PRELIMINARY DATA FROM SASKATCHEWAN GLACIER, ALBERTA, CANADA[1]

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An understanding of glacier flow is not likely to come from field or laboratory studies alone; both are required. Needed from the field are quantitative data on the movement, size, shape, slope, and thickness of flowing ice bodies. Reliable measurements of the amount and absolute direction of movement at points widely distributed over the surfaces of glaciers are sparse. Even the surface velocity map of Hintereisferner (Hess, 1904, p. 136), one of our most thoroughly studied ice streams, shows only the downvalley component of movement. Quantitative data on velocity distribution in depth are even more meagre (Gerrard, Perutz and Roch, 1952; Sharp, 1953), although they are essential to an understanding of the physical properties and behaviour of ice deep within a glacier and to the setting up and interpretation of laboratory experiments. Thickness is known in reasonable detail for a few European valley glaciers (Hess, 1933, p. 43; Renaud and Mercanton, 1948) and along one profile in the Greenland Ice Sheet (Joset, 1950, p. 54), but it has been determined only locally in North American ice bodies (Goldthwait, 1936, pp. 506–7; Poulter, Allen, and Miller, 1949; Littlewood, 1952, p. 124; Allen and Smith, 1953, p. 758; Röthlisberger, 1953, pp. 234–7).

In reaching a satisfactory explanation for the dynamic behaviour of glaciers, the following are a few of the many questions that must be answered. What are the effects in quantitative terms of surface slope, ice thickness, and channel characteristics on flow velocity? What is the influence on velocity of seasonal or other temporal changes in the climatological environment, and what is the exact mechanism by which these changes affect the flow rate? Are variations of velocity within a glacier reasonably uniform in time and space or are they notably heterogeneous and irregular? Do all stream-lines through a cross section accelerate or decelerate in phase, and if not why not? Is there any consistent genetic relation between the mechanics of flow and the various structures visible in glaciers? What orientation patterns do ice crystals in glaciers show, how are these orientations produced, and what relation do they bear to the mechanics of flow? What are the conditions and magnitudes of stress and strain deep within glaciers, and what are the physical properties and behaviours of ice at depth? The present study hopes to provide information bearing on some of these questions.

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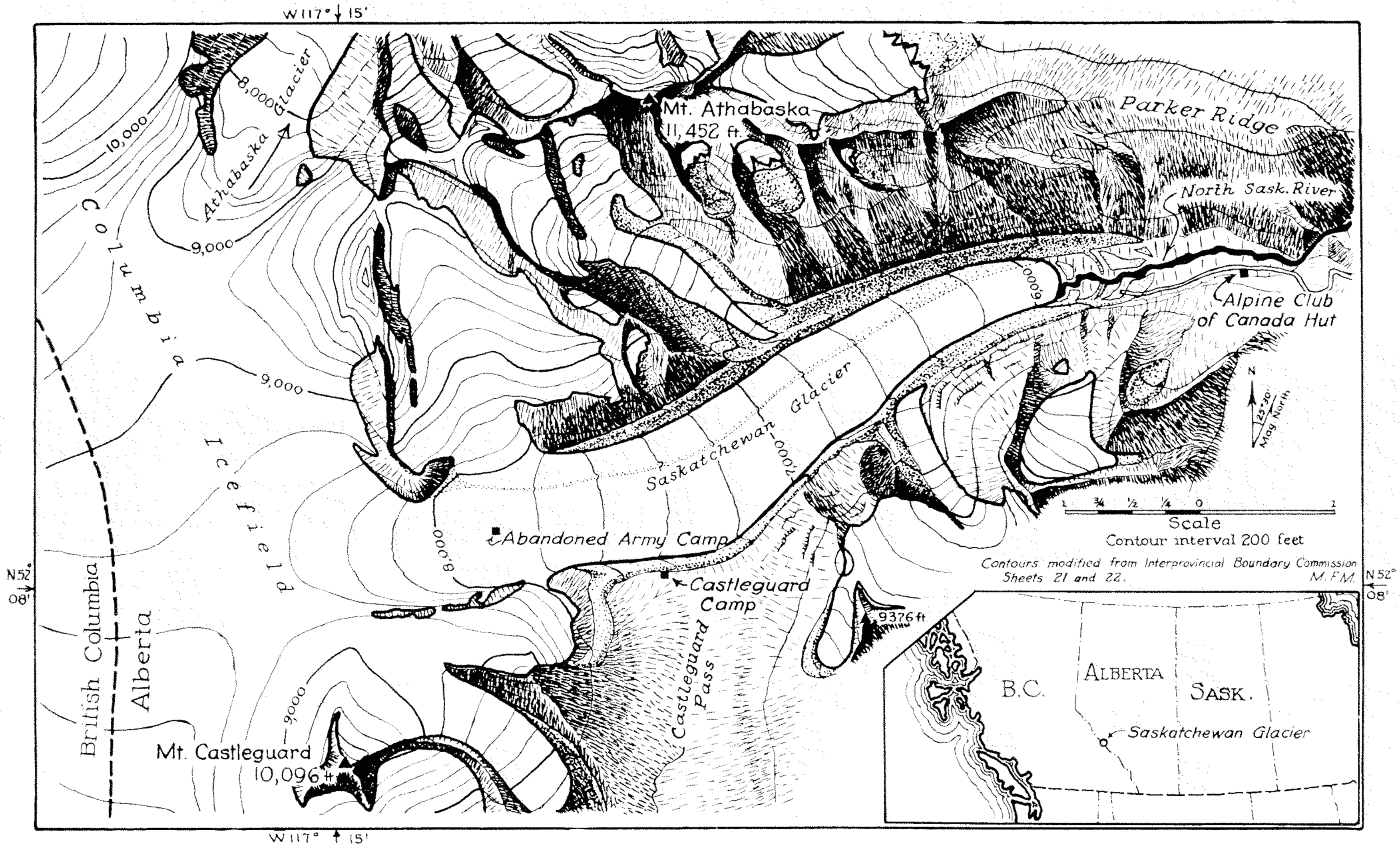

Fig. 1. Sketch-map of Saskatchewan Glacier and surroundings.

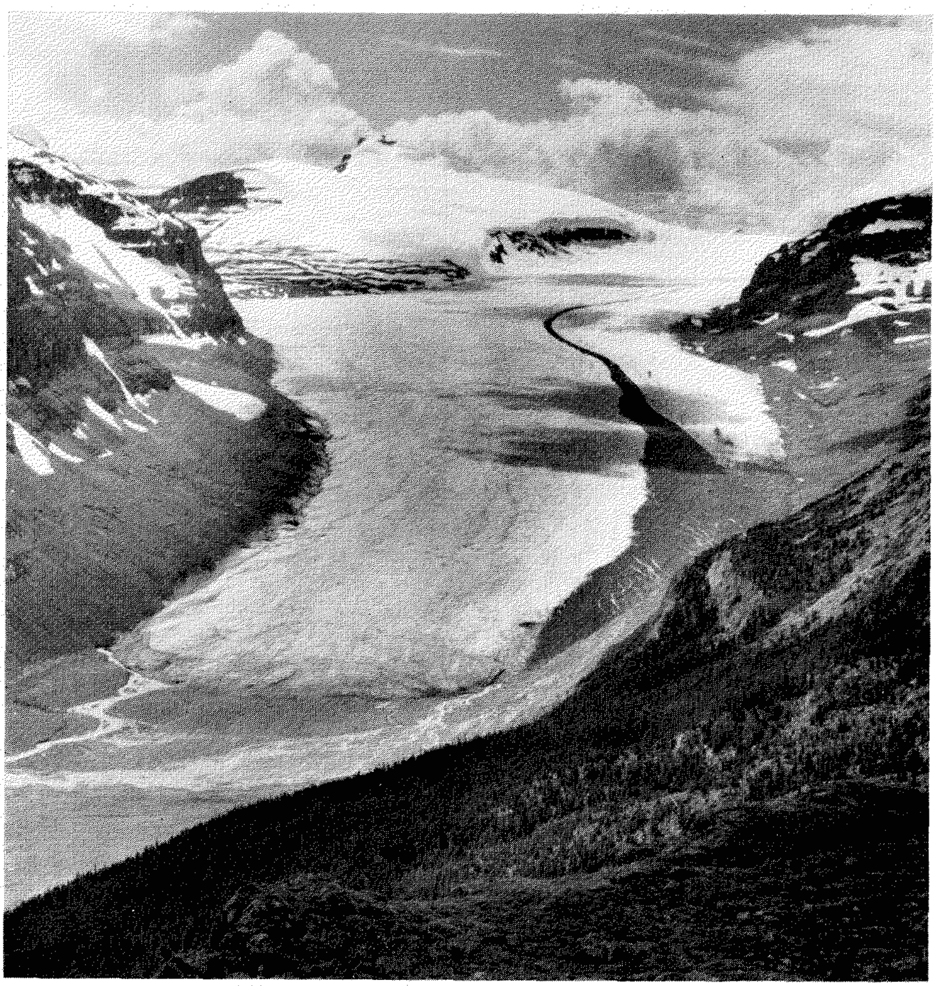

Fig. 2. Looking west-southwest up Saskatchewan Glacier from Parker Ridge, 11 July 1952. Glacier flows from Columbia Icefield partly obscured by rock spur on right skyline. Mt. Castleguard on centre skyline.

This is a progress report on an active project. The investigations are incomplete, the data obtained have not been fully analysed, and the interpretations offered are necessarily preliminary and tentative. Work on Saskatchewan Glacier was initiated in the summer of 1952 under a grant from the Arctic Institute of North America[1] with equipment provided by the U.S. Office of Naval Research. The work continued in 1953 with Office of Naval Research support, and will be carried further in 1954 on the same basis. The government of Canada graciously granted permission for the field study, and the National Parks Administration of Canada has consistently aided the work. Warden William Black of that organization was especially helpful. The

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following persons have served in field parties, and their assistance is warmly acknowledged: Clarence R. Allen, B. Gunnar Bergman, James E. Conel, Donald B. Emerson, Ben F. Jones, Lee R. Magnolia, Jack W. Rocchio, and Gordon D. Seele. The courteous cooperation of Canadian Customs officials, the privilege of storing equipment in a hut owned by the Alpine Club of Canada, and the loan of two toboggan sleds from the Defence Research Board of Canada are acknowledged with appreciation. The Parsons Survey Company of South Gate, California, kindly supplied without charge an inclinometer for survey of bore holes.

Physical setting

Saskatchewan Glacier flows eastward out of Columbia Icefield which lies astride the British Columbia–Alberta border in the Rocky Mountains at the boundary between Banff and Jasper national parks (52°08N., 117°12W.). The Saskatchewan is a geometrically simple ice stream emerging from the ice field at an altitude of 8,500 feet and flowing a little less than 6 miles through a steep-walled valley to a terminus at 5,900 feet (Fig. 1). Its firn limit is near 8,000 feet. The trunk glacier receives only one tributary which enters from the north at about 8,000 feet altitude. Two former tributaries approach the trunk glacier closely on each side about 2 miles above its terminus but do not attain a junction. The low, gently sloping terminus and a relatively smooth ice surface afford easy access to and movement across the glacier (Fig. 2). Camps were maintained near the Alpine Club hut, one mile below the glacier, and outside the south lateral moraine (Castleguard Camp) 4 miles up the glacier (Fig. 1).

Dimensions of the glacier

The width of Saskatchewan Glacier is a little more than one mile in the Castleguard sector, and it tapers to 0.5 mile near the terminus. The surface slope is about 650 feet per mile as the glacier emerges from the ice field, it is 375 to 400 feet per mile over most of the next 5 miles with some gentler and steeper reaches, and it steepens to 840 feet per mile near the terminus. Transverse profiles across the glacier are relatively flat, except for marginal ablation depressions 75 feet deep.

Seismic reflections obtained by C. R. Allen provide information on ice thickness and configuration of the subglacial floor. A portable 6-trace seismic set was used. The shots, consisting of a small fraction of an ounce of dynamite on an electric cap, were detonated in water-filled holes of 1.5 inches diameter bored 25 to 30 feet into the ice by hand drills, as described elsewhere (Allen and Smith, 1953, pp. 756–7). The shotpoints were arranged along two transverse profiles in the upper reach of the glacier and along a longitudinal profile extending to the terminus (Fig. 3).

The greatest thickness recorded, 1,450 feet, is near the centre-line about 5 miles above the terminus. This is not the maximum, for the subglacial floor here slopes gently upvalley and 15 degrees toward the south wall. The south wall of the valley must be extremely steep for probable thicknesses of 1,050

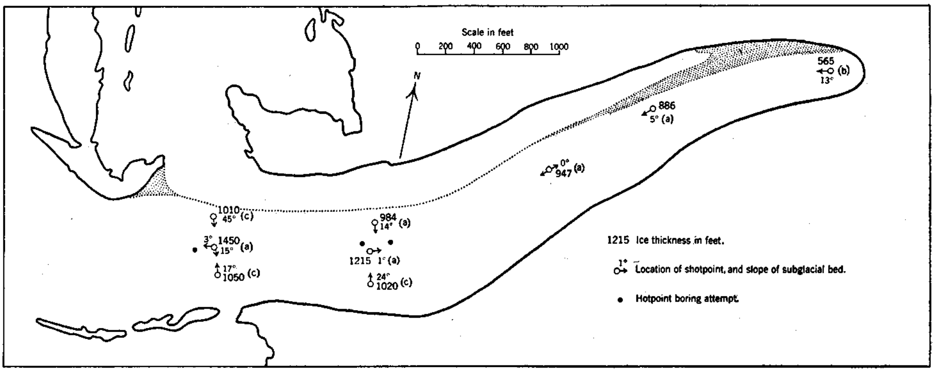


Fig. 3. Map of seismic shotpoints, ice thickness, and slope of subglacial floor, from data obtained by C. R. Allen in 1952: a) good reflection, no doubt about interpretation; b) moderate reflection, some doubt; c) poor reflection, considerable doubt.

and 1,020 feet were recorded within 1,100 feet of the glacier's margin. Thickness decreases progressively along the longitudinal profile toward the terminus, but more slowly than might be expected. A thickness of 565 feet recorded 1,000 feet from the terminus may include some unconsolidated till or other deposits between the ice and the bedrock floor. The flat and, in places, upvalley slope of the bedrock floor at the reflection points along the longitudinal profile affords further testimony to the effectiveness of glacial excavation in overdeepening valleys. After more figures on thickness have been obtained, and the gathering and processing of data on surface slope and velocity have been completed, it may be possible to provide a quantitative evaluation of the relation between surface velocity and ice thickness in this temperate valley glacier.

Observations of surface velocity

Measurements of surface velocity on glaciers usually record only the downvalley component of movement, although it has long been recognized that the motion is not parallel to the surface at most places (Hess, 1933, p. 41). By an excellent bit of deduction, Reid (1896, pp. 917–9) showed that the movement should be obliquely downward in the accumulation area and obliquely upward in the ablation area, and he subsequently confirmed this by actual measurements (1901, p. 750). It follows from flow-lines based on this relation that ice exposed at the surface should be progressively older from the firn limit to the terminus. Reid's deductions have not received the further study they deserve, and observations of absolute direction of motion in glaciers are so few and so scattered as to be more tantalizing than informative (Hess, 1933, p. 41). One objective of the present study is a determination of the amount and absolute direction of motion at many points on the glacier's surface.

When calculations are completed, the movement data from the Saskatchewan should show: (1) the change of velocity along a longitudinal centre-line, thus giving some quantitative measure of compressive flow (Nye, 1952, p. 89),

(2) the actual magnitude of surfaceward movements below the firn limit and the change in angle between the flow vector and the surface along a longitudinal profile, (3) changes of surface velocity in relation to ice thickness, surface slope, and channel cross section, (4) possible convergences or divergences of flow-lines in plan view, (5) velocity differences between individual ice streams within the glacier, (6) variations in winter and summer flow rates in different parts of the ice tongue, and (7) comparative daily velocities at different points on the glacier surface.

Procedure. A total of 52 movement stations has been established on the surface of Saskatchewan Glacier. Each consists of a one-inch wooden dowel 10 feet long set snugly in a hole bored 8 to 9 feet into the ice. Dowels are reset periodically as ablation demands, and all are reset at the end of summer so that most survive the winter in upright position. These stations are arranged in a longitudinal centre-line profile extending from the terminus to above the firn limit and in 8 complete or partial transverse profiles spaced at intervals along the ice tongue (Fig. 4).

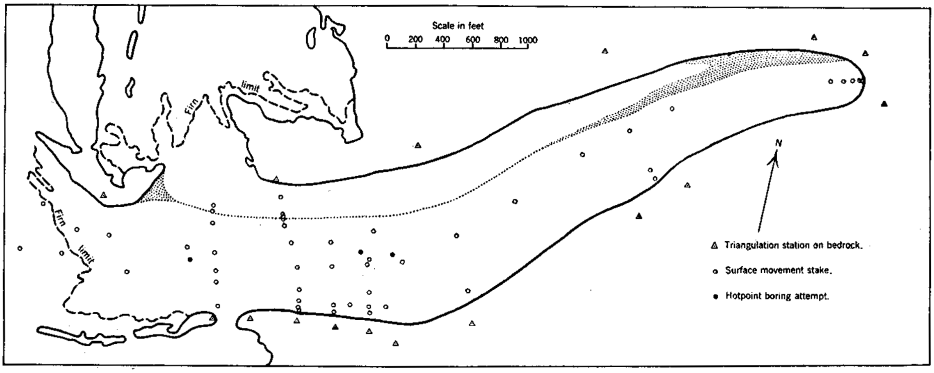


Fig. 4. Map of triangulation stations, surface movement stakes, and hotpoint boring sites.

Readings are made to a fixed mark on each pole by triangulation from a number of established stations on bedrock or stable deposits on the valley walls. Observations in 1952 were by transit and gave locations accurate to ± 0.2 feet horizontally and to ± 0.6 feet vertically for the most distant station, 4,000 feet. Most readings were over a much shorter distance, and the accuracy was considerably greater. A Wild T-2 theodolite used in 1953 gave vertical angles with accuracies several orders higher. Stations were generally observed every two weeks during the summers of 1952 and 1953.

Results to date. Reduction of these triangulation data requires laborious calculations not yet made, but inspection of the readings indicates that movements within the glacier are indeed oblique to the surface. The downvalley component for a few selected stations has been computed from the 1952 data. The maximum velocity recorded was in the mid-glacier position a mile below the firn limit, but it amounted to an average of only 1.2 feet per day over a six-week period. Velocities along the mid-glacier stream-line decreased in a downvalley direction and most markedly in reaches of gentlest surface slope. This produces a shortening of the longitudinal distance between points on the glacier surface, which for the stakes measured amounted to 0.7 foot per day in a distance of 7,600 feet. If velocity is assumed constant over the year, this amounts to 260 feet annually, or a shortening of 3.4 per cent. This shortening could result from surfaceward movement at increasingly steeper angles downvalley, and it is compatible with the concept of compressive flow (Nye, 1952, p. 89). Movement with an increasing vertical component downvalley enables the glacier to maintain its surface profile and elevation in face of increasing ablation.

It was also found that the highest surface velocity gradient along a transverse profile, that is the greatest change in velocity per unit transverse distance, occurs within a 300-foot zone at the glacier's margin. On a transverse profile in the Castleguard sector the velocity 50 feet in from the lateral moraine was only 18 per cent of the maximum, but 100 feet farther in it was nearly 50 per cent. Furthermore, little difference in velocity, averaged through a 6-week period, could be detected across a zone 1,000 feet wide in the centre. A similar velocity distribution is common in valley glaciers, being more marked in some than in others (Hess, 1904, pp. 119, 124; Klebelsberg, 1948, p. 83; Nielsen, 1953, p. 10). An ice stream like the Saskatchewan with steep valley walls and a probable thickness of 1,000 feet close to the edges should have a steeper marginal velocity gradient than a similar ice stream flowing in a channel with gently sloping sides. Extreme marginal velocity gradients are found in glaciers experiencing *Block Schollen*, or "plug flow", in which practically the entire velocity change occurs in a marginal zone a few tens of metres wide, and the major part of the glacier moves as a unit like a solid body sliding down a chute (Finsterwalder, 1937, pp. 96–8; 1950, p. 385).

Short-interval velocity measurements. In 1952 density waves in the air spoiled an attempt to record the movement of this glacier by means of a time-lapse motion picture camera taking exposures at 45-second intervals. This procedure appears to have merit, but an environment different from that of the Saskatchewan, where the air waves caused apparent displacements of the target exceeding the actual amount of movement, is required for its successful application.

Daily measurement of velocity stakes opposite the Castleguard camp in 1952 revealed marked variations in flow rates at different points, so a similar but extended and more carefully controlled study was undertaken the next year. In 1953, 5 stakes arranged in a cross, one arm transverse and the other parallel to the direction of flow (Fig. 5), were observed at 12-hour intervals from August 17 to September 3. Bad weather and low visibility occasionally interrupted the observations. Distances to the stakes were known from triangulation so that measurement of horizontal angles permitted calculation of the downvalley component of movement. The observing instrument, a Wild T-2 theodolite, was enclosed in a tent to guard against wind and incident

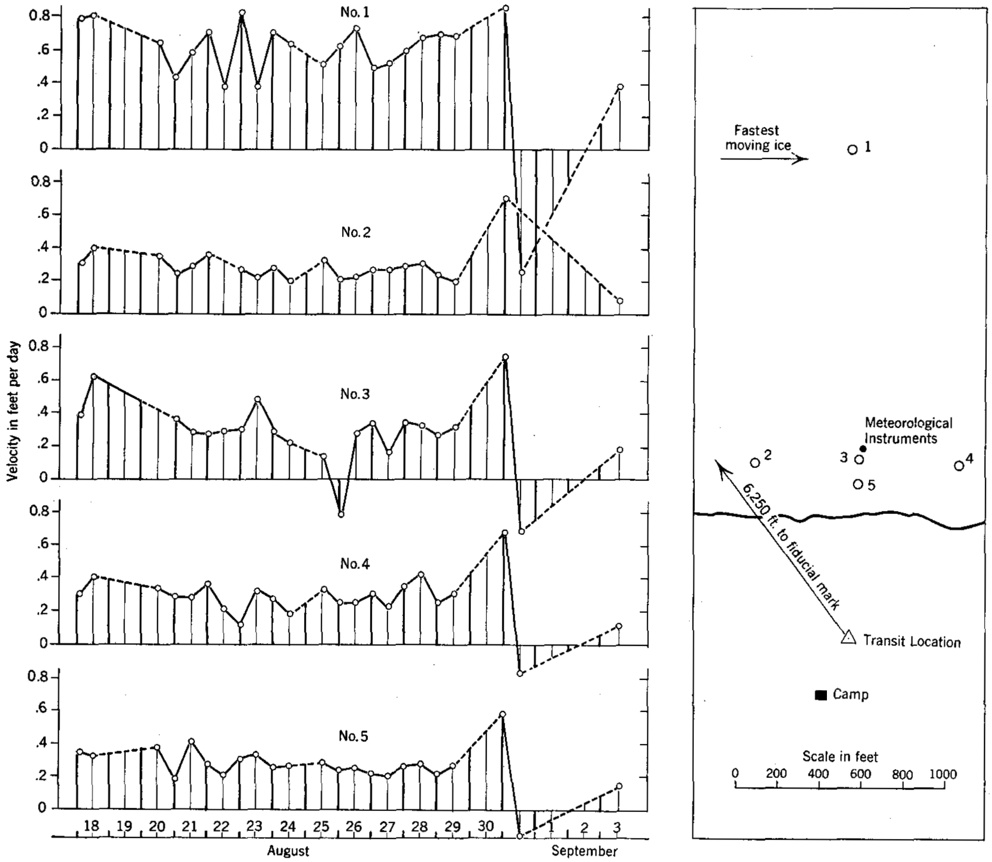


Fig. 5. Map and plot of short-period movement stakes. Readings plotted at mid-point of 12-hour observation intervals, specifically at 1 a.m. and 1 p.m.

radiation, and great care was taken to eliminate effects of diurnal drift in the instrument by turning all angles from a fixed mark on a bedrock cliff across the glacier. The observation point was on a flat area along the crest of an abandoned lateral moraine. Possibilities of creep or slump at this site cannot arbitrarily be dismissed, but the records show that it did not occur in perceptible amount during the study. All the movement stakes lie on the same side of the line from the station to the fiducial mark on the opposite side of the glacier, and a shift in the observing instrument would produce an apparent movement of the velocity stakes all in the same direction and of an amount inversely proportional to their distance from the instrument. No such behaviour was recorded.

The precision of measurement was tested by turning a fixed angle between two bedrock stations across the glacier, a little more than 6,000 feet from the instrument point, 50 times under various different conditions with different operators. The frequency distribution of these angles shows a standard deviation of 3.0 seconds of arc, which means that 95 per cent of the time the measured angle was within 6.0 seconds of the true angle. Six seconds of arc are equivalent to distances of 0.02 and 0.07 feet respectively at the nearest and most distant stakes. An optical plummet on the theodolite reduced set-up errors to less than 0.01 foot, so, including possible inaccuracies arising from wind, it seems safe to conclude that the accuracy of measurement of velocity stake locations is, with rare exceptions, within 0.05 foot.

For purposes of graphic presentation, the velocities recorded for each 12-hour interval are plotted at the mid-point of the observation period, namely 1 a.m. and 1 p.m., and these points are connected by a straight line (Fig. 5). This is somewhat unrealistic as the movement may have occurred any time within the 12-hour interval, but since velocity changes are the matter of principal concern this representation is permissible. Inspection of Fig. 5 reveals some erratic behaviours. Some periods of acceleration affect all stakes simultaneously, but it also appears that rapid movement of one stake is not necessarily matched by proportionately rapid movement of its neighbours in either the longitudinal or transverse position. For example, during the first half of August 18, stake No. 5 decelerated slightly while the others accelerated, especially No. 3. During the second half of August 21 all points accelerated except Nos. 3 and 5 which decelerated, the latter markedly. During the last half of August 25, No. 1 accelerated while all other stakes decelerated, but by the second half of August 26 when No. 1 started to decelerate the other stakes were accelerating, except No. 5 which decelerated slightly. Ignoring for the moment the exceptional behaviour of August 31, individual stakes show velocity changes as large as 0.4 foot in successive 12-hour intervals, and the velocity difference between stakes during a single 12-hour interval is as great as 0.7 foot. The largest fluctuations are displayed by stake No. 1 which is farthest out on the glacier where velocities are greatest. It is clear that studies of this type should be made with respect to a number of stakes and not to just one, as is the usual practice.

In the last half of August 25, stake No. 3 actually receded 0.2 foot up-valley. Isolated reverse movements of this type are probably due to backward slip or rotation of individual blocks within the crust of the glacier. Many of the erratic accelerations and decelerations displayed by the various stakes may be due to shiftings of individual crustal blocks superimposed on a reasonably steady downvalley flow of the whole glacier. This possibility could be tested by comparing the total movement of the various stakes over a longer period of time. It seems unlikely that the erratic behaviours are due to refraction of sight-lines by differences in air density because such effects should apply in a similar sense to all the stakes.

The most striking behaviour recorded was a gradual acceleration followed by a sudden and larger deceleration, involving actual backward movement, shown by all stakes, except possibly No. 2 for which the record is incomplete, over the period August 29 to September 1. This fluctuation occurred during a prolonged period of heavy precipitation, suggesting a meteorological cause. For reasons already given this exceptional behaviour cannot be explained by a shift of the observation point. Washburn and Goldthwait (1937, p. 1,662) recorded periods of deceleration on Crillon Glacier, Alaska, following rain storms but offered no explanation for this behaviour. It may be that additional weight placed on the glacier by the precipitation sets up a longitudinal wave that passes through the ice in a downvalley direction, or perhaps a sudden local movement within the glacier, lubricated by the water, generates a similar wave. Passage of a longitudinal wave through the ice could conceivably produce the type of behaviour recorded. Critical evaluation of these highly speculative ideas is not possible with the scant data in hand.

Regular diurnal fluctuations in glacier velocity have been attributed to meteorological influences, particularly temperature and radiation (Krasser, 1939, p. 302; Klebelsberg, 1941, p. 375; 1948, p. 86; Drygalski and Machatschek, 1942, p. 112), although no wholly satisfactory explanation of the mechanics involved has been offered. During the Saskatchewan study temperature and pressure data were collected by continuously-recording instruments on the glacier's surface, and these data were supplemented by general observations of cloud cover, wind, and precipitation. A statistical analysis of the meteorological data has not been made, so the possiblity of correlation between daily velocity behaviour and the recorded meteorological variations cannot be evaluated. A watch was kept for waves or impulses of increased velocity within the glacier that might be traced successively from station to station along the longitudinal profile, but none was recognized.

Englacial velocities

Knowledge of the magnitude and distribution of velocity within glaciers is essential to analysis of stress–strain relations under the pressures and temperatures attained at depth, to an understanding of the mechanics and mode of glacier flow, and to an evaluation of the role of basal slip in the movement of ice bodies with sloping floors. Saskatchewan Glacier is an excellent subject for investigation of velocity–depth relations because of its relatively simple geometry and the data available on its various dimensions and behaviours.

For this purpose we planned to sink a pipe vertically through the glacier to its floor and to determine subsequent deformation of the pipe by means of inclinometer surveys, a procedure already used in other glaciers (Perutz, 1949; Gerrard, Perutz and Roch, 1952; Sharp, 1953). Aluminum pipe, inner diameter 1.38 and outer diameter 1.65 inches, was selected for the drill stem because this permitted use of a small-diameter inclinometer in which bearings of inclination readings were given by magnetic compass. Boring in the ice was by electrical hotpoints of 1.75 inches outer diameter, power being furnished by a portable 2,500-watt, 220-volt, AC generator driven by a small gasoline engine (Fig. 6). The hotpoints operated at a current of 8.0 to 9.5 amperes, and maximum boring speeds of 14 to 16 feet per hour were attained under normal conditions.

In 1952 borings were attempted at two sites near the centre of the glacier opposite the Castleguard camp (Fig. 4). Both attempts were unsuccessful owing to burned out or shorted hotpoints and to seizures of the drill stem by shearing in the crustal ice. The greatest depth attained was 150 feet. A

Fig. 6. Hotpoint boring operation in 1952, August 3.

resurvey was made in 1953 of the pipe left in this hole, but deformation was too small to be significant.

In 1953 a better site was found higher up the glacier near the abandoned Army camp (Fig. 4). This proved suitable except for minor seizures of the drill stem, possibly caused by freezing within a vestigial chilled zone near the surface. The operation got off to a reasonably good start but ended in failure when the deepest hole, 395 feet, was lost in the process of replacing a shorted hotpoint. Failure of the 5 available hotpoints was the principal stumbling block in 1953. Further attempts will be made to bore through the glacier at this site.

Structures in the glacier

Most structures in glacier ice have been created by or because of flowage so that knowledge of their characteristics and origins can contribute to an understanding of glacier flow.

Sedimentary layering. The problem of distinguishing sedimentary layering from metamorphic foliation in glacier ice has long plagued glaciologists. Initially, this proved a troublesome matter on the Saskatchewan, but continued observation and experience brought conviction that both a relic sedimentary layering derived from the firn and a secondary foliation created by flowage could be recognized. Sedimentary layering is most easily identified close to

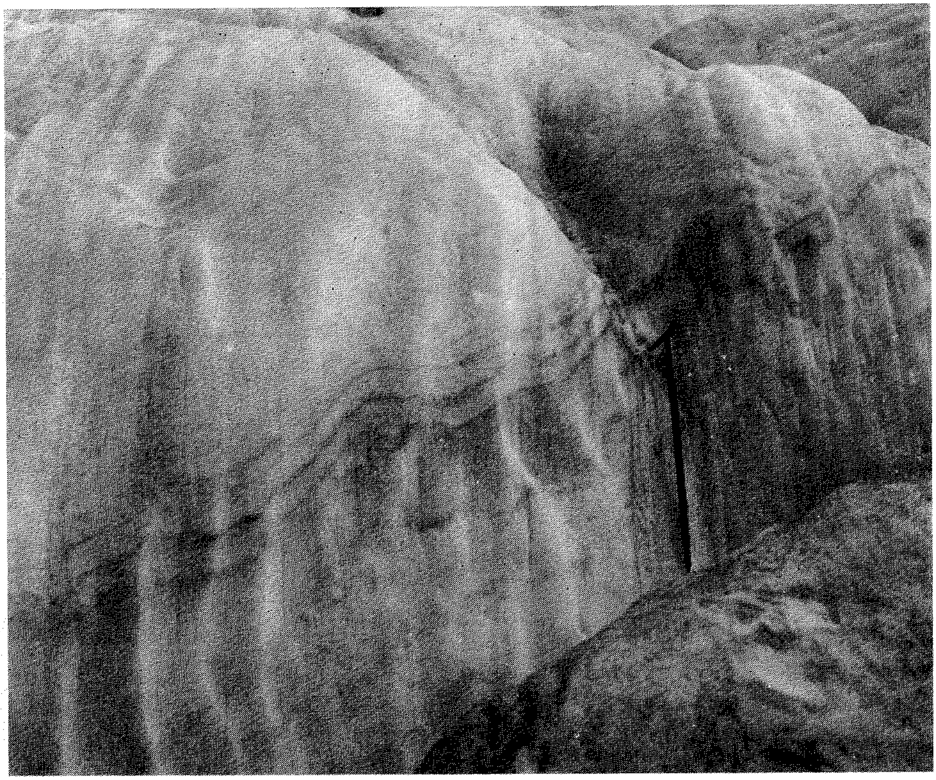


Fig. 7. Gently dipping sedimentary layering cut by essentially vertical flow foliation exposed in wall of crevasse nearly 3 miles below firn limit.

the firn limit where it consists of gently dipping layers of loose, granular, partly reconstituted firn alternating with ice layers and lenses of the type formed by freezing of meltwater percolating into firn (Sharp, 1951, pp. 609–14). These granular layers are much too thick and irregular to be the product of shearing. Farther down the glacier a foliation, which appears to be of secondary origin, becomes progressively stronger, and all doubt of its secondary nature is removed by exposures in which the foliation cuts across sedimentary layers at a high angle (Fig. 7). Isoclinally folded and contorted granular layers seen locally along the lateral margins of the glacier (Fig. 8) are probably sheared-out sedimentary beds, for the folded structures are transected by flow foliation.

In many places, outcrop traces of the granular layers are discoloured by an accumulation of fine silty dirt, although in fresh excavations the granular layers are no dirtier than the adjacent masses of solid ice. It seems likely, though not as yet certain, that the dirt on the granular layers is a surface deposit of wind-borne or water-washed material, secondarily acquired. Debris collects on the granular layers because of the irregular intergranular depressions, while it is washed off the smooth surface of adjacent areas of solid ice, as early recognized by Forbes (1859, p. 21).

Fig. 8. Contorted sedimentary(?) layering near south margin of Saskatchewan Glacier, 28 July 1952.

Outcrop pattern of sedimentary layering. The attitude and structure displayed by sedimentary layers within the Saskatchewan Glacier raise a most puzzling and enigmatic problem. Viewed from a high vantage point, the surface of this glacier displays a faint pattern resembling that of truncated beds in a plunging fold (Fig. 9). Similar patterns are reported on the surfaces of other glaciers (Streiff-Becker, 1952, p. 5). Alternate bands of dirty and relatively clean ice describe broad sharp-pointed curves transversely crossing the Saskatchewan Glacier from the medial moraine on the north nearly to the south margin. These bands are most prominent within the first two miles below the firn limit. They die out toward the margins and become progressively fainter downvalley although recognizable to within one mile of the terminus (Fig. 10). Disappearance of the bands is probably due to recrystallization of the ice with elimination of the granular layers on which dirt tends to accumulate. Spacing and width of the bands are not uniform, and the curves become sharper and of greater axial length downvalley. After considerable ablation in summer, the dirtier bands can be traced across the glacier's surface by an experienced observer on foot, and they are found to be conformable with the sedimentary layering earlier described.

Fig. 9. Looking up Saskatchewan Glacier from south wall, 24 August 1952. Note outcrop pattern of sedimentary layering on surface of glacier. Short dark streak in centre mid-distance is abandoned Army campsite.

As initially laid down in the accumulation area the sedimentary layers were probably inclined gently downglacier, but subsequent flowage according to the concepts of Reid (1896, p. 919) should eventually give them an upvalley dip. The assumed age of the layers and their outcrop patterns would thus lead one to expect an open synclinal structure with the axis nearly horizontal or plunging gently upglacier, but such is not the case. The structure of the sedimentary layers, at least as determined from direct dip and strike readings, is actually an anticline plunging 8 to 14 degrees downglacier (Fig. 10), which is steeper than the slope of the glacier's surface. This means that ice at the surface should be younger toward the terminus and older toward the firn limit. The seeming absurdity of this raises serious doubt as to the validity of the field observations, but repetitions of the observations have not yet produced anything different. This is an enigma for which a suitable explanation has not yet been found.

Flow foliation. Much of the ice exposed on the surface of Saskatchewan Glacier and in the walls of its crevasses displays a well-developed flow foliation consisting of alternating laminae, a fraction of an inch thick, of white bubbly ice and denser bluish ice. This structure is strongest along the margins of the glacier and is prominent near the medial moraine.

In the centre near the terminus, foliation planes are spoon-shaped and dip gently upglacier. One mile above the terminus they are much steeper and trend parallel to the margins even in the centre where dips are essentially vertical. Along the margins dips are mostly 50 to 75 degrees inward. The nature of the transition from transverse spoon-shaped folia to steep longitudinal folia is obscured by an intervening zone of contorted ice. This difference in the shape and orientation of foliation planes has long been recognized (Forbes, 1859, p. 19), but a satisfactory explanation has never been offered. The following comments are wholly speculative.

Flow foliation is most strongly developed where velocity gradient is highest (shear stress is greatest) and where the physical state of the ice is such that it yields primarily along closely spaced shear planes. Throughout the length of the glacier the extreme basal foliation, if such exists, is probably essentially conformable to the valley floor except for divergences related to surfaceward movement of ice. Where the glacier exceeds a certain thickness, foliation planes of similar shape and orientation are not formed clear to the surface of the glacier because the ice at depth yields readily and prevents the drag of the floor from having any noticeable effect on ice at the surface. The surface ice, yielding less readily, is presumably influenced by marginal drag to the extent that a weak foliation is formed essentially parallel to the lateral margins, as actually observed.

The exact mechanics involved in the formation of foliation remains something of a problem upon which the Saskatchewan data throw little light. Thin sections cut across zones of granulated ice, apparently formed by shearing, show some evidence of recrystallization, and this has been proposed as a way in which foliation is formed (Philipp, 1920, pp. 502–3). It has also been suggested that reorientation of crystals involving recrystallization, migration of crystal boundaries, and similar processes acting along specific planar layers during solid flowage create the dense folia (Perutz and Seligman, 1939, p. 354). However, crystal fabric diagrams from the Saskatchewan and some other glaciers, although seemingly related in orientation to the foliation, do not show, so far as our understanding extends, that reorientation and recrystallization have occurred more extensively in the dense folia than in the adjacent ice. Individual crystals actually extend through two or more folia, and it may be that recrystallization following development of the foliation has destroyed differences in crystal orientations that formerly existed between adjacent folia. It can only be said that foliation in the Saskatchewan Glacier appears to have been formed by flowage and to be conformable with the planes of greatest shear strain in the ice. Some exposures show two sets of foliation planes of different orientation, age, and stage of development suggesting that new folia form as the stress orientation changes during flow.

Faults. Faults of small displacement offset the foliation in the terminal and marginal parts of the glacier. Near the terminus are many transverse low-angle thrusts, most of which dip upglacier, suggesting that compressive flow prevails (Nye, 1952, p. 89). A few dipping downglacier are probably

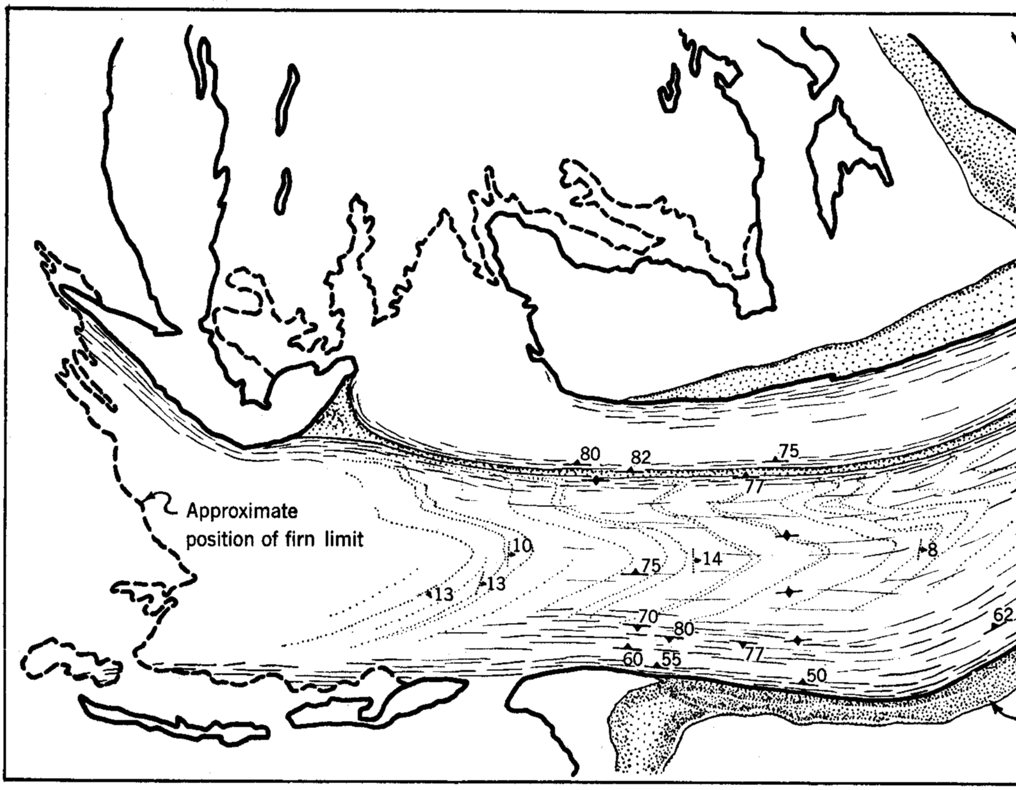
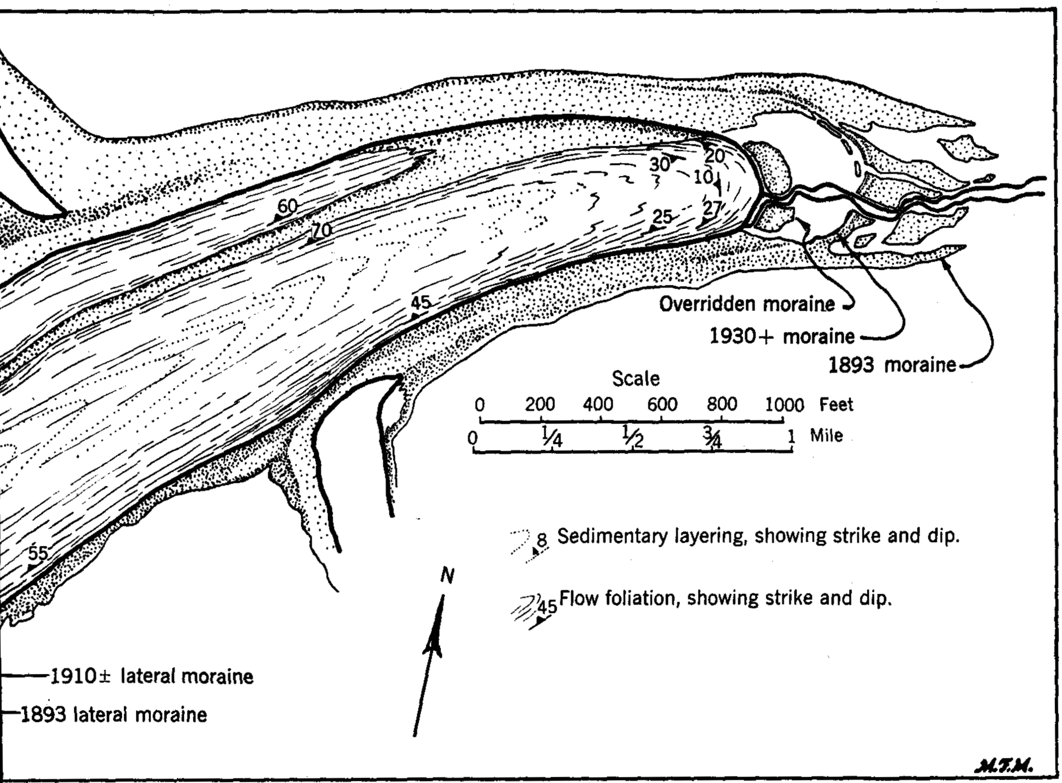


Fig. 10. Structural map of sediment

conjugate to the predominant set. The terminal zone also displays a number of transverse high-angle reverse and normal faults. Although some of the latter resemble structures produced by extending flow (Nye, 1952, p. 89), it seems more likely that they are the product of buckling and settling of crustal blocks.

Marginal faults, studied chiefly along the south edge of the glacier, are mostly strike-slip in nature with apparent left-hand displacements of a foot or less. Depression of the downglacier block in some instances indicates movements that are oblique rather than strictly parallel to the strike. The location, distribution, nature, and orientation of these marginal faults suggest that they are the result of adjustments within the glacier related to slight bends or other irregularities in the valley walls.

Crevasses. Saskatchewan Glacier displays crevasses of 5 different types, all but the last of which can probably be explained by a stress analysis similar to that used by Nye (1952, pp. 89–91). The most common crevasse has a curved trace concave upglacier and passes from a longitudinal orientation near the centre to a 45-degree transverse orientation near the margin (A, Fig. 11). Such crevasses mostly die out within a few hundred feet of the margin, and



layering and foliation at surface of glacier.

only a few extend to the glacier's edge which they intersect at 45 to 75 degrees. Crevasses of this type are abundant in the Castleguard sector where the glacier attains its greatest width. The central longitudinal trend may be related to lateral spreading of the ice in this wider part, and the gradual change to an oblique trend is probably due to the increasing influence of tensile stresses generated by a high velocity gradient near the margin. Relations between these stresses and transverse velocity gradients are being studied quantitatively.

Transverse crevasses convex upglacier constitute the second type recognized and are seen in the steep lower reach of the north-bank tributary (B, Fig. 11). The convex longitudinal profile of the glacier in this reach, related to a steepening of the rock floor, appears to be the cause of these crevasses. Their upglacier convexity is probably related to the marginal velocity gradient which gives rise to tensional stresses oblique to the glacier's margin.

The same tributary shows an interesting change in character, longitudinally, which is largely responsible for longitudinal crevasses in its lower part, the third type of crevasse distinguished (C, Fig. 11). The upper reach of the ice stream is confined in a steep-walled valley, but the lower half flows in an unconfined course down the steep wall of the trunk valley; in effect it becomes a "wall-sided glacier" (Ahlmann, 1940, p. 192). A cross section of the ice stream in the upper reach would show a relatively flat upper surface and a curved bottom. A cross section in the lower reach would be just the opposite, a flat bottom and an upper surface convex to the sky. This transverse convexity plus the spreading of the ice in the unconfined lower reach accounts for the longitudinal crevasses.

The fourth type of crevasse is simply a straight marginal fracture trending obliquely upglacier (D, Fig. 11). Marginal crevasses have probably been integrated into the transverse crevasse systems already described, but they also exist as independent entities in Saskatchewan Glacier. They are the product of oblique downglacier tensile stresses set up within the marginal zone of steep velocity gradient, without perceptible complication by other stress systems.

The fifth and most unusual type of crevasse consists of narrow belts of short *en échelon* cracks, best seen along the south margin near the bend below the Castleguard sector (E, Fig. 11). These subequally spaced belts diverge only a few degrees from the direction of flow, although the individual cracks are about 45 degrees to the flow direction and essentially parallel to neighbouring transverse crevasses (Fig. 12). No satisfactory explanation has yet been evolved for these crevasse belts, but they may be related to shear couples set up by the obstructing effect of the valley wall at the curve. The only other parts of the glacier displaying such crevasses are the north and south margins just below the entry of the north-bank tributary. Here the glacier is crowded forcefully against the valley walls to make room for the additional ice, and this may give rise to the exceptional marginal shear stresses required to produce belts of *en échelon* crevasses.

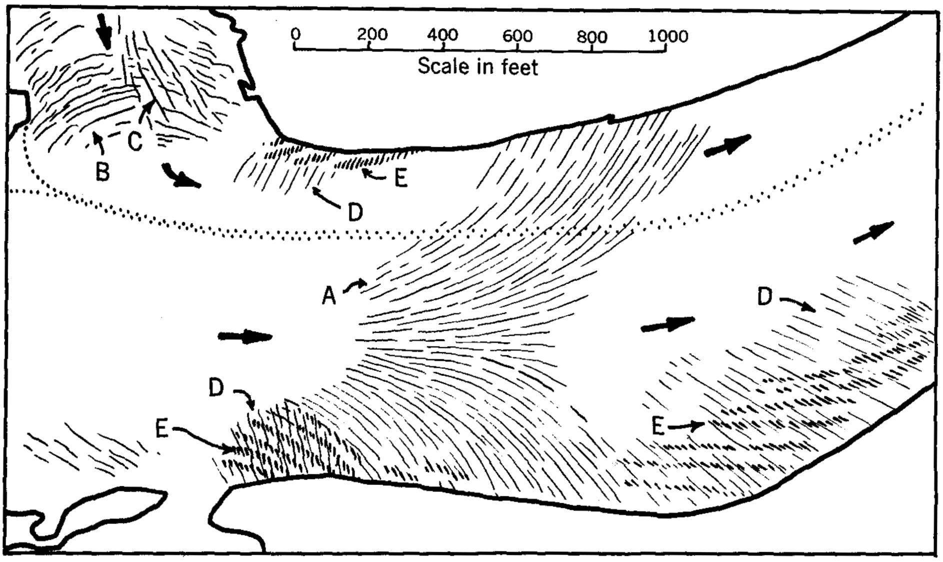


Fig. 11. Map of crevasse types in Castleguard sector. A—transverse crevasses concave upglacier; B—transverse crevasses convex upglacier; C—longitudinal crevasses; D—marginal crevasses; E—belts of *en échelon* crevasses.

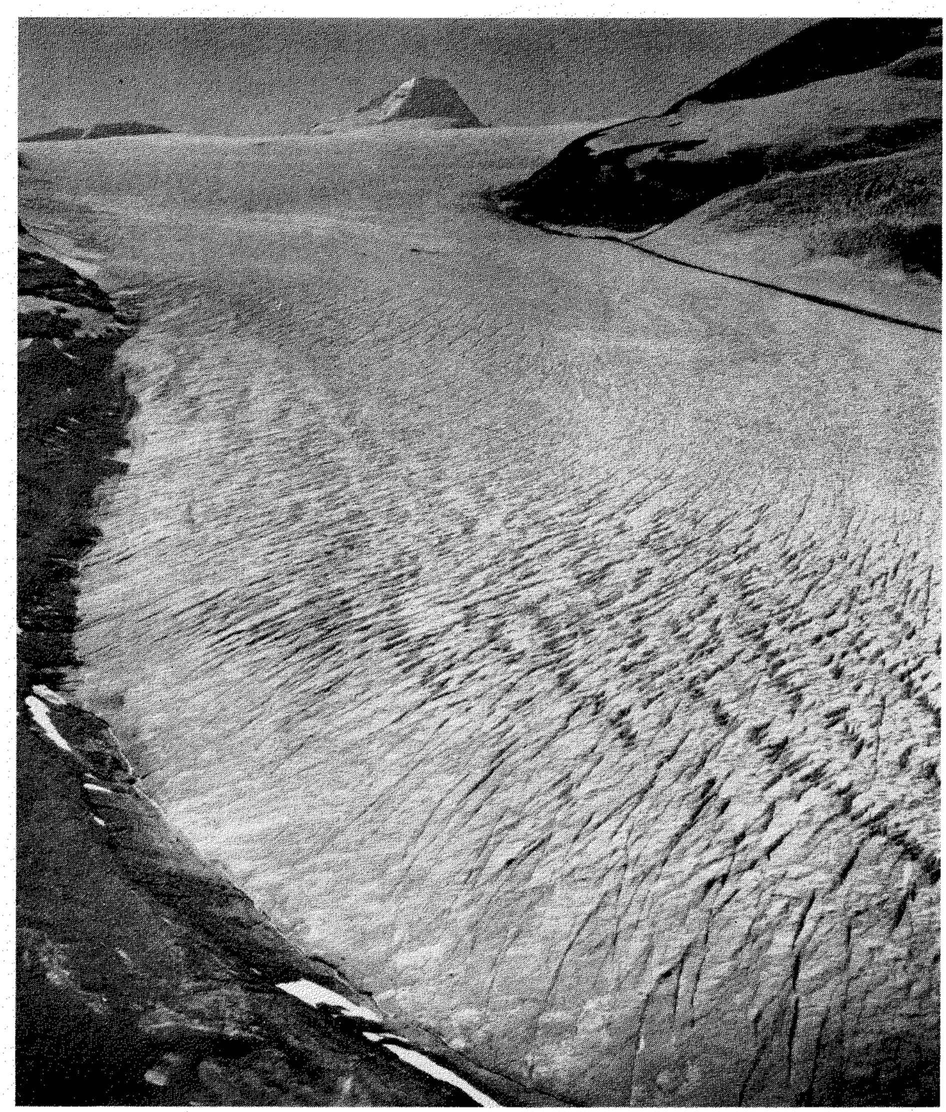


Fig. 12. South margin of glacier in Castleguard sector showing belts of *en échelon* cracks and large transverse crevasses concave upglacier. Transverse and longitudinal crevasses are also visible in tributary at right. Columbia Icefield and Mt. Columbia on skyline, 13 August 1953.

Crystal fabrics

Method and procedure. The optical orientation of ice crystals was determined in the field by means of a large universal stage mounted between crossed polaroid sheets. Fabric diagrams were constructed in the conventional manner by plotting optic axes on the lower hemisphere of a Schmidt equal-area net. All sections were cut in a vertical plane parallel to the

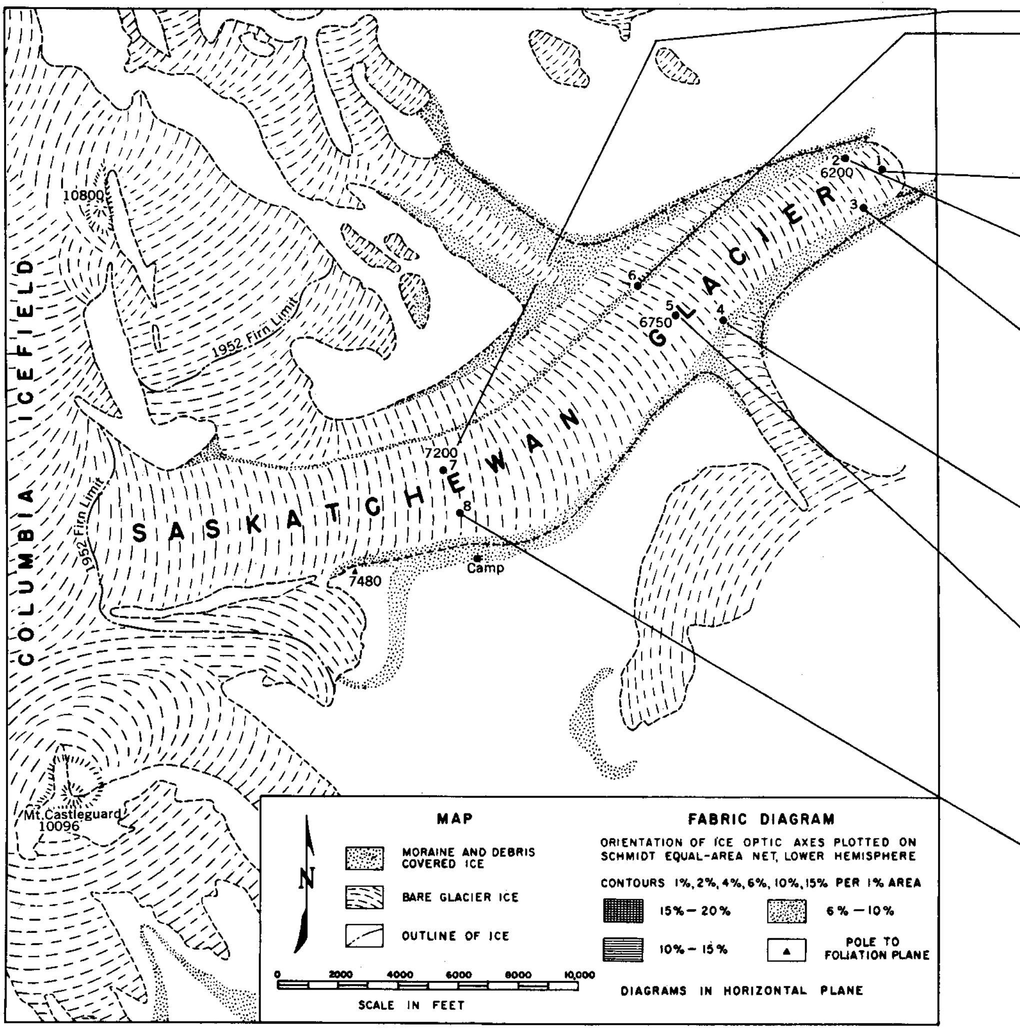
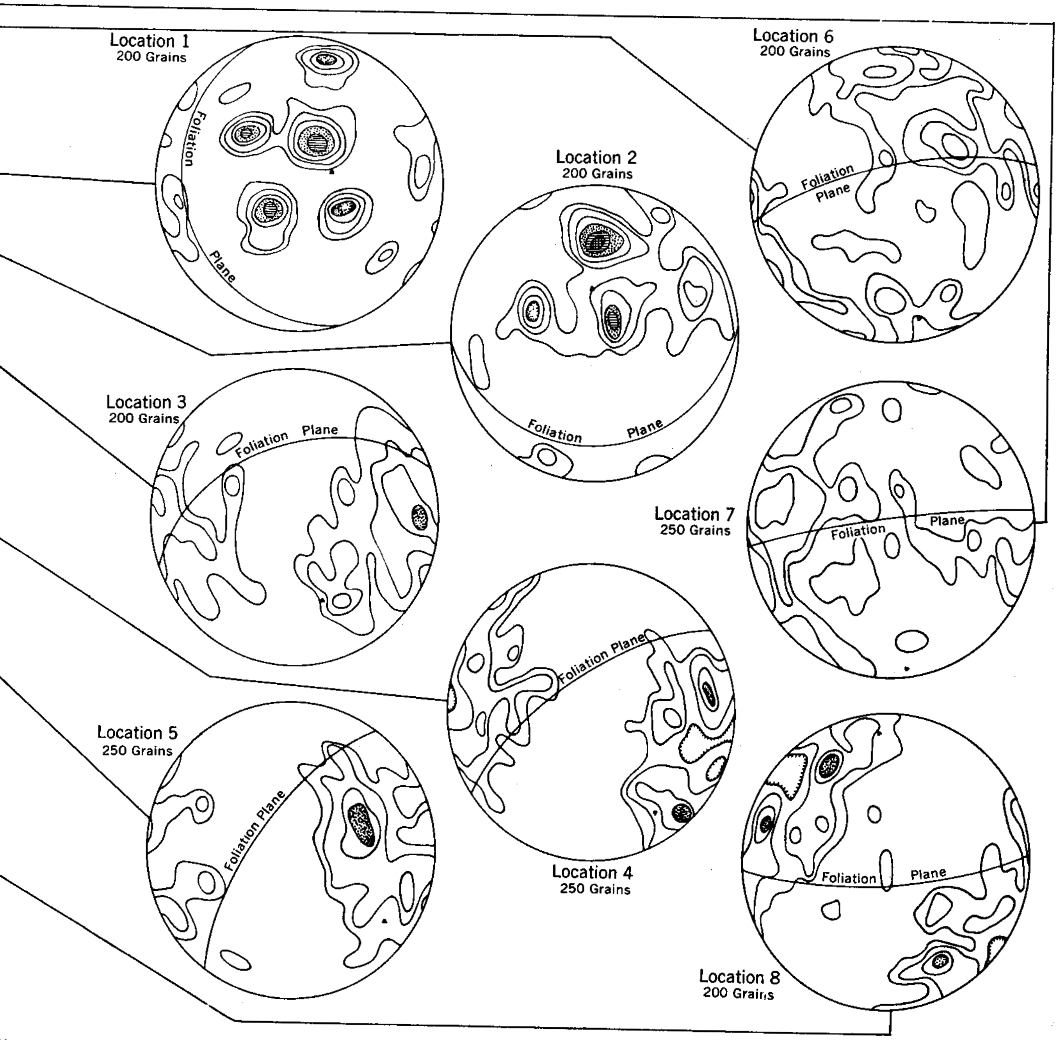


Fig. 13. Map and crystal fa

foliation, but the diagrams were subsequently rotated to a horizontal position to facilitate comparison with other data.

Either 200 or 250 grains were measured at each of 8 locations on Saskatchewan Glacier as shown in Fig. 13. The Saskatchewan diagrams have weaker and more varied patterns than found on either the Emmons Glacier, Mt. Rainier (Rigsby, 1951, p. 594), or the Malaspina Glacier, Alaska (Bader, 1951, pp. 531–2; Rigsby, 1953, p. 17).

Diagrams from locations 1 and 2 near the Saskatchewan terminus have several strong maxima clustered about the pole to the foliation plane and closely resemble the predominant pattern found on the other glaciers. Curiously, the diagram at location 3, also in the terminal area, is weaker and distinctly different, resembling closely those at locations 4 and 5, a mile farther



iagrams for Saskatchewan Glacier.

up the glacier. The diagram at location 6, geographically associated with locations 4 and 5, is still different and is more like that at location 7, 2.5 miles farther upglacier. The diagram at location 8 is more like those at 4 and 5 than that at nearby location 7. It appears that the stronger patterns exist in the more extensively metamorphosed and presumably older ice. Most fabric patterns from Saskatchewan Glacier are weaker than those from the Emmons and Malaspina glaciers because the Saskatchewan ice has been less extensively reconstituted as shown by the preservation of sedimentary structures far below the firn limit.

The patterns at locations 6 and 7 appear too weak to have much significance. The patterns at 3, 4, 5, and 8 all show an offset of the maxima to the same side of the pole to the foliation plane. At locations 3 and 4 this may have something to do with proximity to the glacier's margin, but the actual cause and significance of the relation are not known. A satisfactory explanation for strong patterns like those at locations 1 and 2 in which a number of maxima cluster about the pole to the foliation plane is still being sought. The tentative suggestion that ice crystals may yield along planes other than the basal glide plane (Rigsby, 1951, p. 597) is not supported by any independent data known. A recent suggestion by Schwarzacher and Untersteiner (1953, pp. 122–4), relating such patterns to slip on the basal glide plane of crystals properly oriented along several sets of shear planes within the ice lying at an angle to the foliation, is a possibility that merits further investigation.

The direct relation between strength of crystal orientation and degree of metamorphism and the clustering of maxima around the pole to the foliation plane certainly suggest that the crystal fabrics are related to flowage. However, it may be that the fabrics observed in ice on the glacier's surface are a secondary product formed by a recrystallization that proceeds according to certain laws or controls from a previously well-oriented arrangement established by flowage. Evidence for recrystallization following development of foliation has already been cited. Clearly, the interpretation of fabric diagrams from glacier ice needs the aid of laboratory experimentation, and such work is now being undertaken by Rigsby.

Miscellaneous investigations

Ablation. A determination of total regime was not attempted on this glacier, but gross measurements of surface ablation were made at movement stakes on the ice tongue below the firn limit. These data will be useful as a means of assessing the relative effects of ablation and surfaceward flow upon the glacier's profile. Ablation during summer on the Saskatchewan differs markedly in time and space. It ranged from 0.8 to 4.7 inches of water per day in 1952.

Ablation is greatest near the terminus, an estimated 25 to 30 feet of ice per season, but the progression of average daily ablation from lower to higher points is not an orderly decrease. Ablation during part of the summer of 1952 along the centre-line of the glacier in the Castleguard sector averaged 1.65, 1.99, 1.48, 1.65, and 1.91 inches of water per day from the lowest to the highest station. These figures reflect subtle variations of the local environments as well as differences in the nature of the materials being melted. Still greater differences are found in areas lightly covered by debris, and in proximity to the valley walls where excessive melting produces a marginal depression.

Glacier fluctuations within the last century. The following features are found within the first mile beyond the present terminus of Saskatchewan Glacier: (1) an old, broad, overridden moraine barely 1,000 feet beyond the terminus, (2) a trim-line in mature timber on the north valley wall clearly related to (3) a bouldery moraine 4,000 feet down the valley, (4) a more recent small double moraine 1,800 feet from the terminus, and (5) four distinct terrace levels in the proglacial outwash, each of which is related to a recognized phase of the glacial history. Except for the overridden moraine, the end moraines can be traced upvalley into lateral moraines displaying the same relative position and characteristics.

By means of tree-ring relations along the trim-line, C. J. Heusser (personal communication) demonstrates that part of the trim-line and the outermost moraine are the products of an advance culminating in 1807. The age of the trees outside the trim-line shows that this was the greatest advance in at least a century. A higher part of the trim-line farther up the valley was formed by a later advance culminating in 1893 as previously reported (Field, 1949, p. 109). The overridden moraine described herein may have been formed during recession from the 1807 position and subsequently overridden by the 1893 advance, or it may be still older. Observations of the glacier since 1893 have been frequent enough to show that this moraine was not formed subsequent to 1893. Old photographs indicate that the small double moraine 1,800 feet from the present terminus was formed between 1930 and 1935.

Oxygen isotope studies. Recent investigations of the ratio between oxygen 18 and oxygen 16 in waters of various origins have revealed interesting and significant differences (Epstein and Mayeda, 1953). Snow and the melt-water from glaciers are exceptionally low in 0^{18}. With the thought that differences in the $0^{18}/0^{16}$ ratios within glaciers might be related to the original environment of precipitation or to the subsequent history of the ice, 38 samples of ice were collected on the surface of Saskatchewan Glacier along a traverse from the firn limit to the terminus. Analysis of these samples by mass spectrometry is currently underway, and only preliminary results are available. Significant differences in the $0^{18}/0^{16}$ ratios exist within this glacier, but no consistent trends or relations have yet been worked out.

Five samples of ice were also collected along a profile from the firn limit to the terminus at the request of W. M. Campbell of Atomic Energy of Canada. This organization analysed the specimens for deuterium by mass spectrometer with results suggesting no enrichment of deuterium in the ice over that present in other natural waters.

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