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Surges of glaciers in Iceland

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ABSTRACT. Surges are common in all the major ice caps in Iceland, and historical reports of surge occurrence go back several centuries. Data collection and regular observation over the last several decades have permitted a detailed description of several surges, from which it is possible to generalize on the nature of surging in Icelandic glaciers. Combining the historical records of glacier-front variations and recent field research, we summarize the geographic distribution of surging glaciers, their subglacial topography and geology, the frequency and duration of surges, changes in glacier surface geometry during the surge cycle, and measured velocity changes compared to calculated balance velocities. We note the indicators of surge onset and describe changes in ice, water and sediment fluxes during a surge. Surges accomplish a significant fraction of the total mass transport through the main outlet glaciers of ice caps in Iceland and have important implications for their hydrology. Our analysis of the data suggests that surge-type glaciers in Iceland are characterized by gently sloping surfaces and that they move too slowly to remain in balance given their accumulation rate. Surge frequency is neither regular nor clearly related to glacier size or mass balance. Steeply sloping glaciers, whether hard- or soft-bedded, seem to move sufficiently rapidly to keep in balance with the annual accumulation.

INTRODUCTION

The geographically sparse distribution of surge-type glaciers and the relatively long period between glacier surges combine to limit our understanding of surge behaviour. Repeat investigations of surges on individual glaciers are rare, hence interstudy comparisons are pivotal for generalizing our understanding of surges. The high incidence of surge-type glaciers in Iceland (Fig. 1) and the proximity of settlements to several ice caps have resulted in a rich repository of historical observations (Thórarinsson, 1943). Annual glacier-front variations have been directly monitored on many outlets since 1930 (Fig. 2; Eyþórsson, 1963; Jóhannesson and Sigurðsson, 1998; Sigurðsson, 1998) and have been recorded in regular series of aerial photographs since the 1950s and satellite images since the early 1970s. In the second half of the 20th century, water-level gauges in glacial rivers have provided records of meltwater drainage and sediment transport during several surges (database of the Hydrological Service, National Energy Authority of Iceland; see Pálsson and Vigfússon, 1996). In addition to a synopsis of historical documents on glacier variations, we present the main results of our fieldwork on surging glaciers in the 1990s. Much of this information has thus far only been published in internal reports of the Science Institute, University of Iceland, and the National Energy Authority. Combined with historical accounts, these studies allow us to draw conclusions about the general pattern of surges of ice caps in Iceland (Figs 1 and 3).

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In the 1960s, Thórarinsson (1964, 1969) took up the task of compiling historical records and reviewing the state of knowledge on surges in Iceland. He summarized these records and described several 20th-century surges in detail. Thórarinsson estimated that between 1890 and 1964, a period generally characterized by glacial recession, most of Vatnajökull's outlets (except those in the southeast sector) had undergone surges, affecting approximately 40% of the ice cap. His synthesis led to the conclusion that surging glaciers in Iceland have a geometrical similarity, being characterized by flat ablation zones and shallow spoon-shaped basins that widen toward the glacier terminus. He noted for these glaciers that changes within their accumulation areas are not spread slowly along the glacier; instead stress accumulates until it has reached a

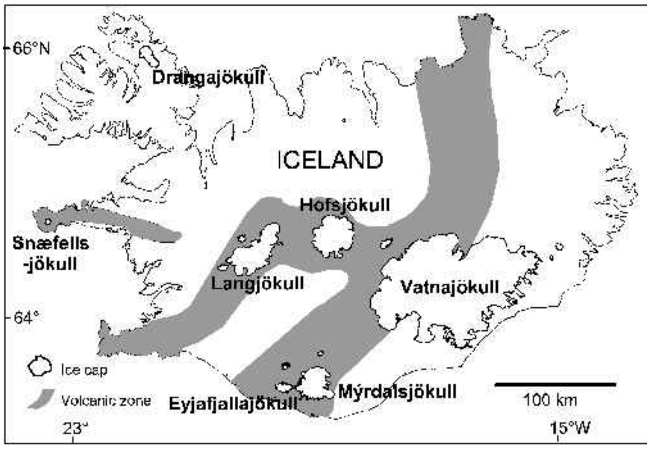

Fig. 1. Location map of the major ice caps in Iceland, with the active volcanic zone shaded in grey.

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Table 1. Reported surge advances in Iceland

Glacier	Area km²	Slope °	Years	Advance km	Area affected km²	Interval between surges years
Vatnajökull	8100					
Síðujökull[1, 2, 3, 4]	380	1.3	1893, 1934, 1963, 1994	0.5–1.2	500	41, 29, 31
Tungnaárjökull[5, 6]	308	1.4	~1920, 1945, 1994	1.2–2.0	550	~25, 49
Sylgjujökull[5, 6]	175	2.2	1945, 1996	0.5	150	51
Köldukvíslarjökull[16]	313	2.6	1975, 1992	0.2	100	17
Dyngjujökull[2, 3, 16]	1040	1.6	~1900, 1934, 1951, 1977, 1999	1.5	400–800	~34, 17,26, 22
Brúarjökull[2, 3, 7]	1500	1.7	1625, ~1730, ~1775, 1810, 1890, 1963	8–10	1500	~105, ~45, ~35, 80, 73
Brúarjökull east			1938			
Eyjabakkajökull[2, 4, 7]	110	2.7	1890, 1894?, 1931, 1938, 1972	0.6–2.8	110	4?, 37?, 37, 34
Skeiðarárjökull[8,9]	1	2.0	1787, 1812, 1857, 1873, 1929, 1985, 1991	1.0	1200	25, 45, 16, 56,56, 6
Breiðamerkurjökull[10, 11, 12]	540	2.5	1794, 1815, 1823, 1861, 1869, 1875, 1892, 1912, 1919, 1954, 1969	0.4–1.0	~600	21, 8, 38, 8, 6, 17, 20, 7, 35, 15
Eastern stream						
Middle stream	210	3.2	1978			
Hofsjökull	920					
Múlajökull[3, 12]	70	3.4	1924, 1954, 1966, 1971, 1979, 1986, 1992, 2002?	0.1–0.4	~10	30, 12, 5, 8, 7, 6, 10?
Blágnípujökull[16]			1975		~10	
Blöndujökull, Kvíslajökull[16]			~1920, 1975		~20	~55
Þjórsárjökull[12, 16]				0.2–0.3	35–100	
South			1991			
Middle			1994			
Langjökull	920					
Hagafellsjökull western[13, 14]	150	2.9	1971, 1980	1.2	200	9
Hagafellsjökull eastern[13, 14]	105	2.7	1974, 1980, 1999	1.0–1.6	100	6, 19
Mýrdalsjökull	600					
Sléttjökull western[16]	60	4.2	1992		50	
Öldufellsjökull[16]	40	4.1	1974, 1984, ~1992		40	10, ~8
Drangajökull	190					
Kaldalónsjökull[3, 12]	37	7.3	1741, ~1860, 1936, 1995	0.2	30	~119, 76, 59
Leirufjarðarjökull[3, 12]	27	7.3	~1700, ~1840, 1939, 1995	1	30	~140, ~99, 56
Reykjafjarðarjökull[3, 12]	22	6.6	1846, 1934	0.75		88
Central north Iceland						
Búrfellsjökull[15]	1	11.9	~1912			
Bægisárjökull	2	13.9	1801			
Teigadalsjökull[15]	0.5	14.3	1971	0.1		

Sources: [1] Nielsen (1937); [2] Thórarinsson (1964, 1969); [3] Eyþórsson (1963, 1964); [4] Thoroddsen (1892, 1905/06); [5] Freysteinsson (1968); [6] Guðmundsson and Björnsson (1992); [7] Todtmann (1955); [8] Jóhannesson (1985); [9] Björnsson (1998); [10] Björnsson (1996); [11] Pálsson (1945); [12] Sigurðsson (1998); [13] Sigbjarnason (1977); [14] Theódórsson (1980); [15] Hallgrímsson (1972); [16] Various unpublished data collected by the Science Institute and the National Energy Authority.

certain limit, when it is suddenly released. Thórarinsson also noted the likelihood of basal water lubrication playing a role in Vatnajökull surges, and correctly dismissed the suggestion of Nielsen (1937) that surges were triggered by subglacial volcanic eruptions. Thórarinsson further rejected suggestions that surges were triggered by earthquakes. This paper constitutes an extension of Thórarinsson's work on Vatnajökull, recognizing as he did that "few, if any, glaciated areas seem more likely than Vatnajökull to furnish us with sufficient data for the solution [for understanding catastrophic glacier advance]" (Thórarinsson, 1964).

POPULATION AND CHARACTERISTICS OF SURGING GLACIERS

Geographic distribution

All major ice caps in Iceland contain surge-type outlet glaciers, and surge-type glaciers occur almost exclusively as outlets of the major ice caps (Fig. 3). These ice caps are widely distributed across the country, and span the full range of climatic conditions represented in Iceland, from the warm and wet southeast coast (Vatnajökull, the largest non-polar ice cap) to the cool and drier northwest peninsula (Drangajökull). All glaciers in Iceland are warm-based, so we can rule out any surge mechanism for these glaciers that relies on a thermal transition. Several surge-type glaciers exist outside of the major ice caps, three of which can be found in the mountains of north central Iceland. Altogether 26 surge-type glaciers ranging in size from 0.5 to 1500 km^2 have been identified in Iceland. About 80 surge advances have been recorded, ranging from tens of metres up to 10 km.

All of Vatnajökull's major outlets are surge-type glaciers. Surge-affected areas of Vatnajökull are delineated in Figure 3a and occupy approximately 75% of the ice cap. The steep and active valley glaciers draining southeastern Vatnajökull and the northwestern flank of Vatnajökull near the ice-capped volcano Bárðarbunga are not known to surge. Surges that have led to advances of the glacier terminus are listed in Table 1. However, glaciers commonly experience uplift, crevasse formation and propagation of surface bulges over a period of weeks that only occasionally lead to a significant advance of the glacier terminus (personal communications from R. Stefánsson, 1991 and S. Björnsson, 2002). The local people

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have recorded these events as "gangur" (e.g. Björnsson, 1956), derived from the Icelandic word meaning "walk". These events have been documented on the small, steep outlets of the ice-capped volcano Öræfajökull in southern Vatnajökull, the steep southern outlets of Mýrdalsjökull ice cap and on surge-type glaciers such as Skeiðarárjökull and Breiðamerkurjökull.

Surges occur in the southern outlets of Langjökull (western and eastern Hagafellsjökull; Fig. 3b), but have not been reported in the steep western and eastern outlets. Surges occur in the relatively flat ablation area of Hofsjökull (Fig. 3c) and in the gently sloping northern outlets of Mýrdalsjökull (Sléttjökull and Öldufellsjökull; Fig. 3d). Hofsjökull and Mýrdalsjökull are the only major ice caps in which surges have not been observed to affect glaciers all the way to their central ice divides. The upper reaches of Hofsjökull and Mýrdalsjökull are relatively steep because the ice caps are draped over a single steep central volcano, and surges tend not to propagate up-glacier past sharp increases in the ice surface slope. Average surface slopes of surge-type glaciers generally fall in the range 1.6–4°, and surges in glaciers steeper than 3° typically only affect the low-sloping ablation areas. Most glaciers in Iceland are steeper than this. The average mean slope of the 45 non-surge-type glaciers in the Iceland Glaciological Society database is 11.8°, ranging from 2.9° to 25.7°. A statistical analysis of an Iceland glacier inventory (253 glaciers, 19 surge-type) found that median surface slope was about 7° for non-surging glaciers, and about 2° for surge-type glaciers (Hayes, 2001). Surge-type glaciers are generally absent in the smaller ice caps (e.g. the individual ice-covered volcanoes such as Eyjafjallajökull and Snæfellsjökull). Exceptions to the low-sloping collection of surge-type glaciers are found only in hard-bedded areas of the Tertiary basaltic region: the three surge-type outlets of the ice cap Drangajökull have slopes around 7°, and the three known surge-type cirque glaciers in north central Iceland (Bægisárjökull, Búrfellsjökull and Teigardalsjökull) have slopes exceeding 10°. In total, known surges affect 60–70% of the ice-covered area in Iceland, with surge-type glaciers outsizing non-surge-type glaciers on average and having no discernible orientational bias.

Subglacial geology and topography

The geological substrate of surge-type glaciers in Iceland is as varied as the geology of Iceland itself, and includes both hard and soft beds, with substantial subglacial mountain ridges extending both along and transverse to the direction of flow. These substrates range from Tertiary basalts in northwest and eastern Iceland to Quaternary and Holocene basalts and hyaloclastites in the active volcanic zone (Jóhannesson and others, 1990). Holocene basalt lavas are highly permeable (hydraulic conductivity $>10^{-5}$ m s^{-1}) but other substrates less so (Sigurðsson, 1990). Tertiary lava is almost impermeable. We infer subglacial bed types and their relative permeabilities from the geology exposed at the ice margins and from glacier drainage characteristics. The flanks of the ice caps are believed to rest on unlithified deposits, while the central regions of the ice caps probably overlie stiff till or solid rock.

The largest ice cap, Vatnajökull, partially overlies the active volcanic zone and covers several of Iceland's largest volcanoes. It straddles a distinct geological boundary coincident with the margin of the active volcanic zone (see Fig. 1). Porous lava beds primarily underlie western Vatnajökull, while the

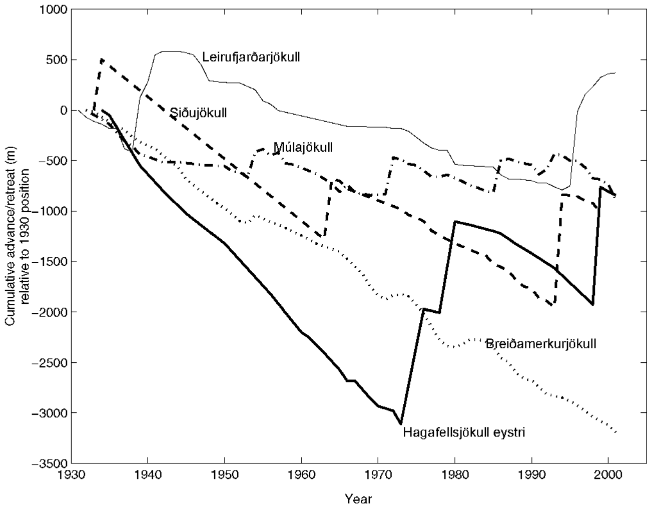

Fig. 2. Front variations of five surge-type glaciers in Iceland since about 1930 (Iceland Glaciological Society database). Prior to 1964, variations of Síðujökull are inferred from Thórarinsson (1964). Breiðamerkurjökull (eastern branch) has been partly a tidewater glacier since 1933.

main outlets of eastern Vatnajökull rest on impermeable unconsolidated till (Björnsson, 1988; Jóhannesson and others, 1990; Sigurðsson, 1990; Björnsson and Einarsson, 1991). The steep non-surging outlets along the southeastern flank of Vatnajökull rest on impermeable bedrock. The other large ice caps, Langjökull, Hofsjökull and Mýrdalsjökull, lie entirely within the active volcanic zone (see Fig. 1). Of those, the northern outlets of Mýrdalsjökull and southern outlets of Langjökull may be partially underlain by porous lavas. The beds of surge-type valley glaciers in northern and northwestern Iceland (Bægisárjökull, Búrfellsjökull, Teigardalsjökull and outlets of Drangajökull) are composed of hard impermeable Tertiary basalts. Despite the presence of numerous geothermal areas beneath the ice caps in Iceland, surge-type glaciers are not spatially correlated with these basal heat sources. Subglacial meltwater produced by geothermal heating tends to accumulate locally and drain episodically, rather than contribute to the ambient basal drainage regime.

All of the major ice caps in Iceland have been mapped by radio-echo sounding (Björnsson, 1988; Björnsson and others, 1991, 1992, 2000). Bedrock overdeepenings of 50–100 m (relative to the proglacial outwash plain) are common beneath the large surge-type outlet glaciers. Skeiðarárjökull and Breiðamerkurjökull, the most active southern outlets of Vatnajökull, excavated troughs 200–300 m below sea level during their advance from 1400 to 1900 (Björnsson, 1996). Our recent hydrologic modelling of Vatnajökull suggests a correlation between the extent of several surge-affected areas and a predisposition for low basal effective pressures. This is probably related to the spoon-shaped basins described by Thórarinsson (1964) and their ability to retain water.

Surge frequency and duration

Major surges, with return periods ranging from several years to a century, have occurred in all of the large lobate outlets on the northern, western and southwestern flanks of Vatnajökull (Fig. 3a). The same applies to some of the broad lobes of Langjökull and Hofsjökull, and to the northern outlets of Mýrdalsjökull (Fig. 3b–d). The timing and duration

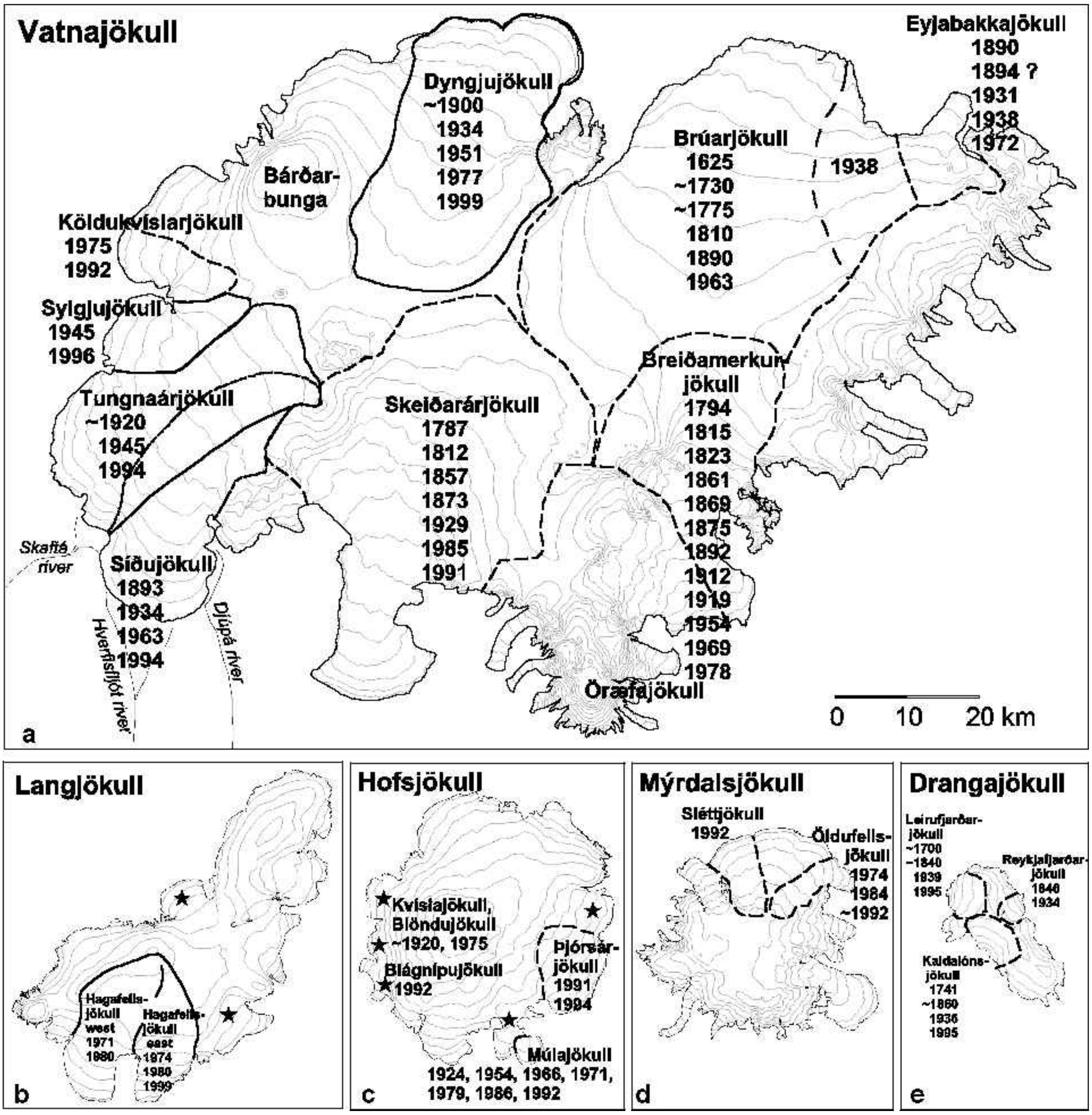

Fig. 3. Geographical distribution of known surge-type glaciers within the major ice caps, and the approximate extent of surge-affected areas. Solid and dashed lines indicate well-known and less certain boundaries, respectively. Major surge-type outlet glaciers are labelled, along with the dates of known surge advances. Stars indicate outlets for which there is some anecdotal evidence of surging. It is likely that some surge-type outlet glaciers have not yet been identified.

of surges that have caused observed advance of the glacier terminus are summarized in Table 1. Surges do occur that do not lead to glacier terminus advance, but because they are more difficult to distinguish in the historical record they are excluded from this list. Reported surge intervals should therefore be taken as maximum estimates.

Observations suggest that certain glaciers surge at fairly regular intervals. Síðujökull has surged at 30 year intervals since the 1930s, Dyngjujökull at 20–30 year intervals, Brúarjökull every 80–100 years and Múlajökull about every 10 years. The recorded surge history of other glaciers points to a variety of surge intervals. Breiðamerkurjökull, one of two major southern outlets of Vatnajökull, illustrates this point: it is known to have surged 11 times between 1794 and 1969, at intervals ranging from 6 to 38 years. The western branch of Skeiðarárjökull, the other large southern outlet of Vatnajökull, surged four times at increasing intervals of 25–118 years between 1787 and 1991. Its middle branch surged four times between 1857 and 1991 at decreasing intervals of 72–6 years. This simultaneous surge-interval increase in the western branch and surge-interval decrease in the middle branch suggests a peculiar relationship between neighbouring glaciers (see Table 1). All southern outlets of Vatnajökull have experienced a similar mass balance during the 20th century (Björnsson, 1980; Björnsson and others, 1998), but their dynamics have apparently been differently affected. Surges in southern Langjökull since the 1970s occurred after at least 40 years of quiescence and may be a response to positive mass balance from the 1960s to 1980s.

Fieldwork on glaciers in Iceland since the early 1990s has revealed that the surge process can take 2–3 years from the first signs of increased sliding and the subsequent down-glacier propagation of a surge wave. Prior to regular monitoring of glacier surface velocity, surge duration was gener-

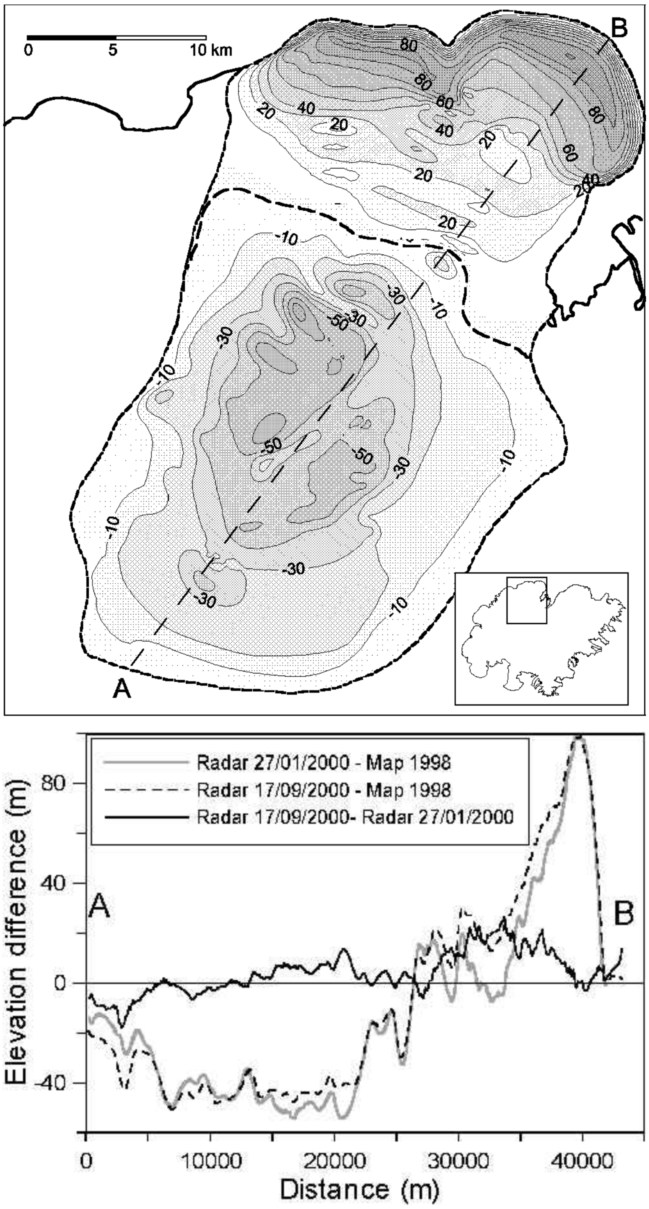

Fig. 4. Changes in the surface of Dyngjujökull during the 1998–2000 surge. Upper panel: Interpolated change in surface elevation from the beginning to the end of the surge. Heavy dashed line separates zones of thickening and thinning. Lower panel: Change in surface elevation along profile A–B in upper panel for three different time periods during the surge.

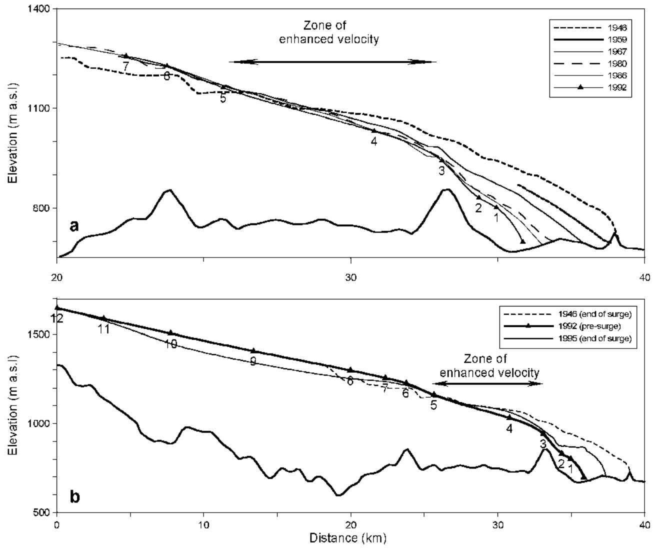

Fig. 5. Observed surface profile of Tungnaárjökull, 1946–95. (a) Surface profiles over 50 years following a surge. (b) Surface profiles between and after two surges. Note the difference in horizontal scales in (a) and (b). The numbered triangles are poles for measuring velocity and mass balance.

ally considered to be less than half a year, as surge onset was associated with increased turbidity in glacier rivers and advance of the glacier terminus. The advance of the terminus, however, lasts in most cases about 2–3 months regardless of the size of the glacier and the distance of the advance. Lingering effects of a surge can often be detected in the accumulation area in the form of crevassing and surface lowering several years after the terminus has stopped advancing.

FIELD MEASUREMENT AND OBSERVATION OF SURGING GLACIERS

Regular monitoring of the mass balance and outlet velocities of Vatnajökull, Hofsjökull and Langjökull has been part of glaciological field campaigns for the last 10 years (Björnsson and others, 1998; Sigurðsson, 2001). About 10 surges have occurred in these ice caps during this period, some of which have been reported elsewhere (Björnsson, 1998; Sigurðsson, 1998). Results presented in these sources show that the known surge-type outlet glaciers of Vatnajökull generally move too slowly to remain in balance with their accumulation rates. The same analysis also shows that glaciers not known to surge, such as the steep northern part of Köldukvíslarjökull, move at velocities comparable to the calculated balance velocities.

Changes in surface geometry and mass flux during surges

Surges have a marked impact on the geometry of the ice caps, typically thinning the accumulation area by 25–100 m. They also play an important role in the mass flux through the outlet glaciers. During the 1990s, 3000 km^2 of Vatnajökull (38% of the ice-cap area) were affected by surges, which transported about 40 km^3 of ice from the accumulation area to the ablation area. This is approximately 25% of the total ice flux to Vatnajökull's ablation area during that period. For some outlet glaciers the contribution of surges to mass transport is higher still. During the 1998–2000 surge of Dyngjujökull, 13 km^3 of ice were transported to the ablation area (Fig. 4) out of the total 20 km^3 of ice accumulated during the preceding 20 year quiescent period.

Repeatedly measured surface profiles on Tungnaárjökull in western Vatnajökull show a classic example of the surge-related cycle of mass accumulation and expulsion (Fig. 5). For approximately 50 years following the surge that ended in 1946, Tungnaárjökull thickened in the reservoir area and thinned and steepened in the receiving area. The next surge of Tungnaárjökull in the early to mid-1990s resulted in a readvance of the glacier terminus relative to its measured position in 1992, and surface drawdown in the reservoir area extending 30 km up-glacier from the terminus.

Surge onset and propagation

After several years of glacier surface steepening, we observe velocity increases in a zone typically 10 km long (in Vatnajökull and Langjökull) and centred in the upper ablation area. We refer to this area as the enhanced velocity zone.

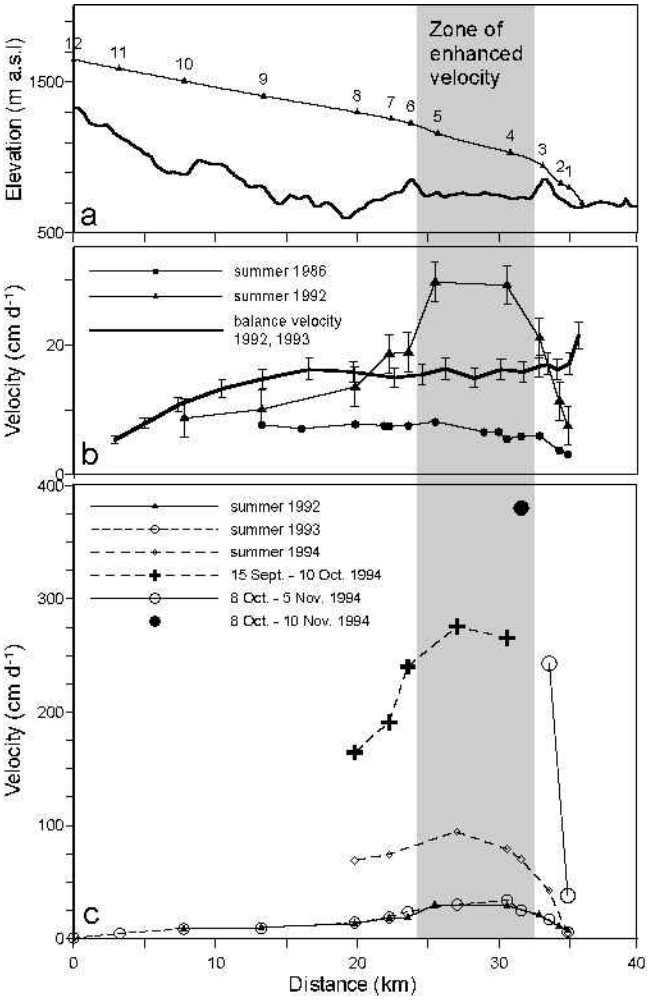

Fig. 6. Changes in surface velocity associated with a surge of Tungnaárjökull. (a) Glacier surface (1992) and bed topography along the profile, with 1992 stake locations. (b) Measured velocity profiles before (1986) and during (1992) the surge, compared to the computed average balance velocity profile for 1992–93. The number of velocity stakes on the profile varied from year to year. Note the error bars. (c) Selected velocity measurements along the profile during the 1992–94 surge. Note the difference in vertical scale between (b) and (c).

Figure 6a shows 1992 surface stake locations and basal topography along a profile on Tungnaárjökull, and Figure 6b shows measured velocity profiles for 1986 (pre-surge) and summer 1992 (at the onset of a surge) compared to the calculated average balance velocity for 1992–93. The 1986 pre-surge velocity profile shows a clear deficit relative to the computed balance velocity profile, while the 1992 profile shows velocities in excess of balance in the lower ablation area. Velocity profiles along a flowline through this zone are usually bell-shaped. Upstream from the velocity peak along the profile (e.g. between stakes 5 and 6; Fig. 6a), crevasses are formed and a slight surface lowering is observed, while downstream from the velocity peak the glacier thickens. Velocities continue to show seasonal variations but generally increase over 2–3 years, while the maximum velocity along the profile remains within the enhanced-velocity zone (Fig. 6c). Hence, the location of maximum surge velocities appears to be relatively stable, rather than migrating up- or down-glacier. Figure 7 illustrates the dramatic velocity increase experienced in lower Tungnaárjökull during this surge that occurred in the early 1990s.

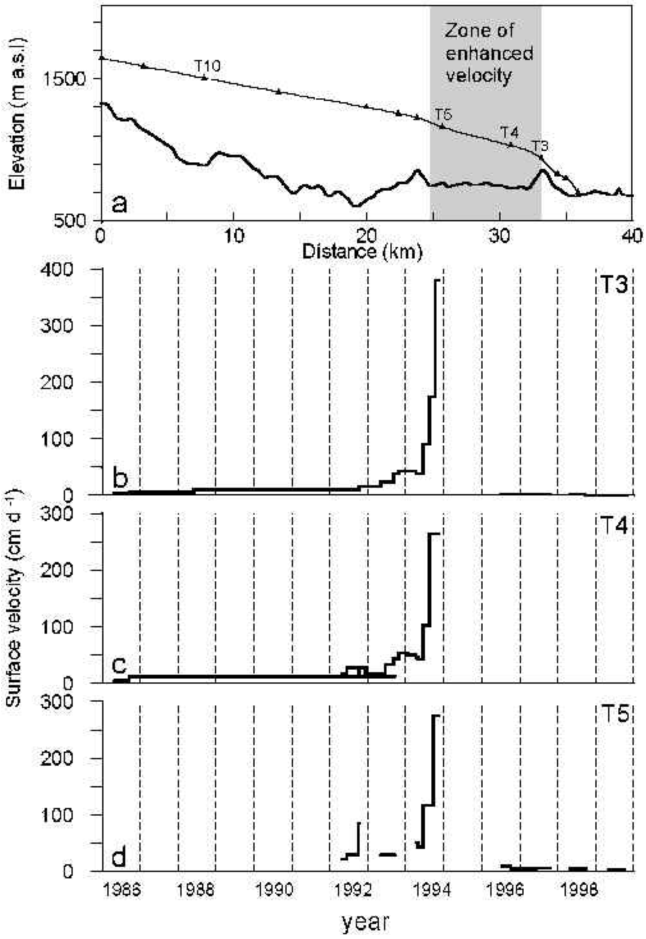

Fig. 7. Temporal changes in surface velocity along a profile on Tungnaárjökull during surge onset. (a) Glacier surface and bed topography along the profile, and stake positions. (b–d) Velocity records from stakes T3 (b), T4 (c) and T5 (d).

Over a period of a few months, a step-like thickening of the glacier develops in the lower part of the enhanced-velocity zone several kilometres from the terminus. This development does not occur preferentially during any particular season of the year. The first unquestionable sign of a surge is the advance of this bulge, usually tens of metres high, at rates measured at 20–80 m d^{-1}. Propagation of the bulge to the glacier terminus typically requires less than 1 year, and often less than 6 months, during which time its crest remains relatively uncrevassed. Once the bulge reaches the terminus, the glacier begins to advance as a vertical front 20–50 m high. For most surge-type glaciers in Iceland, advance of the terminus lasts several months. Notable exceptions to this are the steep (around 7°) hard-bedded outlets of Drangajökull which advance for 5–6 years during a surge. The maximum advance rate measured during a surge in Iceland is 100 m in 24 hours on Brúarjökull in 1964 (Thórarinsson, 1969) and Síðujökull in 1994 (Science Institute, unpublished data). Four of the large outlets of Vatnajökull typically advance by about 1 km (Skeiðarárjökull, Síðujökull, Tungnaárjökull and Dyngjujökull; see Table 1), while Brúarjökull usually advances by 8–10 km.

Surge-related hydrology

The muddying of glacial rivers and the increased number of hydraulic outlets at the glacier terminus (Björnsson, 1998) are two robust hydrological signatures of surging in both hard- and soft-bedded Icelandic glaciers. The latter is a

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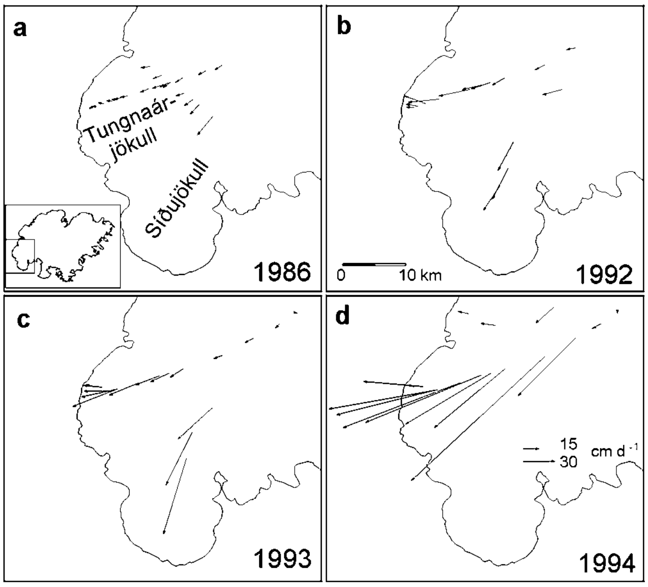

Fig. 8. Surface velocity vectors along profiles on adjacent outlets Síðujökull and Tungnaárjökull in western Vatnajökull, capturing the initiation of surges on both glaciers.

common manifestation of the conduit-to-distributed transition in the basal hydraulic system that accompanies a surge (e.g. Kamb and others, 1985). Distributed drainage promotes basal water lubrication, for which there is much evidence as a facilitator of fast glacier flow (e.g. Raymond, 1987). Well-lubricated basal glide beneath surging glaciers in Iceland is suggested by a general lack of push-moraine formation in front of an advancing terminus. An alternative explanation for this is that surges proceed dominantly by deformation, rather than sliding. However, this notion remains to be reconciled with the very rapid rates of glacier advance combined with low ice surface and bed slopes.

During a surge, long-term dislocation of ice and water divides can take place high on the glacier. Surge-induced changes in the glacier surface structure alter the subglacial hydraulic catchment structure. A natural experiment demonstrated this in 1994 and 1995, when jökulhlaups from central western Vatnajökull partly drained to the river Hverfisfljót instead of the river Skaftá. Surface lowering of Síðujökull (feeding the river Hverfisfljót) during its 1994 surge extended its subglacial water catchment northward, causing this temporary change in flood routing. After Tungnaárjökull stopped surging in 1996, Skaftá jökulhlaups resumed their original course to the river Skaftá (Zóphóníusson and Pálsson, 1996). Although outburst floods from subglacial geothermal areas have occurred during surges (e.g. Björnsson, 1998), surge timing is not related to flood incidence nor has a flood ever been known to trigger a glacier surge in Iceland.

It remains an open question whether surge termination is related to release of stored water. In runoff records during 20 surges, both small and large, no indication can be found of a flood occurring at the termination of a surge. River gauges from which these data were collected tap into large and sometimes multiple catchment basins, which are usually only partially glacierized, so one might expect some surge-related variations to be masked. Floods have been known to occur without interrupting surges, such as took place in 1998 at Drangajökull. Fluvial erosion of deep new watercourses in front of western Hagafellsjökull in Langjökull (Fig. 3b) in autumn 1998 invites speculation as to whether a sudden flood terminated an incipient surge. Whether or not a flood occurs, meltwater production typically increases for several years following a surge, due to the increased surface area of the ablation zone and new crevasses exposed to turbulent heat fluxes and radiation. On Tungnaárjökull and Dyngjujökull, a 30% increase in runoff the first year after the surge can be ascribed to this effect, of which 5% was due to the altered hypsometry.

Muddy discharge appears in the wake of a propagating surface bulge, which acts as a dam to basal water, when it intersects the glacier margin. While the total glacier runoff does not change appreciably during a surge, the sediment load (and hence sediment concentration) increases substantially, especially in the finest grain-sizes (Sigurðsson, 1995; Pálsson and Vigfússon, 1996). During surges in Vatnajökull, the sediment concentration in affected outlet rivers is generally 7–10 kg m^{-3} for 1–2 years (National Energy Authority database, see Björnsson, 1980; Pálsson and Vigfússon, 1996). These concentrations are comparable to those during glacier outburst floods. During the 1963–64 surge of Brúarjökull, the river Jökulsá á Brú had an average suspended-sediment concentration of 6.5 kg m^{-3}. In 1964 the total sediment load was 2.5×10^{10} kg (river discharge of 120 m^3 s^{-1}), equivalent to a denudation rate of 12 mm a^{-1} over the 1500 km^2 affected by the surge. The aftermath of the surge was seen in decreased annual suspended load up to 1980 (Tómasson and others, 1996; Pálsson and others, 2000). During surges of Hagafellsjökull in southern Langjökull, sediment concentrations in the outlet rivers increase on average from 0.2 to 1.0 kg m^{-3}.

Surges in adjacent glaciers

Most surge-type glaciers in Iceland are outlets of ice caps that during surges have sharply defined boundaries with respect to their neighbours, in both the reservoir and receiving areas. Observations from neighbouring Tungnaárjökull and Síðujökull demonstrate that surges can cause migration of ice and water divides in the ice cap. The dynamic relation between these adjacent glaciers is complex. Sometimes they surge separately, as Tungnaárjökull did in 1945 and Síðujökull did in 1963, and sometimes in tandem as in the early 1990s. During their independent surges, Tungnaárjökull and Síðujökull scavenge ice from one another's drainage basins, resulting in a 200 km^2 area that has participated in surges of both glaciers. The same applies for water, as exemplified by the previously mentioned drainage of Skaftá jökulhlaups in 1994 and 1995 to the river Hverfisfljót after the surge of Síðujökull.

Measured surface velocities on western Vatnajökull illustrate the acceleration of Tungnaárjökull and Síðujökull in the early 1990s (Fig. 8). Crevasse formation on upper Síðujökull (personal communication from M. T. Guðmundsson, 1990) and anomalous low-frequency seismic activity in summer 1990 were the first indicators of a surge (personal communication from B. Brandsdóttir, 1990). In January 1994 a 70 m high crescent-shaped bulge was observed on the glacier surface. The river Djúpá, draining the eastern margin of Síðujökull, discharged anomalously turbid water after the passage of the bulge. The bulge propagated down-glacier at a rate of 20 m d^{-1} to the terminus, and turbid water was eventually discharged from all of the major outlet rivers draining Síðujökull. This surge lasted for approximately 4 months after the bulge was first observed, affecting a 500 km^2 area and resulting in a 1150 m advance of the glacier terminus.

Surge activity continued in western Vatnajökull through most of the 1990s. Velocity increases were first detected on Tungnaárjökull in 1992, and in 1993 the measured velocity of central Tungnaárjökull (Figs 6–8) was more than double its 1986 value. By the summer of 1994, velocities had increased significantly over most of the length of the glacier. Figure 8 shows this dramatic increase in speed as well as the southward deflection of velocity vectors toward the depressed basin of Síðujökull. By the autumn of 1995, Tungnaárjökull had advanced a total of 1200 m. In 1994 it became obvious that Sylgjujökull (north of Tungnaárjökull) was surging, and this surge may have continued until 2000 in its northernmost region.

Hagafellsjökull, in southern Langjökull (Fig. 3b), demonstrates similar surge-pattern complexity. Western and eastern Hagafellsjökull started surging separately in 1971 and 1975, respectively. In 1980, both branches of Hagafellsjökull surged together. From summer 1997 to autumn 1998 the surface velocity of western Hagafellsjökull increased from 50 to 250 m a^{-1} in a zone 4–12 km above the terminus, but then suddenly dropped down again. It appears as though an incipient surge in the western branch aborted, while a surge of eastern Hagafellsjökull that had begun around the same time continued.

SUMMARY AND CONCLUSIONS

Surge-type glaciers in Iceland are both hard- and soft-bedded, overlying a variety of volcanic substrates and situated in regions subject to a broad range of climatic conditions. Although many of Iceland's ice caps overlie active geothermal areas, there is no special correlation between the locations of surge-type glaciers and geothermal heat sources. Most surge-type glaciers are gently sloping outlets (typically 1.6–4°) of large ice caps, often located in overdeepened valleys and exhibiting relatively low annual velocities. In ice caps with steep accumulation areas, surges are limited to the more gently sloping ablation zones. Otherwise, surges affect glaciers up to their central ice divides. Surge intervals vary between glaciers from several years up to a century. Several glaciers demonstrate a regular surge periodicity, but most do not.

Measurements of glacier surface velocity often show acceleration over 2 or 3 years prior to other visible signs of surging. Velocities increase first and remain highest in a zone within the upper accumulation area. Development and propagation of a surface bulge often accompanies a surge, with the location of this bulge marking the transition between a distributed basal drainage system upstream and an ordinary (usually channelized) drainage system downstream. Glacier advance occurs when this bulge reaches the terminus, which typically requires 6 months to 1 year. During a surge, sediment concentration in the outlet rivers is very high, especially in the finer grain-sizes. Hydrological data covering 20 surge events provide no evidence for either increases in total runoff during a surge or significant flooding associated with surge termination. In no case has a glacier outburst flood ever been known to induce surging.

Surges have a significant impact on the dynamics and hydrology of Iceland's temperate ice caps. They reduce ice-surface slopes, alter glacier hypsometry through mass transport, and increase glacier surface area and roughness. The latter, along with ice deposition at low elevations, accelerates surface melting from solar radiation and turbulent heat exchange, resulting in increased runoff to glacial rivers following a surge. Surges alter the overall geometry of the ice caps and perturb ice and water divides. In Vatnajökull about 25% of the total ice surplus in the accumulation area during the 1990s was transported down-glacier by surges. During the entire 20th century, surges accomplished at least 10% of the total ice flux to the ablation area.

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