervals for which the cohorts overlap each other near their ends and thus to calculate a continuous record of such cohorts as an isotope chronology. In comparison, the RCS method we applied did not require data in any form of cohorts and, as a consequence, remained insensitive to potential uncertainties inherent to isotopic values over the intervals critically used to adjust the consecutive cohorts. The use of multiple RCS models was also shown to reduce the uncertainty arising from biogeographical aspects i.e. the site longitudes/altitudes.

While these results supported the use of detrending methods conventional to dendroclimatic literature, we also adopted a new equation (Eq. 1) to model the age-related $\delta^{13}C$ trends. In so doing, there was no need to multiply the 'raw' $\delta^{13}C$ data by minus one (Esper et al., 2010; Helama et al., 2015) to be able to fit a negative exponential model, used frequently in tree-ring literature over the past 50 years (Fritts et al., 1969), to isotopic data. Moreover, it was shown that not all the RCS chronologies preserved the $\delta^{13}C$ variations at lowest frequencies. Accounting for the $\delta^{13}C$ series' overall isotopic level (i.e. the RIE) during the RCS produce reduced the amplitudes of the long-term variations in the chronology indicating that the low-frequency signal was actually originating from the temporal distribution of trees with positive and negative isotopic offsets from the grand mean $\delta^{13}C$ value (here -24.9 ‰). These findings concur with those presented previously for tree-ring density proxies from northern Sweden and Finland. It was shown that the number of tree-ring series with overall high latewood density increased during the periods of climatic warmth (i.e. the Medieval Warm Period and the modern warming) whereas the number of series with overall low latewood density increased during the Little Ice Age when climate was predominantly cooler (Matskovsky and Helama, 2016). We anticipate that more research is still needed to detail the potential loss of the low-frequency variability both in the more traditional tree-ring width and density and in the isotope chronologies over similar, multi-millennial scales. Notwithstanding, most of the RCS procedures applied here did preserve the variability at the lowest frequencies, in accordance with the dendrochronological theories (Cook et al., 1995). Yet, the RIE data correlated with site altitudes demonstrating the potential non-climatic bias in the 'raw' $\delta^{13}C$ variability resulting from the well-known altitude-dependent carbon isotope discrimination (Leavitt and Long, 1992; Hultine and Marshall, 2000; Warren et al., 2001; Helama et al., 2018a). Here, the procedure chosen for the reconstruction (i.e. SF-MRCS W/E) did not only remove the biases from tree age but from temporally varying site characteristics. Such results also suggested a removal of biases that may arise from the long-term natural changes in the landscape known to have occurred in the region since the mid Holocene (Eronen, 1979; Kullman, 1992; Helama et al., 2004a, 2010), as reflected in our metadata (Fig. 2) and more generally inferred from circumpolar palaeoecological records (MacDonald et al., 2000; Payette et al., 2002; Binney et al., 2018).

Interestingly, the living trees portrayed an age-related trend in $\delta^{13}C$ data over the fifty biologically youngest rings only, the shape of this trend thus resembling those obtained predominantly for living trees of the same species and region (Gagen et al., 2007, 2008; Young et al., 2011; Loader et al., 2013a), yet differing markedly from the shape of trend defined for the subfossil data. Clearly, the shape of the age-related trend in $\delta^{13}C$ data may change considerably over long periods of time (Helama et al., 2015). Thus, making the assumption that simply leaving out the first fifty rings from each sample may be a procedure sufficient to produce isotopic chronology with no age effects appears valid at most over the recent times and ought not generally be applied at least for data from preceding times with no further information on age-related isotopic variations. In fact, a similar approach of obtaining the mean chronology from tree-ring width proxy data, with any detrending, only removing the fast-growing juvenile phase in each ring-width series has sometimes been used in the old tree-ring literature (e.g. LaMarche, 1974), but its use has not been recommended for decades (Cook et al., 1995). This approach was not applied for our data. Instead, an undetrended 'raw' $\delta^{13}C$ chronology was experimentally calculated and compared with other available records. This record correlated positively with the mean ontogenetic age of the samples, demonstrating the presence of age-related bias in the chronology. The danger of neglecting the age-related biases in the 'raw' $\delta^{13}C$ chronology was notably exemplified over the past centuries when the 'raw' and detrended $\delta^{13}C$ chronologies exhibited a drop in their correlations. Moreover, the component of $\delta^{13}C$ variability removed during the RCS procedure was highly explained by the mean ontogenetic age over the study period. We thus concur with other investigations (Esper et al., 2010, 2015) that have shown the importance for rigorously identifying the potential age-related biases (e.g. noise) in tree-ring stable isotope data and illustrated the essence of detrending for obtaining more reliable isotopic chronologies for palaeoclimate reconstructions.

The $\delta^{13}C$ chronology was used as a proxy for the decadally resolved palaeo-cloud reconstruction validated against the instrumental data and over the pre-instrumental period against the independent proxy evidence, suggesting local to large-scale, possibly even hemispheric connections during the centennial cloud cover minima and coinciding events of widely recognised climate transitions (van Geel et al., 1996, 2014; Matthews and Briffa, 2005; Magny et al., 2006) at around 5.2-4.8 ka, 3.1-2.8 ka and 0.3-0.1 ka. An additional indication of large-scale climate associations was found over the recent decades with positive could cover trend overlapping the era of the Arctic moistening (Zhang et al., 2007; Min et al., 2008). Modelling studies indicate that the increasing Arctic cloudiness may result from the meridional moisture transport from lower latitudes especially during the boreal warm season (Bintanja and Selten, 2014; Vavrus et al., 2011). In the future, the trends in hydroclimate are predicted to depend on frequencies of the governing types of large-scale atmospheric circulation (Santos et al., 2016) that in turn are known to drive the trends in the cloudiness in the study region (Parding et al., 2014). The $\delta^{13}C$ proxy data may effectively track these recent variations in the cloud cover and place them in the context of the past variations.

In comparison to our high-latitude $\delta^{13}C$ data, the tree-ring width and density proxies have been frequently used for reconstructing the past summer temperature variability in northern Sweden and Finland (Briffa et al., 1988, 1992; Melvin et al., 2013; Matskovsky and Helama, 2014, 2016). A palaeoclimatic approach where these conventional tree-ring proxy records are analysed alongside the isotopic data was previously used to demonstrate cool but sunny conditions during the Little Ice Age, and warmer, cloudier conditions during the Medieval Warm Period, as reconstructed for the same region (Young et al., 2010, 2012; Loader et al., 2013a). Such analyses show the value of multiple proxy data to considerably add the information available from any single proxy record and demonstrate the ways the reconstructed climate patterns may be related to large-scale pressure conditions over the North Atlantic sector and Europe. As previously alluded to, cloud cover minima reconstructed from our $\delta^{13}C$ chronology between 0.3 and 0.1 ka concurs with the previously demonstrated Little Ice Age climate patterns. Similar approach was previously used to infer changes in atmospheric circulation from a combination of sedimentology and microfossil based precipitation and temperature reconstructions in central Sweden over the past 10 ka (Antonsson et al., 2008). Compared to such records, however, tree-ring proxies are annually resolved and dated to exact calendar years. Clearly, more detailed analyses using multiple dendroclimatic proxy records, those based on tree-ring width, density and isotopic data, are needed to develop similar comparisons further back in time. While beyond the scope of this paper, we emphasize the potential of such comparisons to systematically analyse the North Atlantic forcing of late and mid Holocene climates.

These connections indicate the importance of tree-ring isotopic proxies in characterizing and dating, with reduced uncertainties, the critical intervals during which the large-scale processes potentially involving ocean-atmosphere dynamics have likely controlled the climate system. Extending the existing living tree and ~ 1 ka long isotope chronologies necessitates that subfossil wood is more frequently and profoundly exploited to produce an increased number of multi-millennial tree-ring isotopic chronologies for several regions where such wooden archives are available. Our results underscore the need for a body of such chronologies and their reconstructions to outline the structure of general instability of past climate and to detail the anomalies punctuating the long-term trends in that variability in the recent geological past and possibly to assess the modelling studies of ongoing and future climates.

Acknowledgements

Fieldwork carried out during the summer of 2012 was greatly aided by Hannu Herva, Jari Hietanen, Esko Jaskari, Tauno Luosujärvi, and Raimo Pikkupeura. We also thank the help in the joiner’s workshop by Pasi Aatsinki, Tarmo Aalto, Pekka Närhi, Aarno Niva and Eero Siivola, in the tree-ring laboratory by Juha Kemppainen and in the isotope laboratory by Hanna Turunen, Aleksi Aalto, and Igor Shevchuk. Two reviewers are thanked for their constructive comments. This work was supported by the Academy of Finland (grants 251287, 251441, 288083, 288267 and 292788).

Data availability

Upon acceptance, the data of time-series representing the palaeoclimate reconstruction i.e. based on the $\delta^{13}C$ data, with the uncertainties estimated, will be placed in an open repository at the National Centers for Environmental Information – National Oceanic and Atmospheric Administration (https://www.ncdc.noaa.gov/data-access/paleoclimatology-data).

References

Anchukaitis K. J., Wilson R., Briffa K.R., Büntgen U., Cook E.R., D'Arrigo R., Davi N., Esper J., Frank D., Gunnarson B.E., Hegerl G., Helama S., Klesse S., Krusic P.J., Linderholm H.W., Myglan V., Osborn T.J., Zhang P., Rydval M., Schneider L., Schurer A., Wiles G., Zorita E., 2017 Last millennium Northern Hemisphere summer temperatures from tree rings: Part II, spatially resolved reconstructions. Quaternary Sci. Rev. 163, 1-22.

Antonsson K., Chen, D., Seppä, H., 2008 Anticyclonic atmospheric circulation as an analogue for the warm and dry mid-Holocene summer climate in central Scandinavia. Clim. Past 4, 215-224.

Becker, B., Kromer, B., Trimborn, P., 1991. A stable-isotope tree-ring timescale of the Late Glacial/Holocene boundary. Nature 353, 647-649.

Berntsson, A., Jansson, K.N., Kylander, M E., De Vleeschouwer, F., Bertrand, S., 2015. Late Holocene high precipitation events recorded in lake sediments and catchment geomorphology, Lake Vuoksjávrátje, NW Sweden. Boreas 44, 676–692

Binney, H., Edwards, M., Macias-Fauria, M., Lozhkin, A., Anderson, P., Kaplan, J.O., Andreev, A., Bezrukova, E., Blyakharchuk, T., Jankovska, V., Khazina, I., Krivonogov, S., Kremenetski, K., Nield, J., Novenko, E., Ryabogina, N., Solovieva, N., Willis, K., Zernitskaya, V., 2017. Vegetation of Eurasia from the last glacial maximum to present: Key biogeographic patterns. Quaternary Sci. Rev. 157, 80-97.

Bintanja, R., Selten, F.M., 2014. Future increases in Arctic precipitation linked to local evaporation and sea-ice retreat. Nature 509, 479–482

Boettger, T., Hiller, A., Kremenetski, K., 2003. Mid-Holocene warming in the northwest Kola Peninsula, Russia: northern pine-limit movement and stable isotope evidence. Holocene 13, 403–410.

Briffa, K.R., Melvin, T.M., 2011. A closer look at regional curve standardization of treering records: justification of the need, a warning of some pitfalls, and suggested improvements in its application. In: Hughes, M.K., Swetnam, T.W., Diaz, H.F. (Eds.), Dendroclimatology, Progress and Prospects. Springer, Dordrecht, pp. 113–145.

Briffa, K.R., Jones, P.D., Bartholin, T.S., Eckstein, D., Schweingruber, F.H., Karlén, W., Zetterberg, P., Eronen, M., 1992 Fennoscandian summers from AD 500: temperature changes on short and long timescales. Clim. Dyn. 7, 111–119.

Briffa, K.R., Jones, P.D., Pilcher, J.R., Hughes, M.K. 1998. Reconstructing summer temperatures in northern Fennoscandia back to AD 1700 using tree-ring data from Scots pine. Arctic Alpine Res. 20, 385-394.

Briffa, K.R., Jones, P.D., Schweingruber, F.H., Karlén, W., Shiyatov, S.G., 1996. Tree ring variables as proxy-climate indicators: problems with low-frequency signals. In: Jones, P.D., Bradley, R.S., Jouzel, J. (Eds.), Climate Variations and Forcing Mechanisms of the Last 2000 Years. Springer, Berlin, pp. 9–41.

Briffa, K.R., Melvin, T.M., Osborn, T.J., Hantemirov, R.M., Kirdyanov, A.V., Mazepa, V.S., Shiyatov, S. G., Esper, J., 2013. Reassessing the evidence for tree-growth and inferred temperature change during the common era in Yamalia, northwest Siberia. Quaternary Sci. Rev. 72, 83–107.

Cook, E.R., Peters, K., 1981. The smoothing spline: a new approach to standardizing forest interior tree-ring width series for dendroclimatic studies. Tree Ring Bull. 41, 45–53.

Cook, E.R., Peters, K., 1997. Calculating unbiased tree-ring indices for the study of climatic and environmental change. Holocene 7, 359–368.

Cook, E.R., Briffa, K.R., Meko, D.M., Graybill, A., Funkhouser, G., 1995. The 'segment length curse' in long tree-ring chronology development for palaeoclimatic studies. Holocene 5, 229-237.

Cook, E., Briffa, K., Shiyatov, S., Mazepa, V., 1990. Tree-ring standardization and growth trend estimation. In: Cook, E.R., Kairiukstis, L.A. (Eds.), Methods of Dendrochronology: Applications in the Environmental Sciences. Kluwer Academic Publishers, Dordrecht, pp. 104–123.

Dorado Liñán, I., Gutiérrez, E., Helle, G., Heinrich, I., Andreu-Hayles, L., Planells, O., Leuenberger, M., Bürger, C., Schleser, G., 2011. Pooled versus separate measurements of tree-ring stable isotopes. Sci. Total Environm. 409, 2244-2251.

Duquesnay, A., Bréda, N., Stievenard, M., Dupouey, J.L., 1998. Changes of tree-ring $\delta^{13}C$ and water use efficiency of beech (*Fagus sylvatica* L.) in north-eastern France during the past century. Plant Cell Environm. 21, 565-572.

Ebisuzaki, W., 1997. A method to estimate the statistical significance of a correlation when the data are serially correlated. J. Climate 10, 2147–2153.

Edvardsson, J., Edwards, T. W. D., Linderson, H., Hammarlund, D., 2014. Climate forcing of growth depression in subfossil south Swedish bog pines inferred from stable isotopes. Dendrochronologia 32, 55-61.

Eronen, M., 1979. The retreat of pine forest in Finnish Lapland since the Holocene climatic optimum: a general discussion with radiocarbon evidence from subfossil pines. Fennia 157, 93–114.

Eronen, M., Hyvärinen, H., Zetterberg, P., 1999. Holocene humidity changes in northern Finnish Lapland inferred from lake sediments and submerged Scots pines dated by tree rings. Holocene 9, 569-580.

Eronen, M., Zetterberg, P., Briffa, K.R., Lindholm, M., Meriläinen, J., Timonen, M., 2002 The supra-long Scots pine tree-ring record for Finnish Lapland: Part 1, chronology construction and initial references. Holocene 12, 673-680.

Esper, J., Frank, D.C., Battipalagia, G., Büntgen, U., Holert, C., Treydte, K.S., Siegwolf, R.T.W., Saurer, M., 2010. Low-frequency noise in $\delta^{13}C$ and $\delta^{18}O$ tree ring data: a case study of *Pinus uncinata* in the Spanish Pyrenees. Glob. Biogeochem. Cycles 24, GB4018. http:// dx.doi.org/10.1029/2010GB003772.

Esper, J., Konter, O., Krusic, P.J., Saurer, M., Holzkämper, S., Büntgen, U., 2015. Long-term summer temperature variations in the Pyrenees from detrended stable carbon isotopes. Geochronometria 42, 53–59.

Etheridge, D.M., Steele, L.P., Langenfelds, R.L., Francey, R.J., Barnola, J.-M., Morgan, V.I., 1996. Natural and anthropogenic changes in atmospheric CO_2 over the last1000 years from air in Antarctic ice and firn. J. Geophys. Res. 101, 4115–4128.

Etheridge, D.M., Steele, L.P., Francey, R.J., Langenfelds, R.L., 1998. Atmospheric methane between 1000 A. D. and present: evidence of anthropogenic emissions and climatic variability. J. Geophys. Res. 103, 15979–15996.

Farquhar, G.D., O'Leary, M.H., Berry, J.A., 1982. On the relationship between carbon isotope discrimination and intercellular carbon dioxide concentration in leaves. Australian J. Plant Physiol. 9, 121–137.

Francey, R.J., Farquhar, G.D., 1982. An explanation of $^{13}C/^{12}C$ variations in tree rings. Nature 297, 28–31.

Franke, J., Frank, D., Raible, C.C., Esper, J., Brönnimann, S., 2013. Spectral biases in tree-ring climate proxies. Nature Clim. Change 3, 360–364.

Fritts, H.C., 1976. Tree-rings and Climate. Academic Press, London.

Fritts, H.C., Mosimann, J.E., Bottorff, C.P., 1969. A revised computer program for standardizing tree-ring series. Tree-Ring Bull. 29, 15-20.

Frumkin, A., 2009. Stable isotopes of a subfossil Tamarix tree from the Dead Sea region, Israel, and their implications for the Intermediate Bronze Age environmental crisis. Quaternary Res. 71, 319-328.

Gagen, M., McCarroll, D., Robertson, I., Loader, N.J., Jalkanen, R., 2008. Do tree ring $\delta^{13}C$ series from *Pinus sylvestris* in northern Fennoscandia contain long-term non-climatic trends? Chem. Geol. 252, 42–51.

Gagen, M.H., Zorita, E., McCarroll, D., Young, G.H.F., Grudd, H., Jalkanen, R., Loader, N.J., Robertson, I., Kirchhefer, A., 2011. Cloud response to summertemperatures in Fennoscandia over the last thousand years. Geophys. Res. Lett. 38, L05701, http://dx.doi.org/10.1029/2010GL046216.

Gagen, M.H., McCarroll, D., Jalkanen, R., Loader, N.J., Robertson, I., Young, G.H.F., 2012. A rapid method for the production of robust millennial length stableisotope tree ring series for climate reconstruction. Global Planet. Change 82–83, 96–103.

Gennaretti, F., Huard, D., Naulier, M., Savard, M., Bégin, C., Arseneault, D., Guiot, J., 2017. Bayesian multiproxy temperature reconstruction with black spruce ring widths and stable isotopes from the northern Quebec taiga. Clim. Dyn. 49, 4107-4119.

Gordon, G.A., Gray, B.M., Pilcher, J.R., 1982. Verification of dendroclimatic reconstructions. In: Hughes, M.K., Kelly, P.M., Pilcher, J.R., LaMarche, V.C. Jr (Eds.), Climate from Tree Rings. Cambridge University Press, Cambridge, pp. 58-62.

Grudd, H., Briffa, K.R., Karlén, W., Bartholin T.S., Jones, P.D., Kromer, B., 2002. A 7400-year tree-ring chronology in northern Swedish Lapland: natural climatic variability. Holocene 12, 657-665.

Hantemirov, R.M., Shiyatov, S.G., 2002. A continuous multimillennial ring-width chronology in Yamal, northwestern Siberia. Holocene 12, 717-726.

Holmes, R.L., 1983. Computer-assisted quality control in tree-ring dating and measurement. Tree-Ring Bull. 43, 69-75.

Helama, S., Arppe, L., Hyvönen, J., Mielikäinen, K., Oinonen, M., 2018a. Isoscapes of plant $\delta^{13}C$ in the northern forests: addressing the question of inter-tree and -site variability in *Pinus sylvestris* L. tree rings from Finnish Lapland. Geochem. J. 52, 287-298.

Helama, S., Arppe, L., Timonen, M., Mielikäinen, K., Oinonen, M., 2015. Age-related trends in subfossil tree-ring $\delta^{13}C$ data. Chem. Geol. 416, 28-35.

Helama, S., Arppe, L., Uusitalo, J., Holopainen, J., Mäkelä, H.M., Mäkinen, H., Mielikäinen, K., Nöjd, P., Sutinen, R., Taavitsainen, J.-P., Timonen, M., Oinonen, M., 2018b. Volcanic dust veils from sixth century tree-ring isotopes linked to reduced irradiance, primary production and human health. Sci. Rep. 8, 1339. DOI:10.1038/s41598-018-19760-w.

Helama, S., Arppe, L., Uusitalo, J., Mäkelä, H.M., Oinonen, M., Mielikäinen, K., 2016. Coexisting responses in tree-ring $\delta^{13}C$ to high-latitude climate variability under elevated CO_2: A critical examination of climatic effects and systematic discrimination rate changes. Agr. Forest Meteorol. 226-227, 199-212.

Helama, S., Eronen,M., Timonen, M., 2010. Dendroécologie des bois fossiles dans le nord de la Laponie. In: Payette, S., Filion, L. (Eds.), La Dendroécologie: Principes, methods et applications. Presses de l'Université Laval, Québec, pp. 709–730.

Helama, S., Huhtamaa, H., Verkasalo, E., Läänelaid, A., 2017a. Something old, something new, something borrowed: New insights to human-environment interaction in medieval Novgorod inferred from tree rings. J. Archaeol. Sci: Rep. 13, 341-350.

Helama, S., Lindholm, M., Timonen, M., Eronen, M., 2004a. Dendrochronologically dated changes in the limit of pine in northernmost Finland during the past 7.5 millennia. Boreas 33, 250-259.

Helama, S., Lindholm, M., Timonen, M., Eronen, M., 2004b. Detection of climate signal in dendrochronological data analysis: a comparison of tree-ring standardization methods. Theor. Appl. Climatol. 79, 239-254.

Helama, S., Melvin, T.M., Briffa, K.R., 2017b. Regional curve standardization: State of the art. Holocene 27, 172-177.

Helama, S., Mielikäinen, K., Timonen, M., Eronen M., 2008. Finnish supra-long tree-ring chronology extended to 5634 BC. Norwegian J. Geogr. 62, 271-277.

Helama, S., Sohar, K., Läänelaid, A., Bijak, S., Jaagus J., 2018c. Reconstruction of precipitation variability in Estonia since the 18th century, inferred from oak and spruce tree rings. Clim. Dyn. 50, 4083-4101.

Hultine, K.R., Marshall, J.D., 2000. Altitude trends in conifer leaf morphology and stable carbon isotope composition. Oecologia 123, 32-40.

Huntington, E., 1914. The climatic factor as illustrated in arid America. Carnegie Institution Publication 192, Washington.

Keeling, C.D., Bacastow, R.B., Bainbridge, A.E., Ekdahl, C.A. Jr., Guenther, P.R., Waterman, L.S., Chin, J.F.S., 1976. Atmospheric carbon dioxide variations at Mauna Loa Observatory, Hawaii. Tellus 28, 538-551.

Kelly, P.M., Leuschner, H.H., Briffa, K.R., Harris, I.C., 2002. The climatic interpretation of pan-European signature years in oak ring-width series. Holocene 12, 689-694.

Kress, A., Young, G.H.F., Saurer, M,, Loader, N.J., Siegwolf, R.T.W., McCarroll, D., 2009. Stable isotope coherence in the earlywood and latewood of tree-line conifers. Chem. Geol. 268, 52-57.

Kullman, L., 1992. Orbital forcing and tree-limit history: hypothesis and preliminary interpretation of evidence from Swedish Lappland. Holocene 2, 131–137.

LaMarche, V.C. Jr, 1974. Paleoclimatic inferences from long tree-ring records. Science 183, 1043-1048.

Laumer, W., Andreu, L., Helle, G., Schleser, G.H., Wieloch, T., Wissel, H., 2009. A novel approach for the homogenization of cellulose to use micro-amounts for stable isotope analyses. Rapid Commun. Mass. Sp. 23, 1934-1940.

Leavitt, S., 2010. Tree-ring C-H-O isotope variability and sampling. Sci. Total Environm. 408, 5244-5253.

Leavitt, S.W., Long, A., 1992. Altitudinal differences in $\delta^{13}C$ of bristlecone pine tree rings. Naturwissenschaften 79, 178-180.

Leuenberger, M., 2007. To what extent can ice core data contribute to the understanding of plant ecological developments of the past? In: Dawson, T.E., Siegwolf, R.T.W. (Eds.), Stable Isotopes as Indicators of Ecological Change. Academic Press, London, pp. 211-233.

Liu, Y., Wang, R., Leavitt, S.W., Song, H., Linderholm, H.W., Li, Q., An, Z., 2012. Individual and pooled tree-ring stable-carbon isotope series in Chinese pine from the Nan Wutai region, China: Common signal and climate relationships. Chem. Geol. *330-331*, 17-26.

Loader, N.J., Young, G.H.F., Grudd, H., McCarroll, D., 2013a. Stable carbon isotopes from Torneträsk, northern Sweden provide a millennial length reconstruction of summer sunshine and its relationship to Arctic circulation. Quaternary Sci. Rev. 62, 97-113.

Loader, N.J., Young, G.H.F., McCarroll, D., Wilson, R.J.S., 2013b. Quantifying uncertainty in isotope dendroclimatology. Holocene 23, 1221-1226.

MacDonald, G.M., Velichko, A.A., Kremenetski, C.V., Borisova, O.K., Goleva, A.A., Andreev, A.A., Cwynar, L.C., Riding, R.T., Forman, S.L., Edwards, T.W.D., Aravena, R., Hammarlund, D., Szeicz, J.M., Gattaulin, V.N., 2000. Holocene Treeline History and Climate Change Across Northern Eurasia. Quaternary Res. 53, 302-311.

MacFarling Meure, C., Etheridge, D., Trudinger, C., Steele, P., Langenfelds, R., van Ommen, T., Smith, A., Elkins, J., 2006. The law dome CO_2, CH_4 and N_2O ice core records extended to 2000 years BP. Geophys. Res. Lett. 33, L14810, http://dx.doi.org/10.1029/2006GL026152.

Macias-Fauria, M., Grinsted, A., Helama, S., Holopainen, J., 2012. Persistence matters: estimation of the statistical significance of paleoclimatic reconstruction statistics from autocorrelated time series. Dendrochronologia 30, 179-187.

Magny, M., Leuzinger, U., Bortenschlager, S., Haas, J.N., 2006. Tripartite climate reversal in Central Europe 5600–5300 years ago. Quaternary Res. 65, 3-19.

Martin-Puertas, C., Matthes, K., Brauer, A., Muscheler, R., Hansen, F., Petrick, C., Aldahan, A., Possnert, G., van Geel, B., 2012. Regional atmospheric circulation shifts induced by a grand solar minimum. Nature Geosci. 5, 397-401.

Matskovsky, V.V., Helama, S., 2014. Testing long-term summer temperature reconstruction based on maximum density chronologies obtained by reanalysis of tree-ring data sets from northernmost Sweden and Finland. Clim. Past 10, 1473–1487

Matskovsky, V.V., Helama, S. 2016. Direct transformation of tree-ring measurements into palaeoclimate reconstructions in three-dimensional space. Holocene 26, 439-449.

Matthews, J.A., Briffa, K.R., 2005. The 'Little Ice Age': re-evaluation of an evolving concept. Geogr. Ann. 87A, 17-36.

Mayr, C., Frenzel, B., Friedrich, M., Spurk, M., Stichler, W., Trimborn, P., 2003 Stable carbon- and hydrogen-isotope ratios of subfossil oaks in southern Germany: methodology and application to a composite record for the Holocene. Holocene 13, 393-402.

McCarroll, D., Loader, N.J., 2004. Stable isotopes in tree rings. Quaternary Sci. Rev. 23, 771–801.

Melvin, T.M., Briffa, K.R., 2008. A 'signal-free' approach to dendroclimatic standardisation. Dendrochronologia 26, 71–86.

Melvin, T.M., Grudd, H., Briffa, K.R., 2013. Potential bias in 'updating' tree-ring chronologies using Regional Curve Standardisation: Re-processing 1500 years of Torneträsk density and ring-width data. Holocene 23, 364–373.

Min, S.-K., Zhang, X., Zwiers, F., 2008. Human-induced Arctic moistening. Science 320, 518-520.

Moré, J.J., 1978. The Levenberg-Marquardt Algorithm: Implementation and theory. Lecture Notes in Math. 630, 105-116.

Naulier, M., Savard, M.M., Bégin, C., Gennaretti, F., Arseneault, D., Marion, J., Nicault, A., Bégin, Y., 2015. A millennial summer temperature reconstruction for northeastern Canada using oxygen isotopes in subfossil trees. Clim. Past 11, 1153-1164.

Nicolussi, K., Kaufmann, M., Melvin, T.M., van der Plicht, J., Schießling, P., Thurner, A., 2010. A 9111 year long conifer tree-ring chronology for the European Alps: a base for environmental and climatic investigations. Holocene 19, 909-920.

Nyman, M., Weckström, J., Korhola, A., 2008. Chironomid response to environmental drivers during the Holocene in a shallow treeline lake in northwestern Fennoscandia. Holocene 18, 215-227.

Parding, K.M., Olseth, J.A., Dagestad, K.- F., Liepert, B.G., 2014. Decadal variability of clouds, solar radiation and temperature at a high-latitude coastal site in Norway. Tellus, 66 (25897) DOI: 10.3402/tellusb.v66.25897.

Payette, S., Eronen, M., Jasinski, J.J.P., 2002. The Circumboreal Tundra–Taiga Interface: Late Pleistocene and Holocene Changes. Ambio Spec. Rep. 12, 15–22.

Pilcher, J.R., Baillie, M.G.L., Schmidt, B., Becker, B., 1984. A 7,272-year tree-ring chronology for Western Europe. Nature 312, 150-152.

Pohtila, E., Timonen, M., 1980. Scots pine plantations and their early development in the protection forests of Finnish Lapland. Folia For. 453, 1-18.

Ruxton, G.D., 2006. The unequal variance *t*-test is an underused alternative to Student's t-test and the Mann–Whitney *U* test. Behav. Ecol. 17, 688–690.

Santos, J.A., Belo-Pereira, M., Fraga, H., Pinto, J.G., 2016. Understanding climate change projections for precipitation over western Europe with a weather typing approach. J. Geophys. Res. 121, 1170-1189.

Solovieva N., Jones, V.J., 2002. A multiproxy record of Holocene environmental changes in the central Kola Peninsula, northwest Russia. J. Quaternary Sci. 17, 303-318.

Sonninen, E., Jungner, H., 1996. Carbon isotopes in tree rings of recent and subfossilScots pines from northern Finland. In: Roos, J. (Ed.), The Finnish Research Programme on Climate Change: Final Report. Edita, Helsinki, pp. 19–24.

Speer, J.H., 2010. Fundamentals of Tree-ring Research. The University of Arizona Press, Tucson.

Stokes, M., Smiley, T., 1968. An Introduction to Tree-ring Dating. University of Chicago Press, Chicago (Illinois).

Stuiver, M., Braziunas, T.F., 1987. Tree cellulose $^{13}C/^{12}C$ isotope ratios and climatic change. Nature 328, 58-60.

Treydte, K.S., Frank, D.C., Saurer, M., Helle, G., Schleser, G.H., Esper, J., 2009. Impactof climate and CO_2on a millennium-long tree-ring carbon isotope record. Geochim. Cosmochim. Acta 73, 4635-4647.

Treydte, K., Schleser, G.H., Helle, G., Frank, D.C., Winiger, M., Haug, G.H., Esper, J., 2006. The twentieth century was the wettest period in northern Pakistan over the past millennium. Nature 440, 1179–1182.

van Geel, B., Buurman, J., Waterbolk, H.T., 1996. Archaeological and palaeoecological indications of an abrupt climate change in The Netherlands, and evidence for climatological teleconnections around 2650 BP. J. Quaternary Sci. 11, 451-460.

van Geel, B., Heijnis, H., Charman, D.J., Thompson, G., Engels, S., 2014 Bog burst in the eastern Netherlands triggered by the 2.8 kyr BP climate event. Holocene 24, 1465-1477.

Van Klinken, G.J., van der Plicht, H., Hedges, R.E.M., 1994 Bone $^{13}C/^{12}C$ ratios reflect (palaeo-)climatic variations. Geophys. Res. Lett. 21, 445-448.

Vavrus, S.J., Bhatt, U.S., Alexeev, V.A., 2011. Factors influencing simulated changes in future Arctic cloudiness. J. Clim. 24, 4817–4830.

Veijola, P., 1998 The northern timberline and timberline forests in Fennoscandia. Finnish Forest Research Institute, Research Papers 672, 1–242.

Warren, C.R., McGrath, J.F., Adams, M.A., 2001. Water availability and carbon isotope discrimination in conifers. Oecologia 127, 476-486.

Wieloch, T., Helle, G., Heinrich, I., Voigt, M., Schyma, P., 2011. A novel device for batch-wise isolation of α-cellulose from small-amount wholewood samples. Dendrochronologia 29, 115-117.

Wilson, R., Anchukaitis, K., Briffa, K.R., Büntgen, U., Cook, E., D'Arrigo, R., Davi, N., Esper, J., Frank, D., Gunnarson, B., Hegerl, G., Helama, S., Klesse, S., Krusic, P.J., Linderholm, H.W., Myglan, V., Osborn, T.J., Rydval, M., Schneider, L., Schurer, A., Wiles, G., Zhang, P., Zorita, E., 2016. Last millennium Northern Hemisphere summer temperatures from tree rings: Part I: The long term context. Quaternary Sc. Rev. 134, 1–18

Yang, B., Qin, C., Wang, J., He, M., Melvin, T.M., Osborn, T.J., Briffa, K.R., 2014. 3,500-year tree-ring record of annual precipitation on the northeastern Tibetan Plateau. Proc. Natl. Acad. Sci. USA 111, 2903–2908.

Young, G.H.F., McCarroll, D., Loader, N.J., Kirchhefer, A.J., 2010. A 500-year record of summer near-ground solar radiation from tree-ring stable carbon isotopes. Holocene 20, 315–324.

Young, G.H.F., Demmler, J.C., Gunnarson, B., Kirchhefer, A.J., Loader, N.J., McCarroll, D., 2011. Age trends in tree-ring growth and isotopic archives: a case study of *Pinus sylvestris* L. from northwestern Norway. Global Biogeochem. Cycles 25, http://dx.doi.org/10.1029/2010GB003913.

Young, G.H.F., McCarroll, D., Loader, N.J., Gagen, M.H., Kirchhefer, A.J., Demmler, J.C., 2012. Changes in atmospheric circulation and the Arctic Oscillation preserved within a millennial length reconstruction of summer cloud cover from northern Fennoscandia. Clim. Dyn. 39, 495–507.

Zhang, X., Zwiers, F.W., Hegerl, G.C., Lambert, F.H., Gillett, N.P., Solomon, S., Stott, P.A., Nozawa, T., 2007. Detection of human influence on twentieth-century precipitation trends. Nature 448, 461-465.

TABLES

Table 1. A short summary of the differences between the methods to construct the $\delta^{13}C$ chronologies. They were constructed as a mean of undetrended $\delta^{13}C$ values (‘raw’) and using the regional curve standardization with single (RCS) or multiple (MRCS) curves and with signal-free RCS (SF-RCS) implementation. MRCS chronologies were produced with separate curves for living trees and subfossil data (LI/SF), series according to their relative isotopic enrichment (RIE) –status, and data originating from the western and eastern sites (W/E).

Method	Division	Model(s) applied to detrent the $\delta^{13}C$ series
'raw' $\delta^{13}C$	All	Chronology is a simple mean of undetrended $\delta^{13}C$ series (no RCS models)
RCS	All	A single RCS model for all the $\delta^{13}C$ series
SF-RCS	All	A single SF-RCS model for all the $\delta^{13}C$ series
MRCS	LI/SF	Separate RCS models: living tree and subfossil $\delta^{13}C$ series
SF-MRCS	LI/SF	Separate RCS models: SF-series of living tree and subfossil $\delta^{13}C$ series
MRCS	RIE	Separate RCS models: $\delta^{13}C$ series with RIE > 0 and RIE < 0
SF-MRCS	RIE	Separate RCS models: SF-series of $\delta^{13}C$ series with RIE > 0 and RIE < 0
MRCS	W/E	Separate RCS models: $\delta^{13}C$ series of western and eastern sites
SF-MRCS	W/E	Separate RCS models: SF-series of $\delta^{13}C$ series from western and eastern sites

Table 2. Parameterisation of the models describing the age-related trends in the regional curve standardization with single (RCS) or multiple (MRCS) curves and with signal-free RCS (SF-RCS) implementation. MRCS chronologies were produced with separate curves for living trees and subfossil data, series according to their relative isotopic enrichment (RIE) –status, and data originating from the western and eastern sites. The values of a, b and c are coefficients of the nonlinear model (Eq. 1) and R^2_{ADJ} is the adjusted coefficient of determination.

Model	Division	a	b	c	R^2_{ADJ}
RCS	All	-24.03	1.25	0.95	0.906
SF-RCS	All	-24.05	1.29	0.95	0.894
MRCS	Living	-24.52	1.25	0.51	0.350
RCS	Subfossil	-24.13	1.23	0.94	0.933
SF-MRCS	Living	-24.72	1.24	0.60	0.816
SF-RCS	Subfossil	-24.30	1.13	0.92	0.932
MRCS	RIE > 0	-24.02	0.99	0.89	0.831
MRCS	RIE < 0	-24.05	1.69	0.97	0.856
SF-MRCS	RIE > 0	-24.05	1.02	0.90	0.875
SF-MRCS	RIE < 0	-23.99	1.86	0.97	0.903
MRCS	Western	-24.32	0.81	0.93	0.884
MRCS	Eastern	-24.14	1.33	0.94	0.915
SF-MRCS	Western	-23.64	1.52	0.97	0.954
SF-RCS	Eastern	-24.19	1.28	0.94	0.869

Table 3. Mean $\delta^{13}C$ levels over the first (5500 BC to 3001 BC), middle (3000 BC to 501 BC) and most recent (500 BC to AD 2010) 2.5 ka periods in the resulting $\delta^{13}C$ chronologies constructed as a mean of undetrended $\delta^{13}C$ values (‘raw’) and using the regional curve standardization with single (RCS) or multiple (MRCS) curves and with signal-free RCS (SF-RCS) implementation. MRCS chronologies were produced with separate curves for living trees and subfossil data (LI/SF), series according to their relative isotopic enrichment (RIE) –status, and data originating from the western and eastern sites (W/E). The differences between the first and most recent periods were compared using *t*-test with statistical significance (p).

	Mean $\delta^{13}C$ chronologies								
	’raw' $\delta^{13}C$	RCS All	SF-RCS All	MRCS LI/SF	SF-MRCS LI/SF	MRCS RIE	SF-MRCS RIE	MRCS W/E	SF-MRCS W/E
First 2500 yr	-24.76	-24.81	-24.77	-24.79	-24.78	-24.88	-24.81	-24.83	-24.81
Middle 2500 yr	-24.83	-24.88	-24.84	-24.86	-24.85	-24.90	-24.83	-24.88	-24.86
Last 2500	-24.93	-25.03	-25.00	-25.03	-25.01	-24.92	-24.87	-25.01	-24.98
Difference	-0.172	-0.223	-0.225	-0.249	-0.223	-0.048	-0.055	-0.177	-0.166
t-test	2.645	3.559	3.575	4.145	3.539	1.234	1.386	2.873	2.66
p-value	< 0.01	< 0.001	< 0.001	< 0.001	< 0.001	> 0.1	> 0.1	< 0.01	< 0.01

Table 4. A list of the highest (a) and lowest (b) $\delta^{13}C$ values and their difference (c) in the $\delta^{13}C$ chronologies constructed as a mean of undetrended $\delta^{13}C$ values ('raw') and using the regional curve standardization with single (RCS) or multiple (MRCS) curves and with signal-free RCS (SF-RCS) implementation. MRCS chronologies were produced with separate curves for living trees and subfossil data (LI/SF), series according to their relative isotopic enrichment (RIE) –status, and data originating from the western and eastern sites (W/E). Calendar years refer to mid years of the decades with extreme $\delta^{13}C$ values (e.g. -905 refers to decade from 910 to 901 BC).

	Mean $\delta^{13}C$ chronologies								
	'raw' $\delta^{13}C$	RCS All	SF-RCS All	MRCS LI/SF	SF-MRCS LI/SF	MRCS RIE	SF-MRCS RIE	MRCS W/E	SF-MRCS W/E
(a) High	-23.95	-24.01	-23.97	-23.98	-23.98	-24.09	-24.02	-24.01	-23.98
Decade	-905	-5445	-5445	-5445	-5445	-4935	-4935	-4635	-4635
(b) Low	-26.15	-26.15	-26.11	-26.10	-26.09	-25.87	-25.81	-26.13	-26.13
Decade	715	715	715	715	715	715	-3545	715	715
(c) Difference	2.20	2.14	2.14	2.09	2.10	1.78	1.79	2.15	2.15

Table 5. Pearson correlations of the $\delta^{13}C$ chronologies (a) and the component of the $\delta^{13}C$ variability removed from the chronology (see Eq. (4)) (b) with the records of mean ontogenetic age, mean latitude, longitude and altitude (according to their modern (M) and estimated past (P) altitudes) of the sampling sites (see Fig. 2). Correlations were calculated between 5500 BC and AD 2010. Statistical significance at levels $p < 0.05$, $p < 0.01$ and $p < 0.001$ are denoted by superscripts a, b, and c, respectively.

		Mean $\delta^{13}C$ chronologies								
		'raw' $\delta^{13}C$	RCS All	SF-RCS All	MRCS LI/SF	SF-MRCS LI/SF	MRCS RIE	SF-MRCS RIE	MRCS W/E	SF-MRCS W/E
(a)	Age	0.231[b]	-0.044	-0.050	-0.071	-0.041	-0.030	-0.057	-0.020	-0.021
	Latitude	0.130	0.178	0.179	0.195	0.184	0.050	0.055	0.206[a]	0.209[a]
	Longitude	-0.037	-0.114	-0.116	-0.130	-0.117	0.000	-0.010	0.013	0.038
	Altitude M	0.037	0.109	0.111	0.126	0.109	-0.016	-0.006	-0.014	-0.039
	Altitude P	0.018	0.083	0.084	0.093	0.084	-0.026	-0.017	-0.042	-0.067

		Component of $\delta^{13}C$ variability removed							
		RCS All	SF-RCS All	MRCS LI/SF	SF-MRCS LI/SF	MRCS RIE	SF-MRCS RIE	MRCS W/E	SF-MRCS W/E
(b)	Age	0.982[c]	0.982[c]	0.931[c]	0.928[c]	0.436[c]	0.467[c]	0.845[c]	0.827[c]
	Latitude	-0.157	-0.157	-0.181	-0.169	0.154	0.146	-0.233[b]	-0.242[b]
	Longitude	0.269[a]	0.269[a]	0.276	0.266[a]	-0.062	-0.048	-0.166	-0.242[a]
	Altitude M	-0.249[a]	-0.249	-0.263	-0.237[a]	0.085	0.071	0.171	0.246[a]
	Altitude P	-0.227	-0.227	-0.225	-0.218	0.065	0.053	0.196	0.272[b]

FIGURE CAPTIONS

Fig. 1. Map showing the study area in northern Finland and the subfossil sampling sites. The sites with living tree-ring data are shown with filled squares and the meteorological stations with stars. See Table S1 for site coordinates.

Fig. 2. Mean 'raw' undetrended $\delta^{13}C$ chronology (a) and the mean ontogenetic age of the same data averaged over all samples in each decade (b), the mean latitude (c), longitude (d) and altitude (e) of the sampling sites, presented according to sites' modern and estimated past altitudes. The number of subfossil (n_{SF}) and living tree $\delta^{13}C$ (n_{LI}) samples through mid and late Holocene time (f).

Fig. 3. Mean and modelled age-dependent change in $\delta^{13}C$ data calculated using the regional curve standardization (RCS) with single (RCS) or multiple (MRCS) curves and with signal-free RCS (SF-RCS) implementation for all $\delta^{13}C$ series (a), separately for living tree and subfossil $\delta^{13}C$ series (b), for the $\delta^{13}C$ series with high and low relative isotopic enrichment (RIE > 0 and RIE < 0, respectively) –status (c) and for the $\delta^{13}C$ data from the western and eastern sites (d). See Table 2 for the parameterization of the models.

Fig. 4. Mean $\delta^{13}C$ chronologies constructed using the regional curve standardization (RCS) with single (RCS) or multiple (MRCS) curves and with signal-free RCS (SF-RCS) implementation for all $\delta^{13}C$ series (a), separately for living tree and subfossil $\delta^{13}C$ series (b), for the $\delta^{13}C$ series with high and low relative isotopic enrichment (RIE > 0 and RIE < 0, respectively) –status (c) and for the $\delta^{13}C$ data originate from the western and eastern sites (d).

Fig. 5. Mean Pearson correlations of each of the $\delta^{13}C$ chronologies with other chronologies (a) and the correlations of unsmoothed (Fig. 4) and smoothed (Fig. 6d) $\delta^{13}C$ chronologies against the similarly filtered relative isotopic enrichment (RIE) record (Figs 6a, 6c) (b).

Fig. 6. Mean record of the relative isotopic enrichment (RIE) (a), the proportion of $\delta^{13}C$ series with RIE > 0 (b), the smoothed RIE record (c) and the smoothed $\delta^{13}C$ chronologies (Fig. 4) through mid and late Holocene time. The records were filtered (c, d) using the spline functions corresponding to 500-year rigidity.

Fig. 7. Pearson correlations of to $\delta^{13}C$ data to climate, calculated using the mean (TAM) and maximum temperatures (TAX), the days of clear sky (NN04), of fair weather (NN09), and of overcast (NN20), the average cloud cover (NNM), minimum cloud cover (NNN), and maximum cloud cover (NNX), separately using the records of the three summer months (June, July and August) and the three-monthly mean (JJA) records. Statistical significance (p) was assessed using the Monte Carlo algorithms.

Fig. 8. Calibration and verification statistics for using the average cloud cover as dependent variable to be reconstructed from the $\delta^{13}C$ data over the instrumental period (AD 1891-2010). Statistics include the coefficient of determination (R^2) (a), the Pearson coefficient of correlation (squared; r^2) (b), and the Reduction of Error (RE) (c). Upward arrows indicate the mean level of the same statistics for the bootstrapped $\delta^{13}C$ chronologies (short-dash line) and that of the full sample size (long-dash line).

Fig. 9. Palaeoclimate reconstructions over the past 7.5 ka. Observed (dotted line) and reconstructed (black line) cloud cover June-August season are shown with the 95% (dark grey line) and 99% (light grey line) confidence intervals of the reconstruction, estimated using a Monte Carlo algorithm (Macias-Fauria et al., 2012). The inlet shows the records over the instrumental period (a). The

reconstruction (light grey line) was filtered using the spline functions corresponding to 500-year (dark grey line) and 1500-year rigidities (black line) (b).