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# Recent, very rapid retreat of a temperate glacier in SE Iceland

TOM BRADWELL, ODDUR SIGURÐSSON AND JEZ EVEREST

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Iceland's glaciers are particularly sensitive to climate change, and their margins respond to trends in air temperature. Most Icelandic glaciers have been in retreat since *c*. 1990, and almost all since 1995. Using ice-front measurements, photographic and geomorphological evidence, we examined the record of ice-front fluctuations of Virkisjökull–Falljökull, a steep high-mass-turnover outlet glacier in maritime SE Iceland, in order to place recent changes in a longer-term (80-year) context. Detailed geomorphological mapping identifies two suites of annual push moraines: one suite formed between *c*. 1935 and 1945, supported by lichenometric dating; the other between 1990 and 2004. Using moraine spacing as a proxy for ice-front retreat rates, we show that average retreat rates during the 1930s and 1940s (28 m $a^{-1}$) were twice as high as during the period from 1990 to 2004 (14 m $a^{-1}$). Furthermore, we show that both suites of annual moraines are associated with above-average summer temperatures. Since 2005, however, retreat rates have increased considerably – averaging 35 m $a^{-1}$ – with the last 5 years representing the greatest amount of ice-front retreat (~190 m) in any 5-year period since measurements began in 1932. We propose that this recent, rapid, ice-front retreat and thinning in a decade of unusually warm summers has resulted in a glaciological threshold being breached, with subsequent large-scale stagnation of the glacier terminus (i.e. no forward movement) and the cessation of annual push-moraine formation. Breaching this threshold has, we suggest, caused further very rapid non-uniform retreat and downwasting since 2005 via a system feedback between surface melting, glacier thinning, decreased driving stress and decreased forward motion.

*Tom Bradwell (tbrad@bgs.ac.uk) and Jeremy D. Everest, British Geological Survey, West Mains Road, Edinburgh EH9 3LA, UK; Oddur Sigurðsson, Iceland Meteorological Office, Bustaðavegi 9, Reykjavik 150, Iceland;* *received 20th July 2012, accepted 13th January 2013.*

Iceland's glaciers are particularly sensitive to climatic fluctuations on an annual to decadal scale. Empirical studies have shown that fluctuations in glacier size (areal extent) are a direct result of the complex interplay between ice melt by radiative flux, ice accumulation through snowfall, and ice-flow rate (e.g. Ahlmann 1940; Björnsson & Pálsson 2008). In most cases, where the ice front does not terminate in a deep lake and surging does not occur, ice-front fluctuations in Iceland have been in sympathy with average air temperature trends since at least the mid-20th century (Jóhannesson & Sigurðsson 1998; Sigurðsson *et al*. 2007). Annual ice-front measurements made since the 1930s show that most of Iceland's glaciers have been in retreat since the 1990s, with many showing a retreat rate increase over the last decade (Sigurðsson *et al*. 2007). An understanding of the high rates of glacier change currently being experienced in Iceland is important in order to place the current period of atmospheric warming and associated glacier retreat in context. For example, is the last 20-year period of glacier recession in Iceland (since *c*. 1990) unusual in the measurement record (since *c*. 1930)? We examine in detail the instrumental, geomorphological and photographic evidence for rapid ice-front retreat at a sensitive maritime glacier in SE Iceland, and by doing so place the most recent period of glacier recession in a longer-term (80-year) context.

Most of Iceland's glaciers have been in general retreat since reaching their Little Ice Age (LIA) maxima, between AD 1780 and 1900 (Thórarinsson 1943; Guðmundsson 1997; Bradwell *et al*. 2006; Geirsdóttir *et al*. 2009). Prior to this, from the time of settlement (*c*. AD 870) to the 13th century, most of Iceland's glaciers were smaller than, or similar in size to, today (Guðmundsson 1997; Kirkbride & Dugmore 2008; Geirsdóttir *et al*. 2009). The intervening period between *c*. AD 1300 and 1700 is not well documented in the historical annals and little glaciological information exists, but most glaciers in Iceland are thought to have been larger than at the present-day; many underwent advances but remained smaller than during the height of the LIA (AD 1850–1900) (Thórarinsson 1943; Björnsson & Pálsson 2008).

Annual moraines occur in front of numerous glaciers in Iceland (Price 1970; Sharp 1984; Boulton 1986; Krüger 1995; Evans & Twigg 2002; Bradwell 2004a). They form at the margins of glaciers when forward ice-front movement during the winter outpaces the negligible ablation. The resulting small advance bulldozes or squeezes sediment at the ice front to create a minor push moraine. Long annual moraine sequences are created when recession during the summer (ablation season) is greater than the



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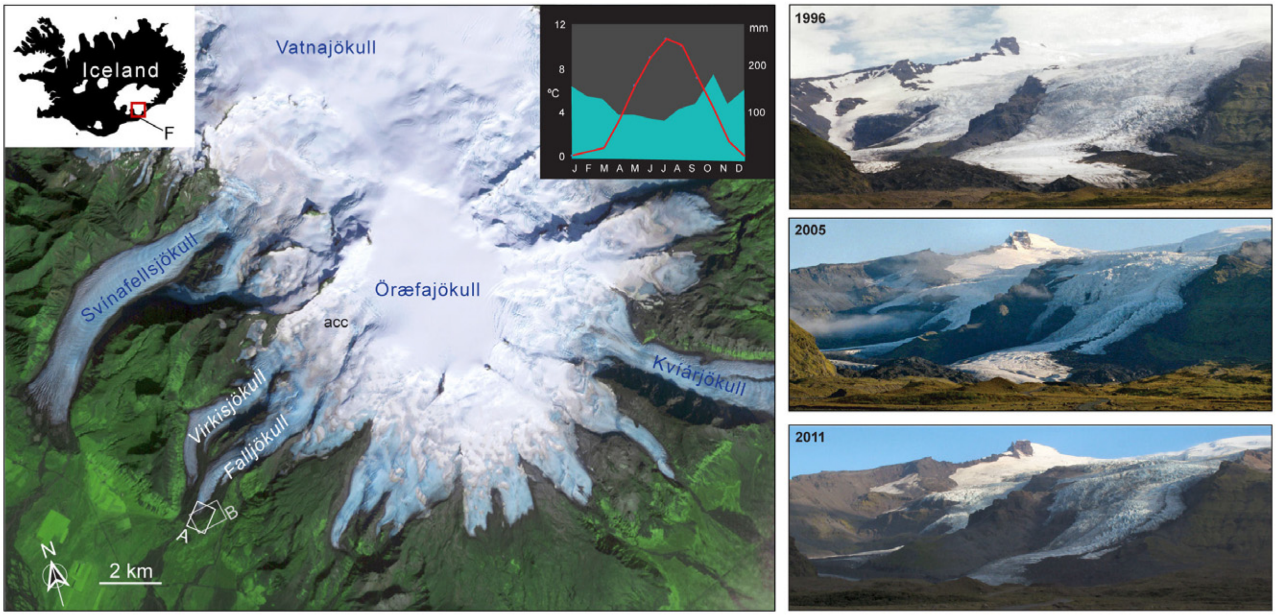


*Fig. 1.* Location of the study site in SE Iceland. Left panel: Satellite map of the Öræfajökull ice cap showing the main glaciers draining the massif. The twin outlet of Virkisjökull–Falljökull is highlighted (white text); note the shared accumulation area (acc). The two areas of minor moraines near to the glacier margin are shown (boxes A and B). Inset shows climatic data from Fagurhólsmyrí weather station (F on inset map); mean monthly air temperature (red line) and mean monthly precipitation (blue shading) for 1961–1990 averages. Right panel: Field photos of Virkisjökull–Falljökull taken in September 1996 (top), September 2005 (middle) and September 2011 (bottom) from the same location. Note the significant change that has occurred, with the glacier decreasing in width, length and thickness considerably over this 15-year period. Ash from the Grimsvötn eruption (of May 2011) has darkened the ice surface somewhat in the most recent picture. This figure is available in colour at http://www.boreas.dk.

advance during the winter (accumulation season) over a number of years (Boulton 1986; Bennett 2001). As such they can provide important sources of palaeoglaciological information and have been associated with periods of elevated ablation-season temperature (Sharp 1984; Krüger 1995; Bradwell 2004a; Beedle *et al*. 2009) and, in more continental settings, with variations in mean winter temperature (Lukas 2012). To be preserved, these small moraines must not be subsequently disturbed by the glacier or be eroded by meltwater streams; hence, their preservation potential in the landscape record is typically low. However, they have been reported from a wide range of glacial environments – at low elevation in maritime settings (e.g. Thórarinsson 1967; Gordon & Timmis 1992; Winkler & Nesje 1999) and fronting high-altitude mountain glaciers in more continental settings (e.g. Beedle *et al*. 2009; Lukas 2012).

This paper focuses on the twin-lobed outlet glacier of Virkisjökull-Falljökull flowing from the mountain ice cap of Öræfajökull, in SE Iceland (Fig. 1). Specifically, this article documents the changes that have occurred in the glacier's conjoined terminal zone over the last ~80 years, since measurements began in 1932. We compare the measurement record with the photographic record and detailed geomorphological record, in the form of annual moraines deposited since *c*. 1990. We then use annual moraines at Falljökull from the first half of the 20th century as a proxy for ice-front retreat; compare retreat rates and styles during these similar periods of environmental change; and discuss the wider implications for glacier behaviour.

## Study area

### Glaciology

Öræfajökull, the southernmost portion of the Vatnajökull ice cap, is a glaciologically distinct ice centre covering the Öræfajökull stratovolcano – Iceland's highest mountain (Fig. 1). Öræfajökull is a high-mass turnover, temperate ice cap with several outlets terminating close to sea level (50–150 m a.s.l.). The ice cap has 13 officially named outlets (Sigurðsson & Williams 2008). Eight of these glacier margins are monitored annually by the Icelandic Glaciological Society; the first measurements were made in 1932 (Sigurðsson 1998). The outlets are all maritime (winter-accumulation type) glaciers, with a strong preference towards mass loss in summer months (Björnsson *et al*. 1998). Although melting in the terminal zone can occur in any month of the year, the ablation season typically lasts from May to late September. Average mass-


*Fig. 2.* Oblique aerial photo of Virkisjökull–Falljökull. Photo taken: 28th September 2002, Oddur Sigurðsson. Rauðikambur (R), medial moraine (MM), and minor moraines (A, B) studied here are all highlighted. X marks the location of photo in Fig. 7C; lines show approximate field of view. This figure is available in colour at http://www.boreas.dk.

balance gradients on the Öræfajökull outlets are the strongest in Iceland. Annual net balance (averaged from 1991 to 2006), calculated using a robust stratigraphic method, ranges from −10 m in the terminal zone to +5 m above 1800 m elevation (Björnsson *et al.* 1998; Björnsson & Pálsson 2008). This mass-balance gradient combined with steep glacier profiles makes these some of the highest-mass-turnover glaciers in Europe (Dyurgerov 2002). Equilibrium-line altitudes based on mass-balance approximations (1991–2006) are between 1000 and 1200 m a.s.l. (Björnsson & Pálsson 2008).

The ice cap flows from the western part of the summit crater of Öræfajökull (2000 m asl) as twin outlet glaciers with a shared accumulation area of ~5 km$^2$ (Fig. 1). Below 1200 m a.s.l. this ice flow splits into two separate glacier arms – Virkisjökull and Falljökull – flowing on either side of a prominent bedrock ridge named Rauðikambur. The glaciers recombine in their terminal zone, which previously extended ~1 km further downvalley (Danish General Staff 1904; Guðmundsson 1997). A wide supraglacial debris band or medial moraine, sourced from the Rauðikambur nunatak, marks the distinction between the two glacier arms below ~300 m a.s.l. (Fig. 2). Much of the terminal zone of Virkisjökull is now debris-covered. Considerable mass loss has occurred at Virkisjökull–Falljökull over the last ~20 years, notably decreasing the areal extent and surface elevation of the glacier in its ablation area. This is evidenced in sequential field photos taken (by the lead author) at multi-annual intervals from the same location (Fig. 1).

Field measurements of glacier-front positions in Iceland have been made routinely every year by members of the Icelandic Glaciological Society (IGS) since the 1930s (Sigurðsson 1998). At Virkisjökull–Falljökull, annual measurements of the northern ice front (Virkisjökull) started in 1932, were discontinuous in the 1970s, and ceased in 2000 owing to the large-scale stagnation of this debris-covered ice margin and uncertainty surrounding its exact location. Measurements of the Virkisjökull ice front since *c.* 1968 are therefore deemed to be unreliable and not representative of the glacier front as a whole (Fig. 3). The southern ice front (Falljökull) has much less supraglacial debris cover than the northern arm. Ice-front measurements at Falljökull started in 1957 and have continued every year to the present day. Measurements are usually carried out in the last week of September or early October. The complete dataset is presented here (Fig. 3), both cumulative and annual ice-front measurements (1932–2011), along with the detailed course of the measurement transect (Fig. 4).

Combined annual data from Virkisjökull–Falljökull reveal several important trends in ice-front fluctuations over the past 80 years, and these can be generally summarized into six periods (Table 1). The average annual advance or retreat rate for these periods ranges from +10 m to −30 m respectively, with maximum values of +35 m a$^{-1}$ and −50 m a$^{-1}$ (excluding unreliable post-1968 data from Virkisjökull) (see Fig. 3). A strong

*Table 1.* Summary of ice-front fluctuations at Virkisjökull–Falljökull (since annual measurements began).

| Date | Ice-front behaviour |
|---|---|
| 1932–1934 | Advancing |
| 1935–1948 | Rapidly retreating |
| 1949–1972 | Slowly retreating or stationary |
| 1973–1990 | Advancing |
| 1991–1997 | Slowly retreating or stationary |
| 1998–2011 | Rapidly retreating |

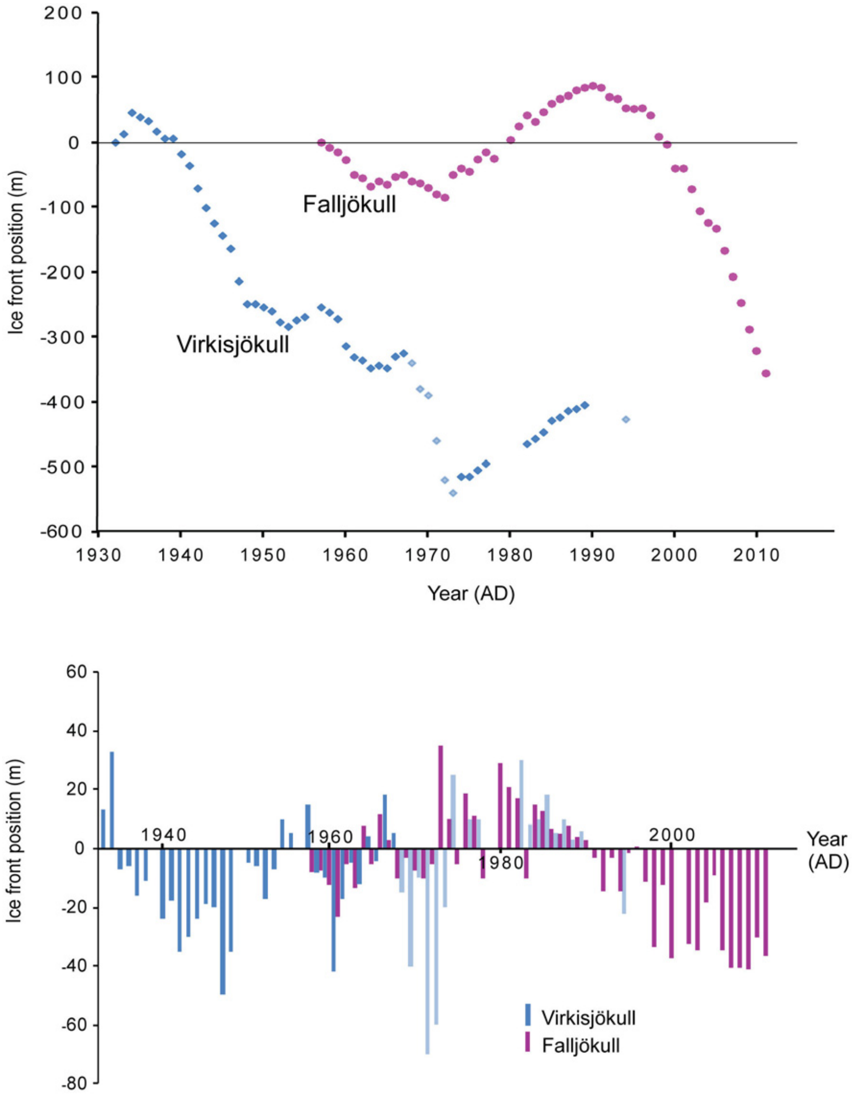


*Fig. 3.* Graphs showing cumulative (upper) and yearly (lower) retreat of Virkisjökull and Falljökull ice fronts. From annual measurements taken by the Icelandic Glaciological Society; Virkisjökull (diamonds) since 1932; Falljökull (dots) since 1957. Retreat measurements of Virkisjökull (1968–1973 and in 1995) are deemed unreliable, owing to thick debris-cover on the ice margin. These data are shown as semi-transparent. This figure is available in colour at http://www.boreas.dk.

relationship has been previously observed between glacier-front fluctuations and air temperatures since 1930, with summer temperature being the primary driver of ice-front recession on a sub-decadal scale (Sigurðsson 2005; Sigurðsson *et al*. 2007).

*Meteorology*

The Öræfi area experiences a relatively mild oceanic climate with a low mean annual temperature range (~11°C) and around 150 rain/snow days a year (Einarsson 1984). Mean summer temperatures (1971–2000) at the nearest long-term weather station (Fagurhólsmyrí; ~100 m a.s.l.) are between 8 and 12°C, but daily maxima of 20°C are not uncommon in summer (Fig. 1). Average temperatures during winter months are typically around 0–4°C, with daily minima below −5°C relatively rare. Snow lies on low ground (<200 m) for only 3–4 weeks a year on average; however, snow above this elevation can last for considerably longer. Mean annual precipitation, although not easy to record in windy conditions, is around 1800 mm immediately south and west of Öræfajökull. By contrast, precipitation on the eastern side of the mountain averages 3000 mm $a^{-1}$, and can locally exceed 7000 mm $a^{-1}$ on the summit plateau (Guðmundsson 2000). There is no strong seasonal trend in precipitation in SE Iceland, although October, December and January are typically the wettest months close to sea level (Fig. 1).

## Methods

Digital high-resolution scans of seven vertical analogue aerial photos (purchased from the National Land Survey of Iceland (LMI)) were georectifed and imported into ArcGIS 9.3. The photos cover the period from 1945 to 2003; their scales vary, as do the

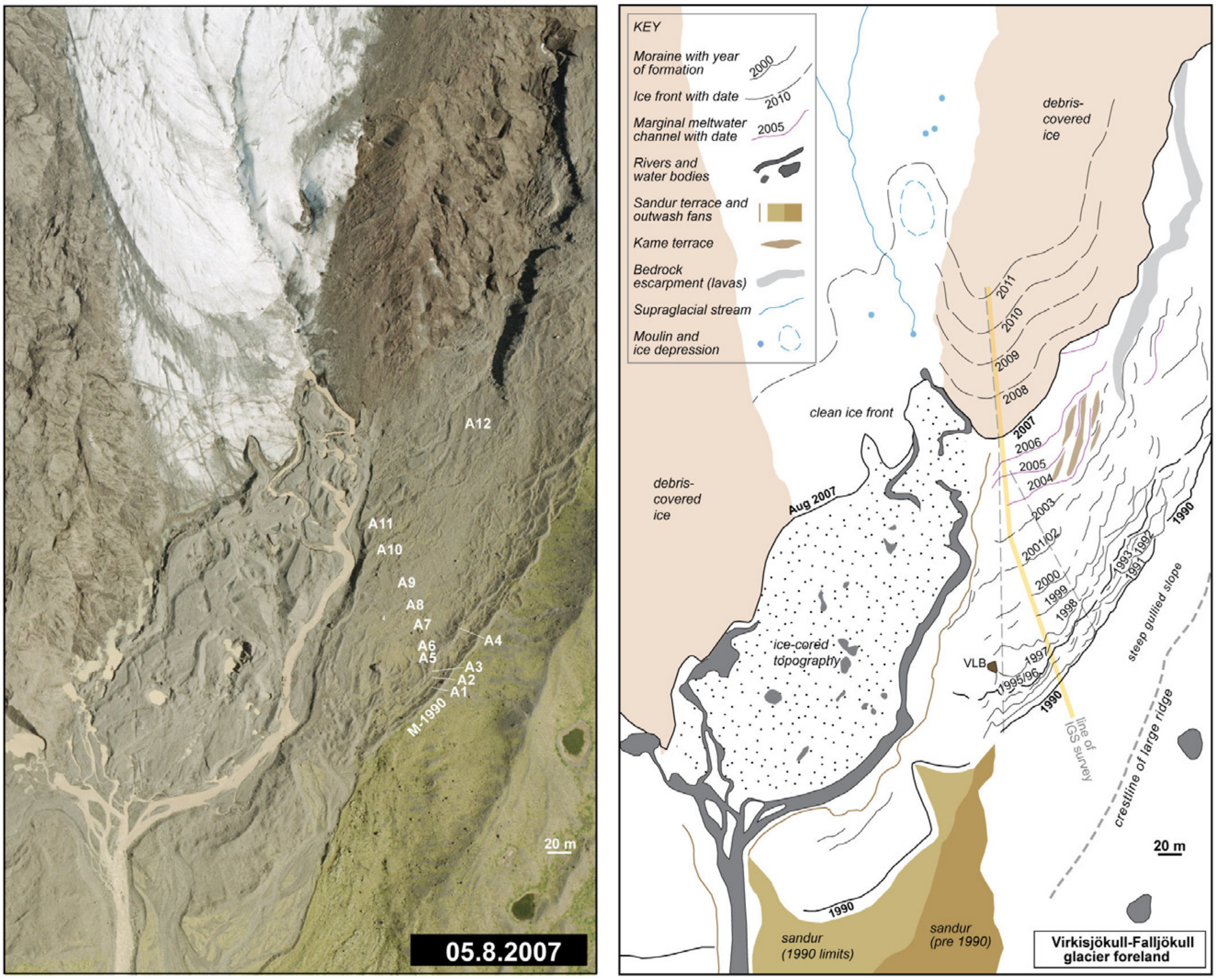


*Fig. 4.* 2007 aerial photo (left) and geomorphological map (right) of Suite A minor moraines, glacier margin and adjacent foreland. Minor moraines are numbered A1–A12 on the 2007 air photo. The line of the IGS ice-front survey is shown as a yellow line. Survey transects for this study are shown as dashed lines converging on the IGS survey line. Moraine formation dates are ascribed based on cross-checks with the IGS survey, aerial photos (see Fig. 7) and repeat fieldwork since 1996. Moraine formation is assumed to take place in late winter or early spring (March–May). The prominent moraines formed in 1990 and 1995–1996 are shown in bold, as is the glacier margin position in August 2007. The long-dashed lines are September ice-front positions since 2007, surveyed in the field, shown for part of the glacier only. VLB = very large boulder. This figure is available in colour at http://www.boreas.dk.

resolution and quality of the images (see Table S1 in Supporting Information). The most recent aerial photo is a high-resolution (0.2-m) digital colour image taken in 2007, part of a regional survey flown by the NERC Airborne Research Survey Facility. This image was used for on-screen precision geomorphological mapping. Positional ground control was collected in the field using a geodetic-grade Leica ViVa dGPS during 2010 and 2011 with an accuracy of 0.03 m in $x$ and $y$ and of 0.05 m in $z$.

A detailed examination of the glacial landforms was made from vertical aerial photos, supplemented by field photos and geological field surveys. From this examination, two areas of interest were identified (A, B; Fig. 1). Both areas display suites of low-elevation closely spaced moraines. Suite A, adjacent to the present-day ice margin, was mapped digitally on-screen from a grey-scale contrast-enhanced 2007 air photo (Fig. 4). Suite B was mapped in the same way; however, a contrast-enhanced enlarged 1954 photo was also used because the moraine morphology is clearer and the preserved moraine sequence is longer on this image, as a result of subsequent glacier overriding (Fig. 5). Both moraine suites were surveyed in the field in 2011 using a geodetic-grade Leica dGPS, to derive ridge heights ($\pm$0.1 m), ridge widths ($\pm$0.5 m) and ridge spacings ($\pm$1 m). For Suite A, where possible, three transects of ridge morphology and spacing were measured; values are expressed as means with maxima and minima (Table 2). For Suite B, ridge spacing was measured

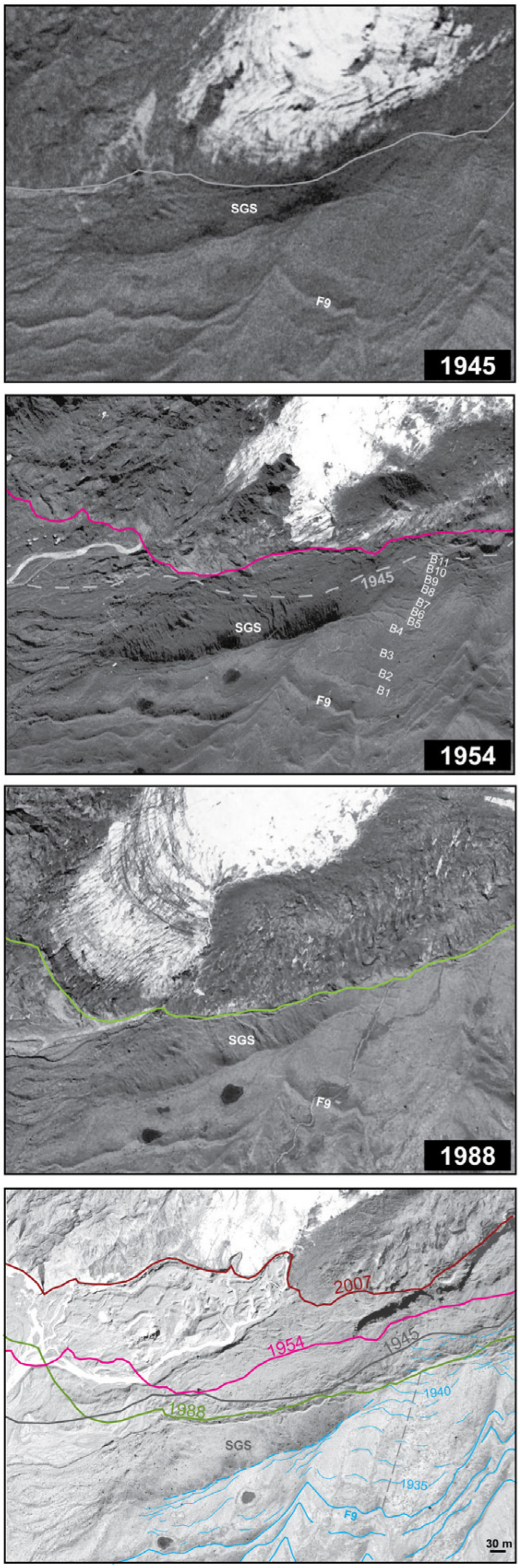


*Fig. 5.* Orthorectified aerial photos of Falljökull ice front (1945, 1954, 1988) and geomorphological map (lower panel) of Suite B minor moraines on 2007 (greyscale) air photo base. Minor moraines are numbered B1–B11 on the 1954 photo. Assuming annual formation, moraine dates can be ascribed to between 1935 and 1945. Lichenometric surveys support this timing. Glacier margins (ice fronts) are shown as coloured lines with dates. SGS = steep gullied slope; F9=moraine ridge just beyond minor moraine suite. See Table S1 for air photo details. This figure is available in colour at http://www.boreas.dk.

along a single representative transect (to the nearest metre); however, where possible, ridge morphology was measured in three places with dimensions expressed as mean values with maxima and minima (Table 3). In both moraine suites, slope angles were calculated by trigonometry using measurements of ridge height, crestline position and ridge width (as shown in Fig. 6).

Analysis and cross-correlation of the IGS ice-front measurements since 1990 with our GPS survey were undertaken to ascertain the formation frequency of the moraines within Suite A (Fig. 4). In addition, field photos taken from the same location over a period of 15 years (1996–2011), supported by aerial photos (taken in 1988, 1992, 1997, 2003 and 2007), were used to determine the exact ice-margin position with respect to moraines formed between 1990 and the present day.

In order to determine the age of moraine Suite B (formed prior to 1945, according to aerial photos) a lichenometric survey was undertaken (*Rhizocarpon geographicum* lichens), following the methodology used previously in SE Iceland (Bradwell 2001, 2004b; Bradwell *et al*. 2006). The maximum diameters of all approximately circular, non-coalescent, Subgenus *Rhizocarpon* thalli over 5 mm were measured within a fixed area (normally ~50 m$^2$), avoiding section *Alpicola* lichens. Lichenometric size-frequency (SF) and largest-lichen (LL) surveys were conducted on the proximal side of two ridges: (i) the oldest moraine in Suite B (F9); and (ii) the youngest preserved moraine in Suite B, believed to date from AD 1945 (Table 3). Uncertainties in the lichenometric technique stem from a range of environmental and operator-related factors, but these can be minimized when LL measurements are supported by the more statistically robust SF approach (Bradwell 2004b). Typically, combined uncertainties are in the 5–10% range when deriving surface ages less than 200 years old (Innes 1988; Noller & Locke 2001).

## Results

### *Moraine geomorphology and sedimentology*

*Suite A.* – Small ridges of glacially deposited sediment, referred to here as minor moraines, occur adjacent to the Falljökull ice margin at the foot of the prominent basaltic-andesitic lava ridge on a relatively steep

*Table 2.* Morphometry and chronology of moraines in Suite A.

| Ridge no. (and date[1]) (AD) | Height[2] (m) | Min, max (m) | Width[3] (m) | Min, max (m) | Spacing[4] (m) | Min, max (m) |
|---|---|---|---|---|---|---|
| M-1990 | 1.6 | 0.8, 3.0 | 7.5 | 6.0, 8.5 | n/a | |
| A1 (1991) | 0.6 | 0.3, 0.8 | 3.5 | 3.0, 4.0 | 3 | 2, 4 |
| A2 (1992) | 1.0 | 0.4, 1.3 | 4.0 | 3.0, 5.5 | 5 | 4, 6 |
| A3 (1993) | 0.9 | 0.6, 1.1 | 4.0 | 3.5, 4.5 | 3 | 1, 5 |
| A4 (1994) | 1.0 | – | 3.5 | – | 2 | – |
| A5 (1995/96) | 1.3 | 0.5, 2.0 | 5.0 | 4.0, 6.5 | 2 | 1, 6 |
| A6 (1997) | 0.5 | 0.3, 0.8 | 4.0 | 3.0, 5.5 | 16 | 6, 24 |
| A7 (1998) | 0.5 | 0.3, 0.8 | 3.0 | 2.0, 4.0 | 25 | 17, 30 |
| A8 (1999) | 0.8 | 0.4, 1.0 | 2.5 | 2.0, 3.5 | 16 | 14, 18 |
| A9 (2000) | 0.7 | 0.4, 1.0 | 3.5 | 2.0, 6.0 | 32 | 30, 33 |
| A10 (2001/02) | 0.6 | 0.4, 1.0 | 3.0 | 2.0, 4.0 | 27 | 23, 30 |
| A11 (2003) | 0.8 | 0.5, 1.2 | 3.0 | 2.0, 4.5 | 29 | 25, 31 |
| A12 (2004) | 0.5 | – | 2.5 | – | – | – |

[1]Age derived from aerial photos and annual IGS measurements; presumed winter–spring formation.
[2]Mean ridge height, rounded to nearest 0.1 m; average of three readings.
[3]Mean ridge width, rounded to nearest 0.5 m; average of three readings.
[4]Crest-to-crest ridge spacing, rounded to nearest whole metre; average of three readings. Only one set of measurements was made on ridges A4 and A12.

concave slope that dips towards the glacier. This lava ridge pre-dates the Neoglacial period and formed >2000 a BP. The moraines comprise a sequence of 12 subdued low-elevation, occasionally overprinted, boulder-strewn ridges (Figs 4, 7). The outer five or six moraines are generally more prominent, larger (>1 m high) and more closely spaced, while the inner six moraines are generally smaller in height (<1 m), more subtle and more widely spaced (Figs 4, 7).

The outermost ridge (M-1990) (not part of the minor-moraine sequence) according to measurements from the IGS was bulldozed during the re-advance of the ice front in the late 1980s. It is a notable asymmetric push moraine up to 3 m high and continuous for over 500 m. It is ascribed a formation date of spring 1990 and now delimits a very marked vegetation change on the glacier foreland (Figs 4, 7).

Detailed measurements across the minor moraines in Suite A revealed that ridge heights range from 0.5 to 2.0 m, ridge widths range from 2.0 to 6.5 m and ridge spacing (crest to crest) varies from 1 to 33 m (Table 2). Most ridges are generally continuous for lengths of between 100 and 400 m. Furthermore, most of the ridges are broadly arcuate in plan form with concavity towards the present-day glacier margin; some have undulating crestlines. Although the plan-form patterns are generally simple, ridge crests occasionally coincide or overlap where the ridges are most closely spaced (near M-1990), indicating that some ridges have been reoccupied or modified on more than one occasion. In cross-profile the ridges are quite steep features with clear slope asymmetry. Distal faces are shorter and steeper, typically 20°–30°, whilst proximal faces are generally longer and more gently sloping, typically 15°–

*Table 3.* Morphometry and chronology of extant moraines in Suite A (also see Fig. S1).

| Ridge no. | Height[1] (m) | Min, max (m) | Width[2] (m) | Min, max (m) | Spacing[3] (m) | Largest lichen[4] (mm) | Date[5] (AD) |
|---|---|---|---|---|---|---|---|
| F9 | 2.0 | 0.8, 2.8 | 8.5 | 6.5, 9.5 | n/a | 41 | 1929±4 |
| B1 | 0.8 | 0.3, 1.5 | 5.0 | 4.0, 7.5 | 48 | 40 | 1930±4 |
| B2 | 0.6 | 0.4, 1.1 | 7.0 | 6.0, 8.0 | 28 | 38 | 1933±4 |
| B3 | 0.4 | 0.2, 0.7 | 6.0 | 4.0, 8.0 | 43 | – | |
| B4 | 0.6 | 0.3, 0.8 | 5.5 | 4.0, 6.0 | 50 | – | |
| B5 | 1.5 | 0.9, 2.1 | 7.5 | 5.5, 8.5 | 23 | 34 | 1940±3 |
| B6 | 1.4 | 0.5, 1.8 | 5.0 | 4.0, 7.0 | 18 | – | |
| B7 | 1.0 | 0.5, 1.2 | 4.5 | 3.5, 7.0 | 17 | 34 | 1940±3 |
| B8 | 0.8 | – | 4.5 | – | 20 | – | |

[1]Mean ridge height, rounded to nearest 0.1 m; average of three readings.
[2]Mean ridge width, rounded to nearest 0.5 m; average of three readings.
[3]Crest-to-crest ridge spacing along marked transect, rounded to nearest whole metre owing to uncertainty surrounding precise crestline position. Only one set of measurements was made on ridge B8.
[4]Long axis of largest lichen measured on ice-proximal slope or crest of moraine.
[5]Moraine formation date based on largest lichen, using corrected age–size curve (see Fig. S1).

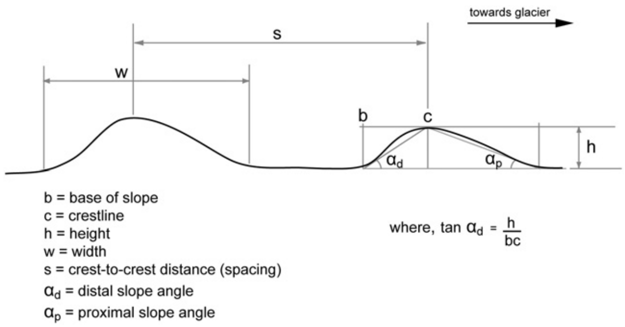


*Fig. 6.* Typical morphometry of minor-moraine ridges: definitions and field relationships.

20° (Fig. 7). The moraines are composed of moderately well-consolidated, poorly sorted subglacial, supraglacial and glaciofluvial sediment ranging from boulder to silt-clay grade. Boulders and cobbles are common on the surface of the ridges; however, most sediment is gravel-grade material (6–120 mm). Clasts are well rounded to subangular, with a high proportion being faceted and striated. Occasional large angular boulders occur on the ridges and strewn between the moraines. Clast lithologies reflect the local geology – typically vesicular and porphyritic basaltic lavas, tuffs and pyroclastic sediments including pumice and scoria. A natural section in the outermost ridge resulting from slope failure revealed almost 1.0 m of poorly sorted moderately well-consolidated sandy gravel diamicton with numerous large clasts (cobbles and boulders). Fines (<100 μm) were present in the sandy matrix, but not as discrete units or lenses. Apart from a crude coarsening-up grading in the diamicton, no sedimentary structures were identified.

*Suite B.* – Minor moraines occur to the east of the prominent ridge that bisects the proglacial foreland on ground that slopes gently away from the glacier. They originally formed a sequence of 11 subdued low-elevation boulder-strewn ridges, mapped as B1–B11; only eight now remain (Fig. 5). Detailed measurements across the minor moraines in Suite B revealed that ridge heights range from 0.4 to 2.1 m, ridge widths range from 4 to 8 m, and ridge spacings range from 17 to 50 m (Table 3). One or two of the ridges are poorly defined and fragmentary (e.g. B2, B3), but most are continuous for between 60 and 250 m. A small meltwa-


*Fig. 7.* Field photos of Suite A minor moraines. A. View looking SW across the outer push-moraine ridges (A1–A9). Moraines from 1990, 1995/96 and 2000 are labelled. VLB = very large boulder, with 1.5-m survey tripod circled. Small white boxes highlight IGS survey posts. B. Typical push-moraine morphology (A2) on reverse-sloping ground. Note the boulder strewn surface; notebook for scale. C. Wide-angle shot of moraine Suite A taken in September 2010 looking to the NE; ice-front positions in 1990, 1995/96, 2000, 2003, 2007 and 2010 are shown. SGS = steep gullied slope; VLB = very large boulder; SB = subhorizontal benches; white arrows = deep ice-marginal meltwater channels. This figure is available in colour at http://www.boreas.dk.

ter stream and associated sandur system have removed part of the central portion of the moraine sequence. The ridges are difficult to map over long distances owing to their subtle relief: in places they merge imperceptibly into the surrounding slopes. Individual ridges are generally linear in plan form, but when mapped out their overall shape is weakly arcuate with concavity towards the glacier margin. In cross-profile the ridges are low-angle features with clear slope asymmetry. Distal faces are shorter and steeper, typically 15°–20°, whilst proximal faces are longer and gentler, typically 10°–15°. The moraines are composed of moderately well-consolidated, poorly sorted sediment ranging from boulder to silt-clay grade. Boulders and cobbles are common on the surface of the ridges; however, the modal clast size is typically gravel-grade material (6–120 mm). Clasts are largely subrounded to subangular, and are commonly faceted and striated. Clast lithologies reflect the local geology, as seen in Suite A. A shallow excavation was made in the distal slope of ridge B7. It confirmed the poorly sorted, coarse grade of the material, and the general lack of fines to a depth of 0.5 m. No sedimentary structures were observed.

### *Chronology*

*Suite A.* – This recent sequence of minor push moraines (A1–A12) formed between 1991 and 2004, based on aerial photo evidence and annual ice-front measurements, with individual moraines forming every year, probably during late winter or spring. Detailed examination of enlarged aerial photos (Fig. 8) shows that the glacier front was still advancing in 1988, prior to the deposition of the large push moraine (M-1990) in 1990; by late July 1992 the ice front was in retreat and two small push moraines had formed inside M-1990 (Fig. 8), probably in late winter 1991 and 1992. In August 1997 the ice margin was only a short distance inside the 1992 position; a small re-advance occurred in 1995/96, forming a push moraine almost up to the July 1992 ice-front position in places (Fig. 8). In August 2003 the ice margin showed an almost exact correspondence with the shape of the minor moraine immediately adjacent to it (~20 m away); this ridge probably formed earlier in that year (Fig. 8). The innermost ridge, a small discontinuous moraine fragment <20 m from the lava escarpment, probably formed in late winter or spring 2004. Since 2003, steep-sided meltwater channels up to 5 m deep have been cut just beneath or adjacent to the spring ice-margin position (Fig. 7). Between 2003 and 2007, small ice-marginal benches (<10 m wide and <50 m long) were also deposited on the slope, akin to small-scale kame terraces (Fig. 7). Since 2005, annual push moraines have not formed at the ice margin, but ice-marginal meltwater channel formation has continued and been enhanced.

*Suite B.* – These minor push moraines (B1–B11) formed in the first half of the 20th century. All the available evidence indicates that the whole sequence formed between about 1935 and 1945 (Fig 5). The oldest air photo shows that the glacier margin was adjacent (<20 m) to the innermost moraine in August 1945. Furthermore, a good-quality higher-resolution air photo from 1954 shows that the moraines were still prominent fresh-looking features (Fig. 5). Their form had become notably more subdued in later photos (1960, 1988 and 2007), suggesting that they formed not long (~years rather than ~decades) before the earliest air photos were taken. The later photos also show that most of this ground has not been overridden by ice since. The inner part of Suite B, however, was bulldozed and overprinted by an advance of the ice margin between *c.* 1980 and 1990. This leaves only eight of the original 11 minor moraines now exposed and undisturbed.

The small size of these moraines combined with their morphological similarity to other features in SE Iceland and elsewhere (e.g. Sharp 1984; Boulton 1986; Bradwell 2004a; Lukas 2012) strongly suggest that these are annual moraines relating to minor oscillations of the ice front during overall glacier recession. The youngest (now removed) ridge (B11) probably formed in the late winter or spring of 1945; older ridges are ascribed annual formation dates with the oldest (B1) probably dating from early 1935 (Fig. 5).

The lichenometric survey conducted on the distal slope of the innermost extant whole ridge (B7) recorded a LL of 34 mm and a single (unimodal) SF population ($n$=250) (Table 3; Fig. S1 in Supporting Information). Unfortunately, a SF analysis was not possible on the innermost ridge (B1) owing to the lack of a suitably large surface area (~50 m$^2$). However, a LL of 40 mm was recorded on this ridge. A SF analysis ($n$=270) on the adjacent moraine (F9), however, did yield a single (unimodal) population and a LL=41 mm (Table 3; Fig. S1). Largest lichens of 38 mm on moraine B2 and of 34 mm on moraine B5 were also recorded, although no SF analyses could be conducted on either moraine, again owing to the small size of the available surfaces. Using the age–size curve constructed by Bradwell (2001) (age corrected to 2011), LL sizes suggest exposure of moraine B1 ~80 years ago (*c.* AD 1930) and of moraine B7 ~70 years ago (*c.* AD 1940) – an age difference of ~10 years (Table 3). It should be noted that even well-calibrated lichenometric dating has an optimum precision of only 5–10% (Innes 1988; Noller & Locke 2001), equivalent to an uncertainty of ±3 to 4 years on these ages. Consequently, bracketing lichenometric ages from Suite B are entirely consistent with the whole suite of minor moraines forming between *c.* AD 1935 and 1945, thus strongly supporting the hypothesis that the moraines formed annually.

Further supporting evidence comes from reliable measurements of the Virkisjökull ice front only 1000 m

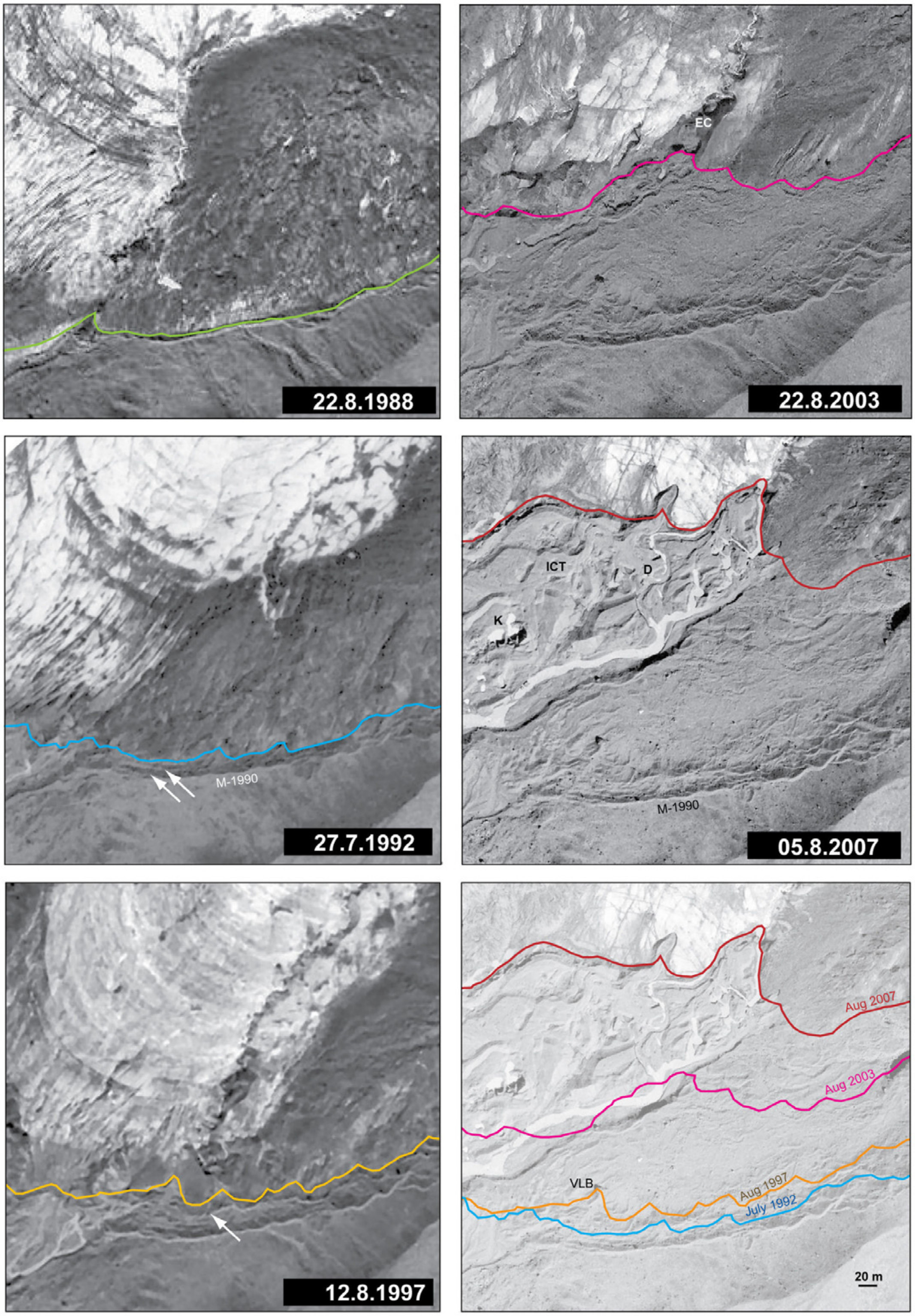


*Fig. 8.* Orthorectified aerial photos of Falljökull ice front covering Suite A minor moraines, spanning the period from 1988 to 2007. It can be seen that the whole suite of moraines formed during this time. Arrows indicate small push moraines formed in late winter 1991 and 1992, and the 1995/96 push moraine. By 2003 the ice front has evolved considerably, with an englacial/supraglacial stream system collapsing to form a large ice-walled chasm (EC). By 2007 a considerable portion of the ice front has become stagnant, forming a large area of sediment-covered ice-cored terrain (ICT). Subsidence features (K) and karstic drainage (D) are clearly visible. The compilation (lower right panel) overlays ice-front positions precisely mapped from four aerial photos to allow a visual comparison of retreat. VLB = very large boulder seen in Figs 4 and 6. See Table S1 for air photo details. This figure is available in colour at http://www.boreas.dk.

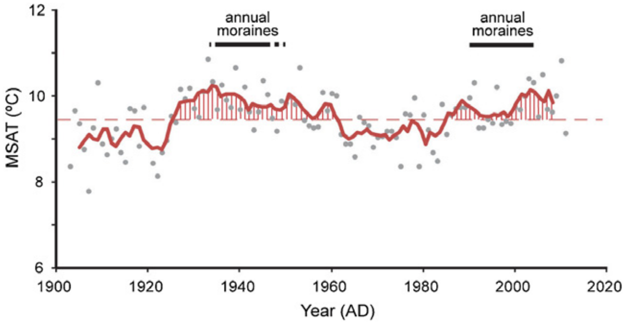


*Fig. 9.* Mean summer air temperature (MSAT) and periods of annual moraine formation at Virkisjökull–Falljökull. Temperature data (grey dots) are from Fagurhólsmyrí (June–September); the red line is the 5-year running mean; the horizontal dashed line is the mean of summer temperatures (1903–2003). Red hatching highlights periods above this long-term average value. This figure is available in colour at http://www.boreas.dk.

away (Sigurðsson 1998). This part of the ice margin receded rapidly (averaging ~30 m $a^{-1}$) during the years from 1935 to 1949, a phenomenon often associated with annual moraine formation in Iceland (cf. Sharp 1984; Boulton 1986; Krüger 1995; Bradwell 2004a).

## Discussion

Previous studies have shown that small-scale push moraines develop during seasons (normally winters) when forward ice movement exceeds ablation, particularly at high-mass-turnover glaciers (e.g. Boulton 1986; Bennett & Glasser 1996; Benn & Evans 2010), with distances between annual moraines representing the net ice-front recession in a single balance year. Moreover, annual moraine spacings have been correlated with above-average ablation-season (summer) temperatures from year to year (Bradwell 2004a; Beedle *et al.* 2009). At Virkisjökull–Falljökull both annual moraine sequences relate to periods of enhanced glacier recession and correspond with periods of elevated summer temperature (Fig. 9). When compared with the temperature data from the nearest long-running weather station, it is evident that Suite B formed during a run of unusually warm summers between *c.*1930 and *c.* 1950 (relative to the 100-year mean); similarly, Suite A formed in the period of predominantly warmer-than-average summers between 1990 and 2005 (Fig. 9). This general relationship between elevated summer temperatures and annual moraine formation is a strong one, and has been found elsewhere in Iceland (Boulton 1986; Krüger 1995; Bradwell 2004a). Interestingly, the temporal coincidence between annual retreat rates and years of increased summer warmth identified elsewhere (Bradwell 2004a; Beedle *et al.* 2009) implies an almost instantaneous response of the glacier front on a sub-annual time scale.

Simple field observations show that the effects of a mass-balance change can be first observed at the glacier front much more promptly (<0.5 years) than the glacier response time, which is usually of the order of decades in maritime glaciers (Jóhannesson *et al.* 1989; Paterson 1994). To distinguish the local short-term behaviour of the ice margin from the integrated longer-term behaviour of the whole glacier, we call the former the ice-front *reaction time*. Clearly, ablation-season temperatures (elevated or depressed) cannot solely account for changes in ice-front behaviour or reaction time from year to year, as measurements show that Virkisjökull–Falljökull was advancing during the warmer-than-average summers from *c.* 1930 to 1935 and *c.* 1986 to 1990. This suggests that the glacier's reaction time has been modified by the cumulative effect of mass-balance changes over the last 80 years, confirming that the link between air temperature and ice-front retreat is not straightforward. Unfortunately, we were not able to undertake a full statistical analysis of summer temperatures and retreat rates (annual moraine spacing) at this glacier because of uncertainties surrounding the formation date of those moraines formed prior to *c.* 1945 (±4 years), and the complexities introduced when a glacier oscillates against a reverse slope (Lukas 2012).

### *Comparison of ice front retreat rates: 1990s–2000s vs. 1930s–1940s*

Our results show that annual moraine spacing in Suite B is generally between 20 to 50 m, and hence the retreat of the Falljökull ice front between *c.* 1934 and 1945 was unusually rapid when compared with the measurement record (since 1957) (Fig. 3). A total glacier recession distance of 310 m occurred in 11 years – an average of 28 m $a^{-1}$. By comparison, the period of annual moraine formation from 1990 to 2004 resulted in ~210 m of ice-front retreat in 15 years – an average of 14 m $a^{-1}$. This shows that the two annual-moraine-forming periods of ice-front retreat are comparable in style but not in magnitude; the 1935–1945 period experienced stronger glacier recession, with average ice-front retreat rates twice as high as those between 1990 and 2004. This comparison may be complicated by the gently normal-sloping ground of Suite B, and the reverse-sloping ground of Suite A, which could suppress the retreat rate of the glacier during the latter period. However, since 2005 (up to September 2011), when no annual moraines have formed at the margin, the ice front has undergone even faster ice-front retreat on horizontal ground: a distance of 230 m in only 7 years, averaging 33 m $a^{-1}$. In fact, ice-front retreat of Falljökull since 2007 – a total distance of 187 m – represents the greatest amount of horizontal retreat within any 5-year period since measurements at this glacier began in 1932. It should also be noted that this accelerated recession (~35 m $a^{-1}$) since 2005 has been accompanied by strong thinning in the glacier's terminal zone.

According to ice-front measurements at Falljökull, every one of the last 20 years has seen net recession, with the exception of 1995–96, when the ice front advanced by only 1 m, and 2001–02, when no change was recorded. Only one other long uninterrupted retreat has been recorded at Virkisjökull–Falljökull: the period from 1935 to 1953 experienced 18 net recession (or no change) years in succession. This mid-20th century ice-front retreat is recorded at most glaciers across southern Iceland, and is thought to be a response to the pronounced summer warmth and lower than average precipitation during the 1930s and 1940s (Ahlmann 1940; Thórarinsson 1956; Jóhannesson & Sigurðsson 1998; Bradwell 2004a). The periods 1935–1953 and 1990–2011 compare quite closely in terms of their overall ice-front behaviour (–17 m $a^{-1}$ cf. –20 m $a^{-1}$) at Virkisjökull–Falljökull (Fig. 3). However, the most recent period is now the longest series of net recession years in the measurement record, and the ongoing ice-front retreat shows no signs of abating (Figs 3, 4). This trend and its magnitude imply that the current very rapid retreat is an exceptional and unusual event.

### *Glaciological implications*

This study has shown that the rate of glacier recession at Virkisjökull–Falljökull has clearly accelerated over the past 10 years, to a point where ~40 m $a^{-1}$ of horizontal retreat is now typical (Fig. 3). However, we have shown that retreat rates of 40 m $a^{-1}$ at this glacier are exceptional, at least in an ~80-year context (since 1932), yet mean summer air temperatures have not exceeded those of the 1930s–1940s, suggesting that something unusual has occurred to promote this behaviour. We suggest that the glacier has probably crossed a dynamic threshold and is now out of glaciological equilibrium, prompting rapid non-uniform changes at the margin as it struggles to re-equilibrate. We suspect that the drivers are internal to the glacier (as explained below), but conditioned by climate and inextricably linked with the 10-year run of warmer-than-average summers (2000–2010) (Fig. 9). Similar conclusions were reached by a team examining the rapid retreat and geometry changes of Alpine glaciers in Switzerland during the 1990s (Paul *et al*. 2004).

Using basic glaciological principles, we propose a simple conceptual model to explain the specific ice-front behaviour at Virkisjökull–Falljökull over the last 20 years. During average-warmth summers from 1990 to 1995 the glacier was generally in check with the climate, melting back steadily from a position of relatively good health in the late 1980s. Forward motion of the glacier by sliding and internal deformation each year was approximately equal to the amount of horizontal retreat during the average-length ablation season (May–September). This resulted in closely spaced (<10 m) relatively large annual push moraines (>1 m high). In fact, in 1995–96 ice-front retreat was insufficient to offset forward motion and a small net advance occurred. Between 1997 and 2004, however, the small amount of net forward glacier movement was more than matched by the considerable ablation-season recession, resulting in small (<1 m high) widely spaced (>20 m) annual moraines. This was probably in response to one or more of the following factors: increased melting during the normal ablation season (May–September); a lengthening of the ablation season (for instance from April to October); a decrease in forward glacier motion. Mean summer temperatures experienced a 50-year high during part of this time (Fig. 9) and hence increased ablation at the glacier margin seems highly likely, and, although we cannot demonstrate a decrease in forward motion during this time, this also seems probable. During the most recent period of very rapid frontal retreat from 2005 to 2011, when mean summer temperatures remained above average and no annual moraines formed, a further change occurred. We suggest that the rate of glacier forward movement has continued to decline over the last 7 years as the ice front has rapidly retreated and thinned, thereby decreasing the forward driving stress and ice deformation rate. As ice thickness and surface slope are the main drivers of glacier flow (Paterson 1994), a thinner glacier flows more slowly than a thicker glacier with the same surface slope. The prolonged run of unusually warm summers since 2000 has set up a type of feedback loop whereby continued ice-front recession and glacier thinning reduce forward driving stress, causing the glacier to flow more slowly. This decrease in forward glacier movement results in a net increase in ice-front retreat, under the same climatic conditions. We propose that a dynamic threshold has now been crossed at Virkisjökull–Falljökull and that the rate of forward motion has decreased to zero – effectively stagnating a large portion of the glacier terminus. A stagnant glacier margin cannot form annual push moraines as it is no longer undergoing forward motion. This glaciological switch from a dynamic to a stagnant glacier front explains the accelerated retreat since 2005 (~35 m $a^{-1}$), the lack of annual moraines over this time period, and the presence and preservation of deep 'annual' meltwater channels. Other evidence for large-scale stagnation of the ice front can be seen in the extensive development of ice-cored (glacier karst) terrain and the formation of a shallow supraglacial lake at the margin of Virkisjökull–Falljökull since *c.* 2005 (Fig. 8). Furthermore, examination of automatic time-lapse photos, taken every day from a fixed camera close to moraine A12, shows considerable glacier surface lowering (2–6 m) in the terminal zone, but no appreciable forward motion over a 13-month period (April 2011–May 2012) (Fig. 10). A network of recently installed GPS receivers (April 2012) on the ice surface will measure glacier motion in detail and seek to test

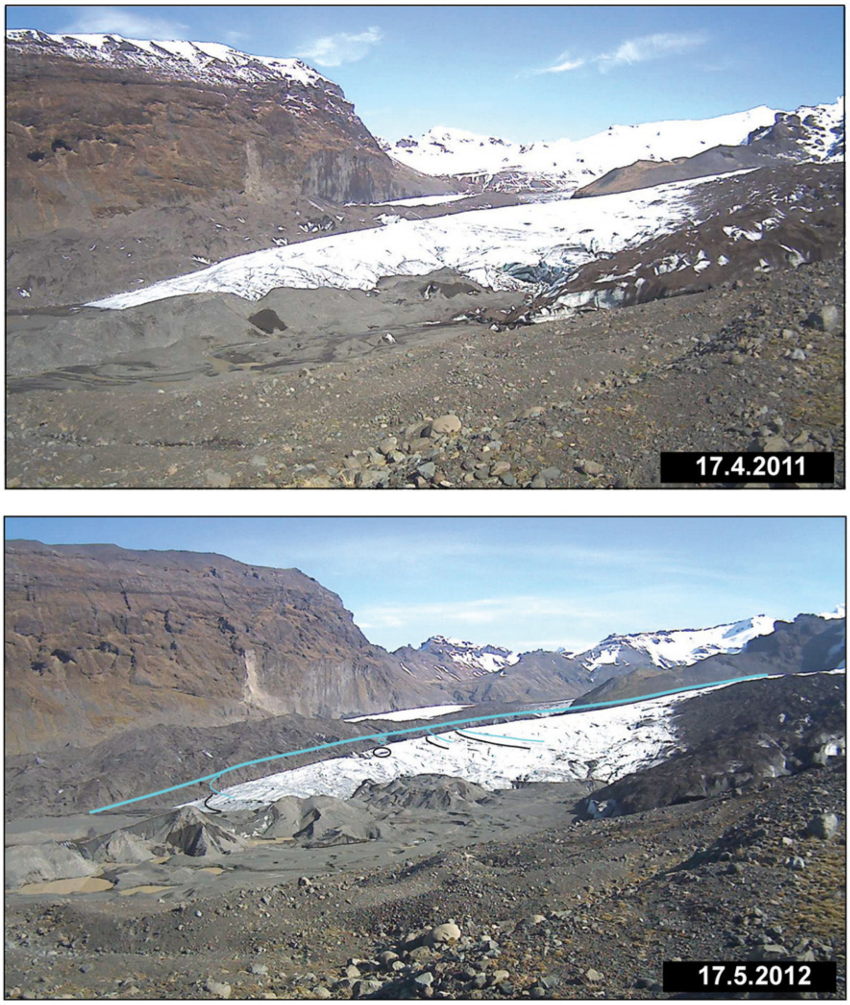


*Fig. 10.* Recent photos of Virkisjökull–Falljökull taken on 17th April 2011 and 17th May 2012 from a fixed time-lapse camera on the foreland. Blue lines show the position of ice-surface features in April 2011; black lines show the same features in May 2012. Note the ice-surface lowering, but no appreciable forward movement of the glacier, indicative of a stagnant margin. This figure is available in colour at http://www.boreas.dk.

this hypothesis further. Now stagnant, and no longer dynamically part of the glacier, we expect the margin of Virkisjökull–Falljökull to undergo enhanced retreat for some years, most probably by processes of collapse and *in situ* decay, until a new position of equilibrium is reached.

## Conclusions

The ice-front position of Virkisjökull–Falljökull – a high-mass-turnover outlet glacier in SE Iceland – has been measured annually since 1932. This has been supplemented by repeat photography (since 1996) and a detailed study of glacier margin evolution and proglacial geomorphology development (since 2009). The glacier has undergone over 500 m of ice-front retreat, punctuated by one major advance of ~200 m during the last 80 years. Specific conclusions from this work are as follows.

- Two suites of annual push moraines occur adjacent to the margin of Virkisjökull–Falljökull: the older probably formed between *c*. 1935 and 1945, based on geomorphological relationships, photographic evidence and lichenometric data; the younger suite formed between 1990 and 2004, based on field measurements and photographic evidence.
- Annual moraines typically range from 0.5 to 2 m in height, from 3 to 8 m in width, and have spacings of 1 to 50 m. Most moraines are continuous over lengths of more than 100 m, and have asymmetric cross profiles with steeper distal slopes.
- The formation of both suites of annual moraines on the foreland reflects periods of sustained ice-front recession over 10 years or more. Both suites formed during periods of predominantly warmer-than-average summer temperatures, suggesting that annual moraines can be generally used as a climate proxy.
- Using annual moraine spacing to deduce ice-front retreat rates, Virkisjökull–Falljökull underwent 310 m of retreat between *c*. 1935 and 1945, and 210 m of retreat between 1990 and 2004. This equates to average retreat rates of 28 and 14 m $a^{-1}$ respectively.

The years from 1935 to 1945 therefore represent the fastest ice-front retreat prior to the most recent period (2005–2012). During years of annual push-moraine formation the retreating glacier was in good (balanced) health and undergoing some forward motion. At this glacier, larger closely spaced annual moraines probably reflect a period of negative glacier mass balance and strong forward motion, while smaller more widely spaced annual moraines probably reflect greater ablation rates and less forward movement at the margin.

- Since 2005, annual push moraines have not formed at the margin of Falljökull even though the ice front has continued to retreat strongly in a period of unusually warm summers. The absence of annual moraines, the formation of 'annual' meltwater channels and the accelerated ice-front retreat rate (>30 m $a^{-1}$) since 2005 all suggest that a behavioural change has occurred at Virkisjökull–Falljökull. The period from 2007 to the present day is exceptional – with the most ice-front retreat in any 5-year period since measurements began (in 1932). We suggest that a dynamic glaciological threshold has now been crossed at this glacier.
- We propose that rapid ice-front retreat and thinning during a decade of unusually warm summers has decreased driving stresses in the glacier terminus sufficiently to cause forward motion to cease and to stagnate the ice margin. Since crossing a glaciological threshold in *c.* 2005, widespread stagnation of the ice front has resulted in the formation of a large portion of ice-cored terrain and a shallow supraglacial lake. The glacier margin is no longer undergoing dynamic retreat but is now undergoing non-uniform downwasting, decay and collapse through a range of geomorphological processes.

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## Supporting Information

Additional Supporting Information may be found in the online version of this article at the publisher's web-site:

*Fig. S1.* Lichenometric data collected from minor-moraine Suite B.

*Table S1.* Photogrammetric information relating to aerial photos of Virkisjökull–Falljökull used in this study.