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HEAT BALANCE AND ABLATION ON AN ARCTIC GLACIER

BY

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WITH 19 FIGURES IN THE TEXT

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CONTENTS

Abstract.

On the Britannia Gletscher in Dronning Louise Land a study is made of the heat balance at the glacier surface during the ablation season. From micro-meteorological observations in a 4-metre air layer bounded by the glacier surface, the mechanisms of heat supply, expressed as a percentage of the total heat, are: radiation 69; convection 29; condensation 2. The heat is used in: melting 54; evaporation 34; conduction into the ice 12. These values are compared with observations on other glaciers. A statistical analysis is made of the profiles of wind speed and temperature above the glacier surface. From a comparison of observations in meteorological screens and ablation, recorded throughout the summer of 1953, and the use of approximate coefficients of transfer of heat and water vapour, it is found that there is little agreement between calculated and observed values of ablation.

THE PROBLEM INVESTIGATED

One of the aims of the glaciological programme of the British North Greenland Expedition was to make a study of the ablation of an arctic glacier and the meteorological elements responsible for it, with a view to determining the heat balance at the surface of the ice. This work, similar to that of Sverdrup (1936a, pp. 5—49) in North East Land, Wallén (1948, pp. 567—575) in Sweden and Eriksson (1942, pp. 29—49) in East Greenland, was carried out in Dronning Louise Land on the Britannia Gletscher (Lat. 77°15′ N., Long. 24°15′ W) an outflow from the inland ice. Dronning Louise Land differs climatically from the other areas mentioned in that it is completely bounded by large glaciers from the inland ice which keep out much of the influence of maritime air; the sea is 90 km further east.

Theory.

The surface of a glacier receives heat from one or more of the following sources:

(1) Short wave radiation from sun and sky;

(2) Molecular conduction from the air above the surface if temperature increases with height;

(3) Eddy convection from the air above the surface if temperature increases with height and the air is made turbulent by the wind;

(4) Condensation of water vapour from the air above if water vapour pressure increases with height; latent heat of condensation is released at the surface. If the air is turbulent the supply of water vapour is increased;

(5) Conduction from underlying layers of ice if temperature increases with depth.

It loses heat by

(a) Outgoing long wave radiation;

(b) Conduction or convection where temperature and vapour pressure gradients are the reverse of those given in (2) to (5) above.

Ignoring molecular conduction of heat and diffusion of water vapour, which are small compared with the other quantities, a heat balance equation for a melting ice surface may be written:

Heat required for melting + Heat required for evaporation + Heat conducted into the ice	=	Incoming short wave radiation — reflected short wave radiation — outgoing long wave radiation + convective heat.

Of the terms in the heat balance equation, incoming and reflected short wave radiation can be measured directly. Outgoing long wave radiation can also be measured or can be estimated from upper air data. The amount of heat conducted into the ice can be calculated from the temperature gradient, density and specific heat.

The flux of water vapour may be written

$$E = -\varrho K_E \frac{\partial q}{\partial z} \tag{1}$$

and the heat flux,

$$H = -\varrho C_p K_H \left(\frac{\partial T}{\partial z} + \Gamma\right) \tag{2}$$

where

ϱ = air density

C_p = specific heat of air at constant pressure

Γ = dry adiabatic lapse rate = 0.986×10^{-4} °C cm^{-1}

$\frac{\partial q}{\partial z}$ = vertical gradient of specific humidity

$\frac{\partial T}{\partial z}$ = vertical gradient of air temperature

K_E = coefficient of eddy diffusion for water vapour

K_H = coefficient of eddy conductivity.

The atmosphere is considered stable, neutral or unstable as the temperature gradient is algebraically greater than, equal to or less than the dry adiabatic lapse rate. In a neutral or near neutral atmosphere over a homogeneous surface it has been shown by many workers (RIDER and ROBINSON 1951, p. 388) that at a specific height the coefficients of eddy transfer are approximately equal, i. e.

$$K_E = K_H = K_M$$

K_M is the coefficient of eddy viscosity, expressed by PASQUILL (1949b, p. 250) and similarly by THORNTHWAITE and HOLZMAN (1942, p. 17)

$$\tau = \varrho K_M \frac{\partial u}{\partial z} \tag{3}$$

where τ = shearing stress = Vertical transfer of horizontal momentum

$\frac{\partial u}{\partial z}$ = vertical gradient of wind speed.

The logarithmic law of wind variation with height

$$\frac{\partial u}{\partial z} = a z^{-1}$$

may be theoretically proved (BRUNT 1939, p. 246) and by many workers has been found, sometimes with small modifications, to give good agreement with observations. In the majority of these cases the observed atmosphere has been near neutral (adiabatic) or unstable. Prandtl developed the logarithmic law for the distribution of velocity in an air layer near the ground, introducing the boundary condition that the velocity should be zero at a distance z_o from the surface. This constant of integration is known as the parameter of surface roughness.

$$u_z = \frac{1}{k} \sqrt{\frac{\tau_0}{\varrho}} \log_e \frac{z}{z_0} \tag{4}$$

where u_z = wind speed at height z.

k = von Karman's constant

τ_o = surface horizontal shear stress

ϱ = fluid density.

With the usual assumption of constant shearing stress in the first few metres of the atmosphere (SHEPPARD 1947, p. 209) the coefficient of eddy viscosity may, by combining (3) and (4), be expressed in the form:

$$K_M = \frac{u_z k^2 z}{\log_e \frac{z}{z_0}} \tag{5}$$

where u is the wind velocity at height z, and other symbols have the same meaning as above. The value of this coefficient derived from the law of wind speed variation with height cannot be used for calculation of heat and vapour transfer unless the transfer coefficients of these properties possess certain regularities of behaviour (PRIESTLEY and

SHEPPARD 1952, p. 505), the first requirement being similarity of the vertical profiles. PASQUILL and ROBINSON and RIDER (loc. cit.) showed that in near neutral atmospheric conditions, the profiles of wind speed, temperature and humidity could, with suitable scale adjustment, be superimposed to follow the same law of distribution with height, only the constants being different.

It is necessary therefore, to measure profiles of wind speed, temperature and humidity and find laws relating these quantities with height.

OBSERVATIONS

Sites

Two stations were established on the Britannia Gletscher, 14 km long by 8 km wide, one near its head at an altitude of 620 metres and the other 4 km nearer the snout at an altitude of 460 metres (figure 1). The upper station was considered typical of the upper portion of the glacier bordering on the inland ice, and the lower station typical of the middle portion of the glacier. Inaccessibility during the ablation season and the extremely hummocky nature of the ice there made observations near the snout of the glacier impracticable.

Both stations were sited so that the prevailing NNW wind had a fetch of at least 4 km over similar surfaces. At the upper station the surface varied from undulating soft snow with sastrugi in the early part of the ablation season, through a period of undulating coarse wet snow in the middle of the season, to rough hummocky ice at its end. At the lower station, early summer was characterised by a surface of fairly smooth hummocky ice with the hollows filled in with very wet snow. This surface lasted only until the surface drainage had been developed when it quickly changed to a rough hummocky surface which was typical of middle and late summer.

The upper station was manned continuously (except for two brief periods dictated by the Expedition's logistics) throughout the ablation season of summer, 1953. The lower station was occupied for three periods of 48 hours each when simultaneous intensive detailed observations were made at both stations. The three periods were chosen to be representative, so far as could be ascertained at the time, of the early, middle and late parts of the ablation season.

Method.

At the upper station, self recording instruments were used throughout the ablation season to record incoming radiation, wind speed at

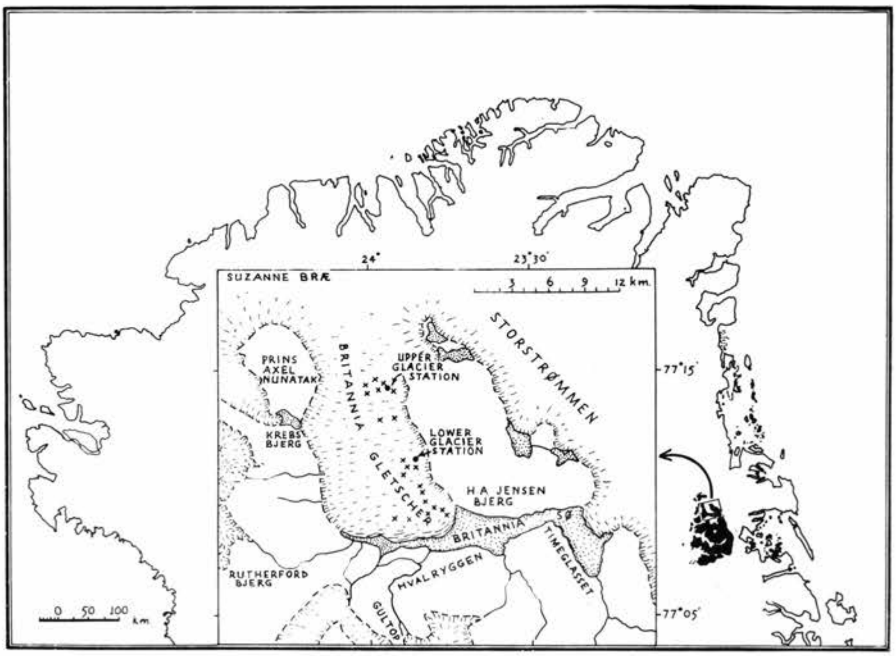


Figure 1. Britannia Gletscher in Dronning Louise Land, North East Greenland.

a height of 100 cm, and temperature and relative humidity at 30 and 300 cm. Cup counter anemometers recorded the run of wind at 30 and 300 cm. Measurements of ablation at stakes, time marks and check readings of temperature and humidity were made two, three or four times daily according to the amount of ablation.

During the three periods of detailed investigation, both stations measured profiles of dry and wet bulb temperature between 2 and 400 cm above the surface, wind speed at heights between 30 and 400 cm, ice temperatures from near the surface to 270 cm depth, ablation, and incoming and reflected radiation. The heights at which temperatures and wind speed were measured were chosen to bracket the heights of the self recording instruments at the upper station since it was hoped to apply the laws derived from the detailed observations to the continuous readings at two heights only.

With the available man power it was not possible to make detailed observations continuously for 48 hours, so readings were taken during every alternate hour of the three periods of intensive investigation. The observed hour was divided into three periods of 20 minutes, 10 of which were occupied in continuous booking of wet and dry bulb temperatures. The other 10 minutes were occupied in measuring temperatures below the ice surface, cloud type and amount, incoming and reflected radiation, wind speed and ablation recorded by stakes and ablatographs. By taking arithmetic means of temperatures and wind speeds at each height recorded over the 10 minute interval, temperature, humidity and wind profiles were obtained which were considered to be representative of the 20 minute period necessary for observation of all the various elements. Ablation was found to be so small that over a 20 minute period the probable error was sometimes greater than the readings. Accordingly it was decided to group the three 20 minute means together to obtain an hourly mean. One hour was considered to be the longest period which could be considered reasonably free from diurnal effects.

APPARATUS

Air temperature and humidity.

Vertical profiles of temperature and humidity up to a height of 4 metres were investigated with an apparatus (figure 2) modified from a design by Pasquill (1949a, pp. 239—248). It comprised a vertical ducting mast with a series of pairs of inlet tubes at 2, 6, 10, 30, 100, 200, 300, 400 cm, above the foot of the mast which rested upon the glacier surface. Air was drawn into the ducting by two electric fans. The inlet tubes were divided into two equal housings, in each of which was a measuring junction of a system of copper-constantan thermocouples. One of each pair of these junctions was arranged as a wet-bulb thermometer fitted with a linen stall, feeding melted ice from a brass reservoir. A reference junction was kept in a vacuum flask of melting ice. The e. m. f. produced by each thermocouple was read in turn on a Tinsley light point galvanometer connected between two selector switches (figure 3) to permit quick reversal of current direction for making check readings and for reading temperature values below the reference temperature. It also permitted the instrument to be used for a range of temperature greater than the scale could accommodate, such as might be met at depth in snow or ice. In that case the switches could be used to select any previously read junction as the reference junction, providing there was not more than 20°C between any two adjacent measuring points.

Temperatures could be read to one twentieth of a degree Centigrade and the galvanometer had a periodicity such that 8 pairs of wet and dry thermocouple junctions could be read in less than one minute.

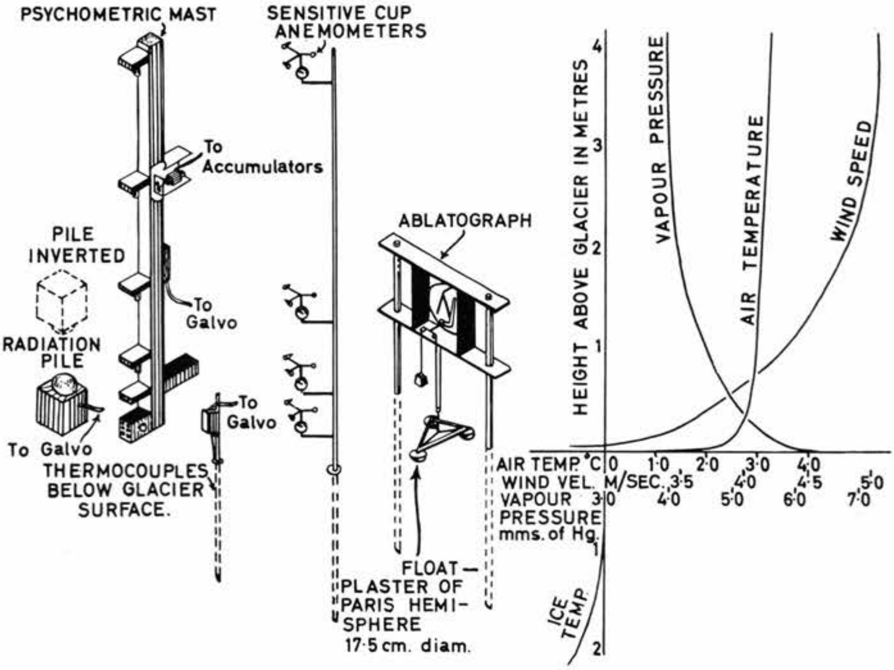


Figure 2. Micro Meteorological investigation of Ablation.

Ice temperatures.

Temperatures in the ice were also measured by thermocouples and the galvanometer. The measuring couples were taped to a bamboo rod which had soft rubber collars attached at frequent intervals to prevent convection currents in the drilled hole in which the rod was placed. Due to heating of the bamboo (principally by direct radiation) the top twenty to thirty centimetres of ice melted and produced a conical, water-filled hole around the rod. Only where the bamboo became frozen into the ice could temperature readings be considered reliable.

Incoming and reflected radiation.

Instantaneous values of radiation were measured at the upper station by a Moll Pile and the lower station by an Eppley Pile. The pile was connected in series with a variable tapped resistance to the galvanometer. Reflected radiation was measured by inverting the pile.

Wind speed.

Four sensitive cup anemometers of a type designed by Sheppard (1940, p. 218) were mounted at heights of 30, 100, 200 and 400 cm. A more efficient investigation of wind velocity profiles could have been made by measuring at more points in the 4 metre air layer but additional anemometers were not available at the time. Indeed, it was considered that the wind profiles would be the least difficult to determine since Bagnold (1941, p. 38), Best (1935, p. 37), Sverdrup (1936b, p. 28), Sutton (1936, p. 126), Pasquill (1949c, p. 126) and Deacon (1949, p. 90) had evaluated wind velocity profiles and all concluded that wind profiles near the ground fitted a logarithmic or a modified form of logarithmic law.

Ablation.

To measure accurately the fall of an ice surface due to ablation two types of instrument have been produced; the Devik ablatograph used by Ahlmann (1935, p. 44) and the Von Huene ablatograph used by Sharp (1951, p. 735).

The Devik type of ablatograph was considered to be the better of the two instruments, but reliance on a single float contact and on only one instrument at any one site of observation was thought undesirable. Six ablatographs, three for use at each station, were designed on the Devik principle. The instrument comprised a box housing a lever system which magnified up to ten times the movement of a float resting on the ice. A graph of the magnified movement was drawn on a 24-hour chart. Each instrument was mounted on two aluminium poles drilled 3 metres into the glacier surface. The float was a light perspex frame in the form of an equilateral triangle, with a single float unit screwed to each corner. Float units of perspex cones, glass tubes, wood hemispheres and plaster of paris hemispheres were tried; the best of these was a plaster of paris hemisphere 18.5 cm diameter.

Insufficient care was taken with the bearing surfaces of the ablatographs so that some of the graphs showed slight jerks instead of a smooth curve. For this reason it was not possible to read off ablation accurately over a very short period. The large flat centrally pivoted circular chart was found to be suitable only in calm or light winds. A drum held chart would be superior since it would stay rigid when the instrument box was opened for adjustment or for re-setting the pen.

Readings from ablation stakes were used where ablation values were required over intervals of many hours, i.e. for periods other than the three detailed investigations.

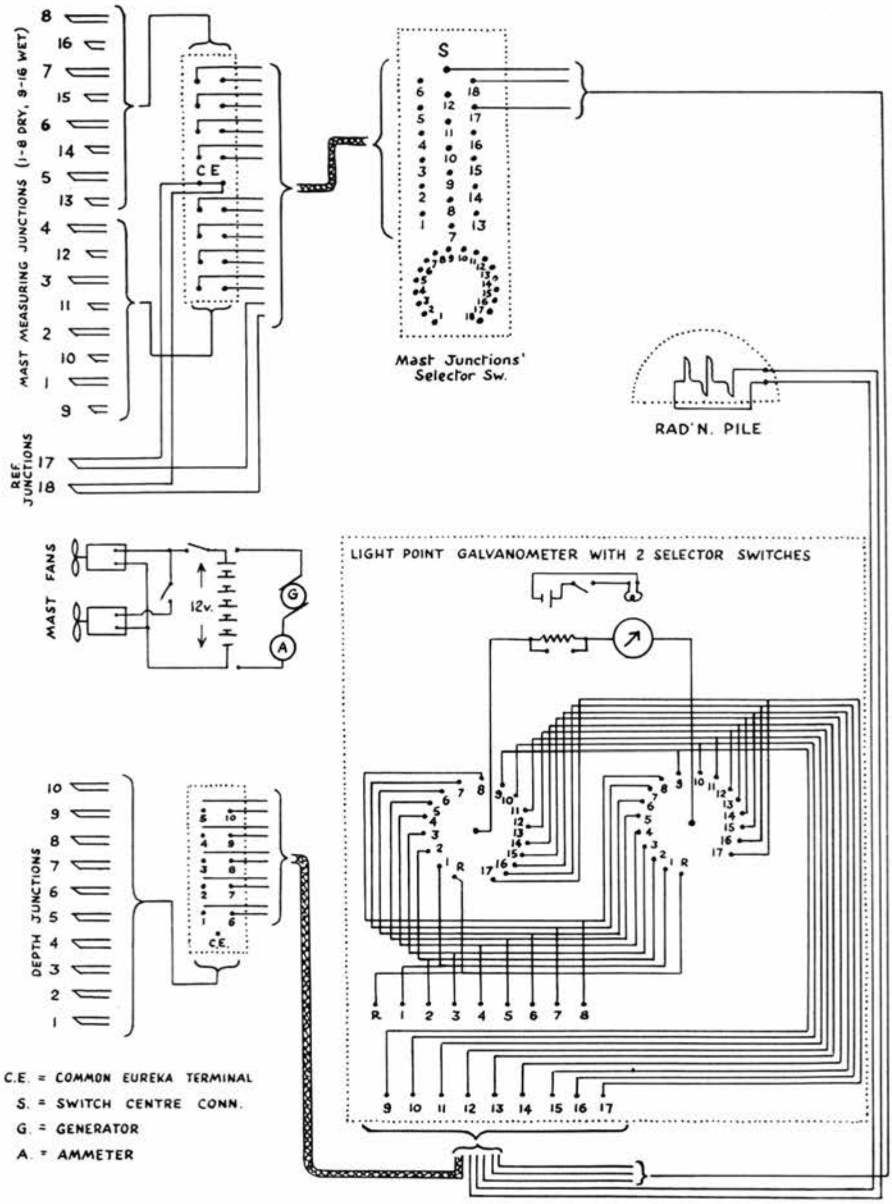


Figure 3. Schematic diagram for apparatus to measure Heat Balance at the ice surface.

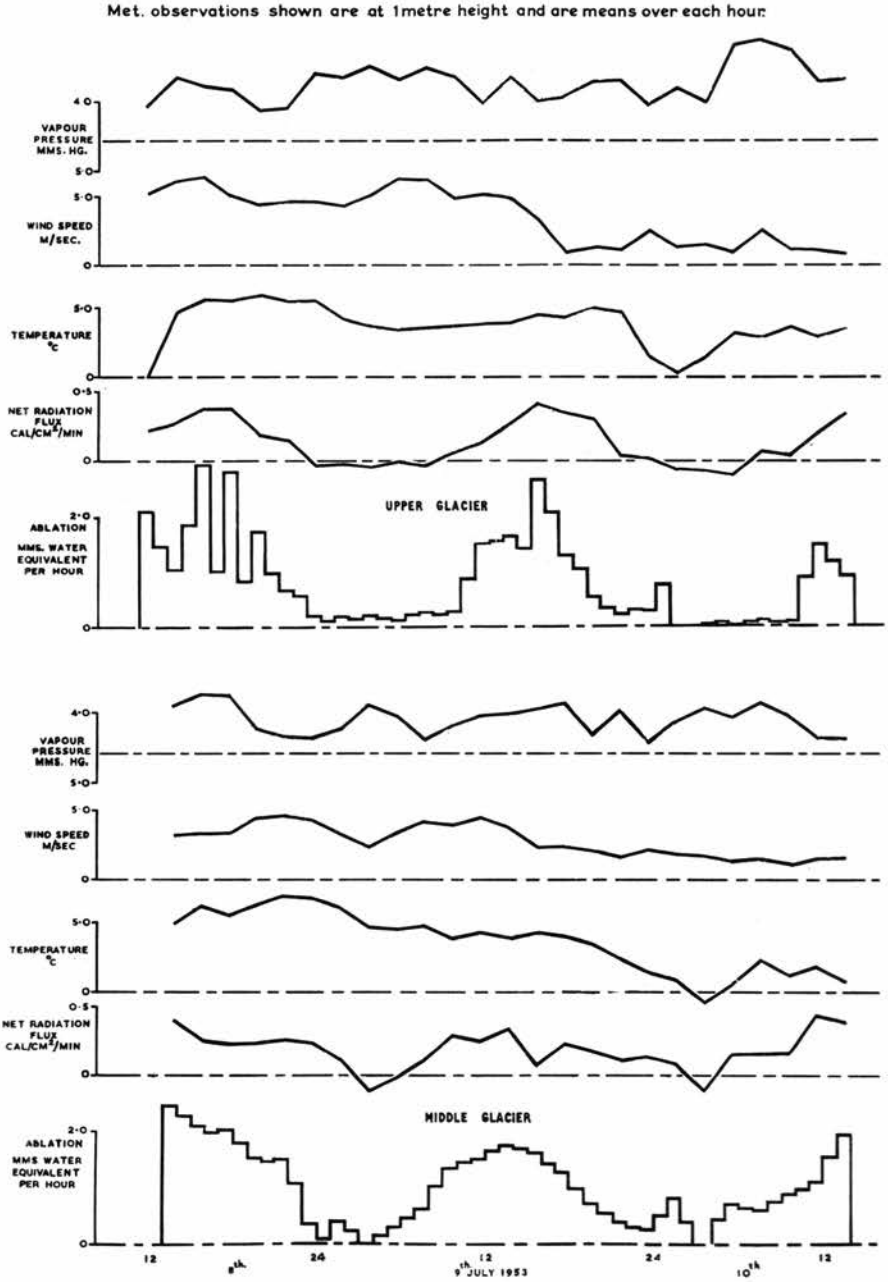


Figure 4. Micro Meteorological and Ablation Values — Middle Summer.

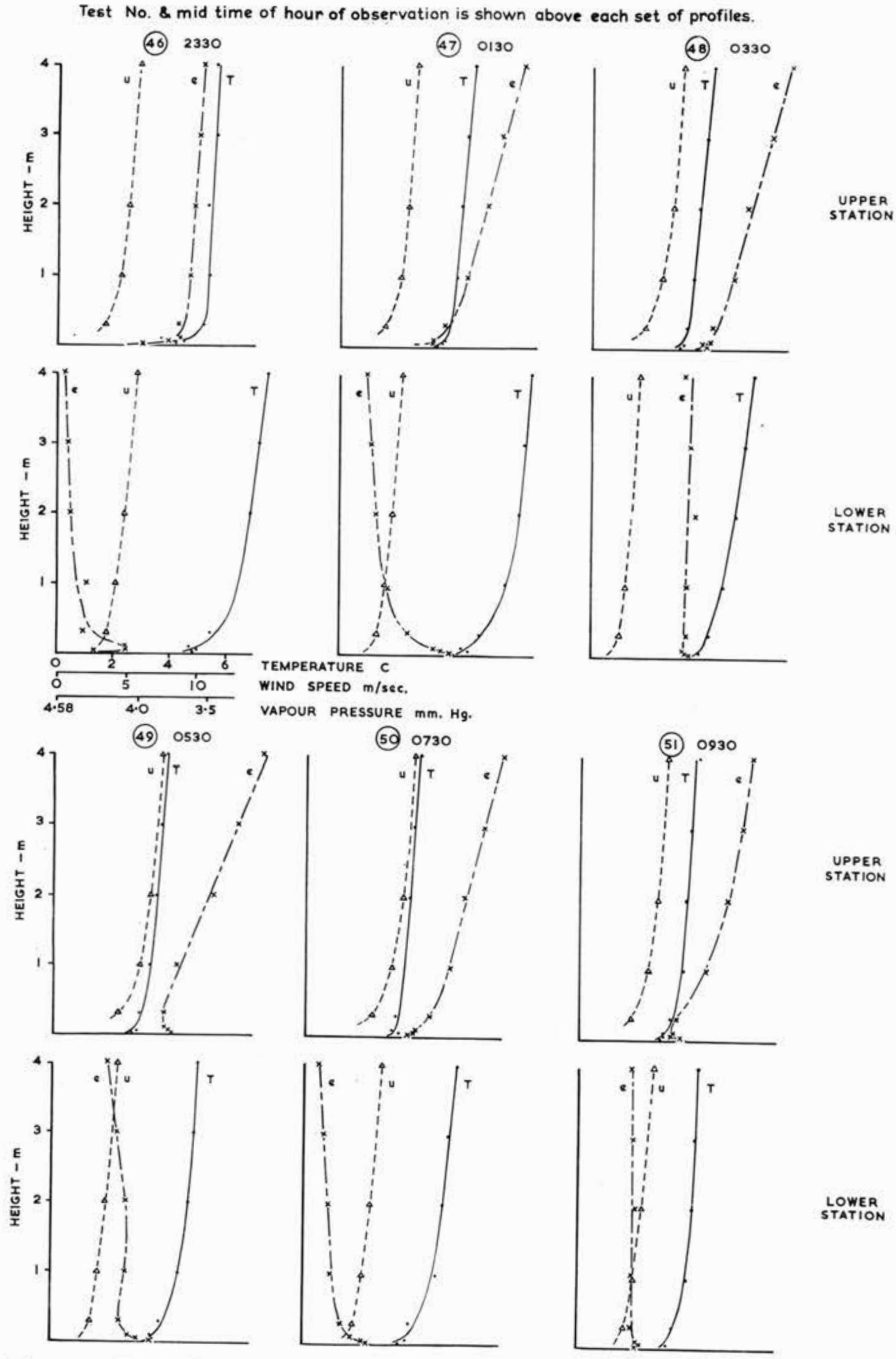


Figure 5. Typical profiles of Wind Speed u, Temperature T, Vapour Pressure e.

OBSERVED DATA

A summary of the data for each of the three periods of investigation was arranged in graphical form but no firm conclusions could be based on these so only that for the middle summer is shown (figure 4).

The middle summer investigation had higher wind speeds and temperatures and lower vapour pressures than the early investigation. Radiation flux, particularly at the lower station, tends to be lower than in the early investigation, which coincided with the summer solstice, although at the time of the early investigation there was much cloud and during the middle summer there was very little cloud. Average temperatures and wind speeds at the lower station are lower than at the upper station, which suggests that the decreased altitude of the middle portion of the glacier did not necessarily ensure higher temperatures by adiabatic warming of the descending air. The more undulating surface of the middle glacier may have been responsible for the lower wind speeds there and, similarly, the fetch of the wind at the middle glacier included more melting ice and thus the air had a greater opportunity to become wet, as suggested by higher vapour pressures at the lower station.

Figure 5 shows typical profiles of wind speed, air temperature and vapour pressure, recorded simultaneously at both stations. The plotted points are mean values for the hour of observation.

The vapour pressure values plotted have been calculated from profiles of the dry and wet bulb readings on the basis of Table 84 in the Smithsonian Meteorological Tables (1939, p. lxxi), due correction being made for the mean barometric pressure of 700 mm Hg.

Many of the plotted values in the lowest few centimetres did not fit any smooth curve. The curves shown have been faired through these points to give a mean profile. In spite of extreme care in making the observations some of these deviations from a smooth profile persisted and are thought to be actual meteorological phenomena and not instrumental errors. Figure 6 shows some of these apparent anomalies observed in summer and winter over cold snow, a melting glacier and a frozen lake. No explanation is offered here of such irregularities in the change in direction of slope of profiles near to a cold surface, which is a matter deserving more detailed study.

Summary of observed data.

During each of the three 48-hour periods of observation there is little diurnal effect, save in ablation and, though to a lesser extent, in net radiation. Radiation would seem to be very important in causing

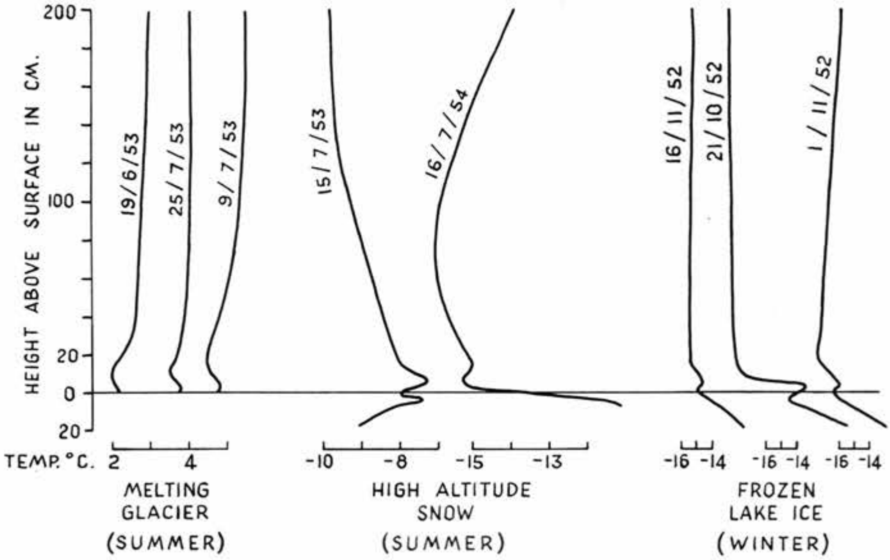


Figure 6. Irregularities observed in Temperature profiles.

ablation, there being no obvious correlation between the graphs of ablation and those of wind speed, temperature and vapour pressure.

The profiles of wind speed, temperature and vapour pressure have a very similar form, frequently becoming linear above one metre height and approaching an infinite slope near to the surface. Departure from the general form of the profiles is most common in the vapour pressure curves, particularly at the lower station where a frequent combination of higher temperatures and lower wind speeds produces a very stable atmosphere; stability is discussed later. Some profiles of vapour pressure display a reversal in direction of slope, which could indicate advection, but this is rarely corroborated by the temperature and wind speed profiles. An increase in the values of vapour pressure with height is uncommon at the upper station and although more common at the lower station, evaporation is more frequent than condensation.

WIND SPEED, TEMPERATURE, AND VAPOUR PRESSURE VARIATION WITH HEIGHT

It has already been noted that the observed values, particularly of temperature, frequently show a linear variation with height in the upper half of the 4-metre air layer.

Values of wind speed, temperature and vapour pressure were plotted on logarithmic and semi logarithmic paper with the height of measurements as the ordinates.

(1) At the upper station,

(a) most of the values of wind speed variation with height below 2 metres could be fitted to a logarithmic law (i.e. wind speed proportional to the logarithm of the height) and frequently were logarithmic over the whole observed range of 30 cm to 400 cm. Only in the early summer investigation could wind speed variation with height be fitted to a power law (wind speed proportional to a power of the height).

(b) Temperature variation with height was generally linear above 1 metre (sometimes above 30 cm) but below one metre, was between a logarithmic and a power law.

(c) The change of vapour pressure with height did not always show a simple relation but below one metre height could best be fitted by a logarithmic law.

(2) At the lower station,

(a) the variation of wind speed with height was usually logarithmic below 2 metres, but frequently followed a power, or a straight line law above 1 metre height.

(b) A logarithmic law could most often be fitted to the temperature variation below 1 metre, but above 1 metre height the dominant relation was linear.

(c) The change of vapour pressure with height was more erratic at the lower station and rarely could be given a simple mathematical relation. Different points of the curves could be fitted by linear, by logarithmic, or by power laws.

The possible error in the observations was added to the plotted points. These errors are: wind speed $\pm$ 5 % of the reading; temperature $\pm$ 0.1°C; vapour pressure, (based upon possible errors in both wet and dry bulb readings) $\pm$ 0.1 mm Hg. It became obvious that the difference between a logarithmic and a power law variation with height was frequently less than the possible errors in the observed values.

From the graphs of wind speed variation against the logarithm of the height, it was seen that most of these curves became a straight line below two metres, some being logarithmic over the whole range of four metres, and it was decided that the logarithmic law gave the best fit. By using the ratio of wind speeds u_2/u_1 at the height ratio z_2/z_1 in the logarithmic form of the law of wind speed, (4) above,

$$\frac{u_2}{u_1} = \frac{\log z_2 - \log z_0}{\log z_1 - \log z_0} \tag{6}$$

and hence

$$\log z_0 = \frac{u_2 \log z_1 - u_1 \log z_2}{u_2 - u_1} \tag{7}$$

Values of z_0 (the constant of surface roughness, Table I) are much higher than expected; evidently the individual small areas of surface have less importance than the undulations of the ice topography. The decrease in z_0 at the upper station and slight increase at the lower station is in accordance with the changes in the mean height of the surface topography as the summer progressed.

NIKURADSE's tests over a pipe surface uniformly roughened with grains of sand (SUTTON 1953, p. 82) indicated that the flow would be fully rough when

$$\frac{u^* z_0}{\nu} > 2{\cdot}5$$

u^* = friction or shearing stress velocity = $\frac{\tau}{\varrho}$

z_0 = roughness parameter

ν = kinematic viscosity

From (4), $u^* = \dfrac{k\,(u_2 - u_1)}{\log_e \dfrac{z_2}{z_1}}$

therefore $\dfrac{z_0 k\,(u_2 - u_1)}{\nu \log_e \dfrac{z_2}{z_1}} > 2 \cdot 5$ for fully rough flow. Taking winds at heights 200 and 30 cm, $z_0 \doteqdot 0.5$ cm, $k = 0.4$, and $\nu = 0.14$ c.g.s. units,

$u_2 - u_1 > 3{\cdot}32$ cm. sec.$^{-1}$ for fully rough flow. In all cases the difference in wind speed is greater (generally greater by a factor more than 10) than the constant required by NIKURADSE and hence the air flow is aerodynamically rough over the snow and over the ice surfaces here considered.

Table I

Changes in the Roughness Parameter of a Melting Ice Surface

Period	Surface	Mean z_0 cm	S. D.	Mean $\frac{U_{200}}{U_{30}}$	S. D.
	Upper Glacier				
Early Summer	Undulating soft snow with sastrugi	1.1	0.25	1.51	0.15
Middle Summer	Undulating coarse wet snow	0.68	0.14	1.50	0.03
Late Summer	Hummocky ice with rough surface	0.58	0.15	1.47	0.06
	Middle Glacier				
Early Summer	Fairly smooth hummocky ice with hollows filled with very wet snow	0.40	0.16	1.39	0.11
Middle Summer	Hummocky ice with rough surface	0.50	0.31	1.47	0.07
Late Summer	Hummocky ice with rough surface	0.57	0.15	1.48	0.04
	Lake Ice				
Early Summer	Flat lake ice with rough surface caused by 'candleing'	0.16	0.02		
Early Summer (3 days later)	,, ,,	0.04	0.005		

The importance of a common profile of wind speed, temperature and vapour pressure in the evalution of the transfer coefficients has already been mentioned. Several previous workers (e.g. RIDER 1954, p. 489) have found their observations to fit one common profile. Profiles of wind speed were plotted on linear scales and the curves of temperature and vapour pressure were superimposed by considering the difference in values between 30 cm and 100 cm height and multiplying this difference by a scale factor for each test. Allowing for the probable error of observation it was generally possible for data at the upper station to draw a curve, below 2 metres height, common to wind speed, temperature and vapour pressure. At the lower station, the vapour pressure profiles, though very similar in the lower layers to those of temperature and wind speed, could seldom be considered coincident.

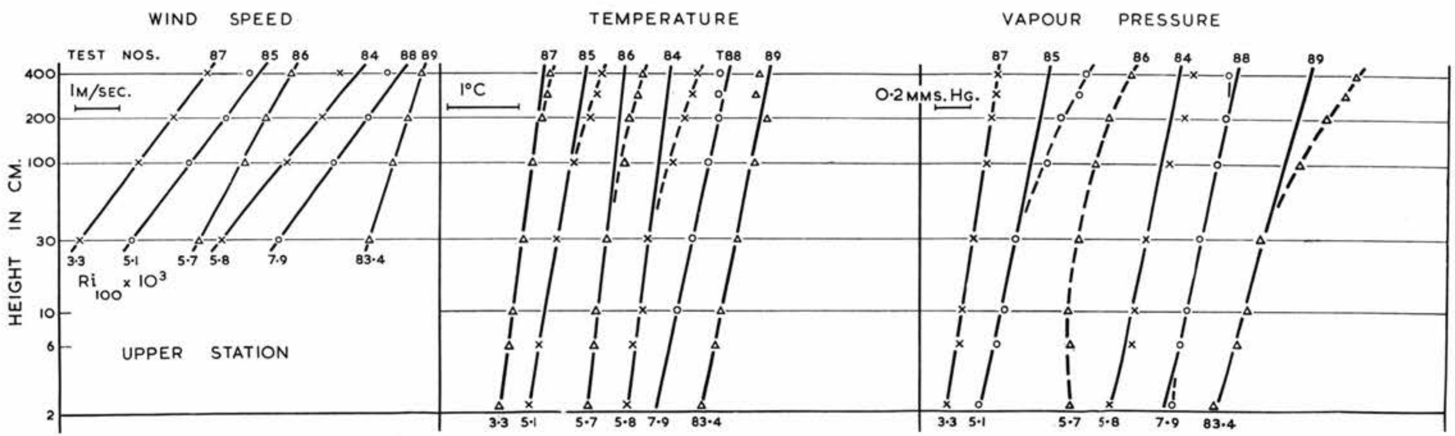


Figure 7. Wind Speed, Temperature and Vapour Pressure profiles over Britannia Gletscher showing deviations from logarithmic form.

It may be concluded that most of the profiles of wind speed, temperature and vapour pressure can be fitted to a common curve below one metre and that this curve closely follows a logatithmic variation with height. Pasquill (1949c, p. 124) found the departure of these properties from the logarithmic law to be related to the stability of the air. Figure 7 shows examples of departures from the logarithmic form for different stabilities. There is a tendency for increased departure from the logarithmic form with greater stability but there are many exceptions to this. Curves of wind speed at the upper station and of vapour pressure at the lower station do not show a curvature related to stability as found by Pasquill, but the average stability considered here is greater than that encountered by Pasquill who was observing over a short grass surface.

Experiments on Britannia Sø.

It was thought that the observations on the Britannia Gletscher may have been complicated by the undulating and hummocky nature of the surface. Furthermore, the glacier observations included only one wind reading in the layer below 1 metre where it had been found that

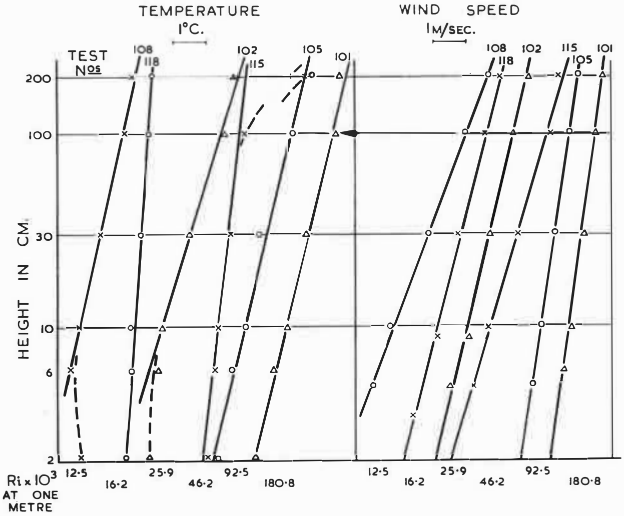


Figure 8. Temperature and Wind Speed profiles over Britannia Sø showing deviations from the logarithmic form.

profiles could be superimposed. A further series of experiments was therefore carried out on the flat ice of Britannia Sø.

The same apparatus was used on the lake ice as had been used on the glacier but the greatest concentration of observations was in the lowest centimetres of a 2 metres air layer bounded by the melting

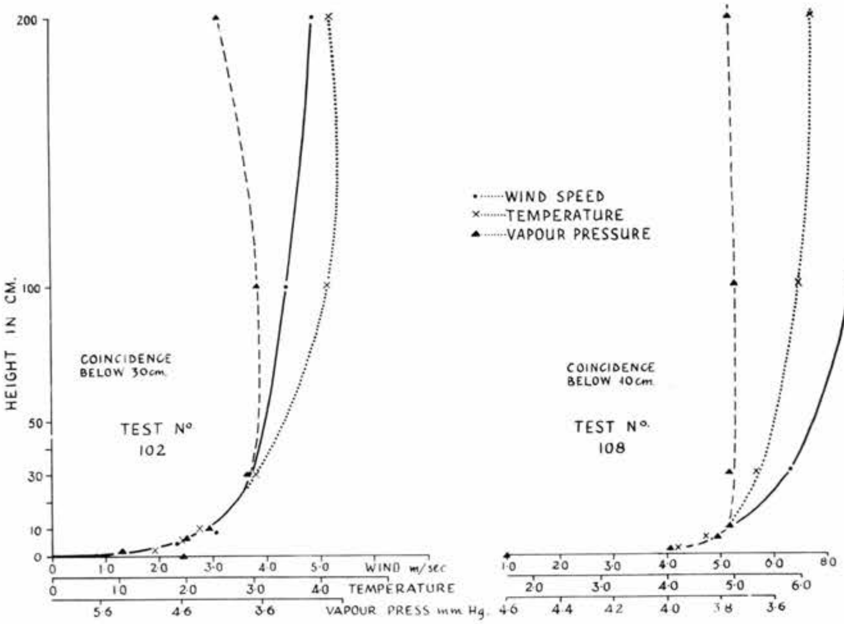


Figure 9. Superimposed profiles of Wind Speed, Temperature and Vapour Pressure.

ice surface. Temperature and vapour pressure readings were made at heights of 2, 6, 10, 30, 100 and 200 cm, with wind speed observations at all these heights except the lowest, at which it was impossible to use a cup anemometer.

During early June, 1954, when the lake ice had started to thaw, six tests of 30 minutes' duration were carried out on each of three separate days. The instruments were set up some 200 metres from the northern shore of the lake and observations were made on days when the wind was from the east or the west, giving a fetch of approximately 2 km over the surface being investigated. The lake ice surface was roughened where melt-water had begun to percolate between the vertical crystals; its appearance was similar to that of the melting glacier ice, but calculations of the roughness parameter (Table I) indicate that the lake surface was aerodynamically less rough than the glacier surface.

On two occasions, anemometers were either incorrectly set or misread and the tests concerned have been omitted from this analysis. Of the remaining sixteen tests there are fourteen which show a logarithmic distribution of wind and temperature up to 100 cm; nine of these have logarithmic profiles up to 2 metres, the maximum height considered (figure 8). Vapour pressure shows a marked change of slope above 30 cm, the change often involving a different sign. In nine of the sixteen cases considered, wind speed, temperature and vapour pressure profiles could, with an adjustment of the scales, be superimposed below 30 cm. Figure 9 shows an example of superposition, Test 102, and an example of divergence, Test 108. There is only one case where wind speed and temperature conform more closely to a power law variation with height than to a logarithmic variation, and here the vapour pressure curve suddenly changes its slope at only 10 cm above the surface.

It seems reasonable to conclude from the above that, in general, wind speed and temperature distribution with height follow the same law below 1 metre, vapour pressure distribution also agreeing below 30 cm.

THE RELATIONSHIP BETWEEN WIND SPEED, TEMPERATURE, VAPOUR PRESSURE AND STABILITY OVER A MELTING ICE SURFACE

Though the interrelation of wind speed and temperature profiles is probably quite complex it is readily appreciated that increase in wind speed will tend to destroy buoyance forces created by a decrease of temperature with height.

The usual parameter of atmospheric stability (Brunt 1939, p. 242) is the non dimensional Richardson number:

$$Ri = \frac{g\left(\frac{\partial T}{\partial z} + \Gamma\right)}{T\left(\frac{\partial u}{\partial z}\right)^2} \tag{8}$$

where

g = acceleration due to gravity

T = absolute temperature

Γ = dry adiabatic lapse rate ($\doteqdot +1 \times 10^{-4}$°C cm^{-1})

$\frac{\partial T}{\partial z}$ = vertical gradient of temperature

$\frac{\partial u}{\partial z}$ = vertical gradient of wind speed.

Thus Ri has the same sign as the temperature gradient, positive values denoting stable stratifications, negative values indicating unstable stratifications, and values near zero indicating neutral buoyancy in the layer considered. The temperature gradient $\frac{\partial T}{\partial z}$ over a melting glacier is generally so large compared with the dry adiabatic lapse rate Γ, that the latter can be ignored. BRUNT states (1939, p. 242), "The critical value of the RICHARDSON number above which turbulence tends to die out and below which turbulence sets in, is established at unity by RICHARDSON, at a half by PRANDTL and a quarter by TAYLOR and GOLDSTEIN".

The profiles of wind speed and temperature over a melting glacier have the most pronounced change of slope very near to the surface;

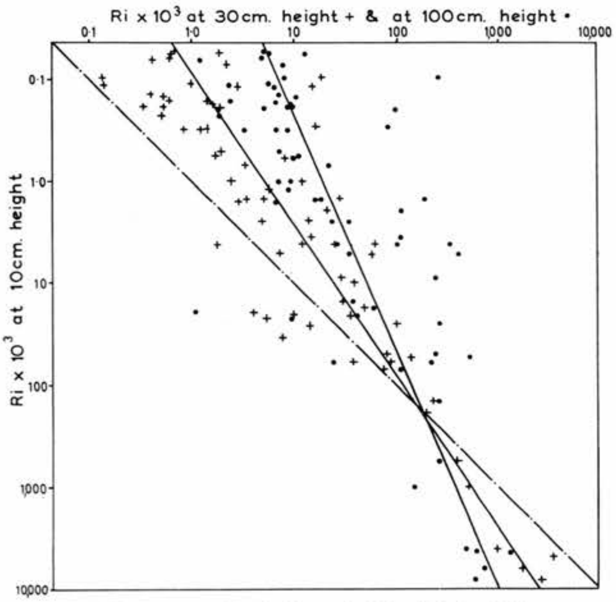


Figure 10. Variation of *Ri* with height.

in the present observations this was in the first 100 cm of height. Hence, to find the change with height of the stability of the atmosphere, the value of the RICHARDSON number was calculated at 10, 30 and 100 cm height. $Ri \times 10^3$ at 30 and 100 cm was plotted against that at 10 cm using logarithmic scales (figure 10). The diagonal of the graph is the line of RICHARDSON number constant with height. Almost all the plotted points lie to the right of the diagonal, that is, *Ri* at 30 and 100 cm tends to be higher than *Ri* at 10 cm. Though there is a fairly wide scatter, a line of best fit has been drawn of *Ri* at 30 cm height against *Ri* at 10 cm height and of *Ri* at 100 cm height against *Ri* at 10 cm height. It can be seen that for small values, *Ri* increases with height. As *Ri* at ten centimetres height increases, this increase of *Ri* with height becomes less marked. Above a critical value of $Ri = 0.2$, where the three lines on the graph can be seen to converge, there is a reversal of the change of *Ri* with height, i.e. there is a decrease of *Ri* over the range 10 cm to 100 cm This confirms what other workers have found, that where conditions favour an increase of stability in the lowest few metres, there is a very marked increase of stability in the lowest centimetres from the surface.

With only one exception, all the values of *Ri* calculated for the tests over melting ice are positive and therefore stable, and almost all the values of *Ri* are less than 0.2, therefore turbulence would seem to be established for the atmosphere here considered.

Statistical relation of observed values.

As a preliminary to the use of observations made at two heights through the ablation season the degree of correlation was found between various expressions of the characteristics of the micro-climate. Correlation coefficients were very low between any pair of the four variables: the difference of wind speed, air temperature, vapour pressure between 300 cm and 30 cm height and the net value of short wave radiation. It seemed possible that the low levels of significance (less than 5%) may have been due to interaction of the different meteorological factors, that is the dependence of one variable upon a third and perhaps a fourth variable. A partial correlation coefficient was calculated between each pair of dependent variables holding a third variable constant but this caused little change to the normal correlation coefficients.

The difference in wind speed over a constant height interval (e.g. $u_{300}-u_{30}$ where the subscripts refer to the height of observation) is a function of the scale of the wind as well as its mean gradient. A true indicator of gradient, reducing values to the same scale is $\frac{u_{300}-u_{30}}{u_{30}}$ Plotting these ratios of the mean gradient for both wind speed and temperature produced such a wide scatter of points that no correlation could be found. It would seem that over the range of observations here considered (300 cm and 30 cm height), there is very little relation between the differences or the mean gradient of temperature and wind speed.

Since most of the change of slope of the profiles occurs near the surface a correlation was sought between wind speed and temperature measured at different heights in this range. Table II lists these correlation coefficients, all of which have a significance level greater than 0.001.

To pursue further the interpendence of wind speed, temperature and vapour pressure, correlation coefficients between pairs of the variables was calculated for tests where the profiles could be superimposed. From a total of 34 such tests from the upper station, correlation coefficients were calculated for differences in wind speed, temperature and vapour pressure between 100 and 30 cm height. In spite of a high degree of coincidence in the profiles of these quantities, only temperature

Table II

Correlation Coefficients between Wind Speed and Temperature.

Height of Observations cm	Correlation Coefficient	Observations
300	+ 0.472	145
100	+ 0.589	155
30	+ 0.658	153
Height of Divergence of superimposed Wind and Temperature profiles	+ 0.586	134
ditto (upper station only)	+ 0.599	71

and wind speed gave a correlation coefficient with a high level of significance.

If full mixing of the air is indicated by the close similarity of the profiles of wind speed and temperature, then dissimilarity of these profiles would be expected to indicate less mixing. The goodness of fit of superimposed profiles was given a value according to the height from the surface to which they could be superimposed. The goodness of fit was then tabulated against the wind speed at one metre height and against the temperature at one metre height (Table III) and a correlation coefficient calculated for each.

Table III

Goodness of Fit of wind speed and temperature profiles at combined upper and lower stations (157 observations)

(a)

Temperature at 1 metre: °C						
5	4	2	1	16	5	4
4	1	3	6	16	6	4
3	3	5	6	14	2	6
2	1	0	6	7	2	0
1	0	2	6	8	5	1
0	6	2	2	2	1	2
Goodness of fit.	0	1	2	3	4	5

Goodness of fit.
$r = 0.188$

(b)

Wind speed at 1 metre: m. sec.$^{-1}$						
5	1	0	2	15	6	6
4	3	3	1	21	2	2
3	3	4	1	7	2	0
2	2	3	8	8	3	1
1	6	4	12	14	3	2
0	1	3	3	3	1	1
Goodness of fit.	0	1	2	3	4	5

Goodness of fit.
$r = 0.365$

The correlation of temperature with goodness of fit is low (but for the upper station taken separately $r = 0.432$), whereas that for wind speed is moderate, both with a high level of significance. Considering the goodness of fit with the lower half of the temperature range, the correlation is much improved. Similarly there is a better correlation of goodness of fit with the higher range of wind speeds (i.e. the upper half of the table.) Lower temperature or greater wind speeds (at one metre height) each give a lower stability and hence the greater possibility of full turbulence.

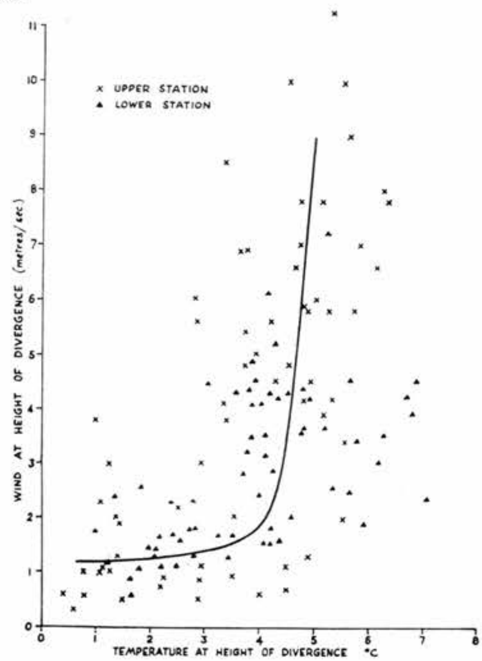


Figure 11. Wind Speed and Temperature at point of divergence of superimposed profiles.

If the profiles of temperature and wind speed may be superimposed only when the air is fully turbulent, then the point of divergence of the profiles may indicate the conditions of breakdown in turbulence. Values of wind speed and temperature at the point of divergence were plotted (figure 11) and a line of best fit drawn through the points which show a wide scatter. The values of wind speed and temperature, at the point of maximum change of slope of the curve, give a value of $Ri = 0.2$. This is the same value of Richardson number as the critical value indicated in figure 10. The correlation of wind speed and temperature, read at the height of the point of divergence of the superimposed profiles, gave a coefficient of 0.586 with a high level of significance. This is very near to the correlation coefficient of values of wind speed and temperature observed at one metre height (table II). The agreement between these coefficients indicates that superposition of the curves can generally be achieved to approximately one metre height.

Summary.

This statistical analysis of the interpendence of wind speed, temperature, vapour pressure and stability has shown that within the range of the observed values there is a high correlation between wind speed and temperature up to three metres height, but the correlation decreases with increasing height. The relationship between mean slopes of the profiles, denoted by differences in the values at two levels, is small. Profiles of wind speed and temperature variation with height can in the lower levels be fitted to a common profile, the height of divergence of the curves being approximately 1 metre. Turbulence, as indicated by the profiles of wind speed and temperature decreases at a value of $Ri = 0.2$. In very stable atmospheres conditions may be erratic and differ from the general relationships summarised here. The higher stabilities encountered at the lower station may be responsible for the less regular temperature and vapour pressure profiles observed there.

THE CALCULATION OF EVAPORATION AND HEAT TRANSFER

It is now possible to proceed with the evalution of the turbulent transfer of heat and water vapour in the lowest half metre of atmosphere using the eddy coefficient derived from the profiles of wind speed. Substituting for K (5) in the expression (1)

$$E = \frac{\varrho u_2 k^2 z}{\log_e \frac{z_2}{z_o}} \frac{\partial q}{\partial z} \tag{9}$$

which, when integrated between two levels z_1 and z_2 to avoid specifying the boundary condition at the surface, yields

$$E = \frac{\varrho u_2 k^2 (q_2 - q_1)}{\log_e \frac{z_2}{z_o} \log_e \frac{z_2}{z_1}} \tag{10}$$

which may also be expressed in the form derived by THORNTWAITE and HOLZMANN (1942, p. 21):

$$E = \frac{\varrho k^2 (q_1 - q_2)(u_2 - u_1)}{\left(\log_e \frac{z_2}{z_1}\right)^2}$$

Since the lowest wind speed reading on the glacier was at 30 cm, equation (10) has been used to calculate evaporation from values of vapour pressure at two heights and of wind speed at one height. Expressing (10) in terms of vapour pressure,

$$E = \frac{u_2 k^2 M (e_1 - e_2)}{R\,T \log_e \frac{z_2}{z_0} \log_e \frac{z_2}{z_1}} \tag{11}$$

where

e_1 & e_2 are vapour pressures at heights z_1 and z_2
M is the molecular weight of water
R is the universal gas constant
T is the temperature in degrees absolute

Equation (11) can be written (using a simplification by PASQUILL 1949b, p. 251) in the form

$$E = B\,u_2\,(e_2 - e_1) \tag{12}$$

where

$$B = \frac{k^2 M}{R\,T \log_e \frac{z_2}{z_0} \log_e \frac{z_2}{z_1}}$$

For any two heights z_1 and z_2 close to a melting ice surface B is effectively constant since the variation in absolute temperature with time in the layer is small. Equation (12) has therefore been used to calculate the evaporation (or condensation) during the three periods of intensive investigation. Separate values of the constant B (Table IX) were used for each period at each station. Results are shown graphically in figure 16.

Calculation of heat transfer by eddy conductivity.

An expression for the eddy heat flux can be developed along the same lines as that for evaporation. Ignoring Γ, the dry adiabatic lapse rate which is small compared with $\frac{\partial T}{\partial z}$ and integrating between levels z_1 and z_2 as before we have from equations (2) and (5)

$$H = -\varrho\,C_p \frac{u_2 k^2 (T_2 - T_1)}{\log_e \frac{z_2}{z_0} \log_e \frac{z_2}{z_1}} \tag{13}$$

where the symbols have the same meaning as before. Or

$$H = -C\, u_2\, (T_2 - T_1) \tag{14}$$

where

$$C = \frac{\varrho\, C_p\, k^2}{\log_e \dfrac{z_2}{z_o} \log_e \dfrac{z_2}{z_1}}$$

Within the layer considered, ϱ may be considered constant and the expression for C is a constant, (Table IX). Equation (14) has been used to calculate the turbulent heat transfer during the three periods of intensive investigation. Results are shown graphically in figure 16.

RADIATION, ALBEDO AND CLOUD

The albedo of a surface is the proportion of the incident radiant energy which the surface reflects. From the readings of incoming and reflected radiation measured by the MOLL and EPPLEY piles, at the upper and lower stations respectively, during the three intensive periods of investigation, this ratio was calculated. It was observed to vary considerably. For each of the three periods of investigation, incoming

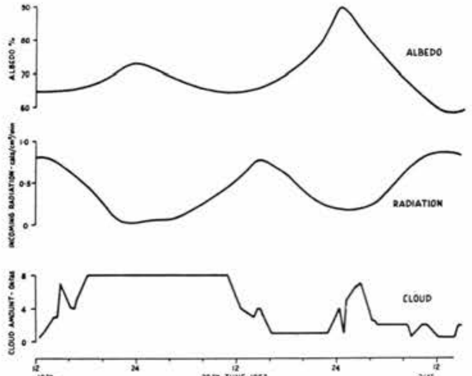


Figure 12. Radiation, Cloud and Albedo. – Upper Station, Early Summer.

radiation, albedo, and cloud amount were plotted on a time base (figure 12 shows the curve for the upper station during the early summer period) and an apparent correlation between albedo and incoming radiation was noticed, low values of albedo corresponding with high values of incident radiation. In some cases the peaks and hollows in the curve of albedo do not exactly correspond with the hollows and peaks in the radiation curve, but occur a few hours later. However, the general trend is for low values of albedo to correspond with high values of incoming radiation. Hubbley (1955, p. 561) found the same result with clear skies but with cloudy conditions there was no apparent relationship between radiation and albedo.

Incoming radiation and cloud.

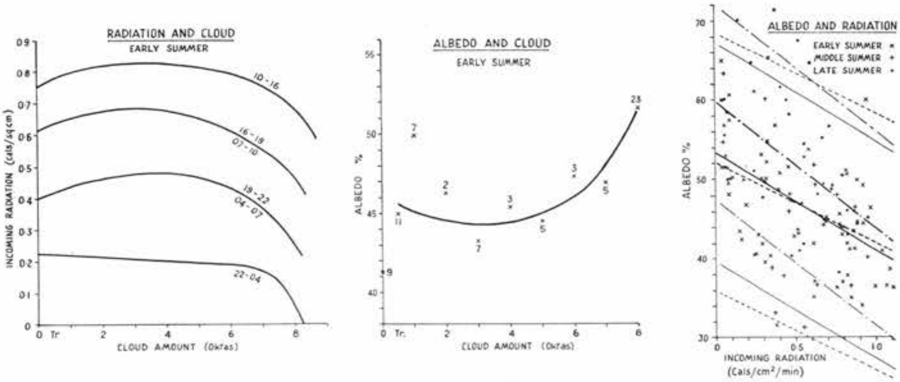
In order to find the extent to which incoming radiation was affected by the amount of cloud present, a plot was made of incoming radiation against cloud amount, measured in eighths of the sky covered by cloud. To allow for the marked diurnal variation in the incoming radiation, the readings were grouped into four six hour periods for this purpose. During the first investigation in early summer there was a good range of cloud cover, but in the middle summer and late summer investigations little cloud was present. However, the plot for early summer (figure 13a) indicates that there is little effect on the radiation if less than five eighths of the sky is obscured, but with a further increase in cloud there is a marked drop in the radiation reaching the surface. It would seem that, provided the sun itself is not obscured by cloud, a small amount of cloud actually increases slightly the radiation raching the surface, probably as a result of multiple reflections between the ice and the lower surface of the cloud. This result is quite commonly remarked in the literature (Wallén 1948, p. 492), though Wallén (p. 487) shows the relationship between radiation and cloud amount to be parabolic.

Albedo and Cloud.

Since it has been demonstrated that incoming radiation is slightly affected by cloud cover it seemed reasonable to suppose that the albedo of the surface would also to some extent depend on the cloud cover. The mean values of the albedo for each of the finite values of cloud cover were calculated and plotted. Figure 13b is typical of the early summer investigation. Weighting the points according to the number of observations in each group, it was possible to draw the curve shown, which indicates a slight decrease in the albedo as the cloud increases from zero to four eighths, and then a marked increase as the cloud increases to cover the whole sky. This diagram is directly comparable with that shown in figure 13a; an apparent correlation between albedo and incoming radiation has already been noted.

Albedo and incoming radiation.

To test the relationship between albedo and incoming radiation, a correlation coefficient was calculated and a regression line for albedo



a *b* *c*

Figure 13. a: Radiation and Cloud; b: Albedo and Cloud; c: Albedo and Radiation.

upon incoming radiation was drawn (as figure 13c) for each of the early, middle and late summer investigations. Correlation coefficients of -0.51, -0.28, and -0.50 respectively were obtained. The early summer investigation with 75 observations had a significance level greater than .001 but the later investigations with fewer observations were less significant. It can be seen from the regression lines that the slight changes in the nature of the glacier surface between each investigation cause less change in the value of the albedo than does the daily range of values of incoming radiation. Thus it would seem that the diurnal variation in albedo is more dominant than the slight variation throughout the summer, provided of course that the reflecting surface is ice in each case. If in early summer the ice is snow covered there will be a much higher albedo, though the daily range will be about the same.

A probable explanation of the diurnal variation of albedo lies in the variation of the intensity distribution of the solar spectrum with the altitude of the sun. At night the curve is flatter and its peak occurs at a longer wavelength, i.e. at night a smaller proportion of the total solar radiation occurs at the blue end of the spectrum than at midday. Since the albedo increases at night it appears that the surface of the ice reflects the longer wavelengths (at the red end of the spectrum) more readily than the shorter wavelengths (at the blue end).

The greatest variation in albedo is found with a change of surface from snow to ice. At the upper station (see Table I) the snow surface which gave a high albedo did not persist for more than a few weeks of the early summer but pockets of wet snow survived until late summer. Table IV shows a decrease in albedo on the upper glacier during the summer and shows a decrease in albedo as we go down the glacier. Both these changes are due to the change of surface from melting snow to melting ice.

It may be concluded from these observations of radiation, albedo and cloud that the variation of albedo with cloud amount and with time (greater than 24 hours) during the summer, is quite small. Over the same surface the greatest variation in albedo is a diurnal variation, albedo varying inversely with incoming radiation. Table IV shows the mean values of albedo for day (09.00–21.00 GMT) and night (21.00–09.00 GMT) measured during three 48-hour periods in early, middle and late summer respectively. During all periods save the early summer on the upper glacier, the surface was melting ice with some very wet snow in the hollows.

Mean values of Albedo.

For calculation of the heat available at the surface during the ablation period as a whole, a mean albedo has been calculated from these values taken over very similar surfaces. Since there is only 8 per cent difference between these means of day and night albedo, and since the standard deviation is 6.6 and 9.6 per cent respectively, it was considered permissible to use one value of albedo for both day and night. This is particularly so since the times of observations of other meteorological factors and of ablation did not perfectly coincide with day and night periods. Hence for purposes of the calculation of heat available at the glacier surface, a value of 50 per cent has been used. This value is the same as that used by SVERDRUP (1935, p. 157) but is greater than the mean of 40 per cent found by WALLÉN (1948, p. 496).

Table IV

Mean Values of Albedo (percentage) over a Melting Glacier Surface.

Period	Upper station		Lower station	
	Albedo	S. D.	Albedo	S. D.
Early summer day	66	8	45	8
night	74	9	51	9
Middle summer day	48	5	47	7
night	66	6	46	8
Late summer day	48	3	49	8
night	52	5	57	5
Summer mean for both stations day	47.4	6.6		
Summer mean for both stations night	55.5	9.6		

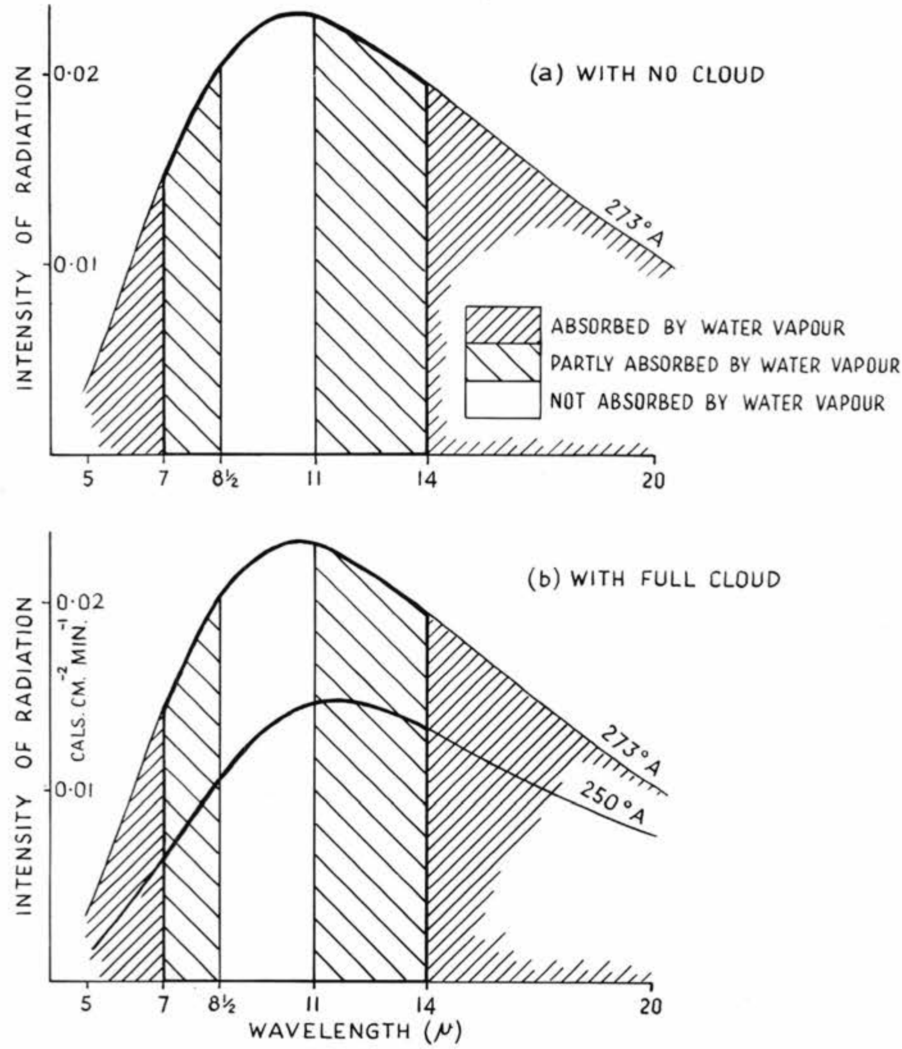


Figure 14. Outgoing, long-wave Radiation. After Simpson.

Outgoing radiation.

Long wave radiation was not measured directly and the intended use of the Kew Chart (Robinson 1947, p. 136) was considered inadvisable because the nearest upper air data was from Danmarkshavn, 150 km distant and in a very different geographical environment. Brunt (1939, p. 108) discusses adaptations of Stefan's Law, for evaluating outgoing radiation, some consideration being given to absorption of certain wavelengths by carbon dioxide and water vapour:—

1. Complete absorption $5^1/_2$ to $7\,\mu$ and greater than $14\,\mu$

2. Complete transparency less than $4\,\mu$ and between $8^1/_2$ and $11\,\mu$.

3. Incomplete absorption from 7 to $8^1/_2\,\mu$ and from 11 to $14\,\mu$. On the basis of Simpson's extension of Hettner's analysis (BRUNT 1939, p. 116 and p. 156), if there is no cloud, it is only radiation in the range $8^1/_2$ to $11\,\mu$, and part of that in the ranges 7 to $8^1/_2\,\mu$ and 11 to $14\,\mu$, which can escape from the surface. Using this approach an estimate has been made of the outgoing long wave radiation from a melting ice surface into clear skies.

Ordinates of the spectrum radiated by a black body at 273°A were calculated from Planck's Law and plotted. The areas between 7 to $8^1/_2\mu$, $8^1/_2$ to 11μ, and 11 to 14μ were measured. The outgoing radiation was taken to be equal to the whole of the area under the curve between $8^1/_2$ and 11μ plus half the areas between 7 and $8^1/_2\mu$ and 11 and 14μ and was found to be 0.102 cals./cm^2/min. (figure 14).

BRUNT (1939, p. 122) shows that cloud, about 50 metres thick, behaves as a black body. Thus with full cloud, the net outgoing radiation is the difference between that portion of the earth's black body radiation not absorbed by water vapour and the corresponding range of wavelengths radiated by the cloud at the temperature of its lower surface. Making a broad assumption that the temperature of the cloud lies midway between the surface temperature of 273°A and the summer stratosphere temperature at latitude 77° N., of 227° A (Brunt 1939, p. 18) the net outgoing radiation for full cloud is .033 cals./cm^2/min. Since the height and temperature of cloud may vary considerably, the figure 0.033 can only be considered as an approximation, though with high alto- and cirrus clouds such as were most commonly observed, it is possibly representative.

Expressions relating outgoing radiation to surface temperature, vapour pressure and cloud amount are given by ÅNGSTRÖM (WALLÉN 1948, p. 497), by BRUNT (1932, p. 402, and 1939, p. 137) and by OLSSON (1936, p. 105). Table V summarises the values derived from these authorities:

It would seem that the estimates of the outgoing radiation based on SIMPSON's analysis are too low, particularly those for cloudless conditions. SUTTON (1953, p. 171) states that in SIMPSON's analysis no use is made of HETTNER's absorption coefficients, the only necessity being the accurate specification of the bands in which water vapour is transparent to radiation. He continues that this circumstance is particularly important in view of the fact that later work has shown the interval of transparency to be greater than SIMPSON estimated. Thus there is reason to believe that the estimates made here are too low,

Table V

Outgoing Radiation in cals. cm^{-2} $min.^{-1}$.

Derived from expression by	Cloud Amount	Outgoing Radiation
Simpson	NIL	0.102
Ångstrøm	,,	.142
Sverdrup	,,	.160
Brunt	,,	.164
mean value	,,	.142
Simpson	FULL	0.033
Ångstrøm	,,	.014
Sverdrup	,,	.040
Olsson	,,	.080 app.
mean value	,,	.042

so the mean of the values quoted in Table V has been used to derive an expression for outgoing radiation similar to that by Ångström (Brunt 1939, p. 144):—

$$R_m = 0.14\ (1—.089\ \mathrm{m})$$

where m is the cloud amount in oktas. This formula has been used with the values of cloud amount observed to find the outgoing radiation from the melting ice surface. The outgoing radiation was subtracted from the net incoming radiation to give the net radiation flux available at the ice surface.

CONDUCTION OF HEAT INTO THE GLACIER ICE

During the three intensive periods of investigation, ice temperatures were taken at two-hourly intervals at both stations. From figure 15, which shows temperature plotted against time for the various depths, it is evident that there is a diurnal variation of temperature to approximately 200 cm below the surface. This variation is superimposed on a slight increase of temperature with time. To assess the amount of heat passing downwards from the surface, the problem has been treated as an elementary conduction problem, the conductivity k of ice being taken as 5×10^{-3} cal. cm^{-1} $sec.^{-1°}$ $C.^{-1}$ (Kaye and Laby 1948, p. 68). To obtain a representative gradient of temperature during the early, middle and late summer, the mean temperature has been taken from the graphs of temperature at depth against time. The resulting mean temperature profiles are also plotted on figure 15. It will be seen that the top 10 to

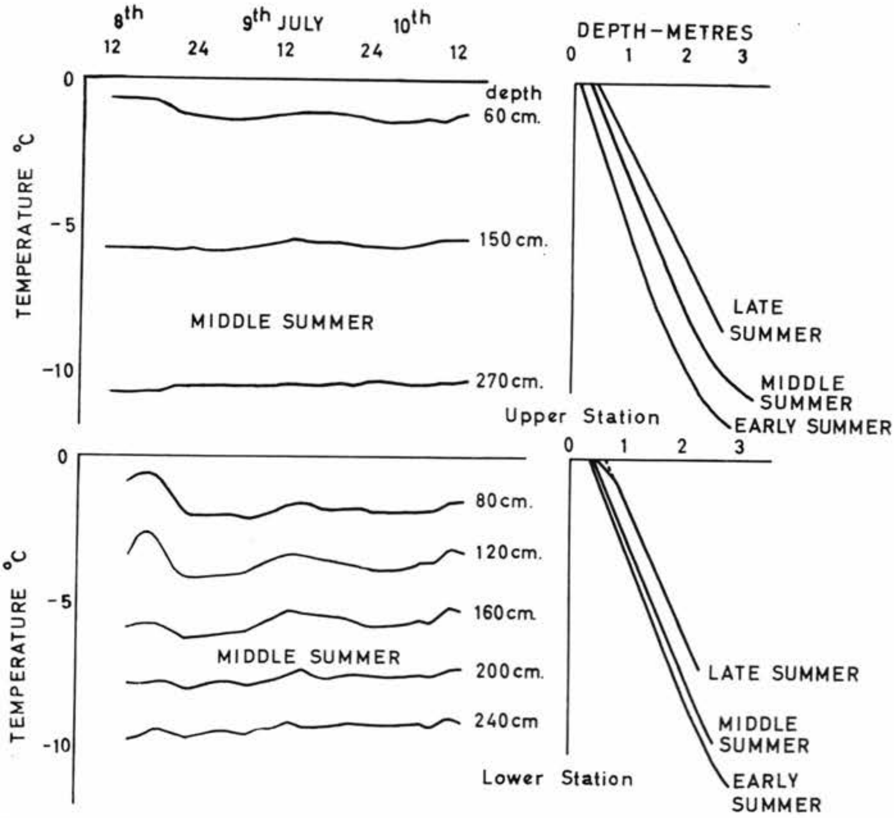


Figure 15. Ice temperatures on Britannia Gletscher, Summer 1953.

60 centimetres of the ice are at 0° C, whilst below that level the temperature decreases almost linearly with increasing depth for about 2 metres. This surface layer is kept at melting point by absorption of radiation, by the percolation of melt water, and by the presence of dust wells, melt water channels and small, water-filled crevasses. The mean thickness of the layer which is at 0° C increases as the summer progresses though it would appear to have a small diurnal variation. From the temperature profiles the heat flux has been calculated and is listed in table VI.

The quantity of heat conducted away from the surface can be seen to be small. There is a small decrease in the downward flux of heat into the ice as the summer progresses, but as the heat involved is small compared with that available by radiation and convection, a mean value can be used for the whole summer. To assess the effect of the diurnal change on this mean value, temperature profiles were plotted for all observation times and the gradients measured. The maximum

Table VI
Conduction of Heat into the Glacier.

Station	Period	Temperature Gradient	Heat Flux		Mean depth at 0°C
		°C. cm^{-1}	Cals. cm^{-2} min.$^{-1}$	Cals. cm^{-2} hr.$^{-1}$	cm
Upper	Early	5.8×10^{-2}	.0174	1.04	10
,, ...	Mid	4.8	.0144	0.86	30
,, ...	Late	3.9	.0117	0.70	40
,, ...	mean	—	—	0.90	—
Lower	Early	5.2	.0156	0.94	35
,, ...	Mid	4.8	.0144	0.86	40
,, ...	Late	4.5	.0135	0.81	60
,, ...	mean	—	—	0.90	—

change from the mean value would seem to be of the order of 10 to 15 per cent with the increased value tending to prevail during the day and the reduced value at night. However, since the values concerned are so small, the mean value for the flux of heat into the ice has been used.

The difficulty of obtaining accurate temperature readings near the surface has been mentioned earlier, and in general no reliable readings were obtained at depths less than about 40 cm. Thus it is difficult to assess the thickness of the isothermal layer (0° C) with any great degree of accuracy, and the small diurnal variation of thickness (of the order of 10 cm) can only be very approximately estimated from extrapolation of the figures available. This variation implies that during the day the surface layer acts as a reservoir for some of the heat available at the surface, and at night this heat is dissipated in maintaining the temperature gradient below. If this is the case it would seem that the mean value of 0.9 cals. cm^{-2} hr.$^{-1}$ for the heat flux downward from the surface could be in error by at least 50 per cent at the hottest and coldest parts of the day.

HEAT BALANCE AT THE GLACIER SURFACE

The different factors in the heat balance equation were evaluated for each alternate hour of the three intensive periods of observation as described in the preceding pages and the calculated source and sink of the heat at the glacier surface were compared. In accordance with the principle of conservation of energy, the source and sink of heat at the glacier surface must be equal, hence the respective blocks are equal,

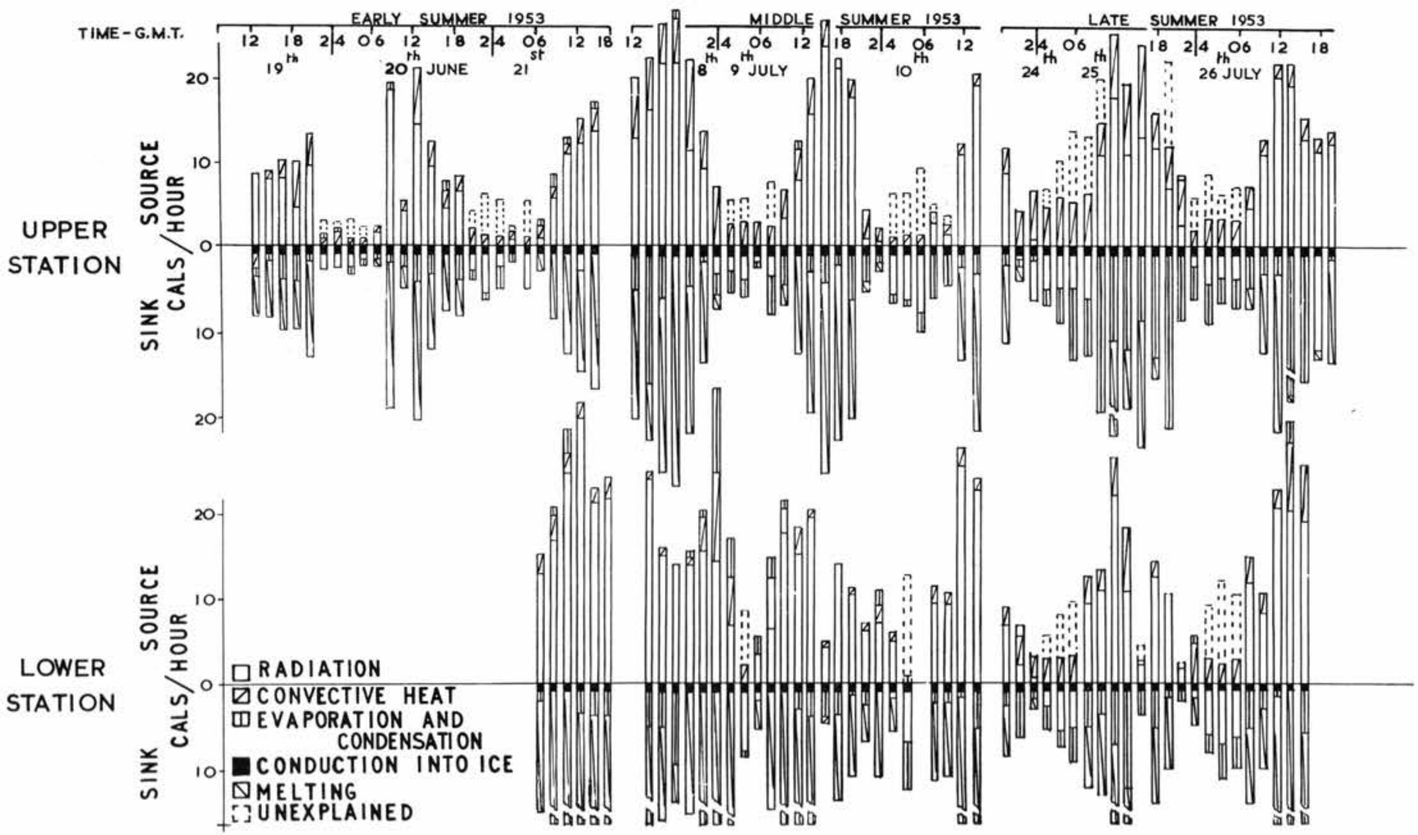


Fig. 16. Heat Balance on Britannia Gletscher.

for each hour shown in figure 16. The calculated values of the source and sink of heat for each hour do not always balance, the latter being the greater. In these cases, the total heat lost in outgoing radiation in evaporation and in conduction into the ice, is greater than the heat gained by convective transfer due to eddying air. The mean values (taken over 48 hours) of the factors governing the supply and use of heat at the ice surface will not be greatly in error since errors during day and night hours have opposite sign. Table VII and Table VIII summarise the sources of the heat and the use made of that heat respectively.

Table VII

Percentage of Heat due to Radiation, Convection, Condensation.

Summer Period	Upper Station			Lower Station		
	Radiation	Convection	Condensation	Radiation	Convection	Condensation
Early.........	74.8	23.3	1.9	82.0	12.6	5.4
Mid	69.1	30.3	0.6	78.3	14.8	6.9
Late..........	57.7	42.3	0	66.5	31.9	1.6
mean	67	32	1	75	20	5
mean .... (both stations)	69	29	2			

Table VIII

Percentage of available Heat used in Melting, Evaporation, and Conduction into Glacier.

Summer Period	Upper Station			Lower Station		
	Melting	Evaporation	Conduction	Melting	Evaporation	Conduction
Early.........	61.6	17.8	20.6	—	—	—
Mid	54.0	34.7	11.3	78.3	13.1	8.6
Late..........	20.2	71.1	8.7	55.6	34.0	10.4
Mean	45	41	14			
Mean .... (both stations)	54	34	12			

Radiation is the most important factor in the supply of heat to the glacier surface during the ablation season. Radiation decreases in importance as the summer progresses and also from the lower tongue to the head of the glacier. The cause of the decrease in radiation as the summer progresses is the decrease in the sun's elevation and the cause of the decrease in radiation with increased altitude is the increase in the albedo higher up the glacier. This is due to increased roughness and pockets of wet snow persisting. With overcast conditions the importance of radiation decreases rapidly but clear skies and low values of cloud amount were very common during the summer of the observations. During the night hours the outgoing radiation is greater than the incoming radiation.

The supply of convective heat increases as the summer progresses and increases with altitude. The meteorological conditions governing the turbulent heat transfer are wind speed and temperature. Both wind speed and air temperature increased from the early summer to the late summer investigation but the increase in wind speed was the most marked, particularly at the upper station. Hence wind speed would seem to be the more important factor in supplying convective heat, though the total heat supplied by this means is less than half the value of that supplied by radiation.

Condensation is small in the early summer and later decreases to a negligible quantity. Melting is the dominant factor in the summer ablation process, using approximately half of the total heat available. Melting decreases in importance as the wind speed and air temperature increase and cause more evaporation. Though more than a third of the total heat available is used in evaporation, the amount of ice removed by this means is small, since evaporation requires eight times the heat required for melting.

The heat lost by conduction into the glacier decreases from one fifth in the early summer to less than one tenth in the late summer. Before recordings of net ablation were begun in the early summer, a larger proportion of the total heat then available must have been conducted down from the glacier surface. A constant heat flux into the ice has been used in the calculations of the surface heat balance. Hence the variation in the proportion of the available heat conducted into the ice is governed by changes in the total heat available at the surface.

Figure 17 shows ablation as measured and ablation calculated from the heat available at the surface for the middle summer period. For the early and the late summer periods the agreement is poor. The hours near local midnight (nearly 2 hours behind the G.M.T. shown) show a negative calculated ablation, which is, of course, impossible. The diurnal change from plus to minus in the differences between calculated and observed ablation suggest that the principal error has a diurnal variation. Radiation alone shows such a marked diurnal variation and since the values of outgoing radiation are estimated, this may well be the principal source of error. The assumption of a constant heat flux into the ice

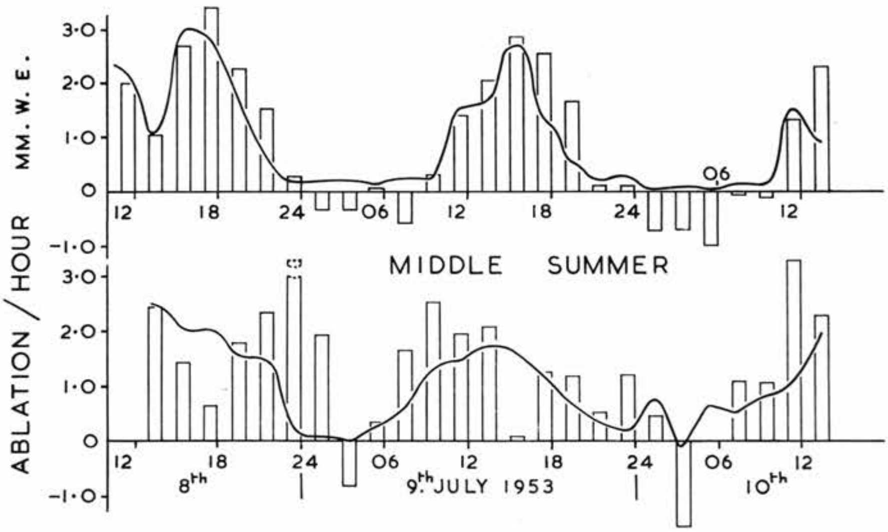


Figure 17. Measured and Calculated Ablation at Upper and Lower Stations on Britannia Gletscher — Middle Summer. Measured Ablation shown by continous line. Calculated ablation shown by blocks.

may also be a contributory factor, though in this case the quantities are small.

Some of the worst discrepancies occur for hours when there is poor agreement between profiles of wind speed, temperature and vapour pressure and the assumption of common coefficients in these cases was probably unjustified. Wrong estimates of evaporation made from such assumptions could result in a large error in the calculated ablation since the heat required for evaporation is much greater than that required for melting.

CONTINUOUS OBSERVATIONS

The evaluation of heat transfer from profiles measured in the four metre air layer bounded by the glacier surface was intended to establish laws which could be used with data recorded at only two heights throughout the summer. The observations made throughout the summer may be summarised:—

Ablation was measured relative to datum marks on each of six stakes drilled three metres deep into the glacier surface. The mean surface was defined at each observation by laying a strip of wood on the north side of the stake and measuring down from the datum to the top of the wood. Generally the ablation was so small that readings made more than twice daily showed too large a probable error, but on occasions when the rate of ablation was unusually high three or four readings were taken each day.

Incoming short wave radiant energy was recorded by a Robitsch Actinograph. Its readings, on an arbitrary scale, were checked from time to time against spot readings on a Moll or Eppley pile.

Air temperature and relative humidity were recorded by thermohygrographs housed in screens 30 and 300 cm above the ice surface. Three or four times daily, check readings were made at both heights with a whirling psychrometer. It was difficult with this instrument to obtain reliable wet bulb depressions when the air temperature was near 0° C despite whirling for from 10 to 15 minutes and removing any solid ice which may have formed on the wet bulb. Direct or reflected radiation falling on to the psychrometer thermometers was another likely source of error; this was overcome to some extent by fitting metal disc radiation screens. From time to time both thermohygrographs were run at the same level as an additional check on their behaviour.

Cup counter anemometers recorded the run of wind at 30 and 300 cm and were read three or four times daily at the same time as the check readings of temperature and humidity.

Figure 18 summarises the ablation and meteorological conditions recorded throughout the ablation season. The gaps in the records during the earlier part of the season are due to instrumental troubles; those at the end of the season are a result of all members of the expedition having to return to the main base for an air supply operation.

The graph of mean incoming radiation shows, as anticipated, the alternation of day and night values and the overall decline after the summer solstice. The average wind speed at 300 cm above the surface is approximately 4 metres per second; it rarely falls below 2 metres per second or rises above 9 metres per second. The average temperature at 300 cm height is approximately 4° C; the temperature seldom rose above 7° C or fell below 0° C. Since wind speed and temperature are both zero at the surface, their gradients could be expected to increase as the values increase. Such an effect would be evident from the increased separation between higher values read at two heights. Figure 18 demonstrates the effect with wind speed, but less so with temperature, since above 30 cm height temperature increases less rapidly than does wind speed.

The graphs of the values of vapour pressure at the two heights cross and recross each other. There is a time lag in the response of hair hygrometers which increases as the temperature falls, becoming effectively infinite at about −40° C (Penman 1955, p. 50). It is thought that the interval over which results have been measured is adequate to eliminate error due to time lag and one must conclude that the height

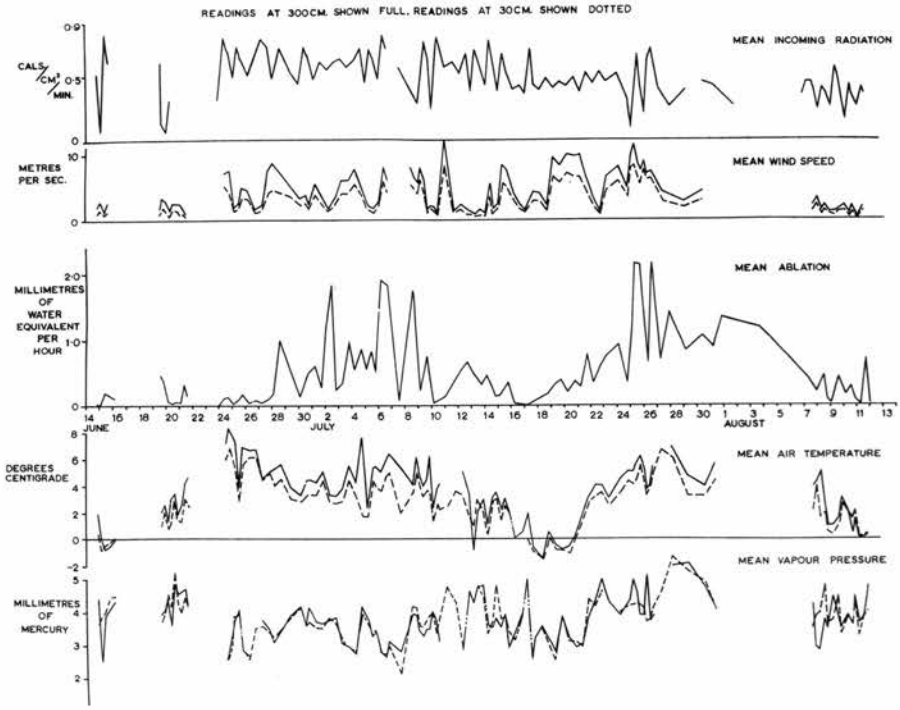


Figure 18. Ablation and Weather, Britannia Gletscher, Summer 1953.

separation of the two instruments was inadequate to give an accurate indication of humidity gradients.

The average rate of ablation over the whole summer was 0.4 mm (water equivalent) per hour and the maximum, late in July, was 2.0 mm. There are two periods of high ablation, centred about 6th July and 29th July with a period of zero ablation in between. The graph of temperature shows a similar 'double summer' effect. In fact, the curves of ablation and temperature show more similarity than any of the other elements observed. On the Frejagletscher in Young Sund Ahlmann (1946, p. 243) found a perfect fit of these two graphs. His observations (at 750 m) also showed that the greatest ablation occurred late in July with a rapid fall off in August, and in this respect, too, it is similar to the Britannia Gletscher.

The detailed observations, made each alternate hour (e.g. figure 4) showed the closest similarity between curves of ablation and radiation, a diurnal effect being evident in both. Over the whole summer the continuous observations show closest similarity between air temperature and ablation, though it should be noted that peak values of ablation usually occur on days with peak values of radiation.

The Evaluation of Data from Instruments in Screens.

The recorded values of wind speed, temperature and vapour pressure at 300 cm height were not used to calculate heat transfer because:

(1) The variations of these values with height were found from the detailed investigations to obey different laws above 1 metre height.

(2) The values of vapour pressure calculated from screen records at 30 cm and 300 cm height appeared to be too inaccurate to permit the difference of these values to be used as a vapour pressure gradient.

Values at 30 cm were used, assuming the same laws to apply as had been observed to fit the detailed observations. In order to solve the equations for observations at one height only it is necessary to specify boundary conditions. In the case of wind speed the boundary condition is specified that $u = 0$ at $z = z_o$. As an approximation, it was assumed that the surface conditions of temperature and vapour pressure should apply at the height z_o. Since vapour pressure profiles often exhibited some degree of supersaturation of the air near to the surface, the assumption that $e = e_o$ (= 4.58 mm Hg) may not be much in error, though the assumption that the air temperature $T = T_o$ (= 0° C) at $z = z_o$ is unlikely to be true.

The modified equation for evaporation (from (11),) is

$$E' = \frac{u_2 k^2 M\,(e_o - e_2)}{RT\left(\log_e \dfrac{z_2}{z_o}\right)^2} \tag{15}$$

or

$$E' = B'\,u_2\,(e_o - e_2) \tag{16}$$

where

$$B' = \frac{k^2 M}{RT\left(\log_e \dfrac{z_2}{z_o}\right)^2}$$

Similarly, (from 13) the equation for the convective transfer of heat may be written,

$$H' = \frac{-\varrho\, C_p u_2 k^2\,(T_2 - T_o)}{\left(\log_e \dfrac{z_2}{z_o}\right)^2} \tag{17}$$

or

$$H' = -C'\,u_2\,(T_2 - T_o) \tag{18}$$

where

$$C' = \frac{\varrho\, C_p k^2}{\left(\log_e \dfrac{z_2}{z_o}\right)^2}$$

Table IX shows the difference in the constants used for calculating the vertical eddy flux in an air layer 30 cm to 10 cm height and 30 cm to the boundary of the surface.

Table IX

Constants for Calculating Evaporation and Turbulent Heat.

$M = 18$ $k = 0.4$
$R = 83.15 \times 10^6$ $z_2 = 30$ cm
$Cp = 0.2396$ $z_1 = 10$ cm

Summer period	Glacier station	Mean T °A	Mean ϱ gm/cm³	z_0 cm	B	B'	C	C'
			$\times 10^{-4}$		$\times 10^{-4}$	$\times 10^{-4}$	$\times 10^{-2}$	$\times 10^{-2}$
Early	Upper	274	11.97	1.10	16.71	5.55	4.548	1.512
	Lower	275	12.16	0.4	12.74	3.24	3.539	0.900
Middle	Upper	275	11.82	0.68	14.53	4.22	3.921	1.117
	Lower	276	12.04	0.58	13.89	3.87	3.832	1.067
Late	Upper	276	11.83	0.58	13.89	3.87	3.765	1.049
	Lower	276	12.05	0.57	13.83	3.83	3.820	1.058
Mean summer	Upper	275	11.77	0.65	—	4.12	—	1.106

Using the constants from Table IX the ratio of the correct to the approximate solutions of the eddy transfer expression was calculated (Table X and figure 19).

There is a decrease in both ratios through the summer. The value near to unity in the middle summer for the ratio of the evaporation expressions suggests that the approximate method is nearly correct for this period, but the extreme value at the upper station for the late summer seems to be in error since it shows a departure from the general trend. The approximate method of calculating the heat transfer seems to yield values which are too big. This may be readily understood when it is remembered that the assumption made, of a surface value being applicable at a height of $z = z_0$ has the effect of steepening the gradient of the profile and indicating a greater transfer of energy.

For the whole summer period, the convective transfer of water vapour and of heat were calculated from the records at 30 cm, using equations (16) and (18), and multiplied by the ratios E/E' and H/H' from the graphs in figure 19.

The flux of heat into the glacier ice has been taken as 0.9 cals. $cm^{-2}\,hr^{-1}$, found from the detailed investigations.

Outgoing radiation has been calculated according to the amount of cloud present in the same way as for the detailed observations. The net incoming radiation has been calculated from the records of the Robitsch Actinograph, using the value of 50 % for the albedo.

Table X

Comparison of Transfer of Heat and Water Vapour Calculated from Profiles between $z = 30$ cm, and $z = 10$ cm and between $z = 30$ cm and $z = z_0$ cm.

Summer Period	Upper Station		Lower Station		Upper Station		Lower Station	
	$\frac{H}{H'}$	Std. Dev.	$\frac{H}{H'}$	Std. Dev.	$\frac{E}{E'}$	Std. Dev.	$\frac{E}{E'}$	Std. Dev.
Early....	0.46	0.05	0.65	0.14	1.44	0.39		
Middle...	0.35	0.08	0.56	0.28	0.91	0.32	1.08	0.31
Late.....	0.19	0.03	0.23	0.05	2.03	0.75	0.84	0.22

H = Flux of heat calculated from equation (14) using temperature values observed at 30 cm and 10 cm height. H' = Flux of heat calculated from equation (17) using temperature values observed at 30 cm height and assuming the temperature = 0°C at height $z = z_0$ cm. E = Flux of water vapour calculated from equation (12) using vapour pressure values observed at 30 cm and 10 cm height. E' = Flux of water vapour calculated from equation (15) using vapour pressure values observed at 30 cm and assuming the vapour pressure = 4.58 mm Hg at height $z = z_0$.

Table XI shows the source and sink of heat at the glacier surface, taken over the whole ablation season.

Table XI

Source and Sink of Heat at Glacier Surface, during Ablation Season 1953.

Percentage of total Heat			
Source		Sink	
Radiation...............	71	Melting.................	23
Convection..............	27	Evaporation.............	69
Condensation............	2	Conduction..............	8

The sources of heat agree well with those found during the detailed observations (Table VII) but the use of the heat (Table VIII) shows less agreement. There is even greater disagreement between the observed and calculated ablation; these can differ by a factor of four. Hence it would seem that calculations of the heat balance at a glacier surface made from observations in a screen at one height can only be very approximate and may be grossly in error, particularly if the observations are made by recording instruments as currently in use.

Previous workers have found some agreement between measured and calculated ablation, generally using a power law with observations of temperature and humidity at up to three heights in a 2 metre air

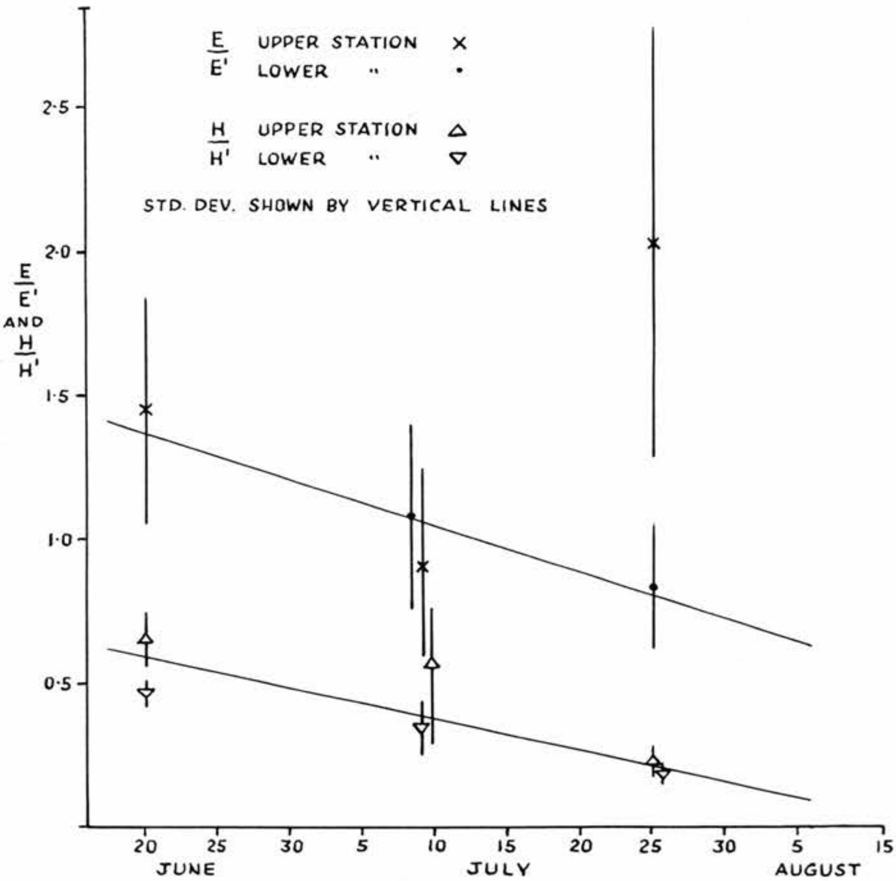


Figure 19. Ratio of Heat and Vapour Transfer calculated from Profiles and from Screen and Surface Values.

layer. In some cases, albeit with caution, values observed in meteorological screens have been used with values of a melting ice surface to calculate heat and vapour transfer, assuming a power law relation of the variables with height. It has been shown that the logarithmic law is generally limited to the lowest 1 metre of the air layer investigated here, and a linear relation is common above 2 metres. Hence the power law is a better approximation than the logarithmic law over a height range of 2 metres. Considering surface values with observations above 50 cm height, the approximate gradients so indicated are less steep than has been shown here by considering observations near the surface. Using such approximate gradients the values of the calculated ablation will be less than those evaluated here, and apparently have been very near to the observed values of ablation.

In this investigation it was found that the profiles of the variables were not always similar above 50 cm height and below this height a logarithmic relation most closely fitted the observed values.

THE METEOROLOGICAL ELEMENTS RESPONSIBLE FOR ABLATION

For comparison of the importance of various factors in the ablation process Table XII lists the data similarly compiled by Wallén (1948, p. 633), Ahlmann (1953, p. 6), Orvig (1954, p. 298) and Adkins (1958, p. 204). The approximate distance of each glacier from the open sea and the results from Dronning Louise Land have been added. Direct comparison of the data must be limited since the period of observation at each glacier varies from two weeks to the whole summer; only on the Kårsa Glacier were observations made for more than one ablation season.

The available data for the two glaciers in North East Greenland are extremely different. The Britannia Gletscher is an outlet glacier of the inland ice cut off from the sea air by glaciers and coastal mountains. The Frejagletscher is from an ice cap on a small island. The fit of the graph of ablation with that of radiation for the former and of temperature for the latter (Ahlmann 1946, p. 243) is in accordance with the percentages of the sources of heat shown in the table. The dates of onset, peak and rapid decline of ablation on the Frejagletscher are almost identical with those observed on the Britannia Gletscher. From these characteristics of the beginning and progress of the ablation, Ahlmann points out the extreme continentality of N. E. Greenland.

From a comparison of meteorological records Hamilton (1958, p. 158) concludes that the summer weather of 1953 was fairly typical. Over the Britannia Gletscher radiation was dominant in causing ablation because of the frequent clear skies and, save in early summer, the absence of a snow cover to reflect the incoming radiation. It was often possible to see clouds rolling in from over the sea, as far as the Storstrømmenglacier and sometimes as far as Britannia Sø. It was unusual to see cloud from the sea reach beyond the lower Britannia Gletscher.

The general trend revealed in the table is an increase in radiation with altitude and with distance from the sea; conversely there is a decrease in convection. These trends may be expected but are modified by the amount of snow cover and by the amount of warming of the air over snow free land. The changes of radiation and convection with

Table XII
Convection, Condensation, and Radiation Factors in the Ablation Process.

Glacier	Year	Position	Elevn. metres	Dist. from ocean km	Conv. %	Con-dens. %	Rad. %	Sur-face
Sveanor snowfield....	1931	79°56′ N 18°18′ E	5	6	58	18	24	snow
Isachsens Plateau	1934	79°09′ N 12°56′ E	870	30	29	15	56	snow
Fourteenth of July Glacier	1934	79°08′ N 12° E	600	8	53*	—	47	snow (ice later)
Britannia Gletscher ...	1953	77°14′ N 23°49′ W	620	150	32	1	67	ice (snow earlier)
Britannia Gletscher	1953	77°12′ N 23°48′ W	470	150	20	5	75	ice
Frejagletscher .	1939	74°24′ N 20°50′ W	453	10	83	9	8	snow
Barnes Ice Cap.....	1950	70° N 72° W	865	150	32*	—	68	snow (ice later)
Penny Ice Cap	1953	66°59′ N 65°28′ W	2050	90	9	30	61	snow
Kårsa Glacier .	1942–8	68°20′ N 18°20′ E	> 1100	80	44	24	32	snow
Kårsa Glacier .	1942–8	68°20′ N 18°20′ E	< 1100	80	29	16	55	ice
Hoffellsjökull ..	1936	64°30′ N 15°30′ W	1000	20	92*	—	8	ice
Salmon Glacier	1957	56°10′ N 130°07′ W	1700	320	15	10	75	snow
Vernagtferner .	1950	46°50′ N 10°45′ E	3000	240	15	4	81	ice

* includes condensation.

latitude are more erratic; topography and proximity of maritime air, are apparently more important than latitude in governing the percentage importance of the meteorological factors causing ablation.

Conclusion.

The calculation of heat balance at the surface of a glacier requires observations over the area of the glacier (similar to the procedure for evaluating mass balance) but a reasonable approximate evaluation can be made from one suitably sited station. For this purpose data recorded by instruments as currently used in meteorological screens are inadequate, particularly in the measurement of humidity. Wind speed, temperature and humidity observations made at heights up to 4 metres can generally be closely approximated by a power law. A logarithmic variation of these elements with height prevails below 1 metre even at comparatively high stabilities; at heights greater than 1—2 metres temperature gradients are frequently linear. Hence profiles are generally dissimilar above 1 metre height and the usual assumption of a common coefficient for the eddy transfer of momentum, heat and water vapour appears to be limited to a layer less than 1 metre height. In the highly stable conditions which are characteristic near to a snow or ice surface there can be marked irregularities of wind speed, temperature and humidity gradients in the lowest few centimetres. An understanding of these phenomena seems to require further work with instruments specially designed for use in polar regions and the development of an inclusive theory for micro meteorology over cold surfaces.

Over the Britannia Gletscher the total amount of heat at the surface is contributed by radiation 69%, condensation 2% and eddy convection 29%. The manner in which this total heat is used, is melting 54%, evaporation 34% and conduction into the ice 12%.

Over the glaciers studied in the Northern hemisphere, the importance of radiation in the total heat supply increases with altitude and distance from maritime air. Convection decreases similarly but generally increases with latitude. However, the local topography and air mass prevailing over a glacier can modify the heat balance interpolated from these broad areal parameters.

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