ENGLACIAL STRUCTURES OF THE VAUGHAN LEWIS ICEFALL, AND RELATED OBSERVATIONS ON THE JUNEAU ICEFIELD, ALASKA, 1967 - 1969

Thesis for the Degree of M. S.

MICHIGAN STATE UNIVERSITY

LOUIS RUTHARDT MILLER

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ABSTRACT

ENGLACIAL STRUCTURES OF THE VAUGHAN LEWIS ICEFALL, AND RELATED OBSERVATIONS ON THE JUNEAU ICEFIELD, ALASKA, 1967-1969

By

Louis Ruthardt Miller

Wave-ogives are defined as the three-dimensional surface wave forms found at the base of many glacial ice-falls. The morphogenesis of these wave forms is in dispute. Two main theories exist to explain their origin: (1) The ablation theory--greater ablation during the summer months forms a trough and the increased accumulation during winter months causes a wave-crest. (2) The compression theory--wave-ogives result from a longitudinal compressive thickening at the base of an icefall, thus forming a wave during late summer presumably when compression is at its greatest.

Measurements of englacial structures on the Vaughan Lewis Glacier, Alaska, including steeply dipping tectonic foliation and transecting thrust surfaces, plus supplemental surface movement surveys obtained during the years 1967-1969 indicate that a longitudinal compressional stress does exist in the ice apron sector at the foot of the

icefall, and indeed in the basal ice of the lower half of the icefall itself. This interpretation reinforces the argument for compression thickening. It does not, however, rule out the possibility of at least some accentuation of surface bulges in the apron area by ablation, though this is considered to be a subordinate factor in basic genesis of these striking surface features.

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By

Louis Ruthardt Miller

A THESIS

Submitted to
Michigan State University
in partial fulfillment of the requirements
for the degree of

MASTER OF SCIENCE

Department of Geology

1970

ACKNOWLEDGMENTS

This study was made possible through financial and material support from the Foundation for Glacier and Environmental Research (FGER) via the Juneau Icefield Research Program (JIRP) in the years 1967, 1968, 1969, and research grants made through the Alaska Glacier Commemorative Project of the National Geographic Society, Washington, D.C. Assistance was also rendered by the Glaciological and Arctic Science Institute of Michigan State University and the U.S. Forest Service with the loan of surveying equipment.

It is with sincere appreciation that thanks are extended to Professor Maynard M. Miller for his stimulation, encouragement, enthusiasm, and guidance both in the field and during preparation of this manuscript. Thanks are also extended to Drs. Hugh F. Bennett, Robert W. Little, and James W. Trow for helpful suggestions and critical review of the manuscript as the other members of the writer's thesis committee.

Deep appreciation must also be extended to many of the members of JIRP, too numerous to mention, who have helped logistically. It is only through the assistance of these many selfless individuals that a topic of this magnitude could be approached and these results accomplished in the main summer field seasons of 1968 and 1969. It is especially true in field research on glaciers that one's success depends on the close help and cooperation of one's colleagues. This is because not only must one work closely with them, but also one must live with and learn from them at times when one's life is wholly dependent on their concerns and their abilities. For this aspect as well, special thanks are extended to my field associates Dr. Alfred C. Pinchak, William Lokey, and John Schutt.

Thanks are also extended by the writer to his brother, William M. Miller, and the writer's wife, Karin, for help in computer programming.

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INTRODUCTION AND PHYSICAL SETTING[1]

"Ogives have puzzled generations of alpinists and their close study is long overdue, especially as their elucidation should advance considerably our knowledge of glacier movement." These are the words of William Vaughan Lewis (1956), of Cambridge University, England, whose name this glacier bears.

The Vaughan Lewis Glacier on the Juneau Icefield (Figures 1 and 2) provides a classic example of the type of features Professor Lewis described. The Vaughan Lewis Glacier (Miller, 1963a), is considered to be a typical, climatologically-controlled normal discharge type of glacier (M. M. Miller, 1969, personal communication). It has been used since 1961 as a prototype on the Juneau Icefield, Alaska for study of glacial behavior and movement within an icefall (Figure 3).

The Vaughan Lewis Glacier is located in Southeastern Alaska at latitude 58° 49'N., longitude 134° 17'W. Its source area or névé lies immediately southeast of Mt. Ogilvie a boundary peak of the Northern Boundary Range

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[1]v. glossary on page 81 for definition of glaciological terms used in this presentation.

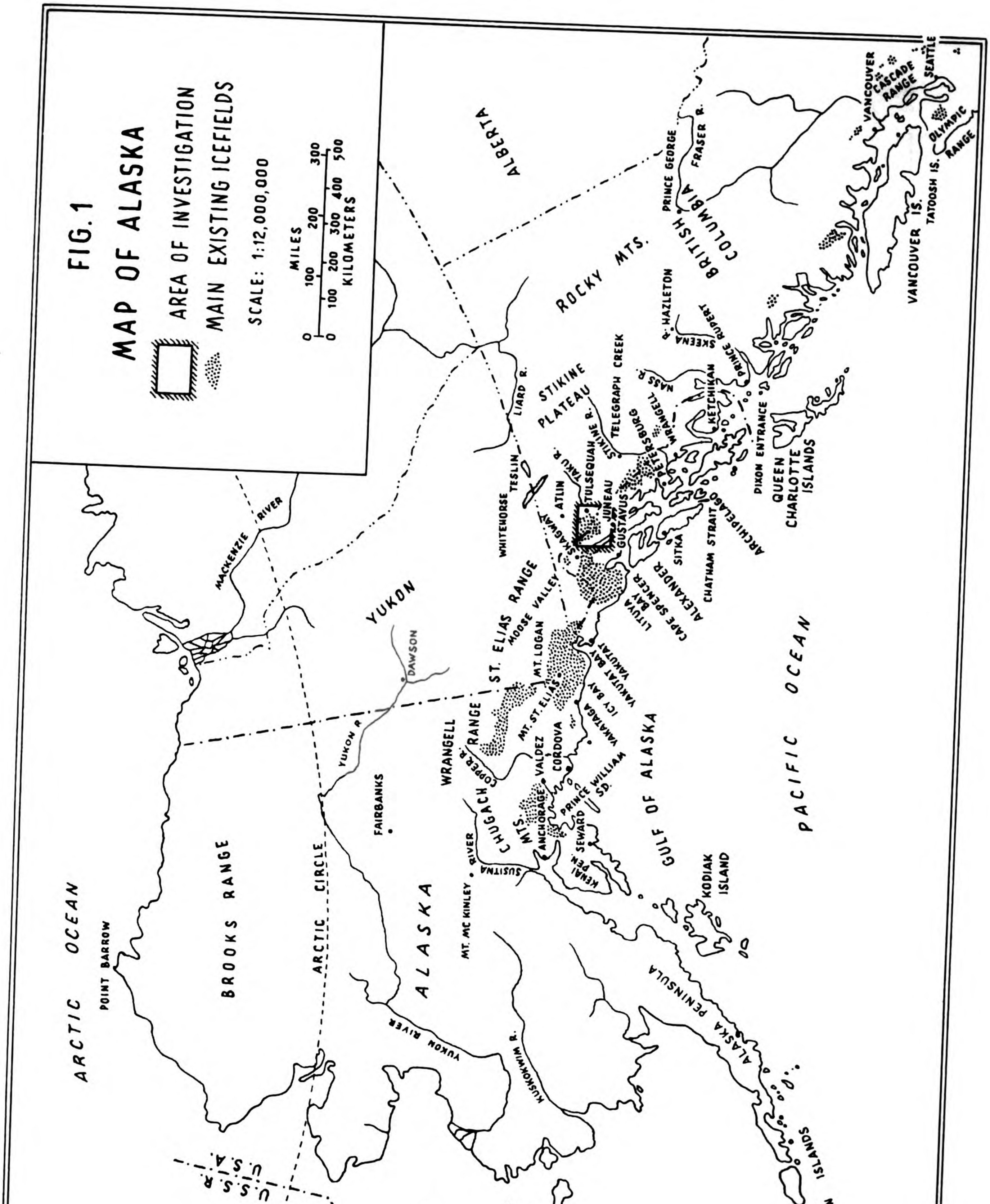

FIG. 1
MAP OF ALASKA
AREA OF INVESTIGATION
MAIN EXISTING ICEFIELDS
SCALE: 1:12,000,000
MILES
0 100 200 300
0 100 200 300 400 500
KILOMETERS
ARCTIC OCEAN
POINT BARROW
BROOKS RANGE
ARCTIC CIRCLE
ALASKA
YUKON
YUKON R.
YUKON RIVER
DAWSON
FAIRBANKS
MACKENZIE RIVER
ALBERTA
ROCKY MTS.
BRITISH COLUMBIA
PRINCE GEORGE
FRASER R.
HAZLETON
SKEENA R.
PRINCE RUPERT
NASS R.
TELEGRAPH CREEK
STIKINE PLATEAU
STIKINE R.
LIARD R.
TESLIN
WHITEHORSE
ATLIN
TAKU R.
TULSEQUAH R.
JUNEAU
GUSTAVUS
SKAGWAY
PETERSBURG
WRANGELL
KETCHIKAN
DIXON ENTRANCE
QUEEN CHARLOTTE ISLANDS
ARCHIPELAGO
ALEXANDER
CHATHAM STRAIT
SITKA
CAPE SPENCER
LITUYA BAY
ST. ELIAS RANGE
MOOSE VALLEY
MT. LOGAN
MT. ST. ELIAS
YAKUTAT BAY
YAKUTAT
ICY BAY
YAKATAGA
WRANGELL RANGE
COPPER R.
VALDEZ
CORDOVA
PRINCE WILLIAM SD.
CHUGACH MTS.
ANCHORAGE
SEWARD
KENAI PEN.
SUSITNA RIVER
MT. MC KINLEY
GULF OF ALASKA
PACIFIC OCEAN
KODIAK ISLAND
ALASKA PENINSULA
ALEUTIAN ISLANDS
KUSKOKWIM R.
BERING SEA
U.S.S.R.
U.S.A.
VANCOUVER
CASCADE RANGE
SEATTLE
OLYMPIC RANGE
TATOOSH IS.
VANCOUVER IS.

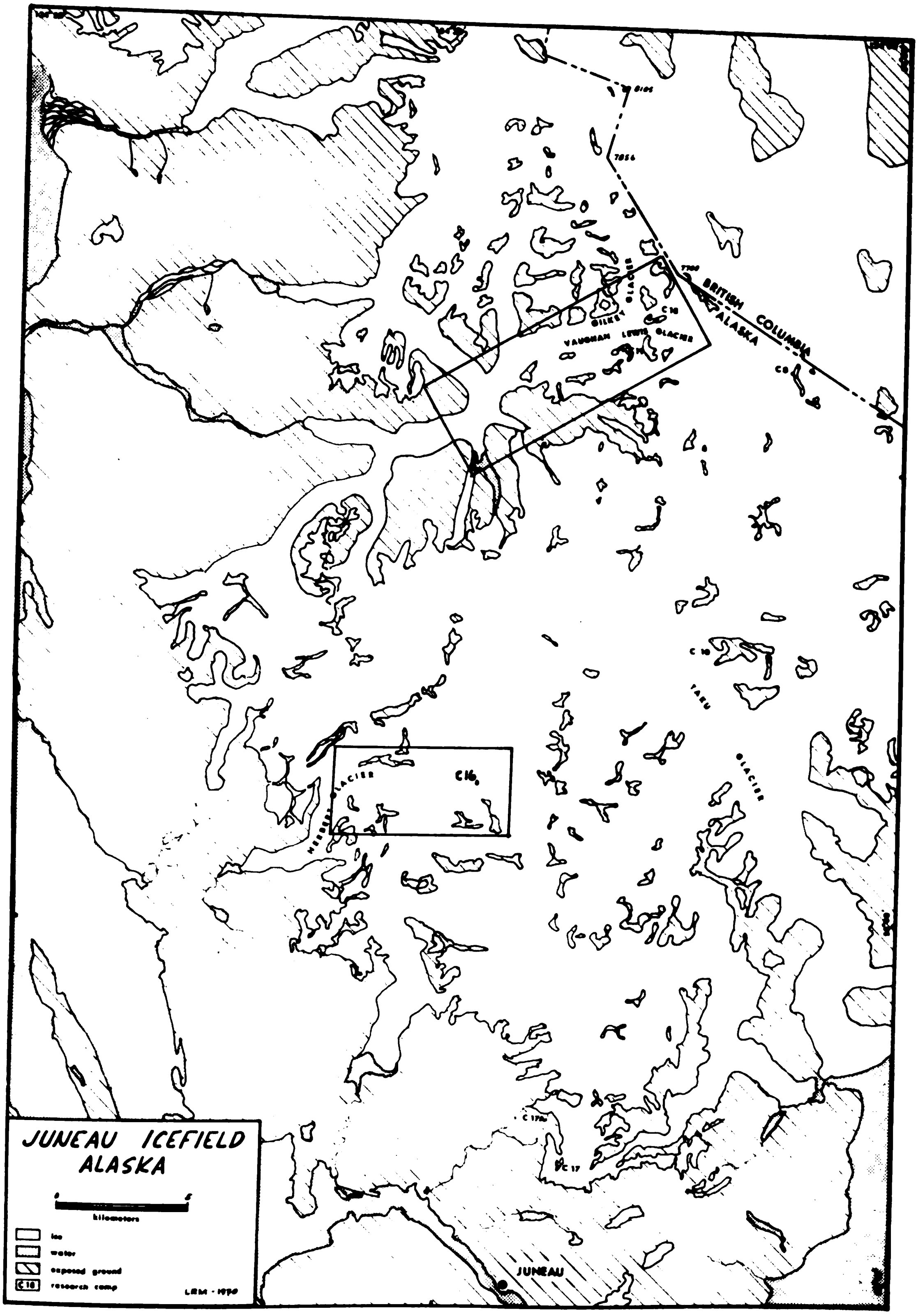

Figure 2.--Southwestern part of the Juneau Icefield denoting field camps used in 1967-69, and also position of Juneau, Alaska. Rectangles show specific location of research described in the report.

Figure 3.--Vaughan Lewis Glacier showing icefall, and apron area. Mt. Ogilvie 2350 m (7700 ft) in upper left. Storm Range in upper right. Camp 18, 1705 m (5600 ft) elev. lies on top of bedrock ridge in middle upper distance. Base of icefall and apron lies at approximately 1100 m (3600 ft). (Photo from Camp 19, by L. R. Miller, 4 August 1968.)

in the high Coast Mountains between northern British Columbia and Alaska.

The Vaughan Lewis Glacier (Figure 3) is of the glaciothermally (geophysically) temperate type with the main structures of concern in this study (surface wave bulges, wave-ogives and ogives) all lying downvalley from the mean névé-line of the 1960's. In the three years of this study the névé-line at end of summer attained an average elevation of 1277 m (4200 ft). It was as high as 1400 m (4600 ft) in the unusually warm and dry summer of 1968. In the unusually cool and stormy summer of 1969, however, it remained as low as 1160 m (3800 ft).

This icefall and glacier is fed by a névé at a mean altitude of 1680 m (5600 ft) with a total accumulation area of approximately 26 km^2 (8.6 mi^2). The icefall is approximately 1 km (0.6 mi) wide, and drops 500 m (1500 ft) at an angle of descent varying between 24° and 40° (Figure 3). The ice depth at the base of the icefall has been measured at somewhere between 180 m (600 ft) and 210 m (700 ft) in depth (Kittredge, 1967; Prather, 1969, personal communication based on 1968 seismic data).

The wave-ogive area lies on the downvalley apron of the icefall where the lower Vaughan Lewis Glacier meets two other main valley glaciers. Adjacent to this icefall and on the south is an "unnamed" glacier. Immediately adjacent and to the north is the Gilkey Glacier. The framing position of these glaciers is shown in Figure 7, page 21. Each coelesces and flows as one continuum down-valley to a terminal position 15 km (9 mi) west southwest of the icefall.

METHOD AND AREAS OF RESEARCH

On the Juneau Icefield the Foundation for Glacier and Environmental Research has constructed a number of permanent research stations in connection with its long-term Juneau Icefield Research Program. These are aluminum-covered and usually well insulated frame structures which serve as main centers from which field parties obtain scientific data regularly each summer, including daily meteorological information at the main stations (Figure 2). The research done on the Vaughan Lewis Glacier, which is discussed in the present report, was carried out during the years 1968 and 1969. Some reconnaissance observations by the writer were also made in the summer of 1967. In all of this, Camp 18 and Camp 18B, and to a lesser extent, Camp 19 served as the main research centers. Camp 18B was a tent camp from which most of this research was conducted. Here two nylon mountain tents and related temporary field facilities, including a meteorological shelter, portable radio, etc. were used. Camp 18 is a more permanent installation, comprised of several buildings. This was used as the radio relay station and permanent logistics base, and as such was revisited about once a week to replenish needed food and supplies at Camp 18B.

Camp 19 was also employed. This camp has a single wood structure and metal-sheeted building 4 m^2 (12 ft^2) situated on bedrock of a high berm at 1160 m (3800 ft) elevation. Its location faces the icefall on the southwest and lies 1.6 km (1 mi) distance from it. This camp was constructed in the spring of 1968 as a base for research not only on the Vaughan Lewis Glacier but also in the downvalley area of the lower Vaughan Lewis-Gilkey Glacier System. In the present study it was used as a station for working the lower glacier area, and especially in connection with movement data taken from the berm of Mt. Grosvenor[1] (Figure 2).

Observations and measurements of the internal structures of the Vaughan Lewis Icefall were taken within natural openings at the glacier's surface, especially in crevasses and stream channels (Figures 9 and 10). Measurements of strike and dip were made using a Brunton compass. Linear distance measurements were taken with a 100 ft

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[1]Provisional place name suggested in honor of Gilbert Hovey Grosvenor, National Geographic Society President, 1920-1950. This is in recognition of the National Geographic Society which has provided prime support for regional glaciological studies of which this research is a part.

(30.5 m) surveyor's chain (tape),[1] and a 10 ft (3.05 m) carpenter's tape.

Movement data were obtained with a Wild -T2 theodolite using triangulating methods from a base line the length of which was determined with an Invar Subtense Bar. Three meter (10 ft) sections of thin-wall electrical conduit were used as movement stakes. These proved to be easier to place in 3 cm (1 in) holes than the previously used 8 cm^2 (2 in x 2 in) wooden stakes. Also by having a galvanized finish they did not ablate out as fast as wooden stakes with a lower albedo. The conduits were placed on the glacier at pre-selected points along each movement profile. They were drilled into the ice or firn pack to depths of 2 m (6.5 ft) either with a 2.5 cm (1 in) SIPRE drill or a 4 cm (1.5 in) Norwegian type hand drill. Later in the field season when bare ice was exposed the Norwegian "Østrem-type" drill was wholly used. Tape measurements were also taken along the slope from movement stake to stake. After a few weeks, some of these stakes had to be reset to their initial vertical position. This was done by deepening the drill holes with an appropriate auger. This, of course, corrected for tipping due to ablation. Movement stakes aligned in strain-diamond patterns were

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[1]Unfortunately these data were taken in feet and fractions thereof. In research of this kind the metric system is much preferred; however, the metric chain anticipated was not at C-18 at the beginning of this study, and so for consistency the English tape continued to be used. Conversions are made in the following pages.

also implanted with measurements taken from carefully marked points on each stake. And similarly about every seven to fourteen days these movement stakes had to be redrilled because of ablation effects.

Various types of smaller pegs in strain-diamond patterns were also installed in horizontal positions on vertical crevasse walls. The data obtained from these proved to be of little value. The reason for disappointment here relates to a number of practical problems, mostly allied to ablation. For example, the installation of wooden dowels into horizontally drilled holes was tried first, then nails were driven into the ice. In each case methods to induce freezing around the markers failed. It was determined that any foreign material, wood or metal, would increase ablation around the marker and thus fall out before adequate records could be made. The best method was found simply to drill a hole and measure from its center. The error encountered, however, was of the same order of magnitude as the movement expected and so the results are so far inconclusive. With more time this method might prove to be of value if one applies a statistical mean movement analysis.

Toward the end of the 1968 field season, a new method was tried for the surface movement surveys. In this a pyramid pylon (Figure 4) was constructed to hold a movement stake more permanently in place, that is without necessitating redrilling the bore-hole. The pylon was also invoked to hold the stake perpendicular in spite of subsequent ablation. The technique was a modification of the movement tripod method used in many previous JIRP seasons. The pylon was constructed from 1.5 m (5 ft) sections of building material, put together in such a way as to form a pyramid with four triangular faces. The device proved to be very effective. In July, 1969 the pylon which had been placed on survey point no. 3 in July, 1968 (wave-ogive K), was still holding its movement stake in a vertical position. Another pylon, however, placed at stake no. 15 in the upper apron area (Figures 4 and 5) could not be relocated during the 1969 field season. This one was assumed to have dropped into a crevasse or to have been crushed by winter snow.

It is probable that the existing pylon at survey point no. 3 was sturdy enough to withstand very high winds[1] during the winter because the complete structure, except for the lower 15 cm (0.5 ft) was extremely abraded from the effects of wind-driven snow. Also, this would seem to indicate that little accumulation occurs on the crests of wave-ogives in the autumn, winter, and spring months, i.e., generally mid-September to late-June. Sequential movement records were taken on stake no. 3. Although not discussed

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[1]Wind velocities of 128 km/hr (80 mi/hr) were recorded at this site in September before the close of the field season. Velocities at various other icefield stations have been recorded in excess of 160 km/hr (100 mi/hr).

Figure 4.--Pyramid pylon marker placed at stake no. 15 in the summer of 1968. (Photo by L. R. Miller, 6 August 1968.)

Figure 5.--Stake no. 15 area late in the summer of 1969. (Note absence of pyramid pylon to right of cloth phototheodolite marker.) Glacier flow is from left to right. (Photo by L. R. Miller, 26 August 1969.)

Figure 4

Figure 5

in detail in this report, these data will be useful for comparison with future measurements.

To better evaluate the nature of glacier movement and surface changes within an icefall, a unique and rather unorthodox program involving crevasse "seeding" was inaugurated on 12 August 1968. In this, sections of damaged triangular radio tower and a total of fifty-two rectangular five-gallon gas cans (numbered with red paint) were dropped 30 m (100 ft) apart in two large crevasses at the top of the fall zone on the Vaughan Lewis névé--about 1620 m (5300 ft) elevation. For future reference a documentation of locations of these markers is given in the 1968 Camp 18 log.

The markers were positioned as a reference for obtaining velocity data at a later time. Everything depends, of course, on the cans subsequently becoming exposed by ablation further down glacier. With a knowledge of their initial position and depth--approximately 27 m (90 ft)--it is hoped that we can then gain a better insight on the structure and ice flow within the icefall, as well as information on the effects of ablation.[1]

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[1]That these markers will eventually become exhumed by ablation is indicated by a unique situation recently reported by M. M. Miller (personal communication) from the American Mount Everest Expedition of 1963. Certain items deposited in a crevasse of the Khumbu Glacier Icefall in March, 1963 were subsequently recovered on the surface of the ramp-slope (apron) of the icefall more than 1.6 m (1 mi) downvalley seven years later, i.e., in October, 1969. It will be highly significant, however, if this does not occur on the Vaughan Lewis Glacier a few years hence.

With respect to englacial structures in the ramp area where these movement data were obtained, apparent dips in the tectonic foliation were measured in the first five wave-ogives found in the ice-apron area below the icefall. These measurements were made in existing openings and fissures in the ice surface, including longitudinal crevasses, splaying crevasses, and meltwater runoff channels. Such features permitted observations to depths of 20 m (64 ft). The mean lower limit of actual measurement was at about 10 m (32 ft). From the tectonic foliation, inferences can be drawn about the structural stress and strain distribution within the ice, as will be discussed later.

The upper Herbert Glacier (Camp 16 arm) on the west side of the Juneau Icefield (v. map in Figure 2 and photo in Figure 6) was also studied in 1969 in a similar manner, although more in a reconnaissance fashion as this was the first field season in which that glacier was explored by members of JIRP. This activity was carried out from Camp 16, at 1590 m (5200 ft) elevation at the upper Mendenhall-Herbert Glacier névé. Aspects of this study too will be discussed in the following pages.

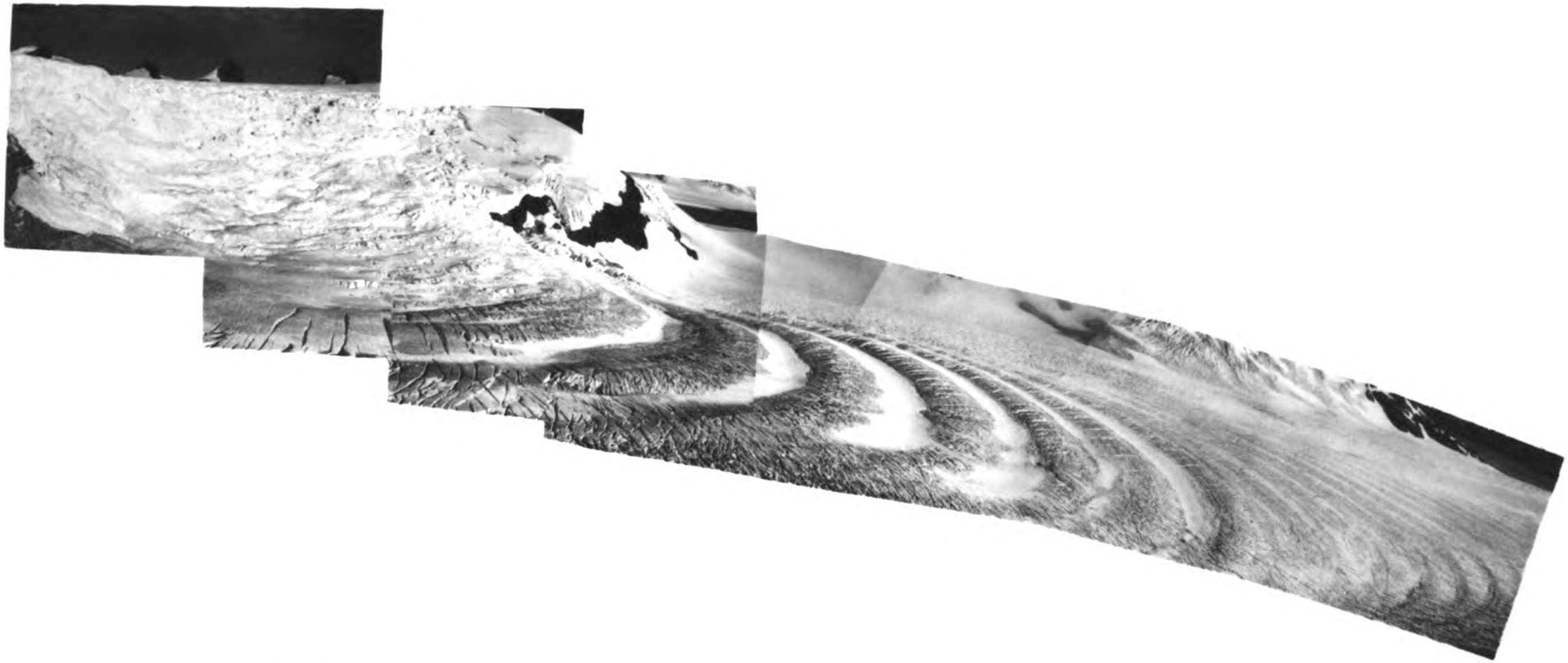


Figure 6.--Upper Herbert Glacier (Camp 16 arm) showing wave-ogives similar to Vaughan Lewis. View taken from bedrock survey station on northern side of glacier, looking toward the southeast. (Photo by L. R. Miller, 4 August 1969.)

INVESTIGATION OF THE VAUGHAN LEWIS GLACIER

The Vaughan Lewis Glacier has been divided into five glaciomorphic sectors--the crestal névé or prime accumulation zone, the icefall zone, the wave-ogive zone (wave-band zone), the ogive zone, and the terminal sector (Freers, 1966).

Observations in the Crestal Névé

The source névê or prime accumulation zone of the glacier lying between 1700-1800 m (5600-5900 ft) is not of major concern in the present study. Information concerning firn stratigraphy and glaciothermal conditions has been reported by other researchers for this and the adjoining zone of the Taku Glacier (Miller, 1952, 1963b; Freers, 1966). In addition, surface movement data have been obtained in this zone but are not discussed here as they do not bear directly on current studies of structural features below the icefall (Miller, *et al.*, 1968).

In 1968, however, a detailed study in the lower névé which adjoins the upper icefall zone was made emphasizing the sector just above the change of gradient in the fall zone. The purpose here was to determine the type and proportion of firn and glacier ice and to gain some insight into the nature of primary structures in the glacier. Crevasses were examined down to a depth of 27 m (90 ft), as near to the fall zone as practical and safe. This was done to determine if the structures observed below the icefall and especially within the wave-ogives are of subsequent or secondary deformational origin or of primary stratigraphic origin. The material observed varied from new to old surface snow to dense firn. It is of significance that no glacial ice or foliation structures were observed, even on exposed crevasse walls. The only primary structures found were dust layers (annual ablation surfaces) and various diagenetic structures such as ice strata (bands), ice glands, and ice lenses formed generally parallel to initial bedding in the firn by the refreezing of percolated surface water. It is emphasized again that nearly all of these features were planar in form, having horizontal orientations and distinguished quite easily from the tectonic folia found in wave-ogives well below the icefall.

Investigation of the Icefall Zone

The icefall zone is defined by Freers (1966) as that area below a point where crevasses are seen with their downglacier walls lower than their upglacier walls, i.e., essentially the serac and bergshrund zone. This zone has been noted to lie within the 1675 m (5500 ft) to

1225 m (4000 ft) range, and as such has been the object of intensive study by Pinchak (1968) with respect to the periodicity and cause of diurnal avalanches.

Between 1964 and 1968, photogrammetric surveys have been made of the Vaughan Lewis Icefall including, of course, the acquisition of comparative photographs. These photos also cover the wave-ogives and ogives from selected base-lines and glacier-surface stations (Chrzanowski, 1968). They have been obtained with a Wild P-30 phototheodolite and are currently the base for a large-scale map of the icefall and apron area being plotted by Dr. Adam Chrzanowski and Professor Gerhard Gloss at the University of New Brunswick. The glass plates have also been helpful in the writer's study of annual changes in the nature of the morphogenic features below the icefall.

At the top of the Vaughan Lewis Icefall there are large curvilinear tension crevasses which degenerate downward into a pronounced zone of seracs and ice pinnacles. Within this area much avalanching occurs. During 1968 and 1969 as many as seventy-five avalanches were recorded in a single twenty-four hour period. Although the crevasses at the head of the icefall are wide and deep and so could be investigated directly, because of avalanching none could be investigated within the icefall (Figure 3). A similar role for avalanches was found to characterize the upper Herbert Glacier (Camp 16 arm) icefall near Camp 16 (Figure 6), proving that both of these glaciers are vigorously discharging through their icefall outlets from their source névés.

King and Lewis (1961), in reference to the Odinsbreen of the Norwegian glacier Austerdalsbreen, have suggested that in such instances crevasses may extend downward to the bottom layers of ice and into less disordered ice masses which survive ablation in the icefall and thence are moved downslope to the ice apron area below. This has not been observed by the writer or others on the Vaughan Lewis or the Herbert Glaciers (Freers, 1966; Kittredge, 1967). Actually what this entails is that only shallow crevasses survive in the icefall zone and that these become largely filled with slump material, and hence in general obscured. Freers (1966) has cited evidence that a few do survive into the apron and wave-ogive zone. This substantiates the writer's view that such crevasses do not extend into ice at the bed of the glacier.

Morphogenetic and Structural Features in the Wave-Ogive Zone

The three-dimensional wave-ogive forms in the apron below the icefall (Figure 3) are the main focus of attention in this study. Downglacier these surface waves or bulges attenuate to form two-dimensional ogives (Figure 18).

For reference purposes the wave-ogives are lettered in a downglacier sequence (Figure 7), starting

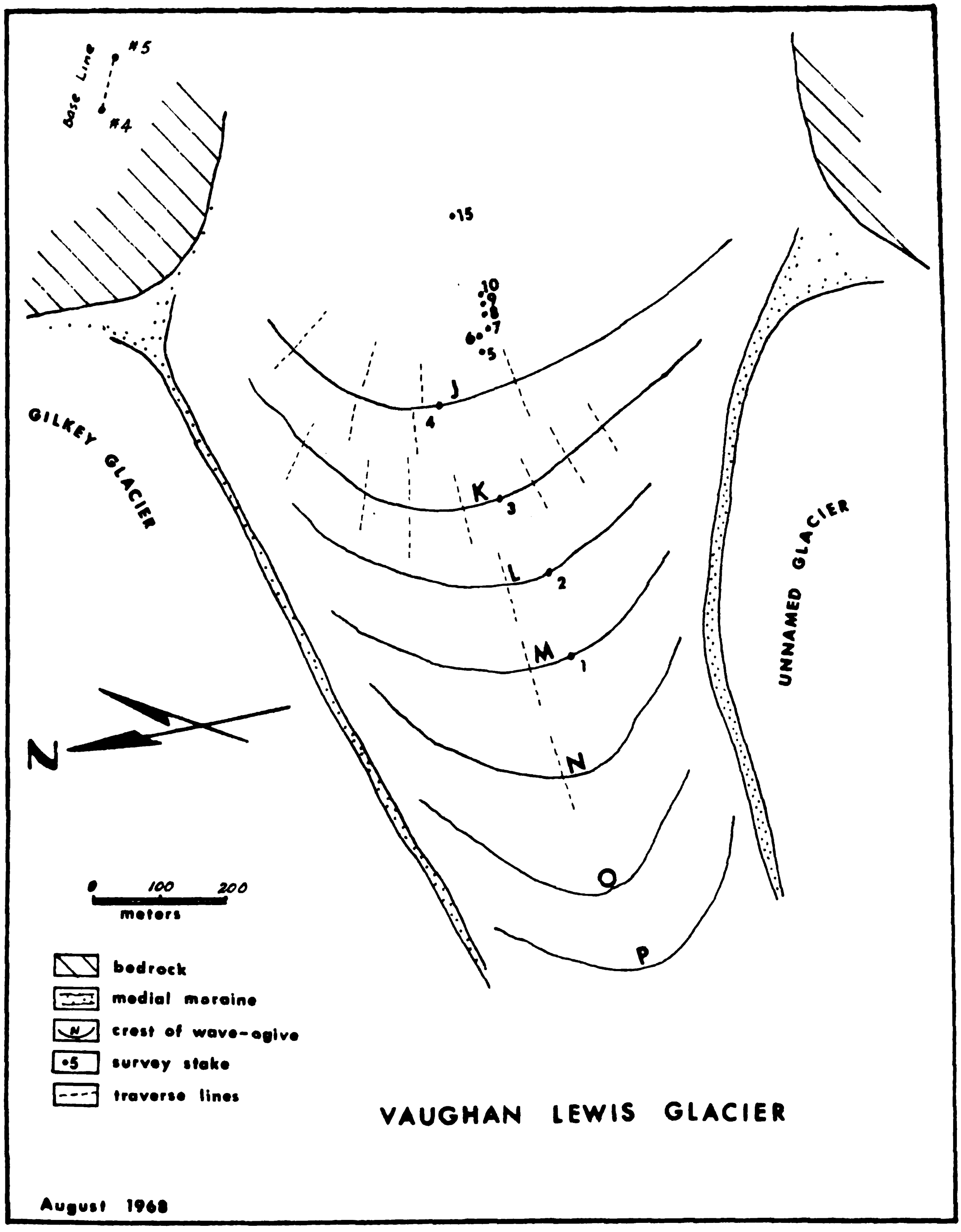


Figure 7.--Sketch map of wave-ogive crests on the Vaughan Lewis Icefall as observed in August 1968.

with J, which designates the uppermost wave-ogive studied in 1968-69. By starting at a midpoint in the alphabet it will be possible in the next few years of this long-term study to extend the series by adding I, H, etc., as new wave-ogives develop at the base of the icefall and without changing a numbering system each year. In Figure 7 the present sequence downglacier is labeled as J, K, L, M, N, O, and P. Wave-ogive P refers to the end of the dominant surface bulges designated as wave-ogives observed in August 1968.

The amplitude of the wave-ogives on the Vaughan Lewis varied from 25 m (82 ft) at the base of the icefall to 0 at a point about 2 km (1.2 mi) downglacier (Figures 8, 20, and 23). The wave length of the wave-ogives varied from 90 to 150 m (288-480 ft). Maximum amplitude could at no time be accurately measured without drilling a test hole because winter snow in the intervening troughs had not totally melted, even by the mid-September end of the ablation season in 1968 (v. Figure 23). In 1969, the situation was worse because new snow arrived and remained at this level by mid-August--1969 being the year of heaviest summer accumulation in the whole of the past twenty-five years of record on the Juneau Icefield.

As originally suggested by Miller (1952) and supported by Kamb (1964), the key to understanding wave-ogives and ogives will come from studies of their internal

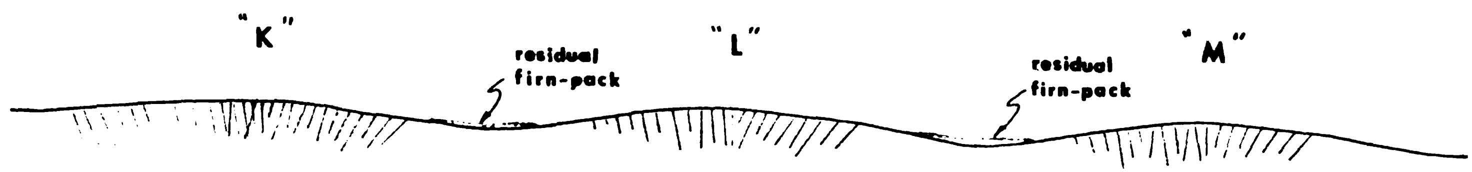


Figure 8.--Diagrammatic reconstruction of surface detail on longitudinal traverse across wave-ogives K, L, and M (v. Figure 7) showing dip of tectonic foliation to depth of 10 meters relative to position of wave crests and troughs.

structure, particularly at depth. Thus in this present study, primary attention is given to the nature and origin of the tectonic folia.

Tectonic foliation is a term first suggested by Miller (Glacier bands, conference on terminology, 1953; also v. Miller, 1952, 1955; and Taylor, 1963). As defined in this study these are planar or foliation structures produced by shear or compression and consisting of alternate zones or seeming layers of fine and coarse-grained ice. In some cases, grains of ice and trapped air bubbles are elongated in the plane of the layer (Figures 9 and 10).

Rigsby (1958) and Kamb (1959b) have indicated that both polar ice and temperate ice show a single maximum crystal orientation (rather than the two areas of concentration from the possibility known in biaxial ice crystals). In other analyses, Gow (1964) has shown two, three, and four maxima, but for the present analysis the probability of a single maximum is emphasized because of the strong unidirectional stress suggested by the geometry of the present case.

In this analysis (v. Kamb, 1959a) the c-axis orientation would be centered about the pole. Thus a single area of concentration of the c-axis perpendicular to the basal plane or plane of foliation is determined and plotted. The point where the perpendicular to this plane intersects the projection sphere is called the pole of that plane.

Figure 9.--Tectonic folia in vertical wall of stream channel cross-cutting β-sector of wave-ogive M. Note ice axe hanging vertically from ice screw. (Photo by L. R. Miller, 27 August 1968.)

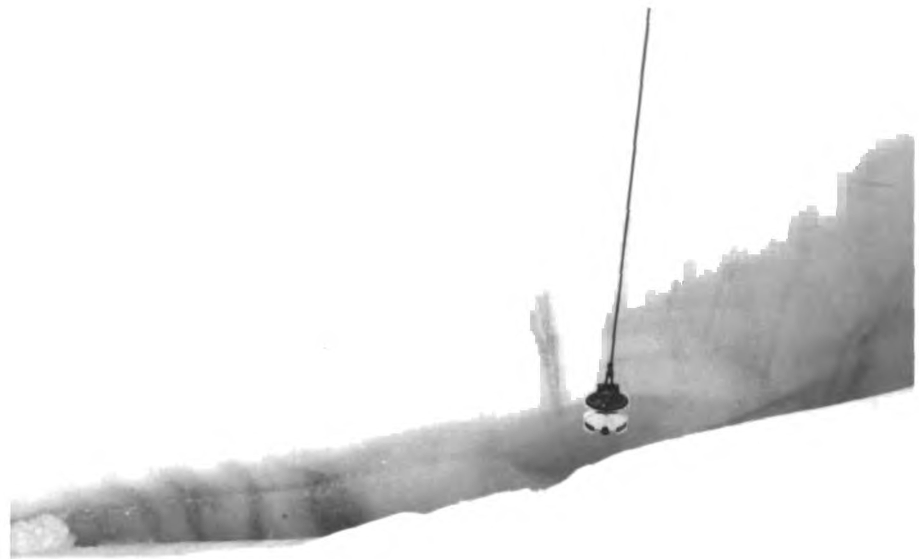


Figure 10.--Tectonic folia as exposed on cross-cutting crevasse wall α-sector wave-ogive K. Note that vertical line from compass hanging on string reveals significant dip of folia. (Photo by L. R. Miller, 4 August 1968.)

Kamb further states that the recrystallization of glacier ice in situ during glacier flow will give a preferred orientation of the c-axis normal to the foliation planes. On this basis the basal planes of the individual hexagonal crystals would be parallel to the folia. Such orientation minimizes the chemical potential which is required for equilibrium across the plane normal to the greatest principal axis of stress.

Such a comparative relationship has been studied previously in this region, in fact on the adjoining Taku Glacier where deep drill core samples were analyzed by Bader in 1950 (Miller, 1963b). From this three-dimensional field study of crystal fabric it was determined that from the surface of the Taku Glacier to 42.5 m (140 ft) the c-axes of the natural ice crystals were preferably at low angles to the horizontal, but without any azimuth. Below 42.5 m (140 ft), the c-axes were still preferably horizontal but in successively deeper samples to the base of the bore, a progressive crowding of azimuth values was observed towards a line of unknown orientation (presumably, according to the above noted discussion, this would be normal to the direction of "flow").

From this morphogenic relationship given by previous investigators of basal plane foliation (tectonic folia) and related planar structures (Bader, 1951; Risgby, 1958; Kamb, 1959b; Miller, 1963), it was concluded that fabric data from thin sections of Vaughan Lewis ice are not mandatory for purposes of the present study. Also it was realized that the effort required in obtaining such information in the field time available would have limited the areal extent of the ice structure investigation. Thus, by using this seemingly relevant c-axis to foliation relationship, an attempt is made to plot the orientation of measured tectonic foliation. In turn some apparent conclusions are discussed.

In Figures 11 through 16 strike and dip orientations of measured folia are plotted using the lower hemisphere of an azimuthal equal-area (Schmidt) projection (Billings, 1954). Such a technique has been successfully employed by Untersteiner (1955) on the Pasterze Glacier of the Grossglocker in the Eastern Alps.

Wave-ogive J on the Vaughan Lewis Glacier (Figure 7) is referred to as a "pre-ogive" because it did not have a well-developed trough on the upglacier side. Here the orientation of folia varied in true dip from 34° downglacier to 17° upglacier; however, it must be stated that most were very steep dips, i.e., within 20° of perpendicular. Below this, each wave-ogive was arbitrarily divided into two sectors using the crest of the wave as the dividing point. With this point of separation, structural relations on each half of a wave could be more accurately determined with respect to changes in the surface configuration of each limb of the wave.

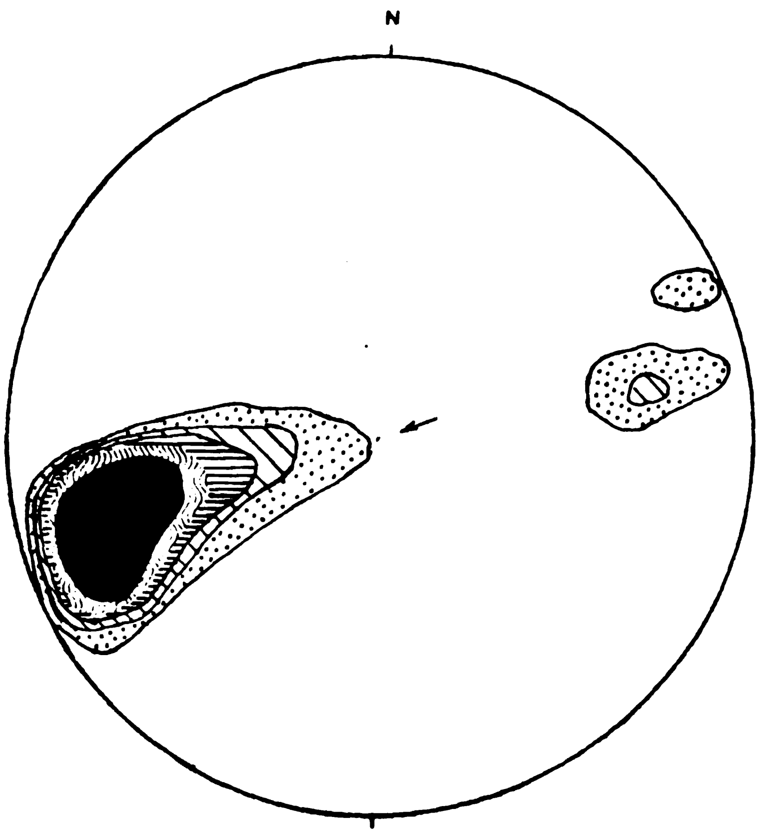


Figure 11

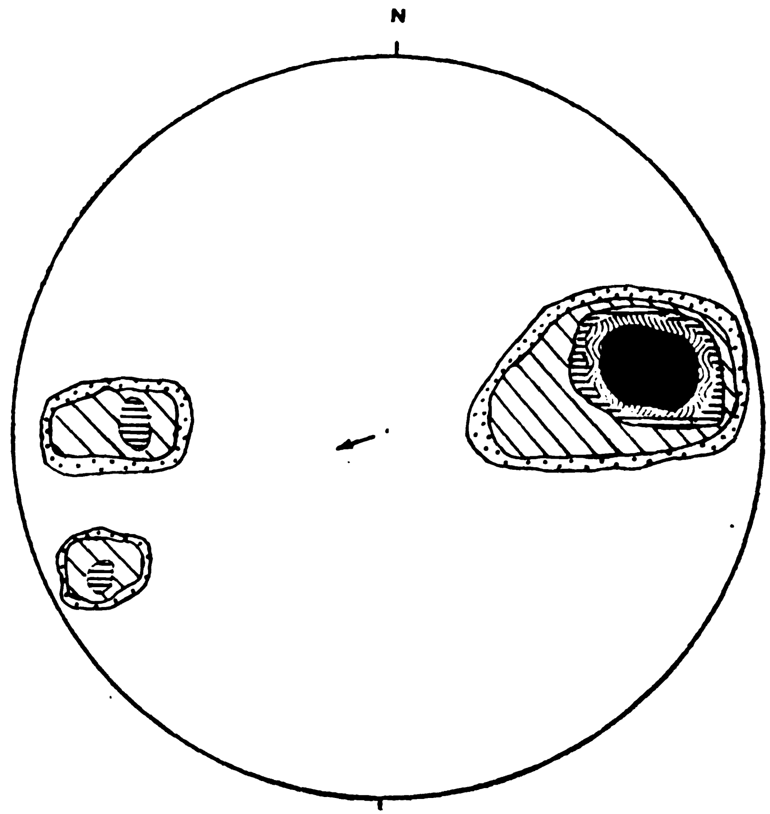


Figure 12

Percent concentrations of point poles contoured for foliation planes.

Distribution of strike and dip orientations of folia in wave-ogive J.
Figure 11, Crest of wave--65.1 m downglacier.
Figure 12, Crest of wave--20.1 m upglacier.
Flow direction indicated by arrow.

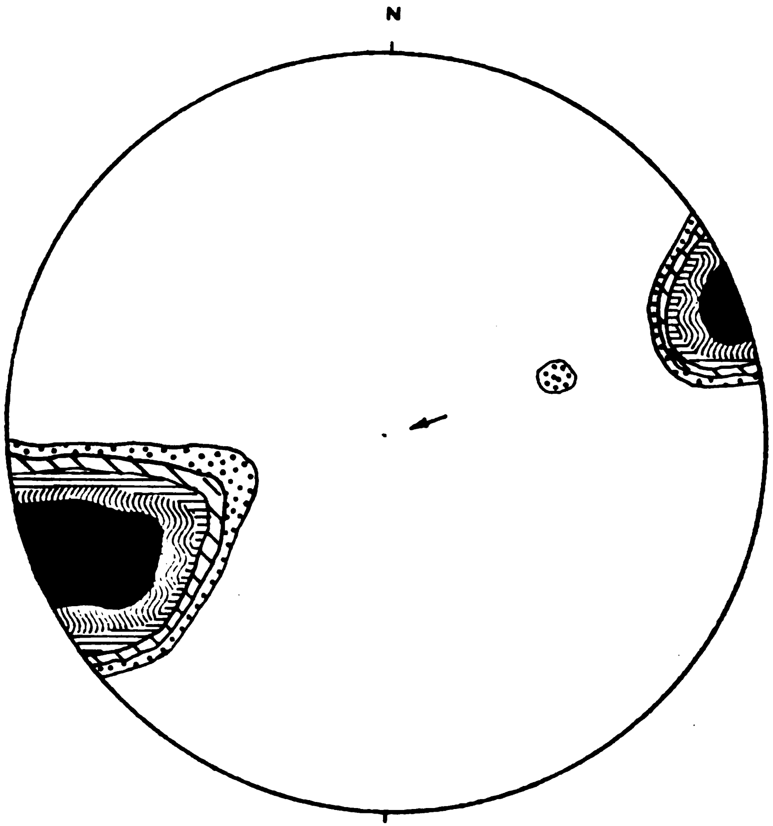


Figure 13

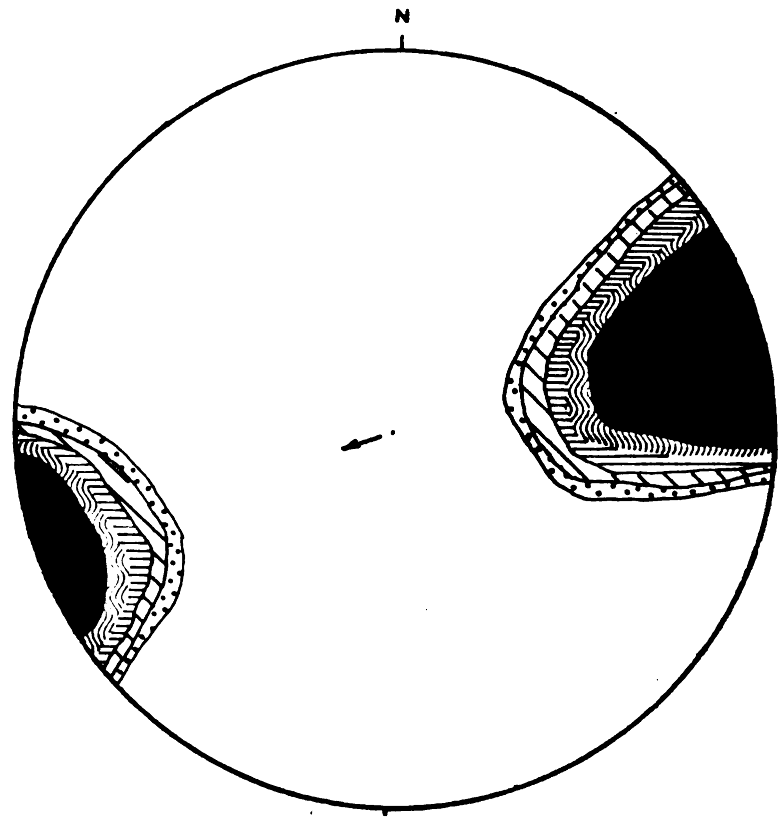
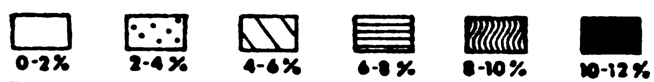


Figure 14



Percent concentrations of point poles contoured for foliation planes.

Distribution of strike and dip orientations of folia in wave-ogive K.
Figure 13, Crest of wave--41.4 m downglacier.
Figure 14, Crest of wave--46.6 m upglacier.
Flow direction indicated by arrow.

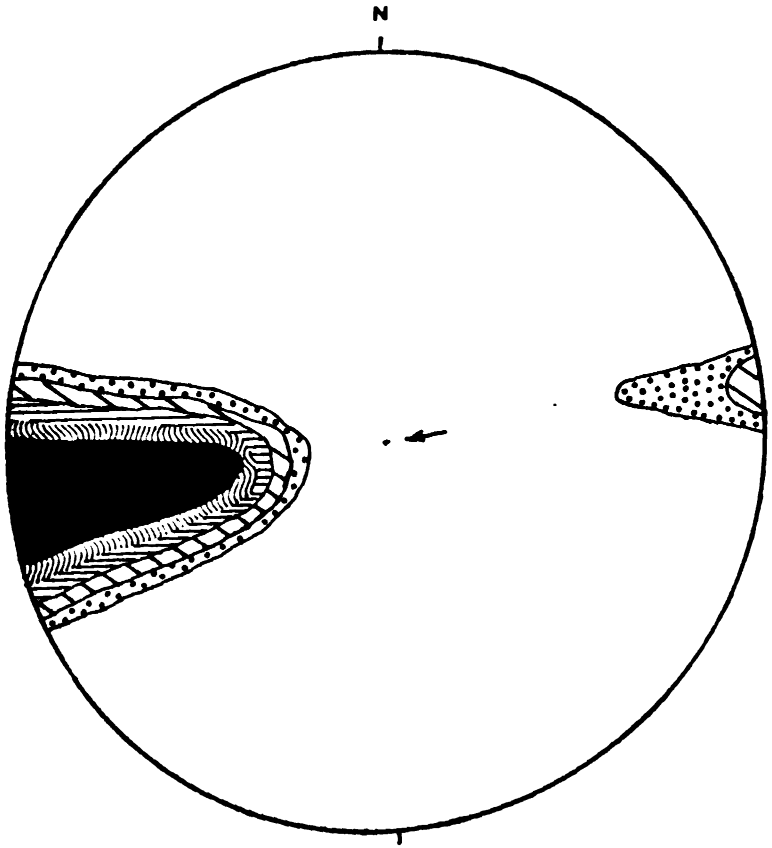


Figure 15

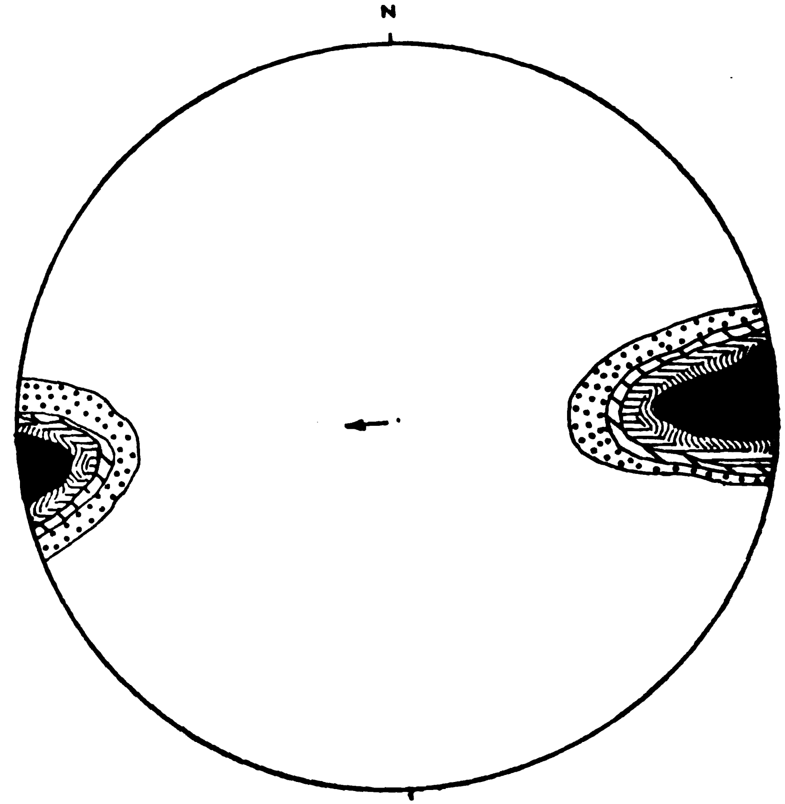
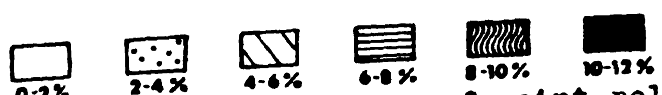


Figure 16



Percent concentrations of point poles contoured for foliation planes.

Combined distribution of strike and dip orientations of folia in wave-ogives L, M, and N.
Figure 15, Crest of wave--73.2 m downglacier.
Figure 16, Crest of wave--63.1 m upglacier.
Flow direction indicated by arrow.

As noted above, the folia distributions were plotted on a Schmidt net, a projection which is true to area. For clarity each limb of the wave-ogive was considered separately. The data from the wall exposures of several crevasses radiating across the wave-ogives make up the total for a single projection. Thus the strike and dip of either exactly 100 or 150 measurements constitute the point poles used to construct each Schmidt projection (again refer to the following Figures 11-16).

In many cases the bands or tectonic folia were found to be distorted, faulted, fractured, and sometimes bifurcated. In each instance, the most linear and distinct section was used for measurement. Each dip reading was taken with a possible error factor of $\pm 2°$. This figure includes both systematic error and random error, viz., the result of instrumentation and human limitations--and as such, are subject to the laws of probability. By using a percentage distribution, as shown by area on a projection, each single reading represents only 1 per cent or less of the total information, and therefore any extremes, either unique or by personal error, do not affect the general trend.

As can be seen in Figures 11 through 16, the direction of folia seem to be in no way irregular or random. There is a misleading aspect here when one looks at the distributions in Figures 11 through 16. This is because each figure is a representation of one or the other limb of a wave; viz., those folia which are near the centerline of a wave-ogive crest are either on one side or the other of this centerline and thus are divided between each half. This division lessens the total number in each half having a dip near the perpendicular and hence gives a misleading overall percentage near the edge of the Schmidt projection (especially note Figure 11).

For purposes of this analysis the true dip angles are extended artificially to depth as is done in Figure 17. Note in this figure that below the limiting depth of observation the extension lines are dashed. The purpose of extending these lines is to better compare the relationship of adjacent dips and to see whether there is any possible point of focus. This figure combines the foliation dips as a generalized composite of data geometrically representing information from separate but sequential traverse lines. In this, we see that (v. Figure 12) the tendency is for most of the folia to dip downglacier in the trailing or upglacier sector of a wave (hereafter referred to as the α-sector). Conversely in Figure 11 we see the dip to be generally upglacier in the leading or downglacier sector of a wave (hereafter referred to as the β-sector). Again it is stressed that the extension of the folia lines in Figure 17 does not imply that these structures in nature extend in this manner at depth. In fact, it is further noted that indeed this

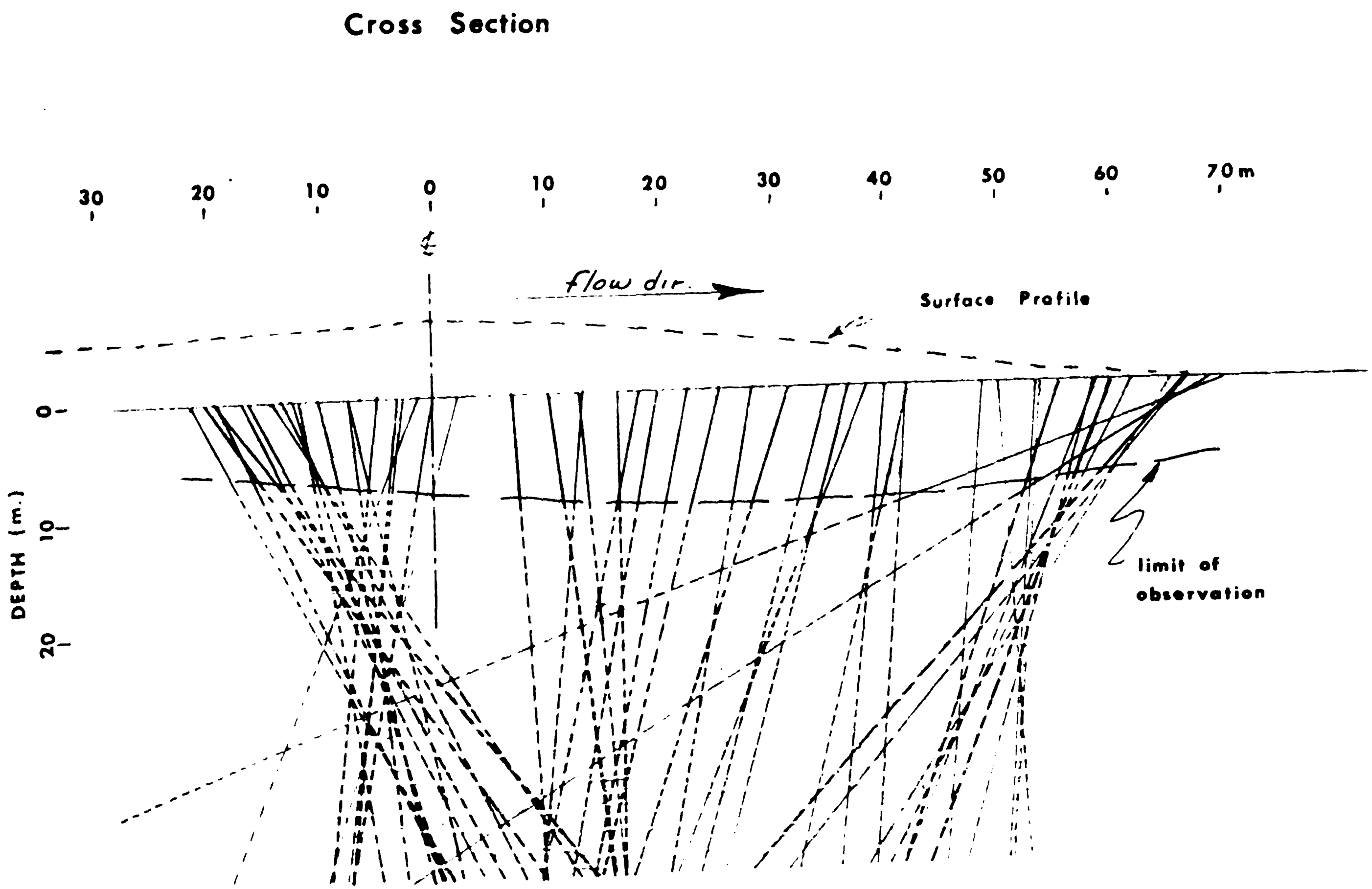


Figure 17.--Wave-ogive J at movement stake no. 4, showing dip of foliation along a crevasse wall. Folia lines measured in surface zone, hence full extension to depth by dashed line below the limit of measurement is purely artificial, and done only for the purposes discussed in the text.

manner of plotting revealed no apparent relationship in nature allied to such hypothetical focal points as the intersections of these lines might imply.

In Figure 17 a wide range of dips is represented. The highest percentages, however, center around 70° dip on each limb of a wave-ogive.

The strike of these folia is essentially parallel to the arcuate centerline of the wave-bulge--i.e., roughly parallel to the trend of the wave-ogive crest. The values recorded do not vary more than ±10° from this centerline, with a mean variance of ±3°. The amount of error in strike is considered to reflect human error in the method of measurement rather than inherent differences in value. Here the mean variance is greater because of the irregular surface upon which measurements were taken. One factor is the direct exposure of these surfaces to atmospheric processes, producing much greater ablation on horizontal surfaces than on crevasse walls. A further consequence is that glacier ice on these top surfaces is more bubbly, making the strike of specific foliation zones more difficult to distinguish and trace.

In the distal or β-sectors the folia in longitudinal cross-section have a tendency to be either concave or convex downglacier; i.e., the deeper folia at the same distance from the centerline of a crest are found to have either larger or smaller angles to the horizontal in comparison with their surface dips. With all folia at a rather high dip to the surface and with lesser dips on distal β-sectors, an asymmetrical appearance is created by the β-sector of each wave-ogive. This asymmetry is greatest in wave-ogive J (which we see in Figure 20, with wave-ogive J and K, left and center, respectively), suggesting overthrusting or underthrusting effects. Indeed on several crevasse wall profiles truncating overthrust surfaces have been observed (Miller, 1958). According to Billings (1954) this would be similar to the layers in metamorphic rock[1] which have yielded the most under a given stress as in the forward β-sector of a wave-ogive, and therefore show greater deformation.

In various places there are non-continuous lines of folia cut by more strongly developed lines of foliation at slightly different angles. Such foliation has been observed by Chamberlin (1928) and also Perutz and Seligman (1939) along the margins of glaciers, and has been related to differential movement and shearing. It also suggests deep ice translation into new stress fields as the glacier flows downvalley. From the predominance of cross folia in the β-sector of most wave-ogives studied, one might conclude that in some cases high angle overthrusts or underthrusts have indeed developed.

There is usually well-marked foliation throughout the entire wave-ogive. In some areas it appears rather

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[1]Glacial ice is technically a metamorphic rock.

jumbled, possibly because of the presence of avalanche material which has not yet become consolidated. This disorientation becomes less as one proceeds downglacier from wave-ogive J to wave-ogive N.

As stated previously, the folia dip angle in wave-ogive J surface ice ranges from 34° downglacier in α-sector to 17° upglacier in the β-sector. The total range of similar dips in α- and β-sectors, respectively, of wave-ogive K; are 51° and 28°, 62° and 42° in L, 70° and 38° in M, and 67° and 71° in N. These are the extreme dip values (i.e., angles from the horizontal) recorded in a total of nineteen traverse cross-sections of the five largest wave-ogives (v. Figure 7).

It is noted again that all folia in the α-sector do not necessarily dip downglacier, nor do all folia in a β-sector always dip upglacier. This can be seen from the distribution of folia in Figures 11 through 16 and in part of the profile of wave-ogive J in Figure 17 (v. anomalus distribution of 2-8% concentrations of point poles on east side of Figure 11 and west side of Figure 12). This again may be the result of local deformation effects.

Also note is made of the trend toward greater angles in the α-sectors with increasing distance downglacier, which in turn corresponds to changes in surface slope and attenuation of the wave-bulges downvalley. In these lower valley sectors, however, the englacial folia remain somewhat normal to the surface. A similar situation, of course, is noted with respect to the mean dip of the tectonic folia described in Figures 11 through 16.

In individual wave-ogives there are notable changes along the crest of the wave as well. In wave-ogive J, for example, the amplitude increases on the order of 5-10 m (16-32 ft) as one walks from the north end to south. This increase in amplitude is also associated with an increase of slope on the downvalley wave surface--i.e., from 30° on the north end to 67° on the south end of the β-sector of this wave-ogive.

Freers (1966) described nine whiter ice folds found in the summer of 1964 above, or upglacier from, the full-formed wave-ogives in the apron sector of the Vaughan Lewis icefall. Most of these were extended continuously across glacier, though some were segmented (Figure 3). These folds have the appearance of symmetrical, double plunging anticlines. Kittredge (1967) and the writer have noted that these small folds seem to appear most commonly on the surface of the larger waves. These smaller superimposed cleaner-ice folds diminish in frequency and size downglacier from wave-ogive J. That is, they appear to have their maximum expression in the upglacier sector of the apron. Far downglacier in the main ogive zone they disappear completely.

Crevasse Observations

Crevasses and fracture patterns on a glacier are principally a reflection of the compressional, tensional, and shear stresses within the upper part of the glacier.

The crevasses at the head of the Vaughan Lewis icefall are, as have been shown, entirely of the tension type. We have also seen that significant or deep crevasses do not occur in the main icefall, but reappear in the sector just above or at the beginning of the wave-ogive development. There are some depressions paralleling the wave-ogive crests which, of course, may be crevasses at depth. But essentially, as can be seen in plan view, the most dominant crevasses are radial. This means they are normal to the arcuate crest axes of the wave-ogives (Figure 18). In some cases these radial fissures are eń echelon. In others, firn in the largest radial crevasses contain smaller fissures with eń echelon pattern (Figure 19). It is not clear whether this smaller pattern of fissures relates to tensional spreading of the large crevasse walls, or to some other stress couple not yet measured. This may even be a result of firn-pack slump within the crevasse, combining effects of ice stress and ablation. Such warrants further examination in a subsequent field season.

Such ice at the base of icefalls is subjected to longitudinal compressional stress (Nye, 1951, 1952, 1959; Glen, 1956). This is because of the abrupt decrease in

Figure 18.--Lower Vaughan Lewis Glacier showing wave-ogives and ogives in compression zone below the icefall. Note position of stream channel and ponded water between wave-ogives. (Photo by L. R. Miller, August 1967.)

Figure 19.--*Eń echelon* fissure pattern within radial crevasse on southern edge of Herbert Glacier. View looking southerly with main section of glacier on right and flow direction is away from viewer. (Photo by L. R. Miller, 4 August 1969.)

Figure 18

Figure 19

bedrock slope (Figures 18 and 20) and the consequent damming effect and slowing of forward velocity. (See effect of longitudinal stress trajectories as controlled by even small changes in bedrock slope in Miller, 1958, as measured on the adjoining Taku Glacier.) This compression tends to bring pre-existing planar features or inhomogeneities of any kind (e.g., basal tectonic folia) within the ice into a more vertical transverse attitude (Allen, et al., 1960). This mechanism has been emphasized for the Vaughan Lewis in a new theory of wave-ogive formation elucidated by Miller (1968); also Miller, et al. (1968). In this regard, Untersteiner (1955) points out further possible complexities by showing that: "corresponding to the two main normal stresses we would have altogether six directions of maximum shear stress in which gliding could theoretically be expected." This would imply also characters of c-axis maximum. Here again reference can be made to the two to four maxima observed in studies by Gow (1964), as noted earlier. Thus, the writer is quite aware that much detailed work on this subject is yet to be done. But again the geometry of the Vaughan Lewis icefall suggests a single maxima.

Regardless of the foregoing, it is clear that wave-ogives M and N have transverse crevasse patterns. According to Nye's initial plastic flow analysis for ice (1951), a transverse pattern can be associated with extended flow and can be further correlated from data

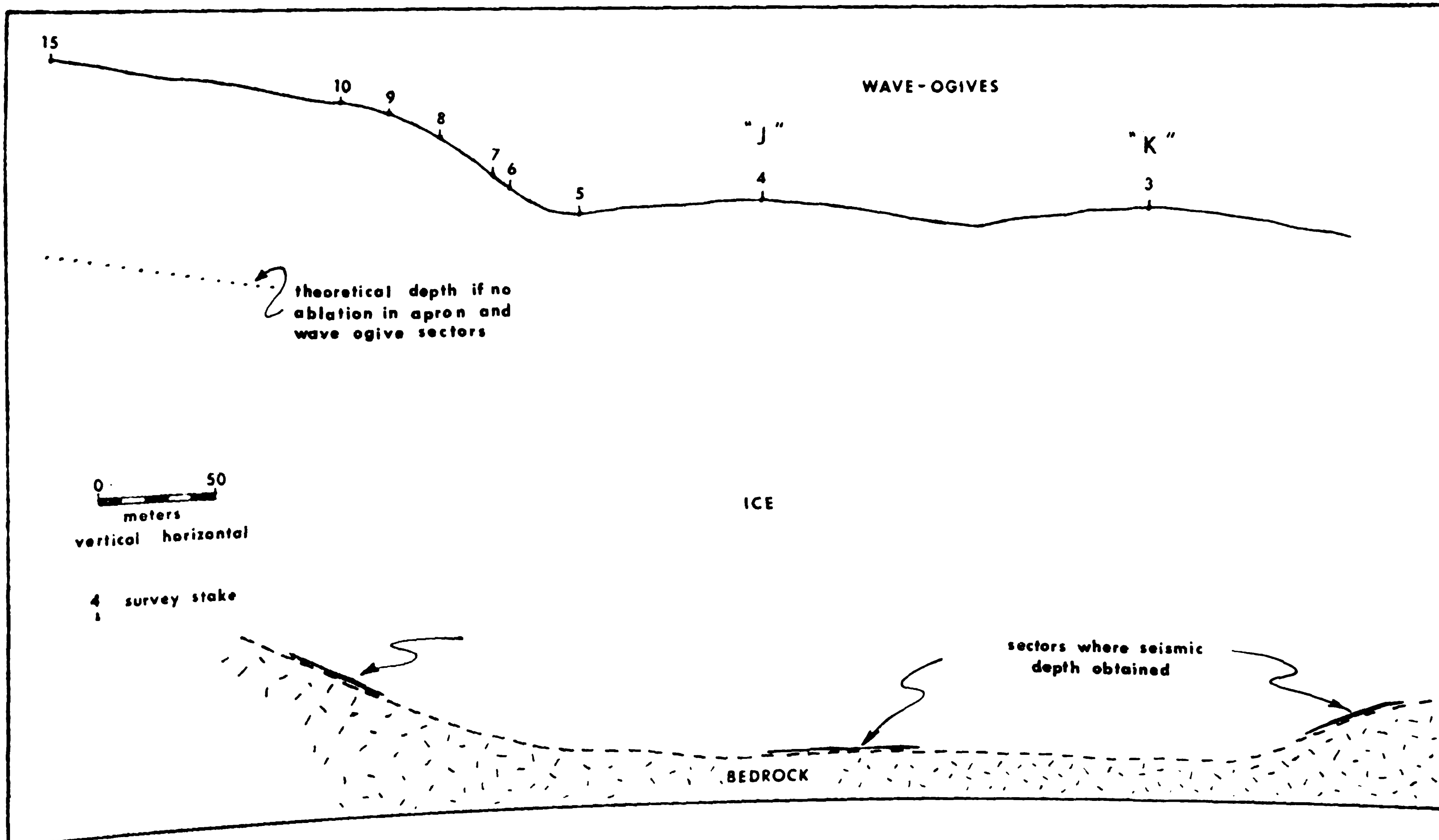


Figure 20.--Longitudinal profile of lower icefall, apron and wave-ogives J and K, Vaughan Lewis Glacier in September 1968. Bedrock depth as suggested by seismic work of Prather, et al. (1968) and Kittredge (1967).

obtained. Attempts have been made, however, to measure the strain rate on the Vaughan Lewis by having four stakes at the corners of a square and measuring the intervals between successive stakes (Nye, 1959), as suggested by Nye (1967, personal communication) and have yet to yield much information. A similar technique for measuring surface strain-rates has been applied by Wu and Christensen (1964) on the Taku Glacier, with not entirely satisfactory results. Many of the problems have resulted from an inability in even solid firn to measure small values of deformation accurately over short periods of time. In other words, it is most difficult to avoid error encountered by settling and ablation of surface markers. Recognizing this, in 1969 R. Little (personal communication) installed a set of 2.5 cm (1 in) pipes, 4.8 m (16 ft) long, in squares on the Lemon Glacier at the southwest edge of this icefield near Juneau, and attached piano wires to a set of sensitive strain gauges placed at the corner of each square. This technique is also currently being organized for use in a similar program on the Vaughan Lewis Glacier planned for 1971.

Surface Movement, 1967-1969

Movement stakes were positioned on wave-ogive J through M as indicated in Figures 7 and 20. In Figure 20 particularly we can see the upper movement stakes in relation to the apron cross-section. Seven movement stakes were also positioned on the apron below the icefall zone. Theodolite resection of these stakes has provided positions from which horizontal components of the movement have been calculated. Longitudinal horizontal components of daily movement have been calculated from a Fortran computer program (Table 1) which reconstructs the triangulation surveys. The resection data were obtained under conditions yielding a ±15 seconds of arc reproducibility, thus giving a maximum movement error of only ±0.05 m (0.16 ft). The results are given in daily movement in Tables 2, 3, and 4; with the plotted rendition of these data shown in Figure 21.

As we look at this figure we can see that the rates of daily horizontal movement do not agree with what Freers (1966) found nearly a decade ago for movement on similar stake placements on the wave-ogive zone (wave-band) in the late summer of 1961. For the interval 8-10 August, 1961, Freers calculated a movement for stake F-4 (compares to the position of 1968 stake 15) of 5.20 m/day (17.1 ft/day). For stake 15 the 1968-69 data actually give 1.06 m/day (3.5 ft/day) as a corrected mean velocity between 8 August and 7 September 1968. On Freers' stake F-7 a 1.68 m/day (5.5 ft/day) movement in 1961 is shown for the stake 10 area of this 1968 study. Again one can see that these values differ from Freers' by a factor of nearly 5. Some of Freers' movement values have been labeled by him as "questionable," due to the small triangulation angles

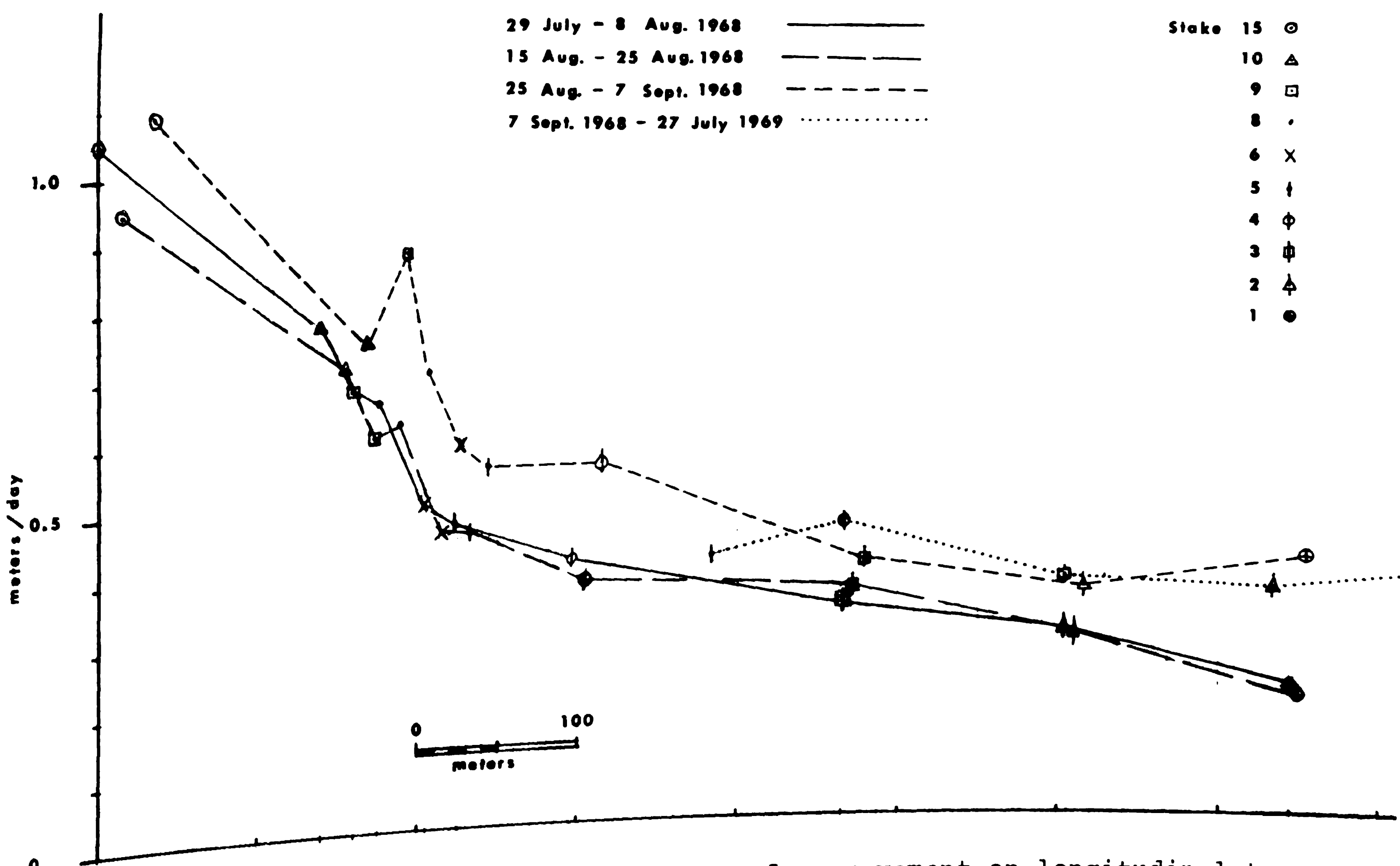


Figure 21.--Plotted horizontal surface movement on longitudinal transect below the Vaughan Lewis Icefall over selected periods of observation. Movement rate vs. relative horizontal position.

from his base line. On this basis, the writer is willing to assume that the 1968-1969 surveys are more representative, though of course there could be a slowing down of the glacier's flow over this ten-year period. This latter conclusion is deemed improbable, though, in view of the consistency of the 1968-1969 trends as revealed in Figure 21. More detailed future movement surveys will help to clarify the correct interpretation.

From Havas' 1964 data as presented by Freers (1966, Havas, 1965) surface movement of 0.35-0.42 m/day (1.15-1.38 ft/day) is noted, although the position relationship is not precisely known with respect to the stake locations for the 1968 data. Havas' strain diamonds, however, were in the upper wave-ogive zone--i.e., just below the apron area--thus giving good correlation with the computed velocities determined for 1968-1969.

Kittredge (1967) with respect to surveys made in August, 1965 states, "The velocity of the ice in the icefall itself was found to be 18 ft/day versus 2.7 ft/day at the apron" (5.49 m/day vs. 0.82 m/day). These positions are not located precisely either with respect to the 1968 positions, but Kittredge's values are partially corroborative.

Freers' (1966) calculated movement was determined from two readings two days apart; Havas' (1965) calculated movement was from two readings forty-eight days apart; Kittredge's (1967) two readings were a month apart.

In comparison to these somewhat sparse earlier data, the current investigation comprised readings taken at nine different times extending over thirteen months. The first was on 23 July 1968; the last on 23 August 1969. Thus again, the writer has more confidence in these computations. At least they are a more detailed contribution to the long-term records obtained through the years by JIRP personnel. The 1968-69 field data are listed in complete form for future reference in Appendix A.

All of these calculated movements are not shown in Tables 2 through 4 and three readings taken in 1969 do not appear in the total. The reason for this is that in 1969, some stakes were missing, by being buried under the winter snow-pack and by sliding into crevasses. Thus the computer program used in the analysis has to take this into account by substituting false data as fill-ins in the program cycle.

The initial setting of these stakes was on 25 June 1968. Due to subsequent ablation effects they were reset between the 8th and 15th of August in that year. At this time, because of distorted holes resulting from ablation, it was necessary to make slight changes in position of some stakes. For example, stakes 1 and 2 were repositioned, with stake 2 being moved $\pm$0.15 m (0.5 ft) north. (The exact change in stake 1 is not known because the original reference hole became ablated out.) Stake 4 was moved 0.45 m (1.5 ft) NNE and stake 15 was moved 4.4 m

(14.6 ft) east of its original position. A field check of the 23 July-29 July movement of stake 15 showed that a point a short distance upglacier might be useful as an indicator of the position of the first wave-ogive to be developed for the summer of 1969 (wave-ogive I). The position change of stake 15 was, therefore, to check this possibility. (It is noted that the stake 15 position in Figure 7 represents this later location. Also at this time the marker was changed from a flag and pole arrangement to a pyramid pylon as discussed early in this report, and as shown in Figure 4.) Thus by adding 4.4 m (14.6 ft) to the total movement of 3.26 m (10.33 ft) for stake 15 (v. Table 2), and calculating for a mean daily movement, one derives a horizontal movement of 1.09 m/day (3.49 ft/day), thus putting it back within the expected range of values obtained for stake 15 during other time intervals. This, again, is shown in Tables 2 through 4.

The subsequent field check of stake 15 showed movement to be about 1.6 m/day (5.8 ft/day). By early September it became apparent, even from empirical observations, that this stake had been repositioned inappropriately and would end up after a year of movement somewhat upglacier from the crest of the developing 1969 wave-ogive I. More accurate movement information determined by computer indicated that ice movement in the apron area is not as great as was originally assumed. Thus the adjustment made in stake 15 was actually in the wrong direction. It should have been moved downglacier to reposition it on the developing crest of wave-ogive I.

From data in Tables 2 through 4, it is clear that there is a change in velocity in the vicinity of the forming wave-ogive I. In corroboration, when stake 10 on the upper apron was resurveyed in 1969, a smaller velocity was calculated than for 1968. Whether this is a significant difference cannot be determined.

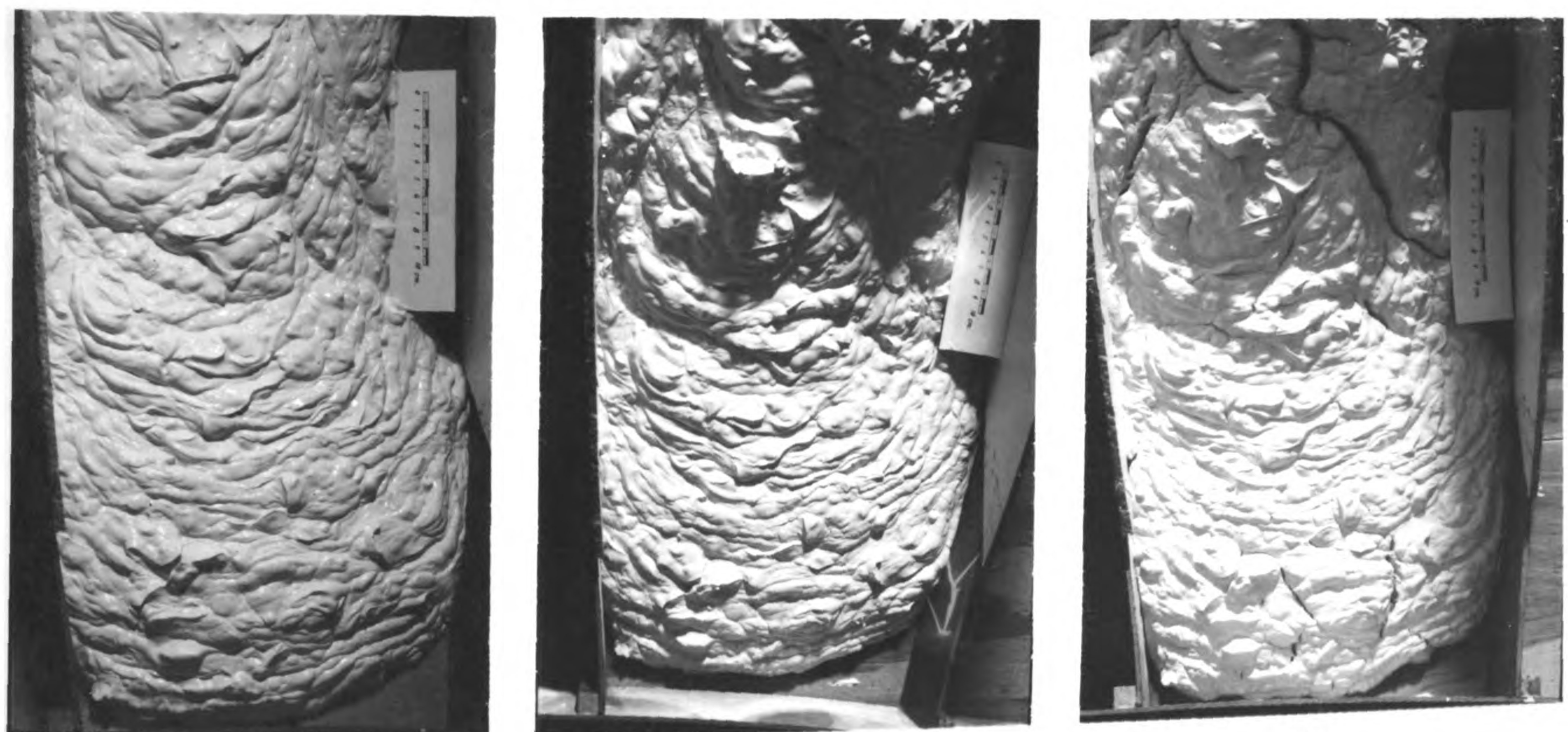
As seen in Table 4, the velocities in the bottom set of data include movement for the whole preceding year and hence give an approximation of annual surface mass transfer. Whether in fact movement in the colder months is in agreement with those found for the warmer months of July and August cannot from these records be inferred with certainty. This is because the period of comparison averages ten months (1968-1969) against two months (summer of 1968) and hence may be subject to critical question. It nevertheless tends to agree with an assumption by Nye (1967, personal communication at time of his visit to this glacier) that the velocity in such outflow glaciers may be constant within the icefall during any given year. This poses some question about ideas concerning seasonal changes in velocity as discussed by Kittredge (1967). It also enhances the current idea concerning local position changes in velocity; i.e., a consequent damming or abrupt change at the base of the icefall, thus giving further credence to the concept of longitudinal stress.

The writer does not note a significant increase in velocity downglacier as is stated by Kittredge (1967) concerning a wave-ogive leaving the compression zone and proceeding downvalley. The foregoing interpretation can, of course, be checked by more frequent periodic measurements of longitudinal flow and surface deviation changes throughout a whole year and especially surveys in the autumn, winter, and spring. In this context it is of interest that year around measurements on the flat intermediate elevation névé of the Taku Glacier (Miller, 1958, 1963b) have revealed a 10 per cent increase in surface velocity during the winter months. This indication of greatest movement in the winter months is in contrast to what may be suggested by the all too scanty data in Table 4. It may be significant, however, that the data on the Taku Glacier are from the broad upland zone of perennial accumulation at about 1075 m (3500 ft) elevation in the vicinity of Camp 10, a region of entirely different stress distribution than to be expected on the Vaughan Lewis apron area. The possibility is, however, that such seasonal variation in "flow" in the Vaughan Lewis névé, may, still be reflected in and below the icefall zone, though perhaps in an attenuated fashion. Therefore, the validity of any interpretation suggesting uniform flow should, of course, be checked by subsequent winter surveys. Nevertheless, at this juncture we are reminded of the consistency in the thirteen-month record covered by Tables 2, 3, and 4.

An explanation may be in the possibility that this represents an example of a dynamic self-regulating steady state in a classical open system--a concept now well recognized in hydrological systems (Chorley, 1962).

Though the theoretical implications of the foregoing are tantalizing, with the insufficient data we have available on this glacier to date one cannot really say much more. Because of interest in this question, however, the writer has attempted to develop a laboratory model (Figure 22a, b, and c) to simulate these flow patterns using a mixture of kaolin and water along the lines first demonstrated by Vaughan Lewis and M. M. Miller (1955). This laboratory test was carried out in a scale-model valley of the Vaughan Lewis Glacier itself. In this experiment wave-bulges not unlike those on the Vaughan Lewis apron (Figure 23) were indeed reproduced, though on a strictly qualitative basis. Details of this effort are given in Appendix B, and brief comments on the relevance of its results are made in the final section of this paper.

The periodicity of individual surface bulges below the icefall on this glacier has been considered in detail by Kittredge (1967) and others, and demonstrated via repeated plane-table surveys to be <u>annual</u> in nature. What this means is that the wave-crests of successive wave-ogives in the apron zone move downglacier at a rate approximately one full wave length per year. Thus,



(a) (b) (c)

Figure 22.--Kaolin model of the Vaughan Lewis Icefall showing wave-bulges in the apron sector in various stages of development. (a) Time zero + 0000 hr, (b) T + 0003 hr, (c) T + 1600 hr. (Photos by L. R. Miller, March 1969.)

Figure 23.--Profile view (looking south) of wave-ogive K, stake no. 3. Note pylon marker at crest. (Photo by L. R. Miller, August 1968.)

presumably this is an annual rather than a seasonal phenomenon, though it may attain its maximum development at the end of the period of maximum accumulation load, i.e., late spring. Therefore, although direct seasonal variation may not be apparent, we may conclude that there is a strong development of at least one notable new wave-bulge each year, and hence also a new wave-ogive annually.

Mass Transfer and Theoretical Considerations

Using measurements from Kittredge (1967) and also the current United States Geological Survey map at a scale of 1:63,360, a rough volumetric analysis was made of the mass transfer in the Vaughan Lewis Glacier for the year 1968. The icefall is about 671 m (2200 ft) wide, the apron 854 m (2800 ft) wide and the wave-ogive area about 732 m (2400 ft) wide. By using a conservative depth value of 198 m (650 ft) for the wave-ogive area and a velocity of 0.5 m/day (1.5 ft/day) one would find a volumetric movement of 7.15×10^5 m^3/day (2.34×10^6 ft^3/day) through a given wave-ogive cross-section. If one were to pass a mass of ice through the apron cross-section of 854 m, at the measured apron velocity of 1 m/day, one would need a depth of 76 m (250 ft) to accommodate this same volume. Such a theoretical depth is indicated by dotted line in the middle left part of Figure 20.

If the depth of ice in the apron area is greater than this theoretical depth of 76 m (250 ft), and it presumably is for reasons given below, an effective surface change would be necessary to accommodate such a volume. As there is significant summer ablation in the apron and wave-ogive zones, of course the theoretical depth of the icefall has to be greater. Thus the noteworthy change in velocity--e.g., from 1.0 m/day at stake 15 to 0.5 m/day at stake 4 in a downglacier distance of 300 m--could coincide with a strong compressional force lifting the surface upward into a wave-bulge. Such a deformation is possibly produced by a vertical stress release both by flow recrystallization and even discrete movement along foliation planes, i.e., a form of intra-foliation deformation. The accentuated pattern of foliation itself may even be a result of this phase of the deformation sequence. The upbulging represents about a 10 per cent increase in total ice thickness compared to that beneath the adjoining wave troughs (Figure 23). The wave-bulge would then be similar to the effects of tangential-compression or bench-vise squeezing described by Badgley (1965) in considering the role of orogenic stress in crustal tectonics.

A further consideration here is the possibility that the up-bulging (wave-ogive surface) is not as directly related to the presence of a bedrock threshold (v. Figure 20) as some other observers have suggested. A qualitative basis for this suggestion is given in discussion of the writer's kaolin model experiment noted in Appendix B.

In view of the foregoing discussion of the close association and possible development of tectonic folia contemporaneously with the upbulging of ice where there is an abrupt change of slope at the base of the icefall, it should be mentioned that the development of tectonic foliation has been considered by others to be prior to this stage. Miller (1968a and 1968b) considers the tectonic folia to be initiated by shear stresses at depth some distance upglacier--i.e., in the area of extending flow, within the icefall itself. These are produced first parallel to the bedrock surface, and then in a complicated flow history becoming folded and deformed and eventually to be exposed in the wave-ogives by later ablation. The mechanism he describes relates to the combined factors of rapid downvalley movement, the squeezing of ice in the defile outlet, and to the damming effects of compressive flow at the base of the icefall.

SUMMARY OF RESULTS AND CONCLUSIONS

Movement data corroborate the hypothesis of a compression mechanism existing within ice of the apron and upper wave-ogive zones. To recapitulate the results of this study, a significant change in velocity has been demonstrated to exist along the longitudinal transect between the upper apron zone at the base of the icefall and the main zone of wave-ogives. A marked change in bedrock slope (θ) translating extending (tensile) flow into compressive (compression) flow, is well illustrated by the shear formulae basic to the description of vertical velocity profiles in glacier mechanics (i.e., $\dot{\gamma}$ or rate of flow = $K\tau^n$ in which $\tau = D\rho g \sin\theta$).[1] Thus the marked change in bedrock slope (θ) presumed to be the prime factor resulting in increased longitudinal compression coincident with the foregoing velocity change. This compression effect is considered to be manifest by the described up-bulging at the surface of the glacier

[1] Where τ = shear stress in bars; D = depth of ice in cm; ρ = density; g = acceleration of gravity; and θ = surface slope angle.

in the apron sector just below the icefall. This up-bulging is further supported by the writer's laboratory model tests (Figure 22) and by the cited surface configuration shown in plotted cross-sections in the apron and wave-ogive zones.

As for the periodic nature of surface bulges in this wave-ogive system, it has been demonstrated by previous investigators on the Juneau Icefield Research Program (v. Kittredge, 1967) that the wave crests in the apron zone move downglacier approximately a full wave-length per year. Thus, although direct seasonal variations in stress may not be significant, or at least not apparent, there is presumably one new wave-bulge developed each year which eventually results in a wave-ogive downvalley (v. left hand edge of Figure 23).

These compression effects are also consistent with the strong development and orientation of the tectonic foliation described in earlier pages, further detailed as follows. The dominant shear stress which Kamb (1959a) theoretically has placed parallel to the folia and which Miller (1968) has allied as parallel to the tectonic folia on the margins and bed of this glacier, still may pertain in the icefall zone. But the conclusion of this study suggests that the dominant effect in the lobate apron zone is not shear, but compression, the direction of which lies normal to the folia surface (as shown in Figures 11-16). Thus with the related c-axis of individual ice crystals in this polycrystalline aggregate explained in this way, the c-axis would lie parallel to the longitudinal compression axes which extend radially outward from the center of the icefall where the apron begins and the wave-ogives originate. Thus further, the folia develop normal to this dominant radial stress field in this critical sector of the glacier. The existence of this significant zone of compression is also corroborated by the radial development and systematic distribution of crevasses, dominated by lobate radial fissures in the ice apron and upper wave-ogive zones.

With respect to other factors, including the always controversial role of ablation, the results of reconnaissance meteorological observations by the writer and others reveal that substantial ablation does occur. Though detailed discussion of this, however, is beyond the quantitative scope of this present report, it reveals that there is upwards of 20 m (64 ft) or more of downwasting per average ablation season. Because of the dimensional characteristics of the wave-ogives, and further because of the results of the writer's kaolin model experiment described in Appendix B, ablation is considered as no more than a subordinate co-factor in the morphogenesis of wave-ogives. In concert the various data here discussed seem to indicate that longitudinal compression plays the dominant role. If one is willing to assume that the measured changes in surface velocity are accompanied also by gradationally related deformation of ice at depth in turn related to a total stress field, then one could conceive of positive overall downvalley compression resulting in a combined but not necessarily contemporaneous continuous and discontinuous compressive deformation along either pre-developed shear foliation or (and) currently produced radial compressive foliation, with a strong upward component. To such a mechanism is attributed the striking development of folia which are suggested further to be genetically associated with the pronounced development of wave-bulges. Together these internal and surface phenomena constitute what has been described as wave-ogives.

SUGGESTED ADDITIONAL INVESTIGATIONS

Evidence indicates a strong causal relation between stresses that cause faulting and crevassing near the surface and those causing recrystallization and foliation well below the ruptured zone. Thus the ice fabric measured near the surface is indicative of fabric below this point or certainly at depth some distance upglacier from it. However, one should have some evidence that this does exist at depth and to what extent. Consideration should be given to exploration at depths greater than 18-20 m (57-64 ft). A bore-hole with core samples retrieved from depth and a pipe placed in such a bore-hole, with its deformation to be subsequently measured below the crest of a wave-ogive to a depth of 30-60 m (96-192 ft), could yield needed information. Ideally such information should be obtained in the glacier above the icefall zone as well as in the apron sector below it.

Surface strain-mechanics measurements in greater detail are also needed using the surveyed square or strain-diamond method. Such short term data will hopefully be obtained on this glacier in the future from instrumentation and measurement method currently being developed by R. W. Little, G. Cloud, and others in refinement of the technique invoked by T. Wu and R. Christensen in earlier JIRP work.

The further use of depth measuring seismic units would also yield much further data concerning details of bedrock configuration, especially in the icefall and apron zones.

The program concerning "seeded" markers should be pursued with the resurveying of markers as they are found and the evaluation of their three-dimensional movement and the role of ablation.

Very detailed micro-ablation measurements are needed using large arrays of stakes across much of the ice apron and ogive zones. Such records should extend over the melting season from April through September.

One should continue detailed periodic theodolite surface movement stake surveys, throughout the year to determine whether or not there are significant differences in surface movement from summer to winter. Also time-lapse photography would yield helpful information concerning significant seasonal changes in the icefall and possibly even direct evidence of "up-bulging" of the wave-ogives. This kind of information can be obtained without great expenditures of money. An expanded program of P-30 mapping for volume comparisons also should be invoked.

Added to the above, is the excellent potential for significant results via either laser range finder, or (and) micro-wave surveying equipment using continuous 24-hour-a-day monitoring of surface glacier flow. Allied to this could be continuous recording, portable seismograms in the ice, as well as in adjoining bedrock, to delineate discontinuous or spasmodic flow for comparison with the aforementioned continuous micro-wave survey data. Such would well lead to a break-through in understanding the true nature of continuous versus discontinuous processes in glacier mass transfer.

Basic to the success of any such research projects, is a small yet dedicated team of field men deeply interested in this fascinating phase of glaciology. This team must, however, be willing to spend a consecutive six-month span from spring to autumn while conducting the kind of systematic observations and measurements suggested here, and of course in much more detail than the present study has permitted.

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TABLES

TABLE 1.--Vaughan Lewis Glacier Horizontal Movement Program, 1-20-70.

```
100     REAL METER
110     DIMENSIØN FILE (9)
112     INTEGER DAY (8)
113     DATA DAY/6,10,7,10,11,325,4,22/
120     FILE (1) = "D1  "
130     FILE (2) = "D2  "
140     FILE (3) = "D3  "
150     FILE (4) = "D4  "
160     FILE (5) = "D5  "
170     FILE (6) = "D6  "
180     FILE (7) = "D7  "
190     FILE (8) = "D8  "
200     FILE (9) = "D9  "
230     J = 1
240     DØ 80 I = J,8
250     PRINT,+,"        FT          M          FT/DAY       M/DAY"
260     RFILE=FILE(I)
270     SFILE=FILE(I+1)
280     CALL ØPENF (1,RFILE)
290     CALL ØPENF (2,SFILE)
300 10  READ (1,,END=70),AD2,AM2,AS2,AD1,AM1,AS1,BD1,BM1,BS1,BD2,BM2,BS2
310     READ (2,,END=70),CD2,CM2,CS2,CD1,CM1,CS1,DD1,DM1,DS1,DD2,DM2,DS2
320      A1=AD1+((AM1+(AS1/60))/60)
330      A2=AD2+((AM2+(AS2/60))/60)
340     ALPHA=180-((A1+A2-180)/2)
350      B1=BD1+((BM1+(BS1/60))/60)
360      B2=BD2+((BM2+(BS2/60))/60)
370     BETA=(B1+B2-180)/2
380     C=330.0
```

```
    RADIAN=57.2957795
    GAMMA=180-(ALPHA+BETA)
    CØN=SIN(GAMMA/RADIAN)/C
    A=SIN(ALPHA/RADIAN)/CØN
    B=SIN(BETA/RADIAN)/CØN
    C1=CD1+((CM1+(CS1/60))/60)
    C2=CD2+((CM2+(CS2/60))/60)
    ALPHA2=180-((C1+C2-180)/2)
    D1=DD1+((DM1+(DS1/60))/60)
    D2=DD2+((DM2+(DS2/60))/60)
    BETA2=(D1+D2-180)/2
    GAMMA2=180-(ALPHA2+BETA2)
    CØN2=SIN(GAMMA2/RADIAN)/C
    B2=SIN(BETA2/RADIAN)/CØN2
    A2=SIN(ALPHA2/RADIAN)/CØN2
    X=ALPHA-ALPHA2
    Z=SQRT(B↑2+B2↑2-2*B*B2*CØS(X/RADIAN))
    METER=(Z*.3048)
    DZ = Z/DAY(I)
    DMETER = METER/DAY(I)
    PRINT 60,Z,METER,DZ,DMETER
60  FØRMAT (4F10.2)
    GØ TØ 10
70  CALL CLØSEF (1)
    CALL CLØSEF (2)
80  CØNTINUE
    END
```

TABLE 2.--Movement Computed for the Periods 23 July 1968--29 July 1968 and 29 July 1968--8 August 1968.

Vaughan Lewis Glacier Movement
23 July 1968--29 July 1968

STK. NØ.	FT	M	FT/DAY	M/DAY
1	4.39	1.34	0.73	0.22
2	7.47	2.28	1.25	0.38
3	8.73	2.66	1.46	0.44
4	11.43	3.48	1.90	0.58
5	11.19	3.41	1.86	0.57
6	12.72	3.88	2.12	0.65
8	17.01	5.18	2.83	0.86
9	16.71	5.09	2.79	0.85
10	17.97	5.48	3.00	0.91
15	24.77	7.55	4.13	1.26

29 July 1968--8 August 1968

STK. NØ.	FT	M	FT/DAY	M/DAY
1	6.62	2.02	0.66	0.20
2	9.66	2.94	0.97	0.29
3	10.83	3.30	1.08	0.33
4	13.39	4.08	1.34	0.41
5	15.53	4.73	1.55	0.47
6	16.42	5.00	1.64	0.50
8	21.70	6.61	2.17	0.66
9	22.08	6.73	2.21	0.67
10	25.35	7.73	2.53	0.77
15	34.56	10.54	3.46	1.05

TABLE 3.--Movement Computed for the Periods 8 August 1968--15 August 1968 and 15 August 1968--25 August 1968.

Vaughan Lewis Glacier Movement
8 August 1968--15 August 1968

STK. NØ.	FT	M	FT/DAY	M/DAY
1	18.00	5.49	2.57	0.78
2	14.00	4.27	2.00	0.61
3	8.71	2.65	1.24	0.38
4	13.35	4.07	1.91	0.58
5	10.06	3.07	1.44	0.44
6	12.71	3.87	1.82	0.55
8	14.16	4.32	2.02	0.62
9	16.47	5.02	2.35	0.72
10	16.57	5.05	2.37	0.72
15	10.68	3.26	1.53	0.47

15 August 1968--25 August 1968

STK. NØ.	FT	M	FT/DAY	M/DAY
1	5.86	1.79	0.59	0.18
2	9.22	2.81	0.92	0.28
3	11.70	3.56	1.17	0.36
4	12.23	3.73	1.22	0.37
5	15.01	4.58	1.50	0.46
6	15.23	4.64	1.52	0.46
8	20.52	6.25	2.05	0.63
9	20.07	6.12	2.01	0.61
10	23.40	7.13	2.34	0.71
15	31.19	9.51	3.12	0.95

TABLE 4.--Movement Computed for the Periods 25 August 1968--7 September 1968 and 7 September 1968--27 July 1969.

Vaughan Lewis Glacier Movement
25 August 1968--7 September 1968

STK. NØ.	FT	M	FT/DAY	M/DAY
1	4.81	4.52	1.35	0.41
2	3.29	4.05	1.21	0.37
3	4.87	4.53	1.35	0.41
4	[illegible]0.28	6.18	1.84	0.56
5	[illegible]0.19	6.15	1.84	0.56
6	[illegible]0.82	6.35	1.89	0.58
8	5.84	7.88	2.35	0.72
9	1.71	9.67	2.88	0.88
10	6.96	8.22	2.45	0.75
15	9.51	12.04	3.59	1.09

7 September--27 July 1969

STK. NØ.	FT	M	FT/DAY	M/DAY
1	417.53	127.26	1.28	0.39
2	385.74	117.57	1.19	0.36
3	397.09	121.03	1.22	0.37
4	487.52	148.60	1.50	0.46
5	451.00	137.47	1.39	0.42
6	MISSING DATA			
8				
9				
10				
15				

APPENDICES

APPENDIX A

VAUGHAN LEWIS GLACIER COMPUTER PROGRAM INPUT DATA, TAKEN FROM SURVEY POINTS FFGR #5 (UPPER) AND FFGR #4 (LOWER) ON CLEAVER WEST OF CAMP 18

APPENDIX A

Part 1

VAUGHAN LEWIS GLACIER COMPUTER PRØGRAM INPUT DATA

MØVEMENT STAKE
NUMBER X 10

DEG,MIN,SEC,DEG,MIN,SEC,DEG,MIN,SEC,DEG,MIN,SEC

UPPER SURVEY PØINT — FFGR # 5 — RIGHT, LEFT

LØWER SURVEY PØINT — FFGR # 4 — RIGHT, LEFT

DATA 1 (D1) 23 JULY 1968

```
10  341,26,27,161,26,24,159,28,27,339,28,33
20  336,23,59,156,23,32,153,37,13,333,37,02
30  331,36,30,151,36,30,147,49,01,327,48,42
40  327,44,34,147,44,32,142,18,51,322,19,01
50  318,26,04,138,26,17,132,05,00,312,05,13
60  316,20,11,136,20,24,129,36,51,309,36,47
80  312,21,56,132,22,22,125,03,48,305,03,33
90  310,27,38,130,28,04,122,48,38,302,48,29
100 308,30,19,128,30,40,120,29,31,300,29,34
150 298,59,31,118,59,28,109,00,32,289,00,12
```

D2 29 JULY 1968

```
10  341,30,11,161,30,00,159,32,42,339,32,34
20  336,29,19,156,28,56,153,43,29,333,43,22
30  331,44,30,151,44,17,147,58,19,327,58,14
40  327,57,03,147,56,48,142,34,39,322,34,22
50  318,40,18,138,40,28,132,22,06,312,22,10
60  316,35,08,136,34,59,129,55,07,309,54,48
80  312,42,03,132,42,03,125,28,12,305,28,12
90  310,48,47,130,48,49,123,14,04,303,14,10
100 308,53,26,128,53,30,120,57,35,300,57,37
150 299,37,12,119,37,38,109,45,54,289,46,05
```

Part 2

D3

8 AUGUST 1968

10	341,36,29,161,36,21,159,39,45,339,39,37
20	336,37,58,156,37,58,153,53,45,333,53,23
30	331,56,47,151,56,31,148,12,38,328,12,18
40	328,18,03,148,17,12,142,59,05,322,58,22
50	319,03,14,139,03,11,132,48,21,312,48,32
60	316,59,54,136,59,42,130,23,28,310,23,16
80	313,14,09,133,14,25,126,05,30,306,05,30
90	311,22,14,131,22,29,123,52,54,303,52,50
100	309,30,32,129,31,20,121,41,02,301,41,58
150	300,38,39,120,38,46,110,55,41,290,55,33

D4

15 AUGUST 1968

10	341,40,37,161,40,25,159,44,48,339,44,55
20	336,46,17,156,46,04,154,03,28,334,03,17
30	332,06,32,152,06,29,148,24,03,328,23,42
40	328,36,01,148,35,37,143,20,51,323,20,28
50	319,18,24,139,18,24,133,05,36,313,05,54
60	317,18,52,137,18,41,130,45,20,310,45,07
80	313,34,52,133,34,39,126,29,22,306,29,24
90	311,47,05,131,46,58,124,21,33,304,21,24
100	309,58,02,129,57,52,122,11,54,302,11,52
150	300,58,28,120,58,20,111,17,21,291,17,31

D5

25 AUGUST 1968

10	341,46,26,161,46,23,159,51,23,339,51,21
20	336,55,08,156,55,12,154,13,45,334,13,28
30	332,18,19,152,18,43,148,37,50,328,38,04
40	328,54,21,148,54,29,143,42,32,323,42,28
50	319,39,15,139,39,18,133,30,28,313,29,56
60	317,40,58,137,40,42,131,11,03,311,10,43
80	314,05,51,134,05,31,127,05,11,307,04,41
90	312,19,03,132,18,52,124,58,06,304,52,55
100	310,32,37,130,32,34,122,52,34,302,52,15
150	301,53,52,121,53,44,112,20,29,292,20,05

Part 3

D6

7 SEPTEMBER 1968

```
10   341,54,50,161,54,40,160,01,09,340,00,51
20   337,07,20,157,07,08,154,27,34,334,27,18
30   332,35,08,152,34,54,148,57,14,328,56,55
40   329,21,03,149,20,46,144,14,42,324,14,34
50   320,07,53,140,07,34,134,03,26,314,03,19
60   318,12,26,138,12,23,131,47,29,311,46,29
80   314,43,09,134,43,07,127,49,00,307,48,03
90   312,57,45,132,57,34,125,43,18,305,42,30
100  311,15,46,131,15,37,123,41,46,303,40,52
150  303,02,54,123,02,42,113,39,48,293,38,30
```

D7

27 JULY 1969

```
10   344,26,12,164,25,54,162,58,12,342,58,51
20   341,10,04,161,09,23,159,09,40,339,10,37
30   338,20,10,158,19,39,155,42,07,335,42,47
40   332,48,42,152,48,26,149,06,54,329,07,40
50   329,41,27,149,41,09,145,18,59,325,19,26
60   327,50,01,147,50,01,143,50,01,323,50,01
80   326,58,02,146,58,08,142,12,06,322,12,50
90   326,01,06,146,01,09,141,03,20,321,03,47
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150  319,51,46,123,05,01,133,44,21,293,40,01
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D8

1 AUGUST 1969

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20   341,12,28,161,12,19,159,13,23,339,12,50
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50   329,47,37,149,27,54,145,27,21,325,27,15
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80   327,06,11,147,06,01,142,21,30,322,22,03
90   326,09,11,146,09,29,141,13,14,321,13,10
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Part 4

D9 23 AUGUST 1969

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30	338,42,22,158,42,28,156,07,44,336,07,55
40	338,38,03,158,38,14,155,27,45,335,27,47
50	333,20,27,153,20,29,149,44,28,329,44,20
60	330,19,37,150,19,35,146,04,10,326,04,06
80	330,19,36,150,19,34,146,04,09,326,04,05
90	326,47,02,146,47,19,141,57,05,321,57,08
100	325,44,54,145,45,05,140,40,27,320,40,35
150	323,54,11,143,54,22,138,13,35,318,13,44

LIST ØF FALSE DATA - SUBSTITUTING FØR BLANKS FØR COMPUTER PRØGRAMING.

27 JULY 1969
D7 STAKE 6
STAKE 7

1 AUGUST 1969
D8 STAKE 6
STAKE 7

23 AUGUST 1969
D9 STAKE 7

APPENDIX B

LABORATORY SIMULATION OF VAUGHAN LEWIS ICEFALL SHOWING MECHANICALLY DEVELOPED WAVE-BULGES IN A VISCO-PLASTIC MATERIAL IN A SCALE MODEL

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Insofar as the movement of ice in glaciers is partly a mechanical phenomenon, one can refer the general mode of flow to models (Lewis and Miller, 1955). Although the peculiar rheological properties of ice derives fundamentally from the unique properties of water, the internal bodily deformation of glacial ice is not in fact a fluid phenomenon. Rather it is an "elasto-plastic" flowage phenomenon involving in part the reorientation (flow recrystallization) of individual ice crystals in a polycrystalline aggregate, and related shear displacement within and between the ice crystals both idio-molecularly and on discrete movement surfaces. The microscope manifestation of this complex deformation process is often an ice fabric and even distinct foliation. Microscopic effects of glacial deformation are often shown by changes in surface configuration and large scale fractures.

In an effort to demonstrate the role of the latter in a scale model of the Vaughan Lewis Glacier, a brief laboratory experiment was conducted. The "flow" material employed in the model was a mixture of kaolin (pure china clay) and water. This material was not elasto-plastic as is glacier ice in reality, but more truly visco-plastic as revealed by the tests.

In this laboratory simulation wave-bulges were produced simular to those observed on the Vaughan Lewis Icefall, Alaska. The significant aspect here is that they were produced without any bedrock "threshold" influence. Possibly more important, they were produced without any ablation effects to modify or exaggerate the effects. (The possible significance of this has been noted briefly on page 55 of the text.)

In nature wave-ogives on the Vaughan Lewis Glacier are periodic formations, though no attempt was made to make a time comparison with the model. In this simulation gravity was the only force involved beyond slight basal and marginal friction effects. In nature the angle of descent of the glacier material varies between 24° and 40°, therefore in the model a mean slope angle of 40° was used. The model was built at a scale of 1:2400 which puts the value well within the accepted range of kaolin model glaciers given by Lewis and Miller (1955), viz., 1:100 to 1:10,000. It reconstructed as well as possible the icefall and valley configuration from the constraints of width and slope angle. It is important to note that no bedrock "threshold" or rise in slope was built into the model downvalley from the headwall sector. Statistics and dimensional values used in this laboratory model were as follows: width of original 600 m, width of model 25 cm, model ratio 2.4 x 10^3; length of "fall" 44 cm; depth of kaolin 5-8 cm.

Although measurements were made of volume and density to stay within the range of accuracy required of a model, no attempt was made to make a quantitative measurement of the results.

The purpose of the experiment was strictly to condense the dynamic morphogenesis of wave-ogives into a small unit of time, hopefully giving more insight into possible related and nonrelated events leading to a better understanding of the surface bulges on the Vaughan Lewis Glacier as seen in the field. The model proved to be a helpful tool for demonstrating the mechanical development of wave-bulges, without any influence from ablation or the presence of a "classical" type bedrock threshold.

Among the features observed during flow were transient bergschrunds, tensional crevasses, compressional crevasses, splaying crevasses, and wave-bulges. After flow ceased most of the "crevases" resealed as a result of the "viscous" or visco-plastic quality of the mixture, hence, the similarity in this respect was not quite like a glacier. But surface irregularities and wave-bulges did persist, quite similar to a glacier.

Two puzzling morphogenic processes cited by previous authors in reference to the Vaughan Lewis Glacier were observed on the model: (1) small folds forming in the apron area coalesced to form the crest of a wave-bulge (simulated wave-ogive as seen in nature); and (2) no transverse crevasses developed in the lower icefall and apron sectors. This latter aspect may, of course, be related to the more viscous character of the material.

Again, however, the significance of the simulation is the strong evidence of a sequence of surface irregularities in the apron area quite similar to what is seen in the field on the actual Vaughan Lewis and Upper Herbert Glaciers; plus the very distinct development of wave-bulges below the apron. The significance of the latter morphologic unit is that these bulges appear quite similar to those relating to the wave-ogives on the Vaughan Lewis and Upper Herbert Glaciers, and that they developed without benefit of a downvalley bedrock "threshold" (v. discussion relating to possible relevance in Vaughan Lewis Glacier situation as noted in the text, page 54).

GLOSSARY OF SELECTED GLACIOLOGICAL TERMS

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Ablation. The wasting or surface-lowering of a glacier by the combined processes of melting, evaporation, and sublimation. On the Vaughan Lewis Glacier ablation is primarily by melting.

Albedo. The percentage of incident radiation that is reflected; hence, the degree of whiteness.

Bergschrund. The crevasse occurring at the head of a glacier or margins of an icefield, which separates the moving firn and ice of the glacier from the relatively immobile firn and ice adhering to the headwall or nunatak. This crevasse commonly penetrates to the headwall or bed of the glacier. It usually has a lower lip on the downvalley side.

Berm. Terraces which originate from the interruption of an erosion cycle. Remnants of earlier valley floor above present glacier surface--usually in bedrock; i.e., a bedrock sholder.

Bubbly glacier ice. The main material of glaciers variably containing air pockets and entrapped water bubbles, and having a density approximating 0.88-0.90 gm/cm^3.

Diagenesis. A process of across-stratum changes which take place in firn due to accumulation above it, or percolation of rain and melt-water through it; i.e., via compaction and recrystallization, and by refreezing of liquid water in selective zones.

Firn. Compacted, granular, but still pervious material metamorphosed from old snow, and in transition to glacier ice. Characterized by a density approximating 0.50-0.75 gm/cm^3.

Névé. An area of old snow partly converted into ice (i.e., area of firn). The net accumulation area of a temperate glacier above the névé-line.

Névé-line. The elevation of the most stable position of the lower limit of firn or the névé. The demarcation line dividing the areas of net accumulation and negative dissipation or wastage.

Ogive. See wave-ogive.

Regime. The material balance (budget) of a glacier involving the total accumulation and the gross wastage in one or more budget years. The long-term state of health of a glacier.

Seracs. When two or more sets of crevasses intersect, the surface of the glacier is torn into a broken mass of jagged ice pinnacles known as seracs.

Tectonic foliation. Planar appearing structures produced by shear or compression and consisting of alternate layers of fine and coarse-grained ice. In some cases grains of ice and trapped air bubbles are elongated in the plane of the layer. May not always represent discrete zones of discontinuous shear but rather differential zonation of continuous (quasi-plastic) deformation.

Wave-band. See wave-ogive.

Wave-bulge. See wave-ogive.

Wave-ogive. The curved light and dark bands, arched or convex in plan view downglacier, that in cross-section show a pronounced surface wave with measurable wave length and amplitude and are that found below icefalls on some glaciers. Where the wave character dominates, the term wave-band or wave-bulge is applied; and where the surface wave is not apparent and the inlaid three-dimensional or englacial structures (usually pronounced tectonic foliation) dominate, the term ogive is applied. The alternating dirty and light or clean ice zones not yet satisfactorily explained, though appear to relate to alternations between originally summer versus winter snow.

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