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DEPARTMENT OF NATURAL RESOURCES

DIVISION OF GEOLOGICAL AND GEOPHYSICAL SURVEYS

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Report of Investigations 88-9
RECENT GLACIER-VOLCANO INTERACTIONS
ON MT. REDOUBT, ALASKA

by
Matthew Sturm, Carl S. Benson,
and Peter MacKeith

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RECENT GLACIER-VOLCANO INTERACTIONS ON MT. REDOUBT, ALASKA

by
Matthew Sturm[1], Carl S. Benson[2], and Peter MacKeith[3]

ABSTRACT

Mt. Redoubt, a volcano located west of Cook Inlet, Alaska, erupted most recently from 1966 to 1968. This eruptive cycle removed about 5.7×10^7 m^3 of glacier ice from the upper part of the Drift Glacier and decoupled it from the lower part in a sequence of jökulhlaups that originated in the crater of Mt. Redoubt and flooded the Drift River. The eruptions blanketed the lower part of the glacier with sand and ash, reducing ice ablation. Normal snowfall, augmented by intense avalanching, regenerated the upper part of the glacier by 1976, 8 yr after the last eruption. When the regenerated segment connected with the rest of Drift Glacier, a kinematic wave of thickening ice was triggered, and surface velocities accelerated in the lower part of the glacier. Surface velocities increased by an order of magnitude, and thickening of 70 m or more occurred in the lower part. At the same time, upper parts of the glacier thinned by 70 m. These events advanced the glacier along its lateral margins, and threatened to cause an advance of the terminus. Had this occurred, the Drift River would have been dammed, and an unstable, ice-dammed lake would have been created, posing a catastrophic flood hazard for the oil-tanker terminal located downstream. However, the glacier appears to be returning to pre-eruption equilibrium.

INTRODUCTION

Mt. Redoubt (3,108 m), located 175 km southwest of Anchorage, is one of more than 40 active volcanoes in Alaska. It is an isolated stratovolcano located on the west side of Cook Inlet. Heavy precipitation on the volcano feeds over 10 distinct glaciers that radiate from the summit region. The youthful volcano is not heavily dissected by valleys, and the glaciers form an ice-cap complex described by Vinogradov (1981) as the atrio-valley glacier type. The 'Drift Glacier' (unofficial name), which flows from the Summit Crater, is the largest, most deeply incised glacier on the volcano. It has complex geometry, including steep icefalls over large cliff bands at the confluence of its two main branches. For convenience in this discussion the glacier has been divided into six zones (fig. 1). Intense avalanching and trapping of wind-blown snow in the summit crater and deep canyons through which the Drift Glacier flows create a vigorous system with high ice flux. The Drift River flows past the terminus of the glacier in a narrow gorge with ice on one side and steep rock walls on the other.

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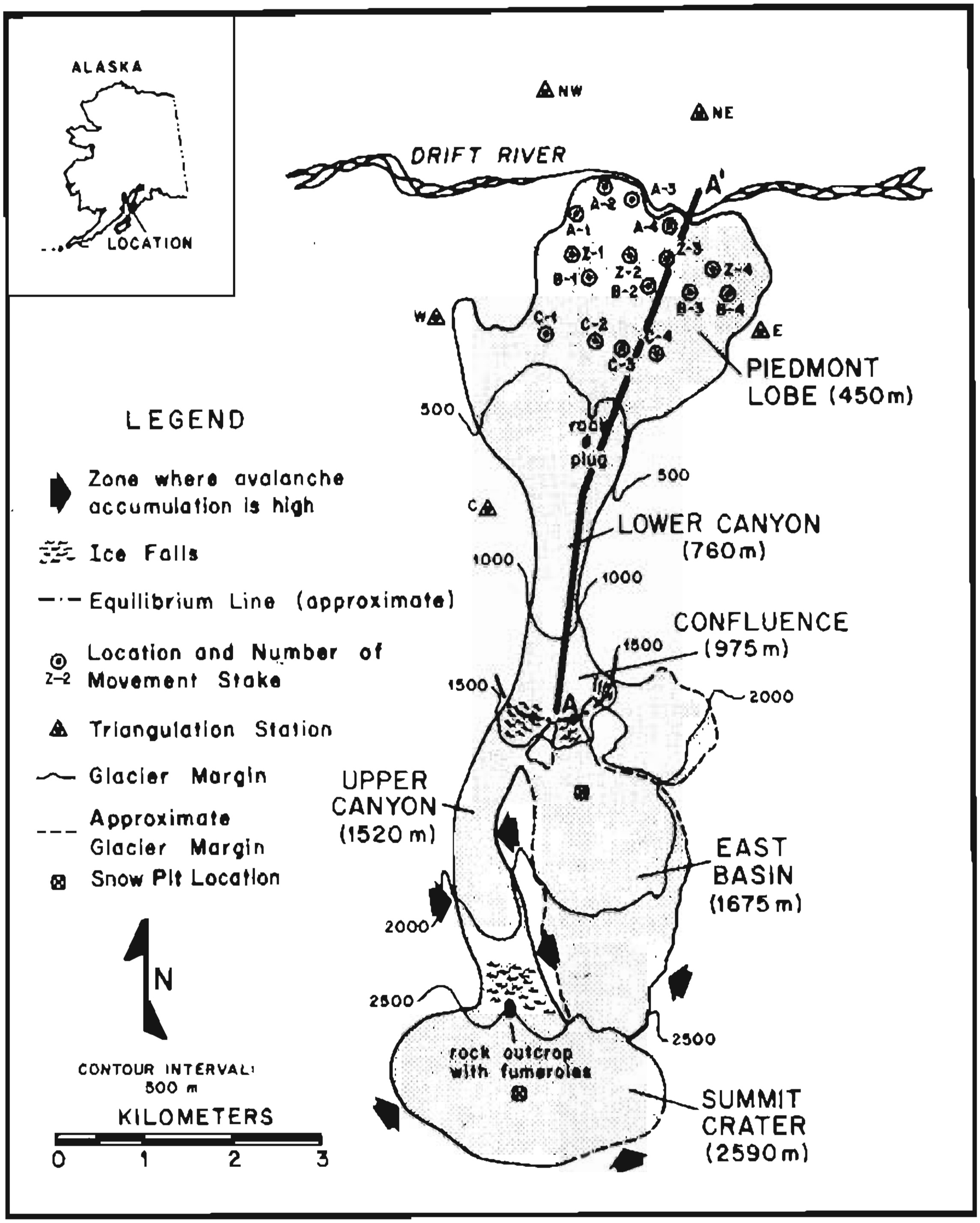


Figure 1. Drift Glacier, Mt. Redoubt, Alaska, showing study area and zones referred to in text. The Drift River flows past the terminus from west to east.

Mt. Redoubt has a history of vigorous eruptive events accompanied by massive lahars. The voluminous lahar deposits in the Crescent River Valley on the south side of Mt. Redoubt were formed about 3,500 yr ago. Prehistoric lahars also built deposits on the north side of Mt. Redoubt in the Drift River floodplain; this is where the most recent (1966-1968) lahars occurred (Riehle and others, 1981). Meltwater from the glaciers on Mt. Redoubt facilitates the formation of lahars, and at the same time the glaciers undergo perturbations during eruptions which may cause dynamic changes which persist for decades before a new equilibrium is established (Sturm and others, 1986). The historic record, probably incomplete, lists eruptions of Mt. Redoubt in 1778, 1819, 1902 (large explosion with voluminous ash ejecta), 1933, and most recently, an eruptive cycle from 1965 to 1968 (Simkin and others, 1981; Riehle and others, 1981).

The most recent eruptive cycle had profound effects on the Drift Glacier in 1966: voluminous steam and ash eruptions produced a jökulhlaup from just below the summit crater (Post and Mayo, 1971). Other jökulhlaups, including a large one on February 4, 1966, occurred between 1966 and 1968 (John Finch, personal commun., 1983). Aerial photographs taken after the jökulhlaups in 1968 show deeply incised gullies, large moulinlike holes, and cauldron-shaped collapse features in the ice. Deltaic sediment deposits originating from ice-tunnel exits can be distinguished clearly in the photographs (fig. 2). Water released during the jökulhlaups apparently traveled over the surface and through tunnels in or under Drift Glacier, spilled out over the terminus, and flooded the Drift River. A jökulhlaup on January 25, 1966 flooded the site on Cook Inlet at the mouth of Drift River where an oil-tanker terminal is now located; this forced the evacuation of a seismic crew (Anchorage Daily News, 1966; Riehle and others, 1981). The flood carried large icebergs and deposited a heavy mantle of sand and ash, over 5 m thick in places, on the Piedmont Lobe of the glacier.

We estimate from examination of aerial photographs that over 5.7×10^7 m^3 of ice was blasted, melted, scoured, or washed away by the cumulative events of 1966-68. Most ice loss occurred in the upper canyon, where large glacier sections were removed and bedrock was exposed; so little ice remained that one branch of the glacier was essentially removed (fig. 2). Large collapse features and deep gullies scoured into the surface of the remaining parts of the glacier suggest that both melting (subglacial and surface) and erosion (the jökulhlaups were heavily charged with sand and debris) were involved. These processes disconnected the lower part of Drift Glacier from ice flowing out of the Upper Canyon and Summit Crater, causing a major reduction in the total flux moving through the Lower Canyon and Piedmont Lobe. A similar phenomenon occurred on the Shoestring Glacier when Mount St. Helens erupted (Brugman and Meier, 1981).

Regeneration began immediately after the eruptive cycle, and the process was essentially completed by 1976. The amount of ice accumulated during this rebuilding stage apparently exceeded that which had been removed by the eruption. When rebuilding was complete and the regenerated glacier connected with the lower part of the glacier, the increased ice flux produced a bulge of thicker ice that moved down through the lower glacier as a kinematic wave at speeds faster than the normal surface-flow speeds. ('Kinematic wave'

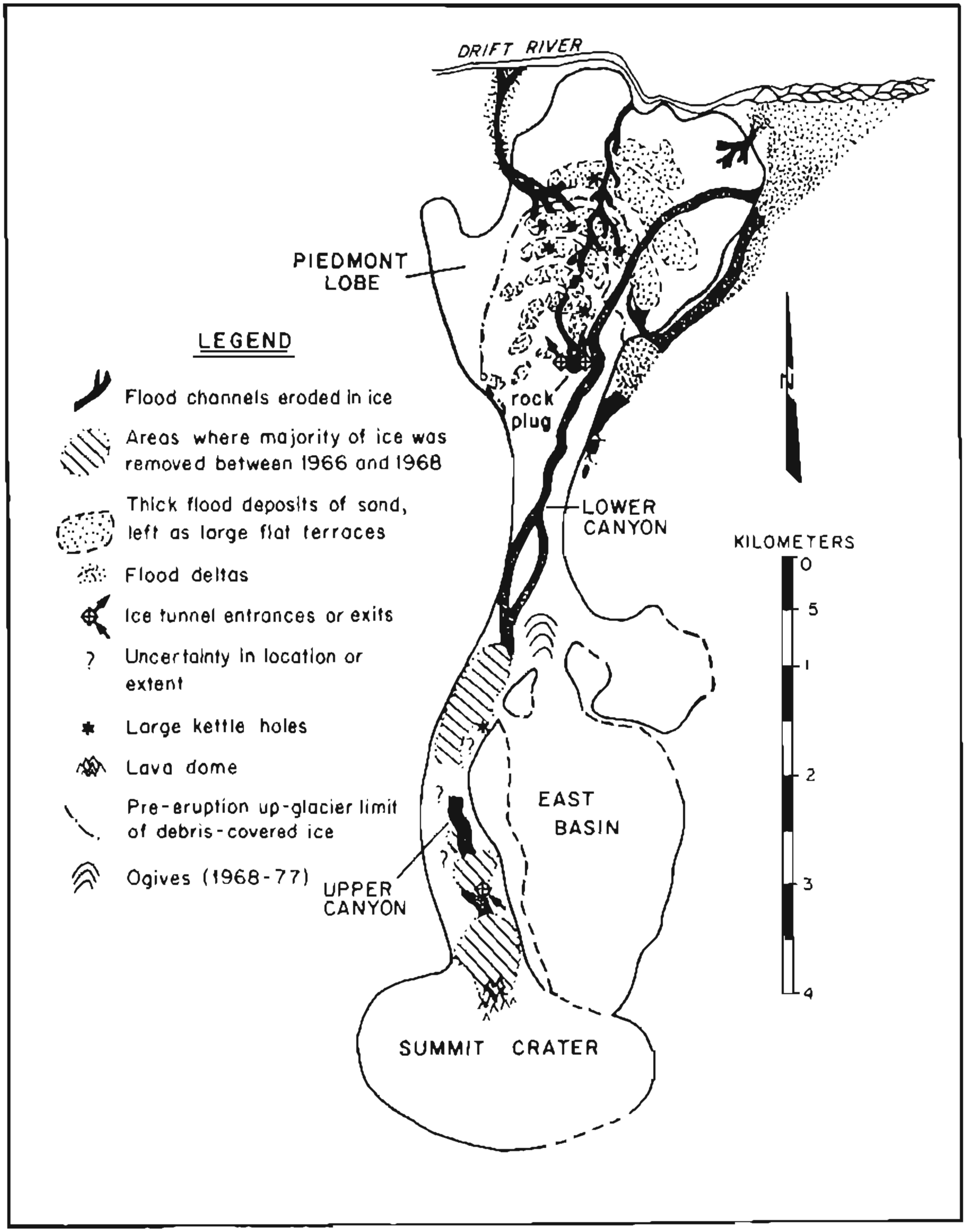


Figure 2. Map compiled from aerial photographs, showing erosional and depositional features left on the Drift Glacier from jökulhlaups and eruptions of Mt. Redoubt between 1966 and 1968.

refers here to the general class of waves that result because changes in flux necessitate changes in thickness and velocity.)

Following the development of kinematic-wave theory by Lighthill and Whitham (1955), there have been a number of theoretical applications to glaciers, beginning with Nye (1960) and summarized by Paterson (1981). Theory has developed more rapidly than field observation, because kinematic waves are difficult to observe on glaciers. Paterson (1981) cited three examples of glacial movement phenomena that appeared to be kinematic waves. The Drift Glacier provides a fourth example.

METHODS

Three methods have been used to investigate the glacier-volcano events on Mt. Redoubt.

Photogrammetry

Ground-control points for aerial photogrammetry were surveyed in 1977 and 1978. Vertical aerial photographs were taken of Drift Glacier in 1977, 1978, 1979, and 1982. Photogrammetric maps at a scale of 1:10,000 and contour interval of 5 m were made from these controlled aerial photographs. From these maps, and from existing USGS maps based on photographs taken in 1954, we prepared longitudinal surface profiles of the glacier accurate to ±2.5 m, and computed recent changes in thickness and volume of the glacier.

Over 120 oblique and vertical aerial photographs of Drift Glacier taken between 1938 and 1984 were examined to compile a recent glacier-volcano history. The photographic history was supplemented by interviews with eye-witnesses of the 1966-68 eruptions of Mt. Redoubt, including people who witnessed the jökulhlaups on Drift River and colleagues who studied the eruption (Wilson and others, 1966; Wilson and Forbes, 1969).

Precision Surveying

A stake network (fig. 1) was established in 1978 so we could measure surface flow on the Piedmont Lobe of Drift Glacier. It was surveyed one to four times a year between 1978 and 1986. The stakes, made of 3-m lengths of pipe set in cairns on the thick supraglacial debris, were placed in four rows (A-D). They required little maintenance because there was little ablation or surface disturbance. The D row was an exception; it survived less than a year because of heavy crevassing and ablation. A fifth row of stakes (Z) was added between the A and B rows in 1982, for more complete coverage. The stakes were located by triangulation. Survey errors in horizontal position have not exceeded ±0.2 m. Errors in vertical position, mainly due to refraction, have not exceeded ±0.5 m.

Snow Pits and Cores

A pit 4 m deep with a core to 10 m was excavated in the Summit Crater in 1977, and in 1985 a second pit with a 26-m core was completed. Pit studies were conducted in 1983 in the East Basin; analysis of these pits and cores, together with the ablation measurements made in 1978, have enabled us to estimate the mass-balance gradient for Drift Glacier.

RESULTS

Chronology

Figures 2, 3, and 4 illustrate melting and erosion of the Drift Glacier in the Upper Canyon during the recent volcanic eruptions and its regeneration in less than 8 yr.

The destructive phase (1966-68) occurred in three main stages:

1. The removal of approximately 5.7×10^7 m^3 of ice eliminated ice flux from the Upper Canyon to the rest of the glacier (figs. 1 and 2) and reduced the flux through the lower glacier by more than half. Some ice that was removed became lodged or buried in the sand and ash covering the Piedmont Lobe and slowly melted out during the next few years, forming kettle holes. Most ice, however, either melted or was washed out to sea in jökulhlaups.

2. Approximately 10^6 m^2 of the Piedmont Lobe, according to estimates from vertical aerial photographs, was covered by flat terraces composed of sand and ash deposited by jökulhlaups and eruptions. These deposits, usually greater than 1 m thick, insulated the ice and eliminated ablation. Similar insulating effects are discussed in Muller and Coulter (1957), Driedger (1981), and Nakawo and Young (1981). Before the debris was deposited, the ablation rate on the Piedmont Lobe was about 5 m/yr, determined on the basis of 1978 measurements of bare ice cliffs at the terminus.

3. The internal and surface-water system of the Drift Glacier was profoundly altered by large channels and tunnels cut during the jökulhlaups (fig. 2). For example, water poured out from the base of the rock plug (figs. 1 and 2) that projects through the lower glacier and carved a new surface channel that persists today as one of the dominant drainages on the Piedmont Lobe.

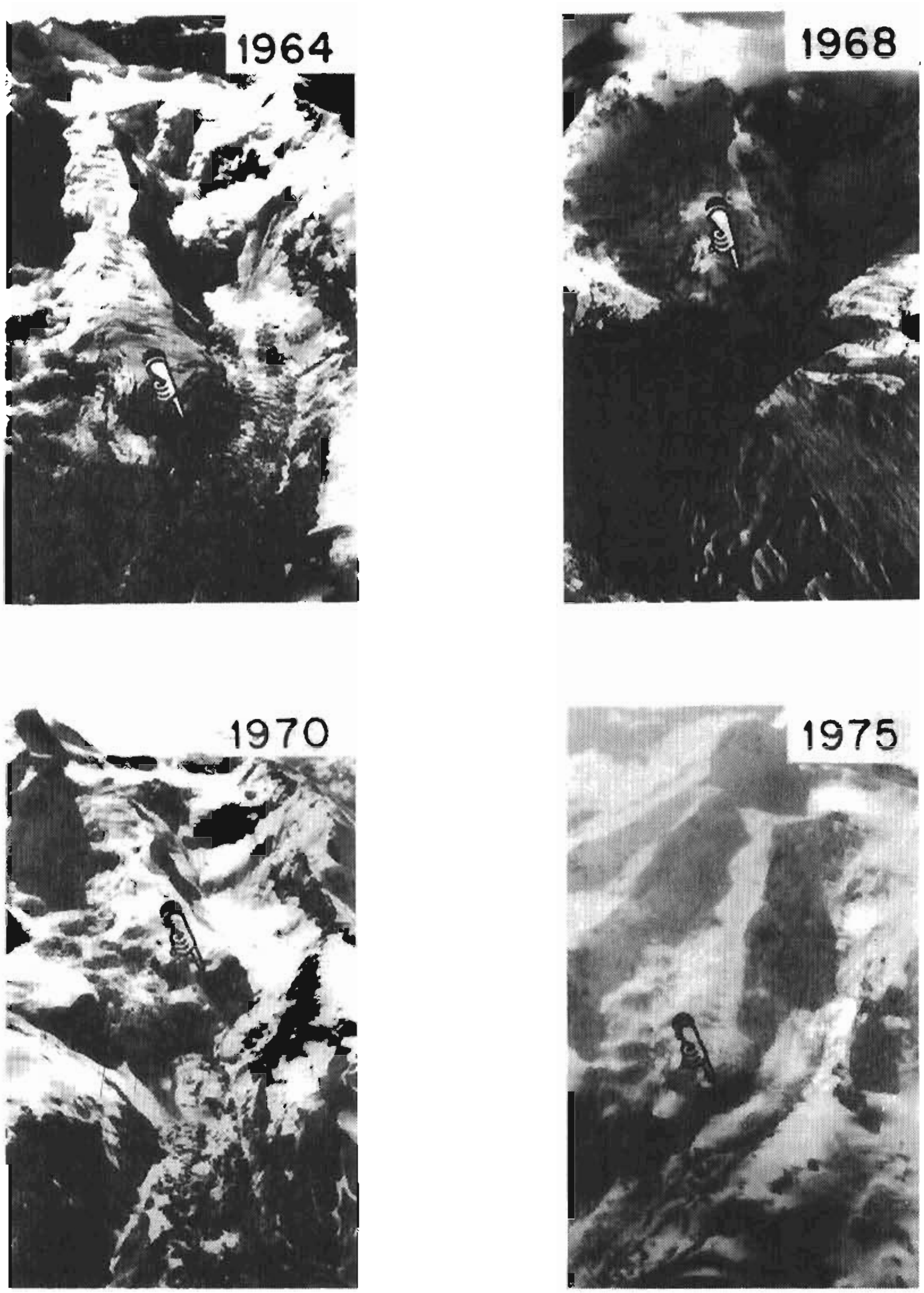
Figures 3 and 4 illustrate how quickly the glacier re-formed in the Upper Canyon and reconnected with its lower segment:

- 1954-64 (figs. 3a and 4a). The condition of the glacier was 'normal,' with heavy crevassing throughout the upper parts, particularly at the confluence of the two main branches. Ogives were absent below the prominent cliffs in the confluence area.

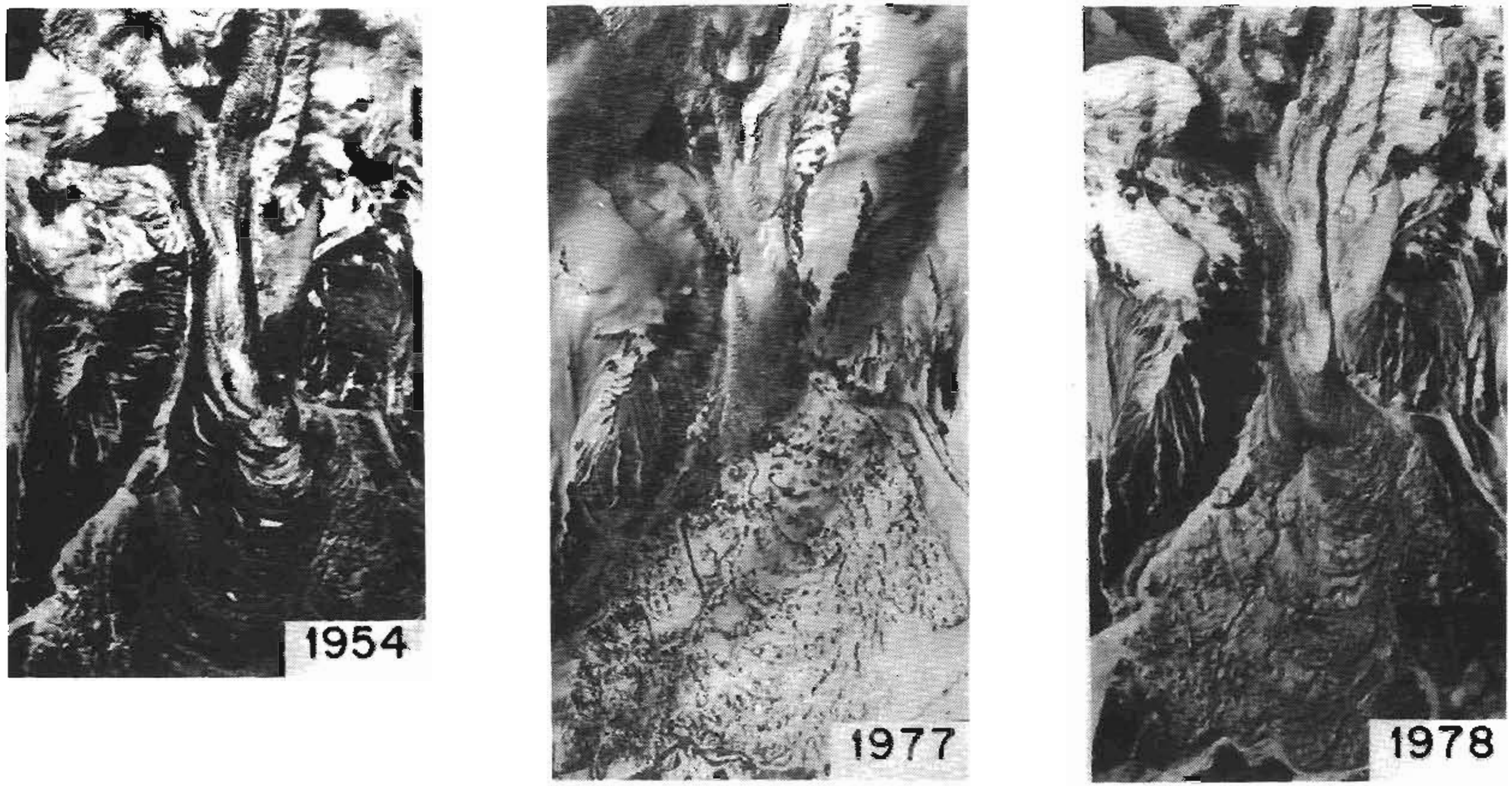
- 1966-68. Eruptions and jökulhlaups occurred.

- 1968 (fig. 3b). The glacier began to re-form in the Upper Canyon.

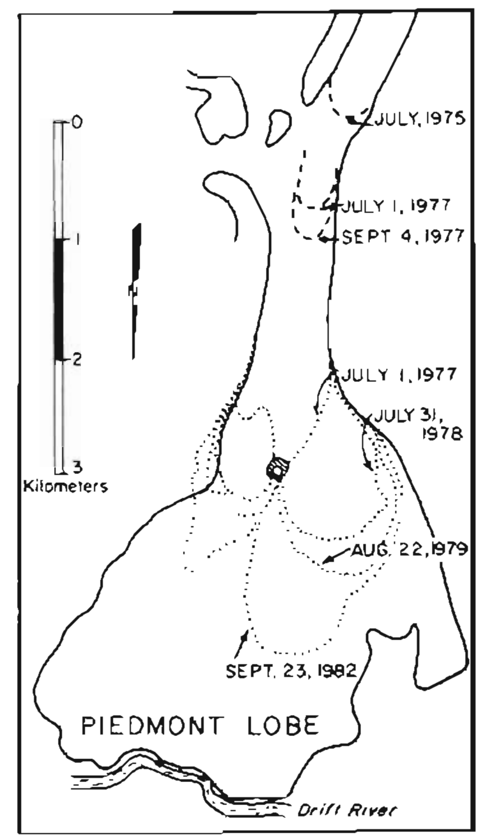
- 1970 (fig. 3c). Avalanching had refilled part of the Upper Canyon, but bedrock was visible at the confluence of the Upper and Lower Canyons.



Figures 3a-d. Oblique aerial photographs taken by Austin Post (1964-70) and authors (1975) showing the destruction and regeneration of the Drift Glacier in the Upper Canyon. The pointer is in the same location in each photo.



Figures 4a-f. Vertical aerial photographs (4a-e) of Lower Canyon and Piedmont Lobe of Drift Glacier, showing pre-eruption condition of the glacier (1954) and subsequent changes caused by propagation of a kinematic wave through the system (1977-82), and a map (4f) of wave front locations between 1975 and 1982. On figure 4f, (below) dashed lines indicate where the front coincided with the terminus of the glacier, which re-formed in the Upper Canyon. Dotted lines indicate the front as manifested by the downglacier limit of new crevassing. Note the rock plug that divides the flow; cross-hatched area of the rock plug was icefree in 1977; the white and black parts of the plug were icefree in 1979, and the black part of the plug was icefree in 1982. Note also the similar appearance of the glacier in 1954 and 1982 photographs.



Figures 4a-f. - Continued.

- 1975 (fig. 3d). The glacier in the Upper Canyon had re-formed. It had a steep, lobe-shaped terminus that appeared to be advancing over bedrock but was separated by 200 m from the main glacier system. Five or six ogives were present below the cliffs on the main glacier in the ice stream from the East Basin.

- July 1977 (fig. 4b). The terminus of the glacier from the Upper Canyon had reconnected and advanced over the glacier in the Lower Canyon. Previously exposed areas of bedrock were covered by crevassed glacier ice. Eight ogives were noted.

- September 1977 (not shown). The lobe-shaped terminus from the Upper Canyon was still recognizable but had advanced 400 m down the Lower Canyon, overriding all the ogives and nearly obliterating them.

- 1978-79 (figs. 4c and 4d). A kinematic wave of thickening, accelerating ice was propagating down the glacier and was accompanied by surface crevassing that progressively disrupted the sand and ash on the glacier surface and made wave movement conspicuous. Ice originating in the Upper Canyon could still be recognized in the combined ice streams of the lower glacier as a zone of clean, white ice.

- 1982 (fig. 4e). The kinematic wave continued to propagate through the lower glacier. Surface conditions and crevasse patterns over much of the glacier appeared similar to those in the 1954 photographs.

Mass Balance

Mass-balance measurements have been difficult to obtain, but are sufficient to estimate the general balance gradient for the glacier. In July, 1977, a snow pit with core to 10 m in the Summit Crater (2,590 m) did not penetrate to the 1976 summer surface; it revealed a minimum water-equivalent accumulation value of 5 m for the year. A snow pit and core to 26 m excavated in the Summit Crater in 1985 suggested an average balance of 6.5 m water-equivalent per yr for the 2 yr of accumulation it penetrated, although there was some difficulty in interpreting the stratigraphy. In 1983, a pit and core in the East Basin at 1,400 m penetrated more than two annual units and indicated an average annual balance of about 0.5 m water-equivalent per yr. In 1978, an ablation rate of 5 m/yr was measured at the terminus on a clean vertical ice face. By using an accumulation value of 6.5 m at the Summit Crater (2,590 m), an ablation rate of 5 m/yr at the terminus (300 m), and estimating an equilibrium-line altitude of 1,200 m from late-summer photography, we obtained a balance gradient of 0.5 m water-equivalent/100 m of altitude. This gradient is in general agreement with those of maritime Alaskan glaciers such as Wolverine and Columbia (Meier and others, 1971; Mayo, 1984), and the accumulation in the Summit Crater is in the same range as that measured on Columbia Glacier at the same altitude (Mayo, 1984).

From 1968 to 1976, when there was no ice flux out of the Upper Canyon, glacier ice rapidly accumulated there, because accumulation rates in both the Upper Canyon and the Summit Crater are high and are augmented by avalanches. Avalanching in the upper part of the canyon had covered much of the denuded area by 1970 (fig. 3c). By 1975 (fig. 3d), a glacier with a lobe-shaped terminus had re-formed in the Upper Canyon and was within 200 m of reconnecting with the rest of Drift Glacier. In 1977, the reconnection was complete, and the regenerated lobe had actually overridden part of the glacier in the Lower Canyon.

Between 1968 and 1976, an estimated 15 x 10^7 m^3 of ice accumulated in the Summit Crater and Upper Canyon, recreating the upper glacier. Ice volume was estimated by applying the balance gradient of 0.5 m water-equivalent per 100 m of altitude to the combined areas of the Summit Crater and Upper Canyon, and summing the 8 years from 1968 to 1976.

Using the same balance gradient to estimate the annual balance flux and balance velocity through the equilibrium line for the entire glacier, and applying the balance gradient to accumulation areas

Summit Crater	2.2 x 10^6 m^2,
East Basin	3.8 x 10^6 m^2,
Upper Canyon	3.4 x 10^6 m^2,

gives an annual ice flux through the equilibrium line of 3.0 x 10^7 m^3. If we assume a deep, parabolic cross section and plug flow, the balance velocity through the Lower Canyon is 450 m/yr.

Dynamics

Curiously (and fortuitously), during the time when ice flux from the Upper Canyon was cut off from the Drift Glacier system, the remaining flow from the East Basin began to form ogives below an icefall at the head of the Lower Canyon. These ogives, which formed only during this period of time, can be seen in figure 4b. In 1977, we identified eight ogives, corresponding to the 8 yr of missing flux from the Upper Canyon. Similarly, six ogives can be seen in the lower left corner of the 1975 photograph (fig. 3d). The ogives, overrun and obliterated by the lobe of ice from the Upper Canyon in 1977-78, have not been noted since. They are a visible manifestation of the different flow regimes that existed in the glacier of the Lower Canyon when there was no ice flux from the Upper Canyon. When flow was from the East Basin only, the ogives indicate an approximate surface velocity of 315 m/yr, if one assumes that they form annually by seasonal variation in flow. This value for the East Basin component is 100 m/yr less than the balance velocity for the entire glacier.

The speed at which the lobe-shaped terminus of the regenerated glacier advanced has been computed from photographs (figs. 4a-e) to be:

Period	Rate of advance (m/yr)
1975-77	450
July-September 1977	1,200
1977-78	1,300
1978-79	520
1979-82	210

These speeds represent three closely related phenomena. Pre-1977 rate is the speed of the terminus advancing over bedrock. Rates during 1977 and 1978, after reconnection to the main glacier, reflect the speed of the frontal lobe of overriding ice. From 1978 on, however, this front was obscure, and the rates reflect the advance of the downglacier edge of new crevassing. The computed speeds were supplemented by survey data that recorded the onset of ice acceleration and thickening as it passed through the stake network on the Piedmont Lobe. The abrupt decrease in speed of the front between 1978 and 1979 corresponds to the time that it moved out of the confined Lower Canyon. The kinematic wave propagated through Drift Glacier at speeds three to five times the average surface velocity of the glacier in the same area.

Survey data from the stake network give a clear picture of the kinematic wave of thickening ice, accompanied by increasing surface velocity, passing through the Piedmont Lobe (fig. 5). Apparently, the rock plug affected wave propagation, because speed and direction of eastern and western segments differed. The rock plug has also served to indicate ice thickening (fig. 4f); before 1979, helicopters could land on it, but in 1982 it was covered by ice seracs, which indicated a thickening of over 50 m.

Surface velocity throughout the Piedmont Lobe was less than 10 m/yr when first measured in 1978. By 1979 the flow had begun to accelerate at the C row of stakes. This acceleration progressed downglacier through the B and A rows at the rate of about 1 row per yr, a distance of about 1 km. Acceleration at the C row began in 1980-81, although stake C-1, which is located in the western component of flow, began to accelerate as early as 1979. Acceleration at the B row began in 1981, and acceleration at the A row began in 1982. Data for the Z row, which was emplaced in 1982, suggest that acceleration began there in 1982. Overall, flow speeds increased by one order of magnitude or more.

The data suggest that by 1986 surface velocities had peaked and begun to decrease at the C and B rows of stakes. This implies that the kinematic wave took about 5 yr to move past the C row of stakes. However, the summer velocity of Drift Glacier is 20-25 percent greater than its winter velocity, and a measurement taken during spring or autumn surveys could include a component of the high summer-velocity component. Inclusion of a greater percentage of summer-velocity component could account for the 1982 velocity peak in B row, but is insufficient to account for the peak in C row, which shows definite deceleration.

The thickening which accompanied acceleration is plotted and profiled on figures 5 and 6, respectively. Maximum thickening, measured by surveying at stake C-4, was over 40 m, although greater thickening (70 m) was measured photogrammetrically at other points on the Piedmont Lobe.

Intense crevassing took place after the onset of thickening and was preceded downglacier by a noticeable warping of the flat sand and ash terraces (fig. 2) into transverse synclinal folds with wavelengths of tens of meters. The crevassing progressively disrupted the terraces. Initially, small longitudinal cracks formed in the surface debris, but within a year or two the glacier surface became a chaotic jumble of crevasses and blue fins of ice.

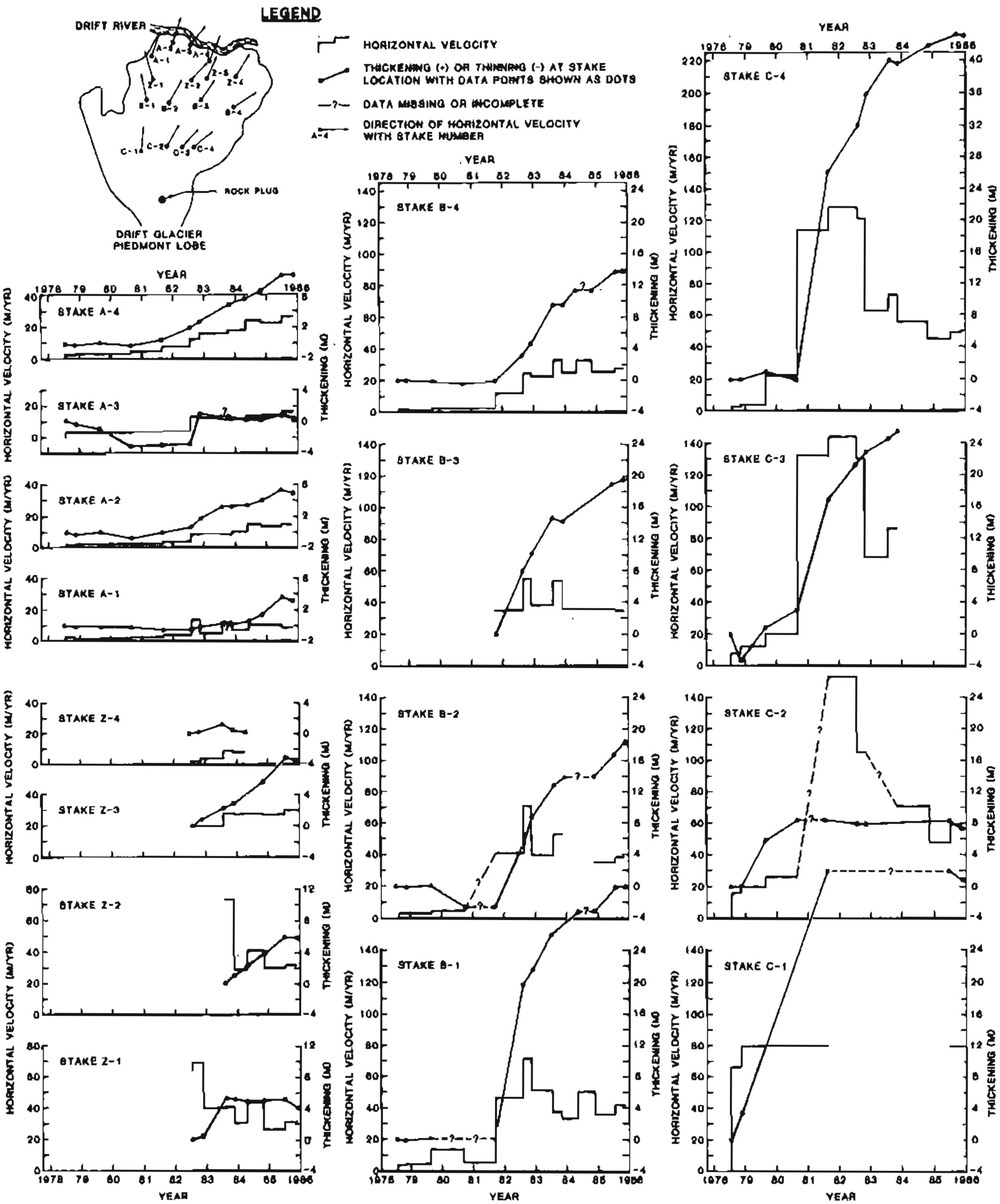


Figure 5. Plots of surface velocity and ice thickening at staked locations on the Piedmont Lobe.

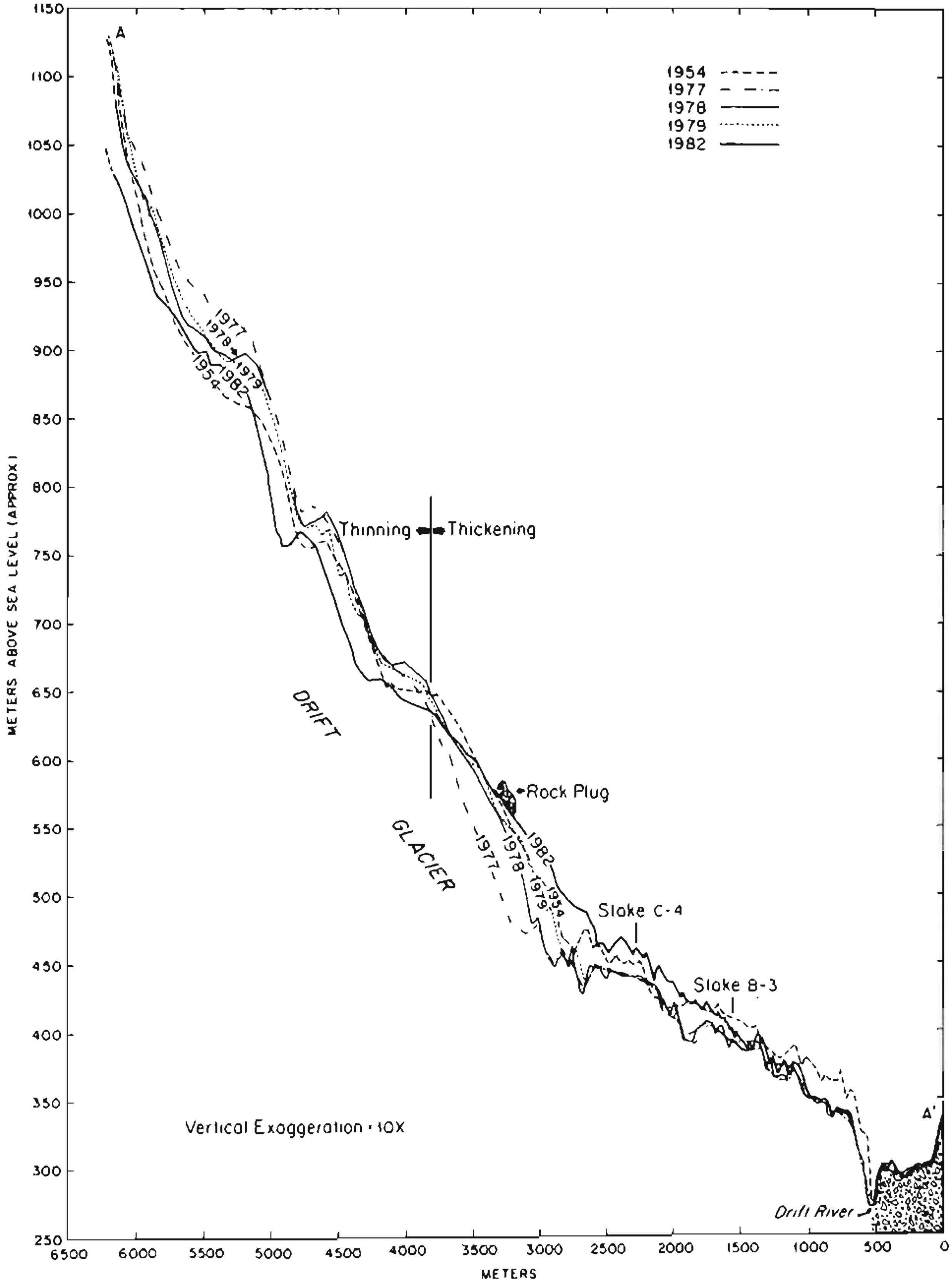


Figure 6. Longitudinal surface profiles of Lower Canyon and Piedmont Lobe of the Drift Glacier.

The debris fell into the crevasse openings and ablation began where debris was removed. The longitudinal and radial crevasses can be seen clearly in figure 4e.

Longitudinal profiles of the glacier for the years 1954, 1977, 1978, 1979, and 1982 (fig. 6) have been extended upglacier to the icefall at the confluence of the East Basin and the Upper Canyon, which is as far as photogrammetric control would allow. The profiles confirm steady ice thickening in the Piedmont Lobe since 1977 but also show thinning in the upper reaches at the confluence of the two glacier branches.

The profile in the Lower Canyon was thickest in 1977 and probably indicates that the kinematic wave was passing through the canyon at that time. Thinning followed rapidly in 1978. By 1982, over 70 m of thinning had taken place at the narrowest part of the canyon, and stranded fringes of glacier ice over 50 m above the glacier surface were noted there, similar to those observed on the Variegated and Muldrow Glaciers after their surges (Harrison, 1964; Post and LaChapelle, 1971; Kamb and others, 1985).

The transition between zones of ice thinning and thickening occurs at the exit of the Lower Canyon, where the glacier begins to spread out on the Piedmont Lobe. This point of zero change in surface elevation has remained nearly stationary throughout the period for which we have data.

A simple model of prismatic volume elements based on the longitudinal profiles was used to calculate the total volume lost from thinning of the upper part of the glacier and the total volume gained from thickening of the lower part from 1978 to 1982. Effects of residual annual snowpack on the volumes were not considered. A total volume of 8.6×10^7 m^3 of ice was added to the lower part, and a total volume of 4.0×10^7 m^3 was lost from the fraction of the upper part covered by our maps.

DISCUSSION AND CONCLUSION

Volcanic eruptions during 1966 through 1968 reduced the flux of the lower part of Drift Glacier by more than half. Ice flux from the Upper Canyon and Summit Crater was eliminated for 8 yr. The glacier in the Upper Canyon re-formed rapidly and reconnected with the main system in 1976. The reconnection more than doubled the flux and propagated a kinematic wave through the glacier which has been observed by our surveying and photogrammetry.

The passage of the kinematic wave appears to be returning the glacier to its pre-eruption equilibrium. This can be seen by comparing photographs of the glacier before the eruption (fig. 4a) with recent photographs (fig. 4e), which show surface crevassing, thickness where ascertainable, and overall longitudinal profile to be similar to pre-eruption conditions. Note, for example, the close similarity between the 1954 and 1982 profiles in figure 6.

The regenerated glacier apparently had more mass when it reconnected than before the eruption. This is suggested (1) by our calculation that about twice as much ice accumulated in the Upper Canyon and Summit Crater during the 8 yr of its rebuilding than was lost during the eruption, and (2) by the data in figure 5, which show that the levels to which ice velocity decreased were still significantly higher than before the ice thickened. These calculations suggest an initial surge of higher flux as the excess mass in the upper glacier was discharged immediately after the reconnection, followed by a slightly reduced flux as the system came to equilibrium.

At the same time that ice was removed from the Upper Canyon, debris was deposited on the Piedmont Lobe, which reduced ablation and allowed the ice to thicken, as flow continued out of the East Basin. Reduced ablation created an ice surplus of 7×10^7 m^3 over what would have accumulated without the debris cover. Thickening during the same time was 9×10^7 m^3. However, the Piedmont Lobe must have been thickened not only by reduced ablation, but by the passage of the kinematic wave as well. The kinematic wave can be identified by its effects on the ice such as crevassing, associated loss of debris cover, and onset of melting.

Changes in velocity on the Piedmont Lobe observed between 1979 and 1986 cannot be explained by increased ice deformation caused by increased glacier thickness and surface slope alone. Some change in basal sliding seems likely. For example, between 1981 and 1982, the velocity at stake C-3 increased from 20 to 115 m/yr. During the same period, ice thickness increased less than 25 m and overall surface slope increased at most a few degrees. Assuming a conservative ice thickness of 200 m, theory (Nye, 1952; Paterson, 1981) indicates at most an increase in velocity of 20 m/yr. Changes in basal sliding may have resulted from changes in internal glacier plumbing as a result of the jökulhlaups, diminished water supply because of reduced ablation on the Piedmont Lobe, or from effects produced by the kinematic wave as it passed.

The events on Drift Glacier were caused by a dramatic external event--a volcanic eruption--but striking similarities exist between these events and those observed during a glacier surge. From 1978 to the present, we observed an order-of-magnitude increase in surface velocity, along with thickening and crevassing of the glacier surface. There was a net displacement of mass from an upper to a lower zone of the glacier. On the basis of these characteristics, it would be possible to classify Drift Glacier as type III in Meier and Post's (1969) system of surge glaciers. Although Drift Glacier is not considered to be a surge glacier, it might have some mechanisms in common with surging glaciers.

Though the kinematic wave appeared to be attenuating by 1986, its passage through the glacier could have created an unexpected geological hazard. If this wave--which caused some advance of the glacier margin between 1982 and 1986--had caused even 25 m of terminus advance, the glacier would have dammed the Drift River and created a glacier-dammed lake. Varved lake sediments found along the Drift River (A. Till, personal commun., 1984) indicate that this has happened in the past. Ice-dammed lakes are unstable and discharge catastrophically; such a lake would have posed a serious threat to the oil-tanker terminal located at the mouth of Drift River.

Generally, active volcanoes covered by glaciers pose unique geologic hazards. In addition to explosive eruptions (Alaska contains 90 percent of the explosive volcanoes in the United States), the melting of ice and snow on glacier-covered volcanoes can produce jökulhlaups and rapidly moving slurries of water-saturated mud, ice, and rock (lahars). Recent examples of the destructive power of lahars were provided by Mount St. Helens in 1980 and Nuevado del Ruiz, Colombia, in 1985. Mt. Redoubt has produced both jökulhlaups and lahars in the past and will produce more in the future. It is likely that both lahars and jökulhlaups require pre-eruptive melting of glacier ice, ponding of meltwater, and saturation of subglacial rock.

Post-eruptive hazards also occur on glacier-volcano systems, as pointed up by our work at Mt. Redoubt. The recovery of Drift Glacier from perturbations caused by the 1966-68 eruptions has taken 20 yr, and nearly resulted in a new flooding hazard 2 decades after the eruption.

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