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Glacier Surge Mechanism: 1982–1983 Surge of Variegated Glacier, Alaska

Barclay Kamb, C. F. Raymond, W. D. Harrison
Hermann Engelhardt, K. A. Echelmeyer, N. Humphrey
M. M. Brugman, T. Pfeffer

Summary. The hundredfold speedup in glacier motion in a surge of the kind that took place in Variegated Glacier in 1982–1983 is caused by the buildup of high water pressure in the basal passageway system, which is made possible by a fundamental and pervasive change in the geometry and water-transport characteristics of this system. The behavior of the glacier in surge has many remarkable features, which can provide clues to a detailed theory of the surging process. The surge mechanism is akin to a proposed mechanism of overthrust faulting.

Glacier surging—popularly called "galloping"—is one of the most dramatic phenomena of glacier motion. A surging-type glacier, after flowing along in an apparently normal manner for years, speeds up for a relatively short time to flow rates as much as a hundred times the normal rate and then drops back to an apparently normal flow state. The process repeats itself with a period (typically in the range 10 to 100 years) that is thought to be roughly constant for any given glacier of surging type (*1, 2*). This phenomenon has intrigued and puzzled glaciologists since it was brought to scientific attention by the 1906 surge of Variegated Glacier, Alaska (*3*). Evidence has accumulated that a good many glaciers are of surging type, even though relatively few have actually been observed in the act of surging (*4*). A number of theoretical explanations of surging have been proposed (*1, 5*), but progress toward a firm understanding of the phenomenon has been slow because of a lack of sufficient field observations providing clear indications of the causative factors (*6*). The surge phenomenon is recognized as one of the outstanding unsolved problems of glacier mechanics. It is also of wider interest, because of the possibilities that glacier surges may impinge upon works of man and that surging of the Antarctic ice sheet may be a factor in the initiation of ice ages and in cyclic variations of sea level (*7*). Glacier surging also bears a relation to overthrust faulting and gravity tectonics in the earth and to landsliding.

We have made detailed observations of the surge of Variegated Glacier that took place in 1982–1983. Our observations reveal many new facts about the surge process and provide the basis for some informed reasoning about the physical mechanism of surging. We present here a summary of the principal results at an early stage in their evaluation.

Background and Presurge Buildup

Surges of Variegated Glacier are known (or believed on the basis of limited evidence) to have occurred in 1906, the late 1920's or early 1930's, about 1947, and 1964–1965 (*4*). The recurrence period is thus about 17 to 20 years, and a surge in about 1981 to 1985 was expected.

In 1973 a continuing program was initiated to monitor the glacier in its "normal" state and to follow the buildup to the anticipated surge (*8*). The ice mass was found to be already at the melting point throughout, ruling out the idea that surging occurs when cold basal ice warms up to the melting point (*5*). Figure 1 shows a map of the glacier and indicates our terminology for its component parts and for the longitudinal coordinate (expressed in "Km") by which we designate the positions of points along the length of the glacier. Measurements revealed that from year to year the ice was thickening in the middle and upper part of the glacier (upstream from Km 12) and thinning in the lower part (downstream from Km 12). Above Km 15 there was a progressive increase in ice flow velocity: in the middle glacier (~Km 11) it approximately doubled between 1973 and 1981, and in the upper glacier (~Km 5) it approximately quadrupled. Summertime flow velocities in 1981 were 0.4 to 1.0 m/day in the upper glacier (Km 3 to 10) and 0.1 to 0.2 m/day in the lower glacier (Km 12 to 15). The terminal lobe, below Km 16, was essentially stagnant.

Until 1978 the flow increase was more marked in summer than in winter. This indicated that the buildup was having a marked effect on basal sliding, which is thought to be the cause of the summertime speedup of "normal" glaciers (*9*). Starting in 1978, wintertime flow velocities began to show an anomalous increase, small but definite, which suggested that increased basal sliding was now occurring in winter as well as in summer.

In 1978 we became aware that short

B. Kamb is a professor in the Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena 91125. C. F. Raymond is a professor in the Geophysics Program, University of Washington, Seattle 98195. W. D. Harrison is a professor in the Geophysical Institute, University of Alaska, Fairbanks 99701. H. Engelhardt is senior research fellow in the Institut für Geophysik, Westfälische Wilhelms-Universität, Münster, Federal Republic of Germany. K. A. Echelmeyer is a research fellow and M. M. Brugman is a graduate student at California Institute of Technology. N. Humphrey and T. Pfeffer are graduate students at the University of Washington.

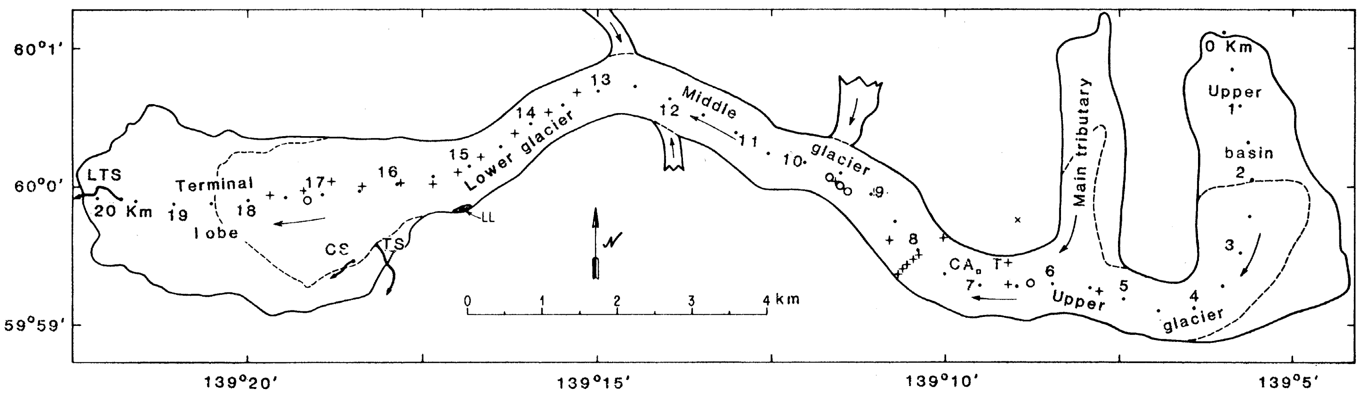
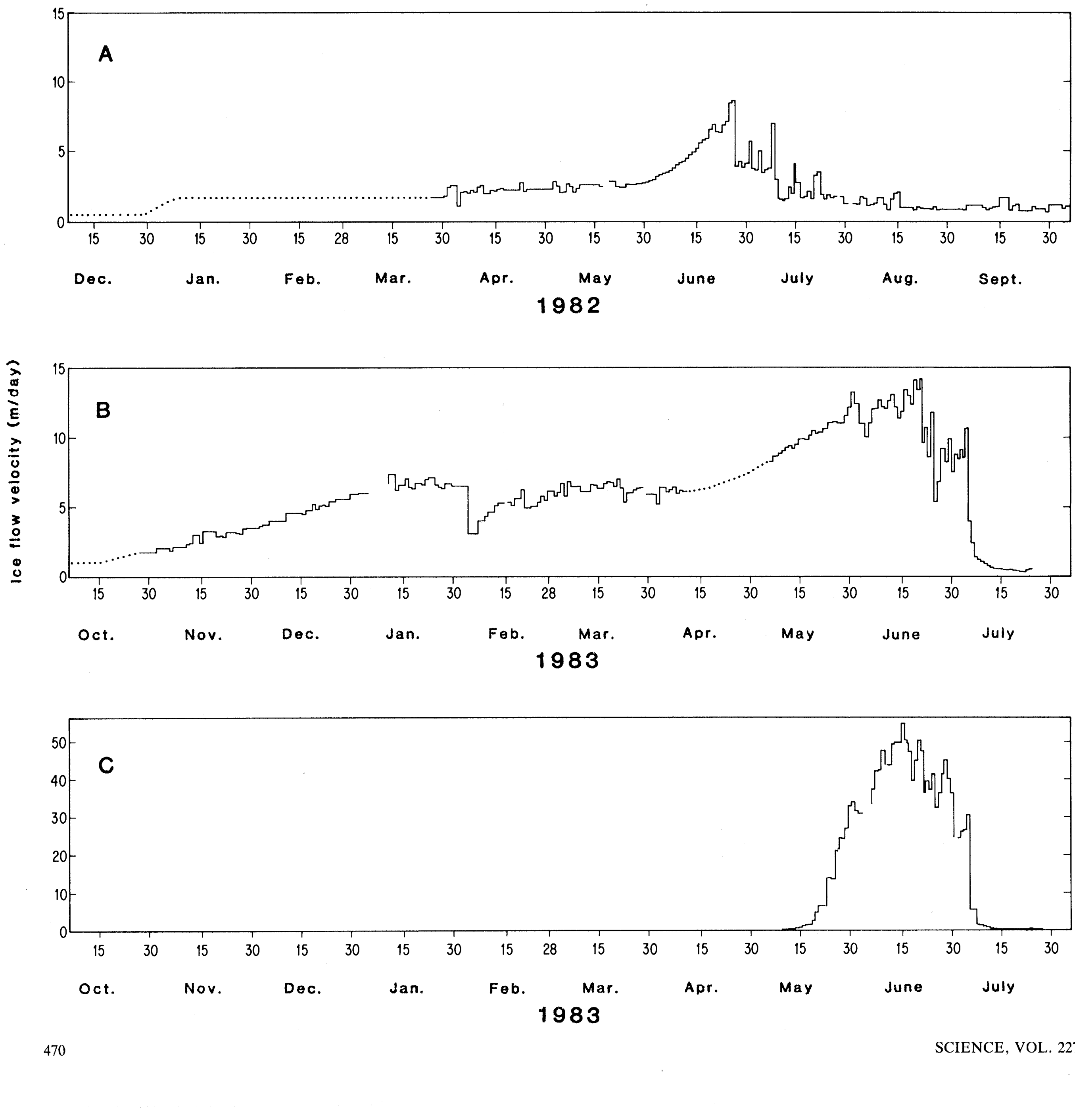


Fig. 1. Map of Variegated Glacier, Alaska, showing the longitudinal coordinate scale (0 to 20 Km) by which locations on the glacier are referenced in the text. Light arrows indicate the general direction of glacier flow. Heavy arrows show outflow streams near the terminus. The dashed line shows the boundaries of the part of the glacier that surged in 1983. Crosses indicate approximate positions of reference markers for which ice-flow velocity data are given in the text. Open circles are approximate positions of boreholes drilled to the glacier bed. The point labeled CA is the location of a crevasse whose water content was observed during February through May 1983. The location of a seismometer on bedrock is shown with an × north of Km 6.5; LL designates the location of the lake shown in the cover photo and discussed in the text; LTS signifies the lower terminus stream; CS and TS are other outflow streams.



spurts of accelerated flow velocity were occurring in Variegated Glacier during the summertime. These events, which we call "minisurges," were documented in 1979 to 1981 (*10*). In an event of this kind, the flow velocity increases rapidly from ~0.4 m/day to 1 to 3 m/day, in an hour or two and then declines to ~0.4 m/day in 10 to 20 hours. Four or five minisurges occurred in the early part of each summer of 1979 through 1981, in June and early July. The flow-velocity peak of each minisurge propagated as a wave down the glacier over the reach from near Km 3 to about Km 9, at a propagation speed of about 400 m/hour.

The velocity increase in a minisurge is accompanied by a simultaneous abrupt rise in the level of water standing in boreholes that reach the glacier bed. In such boreholes, once a hydraulic connection with the basal water system of the glacier has been established, the water level reveals the pressure in the basal water system (*11*). (By basal water system we mean the system of passageways by which water entering the glacier from the surface or sides and finding its way to the bottom is conducted down along the glacier bed and delivered as outflow at the terminus.) In minisurges the peak basal water pressure occurs synchronously with the peak flow velocity and is high enough to "float" the glacier (water level closer than 40 m to the surface in ice 400 m thick). It seems clear that the pulses of high flow velocity are caused by the peaks of high water pressure.

The Surge

The surge began in January 1982. The onset was detected by a marked increase in seismic icequake activity picked up by a seismometer located on bedrock 0.5 km north of the upper glacier (location shown in Fig. 1). (A strong correlation between seismic activity and glacier motion had been established in the minisurges and was later reinforced by observations during the surge.) Flow-velocity measurements beginning on 26 March showed that the upper glacier was moving at surging speed, about 2 m/day. The course of the surge motion is shown in Fig. 2. From late May to late June 1982 the velocity in the upper glacier (Fig. 2A) rose gradually from 2.6 to 9.2 m/day. On the morning of 26 June the velocity peaked at 10.4 m/day and then abruptly dropped to less than half its value in a few hours. The motion thereafter was pulsatory, with five major pulses of movement (peak velocities of 3 to 7 m/day) superimposed upon a decreasing trend. By mid-August the velocity had declined to about 1 m/day. It remained low through August and September, and the surging seemed to have stopped, ending phase 1 of the surge.

The surge motions in phase 1 took place only in the upper glacier. The flow velocity was high and approximately constant from Km 4 to Km 7 and decreased both up- and downstream from this reach. At Km 9.5 the velocity was only about half that at Km 7, and the lower glacier, below about Km 12, was practically unaffected. The surge motion extended headward to within about 2 km

Fig. 2. Glacier flow velocity versus time in the 1982–1983 surge of Variegated Glacier: (A and B) the velocity in the upper glacier in 1982 and 1982–1983; (C) the velocity in the lower glacier in 1982–1983. Velocity values are daily averages where available, otherwise longer term averages. The averages are shown as horizontal lines over the time intervals over which they were measured. Dotted lines show conjectured or interpolated velocities over intervals where short-term measurements are lacking. The data in (C) were measured on markers near Km 15; (A) and (B) are a compilation of data measured on markers over the interval Km 2.8 to 8.3.

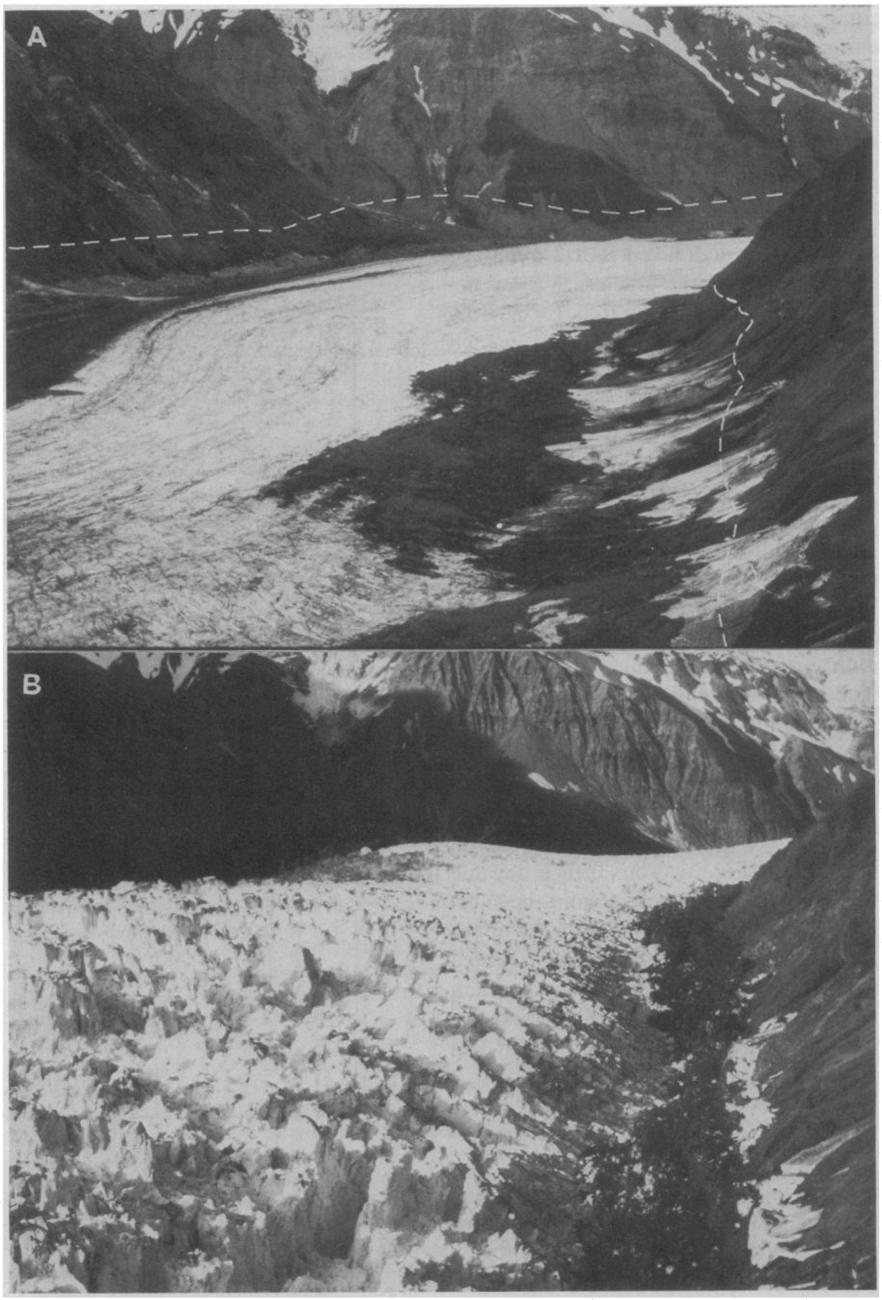


Fig. 3. Views of Variegated Glacier near Km 13 taken from the same point before the surge, July 1982 (A), and during the surge, 4 July 1983 (B). The view is upglacier. The scale, which is identical in (A) and (B), can be judged from the fact that the width of the glacier is approximately 1 km. In (A) the position of the ice surface in surge, as seen in (B), is marked with dashed lines.

of the head of the glacier and of its main tributary, as shown by the dashed lines in Fig. 1.

Phase 2 of the surge began in October 1982 with a gradual increase in seismicity and flow velocity in the upper glacier (Fig. 2B). This continued until early January 1983, when the velocity reached a stable level of 5 to 7 m/day, about three times the wintertime level in phase 1. An abrupt major drop on 3 February was followed by a gradual recovery to the stable level. In April the velocity began a further gradual but substantial increase, reminiscent of the phase 1 increase in May to June 1982, and reaching a broad maximum of about 15 m/day in mid-June, upon which large, complex fluctuations were superimposed, beginning about 1 June. In the afternoon and evening of 4 July the surge entered into an abrupt termination. The velocity dropped in a few hours to about a quarter of its previous value. There was a slight recovery on 6 July, but by 8 July the velocity had decreased to 1.2 m/day or less throughout the glacier. A slow decrease continued thereafter. By 26 July the velocity in the upper glacier was no more than about 0.2 m/day, lower than presurge velocities.

In phase 2 the surge propagated progressively down into the lower glacier. Flow velocities there, which had been very low (0.2 m/day or less) in the presurge condition, rose dramatically as the surge front arrived and reached 40 to 60 m/day during June (Fig. 2C). The highest velocity measured was 65 m/day, on June 9, for about 2 hours, at Km 14.5. The surge termination affected the lower glacier similarly to the upper. By 27 July the velocity was down to about 0.4 m/day and by September to about 0.2 m/day.

The surge produced great changes in the glacier (Fig. 3). The surface became intensely crevassed throughout the surging part (delimited in Fig. 1). In the upper glacier, above Km 8, the ice was thinned from its presurge condition by as much as 50 m. Below Km 8, thickening occurred, up to a maximum of 100 m at Km 16.

The Surge Front

The enhanced downstream transport of ice in phase 1 of the surge produced a general thickening of about 30 m in the lower half of the reach affected by this phase, from about Km 7 to 12.5. In phase 2 this "bulge" began to propagate as a wave downglacier, and, as the thickening increased further, the forward side of the wave steepened and sharpened to form an impressive surge front sweeping ponderously down the glacier. The propagating surge front is depicted in Fig. 4, where it is seen as a downstream-facing topographic ramp about 1 to 2 km long, behind which the ice has thickened by about 100 m from its presurge thickness (which can be judged from the dashed line in Fig. 4B). Within the front there is a high longitudinal gradient of the flow velocity (Fig. 4A), corresponding to extremely large compressive strain rates of up to 0.2 per day. (Compare these with typical rates of ≤ 0.1 per year in nonsurging glaciers.) Ahead of the surge front the flow velocity is about nil, while behind it the ice is moving at full surge speed and the longitudinal velocity gradient is small and in fact becomes locally extensile as a velocity maximum develops around the topographic crest of the surge-front wave from 23 May onward.

The leading edge of the surge front had reached Km 13.5 by 15 February 1982, and by 15 May it had propagated down

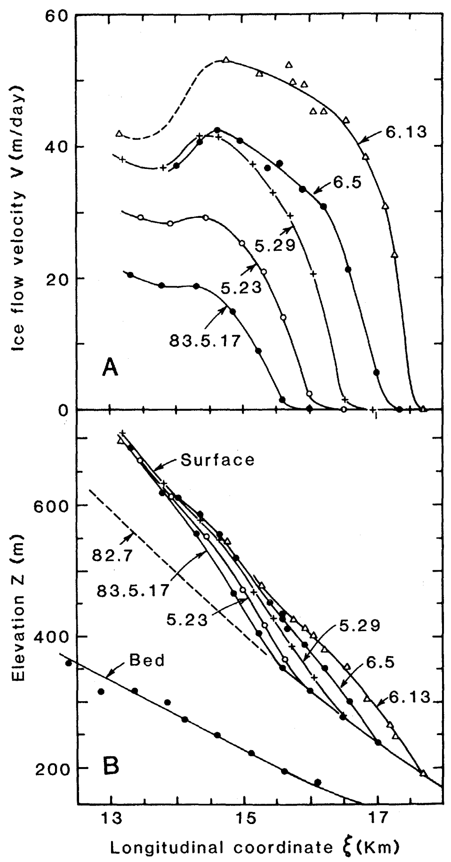


Fig. 4. Forward propagation of the surge front, shown by longitudinal profiles of ice flow velocity (A) and ice surface elevation (B) as a function of the longitudinal coordinate ξ in Km (see curvilinear coordinate scale in Fig. 1). The profiles are shown at five successive times in May and June 1983, at intervals of 6 to 8 days. The position of the ice surface in August 1982 is shown in (B) with a dashed line, and the approximate position of the glacier bed as determined by reflection seismology (*8*) is also shown.

through the lower glacier to Km 15.6, at an average propagation speed of 23 m/day. The forward propagation over the period 17 May to 13 June is documented in Fig. 4. During this time the sharply defined leading edge of the front advanced at a nearly constant speed of 80 m/day. This speed was maintained, with only a slight decrease, up to the surge termination on 4 to 5 July.

The propagation speed of the surge front can be explained on the basis of conservation of ice volume (continuity). If v_2 is the flow velocity at the trailing edge of the front, h_2 is the ice thickness there, and h_1 the thickness at the leading edge (where $v_1 \cong 0$), then continuity requires the front to advance at a speed V given by $V = v_2h_2/(h_2 - h_1)$. For the values of h_1, h_2, and v_2 seen in Fig. 4, this equation gives V values in the range 50 to 100 m/day, in rough agreement with the observed speed of advance. This agreement supports the conclusion that the advance of the surge front is a phenomenon of kinematic-wave type, although different from the kinematic waves previously proposed for glaciers (*12*). The details of the surge mechanism govern the propagation speed V via their effect on h_2 and v_2.

Detailed Flow Features of the Surge

Large pulsations or oscillations in the surge velocity occurred during July 1982 and June 1983; the latter are shown in detail in Fig. 5. Three distinct types of fluctuations are recognizable. (i) Quasi-regular oscillations with a period of about 2 days are very pronounced in the upper glacier (Fig. 5, A and B). From 19 June to 6 July they have a very regular period of about 40 hours; prior to 19 June the period is longer and more irregular. (ii) Oscillations of shorter period, generally somewhat shorter than 1 day, are pronounced in the lower glacier (Fig. 5C). (iii) Approximately every 4 or 5 days there occur major slowdown events, in which the velocity, after first rising to a peak, drops rapidly, in a few hours, to a level distinctly lower than was reached in the previous several days. The surge termination on 4 to 5 July 1983 was an event of this type, as was the major slowdown on 26 June 1982 and the slowdowns that ended the subsequent movement pulses in phase 1. The major slowdown on 3 February 1983 is also similar except that it was not preceded by a peak in motion. Major slowdowns in June 1983 are marked with downward-pointing arrows in Fig. 5. Between slowdown events, there is a tendency toward progressive increase in velocity, superimposed on the oscillations.

There is a very clear correlation of individual velocity peaks, troughs, and slowdowns as observed at different points along the length of the upper glacier (Fig. 5, A and B). The correlation is somewhat confused as we move to the lower glacier (Fig. 5C) because of the increased number of peaks and troughs there, but in many cases a correlation can nevertheless be identified; in particular, the major slowdown events clearly affect the entire surging length of the glacier. Close inspection of Fig. 5 reveals definite shifts in the time of individual peaks, troughs, and slowdowns at different points. In general, these features arrive later at points farther downglacier, indicating downglacier propagation of the velocity fluctuations. From 17 June onward the propagation speed is approximately 600 m/hour. Prior to 17 June the propagation speed is generally faster, and in many cases the fluctuations appear to be almost synchronous throughout the surging part of the glacier. The major slowdown on 3 February 1983, and also minor slowdown events on 21 February and 22 and 31 March, showed downglacier propagation effects as detected in records of strain, tilt, and seismicity that accompanied these events; the propagation velocity for the 21 February event was 500 m/hour.

On 23 July 1983, well after the surge termination, a minisurge occurred in which the flow velocity quickly quadrupled and then dropped back to ~0.4 m/day in ~10 hours. It was detected strongly at Km 9.5 and, 7 hours later, at Km 14.5. The time delay corresponds to an average downglacier propagation speed of 700 m/hour. This is at the upper end of the range of minisurge propagation speeds (250 to 700 m/hour) observed in 1979 through 1981 (*10*).

Basal Sliding in the Surge

In order to determine how much of the surge motion is due to internal deformation within the ice mass and how much to basal sliding, on 8 June 1983 a borehole was drilled 385 m to the bed (*13*) at Km 9.5. The spatial configuration of the hole was measured on 10 June and again on 12 June by means of borehole inclinometry (Fig. 6). Over the 2-day interval the ice moved forward 27.3 m at the surface, while the bottom of the borehole lagged behind by slightly less than 1 m as a result of internal deformation in the ice mass. Thus about 95 percent of the motion is due to basal sliding and about 5 percent to internal deformation, the latter being at least roughly (within a factor of 2) the amount of internal motion that would be expected to take place in the absence of the surging. This result demonstrates that the surge motion is caused by extreme enhancement of basal sliding—a conclusion often previously surmised but never before demonstrated by actual observation.

Indirect confirmatory evidence for the large enhancement of basal sliding in the surge is given by comparison of transverse profiles of flow velocity as observed under nonsurging and surging conditions (Fig. 7). The change from a quasi-parabolic transverse profile of velocity across the width of the glacier under nonsurging conditions (Fig. 7, A and D) to a flat profile, representing "plug flow" (*14*), under surge (Fig. 7, B and C) is interpretable in terms of a great increase in basal and marginal sliding in the surge.

Observations of closely spaced markers near the south margin of the profile in Fig. 7C reveal that there is a large velocity discontinuity within the ice mass near the edge. This discontinuity, marked by shear arrows in Fig. 7C, occurs across a

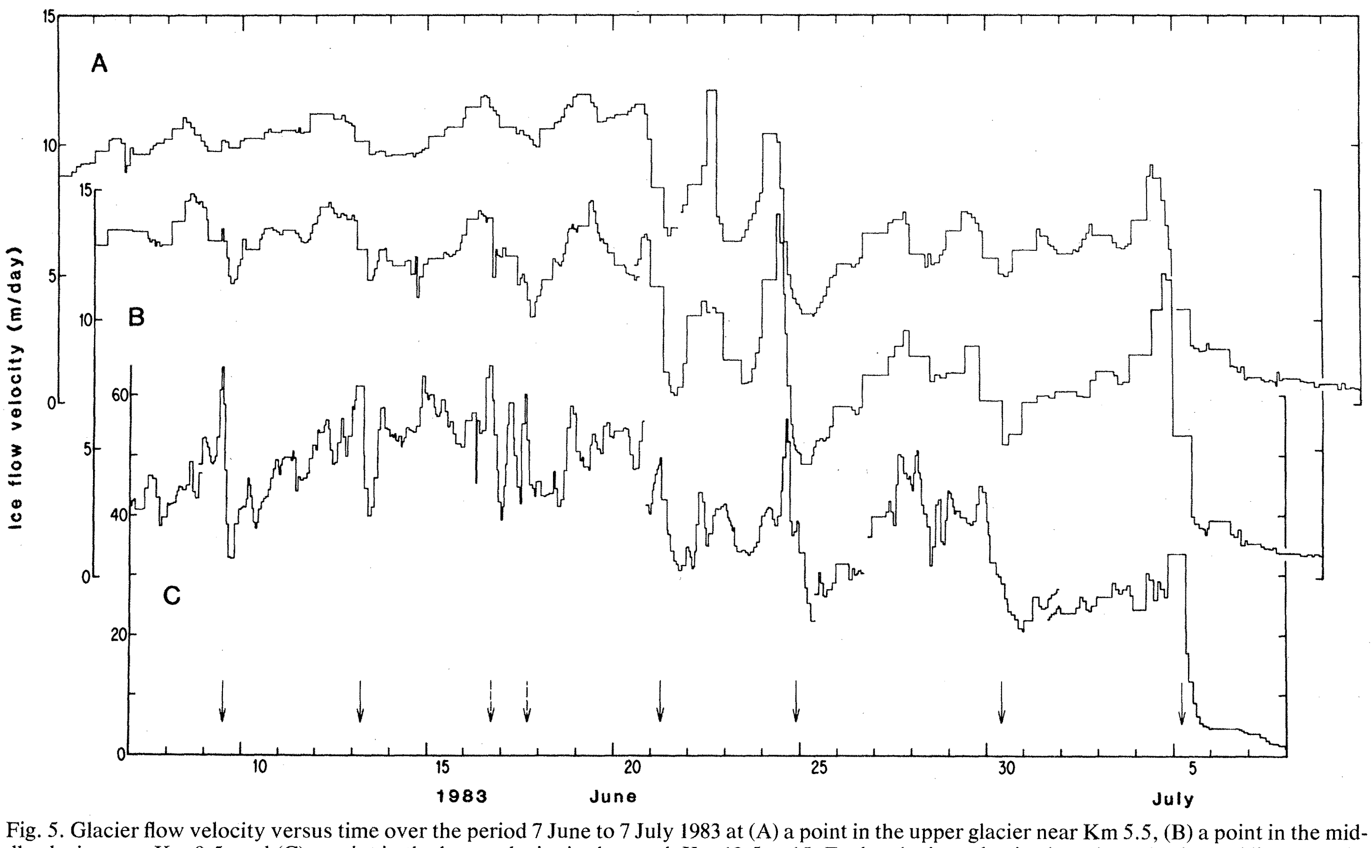


Fig. 5. Glacier flow velocity versus time over the period 7 June to 7 July 1983 at (A) a point in the upper glacier near Km 5.5, (B) a point in the middle glacier near Km 9.5, and (C) a point in the lower glacier in the reach Km 13.5 to 15. Each velocity value is plotted as a horizontal line over the time interval over which it was measured. (The actual velocity as a function of time is doubtless a smooth curve through these data lines.) The graphs have a common time abscissa; the ordinate scales are offset as indicated. The downward-pointing arrows mark major slowdowns in flow velocity, which occurred throughout the surging part of the glacier; dashed arrows are shown for the slowdown on 16 to 17 June because of the complexity of this event. Discontinuities in the data occur where there is a switch between different markers whose motion is being followed.

deep, sharp cleft in the glacier surface, parallel to the margin and about 50 m in from it (Fig. 8). The cleft is the surface expression of a discrete shear surface or wrench fault within the ice and is analogous to such shears sometimes seen in sea-ice packs. A cleft of this kind, usually accompanied by heavy blackening of the ice surface with rock debris, is seen at many places near the margins of the glacier in surge. No such feature was present before the surge. The fault probably extends down vertically to an intersection with the bed, where the internal velocity discontinuity across the shear surface is transferred to the basal velocity discontinuity that is the expression of rapid basal sliding.

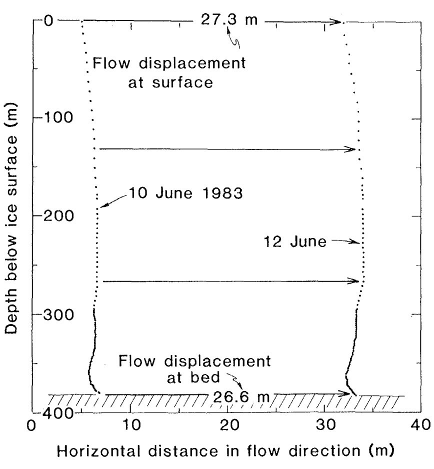
Role of Basal Water Pressure in the Surge

The record of water level as a function of time in boreholes drilled to the glacier bed near Km 9.5 (Fig. 9) shows that the pattern of fluctuating levels underwent a distinct change at the time of surge termination on 4 to 5 July 1983. Before the termination, the fluctuations were relatively modest and the lowest levels reached were at a depth of 80 to 90 m, whereas afterward the fluctuations became much larger and the lowest levels reached dropped to 120 to 195 m. In phase 1 of the surge, continuously high water levels (35 to 80 m) were recorded during June and July 1982 in boreholes at Km 6.3, whereas in prior years the borehole water levels had ranged over 50 to 210 m except for the high peaks accompanying minisurges. As surge phase 1 was ending, during July 1982, water levels in boreholes at Km 9.5 dropped progressively from 40 m to more than 150 m. Superimposed on this declining trend were several high peaks, which coincided with the surge pulses that occurred during this period. Similarly, there is a strong correlation between water-level peaks and flow-velocity peaks in phase 2 (Figs. 9 and 5B).

It thus appears that the glacier under surge is characterized by consistently high basal water pressures, with peaks in pressure that coincide with pulses in surge movement. The pressure levels are best stated relative to the ice overburden pressure, which at the borehole site was about 36 bars, corresponding to an ice thickness of about 400 m; a borehole water level at a depth of 40 m corresponds to a basal water pressure equal to the ice overburden. Relative to this, the basal water pressures during surge are, at a minimum, within 4 to 5 bars of the ice overburden pressure, and there are transient fluctuations often up to within 1.5 bars of overburden and infrequently to as much as 0.5 bar above overburden (as on 27 June and 4 July 1983, see Fig. 9). After surge and also prior to surge (as observed in 1978 through 1980), the pressures are as a whole lower, particularly the minimum pressures, which are 8 bars to as much as 16 bars below overburden, but there are also transient fluctuations to high levels. The highest of these, in which the basal water reaches or exceeds the overburden pressure, coincide with the occurrence of minisurges, as in the minisurge that occurred on 23 July 1983 after surge termination (see Fig. 9).

Further evidence for the role of water pressure in the surge was the formation of a lake along the southwest margin of the glacier at Km 15.2 in early June 1983 (see cover photo and Fig. 1). The lake contained highly turbid water, typical of basal water charged with comminuted rock debris or "glacial flour," and it filled from water that welled up in a large crevasse 100 m in from the margin and then flowed outward to the margin. Thus the main source of the water was the basal water system, not marginal streams. The highest lake level was about 75 m below the laterally averaged ice surface in the cross profile of the glacier through the lake. The lake drained away as the surge terminated. It reappeared for a few hours on the morning of 24 July, as the minisurge was in progress.

There is some evidence that the basal water pressure was high in midwinter 1983. We observed "steam" (condensing water vapor) rising from crevasses near Km 7 on 9 January, suggesting the presence of water in the crevasses at shallow depth. On 17 February we succeeded in lowering a probe to a depth of 42 m in a crevasse at point CA in Fig. 1 (Km 7.5) and brought up samples of dirty water from depths greater than 40 m. Since there is no surface source for water in midwinter and since surface-derived water is generally clean, it seems probable that the water sampled came from the bed and is evidence for the occurrence of high basal water pressure at or before the time of sampling. A recording pressure transducer left in the crevasse at point CA indicated that the water level in the crevasse rose to a depth of about 38 m in March and April.



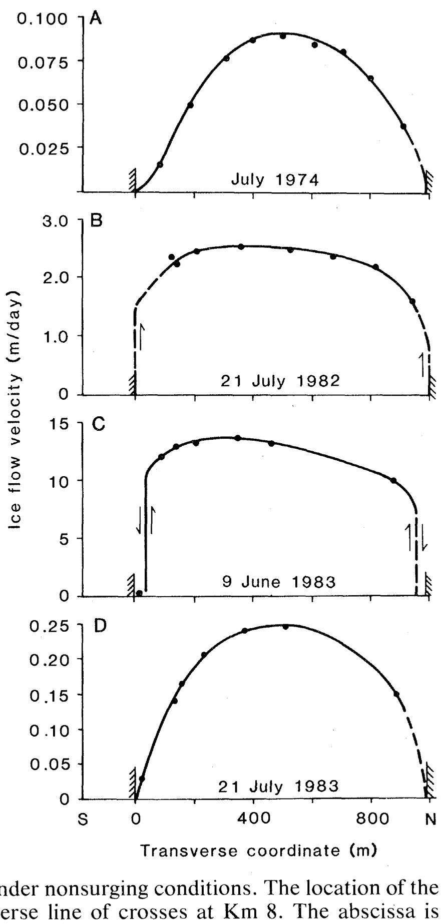


Fig. 6 (left). Ice motion at depth in the glacier, measured from the displacement of a borehole (82-E) drilled to the glacier bed at Km 9.5. The location and shape of the borehole on 10 June and 12 June 1983 are shown, the individual dots representing the individual borehole-inclinometry measurements. The abscissa is horizontal distance in the flow direction (N64°W), with an arbitrary zero. The borehole is plotted in projection onto a vertical plane in this direction. Arrows represent the horizontal component of the forward motion of the glacier over the 2-day period. The displacement at the surface was measured by surveying to fixed bedrock points. Fig. 7 (right). Transverse profiles of ice flow velocity at Km 8 measured at four different times, two (B and C) during surge and two (A and D) under nonsurging conditions. The location of the profile line is indicated in Fig. 1 by the transverse line of crosses at Km 8. The abscissa is distance measured in the direction N50°E from the southwest margin of the glacier. The arrows represent marginal shear zones. In (C) these occur as wrench faults in the glacier about 50 m in from the margins. The profile in (A) is from (8).

The foregoing observations, particularly when considered collectively, demonstrate a well-defined influence of basal water pressure on the motion of the glacier and, taken together with theoretical considerations (see below), make a strong case for the conclusion that high basal water pressure is the direct cause of the high flow velocities in the surge.

Discharge from the Basal Water System

During the surge, floods in the outflow streams at the terminus correlated closely with rapid slowdowns of flow velocity in the surging part of the glacier. This was first observed in the major slowdown on 26 June 1982, which was accompanied by a particularly large flood. The correlation was observed repeatedly in the subsequent surge pulses in phase 1 and throughout phase 2. Figure 10 is a comparison between the flow velocity at Km 16.5 and the discharge of the lower terminus stream (LTS) over the period 17 June to 6 July 1983. Flood peaks occurred on 17, 21, 25, and 30 June and 5 July, in coincidence with major velocity slowdowns. Although the minor slowdown on 28 June (Fig. 10) was not associated with a flood in the LTS stream, there was a flood in the CS stream (Fig. 1) at this time, shown by the dashed line in Fig. 10B. There is a good correlation also between smaller floods and velocity slowdowns earlier in the season, from February to mid-June. In particular, the slowdown on 3 February 1983 was accompanied by a marked flood in the otherwise small and snow-covered TS stream, observed by an automatic camera.

The termination of the surge was accompanied by a particularly spectacular flood. Not only was the flow of the principal outflow streams greatly increased, but, in addition, highly turbid water gushed and spurted from numerous cracks, crevasses, and holes along the leading edge of the surge front, particularly along its south side (dashed line south of Km 17 in Fig. 1). In an ice cliff above the TS outlet (Fig. 1), water spurted out over a vertical range of 20 to 30 m above the stream level.

The flood peaks that correlate with the surge slowdowns and termination in 1982 and 1983 are a manifestation of rapid release of stored water from the glacier; only indirectly do they reflect water input to the glacier from rain or the melting of ice and snow. The water input is variable in accordance with changes in the weather, but the weather-related input did not have high peaks corresponding to the flood peaks in Fig. 10B. There is also a diurnal cycle of water input, which is not noticeably reflected in the outflow discharge.

There does, however, appear to be a relation between weather-related water input and surge motion, although somewhat indirect and complex. A gradual increase in low-level water input as wintertime conditions give way to spring seems to be reflected in the progressive increase in surge motion in May through June of 1982 and 1983 (Fig. 2). Larger water input, however, results in surge pulses followed by slowdowns and ultimate termination, as occurred in June through July of 1982 and 1983 (Figs. 2 and 5). A particularly suggestive instance of this is the major slowdown on 26 June 1982, which occurred after the first 3 days of warm weather and rapid melting of the season. On the other hand, the slowdown on 3 February 1983, and the surge buildup in January 1982 and in November and December 1983, are phenomena outside the above relationship, because they are correlated not with water input but rather with the lack of it.

The surge produced substantial changes in the outflow-stream pattern of the glacier. Prior to the surge, most of the outflow was in stream TS (Fig. 1). As

Fig. 8. Aerial view of marginal wrench fault on the southwest side of the glacier at Km 8.5, looking southeast. The fault is the vertical black line in the lower center. It is a deep cleft blackened by rock debris. Ice to the left is moving toward the observer at ~10 m/day while ice to the right is practically stationary.

the surge front moved past this outlet, its discharge was greatly reduced and the outflow appeared first in the newly formed outlet, CS (Fig. 1) and later, as CS became shut down, in LTS.

During the surge, the turbidity of the outflow-stream water was much higher than before or after the surge, reflecting a much higher content of glacial flour.

Subglacial Water Flow

The flow of water in the basal water system was studied by water-tracer experiments initiated on 9 June 1983, during surge, and on 14 July 1983, after surge termination. The water transit time between a borehole at Km 9.5 and the terminus outflow streams was observed by means of the standard water tracer dye rhodamine WT, detected fluorometrically at the fraction of a parts-per-billion level. This tracer is in common use for hydrologic studies of domestic water systems and has been used for similar measurements in other glaciers (*15*).

A borehole drilled 385 m to the bed at Km 9.5 on 8 June became strongly coupled to the basal water system as evidenced by strong pull-down of instruments into the hole, indicating a rapid rush of water down the hole to the bed. While this condition obtained, at 11:20 on 9 June, the tracer dye was introduced into the borehole. The return of tracer at the outflow stream CS, which was the main outflow stream at this time, was followed by collection of water samples every hour for the first 24 hours after tracer injection and at increased sampling intervals thereafter (Fig. 11A). Tracer first began to appear in detectable amounts in the outflow about 24 hours after injection. The concentration climbed to about 1.7 ppb by about 50 hours after injection and remained in the range 0.7 to 2.2 ppb until about 130 hours, after which it slowly decreased, reaching the detection limit ($\sim$0.05 ppb) about 15 days after injection. The mean transit time for the dye was about 90 hours. The average flow speed of water in the basal system was therefore about 0.02 m/sec over the 8-km interval from Km 9.5 to stream CS. The broad dispersion of transit time (Fig. 11A) was accompanied by a lateral dispersion of tracer in the basal water system, shown by the fact that the tracer appeared in all three outlet streams (CS, TS, and LTS).

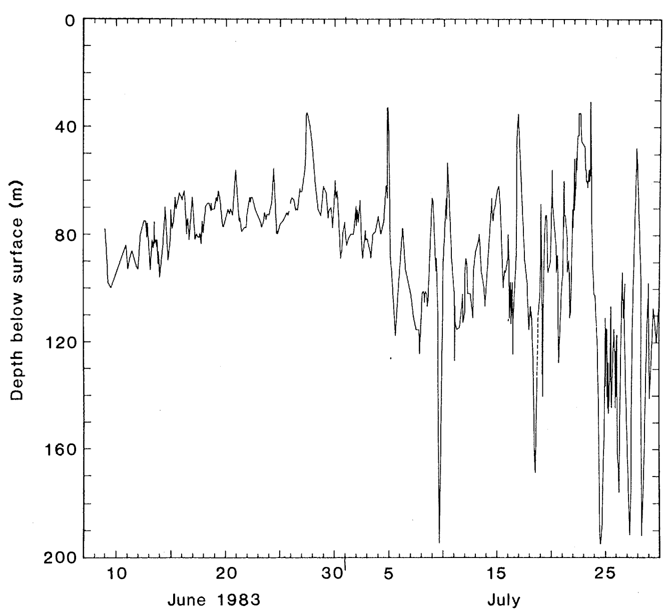


Fig. 9. Water level in boreholes near Km 9.5, measured with downhole pressure transducers. Record over 9 to 26 June is from borehole 83-E, 26 June to 18 July from borehole 83-F, and 18 to 29 July from borehole 83-G.

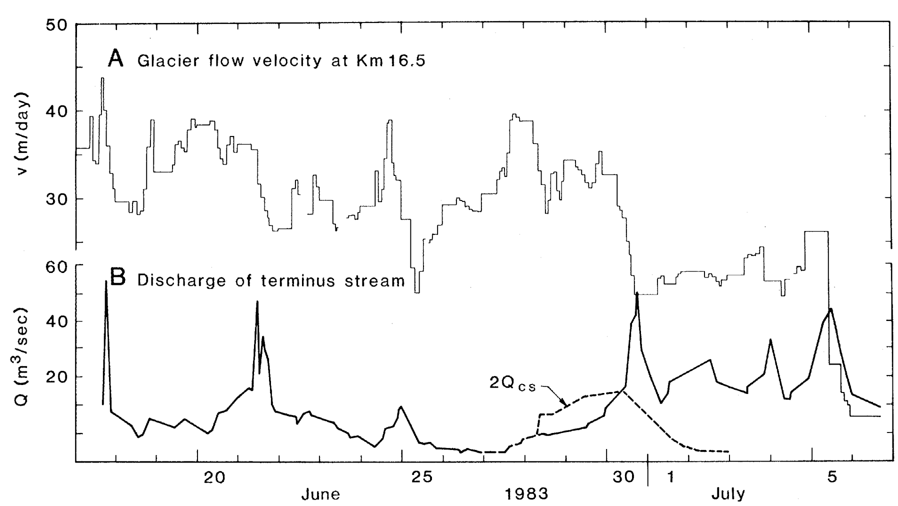


Fig. 10. Time variations of the discharge Q of the terminus outflow stream LTS (B) plotted alongside the variation in glacier flow velocity v at Km 16.5 (A), over the period 17 June to 6 July 1983. The dashed curve in (B) shows the discharge of outflow stream CS, multiplied by 2, over the period 27 June through 2 July; it follows the curve for LTS over 27 June to 10:00 on 28 June.

The foregoing behavior contrasts greatly with that observed in the water tracer experiment after surge termination. Tracer was introduced at 10:00 on 16 July via a borehole of depth 355 m located at Km 9.5. The observed tracer return in outflow stream LTS (now the main stream) is given in Fig. 11B. The tracer appeared at LTS as a sharp peak (of peak concentration 40 ppb) 4 hours after injection. The average flow speed over the 10-km interval from Km 9.5 to LTS was therefore 0.7 m/sec. This resembles flow speeds measured by a similar technique in normal alpine-valley glaciers (*15*). No detectable tracer appeared in streams TS or CS.

These experiments show that a dramatic change took place in the basal water system of Variegated Glacier from the surging to the nonsurging state. In the nonsurging state, the water flow speed was approximately the same as in normal, nonsurging-type valley glaciers, whereas, in the surging state, flow in the basal water system was greatly retarded and lateral dispersion of flow was enhanced.

Basal Cavitation

Measurements of the position of a marker stake in the glacier at Km 8 show that the glacier surface there dropped about 10 cm at the time of surge termination (Fig. 12). Similarly, in the major slowdown on 26 June 1982 the glacier surface at point T at Km 6.5 (Fig. 1) dropped 70 cm. These vertical motions may be the surface indication of the closing up of cavities between the base of the ice and the bed. Such cavities form when a glacier slides over an irregular bed with sufficient speed and at a sufficiently high basal water pressure, by a process known as ice-bedrock separation or basal cavitation (*16–18*). The cavities close by collapse of the ice roof when the water pressure falls or the sliding speed decreases, or both. In the surge, the opening of these cavities is not detectable by the method of Fig. 12, because it takes place too slowly, except possibly for the uplift of about 5 cm between data points 1 and 2 in Fig. 12, which occurred in connection with the speedup just before the onset of surge termination. But in the minisurges observed in detail in 1980 (*10*), each quick speedup was accompanied by a rapid uplift of 5 to 10 cm, followed by a slower drop back to the original level, which constitutes an observation of both the opening and the subsequent closing of basal cavities. Iken *et al.* (*19*) have similarly interpreted surface uplifts and drops connected with flow-velocity fluctuations in Unteraar Glacier, Switzerland.

An interpretation of the drop in Fig. 12 in terms of basal cavitation is subject to the qualification that a drop of comparable amount would be expected on account of the strain wave that accompanied the downglacier-propagating wave of velocity slowdown. In the minisurges of 1980, we observed examples in which the strain that accompanied the uplift and subsequent drop was of such a character that the uplift and drop must have been due to changes in basal cavitation, but in the surge termination in 1983 the drop could have been due to strain, at least in part.

If water-filled basal cavities open, this provides a means for storage of water at the base of the glacier, and, if such cavities close at the surge termination and the major slowdowns, water would be expelled, contributing to the floods that accompany these events. If the areal average of the cavity height is 0.1 m, as the drop in Fig. 12 may indicate, the volume of water stored is about 1.7×10^6 m^3 over the surging part of the glacier, about 17 km long and 1 km wide. Release of this water over a period of 20 hours, as in the outflow-stream flood at surge termination, would add an average discharge of about 24 m^3/sec, which is comparable to the observed flood (peak discharge 50 m^3/sec, Fig. 10). However, this contribution to the flood would be difficult to disentangle from that due to the release of intraglacially stored water. If the intraglacial void space that is connected to the basal water system in such a way that the contained water can drain out amounts to a bulk porosity of 0.1 percent and if the ~50-m drop in water level at surge termination (indicated by the borehole water levels, Fig. 9) is applicable to the entire surging part of the glacier, the volume of water released would be about 1×10^6 m^3, comparable to the amount that might be released from basal cavities.

The closure of basal cavities would increase the area of contact between the glacier and its bed, which would necessarily increase the resistance to basal sliding. This accords with the observed correlation between floods and slowdowns.

The Basal Water System in Surge and Nonsurge

In the nonsurging state, the transport of water in the basal water system is thought to be describable by the model discussed by Röthlisberger (*20*). There is only one main conduit, or at most only a few, carrying most of the subglacial discharge. It can be approximated as a semicircular tunnel, a few meters in diameter, carved by frictional melting into the ice directly overlying the bed. The water pressure is well below (~10 to 20 bars below) the overburden pressure of ice, and the tunnel is kept open by the frictional heat generated by viscous dissipation in the flow of water through it, which enlarges the tunnel by melting at a rate that compensates for the closure of the tunnel by quasi-viscous creep of the ice under the ice overburden pressure less the water pressure. We will call a basal water system of this normal type a "basal ice tunnel system." For Variegated Glacier in the postsurge condition in mid-July 1983, the diameter of the main tunnel was probably about 4.5 m, to account for the observed outflow discharge of about 20 m^3/sec.

Water transport through Variegated Glacier in surge cannot be in a normal tunnel system, because the observed average water-flow speed (0.02 m/sec) is

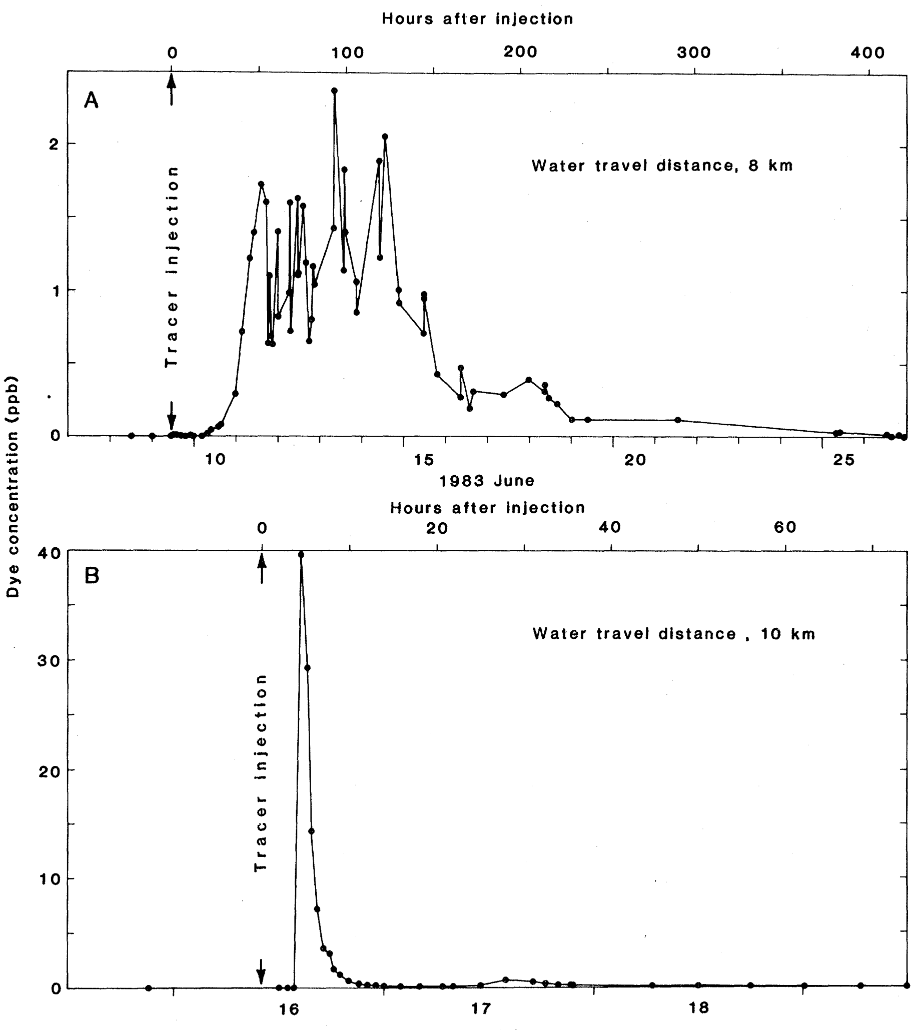


Fig. 11. Results of water tracer experiments under surging conditions (A) and after surge termination (B). The tracer dye (rhodamine WT) was introduced into the glacier in a borehole at Km 9.5 at the times indicated by the arrows. (A) Tracer content of terminus outflow stream CS is plotted over the period 9 June to 1 July, showing the results of the first injection. (B) Tracer content of stream LTS is plotted over the period 15 to 19 July, showing the results of the second injection. Note the different abscissa and ordinate scales for (A) and (B).

much too slow in relation to the outlet-stream discharge at the time of observation (7 m^3/sec). To permit a water-flow speed of only 0.02 m/sec down the actual gradient (0.1), a tunnel would have to be only ~1 mm in diameter and would carry a discharge of only $\sim 10^{-8}$ m^3/sec. Thus there must be many water-conducting passageways, distributed across the width of the glacier bed. A limiting case of this would be a continuous gap, about 1 mm thick, between the ice and its bed, which is similar to the model proposed for surging by Weertman (*21*). However, even in this limiting case the water flux carried by the system would be much too small, only ~0.02 m^3/sec. From this we are forced to conclude that the system consists not of conduits or gaps of uniform width or thickness but rather of relatively large cavities linked by narrow, gaplike orifices that conduct the water flow from one cavity to the next. The overall flow rate (water discharge) is controlled by high-speed flow through the narrow orifices, whereas most of the travel time for water transport through the system is taken in slow-speed flow through the large cavities.

We believe that this system of linked cavities forms by the process of basal cavitation (*16*). The cavities, which form on the lee sides of bedrock protuberances, probably have horizontal dimensions of a few meters and heights of a meter or less. The connections between them, which are in places where cavitation occurs only marginally, must be only a few centimeters high to throttle the flow.

A quantitative theoretical model of such a system has been constructed (*22*), which is able to account for the observed discharge of 7 m^3/sec and transit time of 90 hours. These figures require that the cavities have an average area of 200 m^2 in lateral cross section, which means that their average height is 200 m^2 ÷ 1000 m (the glacier width) = 0.2 m. The opening or closing of such a cavity system would store or release water as discussed earlier and would cause the glacier surface to rise or fall by 0.2 m, which is of the order of magnitude of the drop actually observed at surge termination. The widespread distribution of the passageways across the glacier bed accords with the greatly increased turbidity of the outflow water during surge: by comparison with a normal tunnel system, it allows a much more extensive access of the basal water to the source of comminuted rock debris that is generally present between the base of the ice and the underlying bedrock (*23*).

In the linked-basal-cavity system, by comparison with a tunnel system, the enlargement of the passageways by frictional melting is greatly reduced, because the total ice surface area of the cavity roofs is much greater than the roof area of a tunnel carrying the same discharge. The total frictional heat, which for the same discharge and same hydraulic gradient will be the same in both systems, is thus spread out in the linked-cavity system over a much larger ice surface area and the rate of enlargement of the cavities by melting is proportionately reduced. To keep the conduits open enough to provide the hydraulic conductivity needed to carry a given discharge therefore requires a much higher basal water pressure in the linked-cavity system than in the tunnel system.

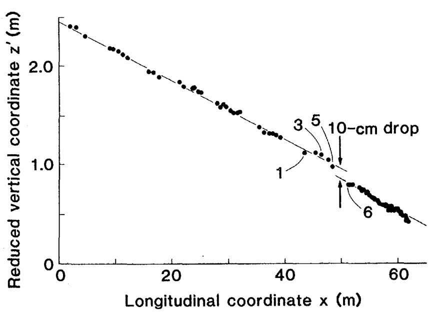


Fig. 12. Vertical displacement of a marker emplaced in ice at the glacier center line near Km 8, plotted against horizontal displacement. The reduced vertical coordinate z' is the actual elevation z reduced by an amount corresponding to motion down a sloping line of constant slope 0.125; thus $z' = z - 0.125x - z_0$, where z_0 is a fixed arbitrary constant; x is the horizontal distance in the direction of glacier flow (N50°W), with arbitrary zero. The drop of 0.1 m that occurs between measurement points 4 and 6, shown by the offset between the sloping lines fitted to the points before and after, may be due to the closure of basal cavities when the surge ended. The measurement times of the numbered points are as follows: 1, 4 July 1983, 10:34; 2, 15:11; 3, 17:24; 4, 20:09; 5, 22:04; 6, 5 July 1983, 12:11; and 7, 13:49.

In the linked-cavity system, an increase in basal water pressure enlarges the cavities and therefore increases the basal water flux, under otherwise fixed conditions (*24*). This behavior is opposite to that of a normal tunnel system, in which an increase in basal water flux results in a decrease in water pressure in the tunnel or tunnels, once transient effects have died out and a steady state is established; this is so because the increased melting caused by increased water flow must be balanced by an increased viscous closure rate for the tunnel, which requires a drop in the water pressure (*20*). One consequence of the inverse relation between water flux and pressure in a tunnel system is that, if there is more than one tunnel, the larger ones will grow at the expense of the smaller, so that the system evolves into one with only a single tunnel or at most very few (*20, 25*). In contrast, because of the direct relation between water flux and pressure, passageways in a linked-cavity system are stable against perturbations in which some passageways are enlarged at the expense of the others; as a result, the linked-cavity system can exist stably as a system of multiple interconnecting passageways distributed over the width of the glacier bed.

The Surge Mechanism

The immediate cause of the high glacier flow velocities in the surge is high basal water pressure, which causes a great increase in basal sliding. The observational basis for this conclusion was given above. According to theoretical discussions of the basal sliding process (*18*), increased basal water pressure causes increased sliding by reducing friction and by increasing basal cavitation, and also by exerting a direct effect of the integral of pressure over the irregular undulating base of the ice mass.

Theoretically, if the basal water pressure were to reach and remain at the ice overburden pressure, the glacier would be floated off its bed, which would cause the sliding velocity to increase practically without limit. A more limited effect of this kind must be produced when the basal water pressure approaches but does not reach the ice overburden pressure. How closely the water pressure must approach the overburden to give sliding velocities of the magnitude observed in the surge (~15 m/day in the upper glacier, ~50 m/day in the lower) is from the theoretical point of view (*18*) a complicated question. Empirically, on the basis of the evidence presented above, we conclude that basal water pressures consistently within 4 to 5 bars of the overburden, and with transient fluctuations close to the overburden, are sufficient to cause these high sliding velocities. In minisurges, the peak basal water pressures are higher, normally at or above overburden, while the peak sliding velocities (1 to 3 m/day) are lower than in surge; this is probably because the high pressure does not gain access to more than a fraction of the area of the bed in the relatively short duration of the high-pressure peak (~10 hours).

The role of high water pressure in causing the rapid basal sliding motion in the surge of a glacier is analogous to its proposed role in the mechanics of overthrust faulting and landslides (26). Landslides sometimes show periods of greatly enhanced movement that are called surges (27).

The key element of the surge mechanism (and of the overthrust and landslide mechanisms also) is what enables the high basal water pressures to be built up and maintained. The difference in the basal water system between nonsurging and surging conditions is crucial. The pressure in the system is the result of a balance between buildup of pressure from water input and drawdown of pressure by outflow. Since increased pressure tends to increase the cavitation and hydraulic conductivity of the basal water system, it seems at first sight paradoxical that the higher basal water pressure in the surging state can be accompanied by increased retention of water in the glacier, as is necessary to maintain the higher water pressure. However, the linked-basal-cavity water system of the surging state has characteristics that can maintain the higher basal water pressure. An essential characteristic is the stability of the linked-cavity system against the preferential growth of larger passageways, as occurs in a normal tunnel system. Without this stability, the linked-cavity system would degenerate into a tunnel system, as indeed it does in the surge termination. There are hints of instability in the major slowdowns prior to surge termination, but in these the linked-cavity system proves itself robust and recovers. Theoretical modeling shows that the primary factor responsible for the stability is the effect of rapid basal sliding, which provides a powerful offset to the enlargement of the cavity-connecting orifices by frictional melting and is able to stabilize the orifices at a size only modestly enlarged from what they would be in the absence of melting (22).

From this perspective, what is required for a glacier to be in the surging state is a kind of "bootstrapping" in which high basal water pressure is needed to cause rapid sliding, while rapid sliding is needed to provide the cavity stability that permits the high basal water pressure to be maintained.

We visualize that in the surge front as it advances the great increase in basal sliding of the ice, shoved forward by the surging mass behind it, destroys the tunnels of the normal basal water system that exist below the front and at the same time causes widespread basal cavitation, thus generating a linked-cavity network that takes over as the basal water system above the front.

Because of the inverse relation between flux and pressure in a normal tunnel system, high water pressures in this system can be maintained only in winter, when the water flux is very low. It therefore makes sense in a general way that the time when the pressure might rise high enough to cause the transition to the surging state is in winter, which is when the surge actually started.

Although the foregoing ideas about the fundamental differences between the basal water systems in the nonsurging and surging states provide a working hypothesis for understanding the surge mechanism, numerous features of the surge remain to be understood. What are the switchover processes in the basal water system that start the surge in winter and terminate it in summer? What is the water source, intra- or extraglacial, that in midwinter provides the water to pressurize the basal water system and initiate the surge? How important in the surge mechanism is the water produced by frictional melting due to basal sliding? How does the surge mechanism control the advance of the surge front? What causes the oscillations or fluctuations in surge motion on various time scales and what controls their propagation downglacier? What is the relative importance of basal water volume versus pressure in causing the large basal sliding rates in the surge? What are the roles of the numerous prominent structural features—wrench faults, thrust faults, folds on various scales, and intense crevassing—produced by the surge and doubtless connected somehow with the surge mechanism (28)?

We hope that our observations of the 1982–1983 surge of Variegated Glacier and the resulting inferences about the surge mechanism will be useful in the further efforts needed to understand the many intriguing aspects of the glacier surge phenomenon. It will be necessary to determine how general the inferred surge mechanism is and whether there are other types of glacier surges in which other mechanisms play a predominating role.

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