Glacier surge at Usherbreen, Svalbard

JON OVE HAGEN

Hagen, J. O. 1987: Glacier surge at Usherbreen, Svalbard. *Polar Research 5 n.s.*, 239–252.

Usherbreen started to surge in 1978, and the front has advanced 1.5 km and covered an area of 4.5 km^2. During the first two years the front advanced more than 1 m/d, and the front was still advancing 0.15–0.20 m/d in 1985, seven years after the start. The mean gradient of the lower 7 km decreased from 3.3 grad. to 1.8 grad. during the surge. The volume of ice transported down the glacier from higher to lower parts during the surge was 815×10^6 m^3, which is almost 20% of the total glacier volume. Old icecored ridges in front of the glacier were reactivated, and the whole ridge system was pushed forward, in the summer of 1985 at a speed of about 0.05 m/d. Parts of the ridge system were moved 200 m during this surge. New ridges were developed on the flat sandur in front of the old ridge system. This demonstrates that the glacier advanced further than in any previous surge.

Jon Ove Hagen, Norsk Polarinstitutt, P.O. Box 158, 1330 Oslo Lufthavn, Norway; November 1986 (revised March 1987).

Surge is a periodic advance of an ice mass with increased velocities up to hundred times the normal. The surge mechanisms are one of the unsolved and challenging topics in glaciology, and field observations are still needed to obtain a better understanding.

Both temperate, subpolar (with temperate and cold parts) and polar (cold) glaciers may surge, but it seems to be most common in subpolar ones. In Svalbard most of the glaciers are subpolar and 90% of them are assumed to be of the surge-type (Liestøl pers. comm.).

A surging glacier will in the quiescent stage have an ice flux that is too low to maintain the steady-state profile. The surface gradient will then gradually increase and when the basal shear stress reaches a critical, but unknown value the surge starts (Meier & Post 1969), and the sliding velocity increases rapidly. The motion usually starts in the upper or middle parts of the glacier resulting in a compressive flow further down. A compressive glacier surface strain rate of up to 0.2 per day has been measured (Kamb et al. 1985), which results in a wave motion of thicker ice that increases its transport capacity.

The ice velocity in a surge varies in different glaciers. 2–5 m/d is common, but up to 100 m/d has been recorded at Brùarjökull, Iceland (Thorarinsson 1969).

So far, the trigger mechanism has not been explained adequately, but most probably it is a combination of increasing basal shear stress and increasing subglacial water pressure (Kamb et al. 1985). However, the water must then be prevented from draining out while the glacier is approaching a new surge. At subpolar glaciers this may happen as the outermost parts are frozen to the bed, preventing the water from draining out, while the central parts are at the melting point. Schytt (1969) has done some measurements on Vestfonna, Svalbard, that indicate such a temperature distribution with a cold ring surrounding a central area of melting ice.

Liestøl (1976) registered a surge at Hessbreen, Svalbard, in 1973–74. Temperatures at 10 m depth were −4.2°C before the surge in the lower part of the glacier tongue where the ice thickness was less than 100 m. This indicates that the tongue has been frozen to its bed, damming the subglacial water and preventing any sliding. However, water was draining year round from the glacier, so parts of the glacier sole must have been at the melting point. The glacier seemed to move as a solid block during the surge. Crevassing only developed in a zone along the edge and in the upper part of the glacier. Liestøl's observations suggest that the surge started at the tongue and not in the accumulation area. Similar block motion has also been reported from the surge of Tindebreen in the Hornsund area of Svalbard by Pillewizer (1939).

An observation from Tillbergfonna in Gangdalen, Svalbard, also indicates subglacial damming of water near the front. Water stored under the glacier tongue drained out in February–March 1970, and the central parts of the glacier tongue developed a marked depression. The outer part

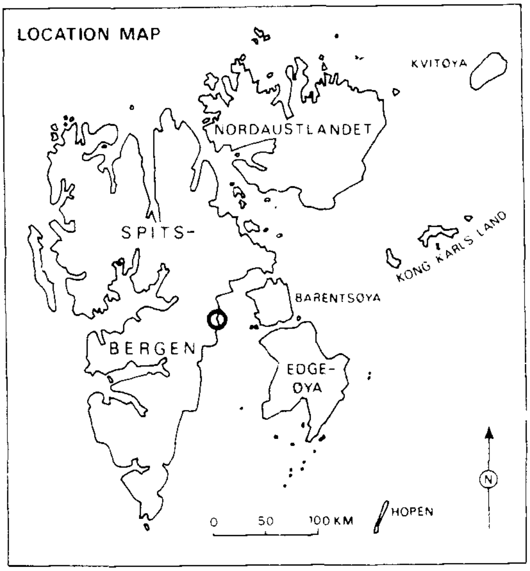


Fig. 1. Location map.

of the tongue was not affected, while there was some crevassing along the depression (Liestøl pers. comm.). Liestøl estimated from air photo studies that about $1 \times 10^6\,m^3$ water was drained. The slope of the glacier surface was 5–6 deg.

It has been proposed that the water pressure at temperate glaciers builds up during wintertime when outgoing channels are closed by freezing and ice deformation (Kamb et al. 1985). If this hypothesis is correct it requires surging of temperate glaciers to start in late winter/early spring when water pressure is at its maximum. This is not verified from observations.

Relatively few glaciers have been observed in active surge. Two examples are Medvezhiy Glacier in the Pamirs in Soviet Central Asia (Dolgushin & Osipava 1973) and Variegated Glacier in Alaska (Kamb et al. 1985).

In Svalbard, surges have been dated for more than 80 glaciers (Liestøl in press), but a lot more have surged at unknown times. The change of longitudinal profiles and front positions has been recorded at some of these glaciers (Liestøl 1969).

Usherbreen

Usherbreen is an outlet glacier that drains eastwards from the icecap Nordmannsfonna, West-Spitsbergen (Figs. 1 and 2). Before the surge the glacier had an area of 29 km^2, it was 12 km long, and the elevation ranged from 550 m above sea level to 10 m a.s.l. at the sandur. Its average slope was 2.5 grad. The bedrock consists of marine schists and sandstones up to about 300 m elevation, and above this Trias layer there are some grey and black Jurassic shales (Flood et al. 1971).

The glacier was in the final act of a surge in 1985. Field work was concentrated on two tasks: 1. recording the glacier change during the surge, and 2. studying the landforming processes in front of a surging glacier. This paper will mostly describe the results under 1, while the second task will be discussed in a later paper.

Results and discussion

Glacier map before and after the surge

Pre-surge air photos from August 1971 were available. Fixed triangulation points were surveyed in the field, and later a glacier map was constructed on a Wild B8 Autograph Stereoplotter at the Department of Geography, University of Oslo to the scale of 1:20,000. The map is here reproduced on the scale of 1:75,000 (Fig. 3).

The exact start of the surge has not been established, but it must have been close to 1977. The upper part of the glacier was crevassed in 1978, and from Landsat satellite images it was observed that the glacier front started to advance this year. Satellite images from 1980 showed that the front had advanced more than 1 km. The front moved more slowly in the following years, and the field work in 1985 showed that the advance had practically ceased, so 1985 can be taken as the maximum position. The assumed minimum position before the surge is obtained from the air photos from 1971 even though the glacier might have retreated slightly further before the surge.

The glacier map from 1985 was drawn manually on the basis of surveyed points and sketches of the surface. 140 points on the glacier surface, 30 points along the front and 40 points on the moraine ridges in front of the glacier were fixed by closed polygon from the triangulation points. The surveying equipment was a Wild T2 theodolite and electronic distance measuring device, Wild Distomate DI4. It was not possible to do any surveying on the glacier surface above 300 m a.s.l. because of the crevassing. The elevation contours in the upper part of the glacier in 1985 are therefore only approximate and dotted on the

Fig. 2. The front of Usherbreen after the surge in August 1985. Photo direction is south–north. Note the steep front and the folded moraine ridge systems.

glacier map (Fig. 4). However, the surveyed part covers the whole receiving area where the glacier surface has risen during the surge.

Both maps from 1971 and 1985 have been digitized at Statens Kartverk. A computer program was used to put a grid net over the maps, and the elevation (z) for each square was calculated by interpolating between the contour lines. The digitized data were used in a computer terrain model. In our display of this model close profiles are drawn with the vertical scale enlarged eight times to the horizontal, to give a three dimensional impression (Figs. 5 and 6). The terrain model for Usherbreen used a grid net with small squares of 20×20 m (400 m^2) in order to give a detailed smooth surface. The horizontal distance between each profile is then 20 m. To avoid a wrong impression of flat areas in the display, some extra contour lines and height points had to be inserted. The profiles are drawn perpendicular to the line of sight, but the view positions can be varied so that the terrain may be observed from any direction. The terrain-model formed the basis for the calculation of the ice mass transport during the surge.

The glacier front advanced 1.5 km during the surge and covered an area of 4.5 km^2. This included the whole area behind the terminal moraines that mark the outer limit of former advances. In the south the glacier advanced over the old ridges and on to ground formerly unaffected by this glacier. The lateral glacier river was pressed outwards and eroded a new course between the glacier and the valley wall. At some places this river eroded tunnels in the permafrozen alluvial debris at the level of the glacier sole. A glacier dammed lake was also formed (Fig. 7).

The glacier seems to have moved over the 1 km long and flat area behind the old moraine ridges without pushing any material up in front of it. As the glacier reached the old ridges, however, the whole ridge system began to move forward and at the same time new ridges developed on the sandur in front of the old system. The outermost limit of the ridges was moved up to 200 m. This displacement, together with the glacier advance out on fresh ground in the south, showed that this advance was greater than in any earlier surge of this glacier.

Due to crevasses after the surge there was little surface drainage which otherwise is common on

Fig. 3. Map of Usherbreen glacier constructed from airphotos from 1971. The scale is 1:75,000. Contour interval is 20 m. In the lower parts of the glacier 10 m contour lines are dotted.

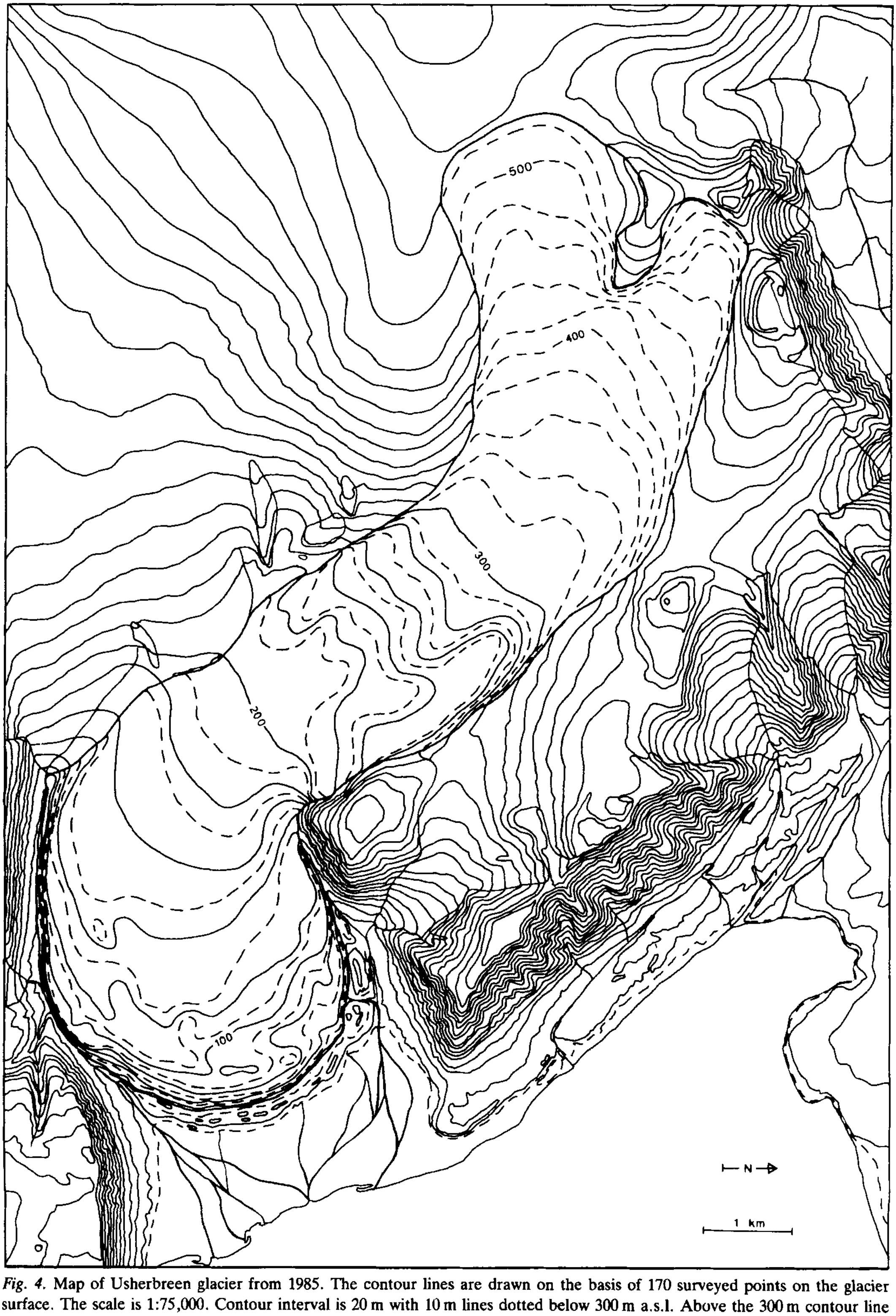


Fig. 4. Map of Usherbreen glacier from 1985. The contour lines are drawn on the basis of 170 surveyed points on the glacier surface. The scale is 1:75,000. Contour interval is 20 m with 10 m lines dotted below 300 m a.s.l. Above the 300 m contour line there are no surveyed points, and the lines are dotted as rough estimates.

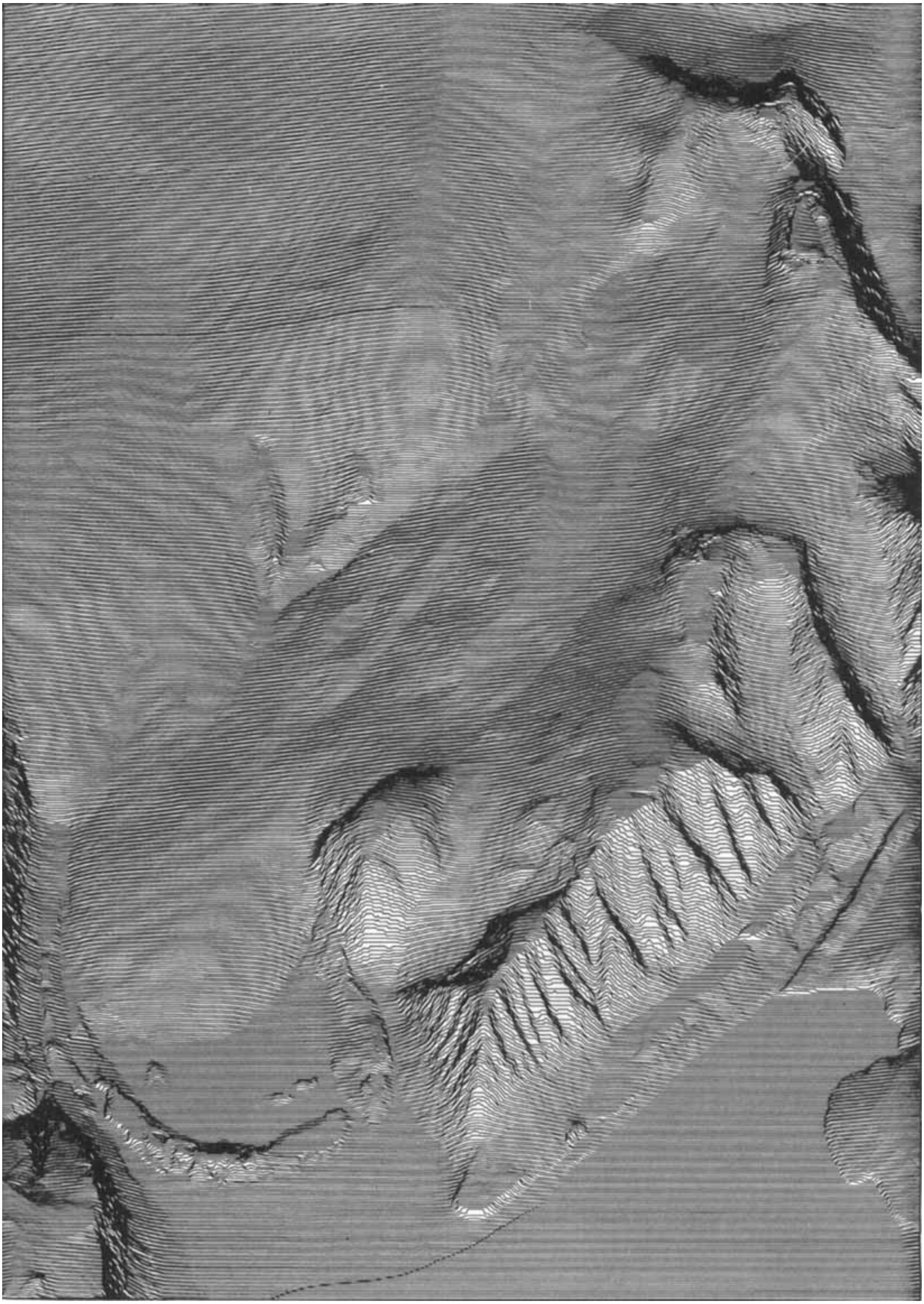


Fig. 5. Three-dimensional display of a terrain model of the glacier before the surge based on the map in Fig. 3 and in the same scale. The vertical scale is eight times the horizontal and the distance between the profiles is 20 m.

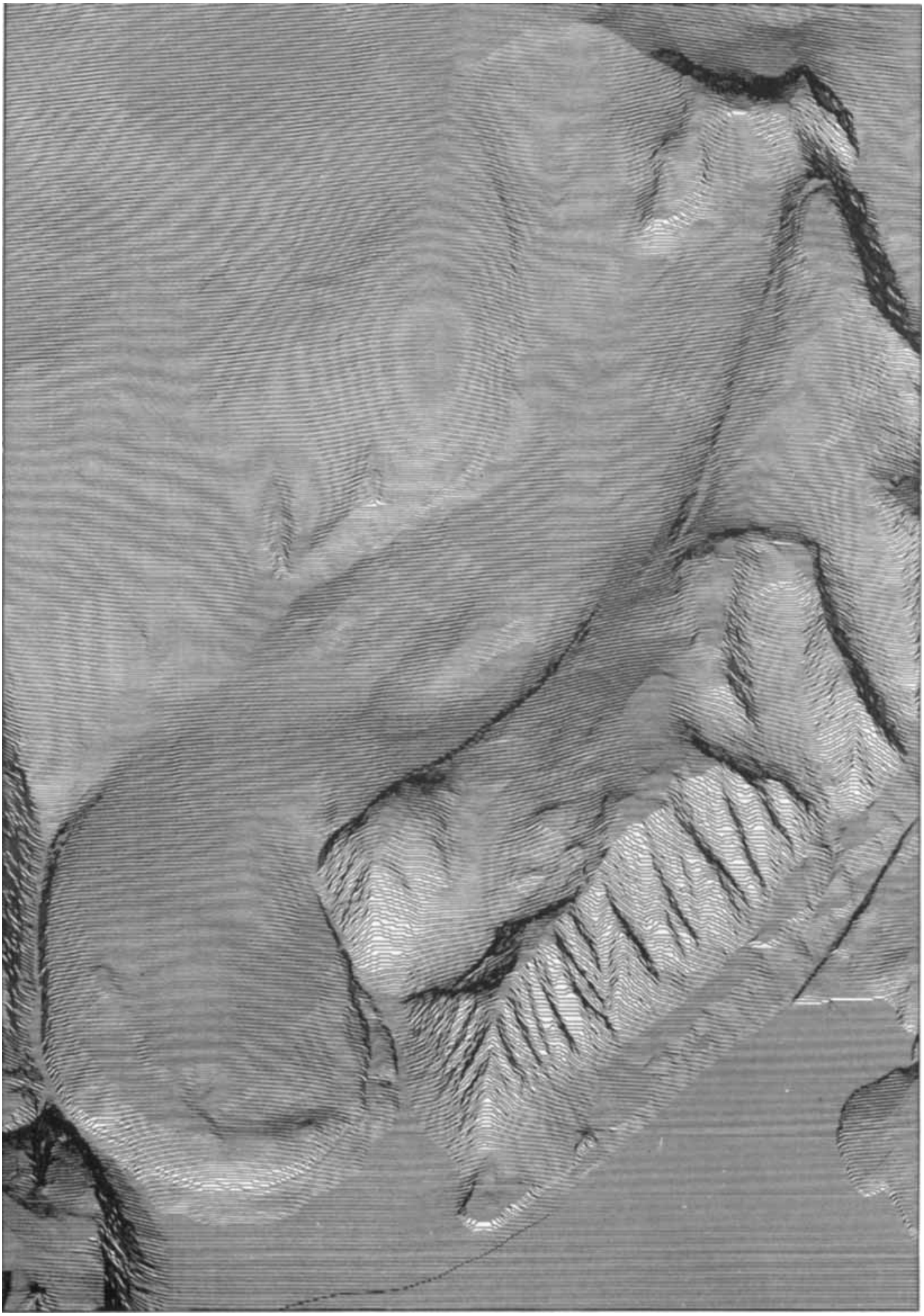


Fig. 6. Three-dimensional view of the model of the glacier after the surge based on the map in Fig. 4. The axes are as in Fig. 5.

(a)

(b)

(c)

Fig. 7. The glacier front was steep and advanced out on undisturbed ground (a). The lateral water stream eroded tunnels in the frozen alluvial material (b). A glacier dammed lake was also formed south of the glacier (c).

cold and subpolar glaciers. The water from the glacier was concentrated in two subglacial channels coming out of the southern and the northern side of the front. Only a few minor waterways were observed englacially or at the surface. The main stream in the north came out under the new advancing glacier, but over the old icecored moraine ridges, eroding a 20–30 m deep canyon in the old ridges (Fig. 8).

Longitudinal profiles and basal shear stress

The slope of the glacier surface changed considerably during the surge. Profiles before and after the surge were drawn as far up as the glacier surface was surveyed in 1985 (Fig. 9). The surface rose more than 100 m near the glacier margin during the surge. The front, which originally was gentle with an even slope, became steep and arching and 10–25 m high. A break to the steep wall (Fig. 2) appeared over a long distance along the outermost front. This must be due to the great ablation in the low-albedo dirt-covered surfaces that absorb a lot of energy from the global radiation. The dirt is melting from thin debris-layers in the basal ice, possibly regelation layers (Weertmann 1961).

Mean longitudinal gradient of the lower 6 km was 3.3 grad. before the surge. After the surge the gradient was 1.8 grad. measured from the top of the steep front. The mean basal shear stress can be found from $\tau = F\rho gh \sin\alpha$, where F is a shape factor dependent on the cross section. Here the width is about 2 km, and estimated mean thickness is about 150 m. This gives an approximate F-value of 0.9 (Nye 1965). The mean shear stress was then τ = 61 kPa (0.61 bar) at the gradient of 3.3 grad. just before the surge and τ = 33 kPa (0.33 bar) at the gradient of 1.8 grad. after the surge.

It is surprising that the critical shear stress value before the surge is as low as 0.6 bar. From registrations of change in the longitudinal profile of Finsterwalderbreen during a surge (Liestøl 1969) a mean basal shear stress of 170 kPa (1.7 bar) was calculated in the steepest parts of the glacier 3–5 km from the front in 1898 (Robin & Weertmann 1973). The surge started in 1910, and based on profile measurements in 1920 the mean shear stress after the surge was calculated to 60 kPa

Fig. 8. The glacier stream coming out in the north was forced into a new course and eroded a 20–30 m deep canyon into the old moraine ridges. These ridges nearest to the glacier contained old glacier ice.

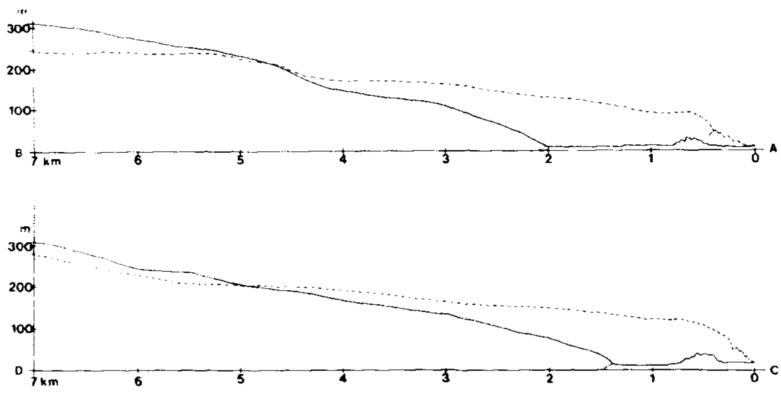


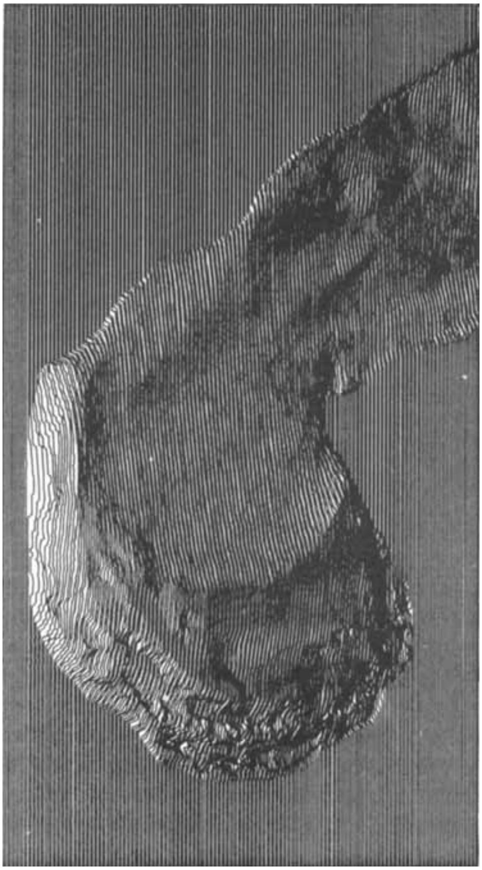
Fig. 9. Profile changes in the lower 7 km of the glacier during the surge. Dotted line is after the surge in 1985. Locations of the profiles are shown in Fig. 11.

(0.6 bar) in the lower 5 km. A common value for the basal shear stress in a temperate valley glacier is 150–200 kPa (1.5–2.0 bar).

The heat of friction (Q) is proportional to the sliding velocity (v): $Q = \tau A v$, where A is the area of the glacier sole. Increased heat of friction results in increased production of water that works as lubrication and keeps the high sliding velocity to maintain and lower the slope even more than to the steady state profile.

Ice transport

Transport of ice from the higher to the lower parts of the glacier was calculated from the increase in volume in the lower parts of the glacier. The change of volume was derived from the terrain model with the same grid net of 400 m^2 squares. The change of volume is calculated for each square (i): $\Delta V_i = \Delta h_i \times 400\ m^2$. In total the change becomes: $V = \Sigma \Delta V_i = \Sigma \Delta h_i \times 400\ m^2$.



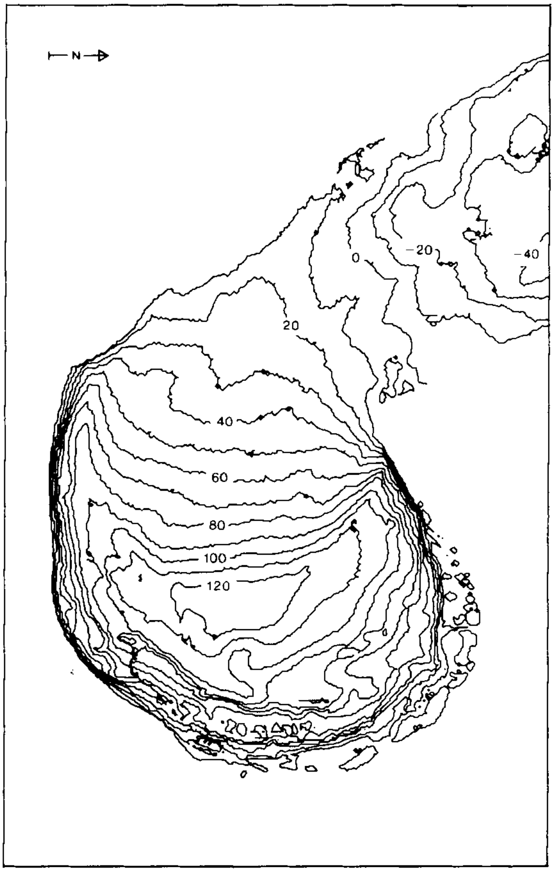


Fig. 10. Distribution of the change of ice volume in the lower parts of the glacier. Increased thickness is drawn as 10 m contour lines. To the left is a three-dimensional display. The glacier front before the surge is easily seen in the display. The horizontal scale is 1:65,000.

The result was that a total volume of ice transported down the glacier during the surge was $815 \times 10^6\,m^3 = 0.815\,km^3$. The total glacier area was 29 km^2 before the surge. Liestøl & Roland (pers. comm.) found an empirical formula for the area-thickness relationship from analysis of data from radio-echo soundings at Spitsbergen glaciers. They found that a mean thickness estimate can be found by $h = 33\,m \times \ln A + 25\,m$ in which A is the numerical quantity of the area given in km^2. Assuming this relation for Usherbreen gives an estimate of the mean thickness of 137 m and 4 km^3 in the volume. This may indicate that almost 20% of the ice mass was transported to the lower part of the glacier during the surge.

Annual accumulation on the glacier is not known. The glaciers near Ny-Ålesund in northwest Spitsbergen show a surplus in the corresponding height level in the accumulation area of about 500 kg/m^2 (Liestøl 1983). The snowfall is higher on the east coast, but if the value from Ny-Ålesund is used we can estimate the total yearly surplus in the accumulation area. The area is about 20 km^2 which gives an estimated yearly surplus of about $10 \times 10^6\,m^3$. The ice mass transport of 0.815 km^3 during the surge may correspond to 80 years of accumulation. There will be some transport of ice between each surge so the surplus will be less than $10 \times 10^6\,m^3$, and the surge transport of ice will then correspond to a somewhat longer period of accumulation than 80 years. Thus these considerations suggest a surge period of the order of 100 years.

Velocity measurements

The glacier front moved more than 1 km between 1978 and 1980. That is a yearly motion of more than 300 m, i.e. about 1 m per day. In its most active stage the motion was probably considerably larger than 1 m/d.

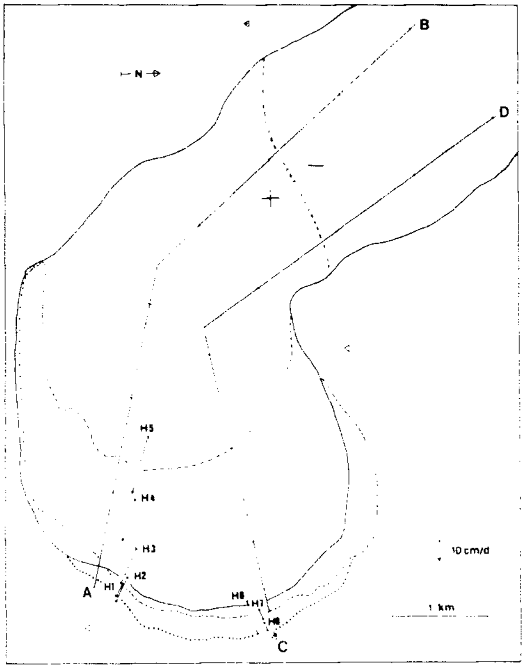


Fig. 11. The lower parts of the glacier with the points for velocity measurements and the profiles AB and CD.
~ glacier front in 1985
– – – glacier front in 1971
= = = dividing line between the area where the glacier surface has lowered and risen during the surge
·–·– front line of the old moraine ridge system before the surge
····· outer limit for new developed ridges
△ triangulation points

Table 1. Results of the velocity measurements in mm/d.

Profile 1: Stakes	H1	H2	H3	H4	H5
Distance from the front (m)	−30	70	370	820	1470
Period in 1985					
15.08–16.08	92	202	264	310	436
16.08–23.08	76	144	195	219	313
23.08–27.08	52	112	143	163	260

Profile 2: Stakes	H6	H7	H8
Distance from the front (m)	−230	−60	60
Period in 1985			
17.08–19.08	52	63	178
19.08–27.08	49	60	179

The velocity was measured in 1985 along two profiles from the moraine ridges in front of the glacier and 1.5 km up on the tongue. The registrations were done by theodolite and electronic distance meter from fixed triangulation points to reflectors mounted on stakes on the glacier. The Distomat DI4 has an accuracy of 5 mm + 5 mm per km. The stakes were surveyed in the period 15.08.85 to 27.08.85. The position of the stakes (H1–H8) and their motion are shown in Fig. 11. The points H1, H6 and H7 were situated on the old ridges in front of the glacier (− on distance in Table 1).

The measurements in 1985 showed that the velocity still was higher than it would have been if the glacier had been in the quiescent stage between two surges. The velocity increased from about 150 mm/d near the margin (H2 and H8) to about 300 mm/d 1.5 km up on the tongue (H5). The speed decreased rapidly from the beginning until the end of the survey. At point H5 the velocity decreased from 436 to 260 mm/d, at H4 from 310 to 160 mm/d, at H3 from 260 to 140 mm/d and at H2 from 200 to 110 mm/d. This may partly be because the glacier was stagnating at the end of the surge, but most probably the great decrease is due to weather conditions that caused the high water discharge and the high sliding velocity to decrease during the observation period. The first days were sunny and the air temperature was 8–10°C. Therefore the ablation and consequently the water discharge were high. The waterways mostly drained out subglacially. The two main streams both had a discharge of about 7–8 m^3/s. During the last four–five days the night temperature was below zero, the sky was cloudy and the temperature in the daytime was not more than 1–2°C. Accordingly, there was hardly any melting on the glacier, and the water discharge was reduced to less than half of the value at the beginning of the period. The sliding velocity therefore decreased rapidly too, and this rapid decrease indicates that the water was drained subglacially in great parts of the glacier. Subglacial drainage together with the high ice velocity showed that the glacier sole must have been at the pressure melting point.

The velocity observations confirmed that the ice in the old terminal ridges was reactivated. The glacier advance was hindered by the ridges, but pressing on them the whole ridge system was reactivated and pushed forward. The speed of the ridges just outside the glacier front was different from that of the glacier. The glacier moved two to three times as fast (see H1 against H2 and H7

Fig. 12. Fresh moraine ridge folding up on the sandur in front of the old moraine ridges.

against H8), and therefore the glacier gradually moved over the old ridges, at least in the periods when the ablation was less than the forward motion.

The motion of the ridges decreased outwards from the glacier front, but at H6, situated on the top of a ridge 230 m from the glacier, the velocity still was 50 mm/d. Motion could also be observed at small shear planes that were formed on the surface of the sandur outside the ridge system more than 300 m from the glacier. Parts of the ridge system were pushed 200 m forward during the surge (Fig. 11).

Only the old ridges nearest to the glacier contained icecores that were reactivated. The ridges more distant from the glacier did not contain any pure ice, but only folded sandur material. The motion causes a pressure that is transmitted through the frozen moraine ridges and on to the frozen sediment layers in the flat sandur area in front of the ridge system. These layers were folded up and new ridges were developed (Fig. 12). The thrust increased the foldings in the old ridges too, and it caused overriding and thrust faulting both in the old and the new ridges. The permafrost and the effect of the salt content on the mechanical properties of the material must be responsible for the deformation and the foldings of the material as the glacier pushed the ridges forward. This process will be described in more detail in a later paper.

Summary

During the surge at Usherbreen the glacier front advanced more than 1 km between 1978–1980 at an average speed of about 1 m/d. The velocity increased very abruptly when the surge was triggered, but the advance was retarding slowly over several years after the main period of the surge. In 1985 the front had advanced 1.5 km and covered an area of 4.5 km^2. Partly the glacier had moved over the old terminal moraine ridges and on to undisturbed ground. The glacier was still advancing in 1985 and was further out than in any earlier surge.

Mean slope of the glacier surface on the lower 7 km decreased from 3.3 grad. to 1.8 grad. This gives a critical triggering shear stress of only 60 kPa (0.6 bar) and a shear stress of 30 bar (0.3 bar) at the end of the surge.

The volume of ice transported from higher to lower parts of the glacier was 815 × 10^6 m^3, which is almost 20% of the total glacier volume.

The velocity measurements showed that the glacier was still advancing in August 1985, in the front 150 mm/d and 1.5 km up on the tongue about 300 mm/d.

The measurements on the ridges in front of the glacier confirmed that the ice on the ridges close to the glacier was reactivated so that the whole ridge system was pressed forward at a speed of about 50 mm/d. This resulted in increased folding and thrust faulting of the old ridges. Parts of the ridge system were pushed 200 m forward. In front of the old moraine ridges new ridges were developed as the frozen sediment layers in the flat sandur were folded up.

Acknowledgements. – These field studies were carried out in August 1985. The work was financed by a scholarship from the Norwegian Polar Research Institute while the author was employed at the University of Oslo, Dept. of Geography. I am grateful to Olav Liestøl who initiated this study and to Nils Flakstad who participated in the field work and parts of the home work. Atle Skarstein at Statens Kartverk gave valuable computer assistance. Helgi Bjørnsson and Olav Orheim commented on the manuscript.

References

Dolgushin, L. D. & Osipava, G. B. 1973: The regime of a surging glacier. *IAHS 107*, 1150–1159.

Flood, B., Nagy, J. & Winsnes, T. S. 1971: Spitsbergen. Southern part. Geological map 1: 500 000. *Norsk Polarinstitutt Skrifter 154A*.

Kamb, B., Raymond, C. F., Harrison, W. D., Engelhardt, H., Echelmeyer, K. A., Humphrey, N., Brugman, M. M. & Pfeffer, T. 1985: Glacier surge mechanism: 1982–1983 surge of Variegated Glacier, Alaska. *Science 227 (4686)*, 468–479.

Liestøl, O. 1969: Glacier surge in West Spitsbergen. *Canadian Journal of Earth Science 6*, 895–897.

Liestøl, O. 1976: Glaciological work in 1974. Hessbreen. *Norsk Polarinstitutt Årbok 1982*, 191–193.

Liestøl, O. 1983: Glaciological work in 1982. *Norsk Polarinstitutt Årbok 1982*, 37–43.

Liestøl, O. 1987: Glaciers of Svalbard, Norway. In: Satellite Image Atlas of Glaciers. *U.S. Geological Survey Professional Paper 1386* (in press).

Meier, M. F. & Post, A. 1969: What are glacier surges? *Canadian Journal of Earth Science 6 (4)*, 807–817.

Nye, J. F. 1965: The flow of a glacier in a channel of rectangular, elliptic or parabolic cross-section. *Journal of Glaciology 5(41)*, 661–690.

Pillewizer, W. 1939: Die Kartographischen und Gletscherkundlichen Ergebnisse der Deutschen Spitzbergen Expedition 1938. *Patermanns Geog. Mitt., Ergänzungsheft 238*.

Robin, G. de Q. & Weertmann, J. 1973: Cyclic surging of glaciers. *Journal of Glaciology, 12(64)*, 3–18.

Schytt, V. 1969: Glacier surge in eastern Svalbard. *Canadian Journal of Earth Science 6(4)*, 867–873.

Thorarinsson, S. 1969: Glacier surges in Iceland. *Canadian Journal of Earth Science 6(4)*, 875–882.

Weertmann, J. 1961: Mechanism for the formation of inner moraines found near the edge of cold ice caps and ice sheets. *Journal of Glaciology 3*, 965–978.