# GLACIOTHERMAL STUDIES ON THE TAKU GLACIER SOUTHEASTERN ALASKA

by

Maynard M. Miller

Juneau Icefield Research Program
and
Lamont Geological Observatory
Columbia University

Publication no 39 de l'Association Internationale d'Hydrologie, IUGG
(Assemblée generale de Rome, tome IV). *Extrait*, pp. 309-327
1954

Publication n° 39 de l'Association Internationale d'Hydrologie
(Assemblée générale de Rome, tome IV). *Extrait*

# GLACIOTHERMAL STUDIES ON THE TAKU GLACIER SOUTHEASTERN ALASKA

by

**Maynard M. MILLER**

## Introduction

The research here described was part of the 1950 to 1953 glaciological program on the Juneau Icefield. It extends certain studies carried forward since 1948 with the support of the American Geographical Society, the Office of Naval Research and other governmental and private agencies (Miller and Field, 1951). The objective of these particular investigations was to obtain information pertinent to an understanding of the regimen of the firn-pack on the Taku Glacier system and an explanation of the anomalous and spectacular advance of this glacier during the past 50 years in a region where general glacier shrinkage is the rule. Systematic measurements were made of englacial temperatures in each season of the year and in different sectors of the icefield. The results now permit comparison of the thermal regime at and near the névé line with that in the critical highland source area lying at the crest of the Alaskan Coast Range.

## Physical Setting

The Taku Glacier (Fig. 1) is the largest ice stream leading from the 1000 square miles of highland ice, known as the Juneau Icefield, in the mountains north and east of Juneau, Alaska. The glacier is 36 miles in length from its tide-water terminus in Taku Inlet to the highest source névé at the summit of the range. It is 2 to 4 miles wide with an average gradient of 3 percent. Its direction of moverment is southerly and a regular streaming type of flow is exhibited at a surface velocity of 1 to 4 feet per day. The Llewellyn Glacier, another large valley glacier, takes its source in the same névé but drains towards the northeast and terminates near the shore of Lake Atlin in northern British Columbia. Together, these ice tongues form a transection glacier comprising the central portion of the main glacial area.

Of the two primary sites chosen for these measurements, the first (Camp 10 B, Figs. 1 and 2) was near the center of the main upland branch of the Taku Glacier at an elevation of 3575 feet, only a few hundred feet higher than the semi-permanent névé line and 16 miles from the glacier's tide-water terminus. The second (Camp 8 B) was at 5900 feet elevation at the center of the glacier and near the highest point of its source area. This location is approximately at the divide between the Taku and Llewellyn névés. At both sites temporary meteorological stations were established to provide information concerning atmospheric changes affecting the englacial temperature regime.

## Equipment Employed

Through cooperation of the U. S. Geological Survey, multiple thermistor cables, in lengths of 35, 50 and 200 feet, were obtained for insertion into bore-holes. The instrumentation is similar to that which has been sucessfully used in geothermal studies of permanently frozen ground at Point Barrow, Alaska. (MacCarthy, 1952; Swartz, 1954).

Each cable contains a number of thermistors vulcanized into the cable at suitably spaced points. The individual thermistor (Western Electric Company Type 17 A) comprises a sintered disc of manganese and nickel oxides, 0.2 inch (5 mm) in diameter, and 0.04 inch (1 mm) thick. Two axial copper leads, 0.02 inch (0.5 mm) in diameter are attached by silver eutectic solder to ceramic silver paste buttons applied to the faces of the disc. The resultant semiconductor has a negative temperature coefficient of resistance with a change of 4.4% per degree Centigrade change in temperature at room temperatures and a 6% change per degree C. in the vicinity of —30°C.

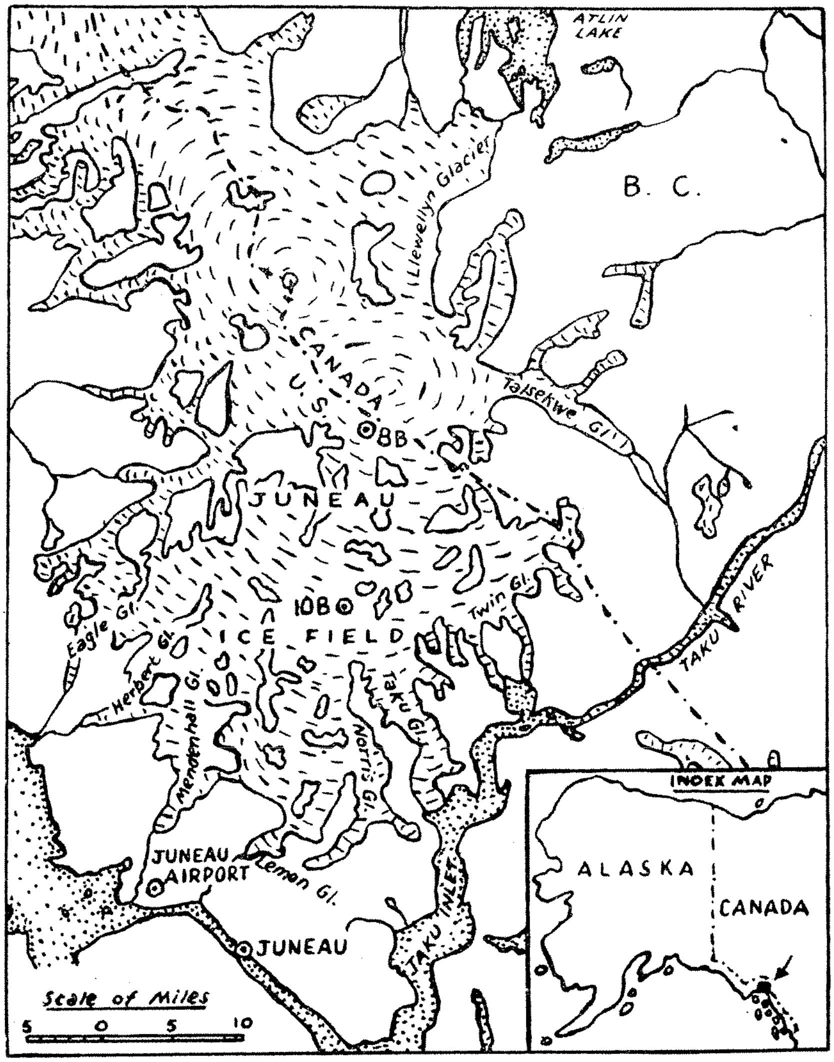



Fig. 1

Measurements are made to an accuracy of 0.01° to 0.03°C. by means of a modified Wheatstone Bridge.

Resistance thermometers of this type are much more sensitive than thermohms and do not require a field reference junction as do thermocouples. Measurements by the U. S. Geological Survey in northern Alaska have indicated that similar thermistors installed at constant temperature points in permanently frozen ground have shown insignificant drift (M. C. Brewer, personal communication). It is believed therefore that, within the limits involved in the present work, the thermistors have suffered little change of calibration with time.

The glaciothermal cables were held taut in the holes by means of lead weights. The 35-foot and 200-foot cables were 20-conductor types with a diameter of 0.70 inch (17.8 mm). The diameter increased to 0.94 inch (23.9 mm) at the vulcanized joints where individual thermistors were incoroporated. The 50-foot lengths were 9-con-

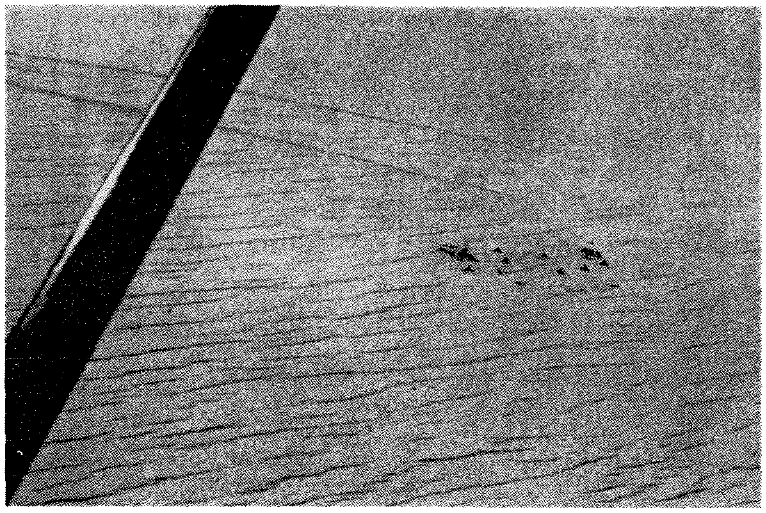


Fig. 2

ductor cables of 0.35 inch (8.9 mm) diameter and 0.63 inch (16 mm) at the vulcanized joints. A 30-contact selector switch permitted direct measurement of any of the 9 to 19 thermistors on each string. The resistance of a unit at the desired depth was then converted to a temperature by means of a calibration chart. For supplementary measurements in test pits and surface layers alcohol-in-glass thermometers with an accuracy of 0.3°C. were employed.

To provide holes in which the englacial cables could be inserted, three kinds of boring equipment were used: (1) a hand auger producing holes of 1.25 inches (31.8 mm) diameter for emplacement of the shallow cables; (2) a motor-driven rotary borer for deep holes of 3-inch diameter (76.2 mm); and (3) electro-thermic boring equipment for holes of 2.5 inches (63.5 mm) diameter to any desired depth. The bore-holes were more than 100 feet from the tents and living quarters so that during periods of measurement there would be no contamination of the record from abnormal sources of heat.

### Field Procedure

Emplacement of the cables at site 10 B is illustrated in Fig. 3. The position of each thermistor to a depth of 170 feet below the fiducial surface is shown.

The 200-foot cable (No. 152 in the figure) had 18 thermistors, spaced 10 feet apart for the first 160 feet and 20 feet apart for the lower 40 feet. It was installed in a 3-inch (76.2 mm) diameter bore (Hole No. 3) and firmly packed with snow held in position by light sheets of rolled paper tied with string. The lower 50 feet were implanted in glacier ice; 107 feet rested in firn; and 13 feet of the cable were placed for measurements in the subsequent winter's snow-pack. The essential purpose was to determine the depth of penetration of the annual chill. 245 feet of 2-inch I. D. (51 mm) aluminum pipe was located in drill hole No. 1 at a horizontal distance of 6.72 feet from hole No. 3. Because of the relatively low thermal diffusivity of firn and ice, as discussed in a later section, it seems improbable that the aluminum pipe materially influenced temperature readings in hole No. 3.

A 50-foot string (Cable 148 in the figure) with 19 thermistors was located in hole No. 2. The lower three thermistors, spaced at 5-foot intervals, were placed in the 1948-49 firn-pack. Above this the thermistors were in the subsequent snow-pack, with 6 thermistors at 2 1/2-foot intervals and the upper 10 of them at one-foot intervals. A third cable (No. 150) in a 50-foot length was hung vertically in a narrow

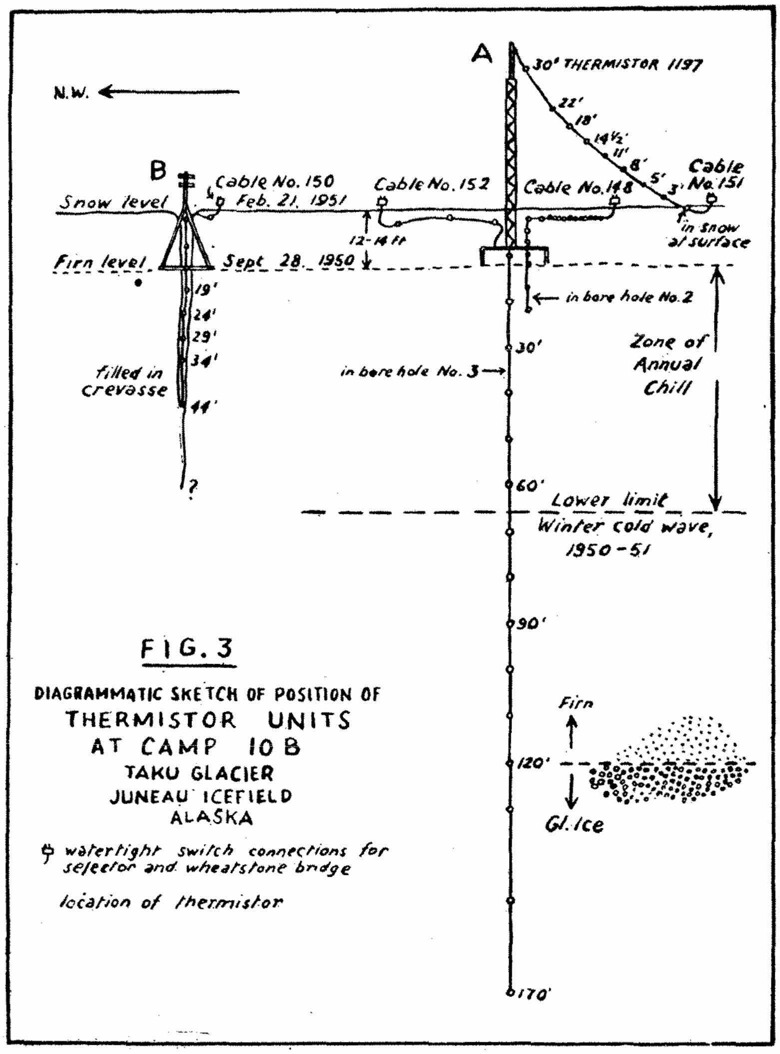



FIG. 3

DIAGRAMMATIC SKETCH OF POSITION OF THERMISTOR UNITS AT CAMP 10 B TAKU GLACIER JUNEAU ICEFIELD ALASKA

crevasse and marked with a wooden tower for future reference. At Camp 8 B another string of thermistors was placed in a bore-hole. The unit spacing on these was the same as noted for No. 148, thus permitting measurement of the temperature gradient in the highest névé at increments of 1 to 5 feet (position below surface shown in Fig. 9).

The first systematic measurements at 10 B were undertaken in February, 1951. Five months had elapsed since installation of the cables so these records are considered to represent natural and undisturbed conditions. Data were obtained at all levels down to 170 feet. Measurements on the upper thermistors were taken three times daily from February 9th to 26th and the deeper readings intermittently at one to three-day intervals. Readings were not obtained from March through May but were resumed on June 4th and continued daily until July 3rd.

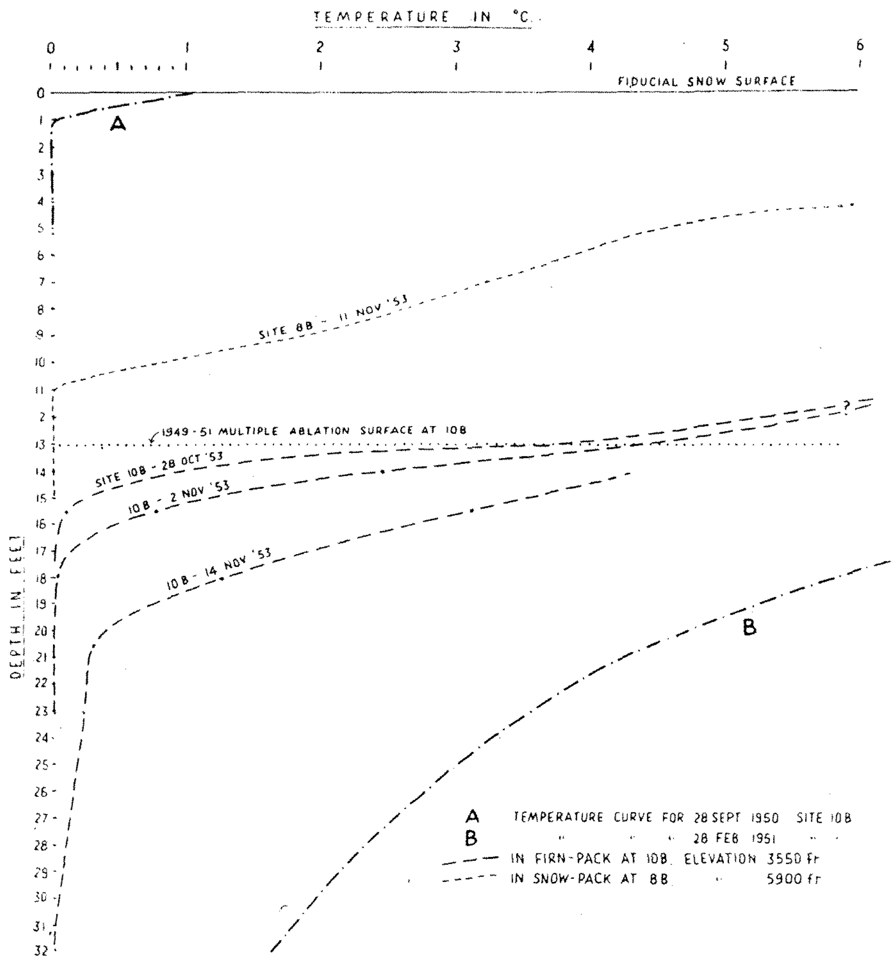



Fig. 4
Temperature profiles at 8 B and 10 B, autumn 1953, showing relationship to limiting curves of 1950-51 thermal year.

## Initiation of the Annual Temperature Wave

In each year between 1948 and 1953 the annual chilling of firn at the elevation of 10 B has commenced variably between the 16th and the 24th of September. With this the summer ablation season has effectively been brought to a close. On the crestal névé a different situation occurs with the cold wave being initiated some weeks earlier. In 1951 dominantly sub-freezing conditions were found at Camp 8 on and after September 3rd. This was at least two weeks earlier than at site 10 B. The radiosonde records for this height above the Juneau station show a preponderance of negative temperatures after September 9th *. In 1952 the cold wave

* A high correlation coefficient has been noted between surface air temperature records at icefield camps and upper air data from the Juneau Airport Station (Miller 1954*b*).

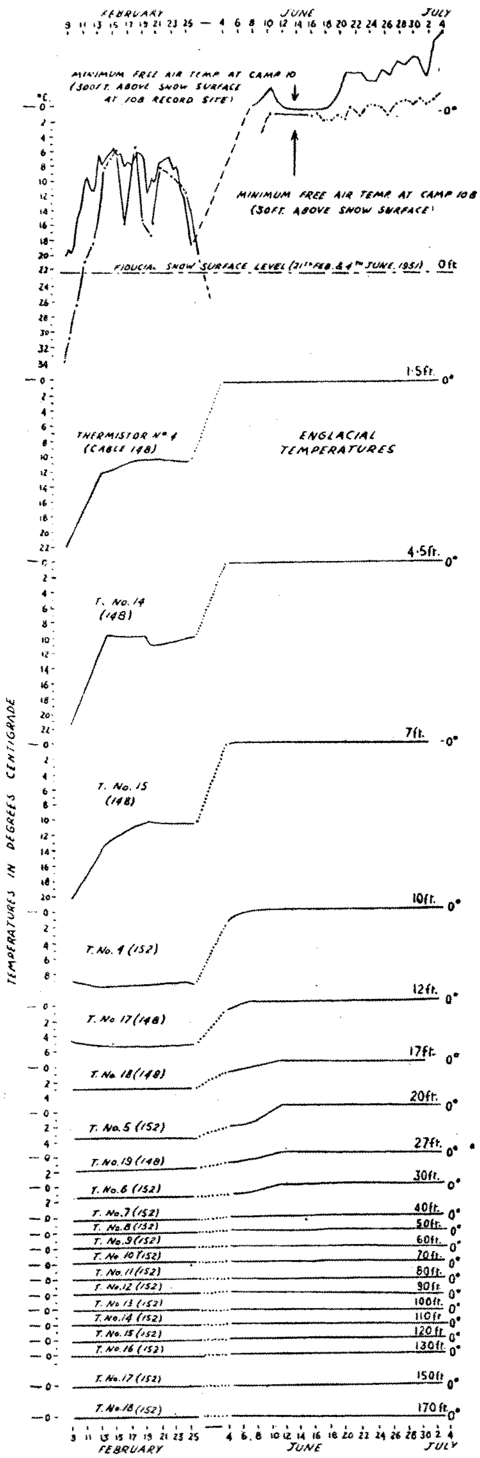



Fig. 5

The winter to summer temperature progression on a vertical profile in the Taku Glacier, Juneau Ice Field, Alaska.

Comparison of Thermistor Values. Camp 10 B. Elev. 3575 ft. 1951.

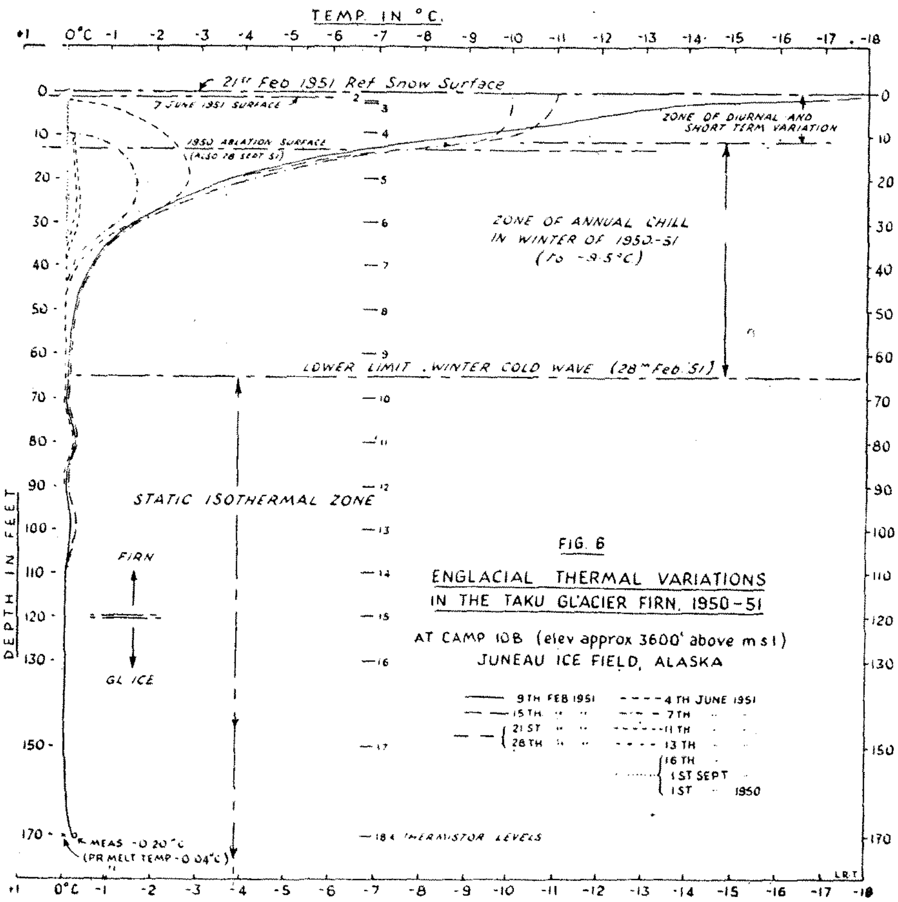



FIG. 6

ENGLACIAL THERMAL VARIATIONS IN THE TAKU GLACIER FIRN, 1950-51

AT CAMP 10B (elev approx 3600' above m s l) JUNEAU ICE FIELD, ALASKA

was initiated on August 16th, nearly five weeks earlier than at site 10 B. In the subsequent two weeks negative temperatures persisted with a minimum of —11°C. and a daily average of —4°C. Between August 20th and September 3rd temperatures rose to above freezing at the surface and a slight warming of the uppermost firn resulted (Fig. 9). This effect was shortlived, however, and did not significantly delay the seasonal penetration of cold since on September 4th the temperatures again dropped, with a low of —14.4°C. occurring in the period of observation up to Sept. 12th. In 1953 a somewhat different condition developed in the late summer and autumn when the annual temperature wave was delayed by factors described later in this paper. In each of the last six years of observation noteworthy summer snow storms have occurred in the 8 B sector in the months of June, July and August and have produced a fresh snow-cover of 1 to 3 feet undoubtedly affecting the glaciothermal regime. At the lower record site these storms only produced warm rain, which instead of adding to the accumulation resulted in further ablation. In 1951 and 1952 snowfalls persisted in the highland after the middle of August, and in each instance developed into a thick depth hoar stratum at the base of the snow-pack for the ensuing budget year.

The depth of penetration of autumn cold at representative dates is indicated in Figure 4. Here measurements obtained in 1953 are plotted below the mid-November reference snow surface and for comparison the limiting curves of initial and maximum penetration of negative temperatures are given for the first half of the 1950-51 thermal year. The length and amplitude of the mid-November temperature wave at both record sites are also shown.

### Magnitude and Duration of Chilling Near the Névé Line

The 1950-51 record represents the thermal regime for the coldest month of the year and for the critical spring-to-summer transitional period in which the annual chill is disssipated. The winter to summer progression of internal temperatures is given in Fig. 5. Changes in corresponding minimum air temperature at 30 feet and 300 feet above the snow surface are shown at the top of this diagram. Aspects of the temperature pattern on selected days in the winter and spring are also given for comparison in Fig. 6.

By reference to these two figures it is seen that the cold wave had reached a depth of at least 60 feet below the mid-winter snow surface by February 9th. A slightly deeper penetration occurred during the ensuing two weeks. The maximum depth, as of February 28, 1951, was 65 feet with the cold wave having a steadily diminishing amplitude. Its lower limit is delineated in Fig. 6 as the base of the *zone of annual chill*. This is differentiated from a surface *zone of short-term variation* which is discussed later.

During February the minimum persistent temperature recorded in the zone of annual chill was —9.5°C. Neither deeper chilling nor lower temperatures were observed upon resumption of measurements in the spring and it is improbable that any significantly greater penetration of cold occurred in March, April, or May. This is corroborated by the fact that mean daily temperatures at the nearby Juneau station showed a systematic rise after the middle of March, as also did those from the radiosonde record at the height corresponding to Camp 10 B.

At all levels below the depth of the annual cold wave no change was shown between the measurements of winter and summer. On thermistors at the 80- and 100-foot level, however, a slightly negative temperature condition was observed, i.e. respectively —0.1°C. and —0.2°C. These values are within the recording accuracy of the equipment. The constancy of record proves that the readings were not the result of current surface influences. At the intervening 90-foot depth, the temperature both in February and July was 0°C. Fully temperate conditions were also observed at all other record levels below 110 feet except at 170 feet. Here, another anomalous condition was found, with the record at —0.2°C. both in winter and summer. The pressure-melting temperature at this depth accounts for only one-fifth of the measured depression (Fig. 6). It is probable that these measurements were due to idiosyncracies in the individual thermistors but the possibility that they represent actually colder horizons is not dismissed. Sandwich structures could produce temperature differences by an apparent entrapment of cold which is slow to disappear due to the varying conductivities involved. In this case it would not be possible to assess the rate of thermal leveling because the physical characteristics are not known. It seems unlikely, however, that over periods of many years such cold zones could persist at levels well below the depth of annual chill.

From the record at the 50- to 60-foot depth only a slight increase in temperature is detected between the last measurement in February and the first one in June. At and above 40 feet, however, a marked rise is shown. Deterioration of the persistent winter chill was therefore well underway before the end of May and fully isothermal conditions were attained by the middle of June, 1951.

Due to abnormally heavy snowfalls which buried the cable marking towers, thermistor measurements were not possible in the springs of 1952 and 1953. The following information was obtained with liquid thermometers in freshly exposed test pits in the 10 B snow-pack. (For comparison the data for 1951 are also noted.)

| Date | Depth below surface | Thermal Condition |
|---|---|---|
| June 7, 1951 | 0-10 ft. | sub-freezing |
| June 16, 1951 | 0-20 ft. | isothermal at 0°C. |
| May 27, 1952 | 0-6 ft. | sub-freezing |
| June 17, 1952 | 0-10 ft. | isothermal at 0°C. |
| June 20, 1953 | 0-12 ft. | sub-frreezing |
| June 25, 1953 | 0-14 ft. | isothermal at 0°C. |

These observations indicate that the 1952 cold wave was dissipated from the winter snow-pack before the end of May and, as in the previous year, all remnants of the annual chill were probably destroyed in the underlying firn prior to the middle of June. In 1953, as shown, the snow-pack reached an isothermal state slightly later in the spring, but no later than June 25th.

### Nature of the Heat Transfer Process

It is a convenient coincidence that the level of ablating snow at the time of first measurements in June 1951 was approximately the same as the reference surface in February 1951. Thus the depth and bulk density relationships of the snow-pack at site 10 B for each of these observation times are easily compared in Figure 7.

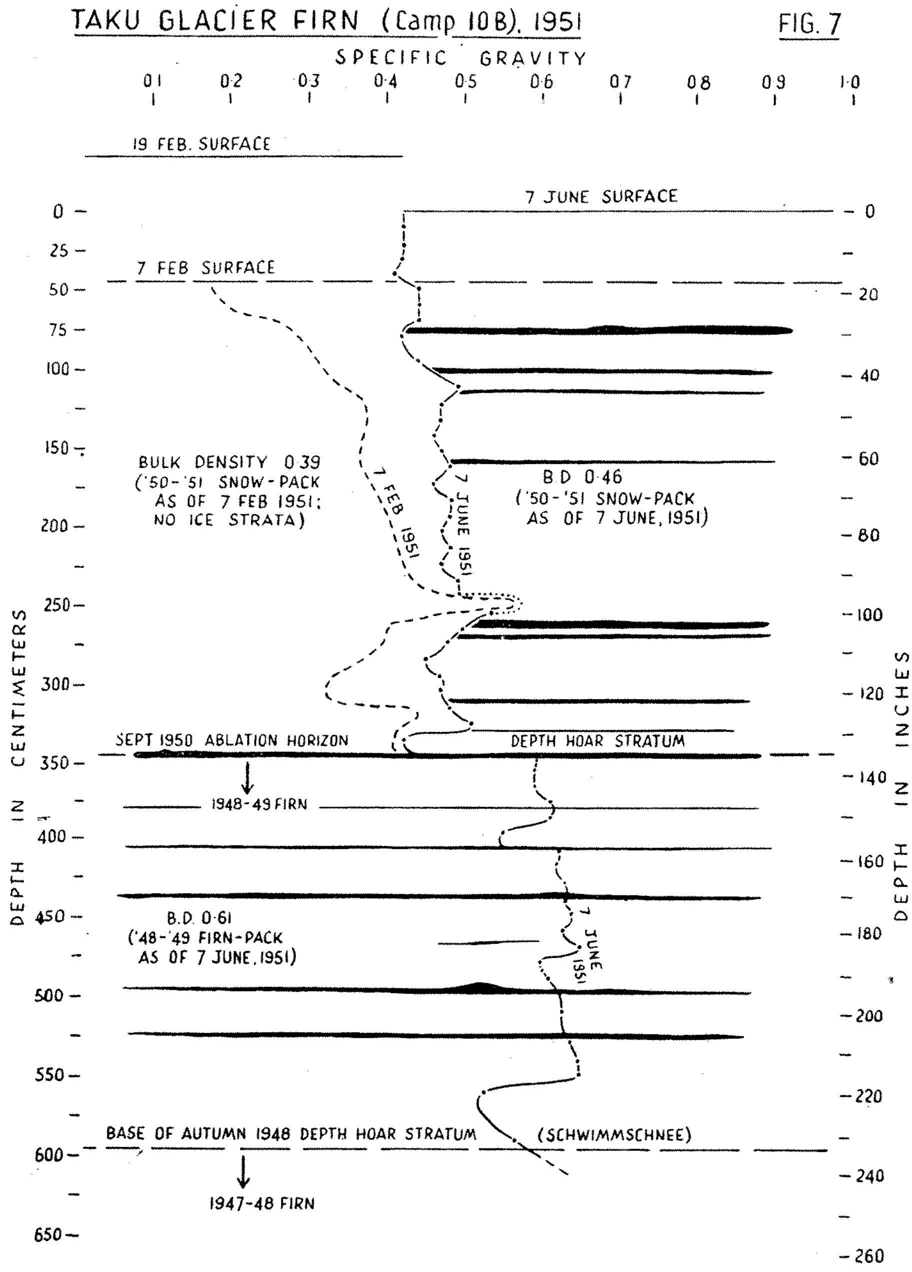
The quantitative difference in firn temperature between February and June is indicated by the upward trending dashed lines in Fig. 5. The later stages of amelioration are shown in those parts of the temperature-depth curves for the first half of June. Fig. 6 gives in more detail the manner of disappearance of the winter cold zone. This is shown ideally as a series of smooth curves for the 4th, 7th, 11th and 13th of June. In detail these curves would probably be much more irregular if more points of measurement were available but in any case the peak of each lies on the same ordinate at a depth of 20 feet. It appears, therefore, that the annual chill was destroyed progressively both from above and below a lingering horizon of cold.

The rate of temperature increase at this depth between June 4th and 11th was on the order of 0.3°C. per day; the firn temperature in that interval having risen from —2.7°C. to —0.22°C. (The minimum at this level in February was —4.12°C.) Between June 11th and 13th, only a slight rise was observed. At 1400 on the 13th, the readings between 10 and 40 feet averaged —0.15°C. By the same hour on the 16th they were 0°C. Therefore the last semblance of the 1951 chill had, on this date, completely disappeared. During the rest of the summer the firn remained at the melting point.

It is significant that in the interval between June 12th and 16th a total of 13 inches of heavy, wet snow, sometimes accompanied by rain, fell at Camp 10 B. The mean air temperature during this storm was 1.0°C. so that the snowfall was also characterized by surface melting. Therefore a quantity of water (upwards of 2 inches) was flushed into the firn over a period of 96 hours. To this is attributed the final attainment of isothermal conditions. It has been shown by other investigators that during the spring, after the surface is thawed, thermal equilibrium at depth is attained primarily through liberation of the heat of fusion of percolating surface water when it re-freezes (Sverdrup, 1935; Hughes and Seligman, 1939). The situation here discussed demonstrates this process. The cold intensity is also diminished by conduction of heat upward from isothermal firn beneath the chill zone, but this is negligible compared to the quantity of heat introduced by percolating water.

Because it is so dependent upon surface meteorological conditions and on local differences in the rate of heat conduction within the glacier, the nature of annual chilling, and its subsequent disappearance, varies greatly in different sectors and elevations of the icefield. Since the rate of cold dissipation is largely dependent upon the volume of water passing vertically downward into the firn, it is also affected by the distribution of horizontal and transverse ice structures. These create barriers to block and channel the water (Fig. 7). From the temperature curves in Figure 6, as already mentioned, it is clear that heating occurs both from above and below. Relating to this it is certain that some heat is conveyed laterally at depth by mobile water. This is facilitated by the drainage and impounding of water in crevasses (Fig. 2), from which source it passes selectively, by capillarity or direct percolation, along the more permeable layers. A controlling factor is the depth of the water table which is seldom greater than the maximum penetration of the cold wave (i.e. 60 to 70 feet) and which at the height of the ablation season fluctuates well up into the zone of annual chill.

In the spring, of course additional heat is absorbed from solar radiation; but this is known to be restricted to the first few feet and to be of minor importance compared to heat transmitted directly by conduction from the outside air. However, when cracks begin to be exposed by partial removal of the snow-cover some of the deeper sections of the chilled zone become warmed by sunlight, and surface air is allowed to circulate down into the fissures. The ordinary process of heat con-




duction can therefore affect portions of the deeper firn whether or not they are directly influenced by water percolation in the intra-crevasse zones. The final destruction of the cold wave is thus hastened, a stage which is well illustrated in the progression of late season curves in Figure 6.

The irregularity in thermal dissipation occasioned by the presence of mobile water and by the various physical inhomogeneities in the glacier precludes close quantitative agreement with hypothetical temperature curves calculated from assumed bulk diffusivity. Computed values are near to observed conditions only in the coldest months when there is no liquid water present to prevent the changes in internal temperature from being controlled mainly by conduction. The detailed mathematical analysis of the heat transfer process is therefore not the concern of this paper, beyond mention of the fundamental thermodynamic properties which pertain to these observations.

In snow and firn, internal temperature changes may be considered as occurring in a semi-infinite homogeneous isotropic solid influenced by fluctuating external temperatures which are a harmonic function of time. The basic physical factors controlling development of the sinusoidal cold wave are thermal conductivity ($k$) of the medium, its specific heat ($c$) and its density ($\rho$). It can be shown (Carlsaw and Jaeger, 1947, pp. 8-9) that the transmission of heat actually depends on the diffusivity, K, defined by a combination of these quantities in the following relationship:

$$K = k/c\,\rho$$

Here the specific heat is taken as constant at the value for ice. The density and conductivity, however, are variable, depending largely on the age of the medium and related genetic factors. The resulting diffusivity is therefore more or less a function of these factors. The relative diffusivity between dense firn and ice is not so large that mass heat transfer and temperature changes in firn are greatly altered by the presence of a few ice strata. On the other hand, a glacier with a substantial covering of snow will display important differences in surface heat transfer compared to one with only solid ice exposed. The general relationships are illustrated in the following table of thermal constants for snow, firn and ice at 0°C. The given conductivities for snow and average firn are based on data from the Snow, Ice and Permafrost Research Establishment (1951, p. 56). The density of firn-ice is chosen arbitrarily between given values for average firn and ice. From this a linear interpolation is made for the corresponding thermal conductivity. For comparison, the thermal properties listed in the International Critical Tables for rubber, steel, aluminum and copper are also noted. The diffusivity figures are rounded off for convenient reference.

| | Conductivity $k$ ($cal^{o}C^{-1}cm^{-1}sec^{-1}$) | Specific Heat $c$ ($cal^{o}C^{-1}gm^{-1}$) | Density $\rho$ ($gm\ cm^{-3}$) | Thermal Diffusivity K ($cm^{2}sec^{-1}$) | Relative Diffusivity to Ice (approx. ratio) |
|---|---|---|---|---|---|
| New Snow | 0.0003 | 0.5 | 0.20 | .0030 | 0.27 |
| Old Snow | 0.0006 | 0.5 | 0.30 | .0040 | 0.36 |
| Average Firn | 0.0019 | 0.5 | 0.55 | .0070 | 0.64 |
| Firn-ice | 0.0038 | 0.5 | 0.75 | .0100 | 0.91 |
| Ice | 0.0050 | 0.5 | 0.92 | .0110 | 1 |
| Water (0°C) | 0.0014 | 1.0 | 1.00 | .0014 | 0.13 |
| Rubber | 0.0005 | 0.40 | 0.92 | .0014 | 0.13 |
| Steel(mild) | 0.1100 | 0.12 | 7.85 | .12 | 11 |
| Aluminum | 0.4800 | 0.21 | 2.70 | .86 | 78 |
| Copper | 0.9300 | 0.09 | 8.94 | 1.14 | 104 |

Applying these values, the seasonal march of surface temperatures may be expected to exert a relatively greater influence on the glacier in the 10 B sector where more ice is exposed and the annual accumulation is much less than in the crestal névé. This may explain why the observed autumn chill, in mid-November, 1953, had reached a deeper level in the firn at 10 B than in the 8 B snow-pack on a corresponding date (Fig. 4). Undoubtedly an important factor is the additional cooling influence of open crevasses in the 10 B sector compared to the highest névé where there is a dearth of exposed crevasses. As the thermal year progresses, however, the trend is reversed and a deeper penetration of the annual temperature wave results at site 8 B. This is to be expected in view of the incidence of much colder mid-winter temperatures here than at lower elevations *. A further reason, as discussed at the end of this paper, is that the rapid development of a thicker snow-pack enhances the intensity of the subsequent temperature wave and also aids in its retention. As a result, although the effects at 8 B and 10 B would not be much different during the first two or three months, they should, theoretically, be quite different seven to ten months after initiation of chilling (Benfield, 1953).

### Thermal Variations in the Winter Snow-Pack

Diurnal temperature changes in the atmosphere during the winter of 1951 caused substantial thermal variations in the shallow winter snow layers at 10 B. The relationship between air and snow temperatures during the coldest month, February, is shown by the shallow curves in Figure 5. The variations were particularly noteworthy at depths of 1.5 to 4.5 feet. The effects of ambient temperature changes over periods of several days were found to be pronounced to a depth of 7 feet; but longer fluctuations of a week or more were required before the influence could be felt at 13 feet. This is well demonstrated by the curve for February 9th in Figure 6. Prior to this date, there had been a week of abnormally cold weather with mean daily temperatures consistently below —18°C. At 7 feet the influence was more immediate; but at 10, 12 and 17 feet there was considerable lag in the time of maximum effect. In fact, at the 17-foot depth, the abnormally low temperatures of the earlier period were not fully reflected for five days, i.e. not until February 14th. At 20 feet the effect was not apparent for eight days and although the mean daily surface temperature had dropped from —21°C. to —35°C. in this period, the englacial temperature at this depth was lowered by only 0.3°C. Subsequent measurements on this thermistor were the same throughout the remaining ten days of observation.

A 26°C. rise in surface temperature between the 9th and the 13th of February was reflected within 36 hours at depths to 7 feet. Warmer surface temperatures prevailed from February 14th to 20th after which date the temperature again dropped off. These variations were also reflected in the shallow snow layers within a few hours. The initial rise, however, did not produce an effect at the 10-foot depth until February 18th. On that date a small change was detected at 12 feet; there was only a very slight one at 17 feet, and below 20 feet the effects again were negligible. The lag increase which is illustrated here may be explained by the Carslaw equations for amplitude diminishment. These demonstrate mathematically how amplitude decreases and lag increases very rapidly with depth when the frequency of surface temperature change becomes larger.

After the first of June the oscillation of surface air temperatures produced no direct influence on thermal conditions within the firn since the surface snow was by then at the melting point and the thermal gradient, requisite to heat transfer by conduction, was no longer present. Sub-freezing nocturnal air temperatures did, of course, facilitate the formation of diurnal crusts but these were usually destroyed by mid-day, after which the warmer air in contact with the snow surface effectively aided in the production of melt-water. Much of the density increase and development of ice strata between February 7 and June 7, as portrayed in Figure 7, is attributed to the refreezing of such melt-water as it percolated downward during this period.

From these observations it is clear that the maximum depth of short-term variation was no greater than the thickness of the 1950-51 snow-pack. It may have been fortuitous that the limiting horizon was so close to the late summer ablation surface of the previous September; however, the marked difference in physical character of the firn below this surface probably helped to increase the diffusivity characteristic (as shown by the table) and to enhance the attenuation of short-term effects. The thermodynamic properties required in a mathematical analysis may be deduced from Figure 7 and the table of comparable heat diffusivities.

---

* In the winter of 1952-53, the comparative minimum air temperatures were: Site 8 B: — 58°F. (—50°C.) and Site 10 B: —42°F. (—41°C.).

## Temperatures in Buried Crevasses

Because of the heavily crevassed nature of the glacier at the lower observation site, the measurements on Cable No, 150 are of special interest. The free air temperature gradient in the crevasse is shown to be less than at corresponding depths in the firn. The magnitude of this difference is illustrated by the data at 1100 hours on February 21, 1951.

| Depth below 21 Feb. Surf. * | Crevasse Temperatures | Depth | Adjacent Firn Temperatures |
|---|---|---|---|
| Feet | (°C.) | Feet | (°C.) |
| 2 | —9.89 | 2 | —9.92 |
| 8 | —6.08 | 8 | —7.82 |
| 13 | —2.02 (top of crevasse) | 13 | —5.70 |
| 19 | —0.47 (in crevasse) | 19 | —3.45 |
| 24 | —0.26 » | 24 | —2.09 |
| 29 | —0.43 » | | |
| | | 30 | —1.64 |
| 34 | —0.09 » | | |
| | | 40 | —0.63 |
| 44 | —0.11 » | | |

The relative uniformity of crevasse temperatures and the existence of nearly isothermal conditions below 34 feet are attributed to the convection of free air within the crevasse. A slight thermal gradient was retained, as shown in the above table; but this was undoubtedly due to the temperature gradient in the nearby firn. The top of the crevasse was completely covered over by a blanket of snow so that convection currents could only be due to the air temperature gradient inside, without influence from winds or the direct introduction of colder air from the glacier surface. The snow-cover thus facilitated the presence of unfrozen water in the bottom of the crevasse and of others in this area which extended below the depth of annual chill. The drainage of sub-glacial outlet streams which flow from the icefield throughout the winter months is favored by this condition.

As in the snow-pack over areas of solid firn, the short-term variations of surface air temperature were also reflected in snow above this crevasse. The influence was most noticeable to a depth of 8 feet. Only a slight change was observed in snow at 13 feet. At 19 feet, within the crevasse, temperatures were distinctly higher than in adjacent firn and did not vary during the fortnight of February observations (Fig. 8).

That the warm front advances irregularly while destroying the zone of annual chill is well illustrated by measurements on and after June 4th. On that date, a thermal gradient no longer existed in the crevasse since all temperatures, including those for some distance inward from the side-walls, had risen to the melting point. This condition was observed while sub-freezing temperatures still persisted at corresponding depths in firn at the bore site 75 feet away.

## Observations on the Crestal Névé

A program of detailed investigations on the 6000-foot névé has not yet been carried out; however, as already mentioned, preliminary studies have shown that the annual cold wave is initiated some weeks earlier at 8 B than at 10 B. Conversely, at the end of a given thermal year the winter snow-pack in the highest sector achieves an isothermal state some weeks later than at the névé line. This is allied to the thicker

* The first two values in each column represent measurements in the continuous snow-pack over the crevasse.

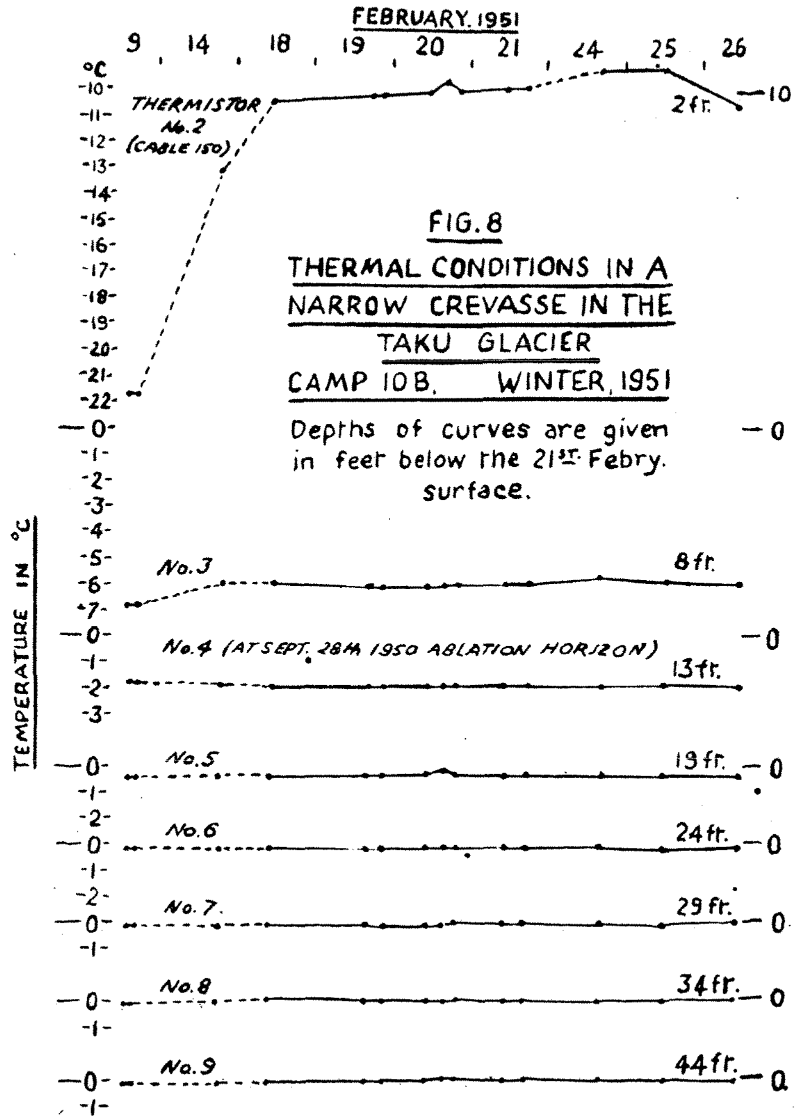
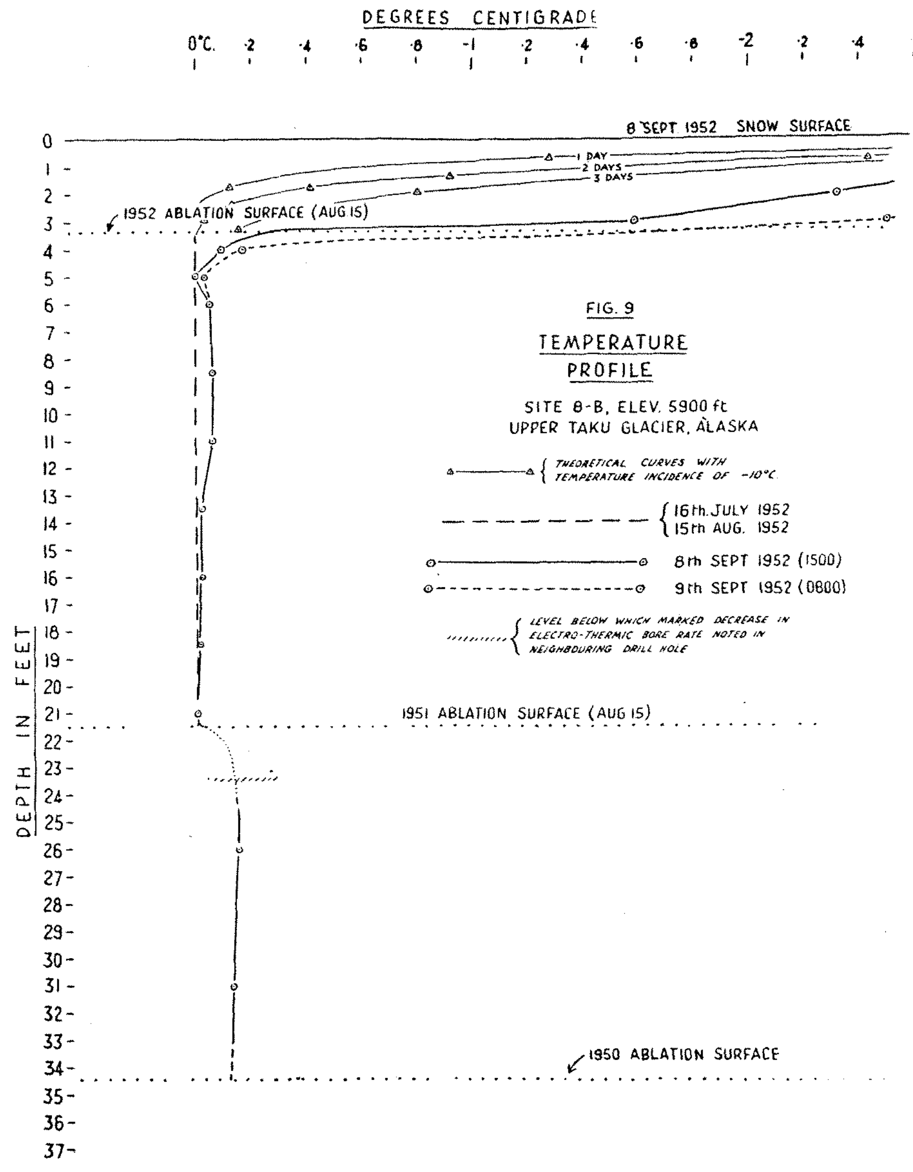



FIG. 8
THERMAL CONDITIONS IN A NARROW CREVASSE IN THE TAKU GLACIER
CAMP 10B. WINTER, 1951

snow-pack as well as to a shorter ablation season, which at the crestal level is usually of less than eight weeks duration.

In the summers of 1951 and 1952, test pits were dug to a depth of 20 feet. In these, it was found that the snow when first uncovered had become isothermal before mid-July, at least down to the previous year's ablation level. The amelioration thus lagged that at 10 B by about three weeks. In each of these years, isothermal conditions existed in the snow-pack until August 15th. After this date, intermittently subfreezing surface temperatures occurred, becoming more or less persistent in the second week of September. In September 1952, the measurements were obatined with thermistors to a depth of 31 feet. The readings were made three days after the cable was installed to allow for any necessary stabilization of the firn's thermal




character which might have been disturbed by the mechanical boring. Part of the record for September 8th and 9th is graphed in Figure 9. Theoretical cooling curves for intervals of one, two and three days are also shown. These are computed on basis of the average snow density and, for purposes of simplification, on an assumed persistent —10°C. cooling at the surface. They are seen to compare favorably with the trend of measured curves and verify the fact that they are chill curves of the new thermal year. The fact that their amplitude falls off near the 1952 ablation horizon is only coincidental.

All thermistors in the upper 3 feet of snow and the subjacent 2 feet of firn showed strongly negative temperatures. At 5 feet, the temperature was 0°C.; but below this the glacier appeared to be slightly sub-freezing to a depth of 11 feet. This was probably near the base of the seasonal chill then underway, since the next 7 to 10 feet of firn were close to the freezing point. The thermistors at 26 and 31 feet, however, were in a definite zone of «cold» firn with temperatures of —0.1° to —0.2°C. The possibility of cold conduction down the cable to this depth is eliminated in view of the warmer zone above; and the probability of thermistor error is reduced since the same order of temperature is indicated on each measuring unit. This was an unexpected condition and one needing corroborative data. It is of interest that at 23 feet, approximately the depth at which the lower temperature depression was noted, a sharp decrease occurred in the rate of advance of an electro-thermic device used for drilling holes in adjacent firn. The velocity change occurred close to the 1951 ablation horizon and was undoubtedly related to density increase. However, the colder firn also played a part in reducing the penetration rate. This is borne out by the fact that the boring rate increased once again at a depth of 40 feet where the drill seemed to pass into a zone of warmer firn. The amount of heat lost to the warming up of the glacier in this cold zone would be rather small since only 0.5 calorie is required to raise the temperature of one gram of ice through 1°C. Therefore, a significant change in boring rate should not be expected from this cause alone. Instead, the slowing down is attributed to the colder firn at this level creating a freezing condition along the unheated upper portion of the guide shaft, which in turn caused masses of partially melted firn to stick to the pipe and increase its frictional drag along the bore wall. A further corroborative fact is that the borer was observed to have frozen in when left in the hole for several hours with the current turned off. This conclusively shows that negative temperatures existed at depth.

A theoretical analysis of this conditions helps our interpretation. It is known that the movement of heat through snow is a very slow process. The rate may be calculated when the thermal constants are known (Wilson, 1941). It can be demonstrated from Carslaw's periodic oscillation equation (Carslaw and Jaeger, 1947, p. 48) that the amplitude decrease in a temperature wave with depth would be much too fast for the oncoming annual chill to have penetrated to a depth of 26 or more feet in the short period involved. Likewise, the small-diameter of the hole in which the drill became frozen would have precluded shortterm cooling due to convection of surface air. The effective conduction of colder temperatures for such a distance along the electrical lead is also improbable because of the low conductivity of its thick rubber insulation compared to ice (see table). From thermodynamic considerations it is therefore unreasonable to expect that the subfreezing temperatures measured at and below this level were the result of the intermittent chilling in the three weeks prior to Septebmer 8th.

From the foregoing it is concluded that the colder zone below the 1951 ablation horizon represents a remnant of the previous year's thermal depression which had not been fully dissipated during the short ablation season of 1952. On this assumption the amelioration in this year reached approximately 23 feet below the September 8th surface, or 20 feet beneath the 1952 late summer ablation surface. This means that melt-water percolating to this depth would be mostly re-frozen and retained in the firn as an increase in bulk density. Under such conditions, the net loss due to ablation would be much less than at lower elevations on this icefield. Of course, if this has occurred over a number of years the effect on the long-range state-of-health of the Taku Glacier has been considerable.

In the autumn of 1953, supplementary measurements were made in the uppermost strata of the Camp 8 snow-pack. From these development of the 1953-54 cold wave is extrapolated. By November 11th, the penetrating chill had attained a depth of only 12 feet as compared with 20 to 30 feet below the mid-November surface at 10 B (Fig. 4). It was not possible to obtain deeper measurements to determine whether in this year a buried cold zone existed, either as a relict condition from the pervious winter or as a result of a colder spell which might have occurred in the early autumn. But it is clear that the 1953-54 cold wave had not much effect prior to mid-November and that probably the ablation season in 1953 extended much later in the autumn than usual. A period of relatively warm weather in September followed by excessively heavy and continuous snow-fall in October, after the initial chilling, is believed to have been responisble for delay in the cold wave's full development. This is substantiated by the meteorological records which show a much warmer and wetter condition than is normal for this time of year.

## Significance of the Measurements

It has been demonstrated that heavy snowfalls on the surface of the uppermost névés of this icefield tend to increase the amplitude and lag effects in the annual cold wave and therefore effectively to delay its penetration. But eventually, the combination of much colder winter temperatures, shorter summer ablation season, and more rapid accumulation in the 8 B sector, results in a more dominant englacial chilling which is retained if not throughout the full year at least for all but a very few weeks in summer. The measurement of persistently sub-freezing temperatures is particularly significant when examined in the light of recent theoretical calculations. For example, Benfield (1951) has shown that in regions of excessive accumulation where the rate of net gain approximates the downward velocity of propagation of the cold wave, we may expect intensification of internal temperatures and an appreciably greater depth of effect than in areas where accumulation is not large. This is due to the fact that the temperature wave is buried under a heavy blanket of new snow so that, in effect, its point of origin is lowered. Additionally the insulation provided by a thick mantle of snow with low diffusivity aids in the retention of cold. It has already been mentioned that accumulation in this sector of the glacier has been excessively positive during recent years. The following comparative measurements, covering the period since 1947, confirm this fact.

**Net Retained Accumulation on the Taku Glacier (in cms. of water)**

| | 1947-48 | 1948-49 | 1949-50 | 1950-51 | 1951-52 | 1952-53 |
|---|---|---|---|---|---|---|
| 10 B Névé (3600 ft.) | +89 | +149 | 0 | —47 | +62 | +55 |
| 8 B Névé (5900 ft.) | +190 * | +270 * | +290(?*) | +220 | +248 | +275 |

Firn often persists to a greater depth in polar than in temperate glaciers, not necessarily because of differences in accumulation but rather because the effects of melt-water are diminished and densification is more largely the result of sublimation and compaction. It is of interest, therefore, to compare the probable thickness of firn-cover on the uppermost névé with the 120 feet known to exist at Camp 10 B (Miller, 1954*a*). A straight linear projection of surface firn densities at 8 B shows a minimum firn-pack depth of 130 feet, while the interpretation of boring rates from the electro-thermic drill used at this site indicates that firn-ice may be encountered at depths in excess of 170 feet (Miller, 1952). The considerable thickness thus implied bears out the inference that the Upper Taku Glacier is influenced by a slightly negative englacial temperature regime. Another supporting fact is that fewer transverse ice structures have been found in this firn than in the broad extensions of the névé at lower elevations. This may be due to the shorter ablation season which, as in polar areas, makes less melt-water available for refreezing at depth.

These studies have shown that at intermediate and low elevations the Taku Glacier is geophysically temperate—i.e. isothermal at all depths during the summer months and with temperatures below the winter chill zone remaining constantly at the pressure melting point. But since the thermal observations in the highland sector are to be followed by more detailed investigations, classification of this portion should perhaps await these further data. Nevertheless, fair evidence has been found that the summit firn, in some years at least, is of sub-polar character. Although a milder climate has developed in Southeastern Alaska since the mid-eighteenth century (Lawrence, 1950), it would seem that the changes have not sufficed to maintain a fully temperate englacial regime at this height. Almost certainly more rain and less snow now falls on most of the valley glaciers in this district, causing less accumulation than formerly; but the result has been just the opposite on the Taku Glacier.

* Provisional values, estimated from bore-hole records and the linear increase in density indicated by test pit measurements to a depth of 10 meters.

This anomalous situation may be due to autumn and spring temperatures over the broad upland névé now being closer to the freezing point, thereby increasing snowfall and offsetting the glacier's ablation losses at lower elevation. Of course, other contributary meteorological factors are also involved.

Recent studies at a corresponding elevation on the Seward Firn Field, 200 miles to the northwest, have shown that effects of summer melt-water on the winter chill zone are so great that isothermal conditions are produced in spite of a mean yearly temperature below the freezing point (Sharp, 1951). This condition, however, need not pertain in the higher reaches of the Coast Range. The Upper Taku Glacier is at least 80 miles farther inland from the outer coast so that a more continental winter climate, with a stronger development of the englacial temperature wave, may be expected *. In addition, the Juneau Icefield is in a somewhat different geographical relation to weather variations resulting from shifts in position of the Aleutian low pressure belt.

In conclusion, the glaciothermal data which have been presented corroborate and amplify the evidence of a present zone of maximum snowfall in the icefield's central névé. They also support the concept that this highest accumulation area on the west side of the range has been undergoing a net gain for a number of years in the recent past. This factor is put forth as one primarily responsible for the noteworthy thickening of the main branch of the Taku Glacier below its névé line and for the paradoxical advance of its terminus since the turn of the century. At the present rate of accretion in the source area, this glacier, beyond the effect of minor fluctuations, may be expected to continue its advance or at least to hold its current position for some decades to come. All of the other glaciers on the Alaskan side of the icefield are not only completely temperate in character but are nourished at elevations well below this zone of maximum snowfall. The Llewellyn Glacier in British Columbia, although stemming from the crestal area, lies entirely on the drier lee side of the range and is influenced by climatological conditions not as favorable to glacial growth. In each of these cases, the regime pattern will probably continue as one of regression.

## Acknowledgements

In addition to the sponsoring groups mentioned at the beginning of this paper, the following agencies provided financial and materiel assistance to make possible these investigations: The E. J. Longyear Drilling Company; the U. S. Geological Survey; the Arctic Institute of North America; the Exploration Fund of the Explorers Club; the Lamont Geological Observatory of Columbia University; and the Foundation for Glacier Research. Especial acknowledgement is given for the assistance of F. A. Milan, F. A. Small and M. C. Brewer. Aid was provided in the mechanical aspects by members of the 1950 and 1951 expeditions of the Juneau Icefield Research Project and in some of the field measurements by members of the writer's 1952 and 1953 expeditions. Gratitude is also extended to the U. S. Geological Survey's Baltimore Geophysical Branch which gave advice on the installation and use of the thermistor equipment and provided information for the preparation of several technical reports concerning this work. Special mention is likewise made of Dr. T. C. Furnas and Dr. J. H. Swartz for their kindness in critically reading this manuscript and of Dr. J. W. Glen and Dr. G. Rigsby for helpful discussions.

## REFERENCES

Benfield, A. E. (1951) «The Temperature of an Accumulating Snowfield», *Monthly Notices of the Royal Astronomical Society, Geophysical Supplement*, Vol. 6, No. 3, pp. 129-147.

—, (1953) «The Effects of Accumulation on Temperatures within a Snowfield», *Journal of Glaciology*, Vol. 2, No. 14, pp. 250-254.

* Negative temperature conditions throughout the summer have been reported in another sub-arctic glacier situated a like distance from the coast and at approximately the same elevation and latitude in Norway (McCall, 1952).

Carslaw, H. S. and Jaeger, J. C. (1947) «*Conduction of Heat in Solids*», Oxford, Clarendon Press.

Hughes, T. P. and Seligman, G. (1939) «The Temperature, Melt Water Movement and Density Increase in the Névé of an Alpine Glacier», *Monthly Notices of the Royal Astronomical Society, Geophysical Supplement*, Vol. 4, No. 8, p. 632.

Lawrence, D. B. (1950) «Glacier Fluctuation for Six Centuries in Southeastern Alaska and its Relation to Solar Activity», *Geographical Review*, Vol. XL, No. 2, pp. 191-223.

MacCarthy, G. R. (1952) «Geothermal Investigations on the Artcic Slope of Alaska», *Transactions, American Geophysical Union*, Vol. 33, No. 4, pp. 589-593.

McCall, J. G. (1952) «The Internal Structure of a Cirque Glacier», *Journal of Glaciology*, Vol. 2, No. 12, p. 127.

Miller, M. M. and Field, W. O. (1951) «Exploring the Juneau Ice Cap», *Research Reviews*, April, Office of Naval Research, Department of the Navy, Washington, D. C.

Miller, M. M. (1952) «The Application of Electro-thermic Boring Methods to Englacial Research», *Arctic Institute of North America Report No.* 2, Office of Naval Research Project 86.

—, (1954*a*) «Juneau Ice Field Research Project, 1950», *J.I.R.P. Report No.* 7, American Geographical Society. (See chapter on Mechanical Core Drilling on the Taku Glacier).

—, (1954*b*) «Juneau Ice Field Research Project, Summer 1951», *J.I.R.P. Report No.* 10 (in press), American Geographical Society. (See discussion in Climatology chapter).

Sharp, R. P. (1951) «Thermal Regimen of Firn on Upper Seward Glacier, Yukon Territory, Canada», *Journal of Glaciology*, Vol. 1, No. 9, pp. 476-487.

Snow, Ice, and Permafrost Research Establishment, Report No. 4 (1951) «*Review of the Properties of Snow and Ice*», Corps of Engineers, U.S. Army.

Sverdrup, H. U. (1935) «The Temperature of the Firn on Isachsen's Plateau and General Conclusions Regarding Temperature of the Glaciers on West Spitsbergen», *Geografiska Annaler*, Vol. 17, No. 1-2, p. 71.

Swartz, J. H. (1954) «A Geothermal Measuring Circuit», *Science*. (in press).

Wilson, W. T. (1941) «An Outline of the Thermodynamics of Snow-Melt», *Transactions, American Geophysical Union*, Part I-B, pp. 182-95.

**Abstract**

Investigations carried out in 1950-53 on the thermal regime of the Upper Taku Glacier of the Juneau Icefield are described. A zone of annual chill to a depth of 65 feet (20 m.) was observed at intermediate elevations (3600 feet) and evidence found of the retainment of sub-freezing englacial conditions in the crestal firn (6000 feet) during the 1951-52 thermal year. This is attributed to the relatively colder temperature and to the shorter summer ablation season in this sector, with the effect being intensified by excessive accumulation. The influence of crevasses and local «water tables» on the thermo-dynamic problem is also discussed. The lower reaches of the glacier are classified as geophysically temperate, while englacial temperature conditions in the primary highland névé are considered as marginal between temperate and polar. The findings are shown to bear significant relationship to the thickening and advance of the Lower Taku Glacier since the turn of the century.