Paper 5

SPRING–SUMMER TEMPERATURE RECONSTRUCTION IN WESTERN NORWAY 1734–2003: A DATA-SYNTHESIS APPROACH

P. Ø. NORDLI,[a,]* Ø. LIE,[b,c] A. NESJE[b,c] and S. O. DAHL[c,d]

[a] *Norwegian Meteorological Institute, Box 43 Blindern, N-0313 Oslo, Norway*
[b] *Department of Earth Science, University of Bergen, Allégt. 41, N-5007 Bergen, Norway*
[c] *The Bjerknes Centre for Climate Research, Allégt. 55, N-5007 Bergen, Norway*
[d] *Department of Geography, University of Bergen, Breiviksveien 40, N-5045 Bergen, Norway*

ABSTRACT

A series of spring–summer (April–August) temperatures was reconstructed for the period 1734–1923 for western Norway based on multi-proxy data. For the period 1734–1842 the long-term variations were based on terminal moraines in front of two southern Norwegian glaciers, whereas the annual variations were based on grain-harvest data extracted from farmers' diaries. For the period 1843–1867 the spring–summer temperatures were reconstructed solely from diaries overlapping instrumental observations. All the results were incorporated into one series for the period 1734–2003 to form the Vestlandet composite series.

The reconstruction method using terminal-moraine sequences was tested against the modern instrumental Bergen series for the periods of moraine formations in front of the glaciers. The agreement with the instrumental series was good, with the mean difference for all periods being only 0.2 °C. Analyses of decadal variations in western Norway revealed three periods of low spring–summer temperatures: around 1740, in the first decade of the 19th century, and in the 1830s. These periods are well known from historic records as periods of starvation, during which the use of bark bread became common. Copyright © 2003 Royal Meteorological Society.

KEY WORDS: temperature; circulation index; farmers' diaries; grain harvest; moraine; equilibrium line altitude; glacier; Little Ice Age; western Norway; stepwise regression analysis

1. INTRODUCTION

Early drawings from the 18th and 19th centuries show glaciers far advanced compared with their present situation, and historical documents about advancing glaciers destroying pastures and even farmhouses were known to early historians and geographers (Hoel and Werenskiold, 1962). The glacial advance culminated in the first part of the 18th century for most glaciers in Scandinavia. The period with considerable glacial advance was later given a specific name: the Little Ice Age (LIA) (e.g. Grove, 1988).

Early Norwegian historians tried to integrate the historical documents with the accounts of the 'topographers' telling of glacier advance. For example, Øverland (1890) interpreted the events as a result of a climatic deterioration leading to much suffering for the Norwegian people. This interpretation is, in itself, problematic, because glacier fluctuations are not a result of summer temperature only, but also accumulation of snow during winter. Large accumulation of snow is commonly connected with mild winters, as demonstrated during the mild 1990s, when the westernmost glaciers in southern and middle Norway advanced (Kjøllmoen, 2001). Climatic variability is greater during winter than summer, so mild winters can often also result in high mean annual temperatures.

---

* Correspondence to: P. Ø. Nordli, Norwegian Meteorological Institute, Box 43 Blindern, N-0313 Oslo, Norway;
e-mail: oyvind.nordli@met.no

There is no accepted definition of the onset of the LIA. According to Grove (2001), the LIA was 'under way' during the 13th and 14th centuries in the North Atlantic region. However, the dominant use of the term LIA has come to mean a widespread cold period occurring most commonly around AD 1500 or 1550 and lasting to 1800 or 1850 (Ogilvie and Jónsson, 2001). More recently, progress in palaeoclimatology has made it possible to reconstruct millennium Northern Hemisphere temperatures derived from proxies, most of them representing the growing season (e.g. Jones *et al.*, 1998; Mann *et al.*, 1999). These show decreasing temperature during most of the millennium, with the coldest periods in the 17th and early 19th centuries. The last minimum was succeeded by a temperature increase up to the present day.

Neither the lower temperatures nor the climate variability were globally synchronous during the LIA (Crowley and North, 1991). Recent studies suggest that it was by no means a single period with below-20th century temperatures, but rather a series of decade-long cool periods separated by warmer intervals (Grove, 1988; Jones and Bradley, 1992; Pfister, 1992; Nesje and Dahl, 2003).

Although some temperature measurements started in western Norway in 1818, a network of modern instrumental observations was not established before the late 1860s with the foundation of The Norwegian Meteorological Institute. To expand the series further back in time, and as a control for the early observations, other sources of information should be used. One approach is to utilize documentary weather-related records. These may be classified into two main groups with respect to the kind of weather information they contain (Ingram *et al.*, 1981):

1. Meteorological data that are directly descriptive of climatic conditions, such as temperature, wind direction, wind strength, etc.
2. Proxy data that are indirect measures of climate. These reflect the effect of weather phenomena on nature and society. Records containing this kind of information are called phenological data (Bradley, 1985) or historical proxy data (Nordli, 2001a) if retrieved from historical documents.

The information from the second group may be related to temperature by regression analysis. Examples are the spring–summer temperature reconstruction based on wine harvests (Le Roy Ladurie and Baulant, 1980) and on the ripening time of rye (Tarand and Kuiv, 1994). Recently, Rutishauer (2001) has related summer temperature in southwestern Sweden to 'korntal' and found that 38% of the variance was accounted for by the regression model. In Norway, the first day of harvest for barley (to some extent also oats and rye) has been related to spring–summer temperature. The variance accounted for by the regression model varied between 61 and 94% (Nordli, 2001a,b).

As historical records of harvest dates rarely cover more than 50 years, records from several farms must be used to expand the period of temperature reconstruction for the LIA maximum in the 18th century. The annual variations seem to be robust, whereas the long-term trends seem to be more sensitive to shifts of location. This is not only caused by different growing conditions at the farms, but also by the use of different cereal varieties. In particular, there are different ripening times for varieties of oats (Nordli, 2002). To overcome this problem, we use overlapping diaries here from western Norway and glaciological data reconstructed from established moraine chronologies at two different glaciers in southern Norway.

2. DATA

Five types of data are used here: (1) modern instrumental, meteorological observations, used for calibration purposes; (2) monthly mean sea-level pressure (MSLP), used for the generation of circulation indices; (3) first day of grain harvest, used as proxies for spring–summer temperature; (4) mass-balance measurements on glaciers, used as climate indicators; (5) dated moraines in front of glaciers. Each data type is briefly described below.

2.1. Instrumental temperature series

The temperature series for calibration of the grain harvest data are chosen on the basis of several criteria. The most important ones are the length and the homogeneity of the series. The distances from the series to the farms and to the glaciers are also essential. The temperature series from Bergen turns out to be the best choice.

The Bergen series consists of several individual series that were homogenized by Birkeland (1928) since the start in 1818, and later by Nordli (1997) since 1868. In the early parts of the series little is known about the time of observation and the instrumentation, and there are many gaps. For calibration purposes, only data later than 1867 are used. Because of the unique topography of the town and the adjustment of the series with data from rural stations, the urban influence on the series is estimated to be 0.2 °C or less (Nordli, 1997).

2.2. Pressure data sets

The MSLP data at the Climatic Research Unit (CRU), University of East Anglia, UK, was used for the years 1780–1995. This data set is solely based on pressure observations. The data set has a quality code for each reconstructed grid point (Jones *et al.*, 1999) and the homogenization techniques used are discussed by Slonosky *et al.* (1999).

The data were updated for another 5 years, 1996–2000, by using the gridded data set for the Northern Hemisphere (1873–2000). This data set can be downloaded from the CRU's homepage (http://www.cru.uea.ac.uk/cru/data/pressure.htm).

2.3. First day of grain harvest

The data are quoted from farmers' diaries originating from western Norway. Their geographical locations are shown in Figure 1 and the length of each series is illustrated in Figure 2. The farms are identified by their names. The first day of grain harvest is used as a proxy for spring–summer temperature. The starting date refers to the earliest ripe cereal; in all cases this is barley.

2.4. Glacial mass-balance data set

The glacier mass-balance data have been provided by the Norwegian Water and Energy Directorate (Noregs Vassdrags- og Energidirektorat, NVE). Since 1963, NVE has reported on measurements carried out on Norwegian glaciers in an annual report series 'Glaciological investigations in Norway'. The measurements are also available as files comprising the measurements through time for each individual glacier.

The studies of mass balance include measurements of accumulated snow (winter balance or accumulation) and measurements of snow and ice removed by melting (summer balance or ablation) (Kjøllmoen, 2001). From these data the equilibrium line altitude (ELA) can be calculated. The ELA marks the area on the glacier where accumulation is balanced by ablation; e.g. see Nesje and Dahl (2000). Normally, it is an average value for the whole glacier at the end of the ablation season.

2.5. System of dated moraines

Nigardsbreen, which is an outlet glacier of the Jostedalsbreen ice cap (Figure 1), has five marginal moraine systems that are dated by historical information (Andersen and Sollid, 1971; Bickerton and Matthews, 1993). This has enabled the construction of a lichenometric dating curve, which has made it possible to date five additional moraine ridges. The age determinations of the moraines in front of Storbreen are based on Matthews (1977).

3. METHODS OF TEMPERATURE RECONSTRUCTION IN WESTERN NORWAY: A BRIEF OUTLINE

The methods used for the reconstruction of a temperature series for western Norway (Vestlandet) differ according to the data that are available. Our aim is a composite series consisting of proxy and instrumental
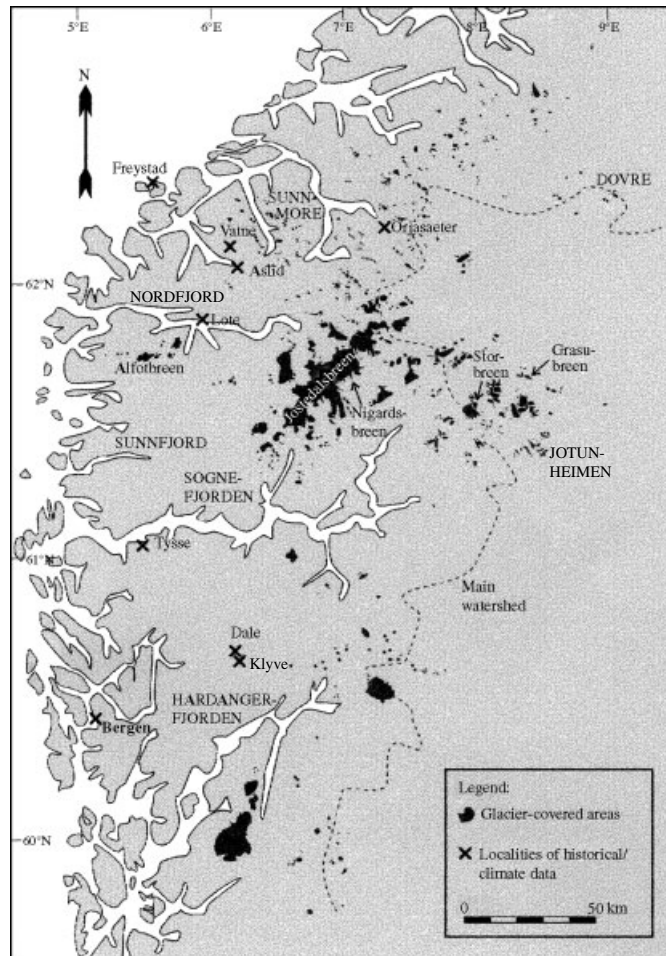
Figure 1. Location of the farmers' diaries in western Norway, the glaciers in the area, and the old temperature series of Bergen
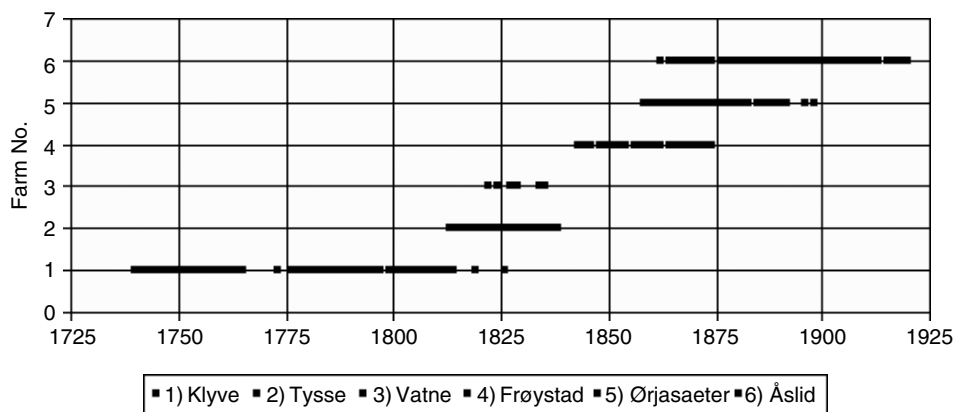

Figure 2. Data coverage of farmers' diaries in western Norway ordered by start year

data. During the period 1868–2003 there exists a homogenized, instrumental series from Bergen, and there is no need for any reconstruction from proxy data (Figure 3, period I).

Scheme for reconstruction of the composite Vestlandet series

Period I: 1868 – 2003:
Instrumental observations from Bergen, 1868 - 2003

Period II: 1843 – 1867:
Data:
- Spring/summer temperature, modern instrumental observations, 1868 - 2003
- Starting date of grain harvest quoted from farmers' diaries

Method:
- Simple regression analysis

Output:
- Spring/summer temperature reconstruction, 1843 - 1867

Period III: 1734 - 1842
Data:
- Spring/summer temperature, modern instrumental observations, 1868 - 2003
- Circulation indices
- ELA based on mass balance measurements

Method:
- Stepwise multiple regression analysis

Output:
- Regression model that can be used on historical data

Model input data:
- Starting date of grain harvest quoted from farmers' diaries
- ELA assessed from dated moraines
- Circulation indices

Model Output:
- Reconstructed spring/summer temperatures, 1734 - 1842
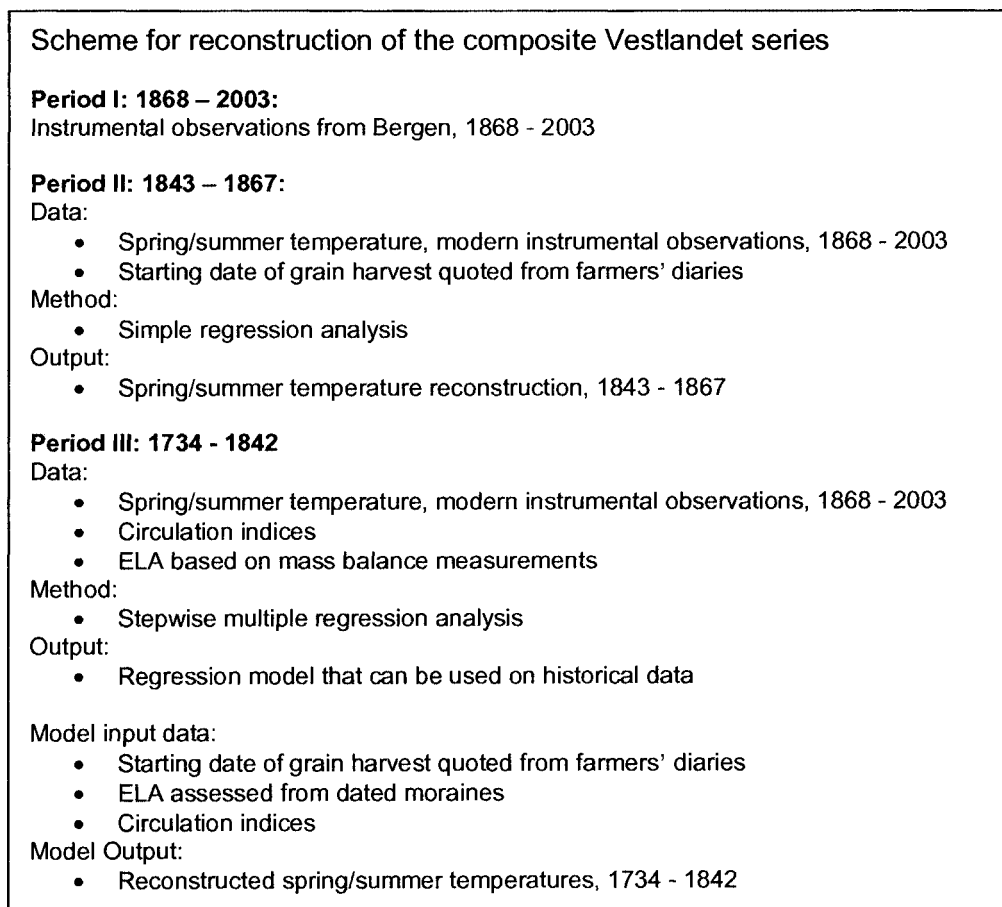

Figure 3. Scheme for reconstructing temperature for western Norway (Vestlandet)

During the period 1868–1923 the series from the Ørjasæter and Åslid farms (Figure 2) overlap the instrumental observations. Simple linear regression analysis is used to reconstruct spring–summer temperature for the period 1843–1867 (Figure 3, period II).

Before 1843 there are no grain harvest data that overlap the instrumental observations. Because of different conditions for cultivation on the farms and different cereal varieties, the regression equations derived for period II cannot be applied directly to the data from the other farms covering the period 1734–1842. In order to overcome the lack of overlapping grain harvest data, additional data are required, as shown in Figure 3, period III.

4. MODELLING SPRING–SUMMER TEMPERATURE FROM GRAIN HARVEST DATA

The first date of rye harvest from Estonia and Finland has previously been used as a proxy for spring–summer temperature (Tarand and Kuiv, 1994; Tarand and Nordli, 2001). Later Nordli (2001a,b) utilized data from southeastern Norway, and from Møre, Dovre and Trøndelag. Not only was rye used, but also barley and oats, which have been (and still are) the most commonly cultivated cereals in Norway. The method used was linear regression analysis.

In order to reconstruct a temperature series that is valid for the climatic region of western Norway (Vestlandet), only data within this region were used (Figure 1). The spring–summer period was defined

as the interval of consecutive months that gives the highest correlations between temperature and harvest dates. In Vestlandet this period consists of the months April–August, whereas in the more northerly districts of Møre, Dovre and Trøndelag this period is May–August, due to the later start of the growing season.

Only harvest data from the two farms Ørjasæter and Åslid overlap modern instrumental observations; consequently, only these farms were used in the regression analysis. After having tested some other possibilities (Nordli *et al.*, 2002), the modern part of the instrumental Bergen series was chosen for the reconstruction. The modern part of the series is considered to start in 1868. The year before, the series was integrated into the network of the Norwegian Meteorological Institute and this improved the quality of the series.

The regression correlations (R) with spring–summer temperature are 0.94 and 0.90 for Åslid and Ørjasæter respectively, implying that 88% and 81% of the variance is accounted for by the regression model (Table I).

In calculating the root-mean-square error (RMSE) of the residuals, a technique called leave-one-out cross-validation was used. The residuals were treated case by case, and different regression equations were used each time. When a residual for a case was derived, that case was deleted from the data, and the regression was based on the remaining $N - 1$ cases. This procedure was repeated for each residual in turn. Thus, the case that was the subject for validation had no influence on the regression used for calculation of that residual. The cross-validated RMSE for Åslid and Ørjasæter are 0.29 °C and 0.39 °C respectively.

The RMSE values show that for individual years the proxy method is less accurate than the 19th century thermometers. However, under the assumption of randomness, the standard error of the mean value of a decade reduces by a factor $1/10^{1/2}$, i.e. less than 0.2 °C, which might also easily occur in instrumental observations if they are not well calibrated for zero-point displacement (Middleton, 1966).

The overlapping periods are long enough to allow grouping of the data into calibration and validation data sets. The first halves of the data were used as calibration data sets and the second halves as validation data sets. For Åslid, the slope coefficients in the two periods were nearly equal, whereas for Ørjasæter the slope was 17% larger in the first period than in the second period. When the equations derived in the calibration data sets were applied on the data of the validation data sets, the bias of the mean values was about 0.2 °C for both series.

The regression equation may be written as

$$T_i = \beta(D_i - D_m) + \Delta T \tag{1}$$

where T_i is the reconstructed spring–summer temperature and D_i is the start of the grain harvest (day no.); the index i represents the year. D_m (day no.) is the mean harvest date for the data period of the farm. The variability of the regression is given by the first term in the equation, whereas ΔT represents the constant term. For convenience, the constant β is hereafter called the temperature *response factor* and ΔT is called the *temperature level*.

Before 1858 there is no harvest series that overlaps the high-quality Bergen series with a sufficient number of years, and regression analysis cannot be used directly. However, regression analysis from different farms indicates that the response factor is about -0.06 °C day^{-1} (Table I). Two farms in southeastern Norway, Hverven and Kollsrud, show response factors of -0.072 °C day^{-1} and -0.064 °C day^{-1} respectively. A response factor of -0.06 °C day^{-1} is used for the data from the farm Frøystad covering the period 1843–57 of the temperature reconstruction. The temperature level for the Frøystad series is calculated by comparison with the overlapping years with the Ørjasæter and Åslid farms.

Table I. Regression analyses between the start date of grain harvest (predictor) and mean spring–summer temperature April–August at Bergen (predictand)

Farm	Regression coefficient	Regression constant	R	Residual cross-validation	No.
Ørjasæter	-0.0542 ± 0.005	24.0 ± 1.3	0.90	0.39	26
Åslid	-0.0621 ± 0.003	26.3 ± 0.8	0.94	0.29	54

5. MODELLING ELA BY CIRCULATION INDICES AND SPRING–SUMMER TEMPERATURE

The ELA has been widely used to infer past and present climatic conditions, and is regarded as the most useful parameter for quantifying climatic effects on glaciers (Porter, 1975; Sutherland, 1984). It marks the line on the glacier where accumulation is balanced by ablation (e.g. Nesje and Dahl, 2000). Normally, it is an average value for the whole glacier noted at the end of any particular monitoring year (end of ablation season). It is sensitive to winter precipitation, wind transport of dry snow, and summer temperature at the glacier. Porter (1977) found that more than 90% of the variance of the ELA was accounted for by summer temperature and winter precipitation. However, surface topography, glacier hypsometry, and aspect can locally complicate this relationship (e.g. Liestøl, 1967; Leonard, 1984; Kuhn *et al.*, 1985; Nesje, 1992; Dahl and Nesje, 1992; Torsnes *et al.*, 1993).

The modelling approach for the ELA was stepwise multiple regression analysis using circulation indices (Chen *et al.*, 1999) and spring–summer temperatures as predictors. The circulation indices were grouped by season, comprising a winter index (October–April) and a summer index (May–August). The spring–summer temperatures (April–August) used in the model were seasonal mean values of the Bergen series.

The circulation indices are inter-correlated, so that parts of the information in one index are repeated in another one. Thus, by using the flow parameters as predictors in multiple regression analysis there is a danger of multi-collinearity and model overfitting. This was prevented by using a stepwise regression procedure that enters or removes variables at each step. An increase in number of predictors was only adopted if its significance was 5% or better according to an F-test. This test may not be sharp enough, and all variables have to pass a tolerance criterion (level: 0.0001) to be included in the regression. Also, a variable is not entered if it would cause the tolerance of another variable already in the model to drop below the tolerance criterion.

The circulation indices are:

$$u = 0.5[p(12) + p(13) - p(4) - p(5)]$$

$$v = 0.499[p(5) + 2p(9) + p(13) - p(4) - 2p(8) - p(12)]$$

$$V = \sqrt{u^2 + v^2}$$

$$\xi_u = 0.529[p(15) + p(16) - p(8) - p(9)] - 0.478[p(8) + p(9) - p(1) - p(2)]$$

$$\xi_v = 0.499[p(6) + 2p(10) + p(14) - p(5) - 2p(9) - p(13) - p(4)$$

$$- 2p(8) - p(12) + p(3) + 2p(7) + p(11)]$$

$$\xi = \xi_u + \xi_v$$

where $p(n)$ is the MSLP at grid point n shown in Figure 4, u and v are the westerly and southerly geostrophic wind components respectively, and V is the resultant geostrophic wind. ξ is the total shear vorticity, and ξ_u and ξ_v are the westerly and southerly shear vorticity components respectively. All indices have units of hectopascals per 10° latitude at 60°N.

The CRU dataset has a quality code running from 1 to 4, where 1 represents the highest quality. During winter the quality code is 1 for all grid points used, whereas in summer the grid point 1 has lower quality before 1850, and grid point 3 has lower quality before 1820. Grid point 2 does not exist in the CRU data set and has been interpolated. The quality is not expected to be any better than for grid point 1. In calculating the geostrophic wind parameters u and v, there is no need to use the grid points of poor data quality. Thus, these parameters are based on quality code 1 during the entire period of the data set, i.e. since 1780.

Mass-balance measurements have been carried out on the Storbreen Glacier (Figure 1) since 1949 (the world's second longest series), whereas similar measurements started on Nigardsbreen in 1962 (e.g. Kjøllmoen, 2001). The results of the regression analyses are shown in Table II. The symbols listed above are completed by an additional (subscript) letter representing the season: 'w' for winter and 's' for summer.

Nigardsbreen: the glacier is an outlet glacier from the Jostedalsbreen ice cap with its main axis oriented towards the southeast. There are only two significant predictors: the straight westerly flow (u_w) during winter
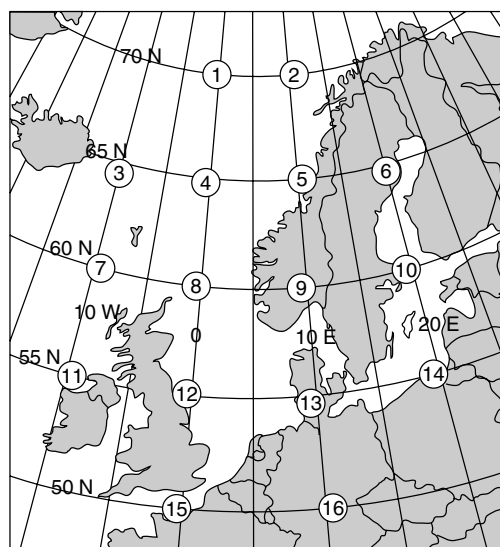
Figure 4. Circulation indices were derived from 16 grid points of MSLP. Their locations are shown on the map

Table II. Regression analysis of the ELA for Nigardsbreen and Storbreen.[a] P = period of regression, N = number of cases, a_0 is the regression constant and a_1, a_2, and a_3 are regression coefficients, R = regression correlation, s = standard deviation of the residuals.

No.	Glacier	P	N	Regression Method[b]	a_0	a_1	a_2	a_3	R	s (m)
Entire data set										
1	Nigardsbreen	62–00	39	S	403.6	$-44.58u_w$	$119.25T$	$-8.90\xi_{u_s}$	0.83	88
2	Storbreen	49–00	52	S	686.4	$111.08T$	$-31.54u_w$		0.82	72
3	Storbreen	49–00	52	E	548.9	$119.23T$	$-27.69u_w$		0.80	75
Calibration and validation data sets										
4	Nigardsbreen	62–81	20	S	1794.1	$-47.65u_s$	$-25.76u_w$		0.77	108
5	Nigardsbreen	82–00	19	S	136.7	$-49.23u_w$	$143.5T$		0.87	82
6	Storbreen	49–74	52	S	239.1	$139.70T$	$-17.34u_w$		0.83	75
7	Storbreen	75–00	27	S	746.7	$115.46T$	$-45.23u_w$	$12.74\xi_{u_s}$	0.87	67

[a] Symbol list of the predictors: u = westerly straight flow, v = southerly straight flow, ξ_u = westerly shear vorticity, ξ_v = southerly shear vorticity, T = April–August temperature in Bergen; subscripts w = winter (October–April) and s = summer (May–August).
[b] S: stepwise entering of predictors; E: direct entry.

that accounts for the accumulation and the Bergen temperatures accounting for the ablation (entry 1 regression parameters in Table II).

Storbreen: the regression equation includes the two terms that were present for Nigardsbreen, plus the cyclonic westerly flow during summer ξ_{u_s}, which has a negative regression coefficient. This type of circulation may occur when a low is present to the north of the area bringing cold air into southern Norway. The flow may be more predominant at higher altitudes and more northerly positions, like Storbreen, than at lower altitudes and more southerly positions, like Bergen; see entry 2 in Table II. This hypothesis is strengthened by replacing spring–summer temperature of Bergen with that of Kjøremsgrende positioned only 70 km away from Storbreen. This leads to a slightly better regression correlation (Nordli *et al.*, 2003) and the vorticity term loses its statistical significance.

The regression correlation for both glaciers is about 0.8, implying that about two-thirds of the variance is accounted for by the regression model. Although significant, the ξ_{us} term (westerly shear vorticity) for

Table III. Modelled ELA by regression analysis using the calibration data set. The calibration data set: 1962–81 for Nigardsbreen and 1949–75 for Storbreen. The validation data set: 1982–2000 for Nigardsbreen and 1976–2000 for Storbreen

	Storbreen ELA (m)		Nigardsbreen ELA (m)	
	Observed	Modelled	Observed	Modelled
N	26	26	19	19
Mean	1751	1752	1446	1474
Standard deviation	115	75	139	97
Minimum	1530	1612	1175	1268
Maximum	1975	1907	1660	1620
RMSE residuals		83		95

Storbreen can be omitted without resulting in a much poorer correlation (compare entries 2 and 3 in Table II). The cross-validated RMSE of the ELA residuals is 88 m and 72 m for Nigardsbreen and Storbreen respectively.

For the Nigardsbreen and Storbreen glaciers the data periods were separated in calibration and validation periods of (near) equal size. The calibration period comprises the years 1962–81 for Nigardsbreen and 1949–75 for Storbreen, whereas the validation periods comprises the years 1982–2000 for Nigardsbreen and 1976–2000 for Storbreen (Table III).

The mean Storbreen ELA in the validation period is successfully modelled by the calibration period. For Nigardsbreen, the mean ELA in the validation period is modelled 28 m too high. This check is important because the westerly flow has been unusually predominant during the late 1980s and the 1990s, leading to a positive mass balance on the maritime glaciers of western Norway, like Nigardsbreen. When using the first half of the series as the calibration period and the last half as the validation period, the regression equations are derived mainly under a weak westerly flow, whereas the verification is performed in a period of frequently strong westerly flow. Despite this, the estimate of the mean ELAs differs by only 28 m at most. The regressions, therefore, appear to be robust concerning changes in westerly flow.

Figure 5 shows the model fit graphically. For most of the years the model performs well for both Nigardsbreen and Storbreen. For Nigardsbreen, it is encouraging that extreme years are well modelled, e.g. the extraordinarily high ELA in 1969 and the extraordinarily low ELA in 1989. Also, for Storbreen, the high ELA in 1969 is well modelled, but not the low ELAs in 1990 and 1995. In particular, the 1990 ELA is badly estimated and is modelled 200 m too high.

The winter precipitation on the western glaciers is linked to the North Atlantic Oscillation (NAO) index (Nesje and Dahl, 2003). This well-known index represents the air flow in the North Atlantic and is positively correlated with precipitation in western Norway. It might, therefore, be argued that the circulation indices could have been replaced by the NAO index in the regression models. However, when this is done, the regression correlation drops from 0.8 to about 0.7 for both Nigardsbreen and Storbreen. This makes the model appreciable poorer, as it accounts for only one-half of the variance, whereas two-thirds of the variance was accounted for by the original model.

The model including circulation indices was adopted for climate reconstruction purposes; see Section 7. The predictors (which were selected by the stepwise regression procedure) could be given physical interpretations and, by splitting the data into calibration and validation data sets, the results indicate that the equations are reasonably stationary.

6. MODELLING ELA FROM MORAINE SEQUENCES

In the middle of the 18th century the Nigardsbreen and Storbreen glaciers reached their maximum LIA extent. At Nigardsbreen, this maximum occurred in AD 1748 (historically dated). Using the accumulation area ratio
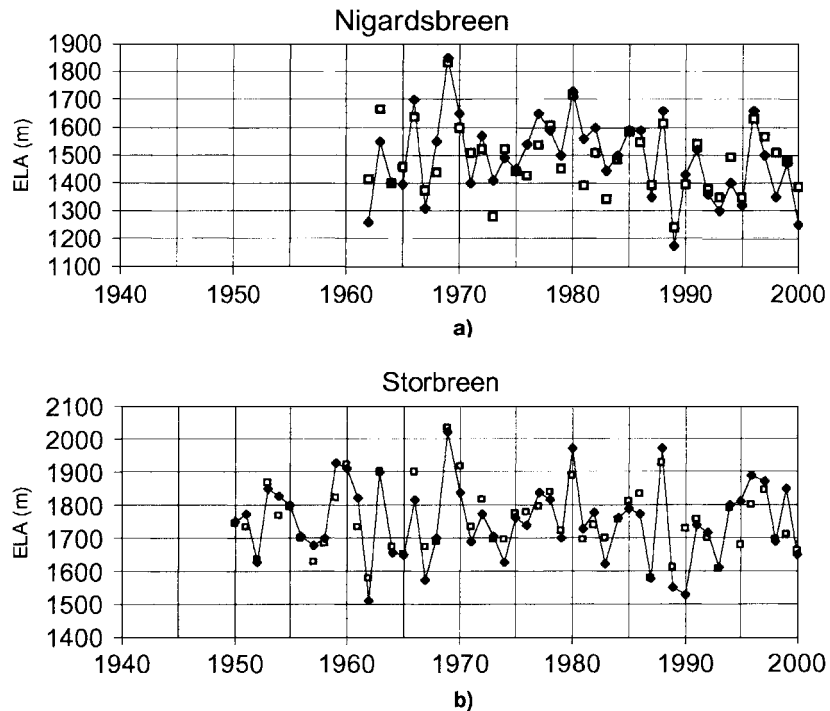
Figure 5. Observed and modelled ELAs for (a) Nigardsbreen and (b) Storbreen

(AAR) method (Meier and Post, 1962; Andrews, 1975; Porter, 1975) on topographically suitable outlet valley glaciers from Jostedalsbreen, the ELA lowering during the LIA maximum extension (LIA$_{max}$) was estimated at about 150 m (Nesje *et al.*, 1991; Nesje and Kvamme, 1991).

The present steady-state ELA represents the altitude at which the net balance on the glaciers is zero. This parameter was calculated by regression analysis using data for the observational periods 1962–2000 for Nigardsbreen and 1949–2000 for Storbreen. The results were 1560 m at Nigardsbreen and 1720 m at Storbreen, as illustrated by the horizontal dotted lines in Figure 6. A lowering of the ELAs of 150 m, corresponding to the situation at the LIA$_{max}$, implies an ELA of 1410 m for Nigardsbreen and 1570 m for Storbreen.

Prior to the LIA$_{max}$, Nigardsbreen advanced *ca* 4000 m relative to the present glacier front (e.g. Grove, 1988). The modern outlet valley glacier at Nigardsbreen slopes gradually away from the central parts of Jostedalsbreen. Just downstream of the present glacier terminus, however, the glacier foreland is almost horizontal. A lowering of the ELA will thus not be correctly reflected in the AAR, despite an increase in ice volume and an advance on the flat valley bottom. The AAR method is thus not suitable for calculating the former ELAs of Nigardsbreen. However, if an advance of 4000 m corresponds to an ELA lowering of 150 m, this can be used to estimate the ELA lowering on Nigardsbreen during the recession phases later than the LIA$_{max}$. A method that is here termed the 'LIR ratio (LR; Dahl *et al.*, 2002) will be used.

Glacier-front variations can be seen as a filtered and delayed signal of the climatic processes determining their mass balance. The delay occurs as variations in net mass balance are transferred by ice dynamics (i.e. sliding and flow) to the terminus. Years of positive mass balance will thus create a front advance, whereas negative mass balance will cause the front to retreat. Owing to the lag of this dynamical response, the response time must be taken into account when calculating the time when the climatic conditions changed the mass balance of the glacier. The response times for Storbreen and Nigardsbreen have been set to 9–15 years and *ca* 20 years respectively.

Glaciers can be seen as being in near steady state when they deposit marginal moraines; thus, good estimates of the ELA can be calculated for these periods using the LR and AAR methods and taking the response time into account. During periods of glacier retreat, however, the mass balance of the glaciers is predominantly
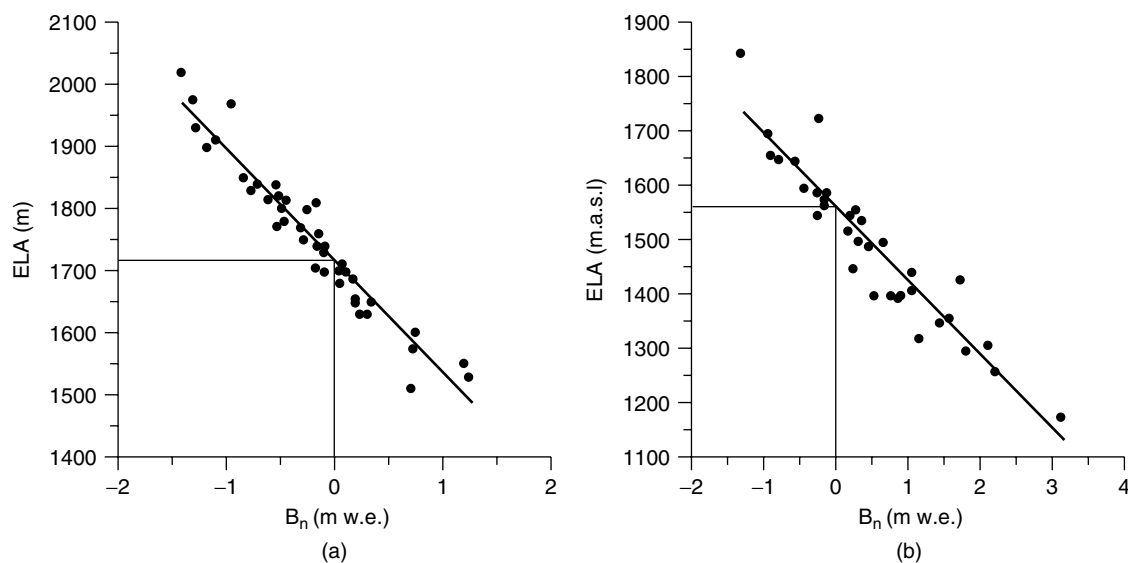
Figure 6. Present steady-state ELAs shown as horizontal lines for (a) Storbreen and (b) Nigardsbreen derived by regression analysis. The net balance is measured in metre water equivalent (m w.e.)
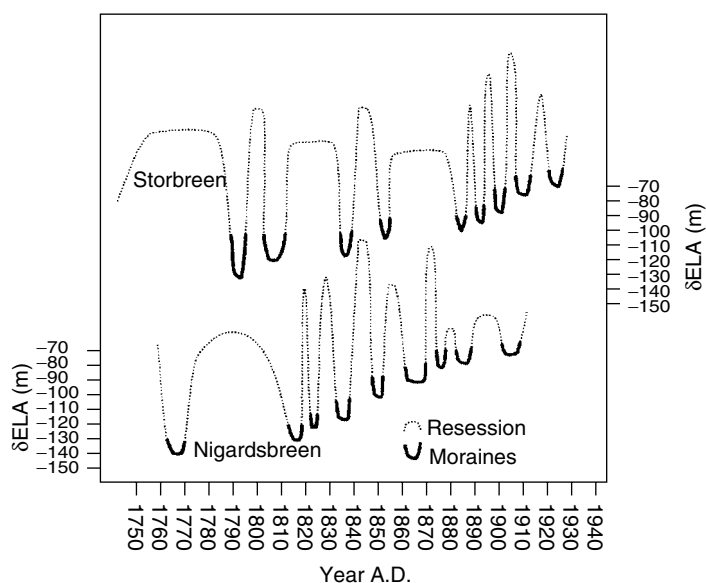

Figure 7. ELA depressions for Nigardsbreen and Storbreen compared with present equilibrium ELAs, for Nigardsbreen of 1560 m and for Storbreen of 1720 m. The dating of the curves is adjusted for the glaciers' response to climate forcing. The dotted curves during the glaciers' recessions indicate less accurate reconstructions than during the time of steady state (solid lines)

negative, and this non-steady-state condition makes the height of the ELA more difficult to quantify. Here, estimates are made of the ELA during the climate depressions causing marginal moraines to be deposited (Figure 7, solid lines), and estimates of the mean recession rate are used to reflect the ELA during phases of retreat (Figure 7, dotted lines).

Nigardsbreen was in a steady state in the following years: 1760–70, 1812–18, 1822–25, 1832–38, 1848–52, 1862–70, 1874–76, 1882–88 and 1900–08. Between 1770 and 1810 no evidence for moraine formation has been found in front of Nigardsbreen. The mean recession rate is, however, only 6.6 m year^{-1},

indicating that the ELA was low for most of the 40 year period. In the periods 1818–21, 1827–31, 1839–47, 1852–61 and 1869–73 the glacier melted back substantially, on average above 15 m year^{-1}.

Storbreen is interpreted to have been in steady state for the periods 1789–95, 1802–11, 1834–39, 1850–55, 1882–86, 1890–93, 1898–1902, 1908–12 and 1920–27. The glacier had a negative mean mass balance in the periods 1750–88, 1796–1802, 1812–33, 1840–49, 1855–81, 1887–90, 1895–99, 1902–07 and 1912–20.

The different behaviour of the glaciers is discussed in some detail by Nordli *et al.* (2003). Whereas Nigardsbreen is strongly exposed to accumulation caused by orographic precipitation during westerly winds, Storbreen has a more sheltered inland position. The shifts between synchronous or asynchronous advances and retreats reflect variations in winter circulation and summer temperature.

7. TEMPERATURE RECONSTRUCTION BY DATA SYNTHESIS

By using the first day of grain harvest as a proxy for spring–summer temperature, a reconstruction could be performed from AD 1843, but the method failed for earlier years due to the lack of overlapping data with instrumental observations (Section 4). In order to perform reconstructions for earlier years, the modelled ELAs for the Nigardsbreen and Storbreen glaciers were also utilized (Section 5). The simplest equations of the ELA (see entries 1 and 3 in Table II) took the form

$$\text{ELA}_i = a_1 u_i + a_2 T_i + a_0 \quad \text{or} \quad T_i = \frac{1}{a_2}(\text{ELA}_i - a_1 u_i - a_0) \tag{2}$$

where less cumbersome symbols are used than in Table II: ELA_i is the equilibrium line altitude in year i; u_i is the straight westerly flow index during winter in year i; T_i is the spring–summer temperature in year i; a_1 and a_2 are the regression coefficients for u and T respectively, and a_0 is the regression constant.

Combining Equation (2) with Equation (1) and rearranging gives

$$\Delta T_k = \frac{1}{a_2}(\text{ELA}_i - a_1 u_i - a_0) - \beta(D_i - D_m) \tag{3}$$

where ΔT_k is the temperature level at farm k (or the regression constant), D_i is the day number for the start of the grain harvest in year i at farm k, D_m is the mean day number for the start of the grain harvest at farm k and β is the temperature response factor, here $\beta = -0.06 \,°\text{C day}^{-1}$.

In the discussion in Section 4, it was shown that the temperature response factor seems to be nearly constant, whereas the temperature level varied from farm to farm. It has been shown to be very difficult to assess the temperature level from the growing conditions at the farms. The existence of cereal varieties with different adaptation to the local climates (Nordli, 2002) is one reason.

The only term in Equation (3) that implicitly includes temperature is the ELA term. This term is also derived from dated moraines, not for individual years, but as mean values for certain periods. Thus, by applying Equation (3) to mean values over selected periods the temperature level can be calculated. Prior to 1843, the grain-harvest data from the farms Klyve and Tysse (Figure 2) and the moraines give two estimated ΔT values for Klyve and four for Tysse. The ELA depression of Nigardsbreen in the 1760s cannot be used because of the absence of relevant circulation data. The average ΔT estimates are 10.8 °C for Klyve and 11.1 °C for Tysse (Table IV).

There are some missing years in the series of harvest data from western Norway. These are interpolated by some early instrumental observations of Bergen or by a temperature reconstruction based on eastern Norwegian diaries adjusted by circulation variations. The interpolations are carefully done to avoid introducing false trends in the western Norwegian reconstruction (for details see Nordli *et al.* (2002)). The missing years are 1766–72, 1774, 1775, 1796, 1798 and 1839–42, i.e. a total of 15 years.

The accuracy of using the moraine data for temperature reconstruction can easily be checked against the modern Bergen instrumental series by using Equation (2). Mean values for the periods of moraine formation were put into the equation and the ELAs modelled were replaced by those derived from the dated moraines.

Table IV. The temperature level ΔT for the farms Klyve and Tysse calculated from Equation (3) by utilizing circulation indices and ELAs derived from the moraines in front of Storbreen and Nigardsbreen

Period	Glacier	ΔT_{Klyve}	ΔT_{Tysse}
1790–95	Storbreen	10.8	
1802–10	Storbreen	10.7	
1813–18	Nigardsbreen		10.2
1822–25	Nigardsbreen		11.3
1830–38	Nigardsbreen		11.0
1835–38	Storbreen		11.9
Mean value		10.8	11.1

Also, the proxy method of overlapping harvest data can be checked against the moraine data for the period 1843–67. There were 10 periods altogether with moraine formation during 1843–1930; eight of these occurred in the modern instrumental period (Table V).

The mean temperature reconstructed by use of the glaciological data does not differ by more than 0.2 °C from the instrumental series. The standard deviation of the difference is 0.3 °C. Also, for the two periods where the reconstruction is based on overlapping diaries (see the first two lines in Table V) there is good agreement with the moraine data. The method of ELA estimation by moraines is thus in agreement with both instrumental observations and reconstructions from overlapping diaries. The method seems to perform well and is expected to give reliable estimates for the temperature levels for the farms Klyve and Tysse. With both the response factor and temperature level known (see Equation (1)), a temperature reconstruction can be obtained back to the first year of the harvest data, AD 1734.

For further analysis of the climate in western Norway the proxy series was chosen for the period 1734–1867, while the Bergen instrumental series was chosen for the period 1868–2003. The whole composite series, 1734–2003, hereafter called the Vestlandet series (means Vestlandet western Norway), is shown in Figure 8 and given in the Appendix. Setting the focus on individual spring–summers, it is clearly seen that the summer of 2002 is by far the warmest one, 14.1 °C. It is remarkable that this summer is as much as 0.8 °C warmer

Table V. Spring–summer temperatures calculated from Equation (2) for selected periods utilizing ELAs derived from glaciers' moraines. The results are compared with the Vestlandet composite series that is reconstructed by overlapping farmers' diaries (1843–67) and instrumental observations (since 1868)

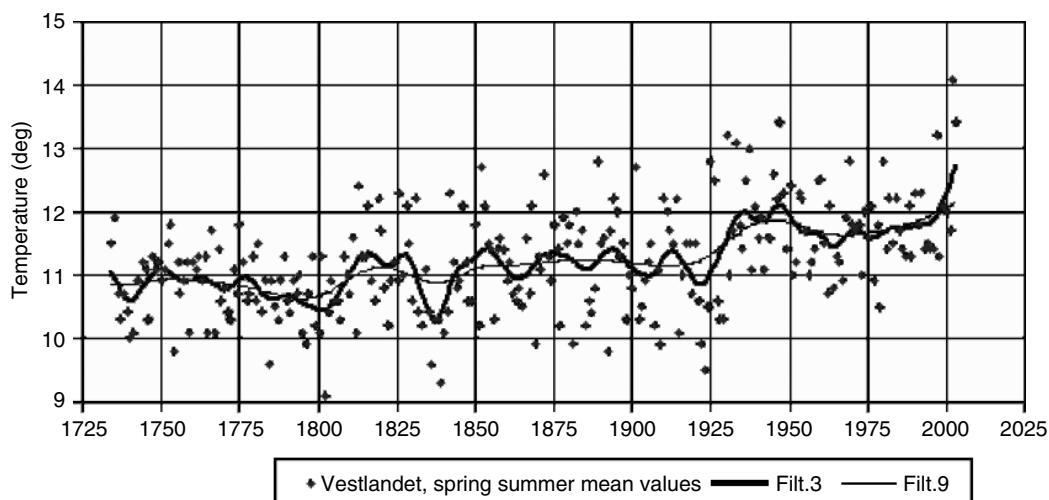| Period	Glacier	By moraines	Composite Vestlandet series
1848–51	Nigardsbreen	10.5	10.8
1862–69	Nigardsbreen	11.1	10.9
1873–76	Nigardsbreen	10.9	11.4
1883–86	Storbreen	10.7	11.0
1890–94	Storbreen	11.0	11.4
1898–1902	Storbreen	11.0	11.0
1882–87	Nigardsbreen	11.2	11.1
1902–08	Nigardsbreen	11.3	10.8
1908–13	Storbreen	10.9	11.4
1920–26	Storbreen	11.2	11.8
Mean value		11.0	11.2

Figure 8. Reconstruction of mean spring–summer temperatures (April–August), 1734–1867, for Vestlandet (western Norway) by proxy data, and since 1868 by instrumental observations. The series is made valid for the currently run station 50540 Bergen–Florida. Individual years are represented as dots (values) in the diagram. The values are filtered by a Gaussian low-pass filter with standard deviations of 3 years (Filt.3) and 9 years (Filt.9) in their distributions

than the next warmest: those of 1947 and 2003. The warmest spring–summers occur mainly later than 1930. Next to the warm spring–summers already mentioned are those of 1930, 1933 and 1937, all of which are warmer than 13 °C.

The spring–summers of 1889 (12.8 °C) and 1735 (11.9 °C) were the warmest in the 19th and 18th centuries (since 1734). Among extraordinarily cold spring–summers, 1802 should be mentioned, it being the coldest one in the whole series. Also, the summer of 1839 was extraordinarily cold. In the 18th century, 1784 seems to be the coldest summer. The low temperatures this year might have been triggered by the Lakagígar eruption (Demarée and Ogilvie, 2001) in Iceland the year before. It should be recalled, however, that in the regression between harvest data and temperature, the standard deviation of the residuals is about 0.5 °C. Sorting of individual years of the reconstructed part of the series might, therefore, be ambiguous. See also the discussion in Section 8 concerning temperature during the last part of the 18th century.

In order to analyse the variations over certain time scales, Gaussian low-pass filters are used with standard deviations in the distributions of 3 years (Filt.3) and 9 years (Filt.9), depressing the variations at time scales smaller than 10 years and 30 years respectively. The main feature of the curves is the increasing spring–summer temperatures during the 270 years represented in the diagram. The increase often seems to have occurred in abrupt shifts, as in the 1810s and 1920s. Since the last shift the temperature has remained high, but it had not increased above the maximum of the 1940s until recently. Analysed on a decadal time scale (Filt.3), the end of the curve has risen above this maximum. The curve has not yet been fixed at the end, and it might be lowered if cool summers are eventually added. But the local maximum during the first years of the 21st century will survive as the highest maximum on the whole curve even if the coming summers turn out to be one standard deviation below the 20th century mean. Another and more likely possibility is that the coming spring–summers will remain high. Then, an abrupt shift in climate may occur like the ones in the 1810s and the 1920s. This is not unexpected when seen from the perspective of global warming, caused by increased amounts of greenhouse gasses.

Three extraordinarily cold periods are shown in Figure 8. The first one occurred around 1740, the next one in the 1800s, and the last one in the 1830s. All of these are well-known famine periods in Norwegian history (Pontoppidan, 1752: 156; Øverland, 1890: vol. 10, 413; Nordli, 2001a). In particular, in mountain villages the grain did not ripen due to cold weather.

For the whole series a linear trend is calculated as 1.2 °C, but a linear model for the temperature increase does not fit the data well. During the last two-thirds of the 18th century that is included in the reconstruction,
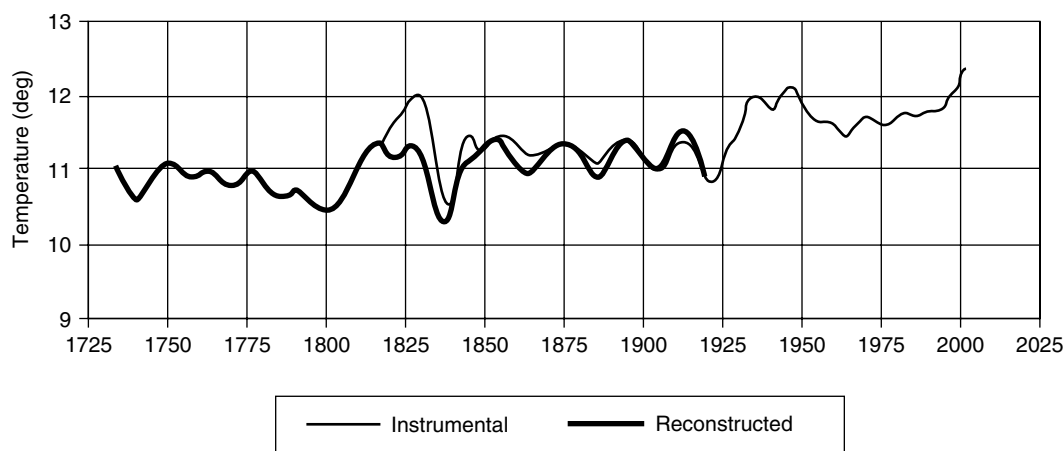
Figure 9. The reconstructed series of Vestlandet and the instrumental series of Bergen. The values are filtered by a Gaussian low-pass filter with standard deviation of 3 years in its distribution

the temperature has decreased by $-0.4\,°C$. This decrease is compensated by an increase through the 19th century, and a further increase of $0.7\,°C$ is present in the 20th century, mainly caused by an abrupt shift that occurred in the 1920s. The significance of the trends was assessed by the non-parametric Mann–Kendall test. The trend of the whole series was statistically significant at the $p = 0.01$ level, as was the trend of the 20th century. However, neither the negative trend of the reconstructed part of the 18th century nor the positive trend trough the 19th century were statistically significant ($p > 0.05$).

The instrumental Bergen series has also been widely used before 1868, when the observations were run by institutions or private individuals outside the Meteorological Institute. Therefore, it is interesting to compare the reconstructed series with the Bergen series before 1868. In Figure 9, the Filt.3 values of the reconstructed series are plotted together with the instrumental Bergen series. The reconstruction seems to fit the instrumental observations quite well later than AD 1840. Prior to 1840, when the instrumental series consists of observations made by teacher Bohr and consul Konow, the reconstructed temperatures are lower than those observed. However, the trend of the instrumental series during this period might not be reliable. Of Bohr's and Konow's observations, only monthly means are known, which makes it very difficult to perform a reliable homogenization of their series.

8. DISCUSSION: COMPARISON WITH OTHER TEMPERATURE SERIES

Available series for comparison are the Swedish instrumental series from Stockholm and Uppsala (Moberg, 1996), the central England series (Parker *et al*., 1992) and a composite series from southeastern Norway (Nordli, 2001b). The Uppsala and central England series start before 1734, the start year of the Vestlandet composite series. The southeastern Norway series (the Austlandet composite series) consists of instrumental observations since 1871 and proxy data for the period 1749–1870. The proxy source is of the same type as used in the Vestlandet series, the first day of grain harvest, which is calibrated against instrumental April–August temperatures. The farms used for reconstruction were located in southeastern Norway or in the bordering Swedish county of Värmland.

A comparison of the Uppsala and Austlandet series is presented by Moberg *et al*. (2003). They suggested that 'the consistently cool summer temperatures before the 1860s indicated by the Austlandet temperature reconstruction for SE Norway, has to be questioned'. The series has also been discussed by Nordli *et al*. (2002), taking into account the different climates of the Vestlandet and Austlandet regions. The temperature difference between the regions was modelled by regression analysis using the same pressure indices as in the present study. They concluded that 'different temperature trends of the two climatic regions do occasionally

occur, but circulation indices for the actual period do not show major differences of the long-term trends. Therefore, this result strongly indicates that the long-term positive trend of the Austlandet series is too strong in the period before 1840'. The Austlandet series should, therefore, be revised if additional diaries can be located.

Recently, the Uppsala and Stockholm series have been tested against homogenized series of cloud cover and circulation indices (Moberg *et al.*, 2003). The indices were similar to those used in this study and also derived from the same data set of gridded MSLP. Based on their tests, Moberg *et al.* (2003) suggested that adjustments should be applied to the Uppsala and Stockholm series. The June and July temperatures should be reduced by 0.7 °C and the May and August temperatures by 0.3 °C for the period before about 1860.

For comparison with the Vestlandet series, the revised version of the Uppsala series was chosen due to its earlier start than the Stockholm series. The Uppsala series correlated well with the Austlandet series ($R = 0.88$) and somewhat poorer with the Vestlandet series ($R = 0.71$) when only the instrumental parts of the series were considered. In the proxy-data parts of the Austlandet and Vestlandet series, the respective correlations fell to 0.61 and 0.43 (Table VI). One reason for the poorer correlation in the proxy-data part was obviously the proxy model itself, with an RMSE of about 0.5 °C in individual years. Inhomogeneities in the proxy series might also have contributed to the reduced correlation.

To resolve this problem, the long-term variations were removed by subtracting the Filt.9 series from the data, thereby eliminating variations over longer time scales than about 30 years. The remaining series reflect variations on shorter time scales, which here are called high-frequency variations. These are shown in the lower part of Table VI. For the instrumental period the high-frequency correlations were equal or somewhat poorer than in the original data, whereas in the proxy-data period the correlations were higher than in the original data. This means that the larger RMSE in the proxy data is not the only reason for the poorer correlation. The proxy parts of the Vestlandet series must also contain different long-term variations or trends compared with the instrumental series.

The series for Vestlandet, Uppsala, and central England are shown in Figure 10 smoothed by Filt.3, reflecting variations on larger time scales than about a decade, and normalized to zero mean during the 20th century. The local maxima and minima of the Vestlandet and Uppsala curves are seen nearly in the same decades, and also the trend in the series seems to be nearly equal from about 1815 to the present. But the Uppsala curve shows a more than 1 °C higher temperature than the Vestlandet series (compared with their 20th century means) during a period of about 50 years prior to about 1815. This contributes strongly to the poorer correlation

Table VI. Cross-correlation table for the series Vestlandet, Austlandet, Uppsala, and central England series shown for the period 1871–2001 (left part of the table) and for the period 1734 (1749 for Austlandet) — 1867 (right part of the table). In the lower part of the table the long-term trend was removed from the data before the correlation coefficients were calculated, whereas in the upper part the whole variability is maintained. In the early periods 1734–1867 and 1749–1870 the Vestlandet and Austlandet series, respectively, are based on proxy data. All other data are instrumental observations

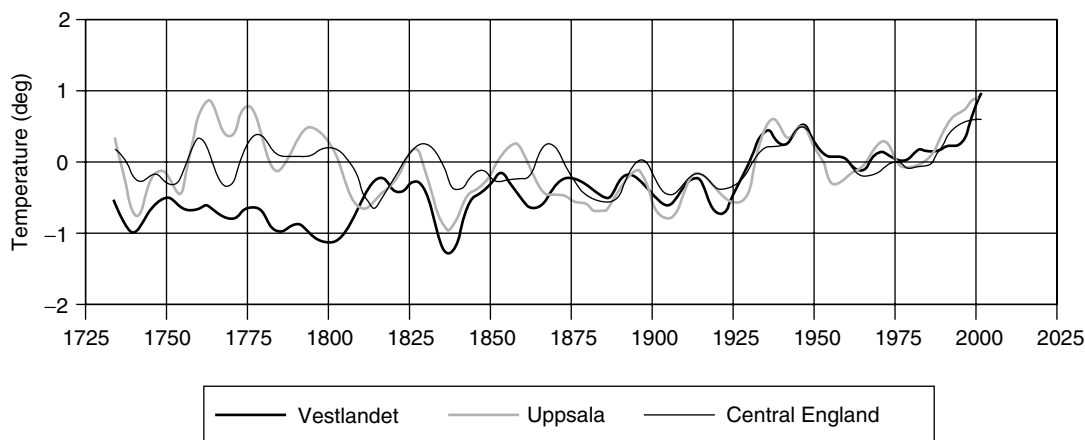| 1871–2001	1734 (1749)–1867			
	Vestlandet	Austlandet	Uppsala	C. England
Vestlandet	—	0.65	0.43	0.25
Austlandet	0.83	—	0.61	0.59
Uppsala	0.71	0.88	—	0.53
C. England	0.49	0.65	0.57	—
High-frequency component 1871–2001	High-frequency component 1734 (1749)–1867			
Vestlandet	—	0.67	0.52	0.32
Austlandet	0.83	—	0.69	0.64
Uppsala	0.67	0.87	—	0.53
C. England	0.41	0.59	0.49	—

Figure 10. The instrumental series of central England and Uppsala and the composite series of Vestlandet. The series are normalized to zero mean during the series normal periods. The values are filtered by a Gaussian low-pass filter with a standard deviation of 3 years in its distribution

in the proxy-data part of the Vestlandet series. In the succeeding period of the proxy series, 1815–67, the high-frequency correlation coefficient between Vestlandet and Uppsala differed very little from that of all frequencies (0.53 versus 0.55). This is consistent within the instrumental part of the series.

It is recalled that between 1814 and 1815 the farm used for the reconstruction of the Vestlandet series shifted from Klyve to Tysse. A too low assessment of the temperature level of the Klyve farm could be the reason for the discrepancy with the Uppsala series (and also with the central England series). The temperature level of Klyve was estimated by two different episodes of moraine formation at the Storbreen glacier and resulted in similar estimates (Table IV).

One source of error lies in the estimation of ELAs. If this is the only reason for the discrepancy of 1 °C, then it requires that the mean ELA was assessed about 120 m too low over the period of moraine formation. This is unlikely, as the ELA should lie well within an error estimate of ±25 m during the LIA. An error of the ELA could, for example, arise by erroneous dating of the moraines. The quality of the lichenometric method used to calculate the age of the moraines has been discussed by Bickerton and Matthews (1993), who proposed error limits of the order of ±20% for the oldest LIA moraines. Thus the age-error of these moraines is less than 50 years. The well-established historical dates of the moraines at Nigardsbreen reduce this error. Another possibility is that the ELA is correctly assessed but that the westerly flow index is too weak in the circulation data set. Although the quality code of the relevant grid points in the MSLP indicates first-quality data, there might exist long-term variations in the flow that are not reflected in the gridded data.

Storbreen is a more continental glacier than Nigardsbreen and accumulates less snow under westerly flow. This is also reflected in the regression model by the regression coefficients (compare entries 1 and 3 in Table II). An increase of 4.3 hPa is required in the pressure difference (between 55 and 65 °N) in order to keep the ELA at the reconstructed level and allow an increase of summer temperature of 1 °C. The moraines that have been used in the assessment of the Klyve temperature level are those formed in the first half of the 1790s and the first decade of the 19th century. It is rather unlikely that an error such as this would be present in the pressure data during several years. In the period between 1770 and 1810, no evidence of moraine formation has been found in front of Nigardsbreen, but a mean recession rate of only 6.6 m year^{-1} indicates that the ELA was low for most of the period.

Moraine formation in front of Storbreen without any moraines in front of Nigardsbreen is a typical reaction of the glaciers during periods of low summer temperature in combination with weak westerly flow (Nordli *et al.*, 2003). If the summer temperature was very low in the first decade around AD 1800, for instance about 1 °C lower than in the period 1760–90, then there must be an inhomogeneity in the Klyve data. This is a possibility that cannot, so far, be tested due to the lack of neighbouring series.

Summing up the discussion above, there are several possible sources of error in the reconstructed spring–summer temperatures, but a large error of 1 °C seems rather unlikely to have occurred as a result of only one source. It is more likely that several sources have been acting in the same direction, resulting in a biased reconstruction during the period of the Klyve proxy data, 1734–1814. A possibility that the Uppsala temperatures are much too high during this period cannot be excluded, but this is considered less likely. High temperatures in the last part of the 18th century are also seen in the central England series (Figure 10) and in many other series on the continent.

In the period 1815–67, where a larger number of moraine formations were available, the agreement with the Uppsala series is excellent. Thus, data from moraine formations in front of western Norwegian glaciers, circulation indices, and harvest data support the use of the adjustments suggested by Moberg *et al.* (2003) for the Uppsala (and Stockholm) series in this period.

We also think that the high summer temperature in the late 18th century needs further study. The Norwegian glaciers remained large during this period. If the summer temperature was higher than at present, then there must have been more accumulation on the glaciers than is modelled. It is possible to go into more detail in the study of weather situations as long as instrumental observations exist. There is also a potential for using proxy data that has not yet been explored. There is also the hope of locating additional western Norwegian diaries containing harvest data. At this stage, the late 18th century paradox of greatly advanced glaciers and high summer temperatures, but without any extraordinary high accumulation, remains unsolved.

9. SUMMARY AND CONCLUSIONS

A series of spring–summer temperatures was reconstructed for the period 1734–1923 for western Norway based on multiple proxy data. In the period 1734–1842, the long-term variations were based on moraines in front of the Nigardsbreen and Storbreen glaciers, whereas annual variations were based on grain-harvest data extracted from farmers' diaries located in western Norway. In the period 1843–67, spring–summer temperatures were reconstructed only by overlapping diaries with instrumental observations. All results were incorporated in one series, 1734–2003, called the Vestlandet composite series.

The reconstruction method using moraine positions was tested against the modern instrumental Bergen series during the periods of moraine formations in front of the glaciers. The agreement with the instrumental series was good, the mean bias being only 0.2 °C. This also strengthens the reliability for the first part of the series where this method was used. The composite series was also compared with the central England series and the Uppsala series. The correlation was strongest with the Uppsala series. Adopting new adjustments for the Uppsala series (Moberg *et al.*, 2003), the Vestlandet series and the Uppsala series exhibit almost equal long-term trends from 1815 to the present.

In the last part of the 18th century, however, the Vestlandet series shows lower temperature than the Uppsala series by about 1 °C. It was not possible to detect any particular reason for this discrepancy, and it was suggested that there might be a combination of several errors that interact to bias the reconstruction in the same direction. A possibility that the Uppsala temperatures are much too high during this period was considered to be less likely, as high temperatures in the last part of the 18th century are also seen in other European series, e.g. the central England series.

Trends in the Vestlandet series were tested by the non-parametric Mann–Kendall test, and a statistically significant positive trend of the whole series was detected. The significance might disappear if the temperatures of the late 18th century have to be adjusted. (For the whole period from 1734 to the present, there were no significant trends in the Stockholm series or in the central England series). The test was also applied to the Vestlandet series for specific centuries. No significant trend was detected during the 18th century (since 1734) or during the 19th century, whereas a positive trend was detected during the 20th century. The temperature trend was mainly caused by an abrupt shift in the series during the 1920s.

Analysis of decadal variations by a Gaussian filter revealed three periods of severe spring–summer temperatures, i.e. around 1740, in the first decade of the 19th century and in the 1830s. These periods are also well known to historians as periods of starvation, during which the use of bark bread became common.

Among individual spring–summers, AD 2002 is remarkable, with a mean temperature 0.8 °C warmer than the second warmest. Even if the 18th century temperatures of the Vestlandet series are adjusted by +1 °C, none of them reach the temperature of 2002. The coldest year in the series is AD 1802, but the error of the reconstruction for individual years is too large for sorting the coldest ones. Other candidates are 1836, 1839 and 1923.

ACKNOWLEDGEMENTS

APPENDIX

The reconstruction of the mean spring–summer (April–August) temperatures for 1734–1867 for Vestlandet (western Norway) by proxy data, and since 1868 by instrumental observations are shown in Table VII. The series is made valid for the currently run station 50540 Bergen–Florida. The values are dated by adding the row number to the column label. The value for AD 1784, for example, is found in row 34 in the column labelled 1750.

Table VII. Reconstruction of mean spring–summer (April–August) temperatures, 1734–1867, for Vestlandet by proxy data since 1868 and instrumental observations (see text)

Row no.	Reconstructed temperature (°C)						
	1700	1750	1800	1850	1900	1950	2000
1		11.1	11.3	10.2	12.7	11.0	11.7
2		11.5	9.1	12.7	10.3	11.2	14.1
3		11.8	10.4	12.1	10.5	12.3	13.4
4		9.8	10.9	11.5	10.9	12.2	
5		11.2	10.6	11.4	11.2	11.7	
6		10.7	10.6	10.3	11.5	11.0	
7		10.9	10.3	11.4	10.2	11.2	
8		11.2	11.3	11.6	11.1	11.4	
9		10.1	11.0	11.4	9.9	12.5	
10		11.2	10.7	10.9	12.2	12.5	
11		11.1	11.6	11.2	12.0	11.5	
12		11.3	10.1	10.7	11.7	10.7	
13		10.9	12.4	10.6	11.5	12.1	
14		11.3	11.3	10.8	12.2	10.8	
15		10.1	11.3	10.5	10.1	11.3	
16		11.7	12.1	11.6	11.0	11.2	
17		10.1	10.9	10.7	11.5	10.9	
18		11.4	10.6	12.1	11.5	11.9	
19		10.6	12.2	9.9	10.7	12.8	
20		10.8	11.7	11.3	11.5	11.8	
21		10.4	10.8	11.1	10.6	11.7	
22		10.3	10.2	12.6	9.9	11.8	
23		11.1	10.9	11.3	9.5	11.0	
24		10.7	11.2	10.9	10.5	12.0	
25		11.8	10.9	11.8	12.8	11.6	
26		11.2	12.3	11.4	12.5	12.1	

(*continued overleaf*)

Table VII. (*Continued*)

Row no.	Reconstructed temperature (°C)						
	1700	1750	1800	1850	1900	1950	2000
27		10.6	11.0	10.2	10.6	10.9	
28		10.7	12.1	11.9	10.3	11.8	
29		11.3	11.5	11.5	10.3	10.5	
30		10.6	10.6	11.8	13.2	12.8	
31		11.5	12.2	9.9	11.0	11.4	
32		10.4	10.4	12.0	11.7	12.2	
33		10.9	10.2	11.5	13.1	11.5	
34	11.5	9.6	11.1	11.7	11.8	12.2	
35	11.9	10.9	10.4	10.2	11.4	11.7	
36	10.7	10.5	9.6	10.6	12.5	11.4	
37	10.3	10.3	10.3	10.4	13.0	11.3	
38	10.7	10.9	10.3	10.8	11.1	12.1	
39	10.4	11.3	9.3	12.8	12.1	11.3	
40	10.0	10.6	10.1	11.5	11.6	12.3	
41	10.1	10.4	10.4	11.6	11.9	11.8	
42	10.9	10.9	12.3	9.8	11.1	12.3	
43	10.9	10.7	11.2	11.7	11.6	11.4	
44	11.2	11.0	10.8	12.2	11.6	11.5	
45	11.1	10.1	10.9	12.0	12.6	11.4	
46	10.3	9.9	12.1	11.3	12.2	11.4	
47	11.3	10.7	11.2	11.5	13.4	13.2	
48	11.1	11.3	10.6	10.3	12.3	11.3	
49	11.2	10.2	10.6	11.0	11.4	12.1	
50	10.9	10.1	11.8	10.8	12.4	12.0	

REFERENCES

Andersen JL, Sollid JL. 1971. Glacial chronology and glacial geomorphology in the marginal zones of the glaciers Midtdalsbreen and Nigardsbreen, south Norway. *Norsk Geografisk Tidsskrift* **25**: 1–38.

Andrews JT. 1975. *Glacial Systems. An Approach to Glaciers and Their Environments*. Duxbury Press. North Scituate.

Bickerton RW, Matthews JA. 1993. 'Little Ice Age' variations of the outlet glacier from the Jostedalen ice-cap, southern Norway: a regional lichenometric-dating study of ice marginal moraine sequences and their climatic significance. *Journal of Quarternary Science* **8**: 45–66.

Birkeland BJ. 1928. Ältere Meteorologische Beobachtungen in Bergen. Luftdruck und Temperatur seit 100 Jahren. Geofysiske Publikasjoner V.

Bradley RS. 1985. *Quaternary Paleoclimatology: Methods of Paleoclimatic Reconstruction*. Allen & Unwin: Boston.

Chen D, Hellström C, Chen Y. 1999. Preliminary analysis and statistical downscaling of monthly temperature in Sweden. C16, SweClim, Department of Physical Geography, Göteborg.

Crowley TJ, North GR. 1991. *Paleoclimatology*. Oxford University Press: New York.

Dahl SO, Nesje A. 1992. Paleoclimatic implications based on equilibrium-line altitude depressions of reconstructed Younger Dryas and Holocene cirque glaciers in inner Nordfjord, western Norway. *Palaeogeography, Palaeoclimatology, Palaeoecology* **94**: 87–97.

Dahl SO, Nesje A, Lie Ø, Fjordheim K, Matthews JA. 2002. Timing, equilibrium-line altitudes and climatic implications of two early-Holocene glacier readvances during the Erdalen Event at Jostedalsbreen, western Norway. *The Holocene* **12**: 17–25.

Demarée GR, Ogilvie AEJ. 2001. Bons Baisers d'Islande: climatic, environmental, and human dimensions impacts of the Lakagígar eruption (1783–1784) in Iceland. In *History and Climate: Memories of the Future?* Kluwer Academic/Plenum Publishers.

Ingram MJ, Underhill DJ, Farmer G. 1981. The use of documentary sources for the study of past climates. In *Climate and History Studies in Past Climates and their Impact on Man*, Wigley TML, Ingram MJ, Farmer G (eds). Cambridge University Press: Cambridge; 180–214.

Grove J. 1988. *The Little Ice Age*. Methuen: London.

Grove JM. 2001. The initiation of the 'Little Ice Age' in regions round the North Atlantic. *Climate Change* **48**: 53–82.

Hoel A, Werenskiold W. 1962. Glaciers and snowfields in Norway. Norsk Polarinstitutt, Skrifter no. 114; 56–79.

Jones PD, Bradley RS. 1992. Climatic variations over the last 500years. In *Climate Since AD 1500*. Bradley RS, Jones PD (eds). Routledge: London; 649–665.

Jones PD, Briffa KR, Barnett TP, Tett SFB. 1998. High-resolution palaeoclimatic records for the last millennium: interpretation, integration and comparison with general circulation model control run temperatures. *The Holocene* **8**: 455–471.

Jones PD, Davis TD, Lister DH, Slonosky V, Jónsson T, Bärring L, Jönsson P, Matheas P, Kolyva-Machera F, Barriendos M, Martin-Vide J, Rodriguez R, Alcoforado MJ, Wanner H, Pfister C, Lutherbacher J, Rickli R, Schuepbach E, Kaas E, Schmith T, Jacobeit J, Beck C. 1999. Monthly mean pressure reconstructions for Europe for the 1780–1995 period. *International Journal of Climatology* **19**: 347–364.

Kjøllmoen B. 2001. Glaciological investigations in Norway in 2000. The Norwegian Water Resources and Energy Directorate (NVE).

Kuhn M, Markl G, Kaser G, Nickus U, Obleitner F, Schneider H. 1985. Fluctuations of climate and mass-balance: different responses of two adjacent glaciers. *Zeitschrift für Gletscherkunde und Glazialgeologie* **21**: 409–416.

Leonard EM. 1989. Climatic change in the Colorado Rocky Mountains: estimates based on modern climate and Late Pleistocene equilibrium lines. *Arctic and Alpine Research* **16**: 245–255.

Le Roy Ladurie E, Baulant M. 1980. Grape harvests from the fifteenth through the nineteenth centuries. *Journal of Interdisciplinary History* **10**: 839–849.

Liestøl O. 1967. Storbreen glacier in Jotunheimen, Norway. Norsk Polarinstitutt Skrifter, no. 141–43.

Mann ME, Bradley RS, Hughes MK. 1999. Northern Hemisphere temperatures during the past millennium: inferences, uncertainties, and limitations. *Geophysical Research Letters* **26**: 759–762.

Matthews JA. 1977. A lichenometric test of the 1750 end-moraine hypothesis: Storbreen gletschervorfeld, southern Norway. *Norsk Geografisk Tidsskrift* **31**: 129–136.

Meier MF, Post AS. 1962. Recent variations in mass net budgets of glaciers in western North America. In *Symposium of Obergurgl, 10–18 September 1962, Variations of the Regime of Existing Glaciers*. IAHS Publication no. 58. IAHS Press: Wallingford; 63–77.

Middleton WEK. 1966. *A History of the Thermometer and its use in Meteorology*. The John Hopkin Press: Baltimore, MD.

Moberg A. 1996. Temperature variations in Sweden since the 18th century. Dissertation no. 5, Department of Physical Geography, Stockholm University.

Moberg A, Alexandersson H, Bergström H, Jones PD. 2003. Were the southern Swedish summer temperatures before 1860 as warm as measured? *International Journal of Climatology* **23**: 1495–1521.

Nesje A. 1992. Topographical effects in the equilibrium-line altitude on glaciers. *GeoJournal* **27**: 383–391.

Nesje A, Dahl SO. 2000. *Glaciers and Environmental Change*. Oxford University Press.

Nesje A, Dahl SO. 2003. The 'Little Ice Age' — only temperature? *The Holocene* **13**: 139–145.

Nesje A, Kvamme M. 1991. Holocene glacier and climatic variations in western Norway: evidence for early Holocene glacier demise and multiple Neoglacial events. *Geology* **19**: 610–612.

Nesje A, Kvamme M, Rye N, Løvlie R. 1991. Holocene glacial and climate history of the Jostedalsbreen region, western Norway; evidence from lake sediments and terrestrial deposits. *Quaternary Science Reviews* **10**: 87–114.

Nordli PØ. 1997. Homogenitetstesting av norske temperaturseriar II [Homogeneity testing of Norwegian temperature series]. DNMI-report, no. 29/97Klima; 1–43 (in Norwegian).

Nordli PØ. 2001a. Reconstruction of nineteenth century summer temperatures in Norway by proxy data from farmers' diaries. *Climatic Change* **48**: 201–218.

Nordli PØ. 2001b. Spring and summer temperatures in south eastern Norway (1749–2000). DNMI-report, no. 01/01Klima.

Nordli PØ. 2002. Old Norwegian cereal varieties used as proxies for temperature. Dei gamle norske kornslaga brukte som indeirekte temperaturdata (bilingual edition). DNMI-klima, report no. 18/02.

Nordli PØ, Lie Ø, Nesje A, Dahl SO. 2002. Spring/summer temperature reconstruction. Western Norway 1734–2001. DNMI-report, report no. 26/02Klima.

Nordli PØ, Lie Ø, Nesje A, Dahl SO. 2003. Glacial fluctuations modelled by circulation indices and spring/summer temperature 1781–2000. In preparation.

Ogilvie A, Jónsson T. 2001. 'Little Ice Age' research: a perspective from Iceland. *Climate Change* **48**: 219–241.

Øverland OA. 1890. *Illustreret Norges Historie [Illustrated History of Norway]*. Folkebladets Forlag. (The volumes were released during the 1890s).

Parker DE, Legg PT, Folland CK 1992. A new daily central England temperature series, 1772–1991. *Journal of Climatology* **12**: 317–342.

Pfister C. 1992. Monthly temperature and precipitation in central Europe 1525–1979: quantifying documentary evidence on weather and its effects. In *Climate since AD 1500*, Bradley RS, Jones PD (eds). Routledge: London; 118–143.

Pontoppidan E. 1752. *Norges naturlige historie [The natural history of Norway]*, Vol. 1. Copenhagen.

Porter SC. 1975. Equilibrium-line altitudes of late Quaternary glaciers in the Southern Alps, New Zealand. *Quaternary Research* **5**: 27–47.

Porter SC. 1977. Present and past glaciation threshold in the Cascade Range, Washington, U.S.A.: topographic and climatic controls, and paleoclimatic implications. *Journal of Glaciology* **18**(78): 101–116.

Rutishauer M. 2001. The Djänkebol farmer diary as proxy data for temperature reconstruction in southwestern Sweden 1761–1861. Degree-Report in Physical Geography, Stockholm University, Degree-Report N-14.

Slonosky VC, Jones PD, Davies TD. 1999. Homogenisation techniques for European monthly mean surface pressure series. *Journal of Climate* **12**: 2658–2672.

Sutherland DG. 1984. Modern glacier characteristics as a basis for inferring former climates with particular reference to the Loch Lomond stadial. *Quaternary Science Reviews* **3**: 291–309.

Tarand A, Kuiv P. 1994. The beginning of the rye harvest — a proxy indicator of summer climate in the Baltic area. *Paleoclimatic Research* **13**: 61–72.

Tarand A, Nordli PØ. 2001. The Tallinn temperature series reconstructed back half a millennium by use of proxy data. *Climatic Change* **48**: 189–199.

Torsnes I, Rye N, Nesje A. 1993. Modern and Little Ice Age equilibrium-line altitudes on outlet glaciers from Jostedalsbreen, western Norway: and evaluation of different approaches to their calculation. *Arctic and Alpine Research* **25**: 106–116.