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Mechanical and hydrologic basis for the rapid motion of a large tidewater glacier
1. Observations

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Abstract. Measurements of glacier flow velocity and basal water pressure at two sites on Columbia Glacier, Alaska, are combined with meteorological and hydrologic data to provide an observational basis for assessing the role of water storage and basal water pressure in the rapid movement of this large glacier. During the period from July 5 to August 31, 1987, coordinated observations were made of glacier surface motion and of water level in five boreholes drilled to (or in one case near to) the glacier bed at two sites, 5 and 12 km from the terminus. Glacier velocities increased downglacier in this reach from about 4 m d^{-1} to about 7 m d^{-1}. Three types of time variation in velocity and other variables were revealed: (1) Diurnal fluctuation in water input/output, borehole water level, and ice velocity (fluctuation amplitude 5 to 8%); (2) Speed-up events in glacier motion (15-30% speed up), lasting about 3 days, and occurring at times of enhanced input of water, in some cases from rain and in others from ice ablation enhanced by strong, warm winds; (3) "Extra-slowdown" events, in which, after a speed-up event, the ice velocity decreased in about 3 days to a level consistently lower than that prior to the speed-up event. All of the time variations in velocity were due, directly or indirectly, to variations in water input to the glacier. The role of basal water in causing the observed glacier motions is interpreted by Kamb et al. (this issue).

1. Introduction

The mechanical basis for the relatively slow, normal flow of glaciers and ice sheets is by now reasonably well understood [*Paterson,* 1981]. In contrast, the cause of the rapid flow that occurs in glacier surges, in tidewater glaciers, and in ice streams within the large ice sheets is very imperfectly known and has therefore become the subject of concentrated research efforts, which were brought to a focus in the 1986 Chapman Conference on fast glacier flow [*Clarke*, 1987]. Rapid flow in grounded tidewater glaciers, which has sometimes been called "continuous surging," is an aspect of the phenomenon that has not been extensively studied. A detailed investigation of Columbia Glacier (large tidewater glacier near Valdez, Alaska) in a multiyear U.S. Geological Survey project by Meier and others [*Post*, 1975; *Meier et al.*, 1980; *Meier et al.*, 1985a, b; *Meier and Post*, 1987; *Krimmel and Vaughn*, 1987; *Walters and Dunlap*, 1987] provided the basis and motivation for the work reported here.

The Columbia Glacier Project furnished extensive data on the surface motion of the glacier and on calving of the terminus, but it did not provide direct evidence of the physical controls responsible for the large observed flow rates, of the order of 3-20 m d^{-1}. These controls operate within or at the base of the ice mass, and to obtain direct evidence for them it is necessary to drill into the glacier. Drilling in Variegated Glacier, Alaska, revealed that basal water pressure played a major role in controlling its surge in 1982-1983 [Kamb et al., 1985] and also in causing the minisurges that occurred prior to the surge [*Kamb and Engelhardt*, 1987]. The present work extends this approach to Columbia Glacier with the objective of ascertaining to what extent basal water pressure controls the rapid motions in a large tidewater glacier and to what extent the basal water pressure is controlled by the internal water budget (input and outflow) of the glacier.

The influence of basal water pressure on glacier motion has been particularly clear in short-term flow velocity fluctuations (surge pulses and mini surges); this suggested that a promising feature for our study of Columbia Glacier was the occurrence of well-marked diurnal and semidiurnal velocity fluctuations and also of occasional large velocity peaks of 2-3 days duration, discovered near the terminus in 1984 by the Columbia Glacier Project [*Meier and Post*, 1987, Figure 2]. For this reason, our approach emphasized frequent measurements of glacier motion and physical variables to detect fluctuations on a semidiurnal timescale and longer.

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2. Columbia Glacier

The terminus of Columbia Glacier, about 30 km west of Valdez, calves into Columbia Bay in Prince William Sound (Figure 1). Icebergs from it occasionally drift into the tanker shipping lanes of Valdez Arm. In 1987 the glacier was about 64 km long; its terminal reach, some 12 km in length, was about 5 km wide (Figure 2). This reach is extremely crevassed, reflecting the high flow speeds of 3-20 m d^{-1} there, as in surging glaciers. From surface elevation and radar sounding data it appears that the glacier is not actually afloat, except locally and temporarily [Meier and Post, 1987; Krimmel and Vaughn, 1987]. The ice of the terminal reach is at the melting point, and its centerline thickness there is mostly in the range of 400-1000 m. The terminus of Columbia Glacier had a long history of stability prior to the early 1970s, when it began to retreat as a result of a calving-instability mechanism [Post, 1975; Meier and Post, 1987]; since then the retreat rate has increased progressively to approximately 1 km yr^{-1} [Krimmel, 1992, Figure 13].

3. Observational Program

Columbia Glacier was studied during the period July 5 to August 31, 1987 (J.D. 186 to 243); all dates are given in Julian days (J.D.). Frequent observations of glacier motion were made at five points on the glacier surface over the longitudinal interval from km 52 to 59 (Figure 2). (The designation "km" refers to a longitudinal centerline

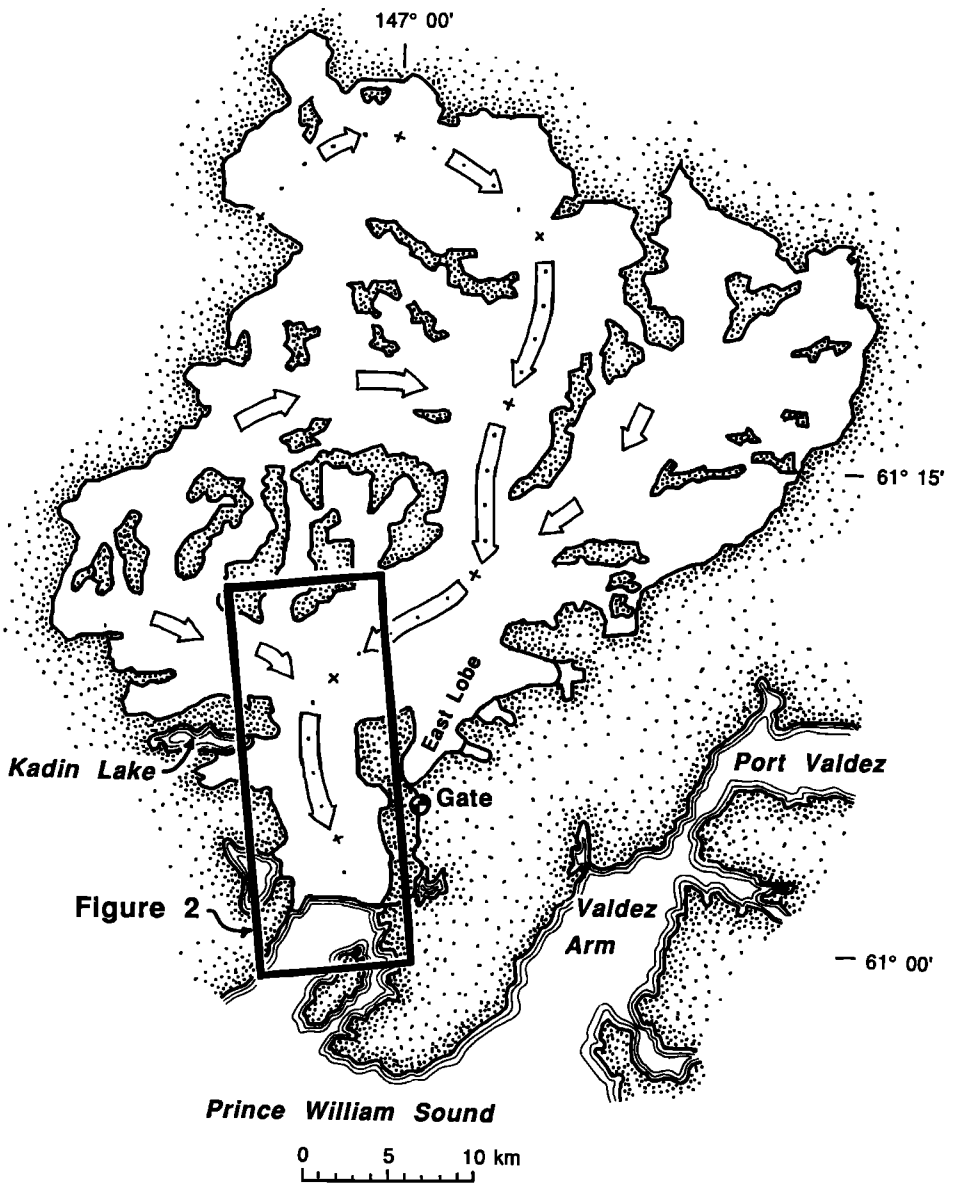
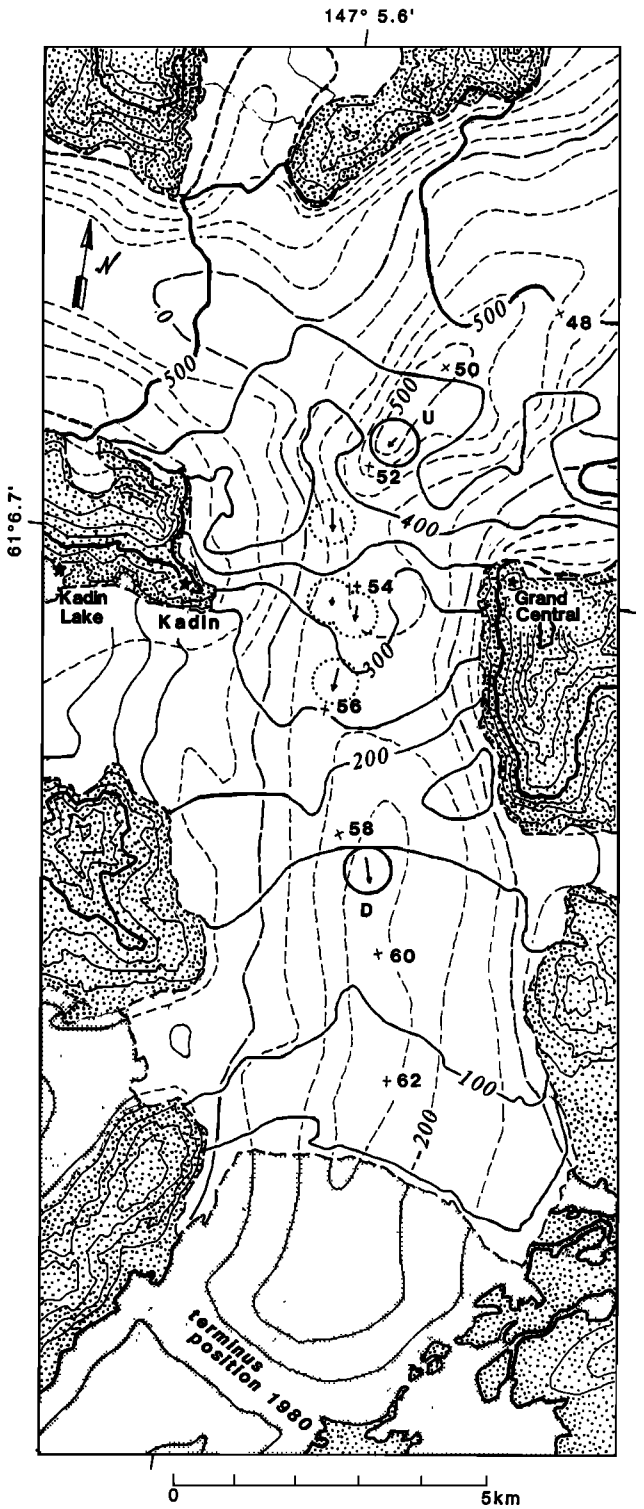


Figure 1. Map of Columbia Glacier showing major tributaries (open arrows, not to scale) and the area of Figure 2, within which the present study was carried out. The calving terminus is in the southwest corner of the map. The longitudinal coordinate system is shown with dots at 2-km intervals and crosses numbered in "km" at 10-km intervals. The map also shows the location of the gauging station "Gate" on Number One River.

coordinate, in kilometers, measured from the head of the glacier, in accordance with the convention used in the Columbia Glacier Project [e.g., *Meier et al.*, 1985b]. This coordinate scale is shown in Figures 1 and 2.) Weather permitting, these observations could detect fluctuations in motion on time scales of as short as an hour or less. Frequent observations of basal water pressure were made in boreholes at km 52 and km 59. Meteorological variables and ablation of the ice surface were recorded as measures of the input of water to the glacier, and water output at an accessible glacier outflow stream was estimated from stream gauge recordings.



Flow velocities. Distances to markers near km 52 and 59, at the borehole sites (see below), were measured from Kadin (Figure 2) every 10 min, weather permitting, using two electronic distance meters (EDM) connected to an automated control and data acquisition system. The motion of these two markers was almost directly toward (in the case of km 52) or away (in the case of km 59) from Kadin. Manual repointing of the EDMs was necessary several times a day. Distances were corrected for atmospheric temperature and pressure using data obtained at Kadin. Angles to the two markers were measured with a theodolite approximately every 2 hours, visibility permitting. After removal of a few wildly aberrant values, distance and angle data were converted to displacements in successive time intervals; the standard error per observation is estimated to be 9 mm. The displacement data were smoothed with a cubic spline function that provides the smallest possible mean squared acceleration, consistent with having a mean squared deviation from the observations less than or equal to that of the measuring system (9 mm). This function was then differentiated to obtain velocity. Several breaks in the record were caused by heavy fog or rain.

Angles to markers at approximately km 53, 54, and 55 were measured from Kadin with a theodolite approximately every 2 hours, visibility permitting. The markers were stakes drilled into the ice. Distances from Kadin were measured at the beginning and end of the observation period and before and after each time the stakes were redrilled. Because of melting or cracking of the seracs and the difficulty of finding a helicopter landing site, the stake at km 54 had to be relocated in midseason at a distance from its original position (see Figure 2). The standard error of a single observation due to surveying error is estimated at <0.1 m, but the actual error could well range up to 0.2 m due to the wobbling of stakes in their holes as melting progressed. Displacements were calculated assuming that the stakes moved in a smooth path through the points determined by both angle and distance

Figure 2. Map of the lower reach of Columbia Glacier including the area studied. Surface contours are shown with solid lines (contour interval on glacier 50 m, elsewhere 100 m), bed contours with light dashed lines (contour interval 100 m), and the sea level contour on the bed with a long-dashed line. Exposed land is shown with coarse stipple and water with fine stipple. The longitudinal coordinate system in kilometers is shown with numbered crosses. Bedrock instrumentation stations Kadin, Kadin Lake, and Grand Central (principal survey reference point) are located with stars. The actual Kadin Lake is south and west of the Kadin Lake station, mostly outside the map. The ice motion markers on the glacier are shown with arrows giving their displacements during the 53-day observation period; near km 54 there are two arrows because the marker was moved to the east early in this period. Each arrow is surrounded by a circle (dotted or solid line). The two borehole sites are enclosed in solid line circles labeled U and D. Surface topography is from aerial photographs taken June 26, 1989, terminus position is from aerial photographs taken January 26, 1988, and bed topography is from Rasmussen [1989].

measurements. Darkness, fog, and rain caused frequent breaks in this record. The record for the markers near km 52 and km 59 is better because the EDM was read more often and at night, and because the marker stakes could be tended and reanchored frequently owing to their proximity to the drill sites.

Water input. Water input to the glacier was determined by measurements of precipitation, ice surface lowering (ablation), and the filling rate of an adjacent glacier-dammed lake, supplemented with observations of air temperature and wind speed. Ablation was measured by two specially constructed ablatographs; each consisted of three dowel crosses as floats suspended from a digital water-stage recorder attached to poles set securely in the ice. These were situated south of Kadin in easily accessible ice, but unfortunately were in the wind shadow of the Kadin ridge and adjacent ice fall at times of the several high downglacier wind events. Readings were taken automatically every 15 min, but because of changes in the ice surface structure (e.g., subsurface melting at times of high solar radiation), only values averaged over several hours are considered significant.

The filling of Kadin Lake (Figure 2) also provided a direct measure of fluctuations in water input to Columbia Glacier. A hydrologic study of this glacier-dammed lake, which included separate analyses of ice melt, snow melt, precipitation, and change in water volume, indicated that outflow from the lake was either zero or very small and constant during the period of observations [*Stone*, 1988]. Thus the rate of change of volume could be used as a measure of water input to this drainage basin and, by analogy, of the water input to the adjacent surface of Columbia Glacier. The stage of Kadin Lake was observed with a pressure transducer connected to a data logger (Figure 2). Although useful in a relative sense, the Kadin Lake filling rate could not be used as an absolute measure of water input per unit area of the glacier surface because Kadin Lake is fed partly by an off-glacier drainage basin and partly from Columbia Glacier; the size of the water collection area on or under the glacier is, however, unknown.

Meteorological data, including precipitation, incident solar radiation, wind speed and direction, and air temperature were taken at Kadin and recorded on a data logger. Additional precipitation data were taken at Kadin Lake and at Gate (Figures 1 and 2).

Water output. This could not be measured directly as the outflow discharges into ocean water at a depth of about 300 m under a calving terminus into a fjord jammed with icebergs. Instead, runoff of Number One River was measured where it emerges from the terminus of the East Lobe of Columbia Glacier at Gate (Figure 1). Water stage was observed with a pressure transducer connected to a data logger. The stage-discharge relation was established using the fluorescent dye dilution method. Although only four discharge measurements at differing stages could be made, the channel (bedrock constriction above a waterfall) and thus the rating curve appeared to be stable. Unfortunately, the hydrologic characteristics of this stream are not identical with those of the outflow from the trunk glacier; the drainage basin of Number One River contains more nonglacial terrain and less greatly crevassed ice and does not extend to as high an altitude as the basin of the main glacier. Also, the division of area between the two basins is unknown.

Boreholes. Two sites for borehole drilling (Figure 2) were found by helicopter reconnaissance. Extreme crevassing made finding appropriate sites difficult. Site U (for "upglacier") was at approximately km 52, at an elevation of 420 m. It was in a small band of relatively moderate crevassing within the broad confluence zone between the trunk glacier (coming from the northeast) and the western tributary. The location was about 2 km upstream from a small (about 60 m high) but prominent icefall that marks the entrance to the terminal reach at the narrows, 4 km wide, between the rock points at Kadin and Grand Central (Figure 2). Site D (for "downglacier") was on a flat-topped serac about 7 km downstream from site U, near km 59, at an elevation of 140 m, in the midst of an impressive serac jumble that extended without interruption from km 53 to the terminus near km 64. Site D is shown in Figure 3. During the period of observation, the longitudinal coordinate of site U varied from km 51.4 to 51.6, and that of site D varied from km 58.5 to 58.9, as did the velocity markers at these sites.

Boreholes were drilled by the hot-water drilling method with equipment developed by *Taylor* [1984]. Water for drilling was pumped from nearby crevasses. After completion, each hole was reamed to a diameter of 75 mm.

At site U three boreholes, designated U-1, U-2, and U-3, were drilled and apparently reached the glacier bed as indicated by cessation of drill penetration at depths 945, 974, and 975 m, respectively. Holes U-1, and U-2, drilled on J.D. 185 and 188, were adjacent, 20 m apart; the shallow depth of U-1 therefore suggests that it stopped at a rock embedded in the ice some distance above the bed. Hole U-3, drilled on J.D. 194, was located 280 m downglacier from U-1 and U-2.

At site D two boreholes, D-1 and D-2, were drilled on J.D. 205 and 206. They were located on the same serac, 5 m apart, and were drilled to apparent bottom at depths 526 and 527 m.

Basal water pressure. Following the technique pioneered by *Hodge* [1976] and by *Röthlisberger et al.* [1979], and used successfully on Blue Glacier [*Engelhardt et al.*, 1978] and Variegated Glacier [*Kamb et al.*, 1985; *Kamb and Engelhardt*, 1987], basal water pressure was measured via the stand of water in boreholes. The recorded pressure is here reported as depth of the borehole water level below the ice surface. It can be expressed as basal water pressure in bars by multiplying the water level depth by 0.1 bars m^{-1} and subtracting from 95 bars (at site U) or 51 bars (at site D). The estimated measurement accuracy is 0.2 m in water level depth or 0.02 bars in variations in basal water pressure. As discussed in section 7, special considerations are needed in interpreting borehole water levels in terms of water pressure in the basal water system.

Basal materials and motions. A penetrometer and small core sampler were used to ascertain the presence and thickness of unconsolidated sediment or rock debris at the glacier bed. Shearing motion of the base of the ice over deformable debris was detected from the bending of a drill stem that penetrated debris at the bottom of borehole U-3. The results of this work are presented in a separate paper [*Humphrey et al.*, 1993].

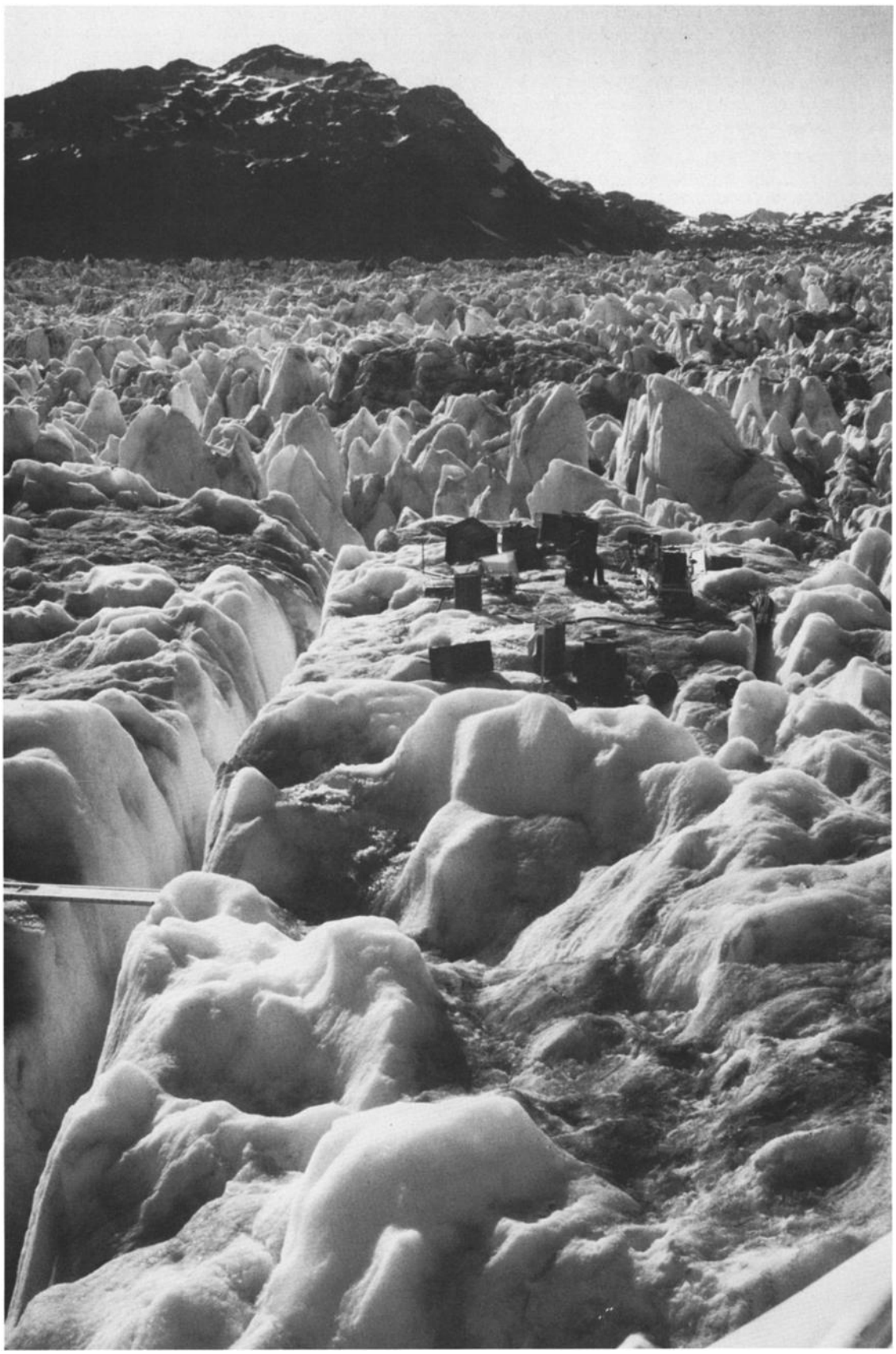


Figure 3. View of drill site D, showing drilling equipment on a small flat-topped serac, surrounded by greatly crevassed ice. View is northeast, from a low-flying helicopter. The ladder bridging the crevasse at lower left was the access route to the camp, on an adjacent serac.

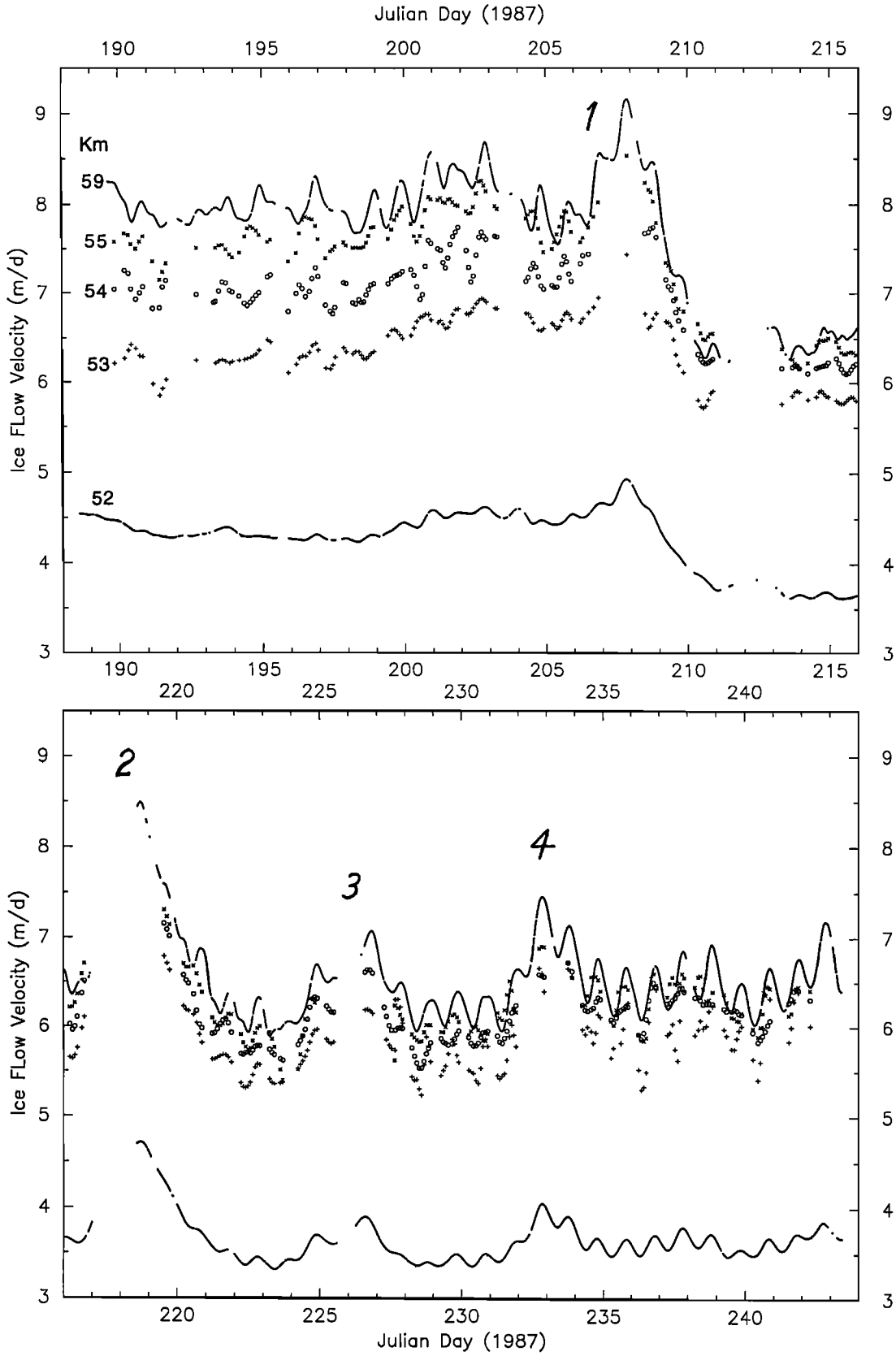


Figure 4. Ice flow velocity versus time for the five ice motion markers, through the period of observation. The data for markers near km 52 and km 59 are shown with discontinuous lines, broken by intervals of data interruption due to clouds or rain. The much less frequent measurements for the other markers are shown as individual data points as follows: near km 53, pluses; near km 54, open circles; near km 55, crosses. Speed-up events are identified by number (1 to 4).

4. General Flow Features of Glacier Flow

Ice velocities at the five markers in the longitudinal interval from km 52 to km 59 over the period J.D. 188 to J.D. 243 are shown in Figure 4. Velocities at km 52 are in the range 3.3-5.0 m d^{-1} and at km 59 in the range 5.9-9.2 m d^{-1}; at the intermediate points the velocities are generally intermediate. The downglacier increase in velocity is mainly associated with the convergence of the flow into the 3.2-km-wide entrance to the terminal reach at km 53, between Kadin and Grand Central (Figure 2); thus the increase from km 52 to 53 is considerably greater than from km 53 all the way to km 59. The velocities are similar to, but somewhat larger than, those observed over 1977-1982 at corresponding locations [*Meier et al.*, 1985b; *Krimmel and Vaughn*, 1987, p. 8965]. The tendency for an overall decrease in velocity over the two-month period is also similar to the earlier observations.

During the period of observation the velocity showed well-defined diurnal fluctuations. These were somewhat irregular early in the season, particularly during rainy weather, but became very regular and pronounced after J.D. 230, especially in the nearly-continuous records at km 52 and 59. After J.D. 230 the average amplitude of the diurnal fluctuation was 0.2 m d^{-1} (5% of the mean flow velocity) at km 52 and 0.6 m d^{-1} (8% of the mean velocity) at km 59. The diurnal fluctuations at the intermediate markers (at km 53, 54, and 55) are less well defined by the data, due to the lack of observations at night and to inaccuracies resulting from stake movement. Thus the occasional lack of synchroneity in the daily velocity pattern for these intermediate markers in Figure 4 may not be real.

5. Flow Speed-Up Events

Superimposed on the diurnal fluctuations are four pronounced peaks, 2-3 days duration, spaced at intervals of 7-11 days (numbered peaks in Figures 4 and 5). We call these "speed-up events." The peaks occurred at both km 52 and 59, synchronously, with amplitude at km 59 scaled up approximately in proportion to the higher general velocity level there (scaling factor of 1.8). The magnitude of the speed up was about 20% in event 1, 30% in event 2, and 15% in events 3 and 4 (Figure 5, curves f and g). Additionally, two or three minor speed-up events are also recognizable in the records.

Events 2 and 3, around J.D. 218 and 226, occurred at times of substantial rainfall, the precipitation totaling about 100 mm in event 2 and about 35 mm in event 3 (Figure 5, curves a). These speed-up events are of the same kind discovered earlier in the Columbia Glacier Project; they are here observed farther upglacier than in the original observations [*Krimmel and Vaughn*, 1987, p. 8966]. Events 1 and 4 do correlate not with rainfall but instead with periods of high wind, as the plots of rainfall and wind speed in Figure 5, curves a and b, show. The strong, warm, foehn-type winds produced enhanced meltwater that substituted for rainwater input to the glacier in generating these speed-up events, which are otherwise similar to the rainfall-induced events. The rainfall peaks and the wind-induced peaks in ice melt are clearly shown in the filling rate of Kadin Lake (Figure 5, curve d), which provides a measure of water input to the glacier in the vicinity of km 55. The wind-enhanced ablation was not registered well by the ablatographs, however, because they were in the wind shadow of the Kadin ridge.

In all instances of rainfall-associated speed-up events, in 1987 and also in 1984 and 1985, the bulk of the rainfall came during the rising part of the velocity peak. For the wind-related events (in 1987) the majority of the wind came during the rising part, but some wind continued in the falling part. High winds on J.D. 242 were accompanied by an increase in velocity, but the observations ended before a complete peak could be observed.

The speed-up events also correlate with water output from the glacier as indicated by peaks in the discharge of Number One River (Figure 5, curve e). The stream flow discharge peaks associated with the speed-up events lag the velocity peaks (Figure 5, curves f and g) by 0.5-1 day. Such a lag may be expected because of the effect of subglacial water storage on the discharge hydrograph.

Rainfall totaling about 50 mm during J.D. 211-212, which resulted in a quite appreciable peak in the Kadin Lake filling rate but only a small peak in river discharge, was associated with only a minor velocity peak, not at all comparable to the speed-ups in events 1-4. (The velocity peak was poorly recorded because of interference of bad weather with the EDM measurements.) Similar instances of incomplete correlation between enhanced water input and speed-up events are seen in the 1984 and 1985 observations of *Krimmel and Vaughn* [1987, pp. 8966 and 8967]. Another instance of incomplete correlation is the low but definite speed-up peak that occurred during J.D. 200-203, unaccompanied by a rainfall peak or high winds. It was, however, accompanied by a pronounced air temperature peak, which was weakly reflected in enhanced ablation (Figure 5, record c).

6. Extra Slowdowns

After speed-up events 1 and 2, the ice velocity dropped to a level consistently lower than before each event. These velocity drops are here termed extra slowdowns. Before event 1, the velocity level was about 4.5 m d^{-1} at km 52 and about 8 m d^{-1} at km 59; after the event the velocity level dropped to about 3.7 m d^{-1} at km 52 and 6.5 m d^{-1} at km 59. Event 2 was followed by a similar but smaller extra slowdown, whose persistence was interrupted by event 3 a few days later. A similar extra slowdown was observed after a rainstorm on August 19-23, 1984 [*Krimmel and Vaughn*, 1987, Figure 6; *Walters and Dunlap*, 1987, p. 8975].

Figure 4 shows that the extra slowdown that accompanied event 1 had a markedly inhomogeneous effect on the velocities of the measured points between km 52 and 59. Although the end points at km 52 and 59 slowed in almost the same proportion, by 18%, the three intervening points slowed proportionately less, so that they all ended up with velocities much closer to the velocity at km 59 than at km 52. The marker at km 53 slowed only 8%, and in fact the difference in velocity between km 52 and km 53 actually increased in the slowdown, from 1.9 to 2.1 m d^{-1}. For a short time at the end of the slowdown, on J.D. 210, km 55 was moving slightly faster than km 59. The slowdown in event 2 did not make a noticeable further change in the relative velocities.

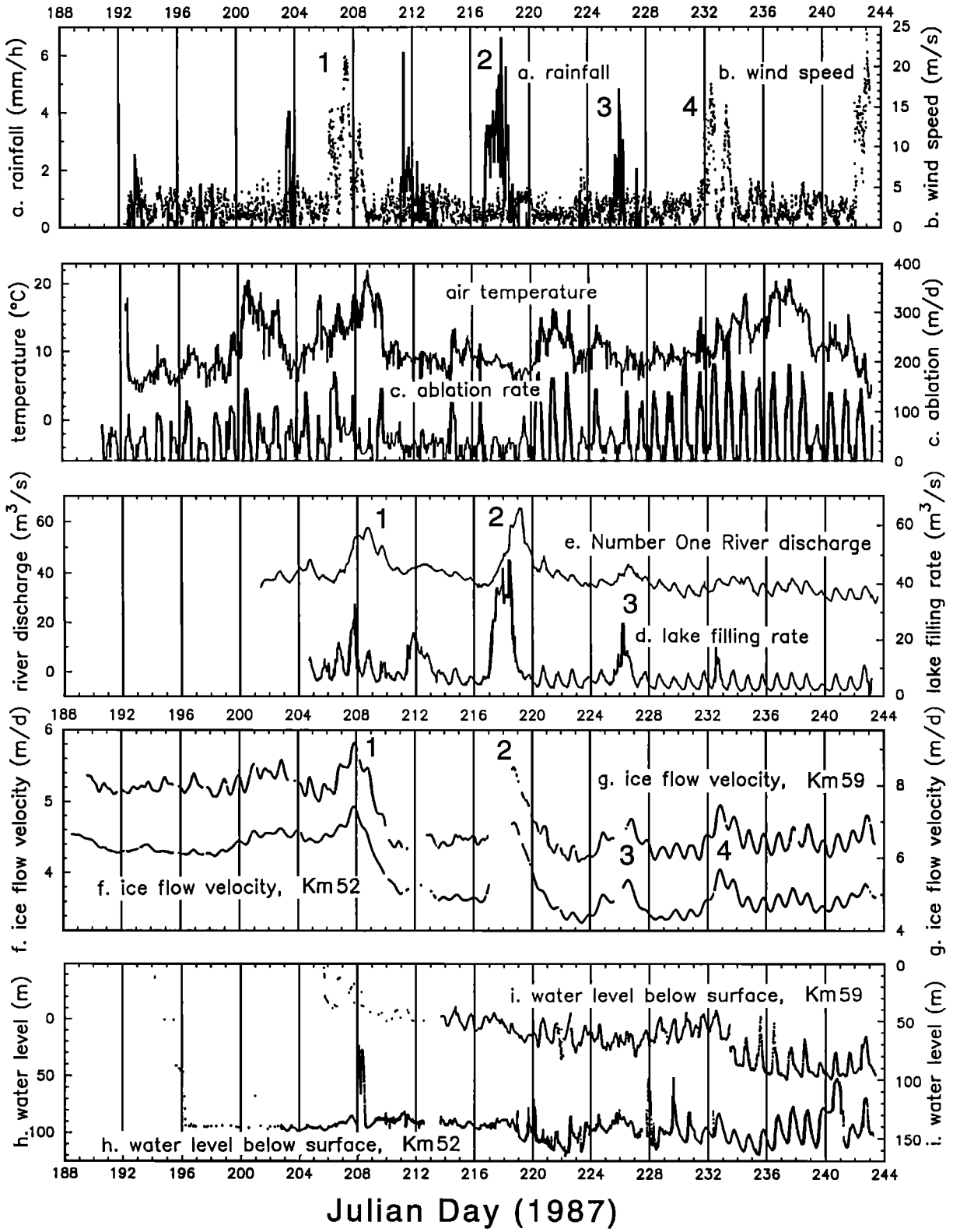


Figure 5. Meteorological, hydrological, and glaciological data for Columbia Glacier through the period of observation in 1987: curve a, rainfall at Kadin; curve b, wind speed at Kadin and air temperature at Kadin; curve c, ice ablation rate on glacier surface south of Kadin; curve d, filling rate of Kadin Lake; curve e, discharge of Number One River at Gate; curve f, glacier flow velocity at km 52; curve g, glacier flow velocity at km 59; curve h, water level depth in borehole U-3 at km 52; and curve i, water level depth at km 59. All variables are plotted as a function of time in Julian days. Dots in the early part of the records h and i are individual water level measurements made with a sounding float; subsequent closely spaced readings were made with pressure transducers. Speed-up events are identified by number (1 to 4).

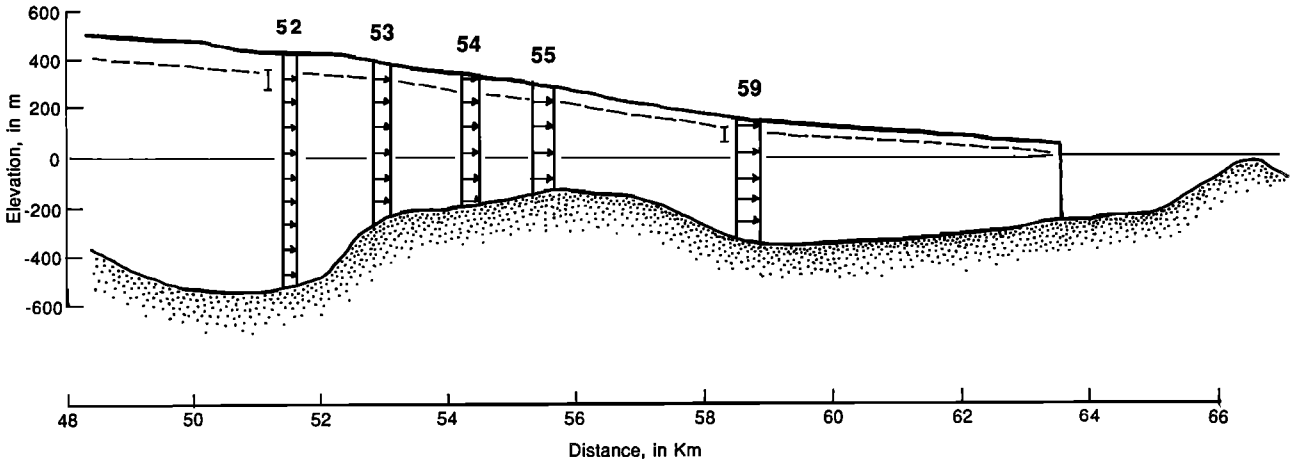


Figure 6. Longitudinal profile of the lower reach of Columbia Glacier showing the ice surface, bedrock (stippled), sea level, and the flotation level (dashed line near the ice surface). Vertical exaggeration is 4:1. The distances traveled by the five ice motion markers during the 53-day observation period are shown by arrows, drawn to represent ice flow velocity on the assumption that all the motion is by basal sliding. The range of water levels observed in the boreholes at km 52 and km 59 is represented by the small "error bars" just to the left of the arrow columns labeled 52 and 59.

7. Borehole Water Levels

Borehole water levels during J.D. 195-243 are shown in Figures 5h and 5i. The water level at km 52 stood, with few exceptions, in the range 80-120 m below the surface, and that at km 59 stood in the range 35-75 m, except after J.D. 233, when it dropped to 50-95 m.

These levels are near the flotation water level, at which the basal water pressure equals the ice overburden pressure. The flotation level, plotted in Figure 6, lies below the ice surface at a depth of about 10% of the ice thickness if the normally assumed 10% difference in the column density of water and glacier ice applies. However, the actual density difference for the specific situations at km 52 and 59 needs to be estimated more carefully because of the effect of crevassing on the bulk density and because an accurate estimate of the flotation level is needed when the observed water level is near flotation. In 1984 a study of crevasse volumes (M. Meier, unpublished data, 1993) was done at km 60.4, 62.2, and 63.7 by means of oblique photography from a helicopter, and it was determined that the crevasse void space corresponded to an effective ice surface lying lower than the mean visible (serac top) surface by 5, 13, and 21 m respectively. The intensity of crevassing increased from 1984 to 1987. We estimate that in 1987 the effective surface lowering at km 59 was 9 m, and at km 52 it was 1 m. The density of ice below 85 m is assumed to follow the curve of density with depth measured in the Byrd borehole, Antarctica [*Gow*, 1970, Figure 4] corrected for thermal expansion from -28.8°C to -0.2°C with an expansion coefficient of 1.5×10^{-4} °C^{-1}; from 85 m to the surface the density is assumed to vary linearly from 0.900 to 0.855 Mg m^{-3}. The resulting column mean densities are 0.912 Mg m^{-3} at km 52 and 0.908 Mg m^{-3} at km 59. Using these mean densities and the crevassing correction given above we estimate a flotation level depth of 89 ± 4 m at km 52 and 57 ± 7m at km 59. Due to the nonparallelism of surface and bed (Figure 6), the flotation level depths decreased to 85 and 55 m, respectively, by the end of the observation period.

Observed water levels at km 52 thus ranged about ± 20 m around the estimated flotation level, with four short excursions up to 50-70 m above flotation. At km 59 prior to J.D. 233 they ranged from about 10 m below flotation to 20 m above, while after J.D. 233 the levels dropped, ranging from about 40 m below flotation to 15 m above.

8. Significance of Observed Water Levels

The extent to which the observed water levels represent "true water levels" that correspond manometrically to the actual pressures in the basal water conduit system depends on how good the hydraulic connection was between each borehole and the basal water system. Observations bearing on this question are the following.

1. In all holes, the water level, which initially stood at or near the ice surface, dropped to near the flotation level at or shortly after the bed was reached in drilling, sometimes even before. The water level drop is seen in Figures 5 (curves h and i) and 7. The drop is usually taken as an indication of connection to the basal water system, although the connection must be via an intraglacial conduit if the drop occurs before the drill reaches the bed. Connection was achieved in hole U-1 even though it probably stopped 30 m short of the bed, and in fact the connection occurred at a drill depth of 785 m, some 190 m above the bed. A similar tendency to early connection was observed in Variegated Glacier in surge. It is quite different from the situation under nonsurging conditions, when often many days pass before a connection is achieved with the basal water system [e.g., *Engelhardt*, 1978, Figures 2, 3; *Kamb and Engelhardt*, 1987, Figures 12-14]. The greater readiness of hydraulic connection in the surging situation is probably due to more pervasive fracturing of the ice and to a more widespread development of the basal

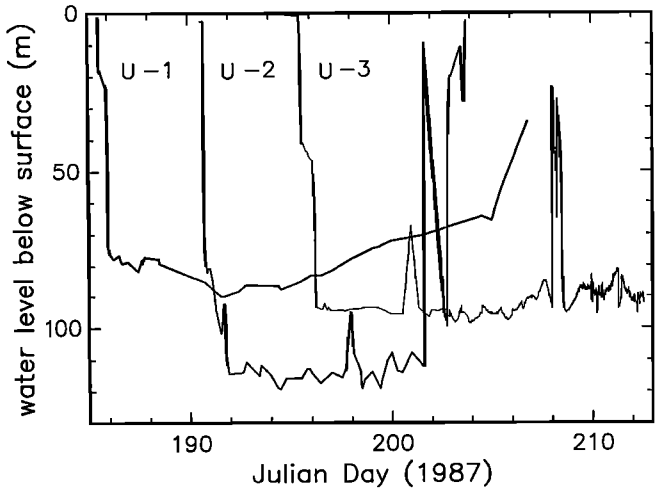


Figure 7. Water levels in boreholes U-1, U-2, and U-3 during the period J.D. 185-213. Measurements in U-1 and U-2, and in U-3 prior to J.D. 202, are by sounding float at irregular times. In U-3 thereafter they are by pressure transducer, giving a continuous record.

conduit system [*Kamb*, 1987]. It may imply that the quality of the connection is generally better in the surging situation. In neither situation are the connections permanent: in Columbia Glacier the connection of holes U-1 and U-2 to the basal system later became sealed off and the holes filled with water (see Figure 7). Holes U-3 and U-2 were remarkable for their long-continued connection (Figure 5, curves h and i).

2. Pumping tests, a recognized method for testing the openness of the hydraulic connection to the basal water system [*Engelhardt*, 1978, p. 43; *Iken and Bindschadler*, 1986, p. 104], were carried out twice in hole U-1, on J.D. 189 and 192. Pumping of water at a rate of 23 L min^{-1} into the hole for several hours caused a water level rise of 6-7 m; upon cessation of pumping the water level went back to its initial position. The rise during pumping indicates detectable hydraulic impedance in the connection to the basal water system.

3. Heavy downflow of water was recognized from its roaring noise heard in boreholes U-3 and D-2. This constitutes a kind of natural pumping test that is a very favorable indication of good hydraulic connection to the basal water system. The fact that the connection in these holes continued through the observation season is perhaps because of the continuing heavy downflow. The actual drop in hydraulic head due to the downflow is not known, but from previous experience we believe that in such a situation of rapid downflow, the connecting passageway from the borehole to the conduit system is enlarged by melting and the drop in head is small [*Engelhardt*, 1978].

4. In hole D-2, observations with a turbidity meter on J.D. 211-212 showed that turbid water was rising from the bed to a height of 184 m above the bottom and exiting there via an intraglacial conduit. The observation of turbid water upflow is an unmistakable indication of connection to the basal water system. It shows that the observed water level at that time was a minimum measure of the basal water pressure. The water level at the time was high, at depth 40 to 50 m below the surface, well above the flotation level (depth 63 m).

5. Simultaneous observations of water levels in the three boreholes at site U are given in Figure 7. Holes U-1 and U-2, 20 m apart, showed water levels of 85-90 m and 110-119 m, respectively, during the period J.D. 192-195. The "true water level" must always lie at or below the lowest simultaneously observed level in nearby boreholes, except in case of turbidity upflow (item 4). Thus the level in hole U-1 was high by about 30 m. This indication of a poor connection from U-1 to the basal water system is consistent with the similar indication from the pumping tests (item 2), with the indication that hole 52-1 did not reach the bed (item 1), and with the rather gradual transition to slow close-off of this hole during the period J.D. 191-205 (Figure 7). The approximately 20 m discrepancy between water levels in U-2 and U-3 during J.D. 196-202 when both holes were operating (Figure 7) cannot be interpreted as indicating a poor connection of U-3 to the basal water system, because the holes were 280 m apart and may therefore have tapped different parts of the basal water system, at different pressures.

From the above observations we conclude that with the exception of U-1 the boreholes had good connections to the basal water system and therefore give a reasonably reliable indication of "true water levels" that correspond manometrically to the basal water pressure. A situation like that in U-1, where the water level was about 30 m high, is unlikely to have affected the other boreholes, except perhaps momentarily as in the spike in U-2 on J.D. 198 and the spike in U-3 on J.D. 201 (Figure 7). It is not possible to put an uncertainty limit on the conclusion that the observed borehole water levels are "true," but in general the limit is probably better than the 30 m error in U-1.

9. Conclusions

The present observations of velocity variation in Columbia Glacier at subseasonal periods, which reveal three types of variation (diurnal fluctuations, speed-up events, and extra-slowdown events) reinforce the results of the original observations near the terminus of Columbia Glacier [*Meier and Post*, 1987; *Krimmel and Vaughn*, 1987] and extend them to a distance of 12 km from the terminus. The diurnal fluctuation observed here, 5-12 km from the terminus, replaces the predominantly semidiurnal, tidally forced fluctuation previously observed near the terminus. The speed-ups are found to occur not only at times of concentrated rainfall, as originally observed, but also at times of strong, warm, foehn-type winds, which cause enhanced production of ablation-generated meltwater. Some speed-ups are immediately followed by an extra slowdown, while others are not; extra slowdowns do not occur separately.

There is little doubt that the speed-ups are caused by the enhanced water input to the glacier. Each extra slowdown is probably a consequence of enlargement of the basal water conduits in the course of carrying away the extra water input that caused the immediately preceding speed-up [*Walters and Dunlap*, 1987, p. 8975], so that enhanced water input events are indirectly responsible for the slowdowns.

At least four of the five boreholes made good hydraulic connections to the basal water system, and in two (boreholes U-3 and D-2) the good connections persisted to the end of the period of observation. The basal water pressures inferred

manometrically from the water levels in these holes are probably reliable to better than the 3-bar error for hole U-1. They indicate consistently high pressures, near flotation and sometimes above flotation. The water levels also show prominent diurnal fluctuations, especially toward the end of the period of observation. Interpretations of the role of basal water pressure and volume in the motion of Columbia Glacier are developed in the companion paper [*Kamb et al.*, this issue].

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