

The Holocene

<http://hol.sagepub.com/>

Climate and glacier fluctuations at Jostedalsbreen and Folgefonna, southwestern Norway and in the western Alps from the 'Little Ice Age' until the present: The influence of the North Atlantic Oscillation

Patrizia Imhof, Atle Nesje and Samuel U. Nussbaumer

The Holocene 2012 22: 235 originally published online 24 October 2011

DOI: 10.1177/0959683611414935

The online version of this article can be found at:

<http://hol.sagepub.com/content/22/2/235>

Published by:



<http://www.sagepublications.com>

Additional services and information for *The Holocene* can be found at:

Email Alerts: <http://hol.sagepub.com/cgi/alerts>

Subscriptions: <http://hol.sagepub.com/subscriptions>

Reprints: <http://www.sagepub.com/journalsReprints.nav>

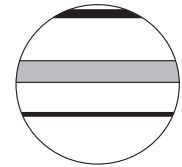
Permissions: <http://www.sagepub.com/journalsPermissions.nav>

Citations: <http://hol.sagepub.com/content/22/2/235.refs.html>

>> [Version of Record](#) - Jan 3, 2012

[Proof](#) - Oct 24, 2011

[What is This?](#)



Climate and glacier fluctuations at Jostedalsbreen and Folgefonna, southwestern Norway and in the western Alps from the 'Little Ice Age' until the present: The influence of the North Atlantic Oscillation

The Holocene
22(2) 235–247
© The Author(s) 2011
Reprints and permission:
sagepub.co.uk/journalsPermissions.nav
DOI: 10.1177/0959683611414935
hol.sagepub.com


Patrizia Imhof,¹ Atle Nesje^{2,3} and Samuel U. Nussbaumer^{1,4}

Abstract

Glacier length, though an indirect and delayed signal of climate conditions, can be used to determine the relationship between climate and glacier response. This study discusses glacier length change of eight outlet glaciers of Jostedalsbreen and Folgefonna (southern Norway), from the 'Little Ice Age' (LIA) until the present. A climate index was calculated from meteorological data from Bergen to determine the specific frontal time lags of the individual glaciers. Short and steep outlet glaciers, such as Briksdalsbreen, react rapidly to changes in climatic conditions, whereas long and gently descending glaciers, such as Nigardsbreen, need longer time to adjust to changes in temperature and/or precipitation. The time lag of Briksdalsbreen was about twice as long during the LIA as today. The correlations between North Atlantic Oscillation (NAO) and climate conditions and glacier fluctuations in Norway and the European western Alps were analysed. As the influence of the NAO on glacier fluctuations is most pronounced during winter, only the winter NAO index was considered. Fluctuations of maritime Norwegian glaciers are highly correlated with the NAO, whereas variations of more continental glaciers in the European western Alps are only partly influenced by the NAO and tend to be anti-correlated. However, the (anti-)correlation with the NAO is not constant during the record, and significantly weaker or even inversed during some periods. A comparison of the LIA glacier fluctuations in southern Norway and the European western Alps suggests that the asynchronous LIA maxima in the two regions may partly be attributed to multidecadal trends in the NAO.

Keywords

climate index, European western Alps, Folgefonna, Jostedalsbreen, North Atlantic Oscillation, time lag

Introduction

During the twentieth century the mean global Earth surface temperature has risen by $\sim 0.6^\circ\text{C}$. Depending on the different CO_2 emission scenarios, global warming may cause an increase in temperature of $1.8\text{--}4.5^\circ\text{C}$ by the end of the twenty-first century (Intergovernmental Panel on Climate Change (IPCC), 2007). When making scenarios for future development of glaciers and to judge whether glacier variations under the current climate lie within natural variability or not, an understanding of past glacier fluctuations is crucial. Glaciers are good climate indicators, as they are sensitive to climate variability and their mass balance and length fluctuations are mainly controlled by climate. Regional climate determines the amount and phase (rain or snow) of precipitation, as well as the solar radiation and air temperature, which are converted into the energy available for melting. Locally, climatic conditions are influenced by the topography and the glacier aspect and elevation. The local climate determines accumulation (e.g. snow, avalanches, rime formation, freezing of rain) and ablation processes (e.g. melting and run-off, evaporation, wind deflation, calving), which are summed up in the net mass balance of a glacier (e.g. Paterson, 1994).

A glacier responds to different climate conditions in a change of size or front position, detectable as glacier advance or retreat.

If the glacier is under balanced conditions (net mass balance = 0), the glacier reflects the current general climate conditions (Paterson, 1994).

Glacier length change is an indirect and delayed signal of a climate perturbation, but much easier to determine than mass balance. Moreover, Oerlemans (2005) showed that data on glacier length from all over the world reflect a distinct global temperature signal. The study confirmed that glacier length can be used as a climate proxy independent of instrumental data and other proxies.

Evidence is becoming stronger of accelerated glacier shrinkage if not even disappearance at a global scale (Haeberli, 2009).

¹University of Bern, Switzerland

²University of Bergen, Norway

³Uni Bjerknes Centre, Norway

⁴University of Zurich, Switzerland

Received 14 February 2011; revised manuscript accepted 17 May 2011

Corresponding author:

Patrizia Imhof, Institute of Geography, University of Bern, Hallerstrasse 12, CH-3012 Bern, Switzerland.
Email: patrizia.imhof@gmail.com

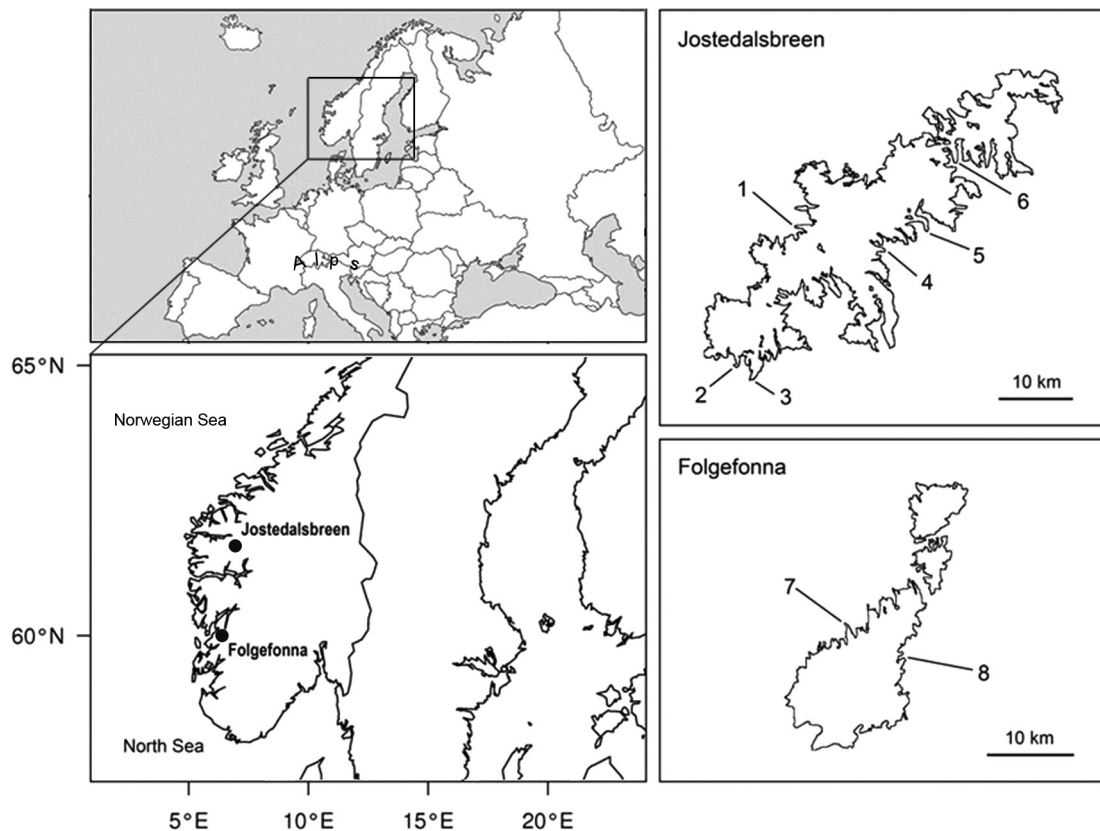


Figure 1. Geographical location of Jostedalsgreen and Folgefonna and the studied outlet glaciers: Briksdalsbreen (1), Bøyabreen (2), Store Supphellebreen (3), Bergsetbreen (4), Nigardsbreen (5), Lodalsbreen (6), Bondhusbreen (7) and Buerbreen (8) (adapted from Nussbaumer et al., 2011)

According to Haeberli and Beniston (1998), glaciers in the Alps have lost about one-third in glacierized surface area and around 50% in ice volume between 1850 and 1980 (see also Paul, 2006; Zemp et al., 2008). They state that the ongoing volume reduction and frontal retreat is mainly the result of warm summers and low winter precipitation. Maritime glaciers in western Norway, on the other hand, experienced positive mass balance and large frontal advance during the 1990s because of increased winter accumulation (Andreassen et al., 2005; Nesje et al., 1995). The same asynchrony between Alpine and Scandinavian glaciers as in the twentieth century (Six et al., 2001) can be traced back to the LIA. At this point, the role of regional weather modes, such as the North Atlantic Oscillation (NAO), has to be considered, as they play a significant role with respect to decadal and multidecadal climate variability (Wanner et al., 2008).

This paper analyses and compares historical glacier fluctuations at Jostedalsgreen and Folgefonna, southwestern Norway, with the focus on the frontal time lag of length changes to climatic variations. The relationship between climate and glacier response is analysed with a climate index calculated from summer temperature and winter precipitation data in Bergen (first in detail for Briksdalsbreen, then also for the other glaciers). For the comparison with Alpine glacier fluctuations Pré-de-Bard Glacier was chosen, for which a newly established glacier length change curve (Imhof, 2010) is first presented here. The fluctuations of the Pré-de-Bard Glacier, together with the newly calculated climate index of Great St Bernard Pass, are compared to the fluctuations of Briksdalsbreen and the climate index of Bergen. Finally, this paper studies the influence of the NAO on glacier fluctuations in southwestern Norway and the western Alps.

Study area and glaciers

This study comprises eight temperate outlet glaciers from the Jostedalsgreen and Folgefonna ice caps, both situated in southwestern Norway (Figure 1). Jostedalsgreen is, with a surface area of 487 km², the largest ice cap of mainland Europe (Østrem et al., 1988). Its highest point is Høgste Breakulen at 1957 m a.s.l. The outlet glaciers studied are Briksdalsbreen, Bøyabreen, Store Supphellebreen, Bergsetbreen, Nigardsbreen and Lodalsbreen.

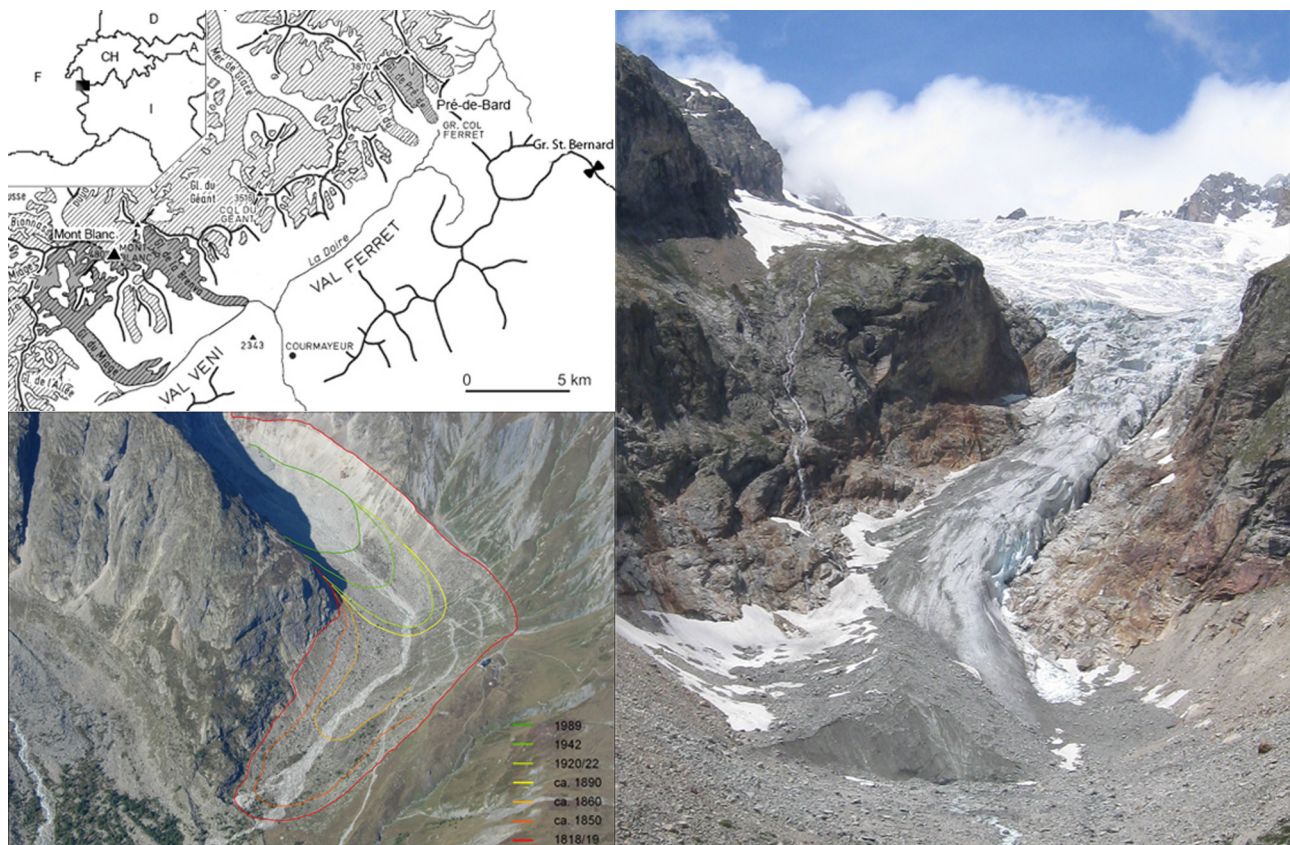
Folgefonna is the third biggest ice mass on mainland Norway. It is divided into three distinct ice fields, of which the southern (Søndre Folgefonna) is the largest (168 km²) (Østrem et al., 1988). The two outlet glaciers of southern Folgefonna studied here are Bondhusbreen in the west and Buerbreen in the east. The characteristics of the individual glaciers are listed in Table 1.

Jostedalsgreen and Folgefonna are maritime glaciers, fed by high rates of snowfall brought by the humid and mild southwesterly winds from the North Atlantic. Owing to their high mass turnover, they react sensitively to changes in the predominant weather and climate conditions. Their net mass balance is mainly controlled by the winter balance (accumulation season) (Andreassen et al., 2005; Nesje et al., 1995). Like most glaciers in southern Scandinavia, the outlet glaciers of Jostedalsgreen attained their LIA maximum position during the mid eighteenth century (e.g. Nesje and Dahl, 2003; Nussbaumer et al., 2011). According to Nussbaumer et al. (2011), the LIA maximum at southern Folgefonna was attained in the late 1870s (with a second peak around 1890).

For the comparison with Alpine glacier fluctuations, the Pré-de-Bard Glacier was chosen. The Pré-de-Bard is a temperate valley glacier located in the Italian Mont Blanc massif, western

Table 1. Overview of the characteristics of the studied glaciers of Jostedalbreen and Folgefonna (tsl, transient snow line. Data from glacier inventory of southern Norway; Østrem et al., 1988), and the timing of their LIA maximum (after Nussbaumer et al., 2011)

	Briksdalsbreen	Bøyabreen	Supphellebreen	Bergsetbreen	Nigardsbreen	Lodalsbreen	Bondhusbreen	Buerbreen
Location	61°65'N/6°92'E	61°29'N/6°46'E	61°52'N/6°8'E	61°65'N/7°03'E	61°72'N/7°13'E	61°78'N/7°24'E	60°03'N/6°33'E	60°02'N/6°4'E
Exposure	W	S	S	E	SE	SSE	NNW	NE
Surface area (km ²)	11.9	13.9	11.8	10.5	48.2	12.2	17.3	15.2
Length (km)	6.0	5.7	8.4	4.8	9.6	6.0	7.8	7.5
Elevation (m a.s.l.)	350–1910	490–1730	720–1730	560–1960	1950–355	860–1960	480–1660	620–1640
tsl (m a.s.l.)	1550	1270	1300	1480	1300	1160	1370	1320
LIA maximum	1760–1765	1740–1750		1743	1748	1750 (?)	1875	1878–1879

**Figure 2.** Top left: Geographical location of the Mont Blanc massif and Pré-de-Bard Glacier, as well as the meteorological station at Great St Bernard pass (adapted from Grove, 2004). Bottom left: Aerial view of the Pré-de-Bard frontal area with approximate glacier outline in 1818/1819, c. 1850, c. 1860, c. 1890, 1920/1922, 1942 and 1989 (Imhof, 2010; Photograph S.U. Nussbaumer, 30 September 2009). Right: Proximal view of glacier snout and ice fall of Pré-de-Bard Glacier (Photograph P. Imhof, 27 June 2009)

Alps (45°54'N, 7°03'E). Its elevation ranges from 3750 to 2081 m a.s.l., with a mean average slope of 37%. The glacier is exposed in a southeastern direction and covers an area of 3.1 km² over a length of 3.4 km (glacier cadastre 2005, adapted from Fondazione Montagna Sicura; Imhof, 2010). The Pré-de-Bard Glacier reached its LIA maximum in 1818/1819 (Imhof, 2010). The location of Pré-de-Bard Glacier and the Great St Bernard weather station (~23 km east of Mont Blanc), used for the calculation of the climate index are indicated in Figure 2.

Data and methods

Climate index

Glaciers are sensitive indicators of climate variability and respond to both summer temperature (ablation season: Norway, 1 May–30 September; Alps, 1 April–30 September) and winter precipitation (accumulation season; Norway, 1 October–30 April; Alps, 1

October–31 March). To examine the glacier–climate relationship, records of winter precipitation (P_w) and summer temperature (T_s) at the Bergen meteorological station (12 m a.s.l.) back to 1862 were used (meteorological data from Climate Division, Meteorological Institute, Oslo). The meteorological record of Bergen is regarded as representative for the climate trends of southwestern Norway (Nesje, 1989). For the length change of a glacier, relative changes in the meteorological conditions are important rather than the absolute values. Thus, a climate index was generated by adding standardized records of P_w and T_s (inversed). For the standardization, the difference between annual and mean values was divided by the standard deviation ($(X - X_{\text{mean}})/\text{std}$). Based on the relative importance of winter balance and summer balance for the net mass balance of Nigardsbreen (Nesje et al., 2000), P_w and T_s were given equal weight. The same procedure was applied to the meteorological data of Great St Bernard Pass (2479 m a.s.l.; data from MeteoSwiss), to calculate a climate index for the

southeastern Mont Blanc region back to 1866. Again, P_w and T_s were given equal weight, as temperature is only slightly more important (52%) for glacier fluctuations of the nearby Mer de Glace than precipitation (Nussbaumer et al., 2007). The climate index indicates periods favourable for glacier advances/positive net mass balance (high P_w , low T_s) and retreats/negative net mass balance (low P_w , high T_s). The climate indices are named BCI (Bergen Climate Index) and GSBCI (Great St Bernard Climate Index).

The frontal time lag

The frontal response of a glacier to changes in climatic conditions is delayed. After a certain 'reaction time' (or time lag), ranging from a few years to several decades, glacier length changes and the glacier finally reaches a new equilibrium after a 'response time' from several years to about 100 years (Haeberli, 1995; Hoelzle et al., 2007). Basically, the reaction and response time of a glacier to a climatic forcing is proportional to its size (Hoelzle et al., 2007), but it also depends on relief, exposition, elevation, etc. (Paterson, 1994). The response time depends also on whether the glacier is advancing or retreating; glacier advances are of a more dynamic character than retreats, which are dominated by ablation and thus more dependent on climate parameters (e.g. Nussbaumer et al., 2007; Schmeits and Oerlemans, 1997).

The glacier length change curves of the eight study glaciers in Norway were reconstructed by Nussbaumer et al. (2011). First, the glacier length change record (annual fluctuations) of Briksdalsbreen was compared with the BCI. The time lag was calculated by a Pearson correlation analysis, and additionally analysed by a simple visual comparison of the events dating in the two records. Briksdalsbreen was chosen because of its short and steep valley tongue, which makes this glacier especially sensitive to changing climate conditions. Moreover, Briksdalsbreen is one of the only glaciers with a continuous record of measured annual front variations going back to AD 1900 (Winkler et al., 2009). In the next step, the other outlet glaciers were compared with Briksdalsbreen. The time lag was determined by comparing the timing of characteristic culmination and inflexion points of the cumulative length record of Briksdalsbreen and the other glaciers.

The role of the North Atlantic Oscillation (NAO) for glacier length changes

The NAO describes interannual variability in the atmospheric and oceanic circulation over the North Atlantic and northwestern Europe. Its manifestation is indicated by the NAO index, based on the difference of normalised sea level pressures between Ponta Delgada, Azores and Stykkishólmur, Iceland. The index is most pronounced over western Norway during winter (Nesje et al., 2000). A positive NAO index is generally associated with stronger-than-normal westerlies, above-normal precipitation and mild temperatures across the North Atlantic and northwestern Europe (Reichert et al., 2001; Wanner et al., 2001). During pronounced negative phases of the NAO, opposite patterns of temperature and precipitation anomalies are typically observed.

One of the aims of this paper is to test the influence of the NAO during the peak winter season (December–March) on climate conditions and glacier length fluctuations in southwestern Norway and the western Alps, respectively. For the comparison, Briksdalsbreen (Jostedalsbreen, southwestern Norway) and Pré-de-Bard

Glacier (Mont Blanc, western Alps) were chosen. To produce a winter NAO index ($NAO_{Dec-Mar}$), monthly mean NAO indices back to 1823 were used (URL: <http://www.cru.uea.ac.uk/cru/data/nao/nao.dat>). For the winter NAO index, the winter season (December–March) instead of accumulation season (October–April) was used, as $NAO_{Dec-Mar}$ is better correlated to glacier fluctuations than $NAO_{Oct-Apr}$ (Nesje et al., 2000). As the NAO and climate indices as well as the glacier fluctuations are highly variable on the annual timescale, the records were smoothed by 5-year running mean to remove the annual variability. The correlations between $NAO_{Dec-Mar}$ and the climate indices of Bergen (observation period 1862–2010) and Great St Bernard (observation period 1866–2009) were tested by Pearson correlation analyses. To analyse the correlation between $NAO_{Dec-Mar}$ and the glacier fluctuations, annual fluctuations shifted for the characteristic time lag to a climate signal (Briksdalsbreen, 3 years; Pré-de-Bard Glacier, 5 years) were used. The observation periods of annual glacier fluctuations were AD 1909–2009 and 1893–2010 for Pré-de-Bard Glacier and Briksdalsbreen, respectively.

Results

The relationship between climate and glacier response – the example of Briksdalsbreen

Extensive meteorological experiments on glaciers have shown that fluctuations in the mass balance, and thus glacier length, are mainly due to temperature during the summer ablation period (T_s) and precipitation during the winter accumulation period (P_w) (Oerlemans, 2005). Nevertheless, it must be noted that also spring and autumn precipitation and temperature may have significant effects and that various combinations of temperature/precipitation characteristics can lead to advances/retreats of a glacier in historical times (Steiner et al., 2008). Figure 3A shows the individual P_w and T_s records of Bergen as well as the BCI. Note that the temperature record has been inverted, such that positive index values correspond to low summer temperature signals. The correlation of the two meteorological parameters is $r=0.42$. A high climate index and thus conditions favourable for a glacier advance are achieved by a combination of high P_w and low T_s values. The meteorological record can be classified into five distinct periods. The first period from 1862 until 1883 was characterized by very low winter precipitation (20% below the long-term mean of 1862–2010) and a corresponding low BCI. The following phase (1884–1930) corresponds to the end of the LIA and includes three BCI peaks, which were mainly due to low summer temperatures. In the transition phase from 1931 until 1970, summer temperatures reached a higher level and were complemented by rather low winter precipitation, leading to a low BCI. The following phase corresponds to the rapid glacier advance in the 1990s as a response to a winter precipitation peak around 1991. The most recent period (since 1995) was characterized by a strong summer temperature rise (10% above the long-term mean of 1862–2010) and a winter precipitation peak around 2006/2007.

In Figure 3B, P_w and T_s are plotted together with the annual front variations of Briksdalsbreen. The glacier was moderately advancing in the periods 1904–1910 and 1922–1930, both intervals after a period of lower summer temperatures. The following retreat phase lasted until 1955, and was especially pronounced in the 1940s. This retreat phase was characterized by high summer temperatures, combined with low P_w in the early 1940s. However, according to Steiner et al. (2008), glacier retreat in this period was

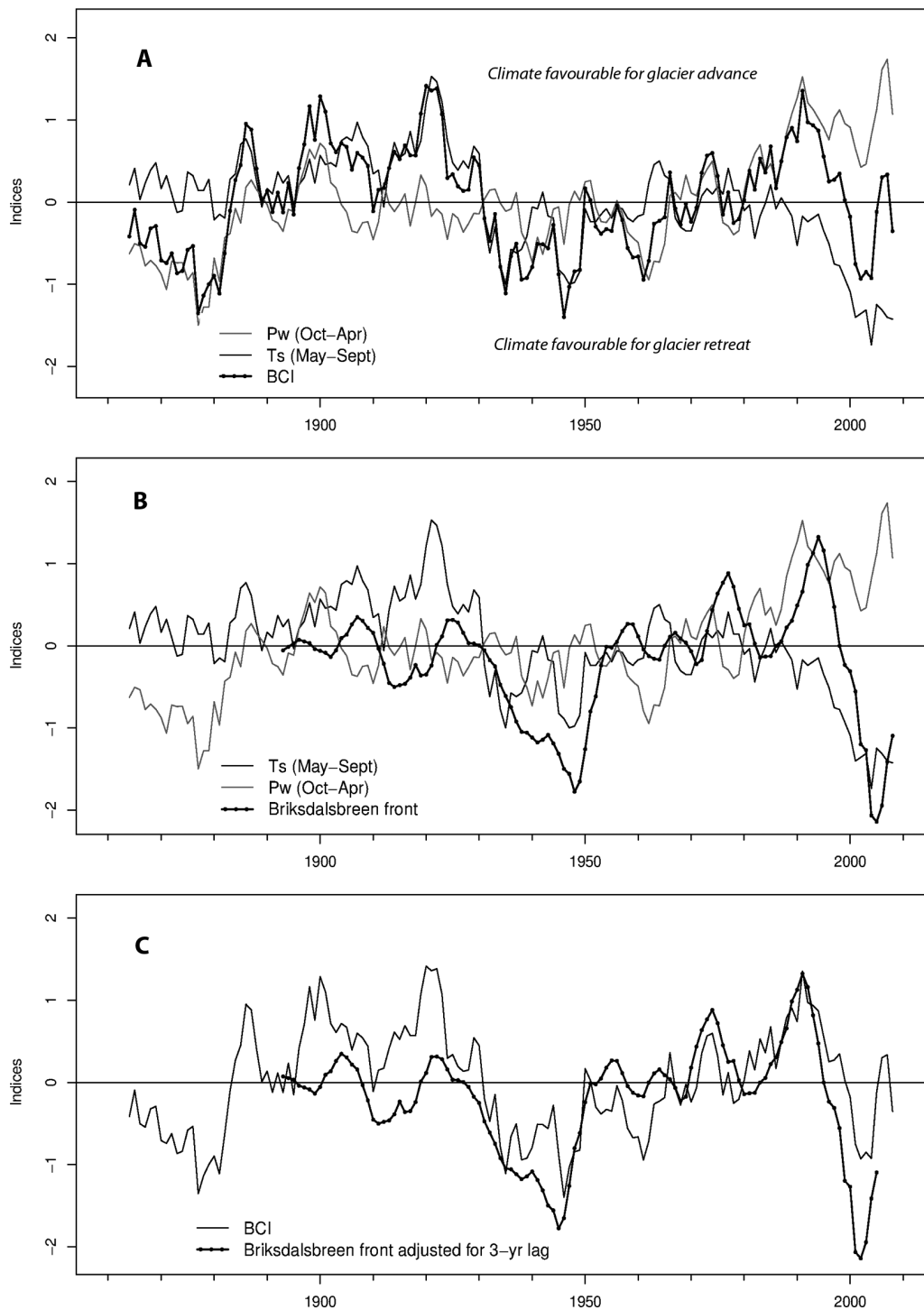


Figure 3. (A) Standardized values for P_w and T_s in Bergen (AD 1864–2010), and the Bergen Climate Index (BCI) calculated (all curves 5-year running mean). Values above the 0-line indicate conditions favourable for glacier advance, values below the 0-line represent conditions unfavourable for glacier advance. (B) Standardized values for P_w and T_s plotted together with Brikdalsbreen annual front variations (AD 1893–2010) (all curves 5-year running mean, Brikdalsbreen curve amplified by factor 10). (C) BCI plotted together with standardized annual front variations of Brikdalsbreen. The glacier front variation curve is adjusted for a 3-year time lag of frontal response (both curves 5-year running mean, Brikdalsbreen curve amplified by factor 10). Meteorological data: Climate Division, Meteorological Institute, Oslo; glacier-front data: NVE, Oslo

mainly driven by low spring (March–May) precipitation and high autumn (September–November) temperatures. After this twentieth-century minimum, Brikdalsbreen recovered gradually: a first advance maximum was reached in 1977, followed by a pronounced glacier advance with the maximum annual advance in 1994 (~22 cm/day; Nesje, 2005). Whereas the moderate advances of Brikdalsbreen in the early twentieth century were mainly due to reduced summer temperatures, the rapid advance of the 1990s was primarily attributed to increased winter precipitation

(positive NAO index), indicating that winter precipitation may be more important for advances of Norwegian maritime glaciers than summer temperature. The most recent retreat though, has been caused by a significant rise in summer temperatures. The summer of 2002 was even the warmest in western Norway since 1734 (Nordli et al., 2005).

The BCI and standardized annual glacier-front variations of Brikdalsbreen adjusted for a 3-year time lag are shown in Figure 3C. Periods of high BCI are, with a certain delay, followed by

Table 2. Maxima (max) and minima (min) of the annual front variations of Briksdalsbreen and their time lag to the BCI during the late LIA and after 1930

	Event	Briksdalsbreen	BCI	time lag
LIA	max 7	1896	1886	10
	min 6	1902	1895	7
	max 8	1907	1900	7
	min 7	1914	1910	4
	max 9	1925	1920	5
	Ø lag			6.6
Post-LIA	l/min 8	1948	1946	2
	max 10	1977	1974	3
	min 9	1983	1978	5
	max 11.1	1994	1991	3
	min 10	2005	2002	3
	Ø lag			3.2
overall	Ø lag			4.9
	Ø max lag			5.6
	Ø min lag			4.2

glacier advance. Statistical lagged correlation analysis shows that the best fit of the two curves is achieved with a glacier time lag of 3 years. When we compare the dating of the culmination points of the annual length changes and the BCI, the average frontal time lag is about 5 years (cf. Table 2). Thus, the time lag of Briksdalsbreen is *c.* 3–5 years. Dividing the record into a LIA period (1862–1930) and post-LIA period (1931–2010), characteristic time lags can be distinguished: during the LIA period the time lag was about twice as long as after the termination of the LIA (6.6 years compared with 3.2 years). This can also be observed in Figure 3C, where the fit of the adjusted glacier curve to the BCI is weaker in the period before 1910. The reason is that the reaction time of a glacier to a climatic forcing is proportional to its size; during the rapid retreat after the last major maximum in 1929 Briksdalsbreen shortened considerably, and hence needed less time

to adjust to a change in mass balance. Overall, the glacier is more rapidly reacting to a negative BCI (time lag of 4.2 years, compared with 5.6 years for positive BCI). In terms of climate forcing, glacier retreats are thus more immediate than advances.

Time lags at Jostedalsbreen and Folgefonna

Annual glacier length change measurements at Jostedalsbreen and Folgefonna started around 1900 (Andreassen et al., 2005). The glacier length change curves presented here (Figure 4) cover the last few centuries, and partly reach as far back as the seventeenth century (e.g. Nigardsbreen, 1675; Nesje et al., 2008; Nussbaumer et al., 2011). The historical fluctuations were reconstructed by Nussbaumer et al. (2011) using historical methods (analysis of written documents, drawings, paintings, photographs, topographic maps, etc.) and glaciomorphological evidence (moraines, etc.).

The maxima (max), minima (min) and characteristic inflexion points (I) of the Briksdalsbreen length change curve (cf. Figure 4), and the dates of the corresponding events at the other outlet glaciers are listed in Table 3. Considering the overall time lag, it appears that Supphellebreen and Bøyabreen do not lag behind Briksdalsbreen, but react in advance (time lag -3.6 and -3.5 years, respectively). Bondhusbreen and Buerbreen are in general well synchronized with the Briksdalsbreen fluctuations (time lag -1.3 and 0.2 years, respectively). The time lag of Buerbreen is not uniform over the whole record: until the end of the LIA (~ 1930), Buerbreen was 2 years in advance of Briksdalsbreen, whereas afterwards it lagged about 3 years behind Briksdalsbreen. The fluctuations of Bergsetbreen are more smoothed, and compared with the other glaciers it experienced a considerably weaker retreat after the LIA termination. Except the 1950–1970s period, the fluctuations of Bergsetbreen are, with a lag of about 4 years, well synchronized with Briksdalsbreen. The fluctuations of Nigardsbreen are even more smoothed, and the glacier only experienced minor advances, in a record of general retreat (time lag 15 years). Lodalsbreen, finally, has been continuously retreating during the whole record, and did not experience any maxima and minima until the 1980s (time lag ~ 18 years).

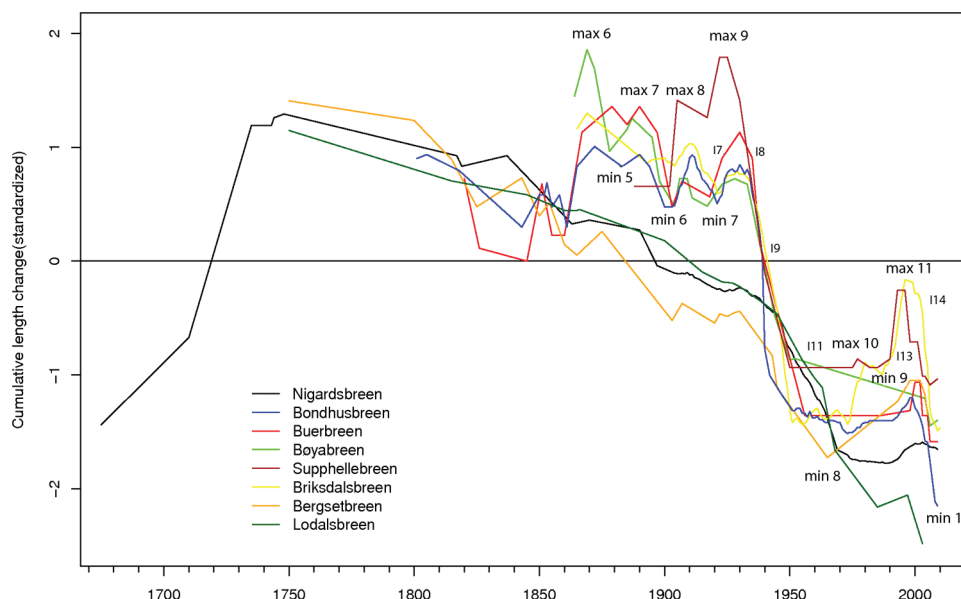


Figure 4. Standardized cumulative length variations of outlet glaciers of Jostedalsbreen and Folgefonna AD 1675–2010 and characteristic maxima (max), minima (min) and inflexion points (I). (Historical data points from Nussbaumer et al., 2011; length change measurement data from NVE, Oslo)

Table 3. Maxima (max), minima (min) and characteristic inflexion points (I) of cumulative length changes of Briksdalsbreen and the corresponding event dates at the other glaciers. Time lag (in years) is given relative to the Briksdalsbreen dates. The last row indicates the time lag to the BCI, calculated on the basis of the Briksdalsbreen time lag

Event	Briksdalsbreen	Nigardsbreen	Bondhusbreen	Buerbreen	Bøyabreen	Supphellebreen	Bergsetbreen	Lodalsbreen	time lag
max 6	1869	1890	1872	1879	1869	0	1875	1900	6
min 5	1893	1897	1883	1885	1878	-15	1903		10
max 7	1900	1909	1890	1890	1887	-13	1907		7
min 6	1904	1924	1900	1903	1904	0	1920	1923	16
max 8	1910	1930	1911	1907	1906	-4	1922	1926	12
min 7	1921		1921	1918	1917	-4	1925		4
17	1925		1925	1923	1923	-2			
max 9	1929	1946	1930	1930	1928	-1	1930	1945	1
18	1934	1946	1933		1933	-1			
19	1945	1963	1939				1943	1963	-2
111	1951	1969	1951	1956	1950	-1		1968	
min / 18	1973	1988	1973				1965	1985	-8
max 10	1980								
min 9	1987								
113	1992		1991	1998			1993		1
max 11.1	1996		1998	2000			1998	(1997)	2
max 11.2	1998	(2003)	1999	2002			2002		4
114	2002		2005	2005	2004	2	2004		2
min 10	2009			2006	2006	-3	2006		
Ø lag									
Ø max-lag			15.4			-3.5			4.2
Ø min-lag			16.8			-4.5			5.3
Ø I-lag			13.0			-5.5			5.5
			16			-0.5			0.3
time lag to BCI	4.9		20.3			1.4			9.1
									17.5
									23.3

The results show that the front of inert ice bodies (e.g. Nigardsbreen, Lodalsbreen) do not react to short-term climate variations. Steep and short glaciers (e.g. Supphellebreen, Bøyabreen, Briksdalsbreen) though, almost immediately adjust their front position to short-term changes. These differences in time lag are responsible for dissimilar reactions of nearby outlet glaciers of the same ice cap. Comparing reaction time for the culmination of advances and retreats, it seems that the time lag is smaller for minima, and the glaciers need more time to reach a maximum.

During the most recent advance, all the short outlet glaciers reached maxima between 1993 and 2002 (max11). Figure 5 shows photographs of Briksdalsbreen, Supphellebreen and Nigardsbreen at the beginning of the advance and during culmination, as well as a photograph of 2010 (for photographic series of Bergsetbreen and Bøyabreen see Winkler et al., 2009). Figure 6 depicts one of the rare photographs of the hardly accessible Lodalsbreen in 1992. After two centuries of more or less continuous melt-back (since the LIA maximum in the mid eighteenth century; Nussbaumer et al., 2011), also Lodalsbreen and Nigardsbreen experienced a maximum in 1997 and 2003, respectively. It is unclear whether this most recent maximum at Nigardsbreen occurred as a response to the same triggering factors as for the maximum at Briksdalsbreen. The time lag of only 5–7 years for Nigardsbreen and one year for Lodalsbreen are instance to the contrary. Alternatively, these maxima could correspond to event max10, which, however, only occurred at Supphellebreen and Briksdalsbreen, but not at the other short outlet glaciers.

The NAO and glacier fluctuations in Norway and the western Alps

It is generally considered that the LIA was a period of cooling after the relatively warm medieval times and before the twentieth-century warm phase (Pfister, 1999). The climate during the LIA was marked by lower than average temperatures, which led to rapid glacier advances in the fourteenth and early seventeenth centuries (Grove, 2004). However, Nesje et al. (2008) showed that the main cause of the early eighteenth-century glacial advance in western Scandinavia was not lower summer temperatures alone, but also mild and humid winters, associated with increased precipitation and high snowfall on the glaciers, most likely as a result of a prevailing positive NAO weather mode (Luterbacher et al., 2002). The climate conditions were not only varying in space, but also in time (Grove, 2004). Whereas the LIA maximum peaks in the Alps occurred around 1600/1640 and 1820/1850 (Nussbaumer et al., 2007; Zumbühl, 1980; e.g. Pré-de-Bard Glacier 1818/1819, Imhof, 2010), the glaciers in southern Norway culminated in the mid eighteenth century (e.g. Briksdalsbreen 1760–1765; Pedersen, 1976), some even around 1870–1890 (Nussbaumer et al., 2011). In the European central Alps, the LIA termination is commonly attributed to the mid nineteenth century (Holzhauser, 2010; Nussbaumer et al., 2007). In Scandinavia, however, glacier front variations and temperature records indicate that the LIA lasted until ~AD 1920 (Nesje and Dahl, 2003).

For the comparison of glacier fluctuations in southern Norway and the western Alps, Briksdalsbreen and Pré-de-Bard Glacier

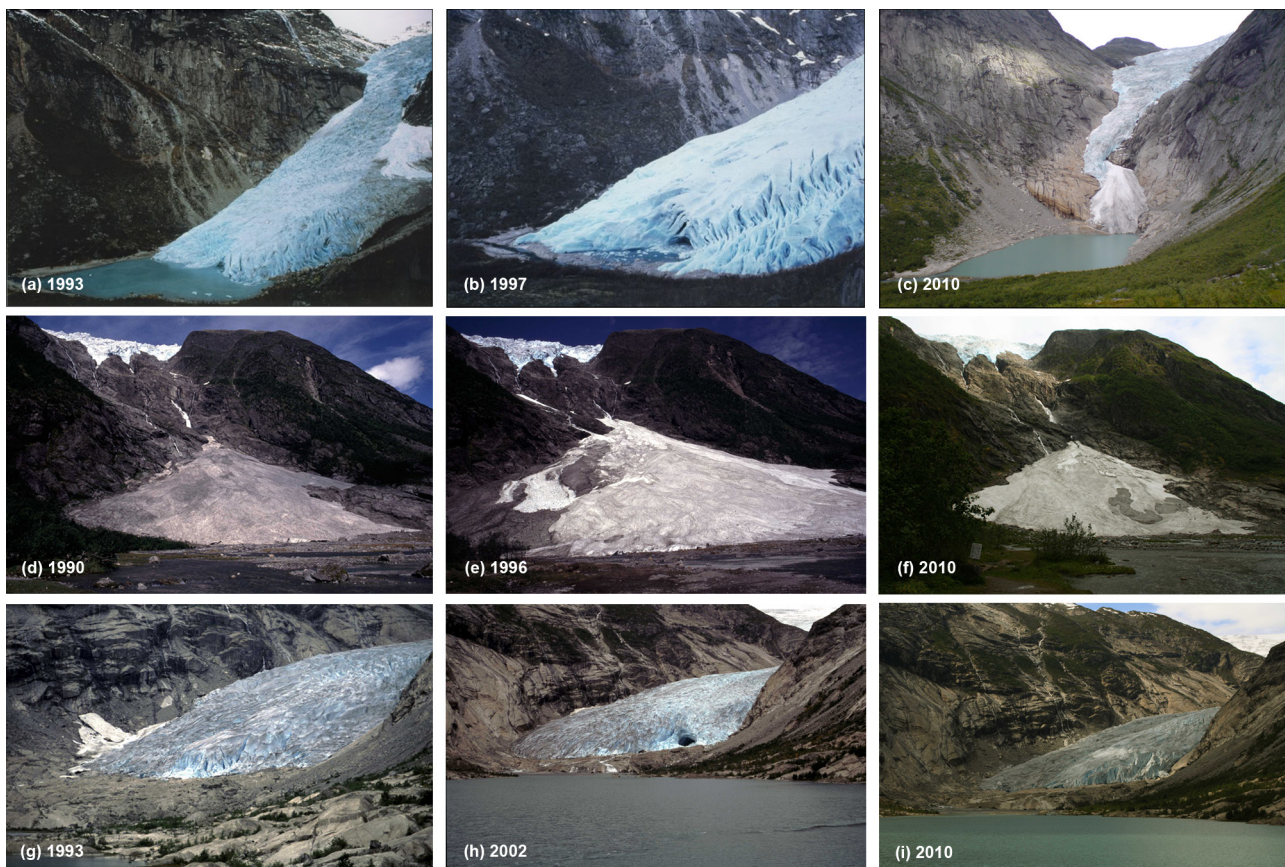


Figure 5. Photograph series of Briksdalsbreen (top), Supphellebreen (middle) and Nigardsbreen (bottom), showing the rapid advance in the 1990s and consequent retreat. Photographs: a,b, Sigbjørn Myklebust; c, Atle Nesje; d–i, Stefan Winkler



Figure 6. The glacier tongue of Lodalsbreen in a photograph from 1992 (photograph Stefan Winkler)

were chosen as examples. Both glaciers are comparatively small, with a short and steep valley tongue (cf. Figure 2, right). Thus they are comparable in their glacier dynamics and have a similar time lag (Briksdalsbreen, 3–5 years; Pré-de-Bard Glacier, 5 years). Note though, that the time lag of the Pré-de-Bard Glacier was determined by a correlation analysis between ablation season temperature anomalies and the corresponding length change (Imhof, 2010), whereas for Briksdalsbreen the correlation to the climate index was analysed. The newly reconstructed cumulative fluctuations of Pré-de-Bard Glacier (Imhof, 2010) and the Briksdalsbreen record are plotted in Figure 7A. For comparison, also the cumulative fluctuations of Mer de Glace (Nussbaumer et al., 2007), a glacier on the Mont Blanc north side, are plotted. On multidecadal scale the fluctuations of Pré-de-Bard Glacier and Briksdalsbreen are mainly anti-correlated; maxima at Pré-de-Bard Glacier coincide with minima at Briksdalsbreen. The last maximum at Pré-de-Bard Glacier was reached in 1989. At the same time Briksdalsbreen, like the other maritime Norwegian glaciers, only started its rapid advance, which culminated in 1996–1998.

These anti-phased glacier fluctuations on multidecadal scale are attributed to the difference in geographical location and climate. The maritime character of the Norwegian glaciers stands for a long melting season, especially on the lower part of the glaciers, with a high sensitivity to changes in temperature from May to October (Reichert et al., 2001). The sensitivity to relative changes in precipitation, on the other hand, is high in winter. The Alpine glaciers (e.g. Rhone Glacier in Reichert et al., 2001) have a high sensitivity to changes in temperature during summer, and nearly constant sensitivity to precipitation throughout the year. Therefore, it has to be assumed that even small changes in summer temperature have a noticeable impact on Pré-de-Bard Glacier, as because of its high elevation a large fraction of summer precipitation falls as snow. Moreover, as a more continental glacier situated in the lee of the Mont Blanc massif, the Pré-de-Bard Glacier receives less precipitation than Briksdalsbreen, which is nourished with frequent winds blowing from the sea onland (mean annual precipitation sum: Great St Bernard, 1858 mm, Schüepf, 1991; Jostedalsbreen, 2500 mm, Østrem et al., 1988). The western Alpine glaciers (Pré-de-Bard) are situated in the transition zone between maritime and continental regimes, which makes summer temperature the dominating factor, whereas Norwegian glaciers (Briksdalsbreen) are more sensitive to precipitation during the accumulation season (Steiner et al., 2008).

The anti-correlation of glacier fluctuations is also reflected in the fluctuations of the Bergen and Great St Bernard Climate Indices (Figure 7B); periods with conditions favourable for glacier advance in Bergen, coincide with periods favourable for glacier

retreat at Great St Bernard. The bi-polar pattern leads to the investigation of possible relationships with the NAO. The correlation between the low-pass filtered (5-year running mean) $NAO_{Dec-Mar}$ index and the BCI is $r=0.44$, whereas for the GSBCI an anti-correlation of $r=-0.23$ was found (both significant at 95% confidence level; for observation periods compare data and methods section). High NAO index winters can be related to above-normal precipitation over Scandinavia (important for Briksdalsbreen) and below-normal precipitation over central and southern Europe (important for Pré-de-Bard Glacier) (Hurrell, 1995). The correlation between $NAO_{Dec-Mar}$ and winter precipitation ($P_{Dec-Mar}$) between AD 1864–2010 in Bergen is $r=0.49$ ($r=0.89$ during 1962–1998). Thus, a strongly positive NAO is associated with conditions favourable for glacier advances in Scandinavia (positive BCI).

At the same time the climate conditions are rather unfavourable for glacier advance in the western Alps, which is reflected in negative GSBCI values. If we calculate the correlations between $NAO_{Dec-Mar}$ and the annual glacier fluctuations shifted for the typical time lag, the coefficient for Briksdalsbreen is $r=0.26$, whereas for Pré-de-Bard Glacier an anti-correlation of $r=-0.23$ was found (both significant at 95% confidence level). These are rather weak correlations, and the explained variance is only 5%. The glacier and $NAO_{Dec-Mar}$ records (Figure 7C) indicate that the suggested relationship was particularly pronounced in the 1962–1998 period. In fact, the correlations for this period are significantly higher: the correlation for Briksdalsbreen increases to $r=0.63$, the anti-correlation of Pré-de-Bard Glacier to $r=-0.73$. Apparently the strength of the relationship NAO–climate and NAO–glacier is highly variable over the whole record. Statistically, the correlations are only significant during some periods (e.g. AD 1962–1998). Before 1940, the correlations between NAO and GSBCI ($r=0.07$) or NAO and Pré-de-Bard Glacier ($r=0.17$) are even slightly positive, but tending towards zero. In the period AD 1945–1965 on the other hand, we find a significant anti-correlation between NAO and BCI ($r=-0.31$) and Briksdalsbreen ($r=-0.56$).

This demonstrates the non-stationarity of the climate system; according to Nussbaumer et al. (2011) the linear (stationary) approach between the NAO index and ‘downstream’ site-specific palaeoproxy reconstructions is limited by the non-stationary behaviour of the NAO system, leading to loss of temporal robustness of such analyses. Recent analyses (Nussbaumer et al., 2011; Ø. Lie, Bjerknes Centre for Climate Research, personal communication, 24 June 2008) indicate that the influence of the NAO index on Norwegian temperature and precipitation was not stationary within the twentieth century. Nevertheless, the analyses in this paper reveal a general correlation between Briksdalsbreen and the $NAO_{Dec-Mar}$. The relationship between the NAO and the Pré-de-Bard Glacier on the other hand, is more complex; the NAO explains but one-third of the climate variability, and only during winter (Casty et al., 2005). Temperature and precipitation in the Alps are controlled by variable NAO influences due to the geographical situation of the mountains, and thus influenced by other atmospheric circulation modes too (Casty et al., 2005).

Discussion

Climate response at Jostedalsbreen and Folgefonna

The described glacier fluctuations during the twentieth century are in good accordance with the observed temperature and precipitation trends. Nordli et al. (2005) showed that the 1930s and

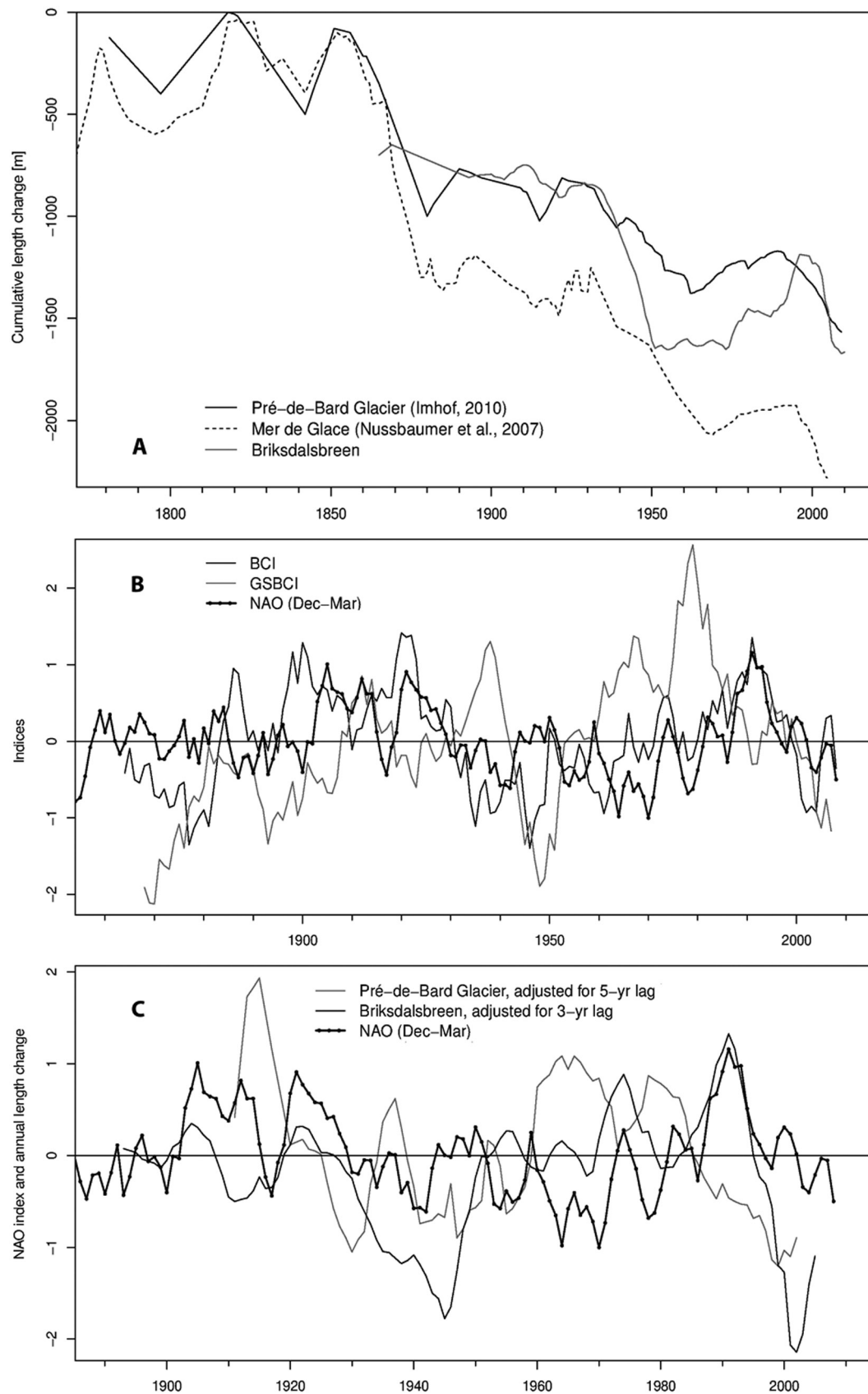


Figure 7. (A) Cumulative length change of Pré-de-Bard Glacier (AD 1781–2009; Imhof, 2010) relative to its LIA maximum extent in 1818/1819 (=0) and Briksdalsbreen fluctuations (AD 1865–2010; Nussbaumer et al., 2011), as well as Mer de Glace fluctuations (AD 1770–2005; Nussbaumer et al., 2007). (B) Climate index at Bergen (BCI) and Great St Bernard (GSBCI) plotted together with NAO_{Dec-Mar} index (all curves 5-year running mean). (C) Standardized annual length change of Pré-de-Bard Glacier (AD 1909–2009) and Briksdalsbreen (AD 1893–2010) adjusted for their typical time lag, plotted together with NAO_{Dec-Mar} index (all curves 5-year running mean).

1940s, or the so called 'early warming period', was characterized by an exceptional ablation maximum, and corresponding particularly high values of the equilibrium line altitude, conditions that led to the pronounced glacier retreat after 1930. According to Nordli et al. (2005) the large winter snow accumulation during the early 1990s, leading to the most recent glacier advance, was unprecedented in the period 1781–2000. In addition, Winkler et al. (2009) state that during the 1990s, the precipitation maximum shifted from autumn into winter. Analyses in this paper showed that the largest glacier advance during the twentieth century was mainly caused by increased winter precipitation. Moreover, the main cause of the early eighteenth-century glacial advance, leading to the LIA maximum at most Scandinavian glaciers, was mild and humid winters associated with increased precipitation and high snowfall on the glaciers, most likely as a result of a prevailing positive NAO weather mode in the previous decades (Nesje et al., 2008). According to Nesje (2005) and Nesje et al. (2000), the net mass balance of maritime glaciers in southern Norway is mainly controlled by the winter balance. However, a strongly positive summer temperature anomaly seems to be the explaining factor for the most recent glacier retreat (Winkler et al., 2009). Based on this event, Winkler et al. (2009) suggest that the previously typical maritime glaciers strongly influenced by precipitation patterns, could become much more sensitive to summer air temperatures and the length of the ablation period. Overall, Norwegian glaciers do not respond uniformly to either summer temperature or winter precipitation, but rather to a complex interplay between different synoptic weather patterns (Nesje et al., 2008).

The rather short time lag of Briksdalsbreen can be explained by the morphology of the glacier basin. The ice masses are accumulated in a wide accumulation basin, and at the ablation zone channelized into a narrow and steep rock escarpment. This topographical channelling effect leads to the amplification of mass balance perturbations. In the narrow, gorge-like section of the icefall microclimatological effects (e.g. long-wave radiation from the valley walls) additionally enhance a meltback signal (Winkler et al., 2009). The longer time lags during the LIA correspond to a more advanced glacier, where the front ended in the plain below the ice fall. Exceeding ice masses had to be transported not only over a longer distance, but also over a sparsely inclined plain. Accordingly, the shortest time lag was achieved around 1950 (min8), when Briksdalsbreen was in a similarly retreated position as today. According to Oerlemans (2007), the large mass turnover (ablation rates on the snout ~ 10 m of ice) is another explanation for the short time lag of Briksdalsbreen.

The study of the frontal time lags showed that Supphellebreen and Bøyabreen have a shorter time lag than Briksdalsbreen. This can be attributed to differences in morphology. Supphellebreen is divided into an active upper part, and a regenerated lower part (cf. Figure 5). From the accumulation basin, exceeding ice mass is rapidly transported to the relatively short active front, where it is transmitted to the regenerated front by ice avalanches instead of flowing down the icefall. Mass balance perturbations are thus directly transported to the valley tongue (Orheim, 1970). Similar mechanisms occur at Bøyabreen, which in the last decades has been separated into two parts (with intermittent reconnection during 1994–2001; Winkler et al., 2009). However, in contrast to Supphellebreen, the active front of Bøyabreen is located half way down the icefall and avalanches fall a much shorter vertical distance.

The results suggest that we have mainly two groups of glaciers. The first group consists of the short and steep outlet glaciers Supphellebreen, Bøyabreen, Briksdalsbreen, Bondhusbreen and Buerbreen. They are characterized by an average time lag of only 1–5 years (observation period 1869–2009), and sensitively react on short-term climatic changes. The long and gently descending Nigardsbreen and Lodalsbreen form another group. Because of their morphology, they have a time lag of about 20 years, and only react on long-term climatic changes. Bergsetbreen, finally, is an intermediate case. With a reaction time of about 9 years, it is not as rapidly fluctuating as the short and steep outlet glaciers, but has nevertheless higher amplitude of fluctuations than the rather strongly smoothed Nigardsbreen and Lodalsbreen. During the most recent retreat, the tongue of Bergsetbreen has been disintegrating and was cut off from the active part in 2006, and is now prone to melt down (Winkler et al., 2009). In the shape of the ongoing climate warming, the disintegration of the tongue is also to be expected at the other steep and short outlet glaciers.

Comparative data concerning time lag is only available for some of the study glaciers. For Briksdalsbreen, Winkler (2009) empirically derived a time lag of 3–4 years. Oerlemans (2007) calculated a response time of 5 years for Briksdalsbreen on the basis of the measured length record. This is very well in accordance with the time lag of 3–5 years determined here. According to model calculations by Laumann and Nesje (2009), the frontal time lag of Briksdalsbreen will also in the future be 4–5 years. For Nigardsbreen, Nesje and Dahl (2003) calculated a frontal time lag of ~ 20 years, which is in good accordance with the time lag of 20.3 years calculated in this study.

It is to be expected that, as the studied outlet glaciers of Jostedalbreen are part of the same ice plateau and their accumulation areas are adjacent or close to each other, the differences in glacier behaviour represent different time lags to similar climate changes. However, Bøyabreen, the glacier with the second shortest time lag, is located in the southwestern part of Jostedalbreen and thus lies ahead in the flow direction of the southwesterly maritime winds. This could be an explanation for the early reaction of Bøyabreen compared with Briksdalsbreen. Supphellebreen, the glacier with the shortest time lag, is located next to Bøyabreen. Orheim (1970) showed that its short time lag is caused by the accumulation through avalanches. Comparing the Jostedalbreen outlet glaciers with Bondhusbreen and Buerbreen reveals parallel deviations, where advances and retreats occurred at all glaciers with no obvious spatial differentiation. This suggests that also the differences in behaviour of Jostedalbreen and Folgefonna outlet glaciers represent different responses to similar climate changes.

Glacier fluctuations and the NAO

The NAO index is most pronounced during winter, and therefore influences glacier mass balance mainly via winter precipitation. According to Reichert et al. (2001) the individual impact of precipitation on mass balance of Nigardsbreen is about 1.6 times higher than the impact of temperature. In a coupled GCM experiment, Reichert et al. (2001) calculated a well-pronounced correlation between the low-pass filtered NAO index and mass balance of Nigardsbreen ($r=0.55$; $r=0.58$ within the period 1962–1998), as well as a significant anti-correlation with Rhone Glacier in the Swiss Alps ($r=-0.64$). These values are in good agreement with the correlation coefficients with the NAO_{Dec-Mar} calculated for Briksdalsbreen ($r=0.63$) and Pré-de-Bard Glacier ($r=-0.73$)

annual length fluctuations over the period 1962–1998. Nesje et al. (2000) showed that the effect of the NAO on the glacier winter/net mass balance in Scandinavia gradually decreases with increasing continentality and increasing latitude (e.g. maritime Ålfotbreen, $r=0.69$; continental Gråsübreen, $r=0.22$).

In the European Alps the NAO affects the climate in a complex manner, as the Alps are situated in the transition zone between northern Europe (NAO positive: warm/wet; NAO negative: cold/dry) and southern Europe, where almost the opposite reactions apply (Wanner et al., 2000). The NAO influence in the Alpine region is less pronounced than in Scandinavia (Six et al., 2001) and often ambiguous (Scherrer et al., 2004). Steiner et al. (2008) showed that winter precipitation, which is the main factor linking NAO and glacier fluctuations, is less important for length variations of Alpine glaciers than for Norwegian glaciers. Moreover, Alpine glaciers are more sensitive to summer temperature, which is less influenced by the NAO (Reichert et al., 2001).

A simulation of the anomaly patterns of precipitation associated with the positive winter NAO over Europe by Reichert et al. (2001) showed that Briksdalsbreen is located near the maximum precipitation anomaly of +40 mm/month, whereas Pré-de-Bard Glacier is located in a region of reduced winter precipitation (~ -20 mm/month). Considering that the Briksdalsbreen is most sensitive to winter precipitation anomalies, this demonstrates the strong influence of the NAO on its glacier mass balance and length fluctuations. A winter-precipitation index for the northern Swiss Alpine foreland between AD 1550 and 1995 (Pfister, 1995) shows, that the time of significant glacier growth in the early eighteenth century in Scandinavia was characterized by a decreasing winter precipitation trend in the Alps. This is in accordance with the NAO north–south dipole pattern in Europe, which partly explains the asynchronous behaviour of LIA glacier variations in Scandinavia and the European Alps (Nesje and Dahl, 2003; Reichert et al., 2001; Six et al., 2001).

Summary and conclusion

1. The net mass balance of maritime Norwegian glaciers is mainly controlled by winter precipitation. Whereas the moderate glacier advances in the early twentieth century were mainly due to reduced summer temperatures, the rapid advance of the 1990s was primarily attributed to increased winter precipitation (positive NAO index).
2. The time lag of a glacier is proportional to its size and thus not constant during the whole record. During the LIA the time lag of Briksdalsbreen for example, was twice as long as in the twentieth century. In general, glaciers react more rapidly to a negative climate signal, and glacier retreats are more immediate than advances.
3. Varying time lags of adjacent outlet glaciers are mainly due to differences in glacier morphology. Short and steep glaciers (e.g. Bøyabreen) almost immediately adjust their front position to short-term climate changes, whereas the fronts of inert ice bodies (e.g. Lodalsbreen) only react to long-term changes and have smoothed fluctuations.
4. Supphellebreen and Bøyabreen react in advance of Briksdalsbreen (average time lags of 1.3 and 1.4 years, respectively; observation period 1869–2009), whereas Bondhusbreen and Buerbreen are in general well synchronized with Briksdalsbreen (average time lags of 3.6, 5.1 and 4.9 years, respectively). The fluctuations of Bergsetbreen are more smoothed, and lag about 4 years behind Briksdalsbreen. Long and gently descending Nigardsbreen and Lodalsbreen finally, have average time lags of 20.3 and 23.3 years, respectively.
5. A strongly positive NAO index leads to enhanced winter precipitation, higher-than-normal positive mass balance and, with a certain time lag, to the advance of Norwegian glaciers. In the Alps, changes in the westerly circulation associated with a positive NAO index can lead to reduced winter precipitation and lower than normal winter mass balance and thus, after a certain delay, to retreating glaciers.
6. The (anti-)correlation with the NAO was not constant during the record. The classical relationship was particularly pronounced between 1962 and 1998 (Briksdalsbreen, $r=0.63$; Pré-de-Bard Glacier $r=-0.73$), but significantly weaker or even inverted during other periods. In the perspective of climate change, especially maritime glaciers in southern Norway will be influenced in a complex way by future fluctuations (frequency and magnitude) of the NAO.

Acknowledgements

The authors would like to thank Mathias Trachsel for help with statistical analyses and Jürg Luterbacher for valuable comments on the manuscript. The paper was improved thanks to the advice of two anonymous reviewers. This is publication A 325 of the Uni Bjerknæs Centre, N-5007 Bergen, Norway.

Funding

This research received no specific grant from any funding agency in the public, commercial, or not-for-profit sectors.

References

- Andreassen LM, Elvehøy H, Kjølmoen B, Engeset RV and Haakensen N (2005) Glacier mass-balance and length variation in Norway. *Annals of Glaciology* 42: 317–325.
- Casty C, Wanner H, Luterbacher J, Esper J and Böhm R (2005) Temperature and precipitation variability in the European Alps since 1500. *International Journal of Climatology* 25(14): 1855–1880.
- Grove JM (2004) *Little Ice Ages: Ancient and Modern*. Second edition. London and New York: Routledge, 2 vols.
- Haerberli W (1995) Glacier fluctuations and climate change detection. *Geografica Fisica e Dinamica Quarternaria* 18(2): 191–199.
- Haerberli W (2009) Accelerated worldwide glacier shrinkage. *Geophysical Research Abstracts* 11. EGU2009-14131. EGU General Assembly.
- Haerberli W and Beniston M (1998) Climate change and its impacts on glaciers and permafrost in the Alps. *Ambio* 27(4): 258–265.
- Hoelzle M, Chinn T, Stumm D, Paul F, Zemp M and Haerberli W (2007): The application of glacier inventory data for estimating past climate change effects on mountain glaciers: A comparison between the European Alps and the Southern Alps of New Zealand. *Global and Planetary Change* 56(1–2): 69–82.
- Holzhauser H (2010) *Zur Geschichte des Gornergletschers. Ein Puzzle aus historischen Dokumenten und fossilen Hölzern aus dem Gletschervorfeld*. Geographica Bernensia, G 84. Geographisches Institut der Universität Bern, 253 pp.
- Hurrell JW (1995) Decadal trends in the North Atlantic Oscillation: Regional temperatures and precipitation. *Science* 269(5224): 676–679.
- Imhof P (2010) Glacier fluctuations in the Italian Mont Blanc massif from the Little Ice Age until the present. Historical reconstructions for the Miage, Brenva and Pré-de-Bard Glaciers. M.Sc. Thesis. Geographisches Institut der Universität Bern, 132 pp.
- Intergovernmental Panel on Climate Change (2007) *Climate Change 2007: The Physical Science Basis*. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (Solomon S et al.). Cambridge: Cambridge University Press, 996 pp.
- Laumann T and Nesje A (2009) The impact of climate change on future frontal variations of Briksdalsbreen, western Norway. *Journal of Glaciology* 55(193): 789–796.
- Luterbacher J, Xoplaki E, Dietrich D, Jones PD, Davies TD, Portis D et al. (2002) Extending North Atlantic Oscillation reconstructions back to 1500. *Atmospheric Science Letters* 2(1–4): 114–124.
- Nesje A (2005) Briksdalsbreen in western Norway: AD 1900–2004 frontal fluctuations as a combined effect of variations in winter precipitation and summer temperature. *The Holocene* 15(8): 1245–1252.

- Nesje A and Dahl SO (2003) The 'Little Ice Age' – Only temperature? *The Holocene* 13(1): 139–145.
- Nesje A, Dahl SO, Thun T and Nordli Ø (2008) The 'Little Ice Age' glacial expansion in western Scandinavia: Summer temperature or winter precipitation? *Climate Dynamics* 30(7–8): 789–801.
- Nesje A, Johannessen T and Birks HJB (1995) Briksdalsbreen, western Norway: Climatic effects on the terminal response of a temperate glacier between AD 1901 and 1994. *The Holocene* 5(3): 343–347.
- Nesje A, Lie Ø and Dahl SO (2000) Is the North Atlantic Oscillation reflected in Scandinavian glacier mass balance records? *Journal of Quaternary Science* 15(6): 587–601.
- Nordli Ø, Lie Ø, Nesje A and Benestad R (2005) Glacier mass balance in southern Norway modelled by circulation indices and spring-summer temperatures AD 1781–2000. *Geografiska Annaler* 87A(3): 431–445.
- Nussbaumer SU, Nesje A and Zumbühl HJ (2011) Historical glacier fluctuations of Jostedalbreen and Folgefonna (southern Norway) reassessed by new pictorial and written evidence. *The Holocene* 21(3): 455–471.
- Nussbaumer SU, Zumbühl HJ and Steiner D (2007) Fluctuations of the Mer de Glace (Mont Blanc area, France) AD 1500–2050: An interdisciplinary approach using new historical data and neural network simulations. *Zeitschrift für Gletscherkunde und Glazialgeologie* 40(2005/2006): 1–183.
- Oerlemans J (2005) Extracting a climate signal from 169 glacier records. *Science* 308(5722): 675–677.
- Oerlemans J (2007) Estimating response times of Vadret da Morteratsch, Vadret da Palü, Briksdalsbreen and Nigardsbreen from their length records. *Journal of Glaciology* 53(182): 357–362.
- Orheim O (1970) *Glaciological Investigations of Store Supphellebre, West-Norway*. Norsk Polarinstitut Skrifter nr. 151, 48 pp.
- Østrem G, Dale Selvig K and Tandberg K (1988) *Atlas over breer i Sør-Norge. Norges vassdrags- og energiverk*. Meddelelse 61 fra Hydrologisk avdeling. Oslo: Norges Vassdrags- og energiverk, Vassdragsdirektoratet, 248 pp.
- Paterson WSB (1994) *The Physics of Glaciers*. Third edition. Oxford: Elsevier Science, 481 pp.
- Paul F (2006) The new Swiss glacier inventory 2000 – Application of remote sensing and GIS. Dissertation University of Zurich. Physische Geographie, 52, 194 pp.
- Pedersen K (1976) Briksdalsbreen, Vest-Norge. Glasiologiske- og glasiologiske undersøkelser. Cand. Real. Thesis, University of Bergen, 2 vols.
- Pfister C (1995) Monthly temperature and precipitation in central Europe from 1529–1979: Quantifying documentary evidence on weather and its effects. In: Bradley RS and Jones PD (eds) *Climate Since AD 1500*. London and New York: Routledge, 118–142.
- Pfister C (1999) *Wettermachhersage. 500 Jahre Klimavariationen und Naturkatastrophen (1496–1995)*. Bern: Haupt, 304 pp.
- Reichert BK, Bengtsson L and Oerlemans J (2001) Midlatitude forcing mechanisms for glacier mass balance investigated using general circulation models. *Journal of Climate* 14(17): 3767–3784.
- Scherrer SC, Appenzeller C and Laternser M (2004) Trends in Swiss Alpine snow days: The role of local- and large-scale climate variability. *Geophysical Research Letters* 31: L13215, doi:10.1029/2004GL020255.
- Schmeits MJ and Oerlemans J (1997) Simulation of the historical variations in length of Unterer Grindelwaldgletscher, Switzerland. *Journal of Glaciology* 43(143): 152–164.
- Schüepp M (1991) The precipitation series of Gr. St. Bernard and Geneva since 1818. *Theoretical and Applied Climatology* 44(2): 139–142.
- Six D, Reynaud L and Letréguilly A (2001) Bilans de masse des glaciers alpins et scandinaves, leurs relations avec l'oscillation du climat de l'Atlantique nord. *Sciences de la Terre et des Planètes* 333: 693–698.
- Steiner D, Pauling A, Nussbaumer SU, Nesje A, Luterbacher J, Wanner H et al. (2008) Sensitivity of European glaciers to precipitation and temperature – Two case studies. *Climatic Change* 90(4): 413–441.
- Wanner H, Beer J, Büttikofer J, Crowley TJ, Cubasch U, Flückiger J et al. (2008) Mid- to Late Holocene climate change: An overview. *Quaternary Science Reviews* 27(19–20): 1791–1828.
- Wanner H, Brönnimann S, Casty C, Gyalistras D, Luterbacher J, Schmutz C et al. (2001) North Atlantic Oscillation – Concept and studies. *Surveys in Geophysics* 22(4): 321–382.
- Wanner H, Holzhauser H, Pfister C and Zumbühl HJ (2000) Interannual to century scale climate variability in the European Alps. *Erdkunde* 54(1): 62–69.
- Winkler S, Elvehøy H and Nesje A (2009) Glacier fluctuations of Jostedalbreen, western Norway, during the past 20 years: The sensitive response of maritime mountain glaciers. *The Holocene* 19(3): 395–414.
- Zemp M, Paul F, Hoelzle M and Haeberli W (2008) Glacier fluctuations in the European Alps 1850–2000: An overview and spatio-temporal analysis of available data. In: Orlove B, Wiegandt E and Luckman BH (eds) *The Darkening Peaks: Glacier Retreat, Science, and Society*. University of California Press, 152–167.
- Zumbühl HJ (1980) *Die Schwankungen der Grindelwaldgletscher in den historischen Bild- und Schriftquellen des 12. bis 19. Jahrhunderts. Ein Beitrag zur Gletschergeschichte und Erforschung des Alpenraumes*. Denkschriften der Schweizerischen Naturforschenden Gesellschaft (SNG), Band 92. Basel: Birkhäuser, 279 pp.