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How Do Glaciers Surge? A Review

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The quasi-periodic oscillations between normal and fast motion exhibited by surge-type glaciers provide the best observational opportunity to determine limiting conditions that allow fast motion. The measurements from Variegated Glacier prove that its surge motion is caused by rapid sliding induced by high water pressure. This arises from a major restructuring of the basal hydraulic system, which impedes water discharge prior to and during surge. Although the evolving glacier geometry and stress distribution play a principal enabling role, the seasonal timing of two distinct surge pulses, each initiated in winter and terminated in summer, indicates a major influence from variable external water inputs. This influence is not considered in existing surge models and should promote caution in the use of data from temperate and subpolar surge-type glaciers to deduce surge potential in polar ice masses. The spatial spreading of surge motion from a zone of local initiation occurs by stress redistribution, which may spread the surging zone rapidly upglacier or downglacier inside a region of active ice, and by mass redistribution with compressional thickening at the surge front, which enables down-glacier propagation into less active ice. The data from other surge-type glaciers, including the extensive data from Medvezhiy Glacier, are not inconsistent with the above processes but are inadequate to establish whether completely different mechanisms operate in some surges. The way by which water accumulates and produces fast sliding is not established in detail for any surge-type glacier and may be different on different glaciers depending, for example, on the presence or absence of unconsolidated debris between the ice and rock.

1. INTRODUCTION

Meier and Post [1969] asked the question, "What are glacier surges?" Their answer has provided the definition of surge behavior. Briefly stated, it is a behavior characterized by a multiyear, quasi-periodic oscillation between extended periods of normal motion and brief periods of comparatively fast motion. Glaciers showing this behavior, here called "surge-type" glaciers, have been identified in various mountain ranges of the world [*Post*, 1969; *Dolgushin and Osipova*, 1975] and sections of ice caps [*Thorarinsson*, 1969; *Liestøl*, 1969]. They represent only a small percentage of all glaciers and are highly concentrated in some glaciated regions and totally absent in others.

The restricted geographical distribution of surge-type glaciers indicates special environmental conditions are required. Surge-type glaciers have been identified in a variety of climates, both maritime and continental, with both temperate and subpolar thermal regimes, and with a wide range of sizes and other geometrical characteristics [*Post*, 1969]. Thus the environmental control is not obvious. A detailed examination of the population of glaciers in the St. Elias mountains of Canada has shown that long glaciers have a greater likelihood of being surge type than short ones, but the meaning of this is not clear [*Clarke et al.*, 1986]. Mass balance has been estimated for a few selected surge-type glaciers; the results suggested that a steady rate of ice transport in these glaciers would correspond to a rate of potential energy loss per unit area that is higher than typical of normal glaciers and lower than typical of continuously fast moving outlet glaciers [*Budd*, 1975]. Recent examination of mass balance estimates for a larger number of normal and surge-type glaciers has not confirmed this idea [*Wilbur*, 1986]. The combinations of environmental factors that cause surge behavior are still unidentified.

Meier and Post [1969] suggested the possibility of several classes of surge behavior depending on glacier size and speed achieved during surge. Some glaciers are also known to show multiyear, periodic, pulselike increases of speed that are not large enough to produce the large ice displacements usually associated with glacier surges and therefore appear to be intermediate between surge-type glaciers and normal glaciers [*Mayo*, 1978]. Many normally flowing glaciers, including surge-type glaciers in their quiescent phase, show seasonal variation of velocity [*Hodge*, 1974; *Aellen and Iken*, 1979]. At yet shorter time scales, complex variations with time have been found [*Iken*, 1978; *Iken et al.*, 1983]. Some of these variations, termed minisurges, occur as short (~1 day long) pulses of increased speed that recur repeatedly at multiday intervals; thus they have some features of the periodic cycle of surges [*Kamb and Engelhardt*, 1987; *Harrison et al.*, 1986*a*]. These phenomena indicate the possibility that surge behavior is an extreme end-member in a continuum of possible pulsating flow behavior. The mechanistic relationship between these diverse time-dependent phenomena is a major unanswered question.

The continuous fast motion of ice streams [*Bentley*, this issue] and of the terminal zones of many grounded tidewater glaciers [*Meier and Post*, this issue] has been described as a state of continuous surge [*Weertman*, 1964]. While this is a very reasonable supposition, it remains as a question to be addressed.

The two central questions about the processes involved in surge behavior have long been identified as (1) what is the mechanism of fast motion during a surge? and (2) what initiates and terminates the fast motion? This paper focuses on available observations that provide answers to these two questions at a broad level. Eventually, detailed understanding of the processes can illuminate related questions, for example, the puzzling geographical distribution of surge-type glaciers; the periodic cycling between normal and surging flow; the relationship between surge behavior, normal glaciers, and the continuous fast motion of ice streams; and the surging potential of polar ice masses.

2. DESCRIPTION OF SURGE CYCLE

2.1. *Characteristics of the Cycle*

Surge behavior as described by *Meier and Post* [1969] is characterized as follows. (1) Surges occur repeatedly. (2) In a single glacier the quiescent interval between surges is fairly con-

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TABLE 1. Comparison of Characteristics of Variegated and Medvezhiy Glaciers

	Variegated	Medvezhiy
General		
Total length, km	20	16
Surging length, km	18	7-8
Surge period, years	≈18	9-14
Mean slope of surging length	0.094	0.11
10-m temperature	temperate	−0.9° C
Surges with measurements	1982-1983	1963 and 1973
Quiescent phase		
Terminus retreat, km	nil	<1.5
Advance of dynamic balance line	1	>4
Elevation change, m	+60 (-40)	+120 (-100)
Maximum τ_b, 10^5 Pa	1.8	?
Maximum u annual average, m d^{-1}	0.6	1.5
Summer velocity increase, %	80	100
Surge phase	1982-1983 surge	1963 and 1973 surges
Number of events/duration	Two, 6 to 8 months	1
Timing of initiation	early winter, late fall	late winter (?)
Timing of termination	early summer	early summer
Advance of surge front, km	12	?
Advance of topographic peak, km	11	?
Thickness change maximum, m	110	120
Advance of velocity peak, km	11	?
Maximum speed, m d^{-1}	65	100
Maximum ice displacement, km	≈2	>1.6

stant (10^1 to 10^2 years). (3) The surge phase is relatively short (several years). (4) During the surge phase, ice speed is 10^1 or more times the speed during quiescence; accumulated ice displacement may be 10^{-1} or more of the glacier length; ice is drained rapidly from an upglacier reservoir area to a downglacier receiving area; large elevation drops and rises (10^1–10^2 m) occur in the reservoir and receiving areas. (5) During the quiescent phase, ice speed is low; accumulated displacement is smaller than during surge; ice builds up in the reservoir area and is lost in the receiving area; there are progressive thickness changes that reverse the thickness changes of the surge and gradually return the glacier to near its presurge state.

These general characteristics have been found on a large number of surge-type glaciers (see *Meier and Post* [1969] and and *Dolqushin and Osipova* [1975] for reviews). Measurements of surface elevation and velocity over many years on several glaciers are now providing quantitative descriptions of various aspects of surge cycles, for example, Black Rapids, Alaska Range (L. R. Mayo, D. Trabant, and others, unpublished data, 1986); Trapridge, St. Elias Mountains [*Clarke et al.*, 1984]; Medvezhiy, Pamirs [*Dolgushin and Osipova*, 1975]; Variegated, St. Elias Mountains [*Kamb et al.*, 1985; C. F. Raymond and W. D. Harrison, Progressive changes in geometry and velocity of Variegated Glacier prior to its surge, submitted to *Journal of Glaciology*, 1986 (hereafter referred to as RH86)]. The measurements from the Medvezhiy and Variegated glaciers are the most extensive and thus far are the only ones covering nearly complete surge cycles. Table 1 compares the characteristics of these two glaciers.

2.2. *Quiescent Phase*

Evolution of geometry and velocity. The evolution of the geometry and velocity found during the quiescent periods of the Medvezhiy (1963-1973) and Variegated (1973-1981) glaciers is summarized in Figure 1. The data from both glaciers show the filling of a reservoir area and depletion in a receiving area. This is accompanied by increasing velocity in the reservoir area. The boundary between thickening and thinning can be located on an annual basis. This has been referred to as the dynamic balance line (DBL) [*Dolgushin and Osipova*, 1978];

According to *Dolgushin and Osipova* [1975, 1978], the evolution of the Medvezhiy Glacier during quiescence between the surges of 1963 and 1973 was dominated by a year-by-year advance of the DBL as a distinct front separating stagnant ice below and upbulging obviously active ice behind. The front steepened with time. This was initiated near the base of the ice fall, which feeds ice to the lower surging part of the glacier.

Although the velocity in the active zone increased dramatically over the quiescent interval (1963 to 1972), the year-by-year changes were not progressive (Figure 1). There was also a strong seasonal variation of velocity. The year-by-year advance of the DBL is described as occurring by a sequence of "wavy surges," and the active zone behind it was severely cracked, similar to, although not so thoroughly as, during surge motions. *Dolgushin and Osipova* [1975] therefore suggest that the evolution during the quiescent phase occurred by processes differing only in degree from the surge motion itself.

Observations cover only the later half of the most recent quiescent phase of Variegated Glacier bounded by surges in 1964-1965 and 1982-1983. In the time span of the observations, the DBL moved only slightly downglacier and could not be easily identified by any dramatic surface morphological features, such as crevasse patterns. The principal features of the evolution are the progressive changes in elevation and steepening of a large fraction of the glacier length. The accompanying velocity changes were also progressive, which according to RH86 can be largely explained by changing ice deformation rates determined by the evolving thickness and stress distribution. Faster motion in summer than winter indicated a seasonal sliding contribution [*Bindschadler et al.*, 1976]. Between 1978 and 1981, the increase in velocity during winter appeared to be too fast to explain by ice deformation, and possibly sliding was also increasing during winter under parts of the upper glacier. Aside from the progressive evolution of geometry and velocity, the dynamic activity of Variegated Glacier during its quiescent phase did not appear to be

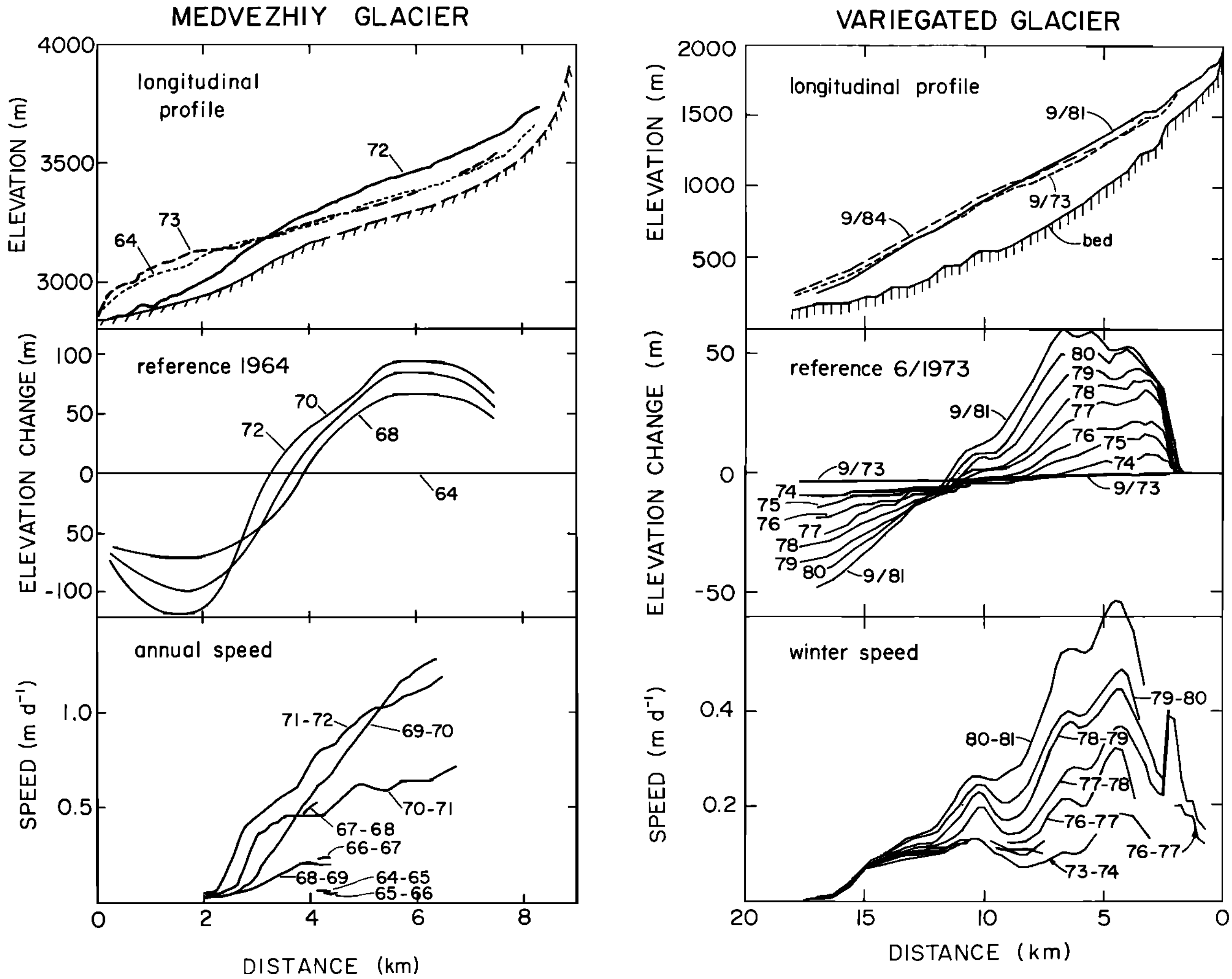

Fig. 1. Evolution of the topography and velocity during the quiescent periods of the Medvezhiy Glacier [*Dolgushin et al.*, 1974; *Dolgushin and Osipova*, 1975] and Variegated Glacier (RH86).

distinct from typical mountain glaciers with normal flow regimes [*Bindschadler et al.*, 1977].

Minisurges. One initially appearing, anomalous behavior found on Variegated Glacier was the occurrence of "minisurges" in the thickening reservoir area [*Kamb and Engelhardt*, 1987; *Raymond and Malone*, 1987; *Harrison et al.*, 1986*a*]. These are repeated velocity pulses having features of surges but on a much smaller scale.

Observed at a fixed location, a minisurge is an abrupt increase in speed over a few hours followed by a slower decay, over about 1 day, to near background speed. Dramatically increased seismic activity, anomalous longitudinal strain rates, surface elevation changes, and large basal water pressure variations occur in association with the speed changes (Figure 2*a*). It is evident that the high basal water pressure is the cause of the fast motion.

In space a minisurge propagates downglacier as coupled velocity and basal hydraulic waves (Figure 2*b*) with wave speed in the range 0.1 to 0.6 km h^{-1}. The anomalous motion also introduces fine rock debris into the water flowing along the bed, which travels as a pulse of highly turbid water to the terminal stream at an average speed of about 1 km h^{-1} [*Humphrey et al.*, 1986].

These minisurges occurred in the early melt season in a sequence of four to six, spaced at several days to 2 weeks. They affected only the upper part of the glacier (the reservoir zone). The first minisurge of a season brought the transition from slower winter to faster summer speed. These minisurge sequences are known to have occurred in the four melt seasons before the most recent surge; it is possible they occurred unnoticed in earlier years. Short-term velocity variations at other times and locations also occurred [*Harrison et al.*, 1986*a*] but were not nearly so dramatic as the minisurges.

Although the fast motion of minisurges may have some mechanistic relationship to surge motion and their occurrence may be premonitory to a surge, similar motion pulses occur on Alpine glaciers that are not known to surge [*Iken*, 1978; *Iken et al.*, 1983]. Therefore even minisurge behavior of Variegated Glacier during the quiescent phase does not definitely distinguish it from normal glaciers.

The relationship between the minisurges of Variegated Glacier and the "wavy surges" of Medvezhiy Glacier is an obvious question with no clear answer at present. On Variegated Glacier the extra motion associated with minisurges was only a small fraction of the summer velocity increase averaged over the full summer (about 30%). Could a similar minisurge phenomenon be more predominant on Medvezhiy Glacier and account for the wavy surges reported by *Dolgushin and Osipova* [1975]? *Kazanskiy et*

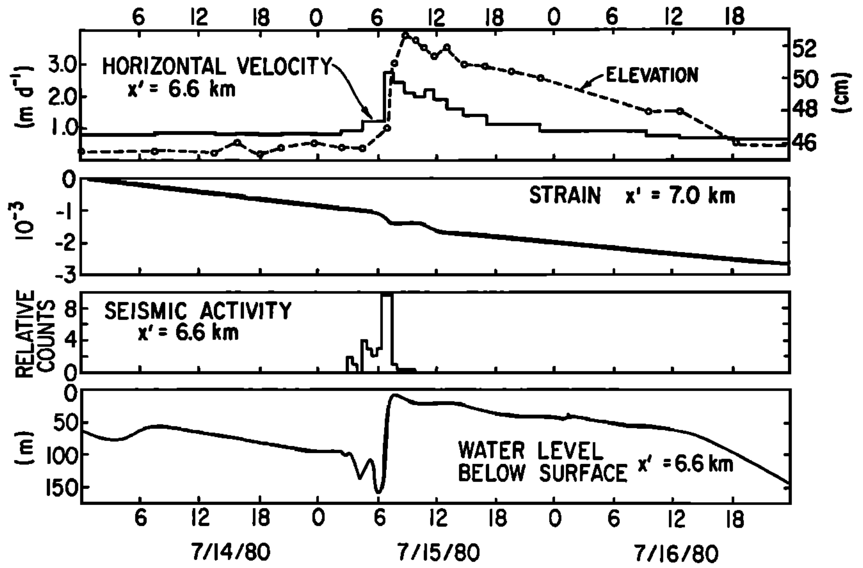

Fig. 2a. Variation of velocity, surface elevation, longitudinal strain, seismic activity, and borehole water level measured near 7 km from the head of Variegated Glacier during the fifth minisurge of 1980 (data from *Kamb and Engelhardt* [1987] and *Raymond and Malone* [1986]).

al. [1984] have examined the short time-scale velocity variations of Medvezhiy Glacier and report about 50% of the motion is associated with a pulsing component.

Flow models. The evolution of geometry and velocity during the quiescent phases of both the Medvezhiy and Variegated glaciers has been simulated by glacier flow models based on parameterization of ice deformation with negligible contributions from basal sliding [*Budd and McInnis,* 1978; *Bindschadler,* 1982]. These models show that the pattern of thickening, the advance of the DBL, and the velocity increase may come about from a normal response to the imposed mass balance. Some of the differences between the Medvezhiy and Variegated glaciers may arise because the reservoir areas are fed differently. The reservoir area of Medvezhiy Glacier is entirely within ablation area, and ice income is fed to it by a steep ice fall at its head. On Variegated Glacier, the reservoir area is mostly within the accumulation area. This illustrates the fact pointed out by *Meier and Post* [1969] that the DBL has no essential relationship with the mass balance equilibrium line.

The models, however, do not have the necessary physics to simulate seasonal velocity variations, minisurges, or wavy surges, nor can they explain certain details of the Variegated Glacier velocity distribution (RH86) or the nonprogressive increase in velocity on Medvezhiy Glacier.

Subpolar conditions. Trapridge Glacier illustrates a somewhat different flow evolution during quiescence [*Clarke et al.*, 1984]. It is dominated by spatial variations of basal temperature (Figure 3). A distinct front developed between warm-based active ice and thin cold-based stagnant ice. The DBL is then located by a boundary between zones of distinct basal slip conditions where ice can slide and where it cannot. This behavior may be common for surge-type glaciers in subpolar environments [*Schytt,* 1969], but the direct temperature measurements to document this are not available except on Trapridge Glacier. Some surge-type glaciers in subpolar environments, for example, Steele Glacier [*Clarke and Jarvis,* 1976] and Black Rapids Glacier [*Harrison et al.*, 1975], appear to have temperate wet bases and are, perhaps, not affected by the kind of basal temperature distribution found on Trapridge Glacier.

2.3. *Surge Phase*

In broad outline, the surge may be described by the rapid reversal of the geometrical evolution during quiescence (Figure 1). The net changes caused by a surge have been observed on a number of surge-type glaciers, and these lead to the setting down of the general characteristics of a surge by *Meier and Post* [1969] (see section 2.1 above). A detailed quantitative picture of the evolution during a surge is known only from the 1982-1983 surge of Variegated Glacier [*Kamb et al.*, 1985].

Velocity variation during a surge. The day-by-day variations of velocity found on the upper and lower parts of Variegated Glacier in 1982 and 1983 are shown in Figure 4. The surge motion started on the upper part of the glacier in midwinter 1982, accelerated smoothly and rapidly in the spring of 1982, and then terminated in late June and early July. The surge motion reinitiated again on the upper glacier early the following winter 1982-1983 with a similar pattern of acceleration in spring 1983 and termination in early summer 1983. This second surge pulse spread downglacier to involve nearly the full length of the glacier.

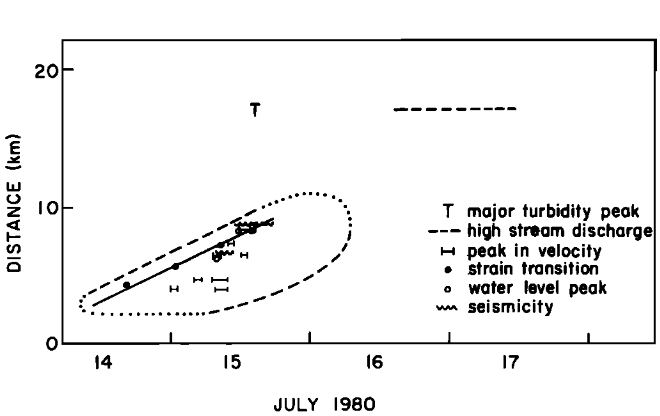

Fig. 2b. Timing of events at various longitudinal positions on Variegated Glacier and its discharge stream during the fifth minisurge of 1980. The dash-dot curve encloses the approximate space-time limits of anomalous activity associated with the minisurge (data from *Kamb and Engelhardt* [1987], *Raymond and Malone* [1986], and *Humphrey et al.*, [1986]).

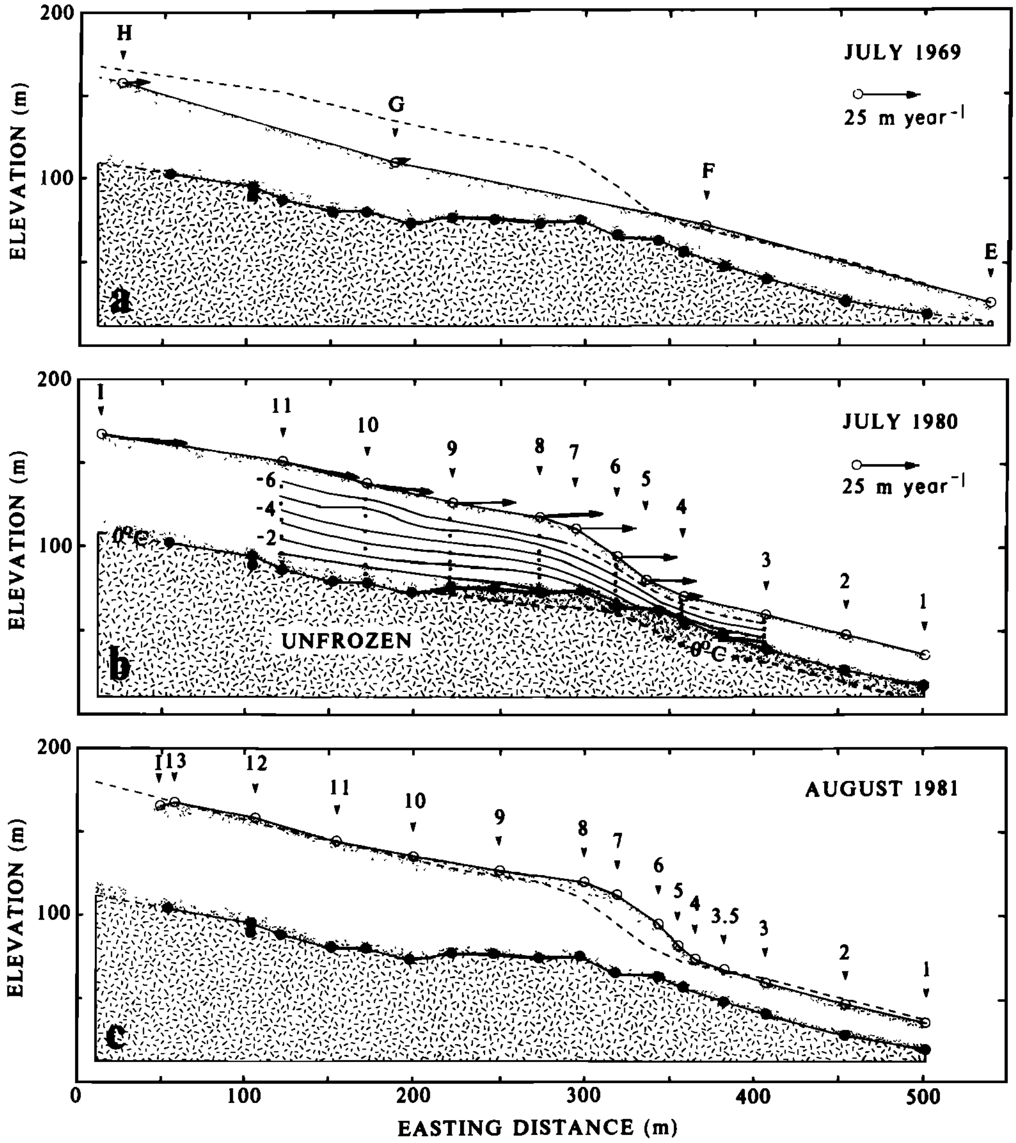

Fig. 3. Changes in Trapridge Glacier between 1969 and 1981 [from *Clarke et al.*, 1984].

The highest speeds occurred on the lower part of the glacier in association with the propagating frontal edge of the surging ice.

The broad-scale space and time evolution of the surge motion and associated ice surface topography are shown in Figure 5. The initial surge motion in early winter 1982 was centered in the upper part of the reservoir area between 3 and 6 km from the glacier head. This zone lay on the upglacier side of the peak in thickness change accumulated in the prior decade (Figure 1).

Propagation of surge motion. After initiation, the surging zone spread by a cooperative downglacier propagation of velocity and topographic disturbances. The leading edges of these disturbances closely coincided and were followed by peaks in velocity and elevation increase. As they propagated downvalley, these peaks increased in amplitude (Figures 5*d* and 5*e*) and approached the leading edge of the surging ice (Figure 5*c*), resulting in an increasingly dramatic shocklike front of steep surface slope (Figure 5*d*) and extraordinarily high longitudinal compression (Figure 5*e*). The motions in this advancing frontal zone were measured in detail by *Raymond et al.* [this issue]. Locally, the front slope exceeded 15°, and the compressional strain rate reached 0.2 d^{-1} for periods of several hours. The peak in thickness change led the peak in velocity, and the separation decreased with time (Figure 5*c*). Behind the peaks there were extended zones of reduced surface slope and extensile strain rate (Figures 5*d* and 5*e*). At the head of the surge there was a zone from which ice had been evacuated and the net surface elevation had dropped. An apparent increase in the glacier volume was evident (Figure 5*a*). This arose partly because the surge motion produced convex upward surface profiles and the center was not representative of width-averaged elevation. However, there was a real volume increase, presumably caused by generation of void space. This amounted to about 10% in the lower part of the surging zone.

Similar features of surge evolution are known from other surges, but the information is less quantitative. The propagation of a surge front can be tracked in satellite images on large glaciers [*Post et al.* 1976]. Ice left hanging on valley walls above the presurge ice level is a common observation indicating the passage of an ice thickness peak [*Post and LaChapelle*, 1971]. Evolving crevasse patterns have been used to identify the moving location of a velocity peak separating zones of compression and extension

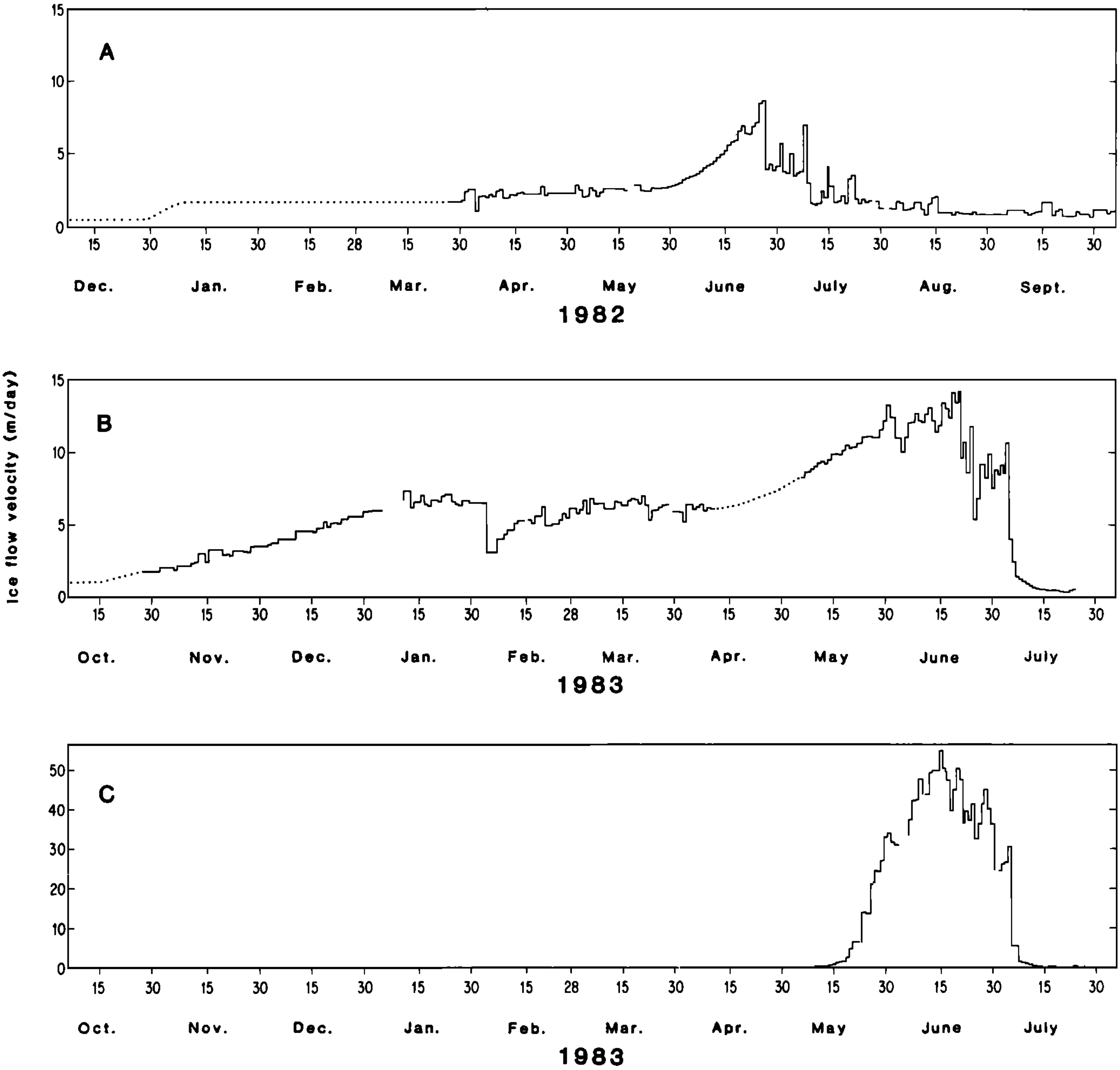

Fig. 4. Day-by-day variation of velocity measured on (*a*, *b*) the upper and (*c*) lower parts of Variegated Glacier [from *Kamb et al.*, 1985, Copyright 1985 by the AAAS].

[*Harrison*, 1964]. *Harrison* [1964] described the coordinated propagation of these features, which he called phases, down the Muldrow Glacier in its 1956 surge. Volume increases during surge motion were especially well documented by photogrammetric measurements on Medvezhiy Glacier [*Dolgushin and Osipova*, 1978].

The relative positions of the thickness change and velocity peaks during a surge (Figure 5*c*) follow rather simply from continuity. The area of high speed probably starts as a small nucleus, which spreads rapidly under the influence of stress concentrations at its edges set up by the shifting of load from the nucleus to the surrounding ice upglacier and downglacier [*McMeeking and Johnson*, 1986]. However, this rapid spreading is apparently limited to a surge initiation zone in the active ice of the reservoir area, where the presurge stress is high and susceptible to initiation of fast motion without major mass redistribution. Once the high speed is set up in this initiation zone, ice is transferred to produce thickening and thinning downglacier and upglacier from the velocity peak. If the velocity distribution were to remain fixed in space, the location of the velocity peak would define the position of the boundary between elevation rise and drop in the surge (i.e., receiving and reservoir areas). In this initial phase of the surge a topographic peak necessarily develops downglacier from the velocity peak.

As the velocity-induced mass redistribution changes the topography, feedback to the velocity distribution starts the propagation. With a steady state propagation at a constant propagation velocity c, continuity determines a relationship between the distributions of velocity $u(x,t) = u(x-ct)$ and thickness $h(x,t) = h(x-ct)$. Overall volume conservation shows that c is given by the flux (uh) and

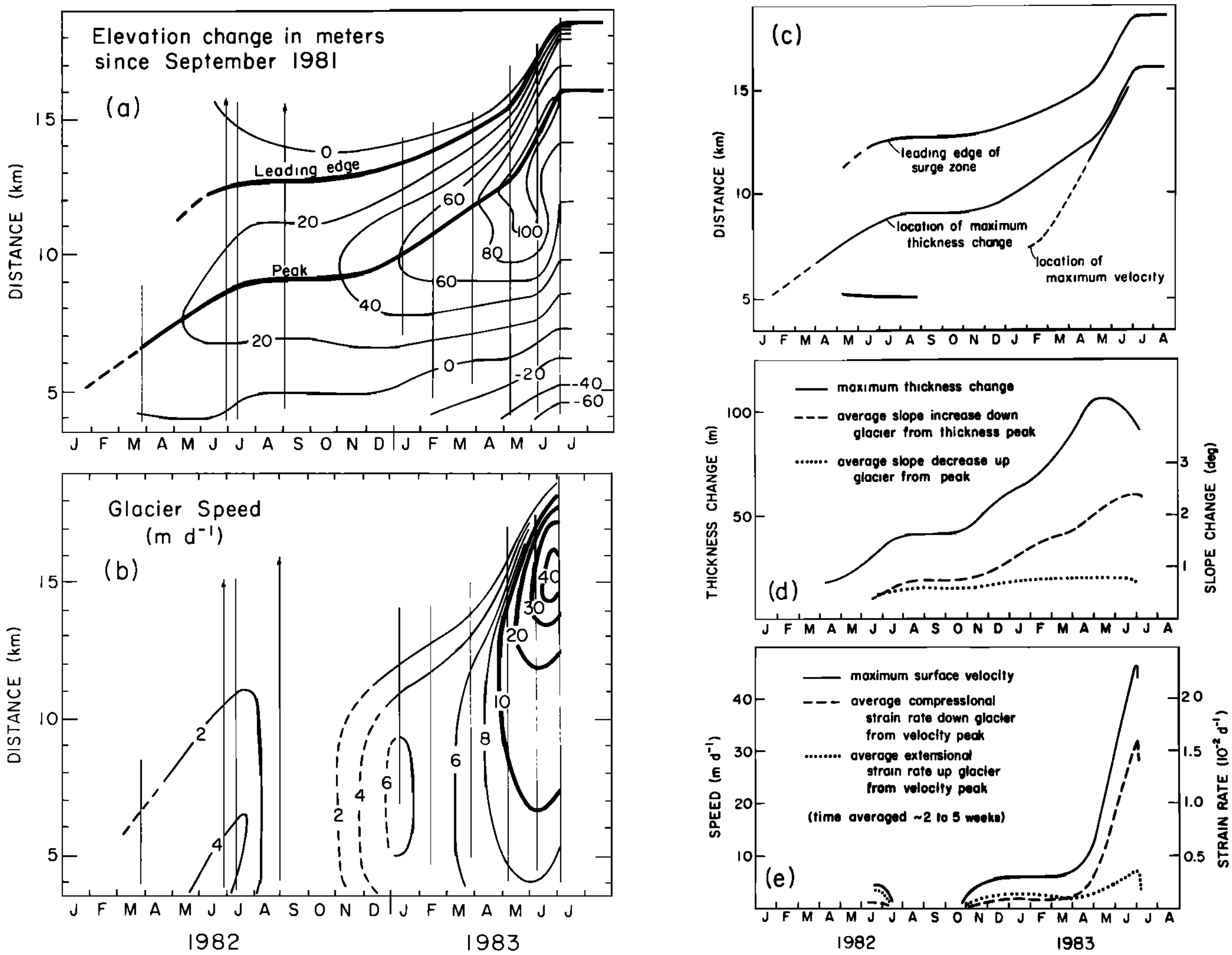

Fig. 5. Smoothed space-time evolution of topography and velocity during 1982-1983 surge of Variegated Glacier. Vertical lines in Figures 5*a* and 5*b* give times of measurements used for the construction (from C. F. Raymond, unpublished data, 1984).

concentration (h) jumps across the surge front as

$$c = \frac{[uh]}{[h]} \frac{u_f h_f - u_i h_i}{h_f - h_i} \approx \frac{u_f h_f}{(h_f - h_i)} \quad (1)$$

where f and i subscripts refer to locations well back in the surging ice (f) and in the nearly motionless ice ahead of the front (i). Similarly, conservation of volume expressed differentially, as in the development of the theory of kinematic waves [*Nye*, 1960], gives

$$0 = \frac{\partial hu}{\partial x} + \frac{\partial h}{\partial t} = h'u + hu' - ch' \quad (2)$$

assuming motion is by sliding and surface mass balance is negligible in the thickness changes. Equation (2) shows that the thickness peak ($h' = 0$) and velocity peak ($u' = 0$) must coincide.

On Variegated Glacier, the relationship between the surge edge, thickness peak, and velocity peak can be qualitatively understood as follows. The initiation phase produces a thickness peak downglacier from the velocity peak. The propagation phase never reaches steady state because of memory of the initiation and the continuous recreation of an initiation-type response by acceleration of the surge with time. The velocity and topographic peaks nearly coincided only when the growth in amplitude had slowed, as occurred in May and June of 1983 (Figures 5*c*, 5*d*, and 5*e*).

Even though the surge propagation was not steady state, (1) predicts the propagation speed found on Variegated Glacier quite well [*Kamb et al.* 1985; *Raymond et al.*, this issue]. A propagation speed in excess of the ice speed behind the front as predicted by (1) has also been found on Tweedsmuir [*Post et al.* 1976] and Muldrow [*Harrison*, 1964] glaciers. A detailed analysis of the surge evolution would, of course, require a relationship that predicts velocity u in terms of thickness profile h and other important parameters [*Fowler*, this issue].

Slowdown events during surge motion. Detailed examination of the velocity within the surging zone of Variegated Glacier shows several types of velocity fluctuation, as in Figure 6 [*Kamb et al.* 1985]. These include regular oscillations of about 2-day periods on the upper glacier, less regular oscillations of shorter periods on the lower glacier, and sequences of slowdown events separated by 4 or 5 days. Some of the fluctuations, especially the distinct slowdowns, affected nearly the full length of the surging part of the glacier.

The timing of the major slowdown events along the length of the glacier shows a downglacier propagation at speeds of about 0.6 to 0.7 km h^{-1} when the surge motion was at its height (Figure

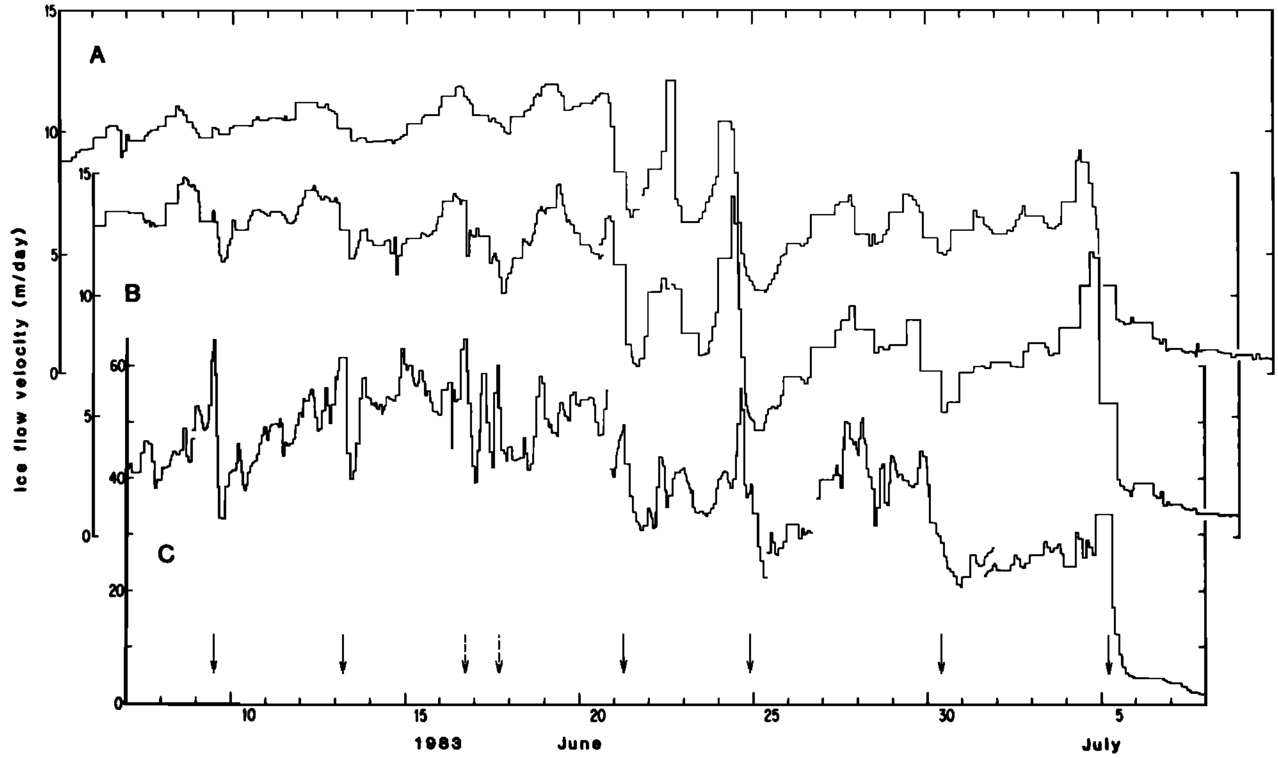

Fig. 6. Variation of velocity with time on Variegated Glacier at three locations [from *Kamb et al.*, 1985, Copyright 1985 by the AAAS]. (*a*) Upper glacier near 5.5 km from head, (*b*) midglacier near 9.5 km from head, and (*c*) lower glacier near 13 ½ to 15 km from head.

7). Because of this propagation and a premonitory increase in velocity to a sharp peak just before the velocity drop, these slowdown events show some similarity to the minisurges of the quiescent phase. Drops in the water level in a borehole located in the central reach of the surging ice occurred at the time of slowdowns and show that the slowdowns were associated with hydraulic waves propagating along the bed. Also, all slowdown events were accompanied by flood discharges in the terminal streams [*Kamb et al.*, 1985]. Slowdown events were most prominent in the early summer, but they also occurred in winter (Figure 3). The termination of each of the two surge episodes in summer 1982 and 1983 occurred by a sequence of slowdowns. The dramatic sudden termination of surge motion of July 4 and 5, 1983, occurred by the downglacier propagation of the penultimate slowdown, which returned the glacier to quiescence.

2.4. *Periodicity and Seasonal Timing*

Based on historical observations of surges and regularity in looped moraine patterns, *Post* [1969] discovered that on a given surge-type glacier, surges are separated by a relatively constant time interval, "the surge period." However, surges are not rigorously periodic. For example, on Medvezhiy Glacier the interval between surges has shortened over the last 50 years from 14 to 10 years [*Dolgushin and Osipova*, 1975; *Dolgushin et al.*, 1974]. Although the Variegated Glacier surges about every 20 years, the history of surges indicates that there are deviations of at least a few years from a regular period (RH86). On Bering Glacier, a surge occurred as two pulses but with the pulses separated by about 6 years [*Post*, 1972], which is shorter than the total cycle period of about 30 years [*Meier and Post*, 1969]. This gives an apparently irregular recurrence of surges. As more is learned about the surge history of various glaciers, one may expect to discover a number of variations and deviations from a simple periodic cycle.

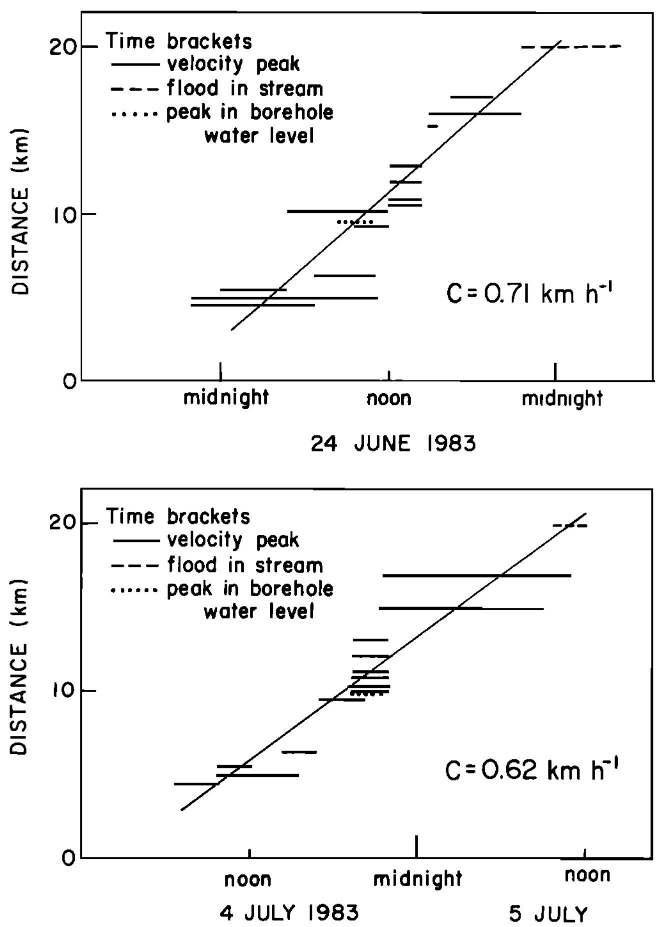

Fig. 7. Relative time of events at various positions along Variegated Glacier during major slowdown events of June 24, 1983, and July 4-5, 1983.

When different surge-type glaciers are compared, there is a substantial range of periods, with known periods ranging from less than 10 years to more than 50 years [*Meier and Post*, 1969].

The seasonal timing of surge initiation and termination provides important clues about the mechanisms involved. This is discussed below. Unfortunately, the information is limited, largely because most observations are from occasional or annual aerial photographs or restricted to termini. Each of the two pulses of the recent surge of Variegated Glacier in 1982 and 1983 initiated in winter and terminated in early summer (Figure 3). The surges of Medvezhiy Glacier in 1963 and 1973 were also consistent with this seasonal pattern [*Dolgushin and Osipova*, 1975; *Dolgushin et al.*, 1974]. Less definite information from other glaciers also seems to support winter initiation and/or summer termination (Tweedsmuir Glacier 1973 [*Post et al.*,] 1976] and Tyeen Glacier 1978 (personal observations)). At present there are no definite exceptions to this seasonal pattern. However, surge motion may sometimes continue through a summer (Muldrow Glacier 1956 [*Harrison*, 1964] and Bruarjökull 1963-1964 [*Thorarinsson*, 1969]) or stop in winter (Black Rapids 1936-1937 [*Hance*, 1937]). Also, fast moving tidewater glaciers may continue their fast motion through the summer [*Meier and Post*, this issue]. Certainly, future observers should attempt to get more precise information about the timing (and location) of surge initiation and termination. Satellite images may provide a means to do this for the large, surge-type glaciers [*Post et al.*, 1976; *Krimmel*, 1978].

3. PROCESSES OF SURGE BEHAVIOR

3.1. *Basic Mechanism of Surge Motion*

For lack of a more plausible explanation, it has long been assumed that surge speeds are achieved by rapid sliding on a water-lubricated base [*Weertman*, 1962; *Lliboutry*, 1968]. This view is now supported by direct observations. On the surging Variegated Glacier, the deformational motion was measured over a large fraction of the depth in a borehole and across the width on a transverse line. The measured shearing could account for only a small fraction of the total speed [*Kamb et al.*, 1985]. Dramatic wrench faults separating relatively rigidly moving, surging ice from chaotically sheared marginal ice have been observed in a number of surges [*Kamb et al.*, 1985; *Dolgushin and Osipova*, 1975; *Post and LaChapelle*, 1971]. These faults apparently represent the extension of a velocity discontinuity up through the thin marginal ice. The role of water lubrication is demonstrated by the water levels measured in boreholes on Variegated Glacier [*Kamb et al.*, 1985]. During its surge motion, water levels indicated high basal water pressure consistently within the range of 4 to 1.5 bars of overburden. After surge termination, basal water pressures, though highly variable, would drop to as low as 16 bars below overburden. The association of slowdowns of the surge motion with flood discharges in the terminal streams proves that the high water pressure and sliding speed were related to water storage in the glacier [*Kamb et al.*, 1985]. The large slowdowns and the surge terminations were associated with water releases of about 0.1– to 0.2–m water thickness averaged over the surging area [*Humphrey*, 1986].

A pivotal question in the surge mechanism concerns the cause of buildup of stored water and high basal water pressure. Results from dye tracing experiments during and after the surge of Variegated Glacier reveal part of the answer [*Kamb et al.*, 1985; *Brugman*, 1986]. During the surge, the water discharge moved relatively slowly (mean velocity $\sim 0.02\ m\ s^{-1}$) through a large total cross-sectional area ($\sim 200\ m^2$). Because of the low water velocity, the large flow cross section must have been composed of numerous, individual small passageways of millimeter diameter. After termination of the surge, water velocity increased to a value ($\sim 0.7\ m\ s^{-1}$) consistent with flow in a tunnel of several meters diameter. The comparison shows that the surging state is associated with a highly constricted drainage system that impedes discharge of water.

3.2. *Role of Longitudinal Stress Gradients in Surge Motion*

It has been recognized that the high velocity and accompanying high strain rates in surge motion may cause significant longitudinal stress gradients that affect the basal stress distribution [*Budd*, 1975]. During a surge, the motion is almost entirely by sliding, so that the velocity u and the corresponding longitudinal strain rate $\partial u/\partial x$ and deviatoric stress τ_{xx} are nearly constant in a cross section. The balance of forces along the glacier length (x axis) is then expressed approximately as

$$<\tau> \approx + f\rho g h \sin\alpha + 2f\frac{\partial h\tau_{xx}}{\partial x} \approx \tau_s + 2\gamma \qquad (3a)$$

In (3a), $<\tau>$ is average basal shear stress over the bed perimeter, f is the hydraulic radius shape factor, ρ is the mean ice density, h is the centerline ice thickness, and α is the surface slope. This is the same longitudinal force equation used by *Budd* [1975], except the longitudinal gradient term on the right-hand side is here multiplied by f. This modification better represents the action of the longitudinal stresses in the full cross-section shape. The two terms on the right-hand side of (3a) may be referred to as the "slope stress," τ_s, and "gradient stress," 2γ.

Figure 8 shows the stress quantities relevant in (3a) as found for Variegated Glacier at the height of its surge motion. Over most of the surging length, γ was quite small, and longitudinal gradients only slightly affected the basal shear stress. Therefore the ice was largely supported locally by shear stress on the cross-section perimeter.

The local zone around the peak in velocity was one exception clearly evident in Figure 8. Compression below this zone and tension above it produced a negative gradient stress that reduced the shear stress. Interestingly, if the ice rheology were nonlinear to zero stress, γ would theoretically be infinitely negative at a transition from extension to compression because of an infinite effective viscosity at zero strain rate. In reality, the zero strain rate viscosity of ice is noninfinite, and its value, together with the reasonable supposition that τ could not be negative, must place a constraint on the sharpness of the velocity peak.

Large longitudinal gradients presumably existed also at the very upper edge of the surging zone and at the front region below the peak in velocity. Measurements were, unfortunately, too sparse at the upper surge edge to say anything about it; the front zone was measured in great detail and is described by *Raymond et al.* [this issue].

Figure 8 gives a picture smoothed in space and time. Local spatial fluctuations are suggested by the velocity data points. Indeed, the time variation of velocity (Figure 6) is not synchronous over the glacier length (Figure 7). This indicates a complexly evolving space-time variation. Consider the velocity peaks just preceding the major slowdowns as an example. Since these transient peaks last several hours and propagate at about 0.6 to 0.7 km h^{-1}, they will have spatial lengths of a few kilometers, which are similar to the time-averaged peak of Figure 8. The amplitudes of 10 to 20 m d^{-1} are also similar. Thus these transient peaks could have substantial gradient stresses associated with them. Instantaneously there may be significant gradient stresses at many locations, but these average to a mean

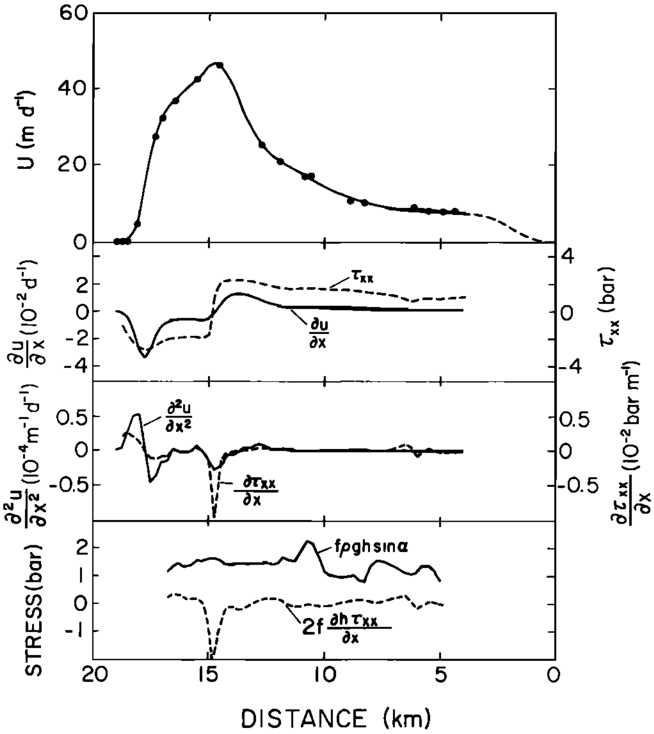

Fig. 8. Variation of time-averaged velocity along the length of Variegated Glacier in June 1983 at the peak of its surge. Longitudinal strain rate and deviatoric stress, their gradients, the gradient stress, and slope stress are also shown. Stress quantities were calculated assuming a nonlinear flow law n = 4.2.

pattern in which the gradient stresses are minor at most locations.

3.3. *Constraints on Sliding Behavior During Surge Motion*

The distribution of velocity $u(x)$ during surge motion of Variegated Glacier (Figure 8) places some constraints on how u would respond to a change in basal shear stress τ. We describe this by $\partial\tau/\partial u$, assuming all other independent variables affecting the sliding are held constant. Although the states of geometry and motion are evolving rapidly with time, the motion is nevertheless in a state of quasi-static equilibrium with zero net forces and no acceleration. This equilibrium state must be stable against all small fluctuations in velocity; that is, a small fluctuation of velocity should result in an alteration in force balance that opposes the fluctuation.

In the absence of a gradient stress γ, it is well known that stability requires $\partial\tau/\partial u > 0$. It is also necessary to account for changes in γ and the consequent effect on the force balance. For example, suppose the velocity distribution $u(x)$ were to be perturbed to $u(x)\,[1 + \varepsilon]$, where $|\varepsilon| << 1$. To first order in ε, the gradient term in (3a) changes from $2\gamma(x)$ to $2\gamma(x)\,[1 + \varepsilon]$. Similarly, the base stress would change from $\tau(x)$ to $\tau(x)\,(1 + \varepsilon u(x)\partial\tau/\partial u)$. From these evaluations and (3a), stability more generally requires

$$\varepsilon \left[u(x)\,\frac{\partial\tau}{\partial u} > 2\gamma(x) = 2f\,\frac{\partial h\tau_{xx}}{\partial x} \right] \tag{3b}$$

In the zone near the velocity peak (kilometers 14 to 18), the velocity profile is concave downward, compression increases downglacier, and $\gamma(x)$ is negative. In this zone, stability against the velocity perturbation would allow an inverse relationship between τ and u (i.e., $\partial\tau/\partial u < 0$).

The zone between kilometers 5 and 14 is more restrictive. There the velocity profile is concave upward, tension increases downglacier, and $\gamma(x)$ is positive. For this profile to exist as a stable state of quasi-static equilibrium, it is necessary that $\partial\tau/\partial u$ be positive. This includes a velocity range from about 5 to 35 $m\ d^{-1}$. In rough numbers, $\partial\tau/\partial u \geq 10^{-2}$ bar $m^{-1}\ d \approx 10^{-4}$ bar $m^{-1}\ yr^{-1}$ or, expressed inversely, $\partial u/\partial\tau \leq 10^2\ m\ d^{-1}$ bars^{-1} $\approx 10^4\ m\ yr^{-1}$ bars^{-1}.

One early, speculative view of surge behavior is that it is a consequence of a multiple-valued relationship between velocity u and base stress τ [*Weertman*, 1964; *Lliboutry*, 1968]. Such a sliding law $\tau(u)$ could have unstable ranges in which τ decreases with increasing u. The above results indicate that the velocity ranges in which shear stress and velocity are inversely related are very narrow or at speeds higher than 35 $m\ d^{-1}$. The rather smooth acceleration of the surge motion in its initial phases (Figure 3) suggests that even narrow unstable velocity ranges are absent.

One example of a nonunique relationship between τ and u is found in lab experiments on sliding of ice over rock slabs [*Barnes et al.*, 1971; *Budd et al.*, 1979]. The experiments show that at a certain level of τu of about 500 bars $m\ yr^{-1}$, there is a transition from a stable friction in which $\partial\tau/\partial u > 0$ to an unstable one for which $\partial\tau/\partial u < 0$. This concept can be employed in a numerical ice flow model to simulate many aspects of surges rather realistically [*Budd*, 1975; *Budd and McInnis*, 1974, 1978; *Budd and Smith*, 1986]. However, it does not appear to be consistent with Variegated Glacier, where $\partial\tau/\partial u > 0$ for τu at least up to ~ 1.5 bars $x\ 35\ m\ d^{-1} \sim 2\ x\ 10^4$ bars $m\ yr^{-1}$.

A number of proposed sliding laws relate sliding velocity u to basal shear stress τ and effective normal stress N (overburden minus water pressure) such that τ increases monotonically with u and N [*Lliboutry*, 1979; *Budd et al.*, 1979; *Jones*, 1979; *Bindschadler*, 1983; *Fowler*, this issue]. To test these sliding laws $\tau(u, N)$ thoroughly and quantitatively would require much more extensive data on basal water pressure than are available.

Qualitatively their forms are, at least, not obviously inconsistent with the behavior found on Variegated Glacier. Furthermore, by the introduction of effective normal stress, they account for a nonunique relationship between u and τ in a natural way and may contain basic ingredients to explain surge behavior. For example, the peaks in velocity both in time average (Figure 8) and the transient propagating ones (Figures 6 and 7) occur in association with low basal shear stress as described above. This does not represent an intrinsic inverse relationship between u and τ but may be explained as a driving of the high speed by high water pressure (i.e., low N).

An additional and equally fundamental ingredient is a means to predict N for various types of drainage systems and the conditions that promote transitions between them. The observations from Variegated Glacier indicate that the distinction between normal (slow) and surge (fast) motion states may be best understood as arising from distinct hydrological regimes of relatively low and high basal water pressures [*Kamb et al.*, 1985]. An example of this approach in an analytical model is presented by *Fowler* [this issue].

3.4. *Initiation and Termination of Surge Motion*

The initiation of surge motion apparently requires some critical condition. The tendency for surges to recur periodically suggests that the geometrical evolution of the glacier has overriding control, and year-to-year fluctuations in external conditions are

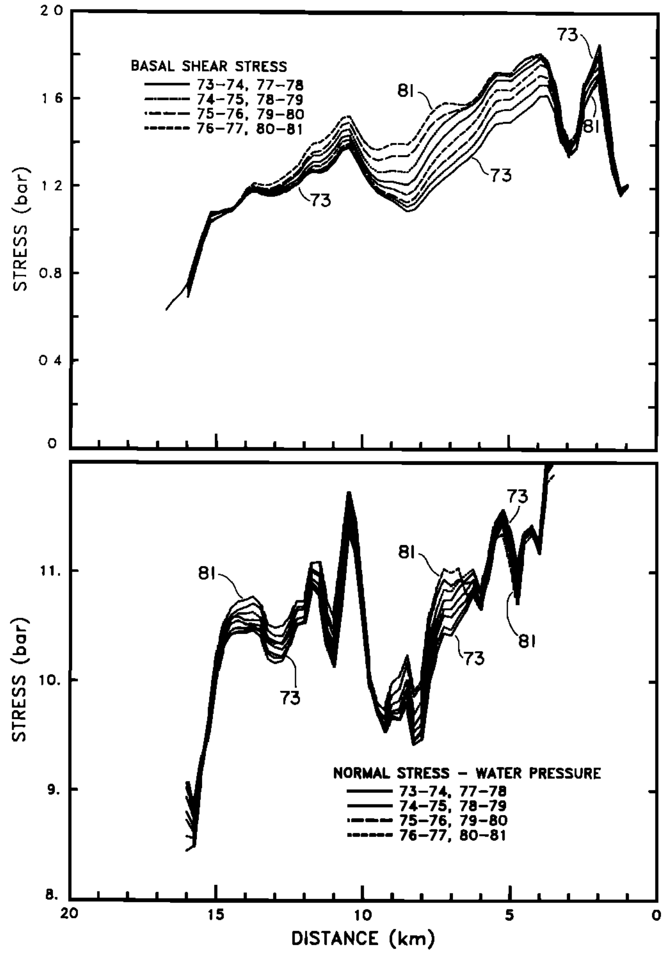

Fig. 9. Evolution of basal shear stress and effective normal stress in Variegated Glacier in the 8 years preceding initiation of its surge in 1982 (from RH86). Effective normal stress is derived theoretically from the theory of *Röthlisberger* [1972] for flow in a tunnel at the base winter discharge.

only secondary [*Post*, 1960]. However, the tendency for a definite phasing of initiation and termination with the seasonal variation of water input indicates the transition between normal and surge flow states is sensitive to short-term effects.

Robin [1969] classified some of the early ideas concerning surge initiation into stress, thermal, and water film instabilities. Although thermal effects may be important in some circumstances [*Clarke*, 1976; *Clarke et al.*, 1984], the existence of temperate surging glaciers shows that this is not an essential feature of surges. Certainly the distribution of stress must play an important enabling role in all surges. However, the hydrological element in the surge process now seems most crucial.

Meier and Post [1969] suggested tentatively that a surge is started when a critical basal shear stress is reached in the lower part of the reservoir area, where the glacier both thickens and steepens. In their own terms, *Dolgushin and Osipova* [1975] use this hypothesis to explain the start of surges of Medvezhiy Glacier. The initiation of surges in the numerical model of *Budd* [1975] is related to a critical value of τu that separates a low-speed regime of stable sliding (τ increases with u) from a high-speed, unstable regime (τ decreases with u). The product τu is related to the rate of basal melting and introduces a hydrological element to the triggering. Although the basal shear stress and sliding speed certainly increase in large areas during the quiescent phase (e.g., Figure 9) and probably play an important role in surge initiation, the above hypotheses are incomplete. This is clear in view of the ability for surge motion to stop and start in essentially the same geometrical condition (as happened in July 1982 and October 1982 on Variegated Glacier), the related seasonal cycle of surge initiation and termination, the large water pressure fluctuations, and the dramatic change in the transmissivity of the basal hydraulic system between surge and postsurge [*Kamb et al.*, 1985]. A full understanding must necessarily involve an explicit treatment of the water flow through the glacier. This was suggested years ago by *Röthlisberger* [1969] and A. Post (unpublished manuscript, 1968), who argued a blocking of the basal drainage system may be the initiating factor in surge motion. A few theories emphasize this aspect of the surge process.

Two very simple views are based on the idea that there are deviations of the local normal stress σ_l from the mean ice overburden σ in the basal zone of a glacier. In the presence of a shear stress, such deviations necessarily exist. The theories of sliding of ice in contact with a rigid rough bed predict that the amplitude of fluctuations in $\sigma_l - \sigma$ is proportional to the basal shear stress τ [*Kamb*, 1970]. The minimum compressive normal stress σ_{min} occurs on the downstream side of bedrock bumps, such that $\Delta\sigma = \sigma - \sigma_{min} \approx \tau/\zeta$, where ζ is related to bed roughness.

Robin and Weertman [1973] proposed that a surge is triggered by a "pressure dam" that blocks water drainage along the bed. Their theory assumes water flows in a linked cavity network and the water pressure is constrained by the overlying ice to be equal to σ_{min} defined above. The pressure dam arises when there is a strong downglacier decrease in τ. This condition yields a downglacier increase in σ_{min} and the corresponding water pressure, which opposes downglacier drainage. The critical negative gradient in τ_b for complete blockage of the water flow depends on the surface and bed slopes and bed roughness. In opposition to the assumption of the theory, the measurements of basal water pressure in boreholes invariably show that large fluctuations in water pressure occur at a given location [*Engelhardt et al.*, 1978; *Hodge*, 1979; *Kamb and Engelhardt*, 1987; *Iken and Bindschadler*, 1986], presumably because the basal water pressure is affected not only by the local ice stress conditions but also by distant inputs to the longitudinally coupled drainage system. Furthermore, a surge may start at a negative shear stress gradient much smaller than required by the Robin/Weertman theory [*Bindschadler et al.*, 1977]. This theory gives no consideration to the seasonal input of water and its effect on surge initiation.

Motivated by the recognition of the seasonal cycle of surge initiation on Variegated Glacier, C. F. Raymond and W. D. Harrison (RH86) suggested a simple mechanism that is parallel to explanations of the seasonal variation of velocity commonly found on normal glaciers [*Röhlisberger*, 1972; *Iken et al.*, 1983]. Their hypothesis concerns the comparison of water pressure p in major tunnels along the glacier axis [*Röthlisberger*, 1972] to the minimum compressive normal stress at the bed σ_{min} described above. If

$$\Delta p \equiv \sigma - p < \Delta\sigma \equiv \sigma - \sigma_{min} \approx \tau/\zeta \tag{4}$$

then water may be pumped from the tunnels to the bed. This circumstance is identified as the condition for bed separation in the theories of glacier sliding [*Lliboutry*, 1968; *Kamb*, 1970], and the parameter $\zeta = \tau/\Delta p$ is referred to as the separation index [*Bindschadler*, 1983].

Two distinct circumstances can produce high pressure p in tunnels [*Röthlisberger*, 1972], possibly sufficient to satisfy the separation condition and drive water out of the tunnels to the bed. The first may arise by a transient, high water input rate from the surface as occurs in the early melt season or storms. Because of an oversupply of water from the upper surface, a tunnel may enlarge by melting and drop p, even though water is pumped away from it to the bed. The second arises as a result of an inverse relationship between p and water discharge required by a balance of melt opening and creep closure in steady state. For example, in winter a low steady state discharge with little surface input can give high p. In this case, tunnels must collapse by a reinforcing feedback loop of decreasing discharge and rising water pressure once the separation condition is reached and water leaks from the tunnels.

The RH86 hypothesis applied to Variegated Glacier presumed that a major drainage tunnel existed along the glacier length at the height of the melt season in all years. A surge was initiated when this tunnel collapsed by the above mechanism at a discharge somewhat greater than the base winter discharge late in the melt season, thus trapping water in the glacier and ground that was later redistributed to the bed in the winter. The first initiation of surge motion in winter 1981-1982, reinitiation in winter 1982-1983, and failure to reinitiate in winter 1983-1984 could be explained in terms of the evolving distributions of basal shear stress and theoretical tunnel pressure distributions (Figure 9), assuming the separation parameter $\zeta \approx 0.15$. Similarly, this hypothesis indicates once a tunnel exists in the summer, its stability at high discharge and low pressure prevents collapse and surge initiation. The possibility of initiation must await a time of low discharge.

Although the condition for surge initiation described by Raymond and Harrison may be a necessary one, it is certainly simplified and not sufficient. Major questions concerning how water flows in a distributed system of basal cavities or other passages and how this water affects sliding need to be addressed [*Kamb*, this issue; *Fowler*, this issue].

Very little theoretical attention has been given to the processes that terminate a surge. The surge of Variegated Glacier was terminated by the downglacier propagation of stopping fronts and the related discharge of large floods in the stream (e.g., Figure 7). This dramatic mode of deceleration has not been explained. The formation of efficient drainage passage ways and release of water are central, but the means are unclear. It is worthwhile to note that initiation and termination are not separate issues. The first phase of the surge terminated in summer 1982, but a second phase could restart in the following winter. Similarly, the final termination of the surge episode was not only by the propagation of a stopping front in summer of 1983, but also by reaching a condition that prevented restart of fast motion the following winter.

Clarke et al. [1984] have proposed the outlines of a theory to explain initiation and termination of surges along somewhat different lines. Their theory emphasizes flow in a porous substrate undergoing deformation. The hydraulic transmissivity of the system will depend on a competition between consolidation, which drops the transmissivity, and shear-induced dilatancy or piping, which increases transmissivity. These may interact in a complex way that could lead to sealing of the flow system, trapping of water, and surge initiation, or the opposite. In their conceptual model the mechanism of surge motion involves failure of the bed material at high pore pressure, which is also affected rather directly by consolidation and dilatancy even without water flow. This theory has been developed further by *Clarke* [this issue].

4. UNRESOLVED QUESTIONS

4.1. *Bed Structure, Sliding Process, and Basal Water Flow*

The central questions in understanding surge behavior are the "basal sliding law" (relationship between $u, \tau, N, ...$) and the basal hydraulic system (relationship between u, τ, N, water inputs,...). The discussion in this paper has focused on the phenomenological aspects of these problems. A satisfying understanding of surges must be based on physical theories of these processes. A major impediment to progress is the lack of a clear picture of the structure of the bases of surge-type glaciers (or glaciers in general). Two distinct end-member views have emerged, the "hard" and "soft" beds.

The most highly developed models of glacier sliding are based on the view that sliding occurs by slip on a discrete interface between ice and rough rigid bedrock, a "hard bed" [*Weertman*, 1957; *Nye*, 1969; *Kamb*, 1970]. *Weertman* [1964] proposed that the high speed of surges may be caused by smoothing of the bed by accumulation of pressurized water between the ice and rock. This idea has subsequently been more thoroughly analyzed [*Lliboutry*, 1979; *Iken*, 1981], and field evidence for the existence of basal cavities that open and close in relation to changes in sliding speed has been found [*Iken*, 1978; *Iken et al.*, 1983, *Kamb and Engelhardt*, 1987]. However, an important involvement of rock debris in the basal ice and between the ice sole and rock is expected based on observations in boreholes on normal glaciers [*Engelhardt et al.*, 1978] and on Variegated Glacier prior to its surge [*Harrison et al.*, 1986*b*]. Such debris was found to be dispersed into the ice several meters above the ice sole and to lie below the ice sole as an ice-free, active subsole drift about 0.1 m thick or possibly thicker. This appears to be an inescapable complication that is only beginning to receive theoretical attention [*Hallet*, 1981; *Kamb*, 1978].

It has also been proposed that a dominant contribution to sliding may come from deformation of an unlithified "soft bed" [*Boulton and Jones*, 1979] and, perhaps, that a soft bed might be a necessary condition for surge motion [*Clarke et al.*, 1984]. *Jones* [1979] suggested that a water-rich slurry just beneath the ice at the top of an unlithified bed could provide a lubricating layer of low viscosity, giving rapid motions. The observations promoting these suggestions come from the terminal zones of glaciers, where deposition occurs and soft beds are expected to predominate. The existence of an approximately 7-m-thick layer of apparently unconsolidated, water-saturated material discovered at the bed of ice stream B deep in the interior of West Antarctica [*Blankenship et al.*, this issue] provides some strong support to the extensive existence of soft beds and their role in fast sliding. However, it is not clear that similar conditions would exist in the reservoir areas of surge-type glaciers, especially the high parts, where, for example, the recent surge of Variegated Glacier started.

The possibility that the fast motion of surges involves concentrated deformation or faulting in the basal ice must also be considered. This was suggested by *Dolgushin and Osipova* [1975] based on observations of apparent slip surfaces exposed in ice walls cut by rivers traversing the lower part of the surged ice of Medvezhiy Glacier. Because there is always a possibility that boreholes do not reach the very bottom of a glacier [*Engelhardt et al.*, 1978], even the borehole measurements from the surging Variegated Glacier cannot exclude the existence of shearing in a zone above the bed. Internal sliding on discrete intra-ice, debris-rich surfaces is known to occur in some circumstances on nonsurging glaciers. In view of the high basal water pressure (often within 2 bars of overburden) and large deviatoric stress (sometimes exceeding 2 bars), hydraulic fracturing of the basal ice is a distinct possibility. The deposits of some surge-type glaciers indicate that bottom crevasses can form [*Clarke et al.*, 1984; *Sharp*, 1985].

The structure of the basal zone is also crucial to the formulation of models of water flow along the base of a glacier, whether this occurs in tunnels melted in the ice [*Röthlisberger*, 1972], in a distributed layerlike network along the bed [*Weertman*, 1972; *Kamb*, this issue] or in permeable bed material [*Boulton and Jones*, 1979; *Clarke et al.*, 1984].

The rapid changes in sediment discharge associated with minisurges [*Humphrey et al.*, 1986], the huge amount of rock debris discharged from Variegated Glacier during its surge [*Humphrey*, 1986], and discharges of muddy water observed during surges of other glaciers [*Thorarinsson*, 1969] indicate a dynamic interaction between the moving ice and the bed that is certainly more complex than the most simple model of clean ice on rigid rock with water passageways at the interface. On the other hand, the hydrological observations also appear inconsistent with water flow through a deformable debris layer. A more probable situation is a mixed type of bed with undeformable humps projecting through debris-rich zones.

The large void space introduced into the ice by surge motion also indicates a less than simple structural picture for the ice thickness. The borehole measurements on Variegated Glacier indicate this does not have a large mechanical significance for most of the thickness. However, it is a problem for calculation of both normal and shear components of basal stress because it introduces uncertainty in the mean glacier density. This becomes significant for estimating the small differences between overburden and water pressure.

4.2. *Surges and Other Fast Flow*

Uncertainty in our knowledge of basal structure and processes extends to the interesting question of the relationship between surges and continuous fast motion in ice streams and tidewater glaciers. Are normal, surge-type and continuous fast glaciers simply an expression of two distinct flow states (slow and fast) with universal applicability? This would be quite appealing, since the three types of behavior (slow, pulsing, fast) could be tied together by a common mechanistic underpinning and distinguished primarily by time-averaged ice flux required by mass balance (low, intermediate, high), for example, in the model of *Budd* [1975] or *Fowler* [this issue]. However, as we learn more about the variety of bed structures that exist in nature, we are likely to discover more than one mechanism for fast motion in both surging and continuous modes. It seems likely we must deal with both hard and soft beds.

One must also be cautious about extrapolating information about surge behavior gained in temperate and subpolar environments to assess the potential for surges of polar ice masses. This obvious caution is reinforced by the recognition of effects from seasonal meltwater input on timing of surges. Even if there is no major difference in basal structure or the principal processes influencing sliding and water flow, the critical geometries for initiation and termination could be quite different, depending on whether water input into the system is from the surface and fluctuating, as in temperate conditions, or from the bed and relatively constant, as in high polar conditions.

Acknowledgments. Although I have attempted to achieve some breadth of information from a number of glaciers, my personal experience and the scope of available data have led me to draw heavily on results from Variegated Glacier. My involvement in the Variegated Glacier Project was through National Science Foundation grants EAR 7622463, EAR 7919424, DPP 7903942, and DPP 8200725 to the University of Washington. The cooperative effort was also supported by National Science Foundation grants to the University of Alaska and California Institute of Technology under the leadership of my valued colleagues Will Harrison and Barclay Kamb. I want to acknowledge all of the many people who have worked hard in sun, rain, and snow on the glaciers I have mentioned.

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