# High sensitivity of North Iceland (Tröllaskagi) debris-free glaciers to climatic change from the 'Little Ice Age' to the present

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## Abstract

The Tröllaskagi peninsula is located in northern Iceland, between meridian 19°30′W and 18°10′W, jutting out into the North Atlantic to latitude 66°12′N. The aim of this research is to study recent glacier changes in relation to climatic evolution of the Gljúfurárjökull and Tungnahryggsjökull debris-free valley glaciers in Tröllaskagi. Glacier extent mapping and spatial analysis operations were performed with ArcGIS (ESRI), using analysis of aerial photographs from 1946, 1985, 1994 and 2000, and a 2005 SPOT satellite image. The results show that these glaciers lost a quarter of their surface area between the 'Little Ice Age' and 2005. In this paper, the term 'Little Ice Age' follows Grove (2001) as the most recent period when glaciers extended globally between the medieval period and the early 20th century. The abrupt climatic transition of the early 20th century and the 25-year warm period 1925–1950 triggered the main retreat and volume loss of these glaciers since the end of the 'Little Ice Age'. Meanwhile, cooling during the 1960s, 1970s and 1980s altered the trend, with advances of the glacier snouts. Between the 'Little Ice Age' and the present day, the mean annual air temperature and mean ablation season air temperature increased by 1.9°C and 1.5°C, respectively, leading to a 40–50 m rise in the equilibrium line altitude (ELA) of the glaciers during this period. The response of these glaciers depends not only on the mean ablation season air temperature evolution but also on other factors such as winter precipitation. The models applied show a precipitation increase of up to more than 700 mm since the 'Little Ice Age'.

## Keywords

climatic change, deglaciation, equilibrium line altitude, Iceland,'Little Ice Age', Tröllaskagi peninsula

## Introduction

The variations in summer temperature (mean ablation season air temperature, Ts) and winter accumulation exert a particularly decisive influence on the dynamics of the debris-free glaciers (Eythorsson, 1935; Liestøl, 1967; Ohmura et al., 1992). For example, Caseldine (1985b) pointed out that the combined effect of a low Ts and normal winter precipitation led to a glacier advance over a 10-year period in Tröllaskagi peninsula. The debris-free glaciers are especially sensitive to climate variations and were extensively studied in the peninsula until the late 1980s (Caseldine, 1983, 1985a, 1985b, 1987; Caseldine and Cullingford, 1981; Caseldine and Stötter, 1993).

For many of the glaciers in the central highlands of Iceland and for the majority of those in the Tröllaskagi peninsula, the maximum glacial advance in the second half of the Holocene was reached during the 'Little Ice Age' (LIA; Flowers et al., 2007; Kirkbride and Dugmore, 2001, 2006; Larsen et al., 2011; Schomacker et al., 2003).

The Holocene temperature variation range is thought to be around 3°C (Stötter et al., 1999): mean annual air temperature (MAAT) during the Holocene thermal maximum (HTM) is estimated to be 3°C higher than that for the period 1961–1990 (Caseldine et al., 2006; Geirsdóttir et al., 2009), and therefore comparable to the warmest decades of the 20th century (Stötter et al., 1999), while conditions during the post-Preboreal Holocene minimum were similar to those during the second half of the 19th century (Stötter et al., 1999). Precipitation is estimated to have doubled between both climatic extremes (Stötter et al., 1999). Caseldine and Stötter (1993) suggest that from the LIA maximum in the late 19th century to the 1980s, the Ts increased by 2°C and winter precipitation by 600 mm (+41%, from 1450 mm) at the equilibrium line altitude (ELA). Climate evolution during the last millennium, and its relationship to sea ice expansion, is relatively well known from the work of Koch (1945), Bergþórsson (1969), Ogilvie (1984, 1996, 2005, 2010) and Ogilvie and Jónsson (2001). Especially cold climatic episodes at the beginning and end of the 13th century, in much of the second half of the 14th century (1350–1380) and during the later years of the 16th century have been reported by Ogilvie (1984, 1991) and Ogilvie and Jónsson (2001). Koch (1945) suggested that between AD 1600 and 1900, the climate was particularly cold in Iceland and East Greenland, but Ogilvie (1984, 2005, 2010) pointed out that there was great variability with some mild periods and different levels of incidence of the sea ice: a more temperate climate occurred during the 1640s, 1650s and early 18th century, while very cold climatic episodes in the late 17th century (1690s), mid-18th century (1740s in the north and 1750s in the south) and during the 19th century (1810s, 1830s and 1880s; Ogilvie, 2010; Ogilvie and Jónsdóttir, 2000; Ogilvie and Jónsson, 2001), coinciding with the maximum sea ice extent. This is in contrast to the 20th century, which was more temperate than the three preceding centuries (Ogilvie, 1984, 1996, 2010; Ogilvie and Jónsson, 2001).

The limits and definition of the LIA are complex issues as they vary between authors, probably because of the differences in regions and the approaches applied (Grove, 2001). The origins and uses of the term 'Little Ice Age' are discussed in detail in Ogilvie and Jónsson (2001). Grove (1988) considered it to have begun earlier and extended from 1450 to 1900. However, it is suggested this period should be expanded, as much evidence has demonstrated that the LIA was under way in the early 14th century along North Atlantic regions as synthesized by Grove (2001), for example, glacial activity found in Iceland during the 13th and 14th centuries. Nevertheless, the period from 1600 to 1900 must have been more important for the glacial activity in Iceland as it was not interrupted by major warming (Guðmundsson, 1997), except a mild period around 1640–1680 (Ogilvie, 2005, 2010). We define the LIA in this paper according to Grove (2001) as the most recent period when glaciers extended globally and remained enlarged between the Medieval period and the warming beginning in the early 20th century (Grove, 1988).

The Gljúfurárjökull and Western Tungnahryggsjökull glaciers reached their maximum extent during the LIA, more precisely during the second half of the 19th century (Caseldine, 1983, 1985b; Caseldine and Cullingford, 1981), coinciding with the Holocene maximum advance of many Icelandic ice-cap outlet glaciers (Geirsdóttir et al., 2009; Kirkbride and Dugmore, 2008). Nevertheless, the maxima of Gljúfurárjökull and Tungnahryggsjökull were not synchronous: Gljúfurárjökull reached maximum extent around AD 1898–1903, while the Western Tungnahryggsjökull reached its maximum in AD 1868 (Caseldine, 1983, 1985b; Caseldine and Cullingford, 1981). Thus, the LIA climate of the Tröllaskagi peninsula was characterized by a Ts 2°C lower than at present and an average winter precipitation of 1450 mm at the ELA (946 m; Caseldine and Stötter, 1993).

Rising temperatures from the end of the 19th century caused the glaciers to retreat from their LIA positions. The retreat of the Western Tungnahryggsjökull (Caseldine, 1985b) started decades earlier than in Gljúfurárdalur, because of the steeper gradient of the Vesturdalur valley and the reduced thickness of the glacier. The retreat throughout the 19th century was interrupted by different advance phases with moraine formation: 1876–1878, 1882– 1887 and 1898–1903 (Caseldine, 1985b).

Gljúfurárjökull retreated 250 m from its LIA position during the first 20 years of the 20th century (Caseldine, 1983, 1987). This retreat was interrupted with moraine formation in 1910 and 1913–1917, and later slowed down between the mid-1920s and 1930. During the early 1930s, the retreat accelerated again (ca. 200 m) and was interrupted by minor advances which enabled moraine deposition around 1935. Once again, in the late 1940s there was a short re-advance, concluding in 1950–1951 (Caseldine, 1983), which left a series of moraine arcs. Marginal measurements of the Icelandic Glaciological Society (IGS) show that Gljúfurárjökull continued to retreat 422 m until it reached its minimum extent in the mid-1970s. Glacial retreat data since the LIA obtained by Caseldine and Cullingford (1981) using photogrammetry and lichenometry show that the terminus retreated 265 m between 1939 and 1972 and 151 m between 1953 and 1960. After reaching its minimum extension, the trend reversed in 1977. The glacier commenced a new, more continuous re-advance of greater scope than the advances which had occurred during deglaciation: the snout advanced 50 m in 1977–1979, slowing in the following years to 30 m from 1979 to 1981 and 25 m from 1981 to 1983 (Caseldine, 1983, 1985a, 1985b, 1988; Caseldine and Cullingford, 1981).

The snout of Gljúfurárjökull was located at altitude 580 m with an ELA ca. 960 m in 1985 (Caseldine, 1985b). The ELA for 48 glaciers in Tröllaskagi was higher, at 992 m (Caseldine and Stötter, 1993). The last publication about Gljúfurárjökull in the late 1980s (Caseldine, 1988) pointed out the end of its advance in 1986. On the other hand, the annual IGS marginal measurements show that the Gljúfurárjökull advance ended in the late 1980s; after that, it began to retreat again, by more than 160 m between 1989 and 2013.

The relationship between the climate and glacier response (glacier termini, mass balance) was first studied in Iceland by Björnsson (1971), who proposed 8°C (at Akureyri) as the Ts threshold which would reflect the change in the mass balance sign in the Tröllaskagi glaciers. The temperature evolution in Iceland was studied by Einarsson (1991), who differentiated six thermal phases between 1901 and 1990, depending on whether these were cold (1901–1925, 1947–1952 and 1965–1971) or warm (1926– 1946, 1953–1964 and 1972–1990).

The study of the above debris-free glaciers in Tröllaskagi enabled the impact of climate change in northern Iceland to be monitored for decades and compared with the evolution of the large ice-caps in central and southern Iceland. The aim of this article is to extend the previous work and record the evolution of the Gljúfurárjökull and Tungnahryggsjökull glaciers up to the present, focusing on termini retreat, area/volume loss and ELA variation in these glaciers. In addition, the trends and relationships of these parameters with climate evolution will be analysed.

## Regional settings

The limits of the Tröllaskagi peninsula in north Iceland are Skagafjörður to the west and Eyjafjarðardalur to the east (Figure 1) between meridian 19°30′W and 18°10′W, jutting out into the North Atlantic to latitude 66°12′N and linked to the central highlands to the south. The peninsula consists of over 4000 km$^2$ of tertiary flat-summit highlands and crests at 1000–1400 m composed of jointed basaltic lava flows often separated by 30–50 cm lithified sedimentary horizons (Jóhannesson and Sæmundsson, 1989; Sæmundsson et al., 1980). The highlands are cut by deeply entrenched valleys with steep slopes and sheer headwalls. These headwall areas host 167 small cirque glaciers (Icelandic Meteorological Office, 2015), a few of which are debris-free and the most sensitive to climatic fluctuations (Häberle, 1991; Kugelmann, 1991).

The Tröllaskagi corrie-glaciers are found in north-facing cirques resulting from the leeward accumulation of snow blowing from the plateau areas (Caseldine and Stötter, 1993) and the solar radiation shadow. Most of the glaciers are debris-covered and rock glaciers because of significant slope activity. The insulating effect of the debris cover makes them static and less sensitive to climate variations (Martin et al., 1991).

The 1961–1990 weather data series show a MAAT of 2–4°C on the Tröllaskagi coasts and −2° to −4°C on the summits (Etzelmüller et al., 2007). At Akureyri (1901–1990), MAAT is 3.4°C, while mean summer (June–August) and Ts reach 9.9°C and 8.4°C, respectively (Einarsson, 1991); in winter (January–March), the mean value drops to −1.6°C (Einarsson, 1991), although it can be higher if the October–April period is considered, with −0.3°C. Precipitation in the Tröllaskagi area oscillates between 400 mm in some lowland areas of Skagafjörður and Eyjafjörður and up to 2500 mm on the summits (1971–2000 data series; Crochet et al., 2007). Two weather stations have been used for the analysis, one of which is located in the capital town of northern Iceland, Akureyri (65°41′N; 18°06′W; 23 m a.s.l.), in inner Eyjafjörður; and the other on the mountain road at Öxnadalsheiði (65°28′N; 18°41′W; 540 m a.s.l.) in southern Tröllaskagi.

Gljúfurárjökull and Tungnahryggsjökull have been selected as the largest glaciers with no superficial debris cover in the study area (Figure 1), with areas between 4 and 9 km$^2$ (Tables 2 and 4). This feature makes them optimal for assessing their level of susceptibility to climatic variations.

## Methods

Glacier monitoring and spatial analysis operations were performed with ArcGIS (ESRI), using analysis of aerial photographs (≈1:30,000 scale) from 1946, 1985, 1994 and 2000 (National Land Survey of Iceland, 2015). A 2005 SPOT satellite image was also used. The glaciers were delimited at different dates by photointerpretation and georeferencing (RMS error 3.1–5.9 m) of the aerial photographs and the previously georeferenced satellite images.

ArcGIS was used to calculate the area and retreat of the glaciers at different dates. The glacier extent during the LIA maximum was delimited over the position of the morainic ridges dated to the late LIA (end of the 19th century) in the bibliographic references cited in section 'Introduction' (Caseldine, 1983, 1985a, 1985b; Caseldine and Cullingford, 1981; Caseldine and Stötter, 1993). The glaciers and moraines were mapped in early publications after previous fieldwork aided by several techniques such as triangulation, tachometry and photogrammetry (Caseldine and Cullingford, 1981), which enabled the authors to make an accurate mapping and contouring of the glacier surface. Regarding dating methods, lichenometry was used in previous works to date the most recent and LIA maximum moraines (Caseldine, 1983, 1985a, 1985b; Caseldine and Cullingford, 1981; Caseldine and Stötter, 1993; Kugelmann, 1991), in combination with radiocarbon (Häberle, 1991), tephrochronology and Schmidt hammer (Caseldine, 1987), where lichenometry was not applicable. Lateral moraines were used to reconstruct the glacier topography and ice thickness during the LIA maximum and for each date analysed.

The ELA of the glaciers was calculated automatically for each date available with the ArcGIS toolbox designed by Pellitero et al. (2015), implementing the accumulation area ratio (AAR; Brückner, 1886, 1887) and the area altitude balance ratio (AABR; Osmaston, 2005) methods. For the AAR method, the ratio 0.67 was applied, previously used by Caseldine and Stötter (1993), as the results obtained in Tröllaskagi were similar to the maximum elevation of the lateral moraines (Stötter, 1990). For the AABR method, the balance ratio (BR) of 1.5 ± 0.4 proposed by Rea (2009) as representative for Norwegian glaciers was used.

The results obtained from the glacial remote studies (ELA depressions) and the climatological data were used to estimate the temperatures and paleotemperatures at the snouts and the ELA, assuming a lapse rate of 0.66°C $100^{-1}$ m. To estimate the current precipitation and the paleoprecipitation, glacio-climatic models were used (Ballantyne, 1989; Braithwaite, 2008; Ohmura et al., 1992). These relate variables such as MAAT or Ts with precipitation or ablation, measured on an annual, seasonal or daily scale. The first model used is based on an exponential relationship existing between the mean ablation season temperature and winter accumulation at the ELA of Norwegian glaciers (Liestøl, 1967; Sutherland, 1984), expressed by equation (1) established by Ballantyne (1989), and later applied by Dahl and Nesje (1992) in southern Norway and by Caseldine and Stötter (1993) in Tröllaskagi

$$A = 0.915e^{0.339Ts} \; (r^2 = 0.989; p < 0.0001) \qquad (1)$$

where $A$ is the winter accumulation (October–April) in metres of water equivalent, and $Ts$ is the mean ablation season temperature (May–September) in degree Celsius.

The second model used was proposed by Ohmura et al. (1992), defined by equation (2), of the best fit polynomial curve obtained through the regression analysis between the mean temperature of the three summer months (June, July and August) and total annual precipitation (winter balance plus summer precipitation) at the ELA for a dataset of 70 glaciers worldwide:

$$P = 645 + 296T + 9T^2 \qquad (2)$$

where $P$ is the total annual precipitation (in mm water equivalent) and $T$ is the mean temperature of the three summer months (in °C). Standard error is 200 mm.

The last method used was the 'degree-day' model (Braithwaite, 2008; Brugger, 2006), based on the existing proportionality between snow or ice melt (ablation, expressed in mm of water equivalent) and the sum of temperatures above freezing point (degree-day sum). The quotient of the two variables obtains the degree-day factor ($d_f$) ratio (expressed in mm day$^{-1}$ °C$^{-1}$). The value for $d_f$ used was the average 4.1 mm day$^{-1}$ °C$^{-1}$, obtained by Braithwaite (2008) from 66 of the 70 glaciers in the dataset included in Ohmura et al. (1992). According to this approach, equation (3) shows that melt ($M_d$) only occurs with positive temperatures, so that:

$$M_d = d_f T_d \qquad \text{when} \quad T_d > 0\ ºC;$$

$M_d = 0$ when, $T_d \leq 0$ ºC (3)

where $M_d$ is the daily melt (in mm water equivalent) and $T_d$ is the mean daily temperature (in °C), obtained from equation (4), based on a sinusoidal distribution throughout the year around the mean temperature so that:

$$T_d = A_y \sin\left(\frac{2\pi d}{\lambda} - \phi\right) + MAAT \qquad (4)$$

where $A_y$ is the amplitude (calculated as half of the annual temperature range), $d$ is the Julian day (1–365), $\lambda$ is the period (365 days), $\phi$ is the phase angle (1.93 to reflect that January is the coldest month). *MAAT* is expressed in °C.

The result of applying the model is the total annual ice or snow melt (in mm of water equivalent), which is equivalent to the accumulation, given that the mass balance at the ELA is 0.

Calculations of glacial volume were then performed following the indications by Bahr et al. (1997), according to which the volume ($V$) of any glacier is related to its surface area ($S$) using exponential equation (5) based on a dimensionless scaling exponent ($\gamma$) which includes the morphometric characteristics of the glacier (width, slope, side drag and mass balance):

$$V \propto cS^{\gamma} \qquad (5)$$

where c is an empirical power law coefficient of 0.2055 (expressed in units of $m^{3-2\gamma}$), derived from Chen and Ohmura (1990), and γ is derived from equation (6):

$$\gamma = 1 + \frac{1+m+n(f+r)}{(q+1)(n+2)} \qquad (6)$$

where $q = 0.6$, $m = 2$ and $f = 0$ are obtained from empirical data, and where $r$ may be 0 for steep surface slopes, or $r = (1 - m + n - nf) / (2\ (n + 1))$ for gentle slopes; and $n = 3$. So $\gamma$ applied is in turn 1.375. Further details on the mathematical basis and its application can be found in Bahr et al. (1997) and Radić et al. (2007).

In parallel to studying the glaciers, the climatological data series from the two meteorological stations presented above was statistically processed. The series from Akureyri were used because it is the longest at the study area, with data recorded since 1882 (Icelandic Meteorological Office, 2015), and from Öxnadalsheiði because it is close to the study area and is located at a higher altitude, with a useful period from 2000 to 2014 (Icelandic Road and Coastal Administration, 2016). The statistical processing analysed the running-means (5 years) and correlation between the two meteorological stations. The correlation analysis was used to reconstruct the MAAT at the Öxnadalsheiði meteorological station using the regression equation obtained.

# Results

## Evolution of the glacier snouts

Studies of aerial photographs and satellite images show that the glacier snouts have retreated by more than 1300 m on average since the LIA maximum (considered to be AD 1898 in Gljúfurárjökull and AD 1868 in both Western and Eastern Tungnahryggsjökull as explained in the publications presented in section ‘Introduction’; Figure 2), with an altitudinal rise of more than 100 m. The retreat accelerated rapidly (15.3 m $yr^{-1}$) during the first half of the 20th century (Figure 2). In the second half of the 20th century, the retreat decelerated considerably, reflected in the lowest values around 1985 (5.2 m $yr^{-1}$) and a trend shift in 1994, with an advance observed in Gljúfurárjökull. The trend then altered again and Gljúfurárjökull retreated in the years 1994–2005.

During the period 1898–1946, the snout of Gljúfurárjökull retreated 635 m, almost two-thirds of the total distance from the LIA maximum (1898–1903) to 2005 (Figures 2 and 3), at an average rate of 13.2 m $yr^{-1}$ (Table 1). The rise of the snout during that period (46 m) was almost half of the total rise. By 1985, the retreat and ascent since 1898 was almost the total for the 1898–2005 period. However, the velocity of the retreat in 1946–1985 was lower than in 1898–1946. The 1994 aerial photograph reveals a change in this trend, with a snout position 20 m more advanced compared with 1985 (Figure 2). Nevertheless, from 2000 onwards, there was a slow but continuous retreat.

The trend in Western Tungnahryggsjökull during the first half of the 20th century was a more rapid retreat, showing the highest average rates of the whole period (19.5 m $yr^{-1}$). By 1946, this glacier had retreated almost 90% of the total recorded between the LIA maximum (1868) and 2005 (Table 1). In the 1946 photograph, this significant retreat of the ice reveals two large moraines in the centre of the deglaciated area. The snout retreat slowed down considerably during the second half of the century, especially in 1985 (1.5 m $yr^{-1}$). By this date, the aerial photograph shows a complex terminus covered with debris, with an uneven retreat, from 60 m in the centre to 150–170 m on the margins, and a vertical rise of more than 200 m since 1946. The 1994 aerial photograph shows a similar snout, although with an advance in the western sector of ≈40 m and a retreat in the eastern sector of ≈20 m (Figure 2). In 2000, the snout, still covered with debris, retreated mainly in the centre. The glacier then continued to retreat, although more slowly than Gljúfurárjökull (6.4 m $yr^{-1}$) preserving the debris-covered snout (Figures 2 and 3).

Just as in the glaciers described above, the retreat of the Eastern Tungnahryggsjökull from its LIA position was more intense during the first half of the 20th century (Table 1), and in 1946 its snout was only 200 m from its current position. The snout then continued to retreat more slowly and by 1985 had already lost its most westerly tongue (Figure 2), where the margin retreated more than 400 m. The 2000 aerial photograph shows that an advance of at least 41 m had taken place since 1985. Nevertheless, between 2000 and 2005, the snout retreated 17 m, even more slowly than Western Tungnahryggsjökull.

**Evolution of ice – area and volume of the glaciers**

During the LIA maximum, the total surface area of the three glaciers exceeded 18 $km^2$, with almost half corresponding to the Western Tungnahryggsjökull. From then until 2005, the glaciers lost a quarter of their surface area (Table 2), with almost 20% lost during the first half of the 20th century. In 1985, the loss rate was considerably reduced, and slight increases in the surface area of Gljúfurárjökull and Western Tungnahryggsjökull occurred in 1994 (Table 2; Figure 2). Since 2000, the surface loss of the glaciers has not reached 2%. A third of the LIA maximum ice volume had been lost by 2005. The greatest volume loss (25%) occurred between the LIA maximum and 1946. The most intense volume loss rate was in Western Tungnahryggsjökull, around 2.5 $km^3$ $yr^{-1}$ $10^{-3}$ (Table 3; Figure 4). During the second half of the 20th century, the losses were lower, with maximum average 7.8% (Table 3), although in 1985 the Eastern Tungnahryggsjökull had lost around 15% compared with 1946. In 1994, the Gljúfurárjökull and Western Tungnahryggsjökull volumes increased. However, the reduction in volume continued from 2000 onwards; the loss rate intensified during 2000–2005 with values similar to or even higher than those of the first half of the 20th century (Table 3).

## Evolution of the ELA

Applying the AAR and AABR methods obtains a mean ELA of ≈1010 m during the LIA maximum and a rise of 40–50 m in the period analysed between LIA maximum and 2005 (Table 4). Using the AAR method, the greatest rise in the ELA (29 m) occurred between the LIA maximum and 1946, coinciding with the most important snout retreat and the greatest surface area and volume losses. From 1946 to 1985, there was a smaller rise (10 m), slightly more intense in Gljúfurárjökull and Western Tungnahryggsjökull. Although 1994 showed a trend shift advancing, the ELAs for the two glaciers remained stagnant. Since 2000, the ELA has remained practically stable around 1050 m, Nevertheless, a sharp intensification can be clearly seen in the ELA rise ratio in the last period 2000–2005, with a mean rate higher than in the period between the LIA maximum and 1946 (Table 4). The results obtained using the AABR method were reasonably close to those obtained using the AAR method, with maximum differences of ±10 m (Table 4).

## Climate evolution

The MAAT calculated for Akureyri (1882–2014) and Öxnadalsheiði (2000–2014) data series was 3.35°C and 0.97°C, respectively. A lapse rate of 0.66°C 100 m$^{-1}$ was obtained from the MAAT data series for the common period 2000–2014. Regression analysis of the two MAAT series for the period 2000–2014 showed strong correlation ($r$ = 0.79; $n$ = 15), which enabled a first approximation of the Öxnadalsheiði series reconstruction for the period 1882–2000. The least squares equation used was $y = 0.9092x - 3.0223$ ($r^2 = 0.63$), with an overall average (MAAT = 0.02°C) only slightly different from the result obtained using extrapolation of the lapse rate (MAAT = −0.05°C).

Using the Akureyri temperature series and the 5-year running- means deviation compared with the overall series average, nine homogeneous periods were identified (Figure 5; Table 5). Thus, four cold periods with negative deviations (1882–1924, 1951–1955, 1966–1973 and 1979–1986) and five warm periods with positive deviations (1925–1950, 1956–1965, 1974–1978, 1987–1999 and 2000–2014). The MAAT was 2.5°C during the period 1882–1924, coinciding with the end of the LIA. This cold period ended in the early 1920s, with a sharp temperature rise and MAAT ≈4°C, maintained until the mid-1960s (Figure 5). This warm period was interrupted by brief cooling, with MAAT ca. 3.6°C, between 1950 and 1955. The temperature increase between the late LIA (1882–1924) and the period 1925–1950 was 1.4°C. However, the MAAT fell to 2.9°C between 1966 and 1986, marking the first cold period with negative deviations since the end of the LIA, interrupted by higher MAAT (3.6°C) between 1974 and 1978. Finally, the climate trend shifted again to warmer conditions during the period 1987–2014. Two sub-periods were identified; during the first period (1987–2002), the MAAT was 3.8°C, while in 2003 an abrupt warming of 0.6°C occurred, marking the onset of the warmest period with an MAAT of 4.4°C from 2003 to 2014.

The Akureyri climatological log shows that increases in the MAAT and Ts were 1.9°C and 1.5°C, respectively, between the LIA and the present (Table 5). The average temperature calculated from the reconstructed Öxnadalsheiði data series suggests that the MAAT was below freezing level above 500 m a.s.l. in the interior of the Tröllaskagi peninsula during the late LIA cold period, 1966–1973 and 1979–1986.

As regards precipitation, the mean annual value for the 1950– 2014 period at Akureyri is 515 mm, while in the winter/accumulation season (October–April) it is 357 mm, that is, 69% of the total.

Running mean analysis (Figure 6c) showed below average values from the late 1950s to the late 1960s (ca. 250 mm minima) and from the late 1970s to the early 1980s (ca. 200 mm minima in 1980). From the mid-1980s onwards, winter precipitation has been above the average, reaching maxima in the early 1990s (ca. 430 mm). Since then, winter precipitation has been relatively regular and close to average, although an increasing trend started in the mid-2000s. Individual values peaked in 1989 and 2014, at over 550 mm.

The temperatures (mean annual, May–September, June–July– August) of the above logs, averaged for the phases identified, and later extrapolated to the successive ELAs through the lapse rate of 0.66°C 100 $m^{-1}$ (Table 6), were input into the glacio-climatic models (Ballantyne, 1989; Braithwaite, 2008; Ohmura et al., 1992). The Ballantyne (1989) model predicted winter precipitation of 2159 mm at the 2005 mean ELA which supposes an increase of 19% 100 $m^{-1}$ during the winter season if the Akureyri mean winter precipitation over the 30-year period 1976–2005 (357 mm) is taken into account. This result suggests an increase of ca. 50% (714 mm) in the winter precipitation compared with around 1445 mm during the late LIA (1882–1924). The results indicate that when the Ts increased by 1.2°C since the LIA, the winter precipitation increased by 714 mm at the ELA. In turn, the Ohmura et al. (1992) model, with an estimated annual precipitation of 2022 mm at the 2005 mean ELA, suggests a smaller increase compared with the late LIA (1578 mm) of around 28% (444 mm). If the estimated annual precipitation for 2005 is compared with that recorded in Akureyri in the period 1976–2005 (517 mm), the pluviometric gradient obtained is lower, with an increase of 14% 100 $m^{-1}$. Finally, the degree-day model (Braithwaite, 2008) obtained 1741 mm annual precipitation at the ELA in 2005 (Table 6), an increase of 518 mm (42%) since the LIA (1223 mm). Using the degree-day model, the vertical pluviometric gradient obtained was lower than with the other models, with 13% 100 $m^{-1}$. This model also estimated precipitation much more sensitive to temperature variations (Table 6). For example, the degree-day model estimates a reduction of 565 mm in precipitation from 1946 to 1985, because of a drop of 1.1°C in the MAAT between these two dates.

From the LIA to the present day, MAAT and Ts at the ELA increased by 1.7°C and 1.2°C, respectively, while in Akureyri the rise was 1.9°C and 1.5°C, respectively. These values are much higher than the temperature increase deduced from the rise of the ELA (0.3°C). For most of the dates, the MAAT in the glacier snouts remained close to freezing level. In Tungnahryggsjökull, the MAAT of the snouts was below freezing level at all the dates and fell below −1°C and −2°C in the coldest periods of the LIA and 1985, respectively. In the Gljúfurárjökull snout, apart from these cold periods, the MAAT was positive, around 0.5–0.6°C (Table 7).

## Discussion

### Interpretation of the results

The results of this research show a gradual climate warming from the end of the LIA, as well as a regressive trend for the northern Iceland glaciers. This process was not uniform, with considerable temperature variations in this region (Einarsson, 1991) which led to important changes in the debris-free glaciers studied.

The most important retreat of the Tröllaskagi glaciers between LIA maximum and the present occurred during the first half of the 20th century. The study of the three glaciers presented here shows that most of the glacier snout retreat, area reduction and volume loss had already occurred by 1946; a similar trend was observed at southeast Vatnajökull outlet glaciers, whose volume loss before 1945 represented the half of the post-LIA total loss (Hannesdóttir et al., 2015). This is reflected in the combination of the field measurements carried out by Caseldine (1983) and the IGS at Gljúfurárdalur. However, the figures are different (Table 8). Our remote measurements on the aerial photographs show that Gljúfurárjökull retreated 635 m in the period 1898–1946. On the other hand, the retreat reported by Caseldine (1983) would be at least 450 m if the retreat from the LIA maximum to 1915–1917 (>250 m) and the retreat during the 1930s (200 m) are considered. The key to this glacier response is found in four main factors: (1) the sharp 1.4°C rise of the MAAT and 1.2°C rise in the Ts (Akureyri) between the cold period at the end of the LIA (1882–1924) and the warm period 1925–1950 (Figure 5). (2) Warm conditions with MAAT ≈4°C and Ts = 9°C were maintained between 1925 and 1950 (Böðvarsson, 1955). (3) The predominant south-westerly airflow after 1920 proposed by Kirkbride (2002), which kept summers warm and caused increased ablation. (4) Other later cold periods did not last longer than 10 years (Caseldine, 1985b). This sharp increase in temperature triggered an ELA rise of ≈30 m compared with the ELA during the LIA maximum. Increased winter precipitation from the LIA maximum (Table 6) did not appear to have a major impact on the termini variation at that moment, but probably in further advances (e.g. mid-1970s to mid-1980s, or early 1990s) by increasing the mass flux and reducing the termini retreat rate (Kirkbride, 2002). In this context, the Western Tungnahryggsjökull glacier seems to be the most sensitive to the increased temperature of the three glaciers, as it presents the highest values for retreat rates, area and volume losses, and the greatest ELA rise.

Stötter et al. (1999) indicate that the coldest period after the LIA was from the early 1960s to the mid-1970s, when temperatures fell to levels equivalent to the warmest recorded in the 19th century. This cooling is the reason given by Caseldine (1983, 1985a, 1985b, 1988) to explain the advance of the Gljúfurárjökull between the mid-1970s and the mid-1980s, which can be clearly seen in Figure 6. This would suggest a time response to Ts cooling close to 10 years. The retreat from 1946 to 1985 calculated using IGS field measurements (322 m) appears to be overestimated if we consider our results of 275 m for the same period (Table 8). This discrepancy can be explained by technical issues such as the accuracy of the georeferencing (RMS error), the change in field measurement procedures (estimates; see Sigurðsson et al., 2007) and the vague and scarce (incomplete) data about termini variations prior to the 1950s provided by Caseldine and Cullingford (1981) and Caseldine (1983). So the GIS measurements over a photograph with snow-free termini (taken at the end of the ablation season) and properly georeferenced can provide the best results, avoiding estimates when fieldwork is not possible. In this paper, two points are mentioned which may clarify the glacial evolution after the 1980s: (1) the 1994 aerial photograph reveals a more advanced position of the Gljúfurárjökull compared with 1985 and (2) between 1979 and 1986 another cold period is identified (with temperatures not as cold as in the previous one, but separated from it by a brief warm 4-year period), characterized especially by a fall in the Ts below 8.5°C and even 8°C in the Akureyri station (Figures 5 and 6b). This cold period between 1979 and 1986 seems to have been the continuation of the cooling which started in the early 1960s and the reason why Gljúfurárjökull continued to advance after 1985. However, by the year 1994 the advance had ended, because of (1) the low advance rate of Gljúfurárjökull inferred from the positions of the snout in 1985 and 1994 (2.2 m $yr^{-1}$) compared with the advance velocities occurring in previous years (Caseldine, 1983, 1985a, 1988; Caseldine and Cullingford, 1981) and (2) the time lag (8 years) from 1994 to the end of the cooling in 1986. These findings confirm the Ts value of 8–8.5°C at Akureyri proposed by Björnsson (1971) and Caseldine (1985b) as the threshold for the trend shift in the glacial mass balance and also suggest that less than 10 years with cold summers may be required for the glacier advance. However, the increase in winter precipitation (671 mm; see Supplementary Material, available online) obtained in this study at the

Gljúfurárjökull ELA between 1985 and 1994, and that obtained in Akureyri also seem to explain the glacier advance during the 1990s when the Ts in Akureyri reached over 8.5°C. It is reasonable to assume that a greater increase in winter precipitation reduced the number of cold summers required for the glacier snout advance. Such a clear advance was not observed in 1994 in the Western Tungnahryggsjökull, because of the difficulty in identifying precisely the lateral margins of the glacier in snow-covered areas. However, different sectors of the snout advanced or retreated compared with 1985. The explanation may be found in the uneven debris cover and the insulating effect it exerted on the ice. From the mid-1980s, there was a gradual rise in the Ts (Caseldine, 1988), which triggered the retreat of the glaciers in 2000 from their position in 1994. A sharp temperature rise occurred around the year 2003, which intensified the reduction in glacial volume and the ELA rise at rates comparable to those in the first half of the 20th century. Nevertheless, the snout retreat did not accelerate dramatically. The retreat rate intensified in the period 2000–2005 compared with 1994–2000, but did not reach the rates recorded before 1946 (Table 1). The glacier evolution in recent years is characterized by continuous retreat, which can be explained by the high Ts above 9°C since 2003. According to Caseldine and Stötter (1993), the effect of the climate warming observed from the LIA to the mid-1980s was a 50-m ELA rise in the glaciers in northern Iceland. This value is similar to the 40–50 m ELA rise obtained in this study for Gljúfurárjökull and Tungnahryggsjökull between the LIA maximum and 2005. The AAR (0.67) and AABR (1.5) methods applied in this paper to calculate the ELAs obtained homogeneous results, suggesting a good adaptation of the application of AABR = 1.5, representative of the Norwegian glaciers (Rea, 2009), to the debris-free glaciers of northern Iceland.

**Climatic implications**

According to Caseldine and Stötter (1993), although the ELA is the parameter which best expresses the relationship between glaciers and climate, the use of its rise or fall to estimate Ts variations may lead to significant underestimation in the results. This has been proved in this study, observing that if the maximum depression of the ELA (41 m) and the lapse rate of 0.66°C 100 $m^{-1}$ are taken into consideration, the rise in the ELA would indicate a lower Ts increase, approximately 0.3°C, assuming the precipitation remained constant. However, the Ts rise recorded in Akureyri (1.5°C) or the rise extrapolated at the ELA (1.2°C) between the LIA maximum and 2005 is considerably higher. This shows that in addition to temperature, other factors may have been decisive in the glacial evolution, such as precipitation and wind (Caseldine and Stötter, 1993).

The model applied by Caseldine and Stötter (1993) and Stötter et al. (1999) suggests that the precipitation in northern Iceland during the LIA was significantly lower than at the present day. Applying the same model in this study shows a similar trend. The model designed for Norwegian glaciers (Ballantyne, 1989) predicted winter precipitation of 1445 mm at the mean ELA (at altitude 1010 m) during the LIA maximum, and an increase of more than 714 mm (twice the modern winter precipitation at Akureyri) from then to 2005. Caseldine and Stötter (1993) estimated practically identical precipitation at the ELA during the LIA (1450 mm) and an increase of around 600 mm until the mid-1980s for the Tröllaskagi glaciers. Dahl and Nesje (1992) using the same model calculated a relatively similar increase of 690 mm in Nordfjord (Western Norway) since the LIA, where the current climate is wetter and milder (Olden, 78 m a.s.l., Ts = 12.2°C, winter precipitation = 812 mm; see Dahl and Nesje, 1992) because of the influence of the North Atlantic Drift and the frequent frontal precipitation associated with the polar front position. The maritime location of the Tröllaskagi glaciers and those (Norwegian) used to devise the Ballantyne (1989) model, and also the good adaptation of a Norwegian AABR for ELA calculation, may postulate this model as the most suitable of the three to infer temperature and precipitation changes in the Tröllaskagi glaciers. This model gives higher values of precipitation than the other models (e.g. Ohmura et al., 1992) at warmer dates (e.g. 1946, 2000 and 2005) because of the exponential nature of its formula (see equation (5) in section 'Methods'). This determines that equal values of temperature as input will give higher output precipitation in the Norwegian model than the Ohmura et al. (1992) one. Only at the coldest dates (e.g. LIA maximum and 1985) were the results inverse with higher precipitation in the Ohmura et al. (1992) model, when the Ts was far below the mean 3-month summer temperature.

The precipitation pattern observed in this paper, lower during the LIA cold period and higher during the warm periods, fully coincides with the model proposed by Stötter et al. (1999) for northern Iceland. This is also coherent with a lower ocean surface temperature (Geirsdóttir et al., 2009) linked to the greater presence of Arctic sea ice (Ogilvie, 1984, 1996; Ogilvie and Jónsdóttir, 2000; Ogilvie and Jónsson, 2001) which weakened the convective processes (Lehner et al., 2013). Nesje and Dahl (2003) and Holmes et al. (2016) link the precipitation changes to the variations in the North Atlantic Oscillation (NAO) phase (Hurrell, 1995) and in the position of the polar front. In this sense, the dates at which high precipitation was obtained in this research (e.g. 1946, 2005) would correspond to a positive NAO phase (Figure 6d) which reinforced the zonal flow of the westerlies and the W–SW winds, coinciding with a northwards displacement of the low-pressure cells and the polar front (Jansen et al., 2016). This situation facilitated the predominance of warm wet sub-tropical masses responsible for warm wet winter weather in Iceland (Holmes et al., 2016). On the contrary, the dates when the calculated precipitation was lowest (e.g. LIA, 1985) would have coincided with the negative NAO phases (Figure 6d) in which Arctic air masses predominated as a result of the southward displacement of the polar front and prevailing N–NW winds (Holmes et al., 2016). This atmospheric configuration would favour cold dry summers (Jansen et al., 2016). Thus, it is reasonable to suppose that the variations in precipitation between the cold and warm periods may also be explained by conditions that either hindered or facilitated convection, respectively (Burn et al., 2016). However, extra accumulation from snow-blowing should also be taken into account in the corrie glaciers studied. In a deeply incised cirque surrounded by a plateau, the snow may deflate from the plateau and accumulate in the cirque, either by direct accumulation or avalanching from the cirque walls (Dahl and Nesje, 1992; Sissons and Sutherland, 1976; Sutherland, 1984). Although most Tröllaskagi glaciers are surrounded by sharp peaks, ridges and summits, they receive snow blown from far out on the plateau mountains. Caseldine and Stötter (1993) suggested that up to 35% of the total winter accumulation could be attributed to the processes explained above (Tangborn, 1980). Based on this relationship between accumulation and snow-blowing, previous authors (Caseldine and Stötter, 1993; Dahl and Nesje, 1992) proposed that the changes in winter accumulation may also reflect changes in the direction of the prevailing wind. According to this reasoning, the increase in precipitation between the LIA and 2005 could be explained by a current predominance of the wind from the plateau (with snow-blowing), coherent with the changes in atmospheric circulation explained above. However, the results from the wind data processing (see Supplementary Material, available online) do not provide strong support for the changes in wind directions based on differences of winter accumulation, at least in the present day, with NW–NE (36%) and SW–SE (35%) as the dominating wind directions above 10 m $s^{-1}$ during winter (October–April) at Grimsey. Further research on wind and snowfall is required to shed light on this issue.

The NAO exerts control over mass balance by influencing temperature and precipitation anomalies (Marzeion and Nesje, 2012), and therefore, a link between NAO phases and termini variations has been suggested on the literature. Bradwell et al. (2006) found that Lambatungnajökull (north-eastern outlet of Vatnajökull, South Iceland) advanced during negative NAO phases and linked this to positive mass balances. On the contrary, Nesje et al. (2000) linked negative mass balance with negative NAO indices. Nevertheless, such relationships are not so clear, at least in Gljúfurárjökull. The continuous retreat from 1950 until the late 1970s in Figure 6a is mostly characterized by negative NAO indices, so it is reasonable to think that, at least for that period, there may have been a link between negative NAO index and negative mass balance. However, this relationship during the 1980s advance is less clear as it coincides with a negative NAO phase interrupted by 3 years with positive indices (1981, 1983 and 1984). We found that this advance started in the mid-1970s, that is, 20 years after the major reversal of the NAO index mode (negative to positive) occurred in 1955 (see Figure 6), in good agreement with the reaction times at the decadal scale reported by Kirkbride (2002) and the theoretical calculations of glacier response times for small mountain glaciers (Jóhannesson et al., 1989). The glacier termini/winter–NAO relationship at this glacier is even fuzzier if we consider the advanced position in 1994 coinciding with positive indices (since 1989) and the retreat since the late 1990s over two positive and negative NAO phases. Furthermore, fuller and more detailed research on the past and future behaviour of the glacier termini and mass balance is needed to determine the future behaviour of the glaciers and elucidate the relationship with the NAO.

This research has shown the high sensitivity of the debris-free Gljúfurárjökull and Tungnahryggsjökull glaciers to climatic fluctuations (Häberle, 1991; Kugelmann, 1991), especially to the *Ts* (Eythorsson, 1935; Liestøl, 1967; Ohmura et al., 1992). Consequently, they experienced an important retreat during the periods characterized by warm summers and advanced during the short periods with cold summers, even when the duration of these periods was shorter than the 10 years proposed by Caseldine (1985b).

## Conclusion

The debris-free Gljúfurárjökull and Tungnahryggsjökull are important indicators of climate change, as the absence of debris and reduced dimensions mean they are highly sensitive to climate fluctuations. As a result, the abrupt climatic transition of the early 20th century and the 25-year warm period 1925–1950 triggered the most important glacier retreat and volume loss since the end of the LIA; meanwhile, cooling during the 1960s, 1970s and 1980s altered the trend, with glacier snout advances.

Calculating the ELAs for the Tröllaskagi glaciers using the AAR and AABR methods showed a good fit of the AABR = 1.5 proposed for Norwegian glaciers. Analysis of the relationships between ELA evolution and climatic data also revealed that the glacier response depends not only on the Ts but also on other factors such as precipitation. The models applied, especially the one obtained from Norwegian glaciers, show a precipitation increase of more than 700 mm since the LIA, compatible with an increase in the surface temperature of the North Atlantic and with a change in the direction of the prevailing wind, currently from the plateau. Nevertheless, the evolution of the glaciers in the last 10 years shows an uncertain trend because of the lack of updated data (except for Gljúfurárjökull), which may become clearer with further monitoring of the glaciers over the coming years. The relationship between glacier evolution and atmospheric circulation patterns remains unclear.

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**Table 1. Glacier advance/retreat and snout elevation shift from the LIA maximum. Values in bold represent glacier advances.**

| **Distance from the LIA maximum position (m)** | | | | | | | |
|---|---|---|---|---|---|---|---|
| Glaciers | LIA | 1946 | 1985 | 1994 | 2000 | 2005 | Total retreat |
| Gljúfurárjökull | - | 635 | 910 | 890 | 916 | 993 | 993 |
| Tungnahryggsjökull (W) | - | 1524 | 1584 | 1610 | 1703 | 1735 | 1735 |
| Tungnahryggsjökull (E) | - | 1027 | 1298 | - | 1257 | 1274 | 1274 |
| *Average* | - | *1062* | *1264* | - | *1292* | *1334* | *1334* |
| **Advance/retreat rate (m $yr^{-1}$)** | | | | | | | |
| Glaciers | LIA | LIA-1946 | 1946-1985 | 1985-1994 | 1994-2000 | 2000-2005 | LIA-2005 |
| Gljúfurárjökull | - | -13.2 | -7.1 | **2.2** | -4.3 | -15.4 | *-9.3* |
| Tungnahryggsjökull (W) | - | -19.5 | -1.5 | -2.9 | -15.5 | -6.4 | *-12.7* |
| Tungnahryggsjökull (E) | - | -13.2 | -6.9 | - | - | -3.4 | *-9.3* |
| *Average* | - | *-15.3* | *-5.2* | - | - | *-8.4* | *-10.4* |
| **Glacier snouts elevation (m a.s.l.)** | | | | | | | |
| Glaciers | LIA | 1946 | 1985 | 1994 | 2000 | 2005 | ↑LIA-2005 |
| Gljúfurárjökull | 512 | 558 | 594 | 591 | 593 | 622 | 110 |
| Tungnahryggsjökull (W) | 540 | 741 | 786 | 759 | 779 | 793 | 253 |
| Tungnahryggsjökull (E) | 597 | 679 | 718 | - | 705 | 711 | 114 |
| *Average* | *550* | *659* | *699* | - | *692* | *709* | *159* |

LIA: 'Little Ice Age'.

**Table 2. Ice surface evolution from the LIA maximum in the Gljúfurárjökull and Tungnahryggsjökull glaciers. Values in bold represent area gains.**

| **Area (km$^2$)** | | | | | | | |
|---|---|---|---|---|---|---|---|
| Glaciers | *LIA* | *1946* | 1985 | 1994 | 2000 | 2005 | ↓LIA-2005 |
| Gljúfurárjökull | 4.372 | 3.540 | 3.407 | 3.413 | 3.384 | 3.335 | 1.038 |
| Tungnahryggsjökull (W) | 8.735 | 6.939 | 6.689 | 6.700 | 6.604 | 6.512 | 2.222 |
| Tungnahryggsjökull (E) | 5.348 | 4.532 | 4.035 | - | 4.054 | 3.940 | 1.408 |
| *Total* | *18.455* | *15.011* | *14.131* | - | *14.041* | *13.787* | *4.668* |
| **Area gain/loss rate (km$^2$ yr$^{-1}$)** | | | | | | | |
| Glaciers | LIA | LIA-1946 | 1946-1985 | 1985-1994 | 1994-2000 | 2000-2005 | ↓LIA-2005 |
| Gljúfurárjökull | - | -0.017 | -0.003 | **0.001** | -0.005 | -0.010 | -0.010 |
| Tungnahryggsjökull (W) | - | -0.023 | -0.006 | **0.001** | -0.016 | -0.018 | -0.016 |
| Tungnahryggsjökull (E) | - | -0.010 | -0.013 | - | - | -0.023 | -0.010 |
| **Area gain/loss (%)** | | | | | | | |
| Glaciers | LIA | LIA-1946 | 1946-1985 | 1985-1994 | 1994-2000 | 2000-2005 | ↓LIA-2005 |
| Gljúfurárjökull | - | -19.04 | -3.75 | **0.18** | -0.86 | -1.46 | -23.73 |
| Tungnahryggsjökull (W) | - | -20.55 | -3.61 | **0.16** | -1.43 | -1.39 | -25.44 |
| Tungnahryggsjökull (E) | - | -15.27 | -10.96 | - | - | -2.79 | -26.33 |
| *Total* | - | *-18.67* | *-5.86* | - | - | *-1.81* | *-25.29* |

**Table 3. Ice volume evolution from the LIA maximum in the Gljúfurárjökull and Tungnahryggsjökull glaciers. Values in bold represent volume gains.**

| **Volume ($km^3$)** | | | | | | | |
|---|---|---|---|---|---|---|---|
| Glaciers | LIA | 1946 | 1985 | 1994 | 2000 | 2005 | ↓LIA-2005 |
| Gljúfurárjökull | 0.278 | 0.208 | 0.197 | 0.198 | 0.195 | 0.191 | 0.086 |
| Tungnahryggsjökull (W) | 0.720 | 0.524 | 0.498 | 0.500 | 0.490 | 0.481 | 0.239 |
| Tungnahryggsjökull (E) | 0.367 | 0.292 | 0.249 | - | 0.250 | 0.241 | 0.126 |
| *Total* | *1.364* | *1.024* | *0.944* | - | *0.936* | *0.913* | *0.451* |
| **Volume gain/loss rate ($km^3\ yr^{-1}\ 10^{-3}$)** | | | | | | | |
| Glaciers | LIA | LIA-1946 | 1946-1985 | 1985-1994 | 1994-2000 | 2000-2005 | ↓LIA-2005 |
| Gljúfurárjökull | - | -1.459 | -0.273 | **0.055** | -0.387 | -0.781 | -0.808 |
| Tungnahryggsjökull (W) | - | -2.502 | -0.663 | **0.125** | -1.632 | -1.863 | -1.744 |
| Tungnahryggsjökull (E) | - | -0.958 | -1.104 | - | - | -1.912 | -0.918 |
| **Volume gain/loss (%)** | | | | | | | |
| Glaciers | *LIA* | LIA-1946 | 1946-1985 | 1985-1994 | 1994-2000 | 2000-2005 | ↓LIA-2005 |
| Gljúfurárjökull | - | -25.21 | -5.12 | **0.25** | -1.17 | -2.00 | -31.10 |
| Tungnahryggsjökull (W) | - | -27.12 | -4.93 | **0.23** | -1.96 | -1.90 | -33.22 |
| Tungnahryggsjökull (E) | - | -20.38 | -14.76 | - | - | -3.82 | -34.30 |
| *Total* | - | *-24.92* | *-7.77* | - | - | *-2.43* | *-33.08* |

LIA: 'Little Ice Age' maximum.

**Table 4. ELAs and ELA changes over variable periods calculated by AAR and AABR methods for the Gljúfurárjökull and Tungnahryggsjökull glaciers.**

| **ELA-AAR (0.67)** (m a.s.l.) | | | | | | | |
|---|---|---|---|---|---|---|---|
| Glaciers | LIA | 1946 | 1985 | 1994 | 2000 | 2005 | ↑LIA-2005 |
| Gljúfurárjökull | 954 | 974 | 985 | 982 | 984 | 988 | 34 |
| Tungnahryggsjökull (W) | 1046 | 1082 | 1092 | 1090 | 1091 | 1094 | 48 |
| Tungnahryggsjökull (E) | 1029 | 1061 | 1069 | - | 1071 | 1073 | 44 |
| *Average* | *1010* | *1039* | *1049* | - | *1049* | *1052* | *42* |
| ELA-AAR rise rate ($m\ yr^{-1}$) | | | | | | | |
| Glaciers | LIA | LIA-1946 | 1946-1985 | 1985-1994 | 1994-2000 | 2000-2005 | ↑LIA-2005 |
| Gljúfurárjökull | - | 0.42 | 0.28 | -0.33 | 0.33 | 0.80 | 0.32 |
| Tungnahryggsjökull (W) | - | 0.46 | 0.26 | -0.22 | 0.17 | 0.60 | 0.35 |
| Tungnahryggsjökull (E) | - | 0.38 | 0.21 | - | - | 0.40 | 0.32 |
| *Average* | - | *0.42* | *0.25* | - | - | *0.60* | *0.33* |
| **ELA-AABR (1.5±0.4)** (m a.s.l.) | | | | | | | |
| Glaciers | LIA | 1946 | 1985 | 1994 | 2000 | 2005 | ↑LIA-2005 |
| Gljúfurárjökull | 960±20 | 975±15 | 986±20 | 988±15 | 990±15 | 994 +15/-10 | 34 |
| Tungnahryggsjökull (W) | 1047 +20/-15 | 1093±10 | 1103±10 | 1101±10 | 1102 +10/-5 | 1105±10 | 58 |
| Tungnahryggsjökull (E) | 1020 +20/-15 | 1062±15 | 1075±10 | - | 1072 +15/-10 | 1079±10 | 59 |
| *Average* | *1009±36* | *1043±50* | *1055±50* | - | *1055±47* | *1059±47* | *50* |
| ELA-AABR rise rate ($m\ yr^{-1}$) | | | | | | | |
| Glaciers | LIA | LIA-1946 | 1946-1985 | 1985-1994 | 1994-2000 | 2000-2005 | ↑LIA-2005 |
| Gljúfurárjökull | - | 0.31 | 0.28 | 0.22 | 0.33 | 0.80 | 0.32 |
| Tungnahryggsjökull (W) | - | 0.59 | 0.26 | -0.22 | 0.17 | 0.60 | 0.42 |
| Tungnahryggsjökull (E) | - | 0.54 | 0.33 | - | - | 1.40 | 0.43 |
| *Average* | - | *0.48* | *0.29* | - | - | *0.93* | *0.39* |

ELA: equilibrium line altitude; AAR: accumulation area ratio; AABR: area altitude balance ratio; LIA: 'Little Ice Age' maximum.

**Table 5. Cold/warm periods at Akureyri and Öxnadalsheiði weather stations and seasonal values.**

| **Period** | **Type** | **Akureyri** | | | | **Öxnadalsheiði** |
|---|---|---|---|---|---|---|
| | | Air temperature (ºC) | | | | |
| | | Annual | Ablation season | Three-month summer | Annual range | Annual |
| 1882-1924 | Coldest | 2.52 | 7.86 | 9.41 | 15.59 | -0.73 |
| 1925-1950 | Warm | 3.95 | 9.05 | 10.38 | 14.80 | 0.57 |
| 1951-1955 | Cold | 3.62 | 8.60 | 9.96 | 14.71 | 0.27 |
| 1956-1965 | Warm | 3.79 | 8.38 | 9.52 | 14.56 | 0.43 |
| 1966-1973 | Cold | 2.81 | 8.12 | 9.47 | 15.44 | -0.46 |
| 1974-1978 | Warm | 3.61 | 8.73 | 10.31 | 15.75 | 0.38 |
| 1979-1986 | Cold | 2.93 | 7.91 | 9.85 | 14.84 | -0.35 |
| 1987-2002 | Warm | 3.77 | 8.95 | 10.27 | 14.71 | 0.45 |
| 2003-2014 | Warmest | 4.44 | 9.32 | 10.92 | 13.89 | 0.96 |

Source: Icelandic Met Office (IMO) and Icelandic Road and Coastal Administration.

Temperatures at Öxnadalsheiði prior to 2000 were reconstructed through the least squares equation obtained from the regression analysis between Akureyri and Öxnadalsheiði MAAT series for the common period 2000–2014.

**Table 6. Temperature and precipitation at the ELA calculated for each year: comparison between different models. All models agree on a wetter climate at the present day than during the LIA maximum.**

| Period | LIA | 1946 | 1985 | 1994 | 2000 | 2005 | ↑LIA-2005 |
|---|---|---|---|---|---|---|---|
| | Mean air temperature (ºC) | | | | | | |
| Ablation season (May-Sep) | 1.35 | 2.35 | 1.13 | - | 2.18 | 2.53 | 1.18 |
| Three-month summer (Jun-Jul-Aug) | 2.90 | 3.67 | 3.08 | - | 3.50 | 4.13 | 1.23 |
| Mean annual | -4.00 | -2.75 | -3.84 | - | -3.00 | -2.35 | 1.65 |
| | Precipitation (mm water equivalent) | | | | | | |
| Winter (Ballantyne, 1989 model) | 1445 | 2029 | 1344 | - | 1913 | 2159 | 714 |
| Annual (Ohmura et al., 1992 model) | 1578 | 1854 | 1641 | - | 1791 | 2022 | 444 |
| Annual (Braithwaite, 2008 model) | 1223 | 1713 | 1148 | - | 1552 | 1741 | 518 |

ELA: equilibrium line altitude; LIA: ‘Little Ice Age’.

**Table 7. Mean annual air temperature (MAAT) extrapolated to the Gljúfurárjökull and Tungnahryggsjökull snouts.**

| **Mean Annual Air Temperature (ºC)** | | | | | | |
|---|---|---|---|---|---|---|
| Glacier | LIA | 1946 | 1985 | 1994 | 2000 | 2005 |
| Gljúfurárjökull | -0.71 | 0.42 | -0.83 | 0.02 | 0.01 | 0.49 |
| Tungnahryggsjökull (W) | -0.89 | -0.78 | -2.10 | -1.09 | -1.22 | -0.64 |
| Tungnahryggsjökull (E) | -1.27 | -0.38 | -1.65 | - | -0.73 | -0.10 |
| *Average* | *-0.96* | *-0.25* | *-1.53* | - | *-0.65* | *-0.08* |

LIA: ‘Little Ice Age’ maximum.

**Table 8. Glacier termini variations of Gljúfurárjökull: comparison between remote and field measurements.**

| Period | Snout variation measurements (m) | |
|---|---|---|
| | Remote (GIS) | Field (IGS) |
| 1898-1946 | -635 | -450[a] |
| 1946-1985 | -275 | -322 |
| 1985-1994 | 20 | 39 |
| 1994-2000 | -26 | -25 |
| 2000-2005 | -77 | -63 |
| *Total* | *-993* | *-821* |

GIS: Geographical Information System; IGS: Icelandic Glaciological Society.

IGS measurements have been summarized for the periods analysed.

[a] Inferred from Caseldine and Cullingford (1981).

**Figure 1. Location of the study area in the interior of the Tröllaskagi peninsula.**

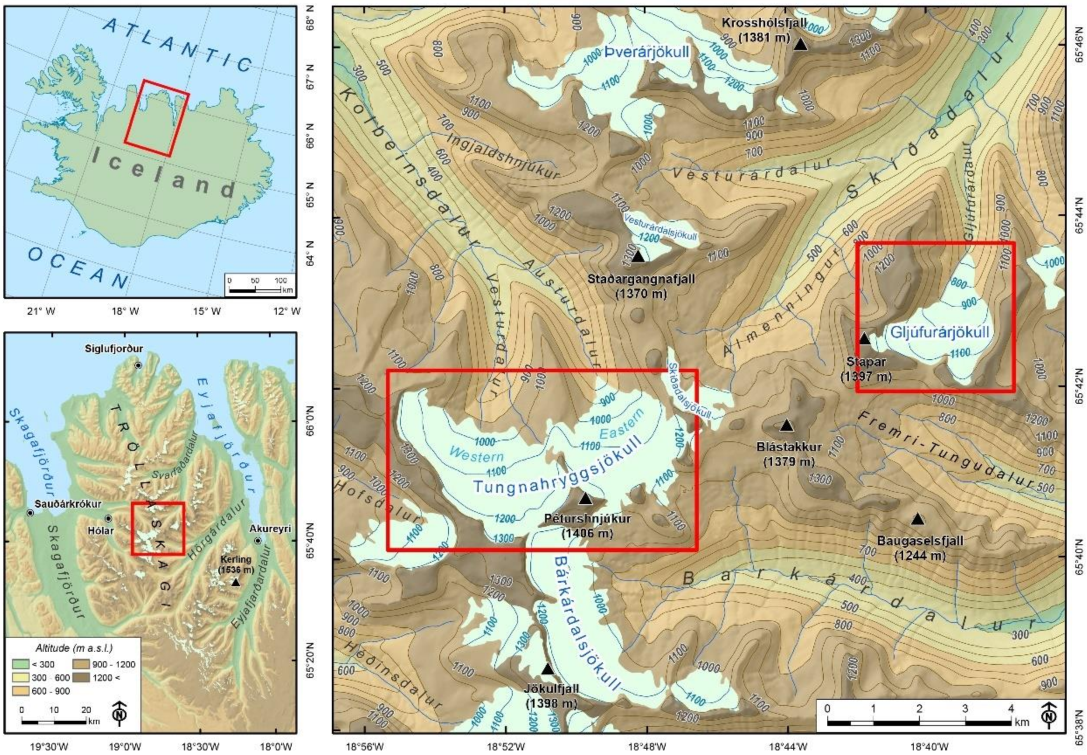

**Figure 2. Evolution of the glacier snouts. The greatest retreat took place between the LIA maximum and 1946 and was especially significant in the Tungnahryggsjökull. LIA: 'Little Ice Age' maximum.**

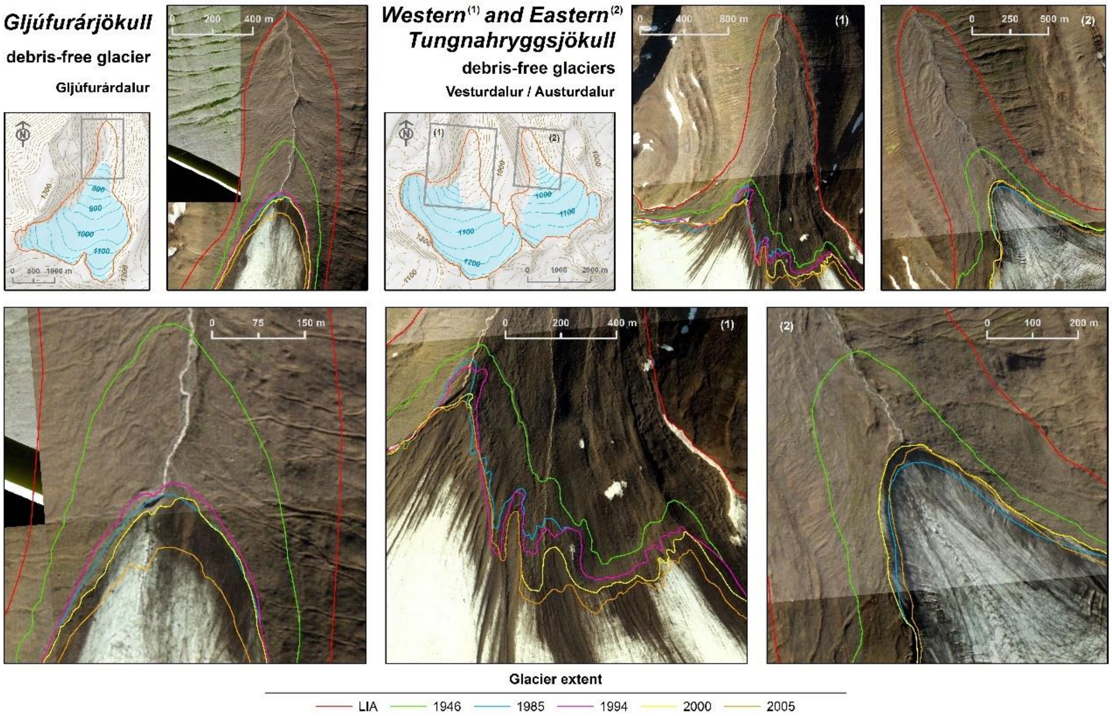

**Figure 3. Snout and moraine positions from field observations. (a, b) The Gljúfurárjökull and (d) Eastern Tungnahryggsjökull LIA moraines are easily recognized from their sharp-crested shape. The debris cover on the (c) Western Tungnahryggsjökull determines its complex snout evolution. LIA: 'Little Ice Age' maximum.**

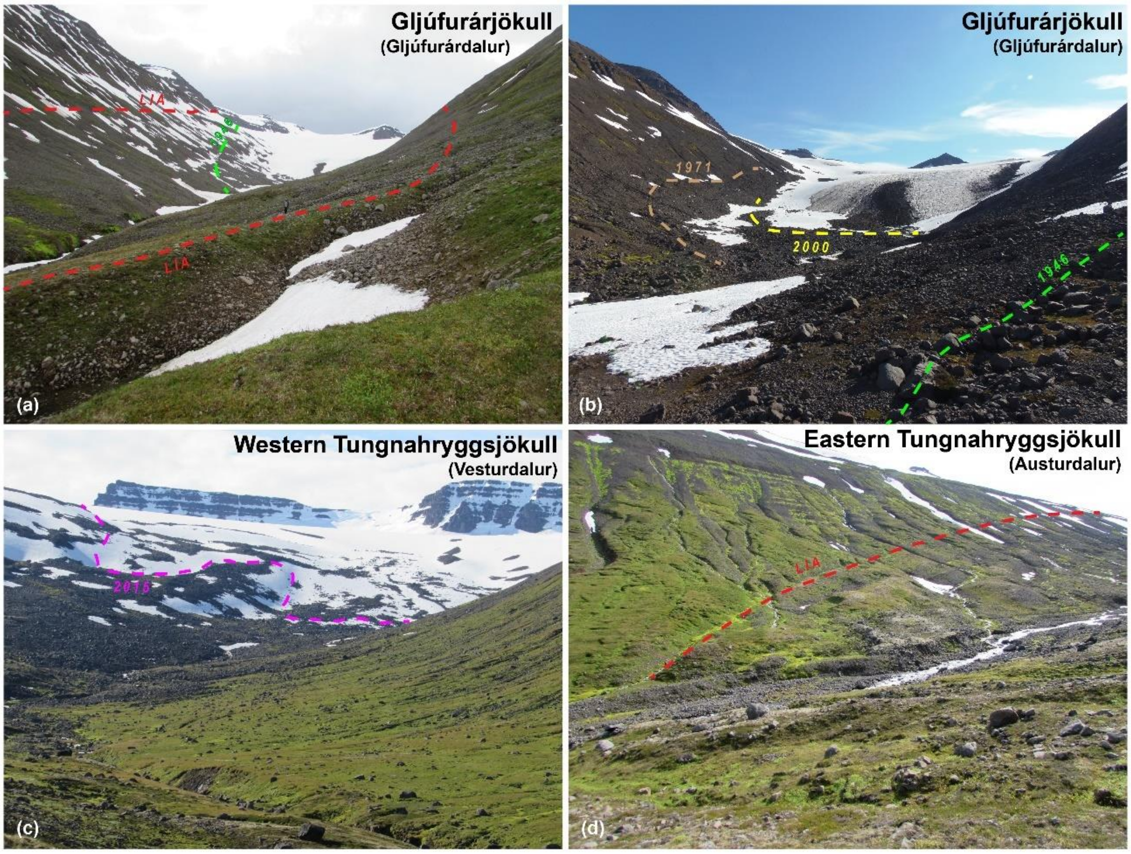

**Figure 4. Evolution of retreat rates and area and volume loss during the different periods analysed. From 2000 to 2005, the rates are close to those recorded in the first half of the 20th century. LIA: 'Little Ice Age' maximum; ND: no data.**

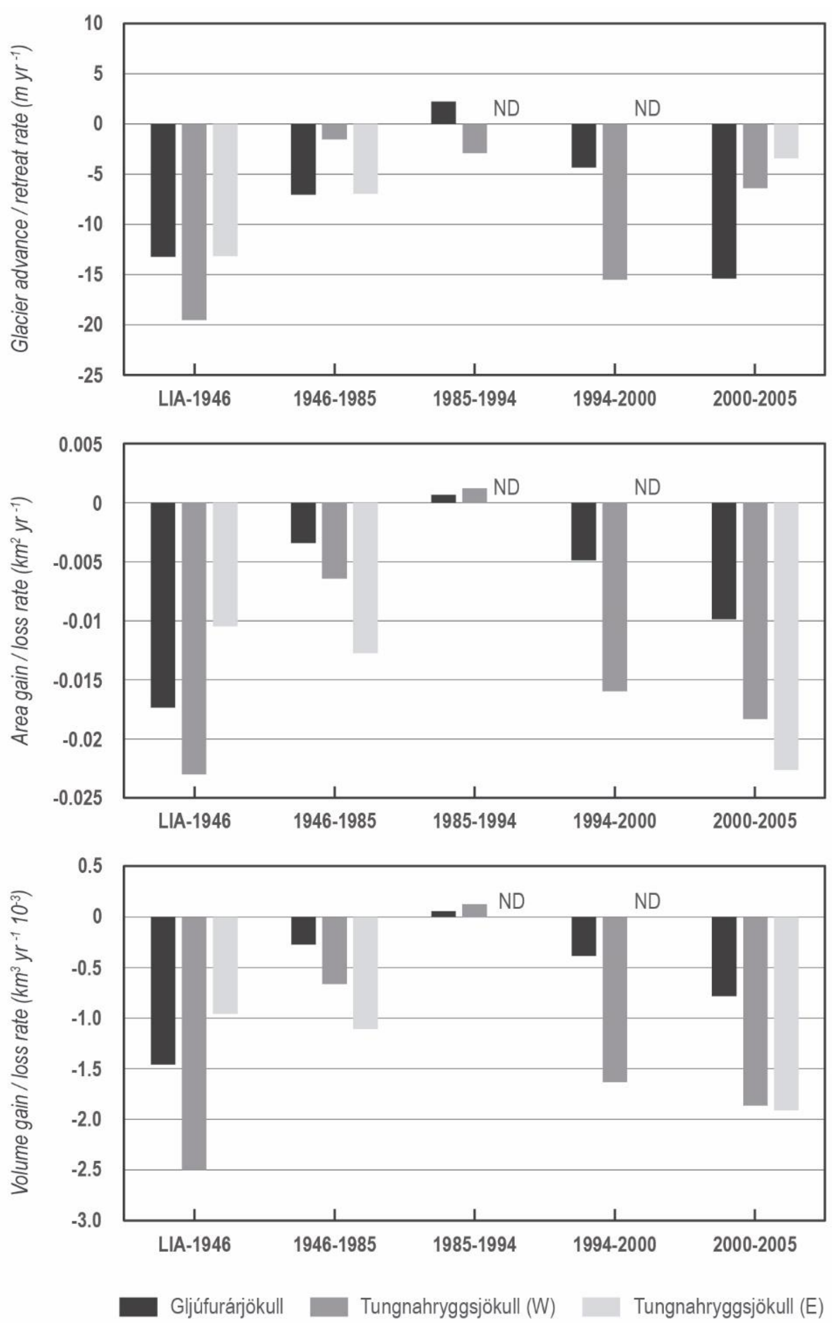

**Figure 5. Evolution of (b) mean annual air temperature (MAAT), mean ablation season air temperature (Ts) at (a) Akureyri and MAAT reconstruction at (c) Öxnadalsheiði. The coloured numbers in the middle of the periods are the mean value of MAAT/Ts for each period.**

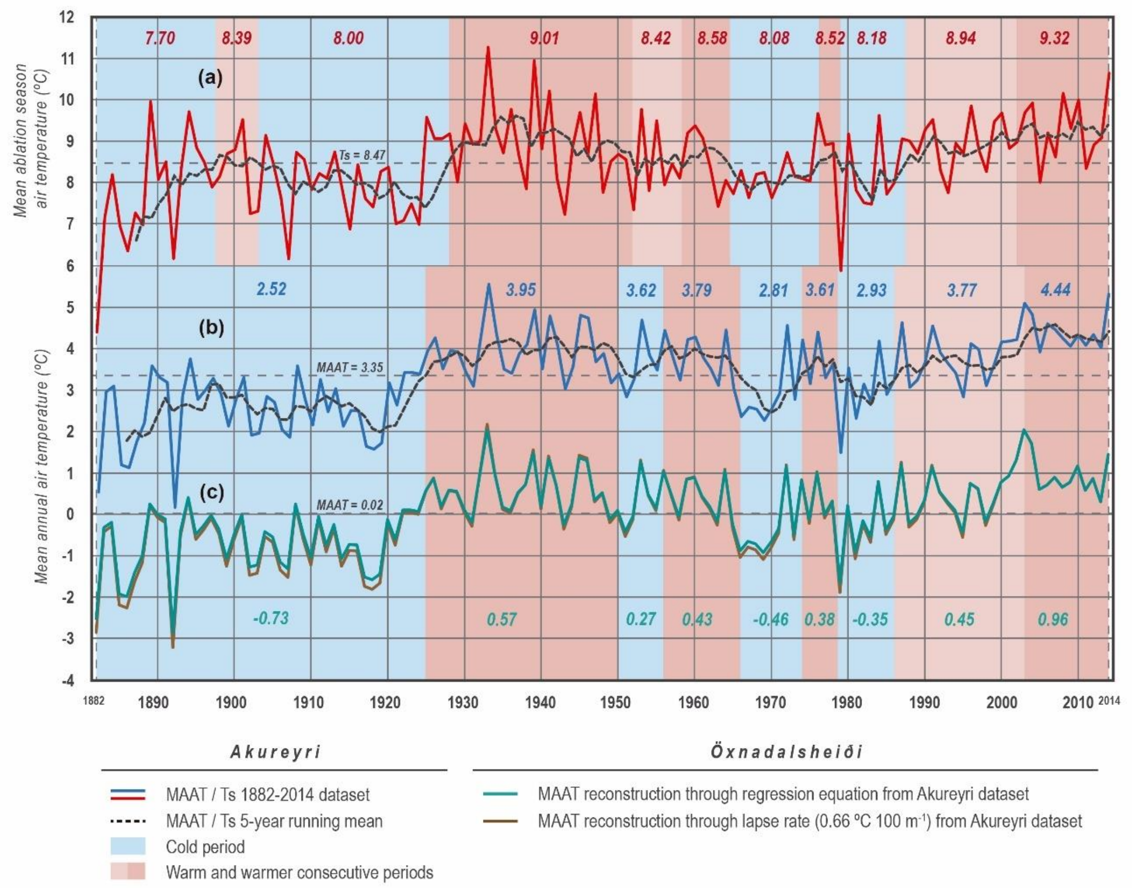

**Figure 6. (a) Relationship between the variations in Gljúfurárjökull snout (taken from the Iceland Glaciological Society, 2016), (b) ablation season temperature, (c) winter precipitation at Akureyri and (d) winter NAO index since 1950. Black dotted lines show 5-year running mean (temperature and precipitation) and LOESS regression in the NAO index (modified from Cropper et al., 2015). Red points in (a) are the years with marginal measurements. There is a clear relationship between the 1980s advance and the previous cooling of the mean ablation season air temperature (Ts). The winter precipitation evolution shows a curve parallel to that of the NAO index (high winter precipitation, positive NAO index phase), suggesting a connection between the NAO mode and the precipitation, especially in the early 1980s and 1990s.**