

# The ‘Little Ice Age’ glacial expansion in western Scandinavia: summer temperature or winter precipitation?

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**Abstract** Reconstructing the temporal and spatial climate development on a seasonal basis during the last few centuries, including the ‘Little Ice Age’, may help us better understand modern-day interplay between natural and anthropogenic climate variability. The conventional view of the climate development during the last millennium has been that it followed a sequence of a Medieval Warm Period, a cool ‘Little Ice Age’ and a warming during the later part of the 19th century and in particular during the late 20th/early 21st centuries. However, recent research has challenged this rather simple sequence of climate development. Up to the present, it has been considered most likely that the ‘Little Ice Age’ glacial expansion in western Scandinavia was due to lower summer temperatures. Data presented here, however, indicate that the main cause of the early 18th century glacial advance in western Scandinavia was mild and humid winters associated with increased precipitation and high snowfall on the glaciers.

**Keywords** “Little Ice Age,” Scandinavian glaciers · Summer temperature · Winter precipitation · North Atlantic oscillation · Sea ice extent

## 1 Introduction

In order to place the recent, apparently anomalous, climate changes in a longer time scale, improved understanding of century to decadal scale natural climate variability on different spatial and temporal scales is of great importance (Wanner et al. 2000; Jones and Mann 2004; Luterbacher et al. 2004; Casty et al. 2005; Moberg et al. 2005a; Xoplaki et al. 2005; Osborn and Briffa 2006). The conventional view of the climate development during the last millennium has been that it followed the sequence of a warm Medieval Period, a cool ‘Little Ice Age’ (LIA) and a warming since the mid-19th century, interrupted by cooling 1940–1975, and continuing thereafter (Lamb 1963, 1965, 1977; Grove and Switsur 1994; Moberg et al. 2005a; IPCC 2007). Research over the last decades has, however, challenged this sequence of climate development in the recent past (Hughes and Diaz 1994; Bradley 2000; Crowley and Lowery 2000). Numerous studies have demonstrated cooler climate and advancing glaciers during the ‘Little Ice Age’ (e.g. Grove 1988, 2004). The ‘Little Ice Age’ term was introduced by Matthes (1939) due to glacial readvance in Sierra Nevada, California, subsequent to the early/mid Holocene Hypsithermal. The term has generally referred to the latest of a series of ‘neoglaciations’ (Porter and Denton 1967). More recently, however, the usage of the term has caused confusion (Matthews and Briffa 2005), depending on whether it is referred to the LIA ‘glacierization’ or ‘climate’, particularly summer temperature. In the European Alps, historic evidence shows that LIA glacial

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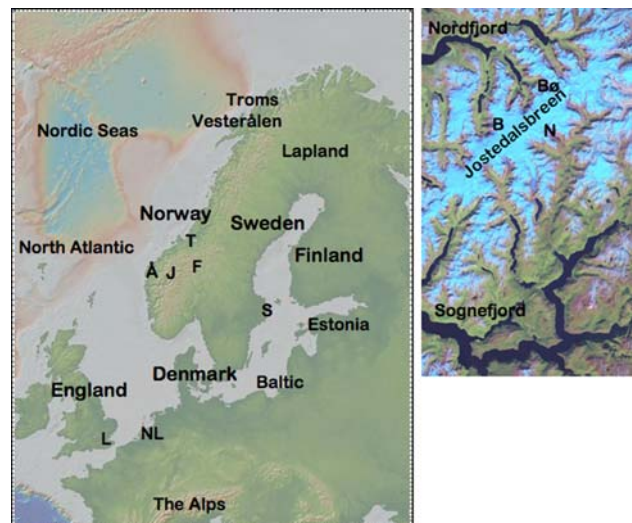
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advances peaked around AD 1350, 1600 and 1850–1860 (Holzhauser et al. 2005). In Scandinavia, on the other hand, most glaciers experienced their maximum LIA position during the mid-18th century (Grove 2004). In Scandinavia, glaciers apparently started the pre-LIA advance during the 14th–16th centuries with a rapid growth in the later part of the 17th and early 18th centuries (Grove 2001), whereas the rapid glacier retreat initiated in the 1930s is commonly considered to terminate the LIA ‘glacierization’ (Oerlemans 2005). Northern Hemisphere and Norwegian temperature records indicate that a LIA ‘climate’ lasted from ~1460 to 1920, however, interrupted by milder periods (Mann and Jones 2003; Nordli et al. 2003). Northern Hemisphere mean annual temperature records for the last millennium (Mann et al. 1999; Crowley and Lowery 2000; Osborn and Briffa 2006) do not show a marked LIA cooling, but rather a gradual temperature decline during the first half of the last millennium.

There were also significant regional temperature variations during the LIA. Several temperature data sets obtained from different archives and proxies (e.g. tree rings, corals, varved sediments, ice cores, glaciers, historical records) show that in the Northern Hemisphere some regions were warm whereas others were cold during the same 30–50 year periods (Jansen et al. 2007: pp. 467–468). Most of the reconstructed climate changes have been linked to external forcing factors (e.g. solar activity, volcanic eruptions) in combination with internal ocean–atmosphere interactions (Crowley 2000; IPCC 2007). The inconsistency between the glacial and climate records over the last millennium has led some commentators to consider that the LIA term is ‘inappropriate’, ‘should be used cautiously’, ‘should be allowed to disappear from use’, or ‘should be avoided because of limited utility’ (Matthews and Briffa 2005 and references therein).

## 2 The ‘Little Ice Age’ glacial expansion in western Norway

The most detailed and historically reliable data on the LIA glaciation in Scandinavia comes from the Jostedalsgreen area in western Norway (Grove 2004) (Fig. 1). There contemporaneous historic evidence (Foss 1750) demonstrates that Nigardsbreen, an eastern outlet glacier from Jostedalsgreen (46°39′N, 8°37′E), advanced 2,800 m between AD 1710 and 1735 (Fig. 2, upper panel), giving a mean annual advance rate of ~110 m (Nesje and Dahl 2003). Between AD 1735 and the historically documented ‘Little Ice Age’ maximum in AD 1748, the glacier advanced 150 m. Thus, between 1710 and 1748 the glacier advanced 2,950 m (mean annual advance of ~80 m). In 2005, a pine (*Pinus sylvestris*) stump with approximately



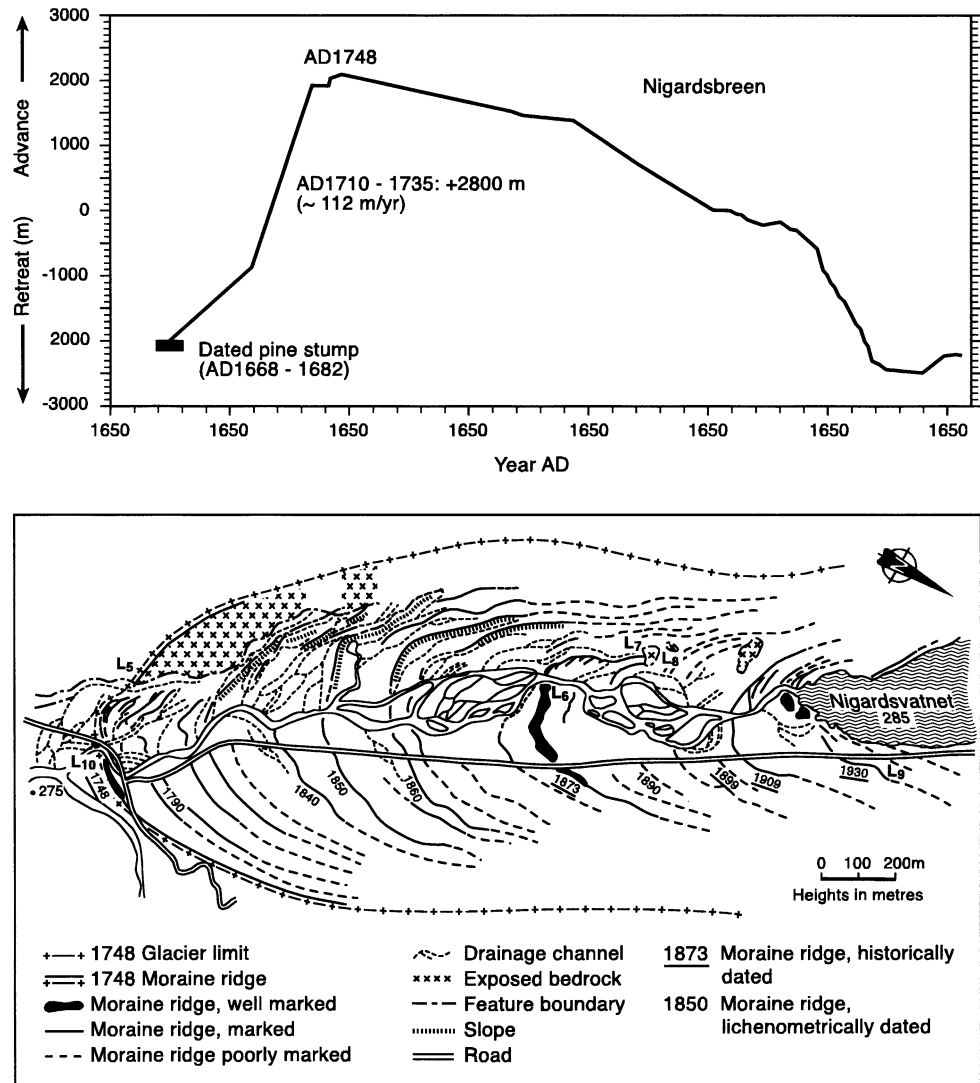
**Fig. 1** Location images. *Left image*: J Jostedalsgreen, T Trøndelag, F Femundsmarka, L London, N The Netherlands, S Stockholm, A Ålfotbreen. *Right image*: B Brenndalsbreen, Bø Bødalsbreen, N Nigardsbreen. *Image*: NASA World Wind

40 annual rings, partly buried in till (in situ) and bearing clear signs of having been overrun by the Nigardsbreen glacier (the subaerial end was splintered), was found at the western end of Lake Nigardsbrevatnet close to the present glacier front. The tree stump was dated at  $175 \pm 25$  radiocarbon years BP (Poz-15936). Age calibration with the CALIB program (Rev5.0.2) (Reimer et al. 2004) yielded several possible age ranges (1 sigma: AD 1668–1682, AD 1736–1782, AD 1797–1804, AD 1935–1951; 2 sigma: AD 1661–1694, AD 1727–1813, AD 1839–1842, AD 1853–1859, AD 1862–1867, AD 1918–1952). Based on the historic evidence, the most likely calibrated age ranges are AD 1668–1682 and AD 1661–1694 with 1 and 2 sigma, respectively (Fig. 2, upper panel). The date gives the approximate age range for when the tree was killed by the advancing Nigardsbreen.

Similar historic evidence comes from Brenndalsbreen at the western part of Jostedalsgreen. Around AD 1650 local farmers said that they saw the glacier as ‘a white cow in the skyline’, meaning that the glacier front was located at the edge of the steep mountain cliff above the present, regenerated glacier. Over a 50-year period during the late 17th/early 18th centuries the glacier advanced ca. 4.5 km, yielding a mean advance rate of  $\sim 90 \text{ m yr}^{-1}$  (e.g. Nesje 1994).

The lichenometrically and historically dated terminal moraines in the Nigardsbreen glacier foreland (Andersen and Sollid 1971; Bickerton and Matthews 1992) demonstrate the rate of retreat after the 1748 LIA maximum (Fig. 2, lower panel). Annual frontal measurements of Nigardsbreen started in AD 1907. No annual frontal measurements were, however, carried out between AD 1964

**Fig. 2** Upper panel historically reported and measured frontal variations of Nigardsbreen. Based on Østrem et al. (1977) with later updates by NVE (Andreassen et al. 2005). The Nigardsbreen glacier record is also based on historically and lichenometrically dated marginal moraines by Andersen and Sollid (1971). Lower panel The glacier foreland of Nigardsbreen (Adapted from Andersen and Sollid 1971)



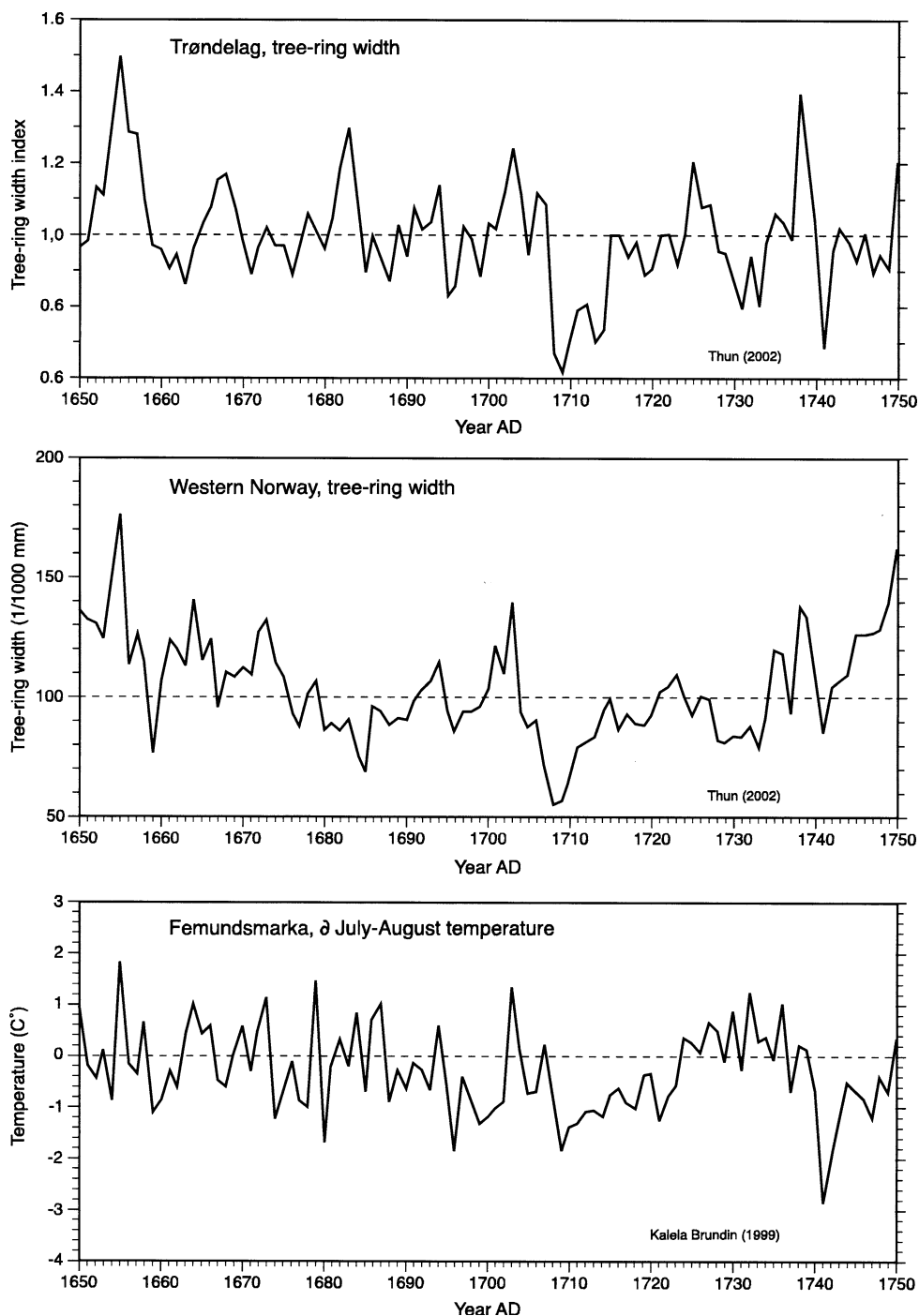
and 1972, but the retreat in this period was photogrammetrically determined to be 515 m (Østrem et al. 1977). The volume of a glacier changes because of variations in mass input (accumulation mainly from atmospheric precipitation) and mass losses (ablation mainly from melting and evaporation) (Dyrgerov and Meier 2000; Lowell 2000). The frontal response to a change in the mass balance is a few years delayed. The frontal time lag at Nigardsbreen has been estimated to 20–25 years (Nesje and Dahl 2003). The question then arises whether the late 17th/early 18th century glacier advances in western Norway were mainly caused by lower summer temperatures and/or higher winter precipitation.

### 3 Early 18th century summer temperatures

Tree-ring width series from Trøndelag and western Norway (Thun 2002) and July–August temperatures reconstructed

from tree rings in Femundsmarka, eastern Norway (Kalela-Brundin 1999) indicate a consistent rise in temperature during the first quarter of the 18th century. The second quarter of the 18th century does not, however, show a consistent rise in the records from Trøndelag and western Norway (Fig. 3). The tree-ring record from western Norway shows a slightly falling trend in the late 17th century. A possible explanation for this falling trend is that the tree-ring material from this period is from buildings of unknown/different origin and/or different age structure of the trees. July–August temperatures in southern Troms and Vesterålen in northern Norway reconstructed from tree rings do not show particularly cold summers during the first part of the 18th century (Kirchhefer 2001). Similarly, tree-ring based reconstructions of April–September temperatures for Northern Europe show relatively mild summers during the first half of the 18th century (Briffa et al. 2002). Reconstructed July mean temperature anomalies going back > 7,000 years based on tree rings in

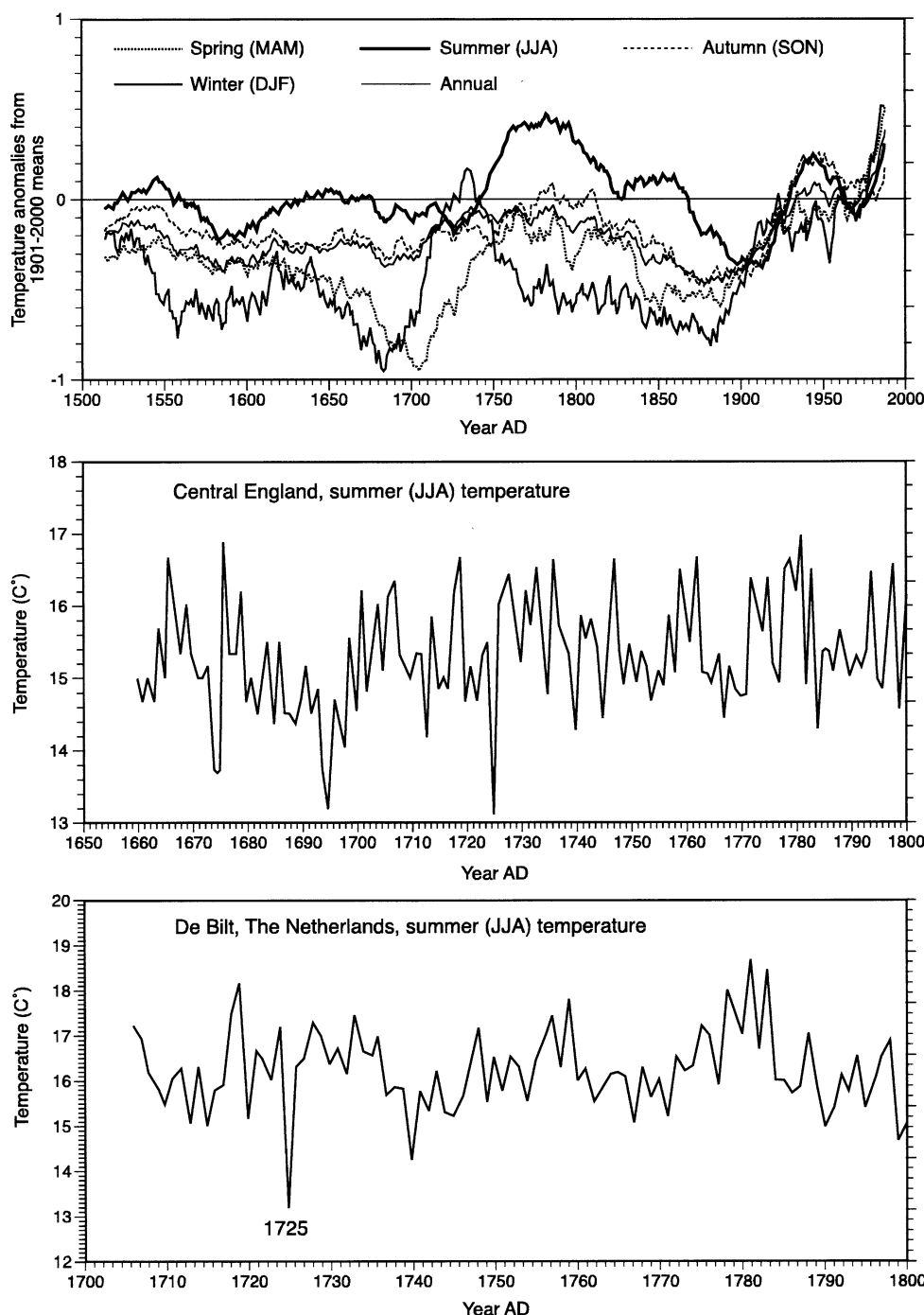
**Fig. 3** Tree-ring width and summer temperature (°C) records from three different regions in southern Norway (Kalela-Brundin 1999; Thun 2002)



Finnish and Swedish Lapland indicates that summers in the 18th century were not particularly cool in northern Scandinavia either (Grudd et al. 2002; Helama et al. 2002). In addition, summer temperatures reconstructed from tree-ring width and maximum density in tree-rings from Lapland, northern Scandinavia do not indicate particularly cold conditions during the first part of the 18th century (Schweingruber et al. 1988). In central Sweden as well, tree-ring data indicate mild summers during the first part of the 18th century (Gunnarson and Linderholm 2002).

Seasonal temperature anomalies in Europe (Luterbacher et al. 2004; Xoplaki et al. 2005) indicate that the summers (JJA) in the last half of the 17th century and first part of the 18th century showed a falling trend (Fig. 4, upper panel). After  $\sim 1730$  the mean summer temperatures over Europe experienced a significantly rising trend. The Central England temperature (CET) series (Manley 1953, 1974; Parker et al. 1992; Parker and Horton 2005) shows that the early 18th century was characterised by relatively high summer temperatures, however, with cold summers in 1725 and

**Fig. 4** Upper panel seasonal and annual temperature anomalies (°C) in Europe (30-year running means) (Adapted from Luterbacher et al. 2004 and Xoplaki et al. 2005). Middle panel a Central England mean summer (JJA) temperature (°C) record for the period 1659–1750 (Manley 1953, 1974; Parker et al. 1992; Parker and Horton 2005). Lower panel mean summer (JJA) temperature (°C) record from De Bilt, The Netherlands (data provided by the Dutch Meteorological Institute: <http://climexp.knmi.nl/getindices.cgi?KNMIData/labrijn+Tdebilt+i+someone@somewhere>)



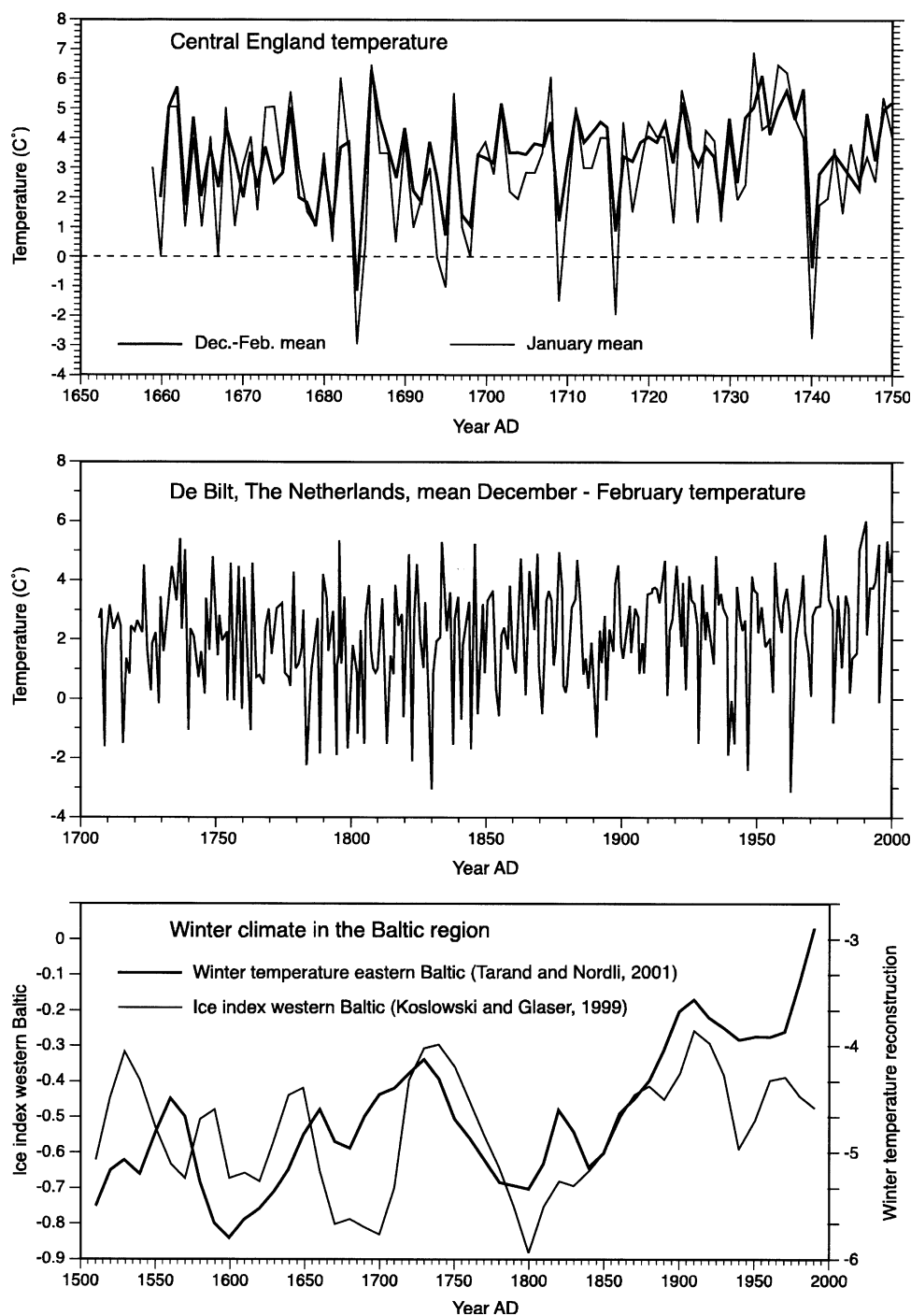
1740 (Fig. 4, middle panel). Cold summer temperatures (AMJJA) in western Norway, three in sequence 1739–1741, are reconstructed by Nordli et al. (2003) and historically documented by Pontoppidan (1752). An instrumental temperature record, starting in AD 1706 from De Bilt, The Netherlands (<http://climexp.knmi.nl>), does not show that the early 18th century summers (JJA) were particularly cold, except for the year 1725 (Fig. 4, lower panel). Finally, the Northern Hemisphere temperature

reconstruction by Osborn and Briffa (2006) indicates a warming trend in the early 18th century.

#### 4 Early 18th century winter climate

Mean winter (DJF) temperatures over Europe showed a markedly rising trend between the 1680s and 1730s (Luterbacher et al. 2004, Fig. 4, upper panel). CET winter

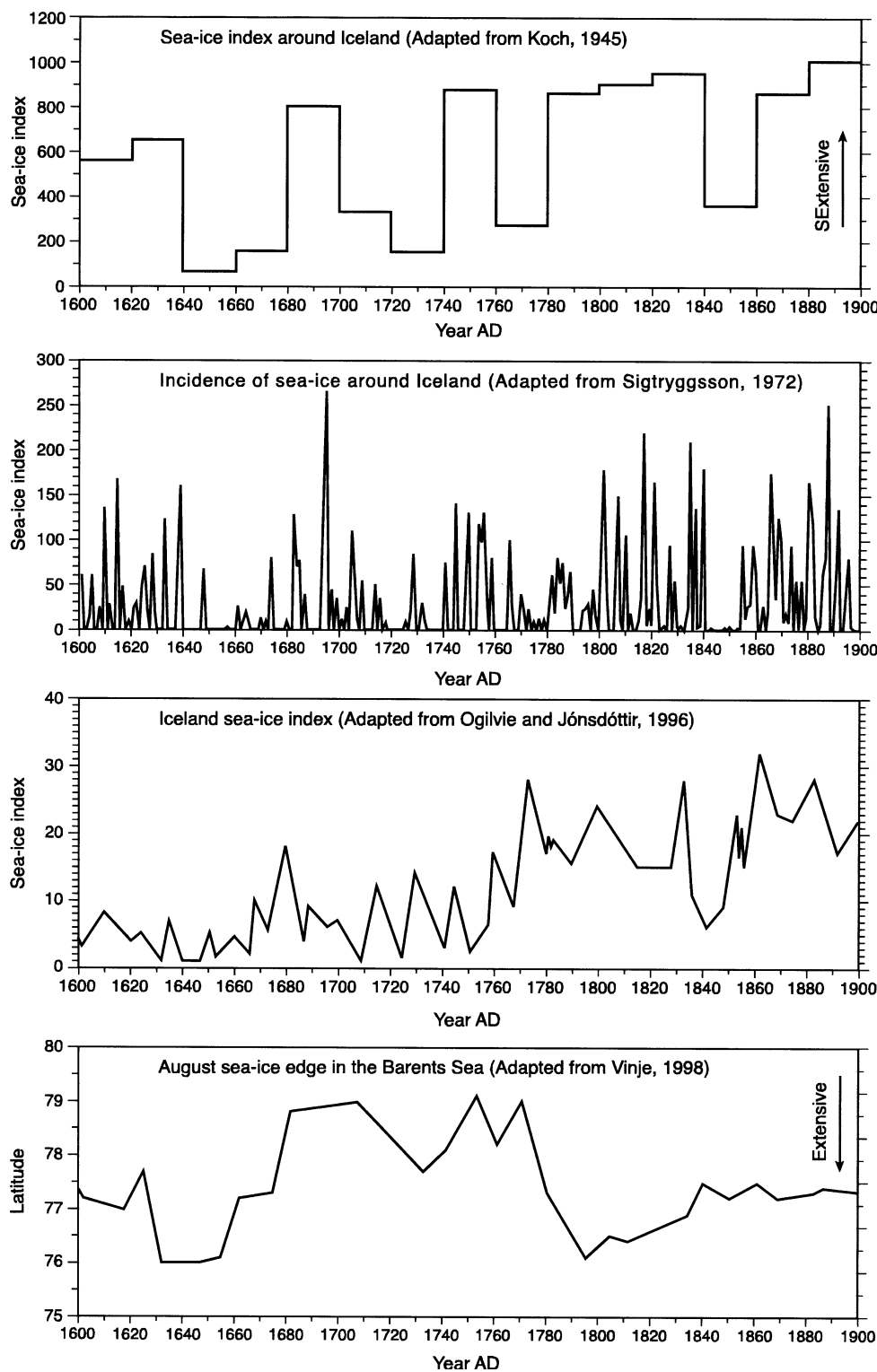
**Fig. 5** *Upper panel* Central England winter (DJF) and January mean temperature ( $^{\circ}\text{C}$ ) records 1659–1750 (Manley 1953, 1974; Parker et al. 1992; Parker and Horton 2005). *Middle panel* mean winter (DJF) temperature ( $^{\circ}\text{C}$ ) record from De Bilt, The Netherlands (data provided by the Dutch Meteorological Institute, KNMI). *Lower panel* Winter climate reconstructions in the Baltic region AD 1510–1990 (Koslowski and Glaser 1999; Tarand and Nordli 2001)



temperatures showed a generally rising trend in the first half of the 18th century, however, with cold winters in 1709, 1716 and 1740 (Fig. 5, upper panel). The instrumental temperature record, starting in AD 1706, from De Bilt, The Netherlands (<http://climexp.knmi.nl>) shows that the winters (DJF) during the early 18th century were generally relatively mild, except for the winters 1709, 1716, and 1740 (Fig. 5, middle panel). Reconstructions of the winter climate in the Baltic region (Koslowski and Glaser 1999; Tarand and

Nordli 2001) shows a positive trend toward milder winters during the first half of the 18th century, the mild winters peaking around the 1730–1740s (Fig. 5, lower panel). Similar mild winters is seen in a reconstruction of late winter/early spring temperatures based on ice break-up of Lake Mälaren near Stockholm in Sweden (Moberg et al. 2005b). A winter–spring thermal index for Iceland, based on analysis of historical documents, also indicates generally mild winters between AD 1600 and 1740 (Ogilvie 1992).

**Fig. 6** Historic sea-ice records from the North Atlantic region (Koch 1945; Sigtryggsson 1972; Ogilvie and Jónsdóttir 1996; Vinje 1998)



**5 Sea-ice extent in the North Atlantic**

Available historic sea-ice records from the North Atlantic (Koch 1945; Sigtryggsson 1972; Ogilvie and Jónsdóttir 1996; Vinje 1998) show that the first half of the 18th century was characterised by less extensive sea ice around

Iceland and in the Barents Sea than in the mid 17th and late 18th centuries (Fig. 6). Forty years (1958–1997) of data were used by Deser et al. (2000) to document Arctic sea-ice variability. The dominant mode of winter (JFM) sea-ice variability exhibits out-of-phase fluctuations between the western and eastern North Atlantic and the time series has a

high winter-to-winter-autocorrelation. The time series is dominated by decadal-scale variations and a long-term trend of decreasing ice cover east of Greenland. The dominant pattern of winter sea ice variability was associated with large-scale changes in surface air temperature and sea-level pressure that closely resemble the North Atlantic Oscillation (NAO).

## 6 The North Atlantic Oscillation

The NAO is one of the major modes of climate variability in the North Atlantic region (e.g. Walker and Bliss 1932; van Loon and Rogers 1978; Kushnir and Wallace 1989; Kushnir 1994; Hurrell 1995; Hurrell and van Loon 1997; Hurrell et al. 2003). The winter (DJFM) temperature and precipitation in Bergen (1861–2006) is highly correlated ( $r = 0.66$ , Fig. 7, upper panel). This relationship holds also for the whole region of western Norway with the mountain range included ( $r = 0.66$ , period 1901–2000). A comparison between the NAO index and winter precipitation between AD 1864 and 1995 in Bergen in western Norway shows that these are highly correlated ( $r = 0.77$ ) (Hurrell 1995). Variations in the NAO index are also reflected in the mass balance records of Scandinavian glaciers (Nesje et al. 2000; Reichert et al. 2001; Six et al. 2001; Nesje 2005), the highest correlation is with winter and net mass balance on the maritime Ålfotbreen in western Norway ( $r = 0.75$  and  $0.66$ , respectively, observation period 1963–2006) (Fig. 7, middle panel). A period of generally high winter precipitation (strongly positive NAO index) between 1988/1989 and 1994/1995 yielded both high winter balance and positive net mass balance on glaciers in western Norway. This resulted in the largest glacier advances in western Norway during the 20th century, and possibly since the early 18th century (Nesje 2005). Reichert et al. (2001) inferred that precipitation is the dominant factor (1.6 times higher than the impact of temperature) for the relationship ( $r = 0.60$ ) between net mass balance on Nigardsbreen and the NAO index (observation period AD 1962–2000). The winter (DJFM) atmospheric temperature in Bergen (western Norway) shows a high correlation ( $r \sim 0.8$ ) with the CET series. The 95 and 99% confidence levels are at 0.27 and 0.35 correlation. This means that the CET, going back to AD 1659, can be used to test whether the significant early 18th century glacier advance in western Norway may have been caused by summer temperature and/or winter precipitation. The cold winters in the late 17th and early 18th centuries when the Dutch canals were frozen (e.g. Grove 1988, 2004) are evident in the CET and De Bilt temperature records (see Fig. 5). The standardised  $[(x - \text{mean})/\text{st.dev.}]$  CET DJFM temperatures show strong coherency ( $r = 0.72$ ) with the winter (DJFM) NAO index

by Jones et al. (1997) with later updates (Osborn 2006) (Fig. 7, lower panel). A climate index was produced to test whether a combination of cold summers and mild winters (normally associated with high precipitation) may explain the late 17th/early 18th glacial expansion in western Norway (Fig. 8, upper panel). The climate index was generated by combining standardised records of JJA and DJF temperatures from the CET series. The standardisation procedure was to subtract the annual values from the mean value and divide by the standard deviation  $[(x - \text{mean})/\text{st.dev.}]$ . The standardised values were given equal weight and summarised. Cold summers/mild and wet winters indicate periods favourable for glacier growth (index above the horizontal line in Fig. 8, upper panel), whereas warm summers/cold and dry winters indicate periods with glacier contraction (index below the horizontal line). The climate index indicates a period favourable for increasing glacier expansion during the late 17th century and three periods favourable for glacial growth during the first half of the 18th century. In addition, periods favourable for glacier growth, as indicated by the climate index produced from the CET record, in the 1770s, 1810s, 1840s, 1850s, 1890s, 1910s, 1920s, and 1990s are represented by marginal moraine formation, based on lichenometric and historic evidence, a few years delayed in the Jostedalbreen area (Bickerton and Matthews 1993; Nesje 2005).

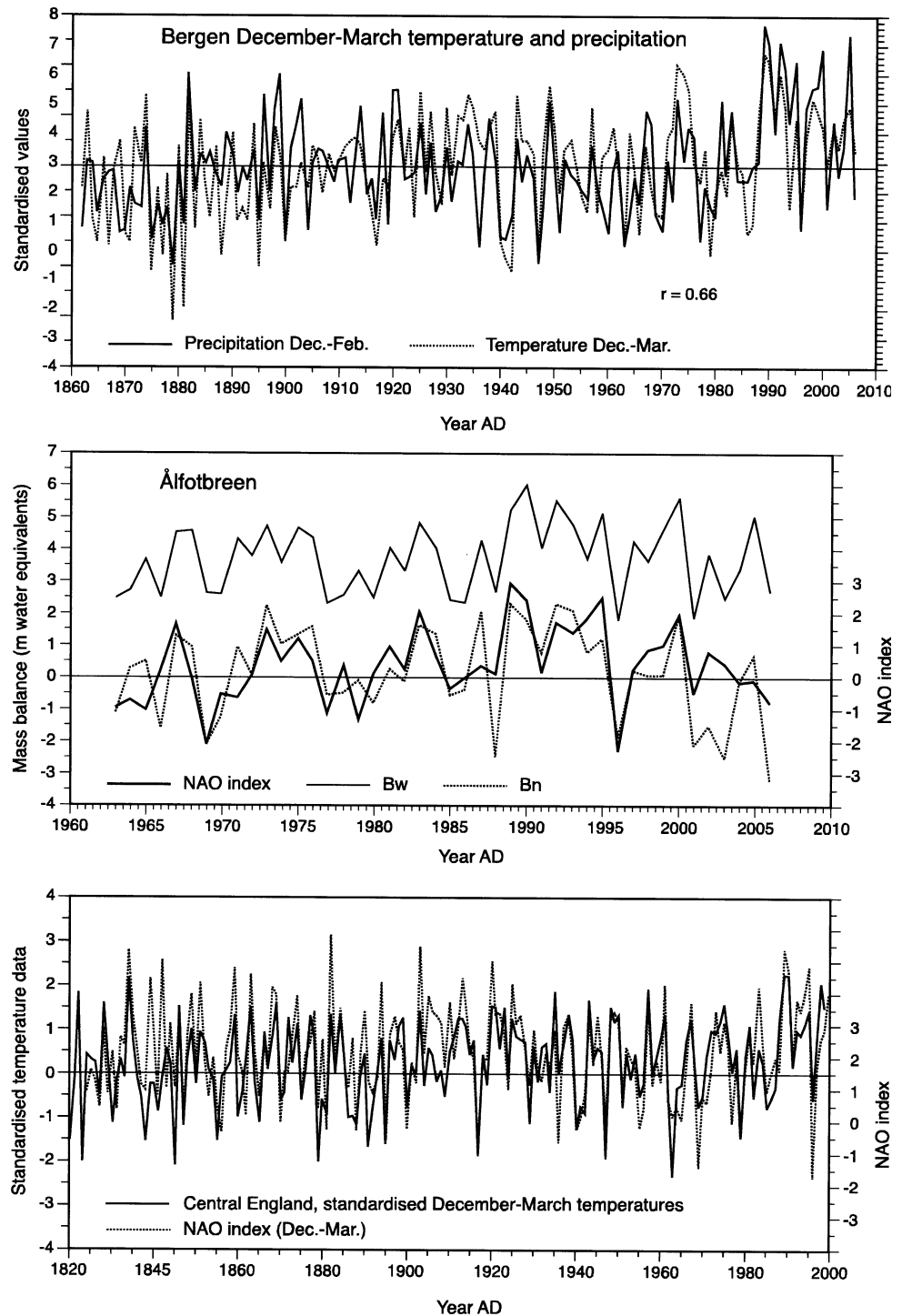
## 7 Discussion

A number of abrupt and widespread climatic variations are recorded around the world during the Holocene (Mayewski et al. 2004). It has been suggested that these changes in the North Atlantic region had a  $\sim 1,500$  ( $1,470 \pm 500$ )-year periodicity (Bond et al. 1997; Campbell et al. 1998). Several Northern Hemisphere temperature reconstructions indicate an irregular temperature decline from AD 1000 to the early 19th century with a subsequent warming trend, especially during the 20th and early 21st centuries (Jansen et al. 2007). Annual to decadal-scale temperature variability were most likely related to variations in solar irradiance and volcanic eruptions (Lean et al. 1995; Mann et al. 1998; Briffa 2000; Crowley 2000). The climatic variations during the LIA caused world-wide growth of glaciers, but the evidence from Scandinavia and the European Alps is best documented (Grove 2004; Holzhauser et al. 2005).

In this paper, it is pointed out that there is only a weak relationship between low summer temperatures in northern Europe and the early 18th century glacial advance in western Norway, whereas the relationship with relatively mild winter temperatures (and hence winter precipitation) is relatively strong. None of the temperature records in Fig. 4 and the most recent tree-ring based summer



**Fig. 7** *Upper panel* standardised [ $(x - \text{mean})/\text{st.dev.}$ ] DJFM temperature and precipitation in Bergen 1861–2006 (Data: met.no). *Middle panel* time series (1963–2006) of winter mass balance ( $B_w$ ) and net mass balance ( $B_n$ ) on Ålfotbreen 1963–2005 (Data: NVE; Andreassen et al. 2005) plotted against the NAO (DJFM) index by Jones et al. (1997) with later updates by Osborn (2006). The correlation between the winter balance on Ålfotbreen and the NAO index is 0.75. *Lower panel* The NAO index (DJFM) by Jones et al. (1997) with later updates (Osborn 2006) plotted against the standardised Central England mean December–March temperatures (Manley 1953, 1974; Parker et al. 1992; Parker and Horton 2005)



temperature reconstructions from the Nordic countries indicate that the early 18th century summers were sufficiently cold to explain the rapid early 18th century glacier advance documented in western Scandinavia. Instead, the records indicate either a general summer temperature rise or not particularly cold summers during the first part of the 18th century. A similar trend is seen in northern Fennoscandian pine chronologies in the first part of the 18th

century (Briffa et al. 1988, 1992; Eronen et al. 1999) and in a Northern Hemisphere (14 chronologies) tree-ring based temperature reconstruction (Esper et al. 2002). A seasonal precipitation reconstruction for European land areas for the period 1500–1900 (Pauling et al. 2005) indicates a period of high winter (DJF) precipitation during the first part of the 18th century. Similarly, European winter temperature records indicate a significant rise in winter temperatures

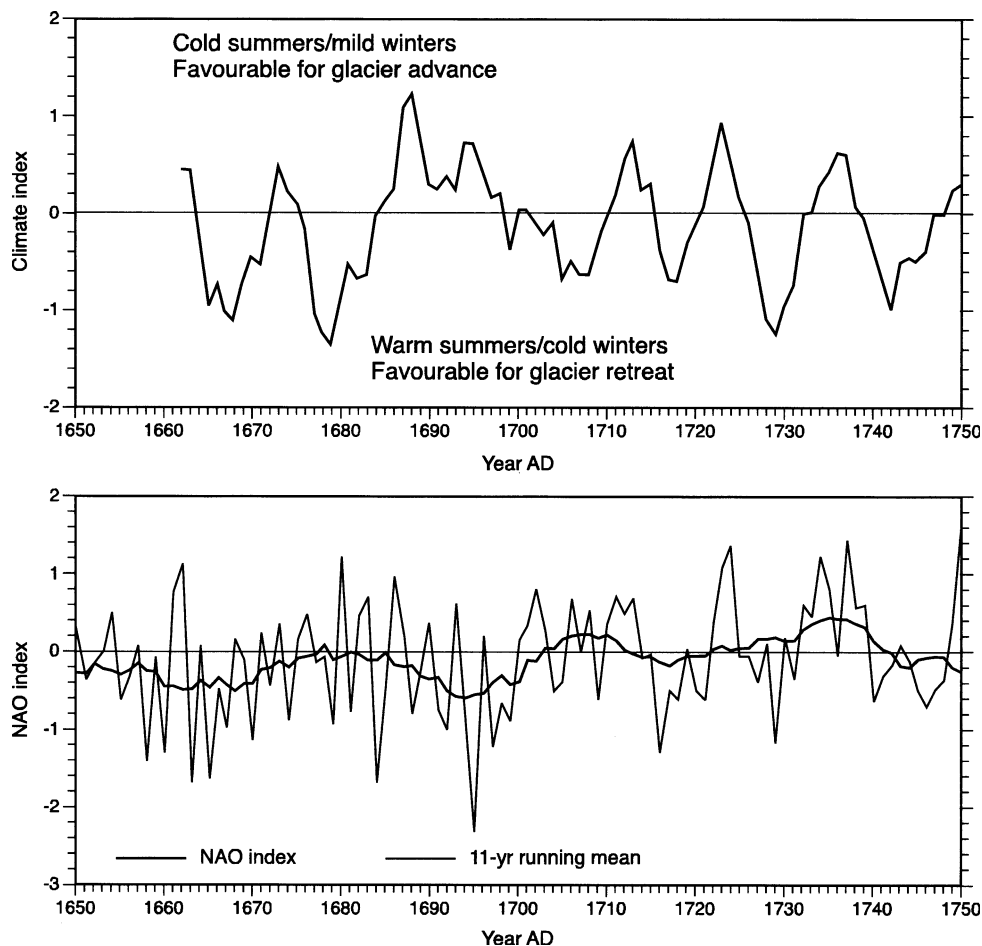
(indicating mild and humid winters and thus a positive NAO-index weather mode) in NW Europe in the first half of the 18th century. A similar pattern is also indicated by a reconstruction of winter (DJF) temperature from three European sites (Central England, Holland and Zürich) for the period AD 1684–1783 (Ingram et al. 1978), reconstructed winter temperatures at De Bilt, Holland, historical records of canal freezing (van den Dool et al. 1978), and a reconstruction of the NAO index back to AD 1429 (Glueck and Stockton 2001) and AD 1500 (Luterbacher et al. 2002). The NAO index of Luterbacher et al. (2002) indicates a positive trend in the NAO index between the mid 1690s and the 1730s (Fig. 8, lower panel). A reconstruction of winter temperature in Tallinn, Estonia, also indicates an increasing winter temperature trend in the southern Baltic region during the first half of the 18th century (Koslowski and Glaser 1999; Tarand and Nordli 2001). This is further supported by an increase in the number of historically reported incidents of major physical hazards related to the autumn, winter, and spring seasons (e.g. snow avalanches and river floods), especially between AD 1650 and 1750, leading to tax reduction for farms in the vicinity to glaciers in western Norway (Grove and Battagel 1983; Grove 1988,

2004; Nesje 1994). A modern ‘analogue’ for such a rapid glacial advance due to increased snow fall and positive glacier mass balance occurred in the early 1990s (Fig. 7, middle panel), when short outlet glaciers at the western side of Jostedalsbreen advanced  $\sim 250$  m over a 10-year period (Nesje 2005).

## 8 Conclusions

Available instrumental and proxy records from NW Europe indicate that summer temperatures alone were not sufficiently low to explain a mean annual glacial advance rate in the order of  $\sim 100 \pm 20$  m over several decades during the late 17th/early 18th century in western Scandinavia. The evidence presented here indicates that the rapid glacier advance that has been historically documented in the early 18th century in western Norway is best explained by increased winter precipitation and thus high snowfall on the glaciers due to prevailing mild and humid winters, most likely as a result of a prevailing positive NAO weather mode during the later part of the 17th century and the first half of the 18th century.

**Fig. 8** Upper panel a climate index produced by combining standardised summer (JJA) and winter (DJF) temperatures from the CET series (Manley 1953, 1974; Parker et al. 1992; Parker and Horton 2005). An index above the horizontal line indicates periods favourable for glacier growth [cold summers/mild winters normally associated with high precipitation (positive NAO index)]. Negative index values indicate periods favourable for glacier retreat [warm summers/cold winters normally associated with low precipitation (negative NAO index)]. Lower panel The NAO index reconstruction AD 1650–1750 (adapted from Luterbacher et al. 2002)



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