Post-print version of the paper by Svarva et al. in *The Holocene* 28(10): 1609-22.

Litte Ice Age summer temperatures in western Norway from a 700-year tree-ring chronology

Svarva HL[1], Thun T[1], Kirchhefer AJ[2], Nesje A[3]

[1]NTNU University Museum, Norway

[2]Dendroøkologen, Skogåsv. 6, NO-9011 Tromsø, Norway

[3]University of Bergen, Norway

Corresponding author:

Helene Løvstrand Svarva

The National Laboratory for Age Determination

NTNU University Museum

Norwegian University of Science and Technology

Sem Sælands vei 5, Gløshaugen

7491 Trondheim

Norway.

Email: helene.svarva@ntnu.no

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Abstract

A ring-width *Pinus sylvestris* chronology from Sogndal in western Norway was created, covering the period AD 1240-2008 and allowing for reconstruction of monthly mean July temperatures. This reconstruction is the first of its kind from western Norway and it aims to densify the existing network of temperature-sensitive tree-ring proxy series to better understand past temperature variability in the Little Ice Age and diminish the spatial uncertainty. Spatial correlation reveals strong agreement with temperatures in southern Norway, especially on the western side of the Scandinavian Mountains. Five prominent cold periods are identified on a decadal timescale, centred on 1480, 1580, 1635, 1709, and 1784 and Little Ice Age cooling spanning from 1450 to the early-18th century. High interannual and decadal agreement is found with an independent temperature reconstruction from western Norway, which is based on data from grain harvests and terminal moraines. The reconstructed temperatures also correlate with other tree ring based temperature reconstructions from Fennoscandia, most strongly with data from central Sweden. Tree growth in Sogndal is correlated to the Scandinavian teleconnection index in the summer months, at least in the last half of the 20th century, and is positively correlated to the summer expression of the North Atlantic Oscillation in the early half of the 20th century. A significant response to major volcanic forcing in the Northern Hemisphere was found, and extreme years seem to be related to the dominance of high and low geopotential height that in turn represents variability in the path of the storm tracks over Fennoscandia. When compared to the variation in frontal positions with time of Nigardsbreen, an eastern outlet glacier from the Jostedalsbreen glacier in western Norway, cold summers in the early-18th century relates to the culmination of a rapid glacial advance that lead up to the 1748 Little Ice Age maximum extent.

Keywords

Late Holocene; Western Norway; Tree rings; Climate reconstruction; Summer temperature; Little Ice Age; Scots pine

Introduction

Fennoscandia is considered an area well-suited for dendroclimatological studies and several millennial or near-millennial temperature reconstructions derive from this region (e.g. Briffa et al. 2008; Grudd 2008; Helama et al. 2010; Linderholm et al. 2010; Gunnarson et al. 2011; Esper et al. 2012; McCarroll et al. 2013; St. George and Ault 2014). These records have improved our understanding of the past temperature development both on local and regional scale and have been utilised in the development of hemispheric scale reconstructions, among the most recent of which include Schneider et al. (2015), Stoffel et al. (2015), Wilson et al. (2016), and Anchukaitis et al. (2017). The link between ring-width chronologies of Scots pine (*Pinus sylvestris* L.) and summer temperature is well established both in the north, and at tree-line sites in the mid latitudes of Fennoscandia, where trees grow in climatically limiting environments (e.g. Briffa et al. 1990; Gunnarson and Linderholm 2002; Helama et al. 2002). In southern Fennoscandia, temperature-sensitive trees are harder to come by and only one published summer temperature reconstruction, based on measurements of maximum tree-ring density, include data from below 62°N (Helama et al. 2014). This results from a switching of the climate signal of Scots pine from temperature sensitive at high latitudes to precipitation sensitive in low altitudes further south (Seftigen et al. 2015). However, in the oceanic climate of western Norway, where summers are relatively cool compared to inland areas (Moen 1998), temperature sensitive trees have been found at 600-950 m a.s.l. as far south as 59°N (Kalela-Brundin 1999b). A reconstruction of mean April-August temperatures from grain harvest data and terminal moraines already exists from western Norway (Nordli et al. 2003), but as historical records are scarce further back in time, this reconstruction only reaches back to 1734 AD. Overall, the understanding of Fennoscandian summer temperature variations would benefit from improving the spatial representation, especially in the period predating 1700 AD and in the southernmost parts (Gouirand et al. 2008; Linderholm et al. 2015; Zhang et al. 2016).

Mean April-September and June-August temperatures have been reconstructed from tree rings in the eastern part of the Scandinavian Mountains in central Sweden (Zhang et al. 2016; Fuentes et al. 2017), and in Norway, mean July and July-August summer temperatures have been reconstructed back to 1500 in Femundsmarka, eastern Norway (Kalela-Brundin 1999a) and at several at sites in northern Norway (Kirchhefer 1999; 2001; 2005). However, on the oceanic side of the Scandinavian Mountains, climate has only been reconstructed from tree rings in northern Norway (Kirchhefer 2001). In south-western Norway, two tree-ring chronologies extend beyond 1700 with sufficient sample depth (Brandt 1975; Thun 2002), but neither is suitable for climatic reconstructions because the samples are from trees at various altitudes and localities and from building timber with unknown provenances.

The term ‘Little Ice Age’ (LIA) has been used to describe the latest and most extensive occurrence of increased glacial activity in the late Holocene (Grove 1988). Most records of frontal variations of Scandinavian glaciers indicate that the LIA glacial maximum was reached during the mid-18th century (Grove 2004), and several sources, both terminal moraine- and historical records describe rapid advances of glaciers in southern Norway and on Iceland during the mid-17th to the early-18th centuries (Grove 1988). Recent studies (Nesje and Dahl 2003; Nesje et al. 2008) have suggested that the early-18th century glacial advance in western Scandinavia was mainly caused by mild, humid winters bringing increased snowfall and that the non-contemporaneous LIA maxima in southern Norway and the Alps can be explained by a prevailing positive North Atlantic Oscillation (NAO) weather mode in the first half of the 18th century. However, the timing and spatial structure of the temperature development during the Little Ice Age are complex, with different reconstructions portraying warm and cold periods for different times, regions, and seasons in the Northern Hemisphere (NH) (Masson-Delmotte et al. 2013). In central Sweden, Zhang et al. (2016) indicated that the LIA spans the mid-16th century to the end of the 19th century. The coldest 100 years of reconstructed April-September temperatures since 850 AD were found from the late-18th to the late-19th centuries, whereas Fuentes et al. (2017) found that the decade centred on 1709 had the coldest reconstructed June-August temperatures for the past 970 years in the same region. Similarly, in northern Fennoscandia and north-western

Russia, McCarroll et al. (2013) identified the 17th century as having the coldest June-August temperatures in the last 1200 years, while in northern Scandinavia, Esper et al. (2012) found the coldest reconstructed 30-year period for the last 2000 years in 1451-1480. LIA variations in summer temperature in an oceanic climate in southern Norway have not yet been explored, and could be of importance when evaluating the influence of summer temperature on the rapid glacier advance in this area in the early-18th century.

Most of the reconstructed climate changes in the LIA have been linked to external forcing factors, such as volcanic eruptions and solar output, in combination with modes of atmospheric variability and internal ocean-atmosphere interactions (e.g. Crowley 2000; Masson-Delmotte et al. 2013). Considering the proximity to the northeast Atlantic, tree-growth in western Norway might reflect an influence of internal dynamics of the climate system in the North Atlantic sector and potential sensitivity to modes of atmospheric variability. These are of high importance to the climate due to their reflection of changes in in the atmospheric wave and jet-stream patterns, and influence on storm tracks and -intensity, precipitation, and temperature over large areas. Recent discussions (e.g. Anchukaitis et al. 2012) have also highlighted the need to improve data coverage and more detailed assessments of the response of temperature sensitive tree-ring records to volcanic forcing. Kalela-Brundin (1999b) related supressed growth in southwest Norwegian Scots pine chronologies to climatic anomalies and ashes associated with the eruption of volcano Laki on Iceland in 1783, indicating that trees in this area should be sensitive to volcanic eruptions in the NH. As a more detailed analysis of the tree response to volcanic forcing in southern Norway is currently lacking, it is explored herein.

The purpose of this study is thus to reconstruct the summer temperature in the LIA in western Norway from tree-ring widths. This dataset is a new contribution to the assemblage of tree-ring proxy series in Fennoscandia. It aims to extend the area covered by regional reconstructions further to the south and west, and thus reduce spatial and temporal uncertainties and improve our understanding of the paleoclimatic variations of northern Europe.

Methods

Study area

In the mountains of southern Norway, elevations surpass the climatic forest line, increasing the probability of temperature being the most limiting factor to tree growth (Fritts 1976). The trees used in this study were sampled near Sogndal in western Norway at 61.16°N 7.09°E (Figure 1). This site is situated at around 750 m a.s.l., close to the local treeline for Scots pine of about 800 m a.s.l., and covers approximately 5 km^2 on a southwest facing slope. The canopy is relatively open, with trees of different ages and sizes distributed evenly. The woodland in the area is composed of Scots pine and *Betula* sp., and the region is classified as weakly oceanic by Moen (1998). The average 1961-90 mean annual precipitation 1543 mm in Sogndal-Selseng, station code: 55730, 20.5 km northwest of the study area at 421 m a.s.l, and the annual mean temperature is 3.4°C and the mean July temperature is 12.5 °C at Sogndal Lufthamn, station code: 55700, 2.3 km southeast of the study area at 497 m a.s.l. Precipitation in the area is highest in autumn and early winter with an average of 197 mm for September to November, whereas spring and early summer are the driest seasons with an average of 79 mm from March to May. April has least precipitation with an average of 57 mm. The snow cover lasts from October/November and into May at 421 m a.s.l., probably remaining somewhat longer at the study site, which is situated at a higher altitude. Adjusting the Lærdal temperatures of the summer months for altitude with a lapse rate of -0.006 °C/m (Tveito et al. 2000) gives a mean monthly temperature above 5.5 °C at the study area from May to August and a mean June-August temperature of 9.6 °C, indicating that the growing season lasts approximately from May/June to August. All values are averages for 1961-90.

[Insert Figure 1.]

Figure 1. Places mentioned in the text. The division between western (W) and eastern (E) Norway (thin black lines) is from Hanssen-Bauer and Nordli (1998). Meteorological stations Lærdal (south of study

area) and Sogndal-Selseng (north of study area) are marked as small red circles. The grain harvest/moraine temperature reconstruction of Nordli et al (2003) is marked as a green rectangle, and tree-ring chronologies as black triangles.

Tree-ring chronology development

One to four cores were taken at breast height with either a 10 mm or a 5 mm increment corer from each of 61 living and 40 dead trees. Only one core from each tree was used in the chronology but many trees, especially snags and logs, required several samples to obtain a core of sufficient quality, i.e. not too damaged by rot. Trees from different age groups were selected to even out the variation in growth variability associated with juvenile compared to ageing trees and trees from a closed canopy, i.e. where the tree crowns touch, were avoided (Konter et al. 2016; Nehrbass-Ahles et al. 2014).

Ring widths were measured to 0.01 mm using an Addo-X measuring device (Parker Instruments, Malmö, Sweden). Cross dating and measurements were quality controlled with the CATRAS (Aniol 1983) and COFECHA (Holmes 1983) softwares. To remove trends due to the decrease of growth rate with aging of the trees, a regional curve approach (RCS) was used (Briffa et al. 1992), with a single 'regional' detrending curve for all series. To investigate potential problems associated with differences in growth rates among trees, detrending runs with separate RCS curves for groups of young and old trees, living and relict trees, and groups of above- and below average growth rate were also investigated. In all cases, detrending was done with the software RCSsigFREE, version 45_v2b (Cook et al. 2014) using an age-dependent spline (Melvin et al. 2007). To minimise removal of the climate signal while removing age trend, detrending curves were calculated by the signal-free method (Briffa and Melvin 2008) and pith offset estimates were used to provide a better fit for each series (Briffa and Melvin 2011). To stabilise the variance, each of the individual tree series were treated beforehand with a power transformation calculated from the relationship between level and spread for each series (Cook and Peters 1997), and the index series were calculated by subtracting the measured value from the fitted detrending function before averaging to produce the final chronology. Expressed population signal (EPS) and Rbar values, which are measurements of the strength of the common signal between the series that make up the chronology, were used to evaluate chronology quality. These were calculated with a 51-year running window with a 1-year lag.

Analysis of the climate-growth relationship

The Norwegian Meteorological Institute provided the instrumental climate data; temperatures from 1870 to present from Lærdal, station code: 54110; homogenised by Andresen (2011), 23.0 km southeast of the study area at 2 m a.s.l, and precipitation data from 1896 to present from Sogndal-Selseng. The stations were chosen due to the length of the records and in the case of Sogndal-Selseng, which only records precipitation, also for being closer to the study area in altitude. Observed SNAO indices were retrieved from Folland et al. (2009), monthly NAO indices from Hurrell et al. (2014), and June-August Scandinavian teleconnection indices from NOAA (ftp://ftp.cpc.ncep.noaa.gov/wd52dg/data/indices/scand_index.tim).

Correlation and response function analyses were computed for mean monthly temperature and total monthly precipitation from September of the previous to August of the current year using the DENDROCLIM2002 software (Biondi and Waikul 2004) to establish the relationship between climate and tree growth. For the response analysis, both the detrended chronology and the meteorological data were converted to first differences, which ensures that the series contain very little autocorrelation or long-term trend. Correlations were calculated from non-transformed data.

Methods of reconstruction and verification

Temperatures were reconstructed by linear regression and verified by splitting the calibration interval in two halves, i.e. 1870-1938 and 1939-2007, and calibrating and verifying on each half using the following parameters: The reduction of error (RE; Fritts et al. 1990), the coefficient of efficiency (CE; Briffa et al. 1988) and the first difference sign test (Fritts 1976). The RE and CE use the mean squared error to compare the accuracy of the reconstruction to the accuracy of average values calculated from the instrumental dataset and the sign test uses first differences to calculate the agreements in sign. The Durbin-Watson test was used to assess the autocorrelation in regression residuals (Durbin and Watson 1951). For the final reconstruction model, calibration was performed over the full 1870-2007 period using average July temperatures from Lærdal (Andresen 2011) as predictand and tree-ring width in year t and t-1 as predictors. Following Zhang et al. (2016), the uncertainty of the reconstruction was estimated as ±2 times the square root of the sum of the squared chronology error and squared calibration uncertainty, where the chronology error was defined as ±2 times the chronology standard error, and the calibration uncertainty as the standard deviation of the reconstruction residuals.

To explore the climate dynamics that control summer temperatures in western Norway, analysis of synoptic patterns related to temperature anomalies was made through composite maps. JJA mean patterns of the 500 hPa geopotential height field (Z500) of Compo et al. (2011) were selected on extreme summer temperatures from observational data (Andresen 2011) and the Sogndal reconstruction (± 1 standard deviation away from a running 21-year median high-pass filter, same treatment for instrumental and reconstructed temperatures). The analysis was extended back to 1659 using the Luterbacher et al. (2002) Z500 reconstruction. In addition, the mean response of the Sogndal reconstruction to major volcanic eruptions were assessed through superposed epoch analysis (SEA; Lough and Fritts 1987) based on the volcanic records of Gao et al. (2008) and Sigl et al. (2015). Following Fuentes et al. (2017), only the 20 largest eruptions in the NH in the last millennium from each dataset was considered (Supplemental Table S1). The analysis, performed with dplR (Bunn et al. 2017), examined 10 years before and after each event and all instances were averaged to determine a mean response with 95% significance thresholds computed through bootstrap resampling with 10 000 iterations.

To put the Sogndal reconstruction in a Fennoscandian context, it was compared to ring width based RCS detrended data covering the same period from Forfjorddalen, northern Norway, standardised chronology index sensitive to June-August temperature (McCarroll et al. 2013; https://www.ncdc.noaa.gov/paleo/study/19943), Torneträsk May-August temperature reconstruction, northern Sweden (Melvin et al. 2013; http://ncdc.noaa.gov/paleo/study/17175) and Jämtland chronology index, sensitive to growing-season temperature, mainly in July from central Sweden (Gunnarson and Linderholm 2002; Gunnarson 2008; https://bolin.su.se/data/Gunnarson-2017) and to a April-August temperature reconstruction AD 1734-1867 from western Norway based on grain harvest dates and terminal moraines (Nordli et al. 2003). The reconstruction was also compared visually to historically reported frontal variations of the glacier Nigardsbreen in western Norway (Nesje et al. 2008) to examine if there is a correspondence between hypothesised extreme low summer temperatures in the late 17th and early-18th centuries and a rapid glacier advance between 1710 and 1735.

Results and discussion

Chronology

The chronology is made up of 83 series with one core from each tree. Seventeen tree-ring series were excluded, for various reasons. One of the excluded trees was cored at an altitude of only 527m and another tree was found unsuitable because it consisted of two trunks that had merged. Five samples had too few rings for reliable cross dating, i.e.<60 years, and the remaining samples were either poorly correlated to the master chronology as indicated by COFECHA, i.e. a correlation between 0.28 and 0.36, and/or showed t-values (Baillie and Pilcher 1973) below 3.9, and affected Rbar and EPS negatively. In some instances, the poor correlation could be attributed to growth depressions indicative of physical damage, i.e. severe and abrupt shifts in growth rate within a single tree that was not reproduced in other trees, and in other cases, rot damage made accurate measurements and the detection of potential missing years difficult. To avoid confounding effects by past defoliation events caused by e.g. insect damage, trees with old dead tops and new dominant crown leaders were avoided. Obvious indications of insect outbreaks were not seen in the tree-ring patterns and none of the observed trees in the study area had fire scars. The Norwegian Institute of Bioeconomy Research (NIBIO; www.skogskader.no) has recorded 17 years with damage to Scots pine forests in Sogn and Fjordane County since 1967. Six of these

(1978, 1979, 1996, 1999, 2002 and 2006) were related to fungi, mainly *Lophodermella sulcigena*, and two (1967 and 1995) were related to defoliation by the pine sawfly (*Neodiprion sertifer*). There are no visual signs of a response to these outbreaks in the Sogndal chronology, and the only year from the NIBIO records in which the sampled trees show growth depressions that seems unrelated to climate is 1984, when damage that was likely caused by winter frost in 1983 was registered on pine growing in the Sogndal municipality.

The mean series length is 199 years with maximum 433 and minimum 67 years, and the mean tree age is 218 years with maximum 448 and minimum 67 years. The mean segment length is relatively even throughout the chronology (Figure 3c). Periods of low regeneration are found in 1350-1600, and 1710-1770. The periods 1590-1620, 1670-1710, and 1770-1785 have slightly elevated regeneration. The sampled trees are situated on the same southwest facing slope and in similar ecological conditions, which in theory should reduce potential problems with trees from different regions or substrates showing different growth rates. However, because detrending with a single RCS curve may introduce bias to the chronology, a double-RCS approach was tested for groups of relict and living trees, for trees with above and below average growth rate, and for young (<218 years, no. 42) and old (>218 years, no.41) trees. Living (no. 46) and relict (no. 37) trees have similar mean growth rates of 1.09 and 0.93 mm, and mean tree ages of 221 and 215 years, respectively. Young trees (no. 42, mean tree age 146 years) have a mean growth rate of 0.84 mm, and old trees (no. 41, mean tree age 293 years) have a mean growth rate of 1.21 mm. Results from attempts with two RCS curves revealed that the two-RCS approach based on groups of young and old trees produced a chronology that was very similar to the one-RCS approach (Supplemental: Figure S1). The two-RCS approach based on living and relict trees, and trees with above and below average growth rate both produced chronologies that gave lower calibration scores in the late half of the calibration interval. To retain sample depth when calculating the growth curve, and because the one-RCS approach produced the best overall calibration and verification results, a single RCS curve was used to detrend the final chronology. The mean chronology spans 769 years from AD 1240 to 2008. Mean EPS and Rbar for the period with >5 samples, i.e. 1343-2007, is 0.902 and 0.402, respectively. EPS values >0.85 are evident back to the start of the 17th century (Figure 3b). They remain reasonably high prior to that, but with a few weaker periods, especially in the early-15th and mid-16th centuries. Rbar remains relatively stable back to the start of the 17th century, when replication begins to drop off. Before that, it has a few periods, especially in the late-14th and early-15th centuries with above average values. Before 1580, replication decreases below ten samples, but EPS fluctuates around 0.85 back to 1343 when sample size drops below five. For the subsequent analyses, only the parts of the chronology with five or more samples is retained, i.e. 1343-2007.

Climate-growth relationships

Correlation analysis of non-transformed data and response analysis of first-differenced data for the previous September to the current August shows that the strongest response and correlation coefficients were found with mean July temperature of the current growing season (Figure 2), which is in accordance with findings elsewhere in Fennoscandia (e.g. Briffa et al. 1990; Grudd et al. 2002; Linderholm and Gunnarson 2005) and Norway (Kalela-Brundin 1999a; Kirchhefer 2001; Linderholm et al. 2003). Significant response and correlation coefficients were also found with mean December temperature of the previous year, and with mean June and August temperatures of the current year. Significant correlations with winter temperatures in the previous year are also found in ring-width chronologies used for reconstructions of summer temperature from northern Finland (Helama et al. 2002), northern Norway (Kirchhefer 2001), central Sweden (Gunnarson and Linderholm 2002), and eastern Norway (Kalela-Brundin 1999a). When the first-differenced data was analysed over an early (1898-1952) and a late (1953-2007) interval, the only parameter that was significant in both intervals was mean July temperature. Mean October temperature of the previous year and mean June temperature of the current year were only significant in the early half, and mean February, April and August temperatures of the current year were only significant in the late half. No significant response to precipitation was detected when examining split intervals. This confirms that mean July temperature is by far the most influential and time-stable factor that influences growth, but that other climate parameters are of importance at shorter time intervals. First-order autocorrelation is 0.82 for the Sogndal chronology. Usually, the growth of Scots pine in Fennoscandia in a specific year also depend on climate in the previous year (e.g. Fritts 1976). This is illustrated for the Sogndal chronology by significant and positive correlations with mean June and July temperatures of the previous year (Figure 2a). To further explore this relationship, stepwise linear regression with July temperature as predictand and growth value for t as well as the three previous and three following years as predictors was used. Two significant ($p<0.05$) predictors were chosen for the following reconstruction of mean July temperature: tree-ring indices in year t and t-1.

[Insert Figure 2]

Figure 2. a) Correlation coefficients of non-transformed data: the Sogndal chronology with Lærdal (Andresen 2011) mean monthly temperature and Sogndal-Selseng (station code: 55730) total monthly precipitation for the previous September to the current August, analysis performed over the period 1898-2007, and correlations with June, July and August temperatures of the previous year. b) Response coefficients of first-differenced data. Filled bars denote significant ($p<0.05$) and empty bars denote not significant correlation and response. c) Climate diagram 1961-90 for Lærdal (monthly mean temperature) and Sogndal-Selseng (total monthly precipitation).

No significant correlations were found between the Sogndal index chronology and monthly NAO indices for 1821-2007, and no significant correlation (r=0.03, p=0.737) was found with the SNAO index over the period 1850-2007. 11-year running correlation analysis (not shown) revealed that the relationship between the Sogndal index chronology and the SNAO index shifts between negative and positive throughout the analysed time interval, but that the coefficients are positive between 1890 and 1950. The correlation in this period is 0.38 (p=0.002), and 0.25 (p=0.057) for first-differences. Both temperature and precipitation in western Norway are positively correlated to the NAO index in winter, especially in the south-western parts, but in summer, the correlations are much weaker (Hanssen-Bauer 2005; Folland et al. 2009). In addition, the correspondence between temperature and NAO indices in the whole of Norway is highly variable with time for all seasons and fluctuates between positive and negative correlation coefficients during summer. Hanssen-Bauer and Førland (2000) showed that more locally defined atmospheric circulation indices might account for more of the local variance in temperature and precipitation in Norway than the NAO, which is likely also the case for tree growth in Sogndal. Barnston and Livezey (1987) described the Eurasia-1 teleconnection pattern (later renamed the Scandinavian pattern), which has the primary centre of action over the Scandinavian peninsula with centres of action of opposite sign over the northeast Atlantic and over central Siberia to the southwest of Lake Baikal. In the summer months, the centre over Siberia is shifted slightly to the north and west (CPC 2005). The correlation between the Sogndal index chronology and the average Scandinavian teleconnection index in June-August for the period 1950-2007 is 0.33 ($p=0.012$), and 0.41 ($p=0.001$) for first-differences. For the same period, the correlation between the Sogndal index chronology and the SNAO index is -0.03 ($p=0.824$), indicating that the Scandinavian pattern accounts for more of the interannual variation in tree ring width in Sogndal than the SNAO, at least in the last half of the 20th century.

Temperature reconstruction

Mean July temperatures were reconstructed by linear regression from the relationship between the Sogndal chronology in year t and t-1 and Lærdal mean July temperature (Andresen 2011). Attempts to reconstruct July-August mean temperatures or June-August mean temperatures did not verify successfully, i.e. they resulted in CE values below zero. The inclusion of the chronology index of the previous year had a positive effect both on calibration and verification statistics. Calibration and verification statistics (Table 1) are comparable to reconstructions based on tree-ring width in Jämtland (Linderholm and Gunnarson 2005) and coastal northern Norway (Kirchhefer 2001). The prediction errors have low serial correlation. The Durbin-Watson statistic is 1.76, which is not significantly different from a null hypothesis of no autocorrelation ($p<0.05$).

Table 1. Verification of the mean July temperature reconstruction. Significance levels of the sign test are all at $p<0.01$ (Fritts, 1976, p. 330) with number of agreements (disagreements) given for each period.

Full interval (1870-2007)	
R^2	41 %
r	0.64
Model	$T=-7.89 + (9.04\ I_t) + (-2.00\ I_{t-1})$
Early calibration (1870-1938)	
R^2	40 %
r	0.63
RE	0.46
CE	0.19
Sign test	49(17)
Late calibration (1939-2007)	
R^2	39 %
r	0.63
RE	0.41
CE	0.23
Sign test	51(16)

The decadal and inter-annual variations in temperature are presented in Figure 3e. Compared to the observation data, the reconstruction has reduced amplitude in the decades around 1900 and fails to estimate extremes in this period, even though the year-to-year variation is in good agreement with the instrumental data (Figure 3a). The standard deviation in the Lærdal record between 1901 and 1915 is 2.0 °C compared to 1.6 °C for the CRU TS4.01 mean July temperatures of the closest grid cell, suggesting an increase in variation in the Lærdal record during this period, a feature not uncommon for the early instrumental measurements (Frank et al. 2007). For the period 1950-2007, the standard deviation is approximately the same (1.1°C) in the Lærdal and CRU measurements.

[Insert figure 3]

Figure 3. a) Observed and reconstructed mean July temperatures for 1870-2007. b) EPS, Rbar, and sample replication. c) Mean tree age and mean segment length. d) Segment time span and estimated germination date e) Reconstructed mean July temperatures expressed as anomalies relative to 1900-2007, with 30-yr smoothing spline. Combined chronology and calibration uncertainty in light yellow shade and extreme cold and warm years are marked at the bottom of the plot area (see text for details). Sample depth drops below 5 in 1343 AD.

The Sogndal reconstruction is positively correlated to gridded mean July temperatures (Harris et al. 2014) in southern Fennoscandia, on Iceland and western Greenland, and negatively correlated to July temperatures in the Iberian peninsula and in western Siberia, an area just east of the Ural Mountains, stretching from Novaya Zemlya in the north to around 50°N (Figure 4). These patterns, i.e. correlations of opposite sign in Scandinavia, and in Siberia and the Iberian peninsula, resemble the correlations between the Scandinavian teleconnection index and monthly surface temperature for the summer months (CPC 2005). The strongest correlations between the Sogndal reconstruction and mean July temperature are found in the mountains and maritime areas of southern Norway, reflecting the predominant westerly airflow over this area. The correlations diminish towards central Norway, but weaker positive correlations extend east into Finland and the Baltic states. Measured temperatures (Hanssen-Bauer 2005) in western Norway in June, July and August have correlations above 0.6 with gridded temperatures in Scotland, on Shetland, and in Denmark. Lower correlations also reach Iceland. Although moderate correlations are present between the Sogndal reconstruction and gridded July temperatures on Iceland, Shetland and in Denmark, they are not strong enough to provide a reliable reconstruction for these sites.

[Insert Figure 4.]

Figure 4. Spatial correlations ($p<0.5$) between the Sogndal July temperature reconstruction and gridded (0.5°) CRU TS4.01 mean July temperature for 1901-2007 (Harris et al. 2014) for (*top*) northwest Eurasia and (*bottom*) Fennoscandia. Maps produced with KNMI Climate Explorer (Van Oldenborgh et al. 2009).

Comparison with other temperature reconstructions

This record is the southernmost tree-ring based reconstruction of summer temperatures in the LIA in western Fennoscandia. Notable warm periods are reconstructed around 1400, in the mid-18th century and the early-19th century, and five major cold periods are evident, centred on 1480, 1580, 1635, 1709, and 1784. The cold periods around 1580, 1709 and 1784 have also been reconstructed for eastern Norway (Kalela-Brundin 1999a), central Sweden (Fuentes et al. 2017; Zhang et al. 2016), and northern

Fennoscandia (Melvin et al. 2013). They seem to have prevailed over a large part of the warm season as these studies target June-August or April-September temperatures. A cold period centred on 1635 is evident in Sogndal, Femundsmarka (Kalela-Brundin 1999a), Rogen, central Sweden (Fuentes et al. 2017) and Scotland (Rydval et al. 2017), but is warmer in the C-Scan reconstruction from central Sweden (Zhang et al. 2016). Likewise, the cold period centred on 1480 is not as clearly defined in blue intensity and density data from central Sweden (Fuentes et al. 2017; Zhang et al. 2016), but is more evident in ring-width records (Gunnarson 2008; Linderholm and Gunnarson 2005). However, these years have been shown to have been cold both in Scotland (Rydval et al. 2017) and south Finland (Helama et al. 2014) and a protracted cold period between 1450 and 1500 is evident in the C-Scan reconstruction, and in northern Fennoscandia (Esper et al. 2014), and in the Alps (Büntgen et al. 2006). In fact, the late 15th century has the coldest reconstructed temperatures in Sogndal and corresponds to the early part of the Spörer Minimum (1460-1550; Eddy 1976). Also, this is a period with closely spaced volcanic eruptions in 1452, 1474, 1476/77 and 1480, among them the eruption of volcano Bárðarbunga on Iceland (Briffa et al. 1998; McCarroll et al. 2013). McCarroll et al. (2013) noted that multiple eruptions in quick succession might have a much greater and more sustained impact on the climate, or tree response, than a single larger eruption, which also seems to be of importance for the climate during the LIA in western Norway. Additional exceptionally cold single years in Sogndal include 1455, which might relate to a volcanic eruption in 1452/3 (Gao et al. 2008; Sigl et al. 2015), 1577, 1741 and 1928. Grove (1988) mentions 1741 and 1742 as years when the harvest failed and death rates peaked over large parts of Norway and 1709 is also known for severe famines and low winter temperatures in Europe (Grove, 2007). Another notable cooling event around 1600, which includes a volcanic eruption in 1601 (e.g. Briffa et al. 1998) seems less severe in the Sogndal reconstruction than would be expected from looking at other tree-ring records. This cold period is often described as lasting for several decades (e.g. Fuentes et al. 2017; Kalela-Brundin 1999a; Gunnarson et al. 2011; Björklund et al. 2013; Zhang et al. 2016), but it is of shorter magnitude and duration in the early-17th century in Sogndal. The 1600 cold is also less pronounced in Forfjorddalen, but this record is out of phase with the Sogndal reconstruction in this period. Although a local disturbance that affected tree growth in Sogndal cannot be ruled out, both EPS and Rbar remains relatively high despite a general decline in sample size and geographical changes in temperature patterns (Gouirand et al. 2008) and the stronger maritime influence at the study area might be the reason for the observed differences.

To investigate the spatial variation of LIA temperatures more closely, the Sogndal reconstruction is compared to RCS detrended ring-width records in Jämtland (Gunnarson 2008), Torneträsk (MJJA temperature reconstruction; Melvin et al. 2013) and Forfjorddalen (McCarroll et al. 2013). These four chronologies represent two sites on the western side of the Scandinavian Mountains, Forfjorddalen and Sogndal, and two sites on the eastern slope, Torneträsk and Jämtland (Figure 5b-d). The chronologies on the eastern slope of the Scandinavian Mountains seems to have the most common variability, both on interannual and longer timescales, and the difference on longer timescales between coastal and inland chronologies is largest in the north although the interannual agreement between Forfjorddalen and Torneträsk is high (Table 2). The new Sogndal reconstruction correlates most strongly with Jämtland, which is as expected considering the distance between the sites. The synchrony between these records is highest in the 18th and 19th centuries and in the mid- and late-15th century (Figure 5e). There is a weaker correlation in the first half of the 16th century, when the Sogndal reconstruction has the best agreement with the tree-ring chronology from Forfjorddalen. The agreement between Forfjorddalen and Sogndal is strongest for medium to long-term variation and weaker for interannual variation. The reconstruction from Torneträsk, however, has a strong correlation to Sogndal on intereannual timescales, but a weaker agreement in the long-term variation. The higher agreement between the smoothed time series of Sogndal and Forfjorddalen might be a reflection of the oceanic influence at these sites. The Sogndal reconstruction portrays LIA cooling from around 1450, with gradual warming from the early- to mid-18th century, where both the smoothed Sogndal and Forfjorddalen records are mostly above the long-term average.

The cold spells around 1480, 1635 and 1709 are most evident at the sites in southern Fennoscandia.

Table 2. Correlation matrix of the Sogndal reconstruction and RCS detrended temperature sensitive ring-width records from Forfjorddalen and Jämtland and May-August temperature reconstruction from Torneträsk for 1343-2005. Black numbers are correlations for non-transformed data, and grey numbers are correlations of first-differenced data. All correlations are significant (p<0.001).

		JÄM	TOR	FOR	
		0.55	0.37	0.27	SOG
JÄM	0.48		0.48	0.35	JÄM
TOR	0.29	0.55		0.62	TOR
FOR	0.41	0.31	0.28		
	SOG	JÄM	TOR		

[Insert Figure 5.]

Figure 5. Sogndal men July temperature reconstruction scaled to the mean and variance of the Nordli et al. (2003) grain harvest/moraine spring-summer temperature reconstruction (a). Sogndal Mean July temperature reconstruction compared to (b) Forfjorddalen standardised growth index, sensitive to JJA temperature (McCarroll et al. 2013), (c) Torneträsk MJJA temperature reconstruction (Melvin et al. 2012), and (c) Jämtland ring-width index, sensitive to JA temperature (Gunnarson 2008). For (b-d), the series are normalised over the common period of overlap, i.e. 1343-2005. (d) 51-year running correlation between the Sogndal reconstruction and the Jämtland, Torneträsk and Forfjorddalen records.

Nordli et al. (2003) reconstructed the mean spring-summer temperatures for AD 1734-1867 in western Norway, using the first day of grain harvest from farmer's diaries. To account for differences in growing conditions at different farms and the use of different cereal varieties, glaciological data reconstructed from established moraine chronologies was used to reconstruct the long-term variation in temperature (only grain-harvest data for 1843-1867). As cereal is usually grown at a lower altitude than the sampled trees in Sogndal, they reflect the temperatures of a longer growing season, i.e. April-August compared to July and because they are grown annually, they are also unlikely to be influenced by the climate outside of the growing season. In addition, cereals might respond differently than trees to climate events unrelated to temperature. For instance, high amounts of precipitation during spring and summer might destroy or delay the harvest date for crops grown in the valleys, but have much less of an impact on trees growing on mountainsides. Taking such considerations into account, a comparison of reconstructed temperatures from tree-rings and grain harvest dates from western Norway is a solid test of the reliability of both records beyond the calibration interval as they are derived from independent sources of data. Both the decadal and the inter-annual agreements between the harvest/moraine reconstruction and the Sogndal reconstruction are apparent, confirming their ability to reproduce temperature, and the temporal stability of the climate signals (Figure 5a). Inter-annual synchrony between the two reconstructions is high, $r=0.44$ ($p<0.01$) for non-transformed, and $r=0.47$ ($p<0.01$)

for first-differenced data. The long-term variation in the harvest/moraine reconstruction was based on terminal moraine sequences, and the agreement with the Sogndal reconstruction indicates that the tree-ring data capture the same low to medium-frequency variation, confirming a link between summer-season-forcing of glaciers in western Norway and ring widths of the Sogndal reconstruction. In Sogndal, a cold period is reconstructed in 1783-1787 and is likely an effect of the eruption of volcano Laki on Iceland in 1783 (e.g. Thordarson and Self 2003), which is known to have caused damage to crops in southern Norway (Kalela-Brundin 1999b). Mean July reconstructed temperatures in this period is 1.97°C below the average of the overlapping years, i.e. 1734-1867, in Sogndal, but in the harvest/moraine reconstruction, the extreme cold is restricted to 1784 although 1783-1787 are all below the average value. Historical and meteorological records show that July 1783 was unusually warm in northern and western Europe (Thordarson and Self 2003). This warmth is not reproduced in either the ring-width or the harvest/moraine reconstruction and it would seem that acidic aerosols or dust from the eruption of Laki inhibited the growth of both Scots pine and *Hordeum vulgare* L. in western Norway in 1783, as proposed by Briffa et al. (1988) in the case of Scots pine.

Synoptic patterns and forcing of extreme years

Banston and Livezey (1987) described the Eurasia-1 (Scandinavia) teleconnection pattern by using monthly mean gridded 700 mb heights as primary data. As the Sogndal chronology is significantly correlated to the Scandinavian teleconnection index in June-August, a spatial composite analysis for June-August was made for extreme years (Figure 3e; Supplemental Table S2) with the 500 hPa geopotential height field. The patterns are similar between observed extreme cold and warm extremes and reconstructed extreme cold and warm extremes (Figure 6). For both positive and negative extremes, they share some of the features of the SNAO, i.e. a dipole over United Kingdom and Greenland (Folland et al. 2009), and of the Scandinavian teleconnection pattern in summer, i.e. a centre of action over Fennoscandia and centres of opposite sign over Siberia and the northeast Atlantic (CPC 2005). The warm extremes in Sogndal are characterised by a dipole pattern with positive geopotential height anomalies over Fennoscandia and negative geopotential height anomalies over the Greenland region, which is similar to that identified by Seftigen et al. (2015) representing drought, and warm extremes (Zhang et al. 2017). This pattern of anomalous anticyclonic circulation over northern Europe is associated with meridional shifts in the northern mid-latitude jet stream and favours negative wind anomalies and drier, warmer and less cloudy conditions over Fennoscandia and the United Kingdom, and cooler and wetter conditions over Greenland, central Europe and

Russia (Linderholm et al. 2011; Zhang et al. 2017). Tree-ring data from central Sweden produces similar patterns with sea level pressure (SLP) and an opposite pattern during extreme cold summers (Fuentes et al. 2017). In Sogndal, however, cold extremes produce a dipole pattern with negative height anomalies over Fennoscandia and positive height anomalies over the Bay of Biscay both for Z500 and SLP (not shown). This pattern is related to anomalous westerly/north-westerly winds directed towards Scandinavia, which are associated with cold air in summer and has been shown to influence temperature variance along the entire west coast of Norway (Hanssen-Bauer and Førland 2000) and extreme cold years in reconstructed summer temperatures from northern Scandinavia (Büntgen et al. 2011). The fields produced with reconstructed geopotential height are similar, confirming the stability of these relationships back to the mid-17th century. Cold extremes were more frequent than warm extremes during the LIA when using a one standard deviation threshold, indicating that negative pressure anomalies over Fennoscandia dominated during this period (Figure 3e). This is especially prominent in the late-15th, late-16th and late-17th centuries, periods that together have only one extreme warm year. The ratio of extreme warm to extreme cold years in from the mid-15th century to the end of the 17th century is 0.46.

[Insert Figure 6.]

Figure 6. Composite fields of 20th century reanalysis (C20C v2c; Compo et al. 2011) June-August geopotential height deviations in the 500 hPa level (m) selected on years with a) Positive (*left)* and negative (*right)* temperature extremes (1880-1999) from observational data (Andresen 2011) b) Positive (left) and negative (right) temperature extremes from the RCS reconstruction (1851-1999). c) Composite fields of reconstructed (Luterbacher et al. 2002) June-August geopotential height deviations in the 500 hPa level (dm) selected on positive extremes (*left*) and negative extremes (*right*) from the RCS reconstruction (1659-1999). Maps produced with KNMI Climate Explorer (Van Oldenborgh et al. 2009). Only significant values ($p>0.05$) are shown. Extreme years are defined as being ±1 SD away from a 21-year median low-pass filter.

The magnitude and timing of responses to major volcanic eruptions in Sogndal varied depending on which dataset was used (Figure 7). Significant growth depressions in the Sogndal reconstruction are found at lags +1 and +5 when compared to the Gao et al. (2008) dataset, yielding a mean cooling response of -0.44°C and -0.50°C respectively. For the Sigl et al. (2015) dataset, there is a significant response in year 2+ with a mean cooling response of -0.49°C. The results show that volcanic eruptions influenced tree growth and summer climate in western Norway in the LIA, which is in accordance with tree-ring based temperature reconstructions from other sites in the NH (e.g. Fuentes et al. 2017; Linderholm et al. 2015; Rydval et al. 2016; Stoffel et al. 2015). It has been demonstrated that prominent climatic episodes such as the Medieval Warm Period and the LIA were also influenced by solar output (Masson-Delmotte et al. 2013), but the Sogndal reconstruction is not long enough to provide accurate estimations of orbitally forced temperature signals.

[Insert Figure 7.]

Figure 7. Superposed epoch analysis of the response to volcanic eruptions in the Northern Hemisphere for the Sogndal reconstruction. *Top*: data from Gao et al. (2008). *Bottom*: data from Sigl et al. (2015). Years outside the 95% confidence envelopes (dotted lines) are marked with an asterisk.

[Insert Figure 8.]

Figure 8. Historically reported and measured frontal variations of glacier Nigardsbreen from Nesje et al. (2008) compared to reconstructed mean July temperatures expressed as anomalies relative to 1650-2007.

Summer temperatures during the LIA glacial maximum

Both temperature and precipitation throughout the year influence glacier growth, but recent studies have emphasised the importance of winter precipitation in accounting for the rapid LIA advance of maritime glaciers, i.e. glaciers with a high mass balance gradient (more accumulation, more ablation), in western Norway (Nesje and Dahl 2003; Matthews and Briffa 2005; Nesje et al. 2008; Rasmussen et al. 2010). The most detailed data on the LIA glaciation in Scandinavia comes from Nigardsbreen (61°42'N, 7°08'E) in western Norway. This is the only glacier in southern Norway with reliable historical data on glacier advances in the late-17th and early-18th centuries and its geographical position close to Sogndal warrants a comparison of reconstructed mean July temperatures to the Nigardsbreen frontal positions in the years leading up to the LIA glacial maximum. Records from this glacier suggest a high mean advance rate during the late-17th and early-18th centuries. Between 1710 and 1735, Nigardsbreen advanced 2800 m and between 1735 and the LIA maximum in 1748, the glacier advanced a further 150 m (Nesje et al. 2008). The estimated frontal time lag at Nigardsbreen is 20 to 25 years (Nesje and Dahl 2003), making the climate in the period between 1685 and 1728 of interest to the rapid LIA glacial advance. Temperatures in Sogndal are -0.89°C below the 1961-90 average in this period, which is colder than the assumptions of Nesje et al. (2008) and Rasmussen et al. (2010) of mild summers during the first part of the 18th century, and a -0.5°C LIA temperature anomaly, respectively. The 1710s is the coldest reconstructed decade in Sogndal after 1650, and from the 1710s and to the late-1720s, temperatures in Sogndal increase towards the long-term average, which matches the end of the rapid glacier advance when the delay in frontal response is taken into account. This however, does not have to mean that the influence of precipitation has been less important than it is today, as it seems that the response of the advance is not linear with temperature when comparing the advance periods 1710-1735 and 1735-1748 to preceeding temperatures from around 1660-1710, which were also low.

Conclusions

Measurements of ring widths on trees from Sogndal were used to reconstruct mean July temperature variability in western Norway in the Little Ice Age. The spatial pattern of correlations of the reconstruction to instrumental temperatures is strongest in southern Norway, especially in maritime areas. The agreement with the Jämtland chronology in central Sweden is strong, reflecting the close proximity between the two sites and the predominant westerly air flow in these areas. Five cold periods were identified, centred on 1480, 1580, 1635, 1709, and 1784 and are supported by findings elsewhere in Fennoscandia, although a cold period in the early-17th century is of shorter duration and magnitude in Sogndal than has been reported at other sites. LIA cooling was found from the mid-15th century, with gradual warming from the start of the 18th century. When compared to a temperature reconstruction for western Norway based on grain harvest dates and terminal moraines from 1734 to 1867, the two independent proxies show common variation both on annual and decadal timescales. Correlations with the SNAO were variable with time and at least for the late half of the 20th century, the Scandinavian teleconnection index seemed to account for more of the variation in tree-ring width. Extreme July temperatures were associated with patterns of 500 mb geopotential height with a centre over Scandinavia and centres of opposite sign in Siberia, the northeast Atlantic (for negative extremes) and over Greenland and the

Mediterranean (for positive extremes). These patterns resembles in part the SNAO, and in part the Scandinavian teleconnection pattern, and might be related to shifts in the storm tracks over Fennoscandia, where positive extremes are associated with negative wind anomalies and a northward shift in the storm track, and negative extremes are associated with anomalous westerly winds directed towards Scandinavia. The patterns appeared to be stable back in time, at least to the mid-17th century and the frequency and distribution of extreme warm and cold years indicate that negative pressure anomalies over Fennoscandia were prominent during the LIA. A significant response to volcanic forcing was detected and the coldest reconstructed mean July temperatures were in the late half of the 15th century, which is a period with closely spaced volcanic eruptions. Finally, a comparison between frontal variations of the glacier Nigardsbreen and reconstructed temperatures shows that a spell of cold summers in the early-18th century matches the culmination of a rapid glacier advance leading up to the LIA maximum extent.

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