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SCIENTIFIC RESULTS NO. 6

TORGNY E. VINJE

ON THE RADIATION BALANCE AND MICROMETEOROLOGICAL CONDITIONS AT NORWAY STATION, ANTARCTICA

NORSK POLARINSTITUTT
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Contents

Abstract

The long-wave net radiation, Q, recorded with the Schulze radiometer is considered. Frequency distributions are discussed, and Q is related to cloudiness, temperature and wind speed. The counter-radiation from cloudless atmosphere is related to the insttument temperature and to the monthly mean screen temperature. Comparisons are made with evaluations from the Elsasser radiation chart.

Discussing the energy exchange between air and snow, one finds evidence that the transport of sensible heat is executed by the frequentation of air in the upper layer of the snow.

The annual heat budget of the snow surface is considered.

A zigzag pattern in the temperature profile near the snow surface (0–20 cm) indicates special features of the turbulence in this layer.

Oscillations in the temperature near the surface, with a period of 4–12 seconds, are considered in connection with observed wavy motions in the air.

Introduction

At Norway Station (70° 30′ S, 2° 32′ W) we made registrations with a Schulze radiometer for about a year and a half (May 1958 to January 1960). In this paper some features of the long-wave radiation components will be discussed, mainly in connection with observations made when the sun was below the horizon.

The thermopiles of the radiometer were covered with un-ventilated domes of polyethylene film. The more recent constructed Schulze radiometers have been ventilated to avoid extra radiation from the domes wich can be heated by absorption of radiation if the wind speed is too low. However, as the wind speed at Norway Station was relatively high any excess radiation from the domes should in all probability be negligible here. The instrument was found to be well fitted for continuous registrations and was operating at mean wind speeds up to 40–50 m/sec. Heavy drifting snow caused a fine scratching on the windward side of the domes (east) and reduced the transparency. This had to be considered when the short -wave radiation calibrations were taken in clear weather. For the long-wave calibration constant, however, this scratching should be of negligible importance because the long-wave radiation components are of a diffuse character.

Hoar-frost forms easily on the domes, especially in clear weather. The radiometer therefore had to be inspected frequently. Generally this inspection was made every three hours when the ordinary synoptic observations were taken. If the domes had been ventilated, as is the case with more recent constructed radiometers, the hoar-frost formation would probably have been reduced.

Norway Station is situated on an extensive flat ice shelf 36 km from the barrier and 56 m above sea level. The horizon is unbroken except of the nearly 400 m high Blåskimen 8 km towards north-west and the faint rise of the inland ice towards south.

Calibration

The calibration factor for the long-wave radiation (LF) has been evaluated by means of the temperature difference between the radiometer and the snow-surface. The deflection produced by the lower cell was generally small, but in the winter of 1959 the sensitiveness of the recorder was increased considerably, and the greatest deflections should then be fairly reliable for calibration purposes. Considering stable situations only, one gets 18 sets of hourly mean values where the deflection is greater than 0.15 scale units, which is the lower limit we have chosen. The mean value of the factor is found to be 0.110 ly/min/scale unit, with a standard deviation of 16 % and a maximum deviation of 27 %. The most important cause for the standard deviation should be the reading error which is ± 10–20 % of the deflections in question.

The factor found above can be compared with the diffuse short-wave radiation calibration factor (DSF), which for this period is found to be 0.108 ly/min/scale unit for the lower cell (see below). It is seen that the former exceeds the latter by only 2 %, and this result is in agreement with what has been found by laboratory measurements. Funk (1960) found for his radiometer that the long-wave factor is about 5 % higher than that for the short-wave radiation, and Fleischer[1] applying another method, finds that the two factors are generally almost equal for the Schulze net radiometer.

The LF for the upper cell was determined by turning the instrument upside down in fairly stable situations during clear nights. From the six comparisons made, we find a mean factor from which the single values deviate less than 5 %.

The DSF for the lower cell of the radiometer has been determined on the basis of comparisons with a calibrated Moll–Gorczynski pyranometer, taking into consideration the observed 0.3 % increase of the sensitiveness per degree decreasing temperature for the latter instrument. (This is nearly the same as found by Bener (1950) and Liljequist (1956), namely 0.2 %). From 14 sets of daily sums, restricted to overcast weather with even registrations, and with fairly high wind speed ensuring no temperature difference between surface and instrument, we find a mean factor from which the single values deviate less than one percent. This comparison shows further that the sensitiveness of the radiometer for short-wave radiation is independent of the instrument temperature. The DSF for the pyranometer has been found by integration over the upper hemisphere (see Liljequist (1956)), and the calibration factor for direct sun radiation received by the upper cell, was found by comparison with an Ångström pyrheliometer. The sensi-

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[1] Private communication.

tiveness of the radiometer increases as the height of the sun decreases, to a similar degree as for the pyranometers. When evaluating the daily sums for the radiation energy fluxes, one generally has to deal with cloudy conditions. The actual calibration factor applied for the short-wave radiation has therefore been estimated by a linear interpolation with respect to the mean cloud-amount. The radiation values are given in The International Pyrheliometric Scale of 1956.

In Antarctica we tried to find the long-wave radiation calibration factor (LF) with the aid of an unventilated black body source of radiation, which in the following will be referred to as "the tin". The space between the double cylindrical walls of the tin was filled with water, the temperature of which could be measured with a thermometer. The insulation was naturally made as effective as possible and it managed to keep the fall in the temperature of the water less than one °C per 10 minutes. When evaluating the factor from the deflection 10 minutes after the tin had been placed on the net radiometer, we get a factor whose mean is 22 % lower than the DSF. This deviation is about the same as HOINKES (1959) found (20 %) with the aid of an unventilated device.

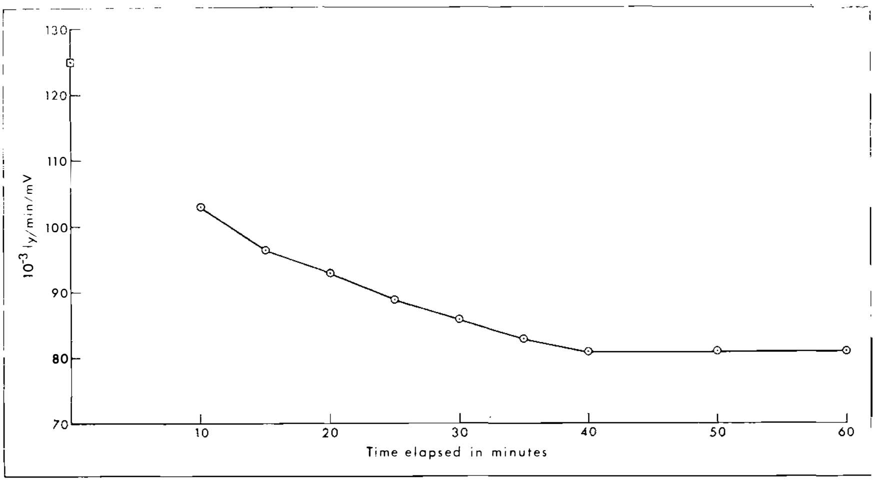

Fig. 1

The variation of the calibration factor with time as found when utilizing an unventilated black radiator. The figure are based on three series which separately shows great similarity as for the variation with time. The square on the ordinate indicates the calibration factor which has been applied in the calculations.

Taking measurements over one hour and making the necessary readings for short-time interval evaluations of the LF, we find, as shown in Fig. 1, a systematic decrease of the factor with time; converging against a constant value after 40 minutes which is about 40 % lower than the DSF. The change in the factor is probably mainly caused by the variation in the magnitude of the conductive current, which evidently must be set up within the instrument, plus extra radiation from the radiatively heated polyethylene dome. It is probable that this influence will be of importance for the factor evaluated after 10 minutes as well, so the measurements with the tin have therefore been disregarded as determinative for the long-wave radiation calibration factor. They can, however, be used in another connection as seen below.

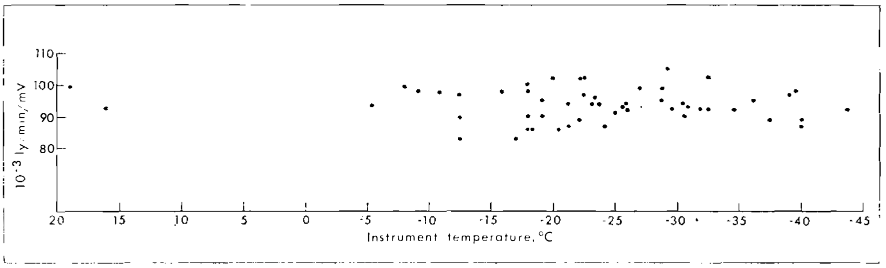

Fig. 2

Calibration factor in relation to the instrument temperature. The values have been calculated from readings taken about ten minutes after the placing of the black radiator on the radiometer.

To show the temperature independency of the factor, we have in Fig. 2 represented 58 measurements of the LF obtained with the aid of the unventilated black radiator, plotted with respect to the instrument temperature. The two measurements for the highest temperatures were kindly performed by Dr. R. FLEISCHER at the Meteorologisches Observatorium, Hamburg. He used our black radiator and also the one at the observatory, and used the same method as we did for the sake of comparability.[1] It can be seen from the figure that there is no detectable dependency between the instrument temperature and the long-wave factor. In Fig. 2 the calibration measurements for wind speed higher than 5 m/sec are omitted. The reason for this is that we find evidence that drafts possibly influence the measurements for higher wind speeds.

[1] Otherwise the long-wave calibrations at the observatory are made with a ventilated black body radiator.

Frequency distribution of net radiation for different stations

Clear weather conditions

For the period when the sun is below the horizon we have selected values of the hourly mean long-wave radiation loss from the surface. The cloudiness has been 1/8 or less and in addition fog and mist cases have been omitted. The radiometer was inspected at least once in the last three hours to be certain of no hoar-frost influence. The synoptic hours have been chosen as we then will have one observation of the cloudiness during the time-interval considered.

The frequency distribution is represented in Fig. 3 by the uppermost left curve. It is seen that about 85 % of the recorded values occur between 0.030 and 0.070 ly/min, with the highest frequency, about 35 %, between 0.040 and 0.050 ly/min. The mean loss is 0.047 ly/min. This is a somewhat lower value than what is found at "North Pole 6" between 78 and 80° N in the North Polar Basin, namely 0.053 ly/min as derived from DRIACKOGS tables (1961).

The lowest value for clear air registered at Norway Station is close to 0.020 ly/min. This is in agreement with what we later find as the evidently lowest possible value in clear weather by studying the energy exchanges between air and snow.

In Fig. 3 we have also represented the frequency distribution for the long-wave net radiation ,Q, in clear weather for some other stations in Antarctica. The values from Mirnyj, Oazis, Pionerskaja, Vostok and Sovjetskaja (SHKOLYAR, 1961, 1962) are based on momentary readings with a Yanishevski radiometer furnished with polyethylene filters, and those from Little America (HOINKES) are based on hourly means obtained with a Schulze radiometer. The observations are taken over snow-covered surfaces, except those from Oazis, which are taken over partially bare rocks. The site of the different stations is shown in Fig. 4.

It is seen that Q is not normally distributed, and further that there seems to be a more or less pronounced tendency of bi-modality for all stations. The graphs representing snow-covered areas reveal a quite narrow interval for the occurrence of the major numbers of the cases, while that for Oazis reveals a more broad distribution, probably due to a variation in the property of the under-lying ground.

The mean net radiation at Mirnyj is considerably higher (72 %) than that at Norway Station. This must in all probability mainly be ascribed to differences in the meteorological conditions prevailing at the two coast-stations. RUSIN (1960) mentions that in connection with cyclonic activity in east Antarctica the moist air which is forced upover the relatively steep coastal slope returns to the coast as a catabatic wind resembling the foehn, dry and relatively warm. He points out that

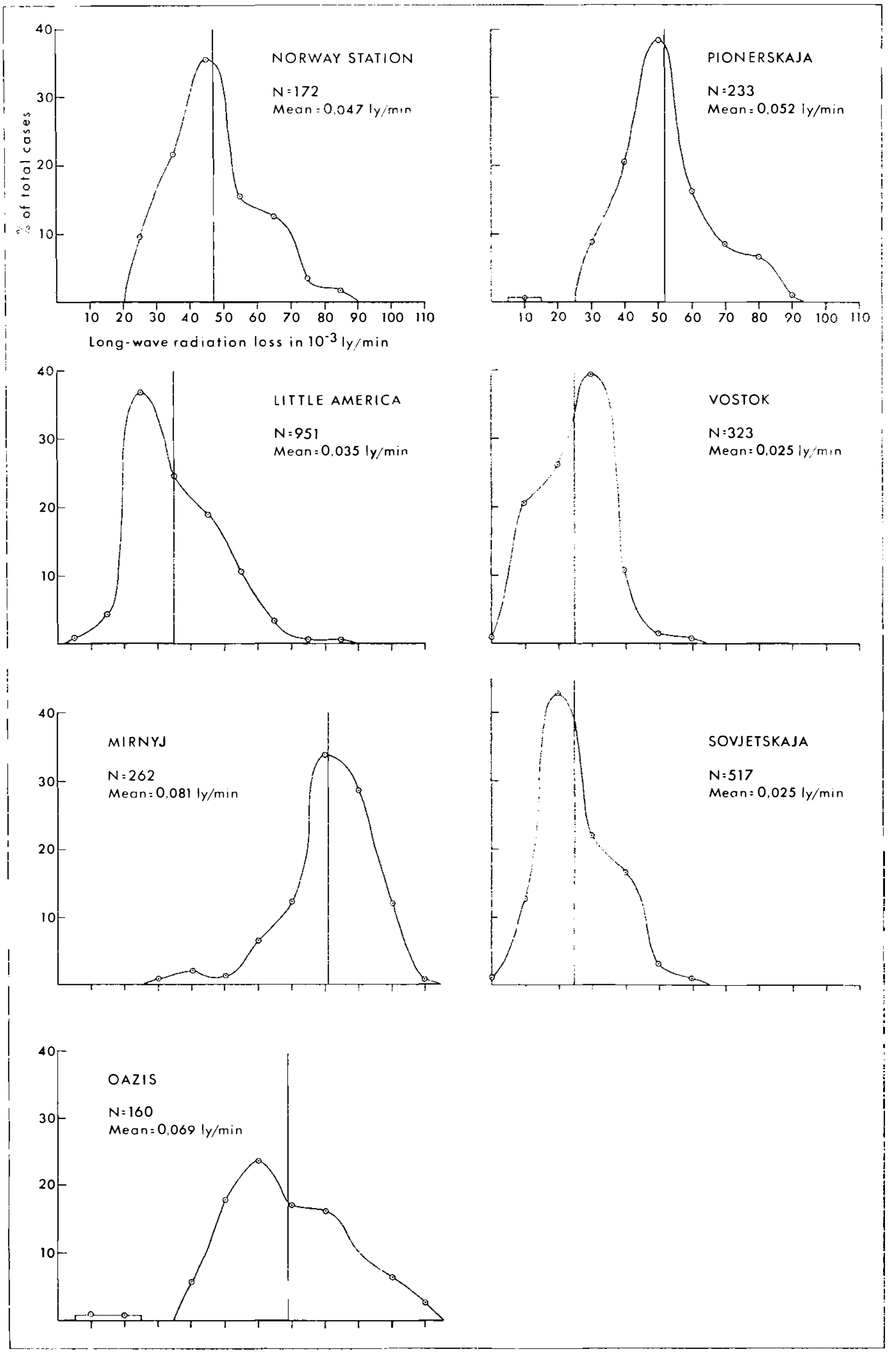

Fig. 3

Frequency distribution of radiation loss values for clear weather. Curves have been drawn to make comparison between the graphs easier. N is the number of observations. Sun below the horizon.

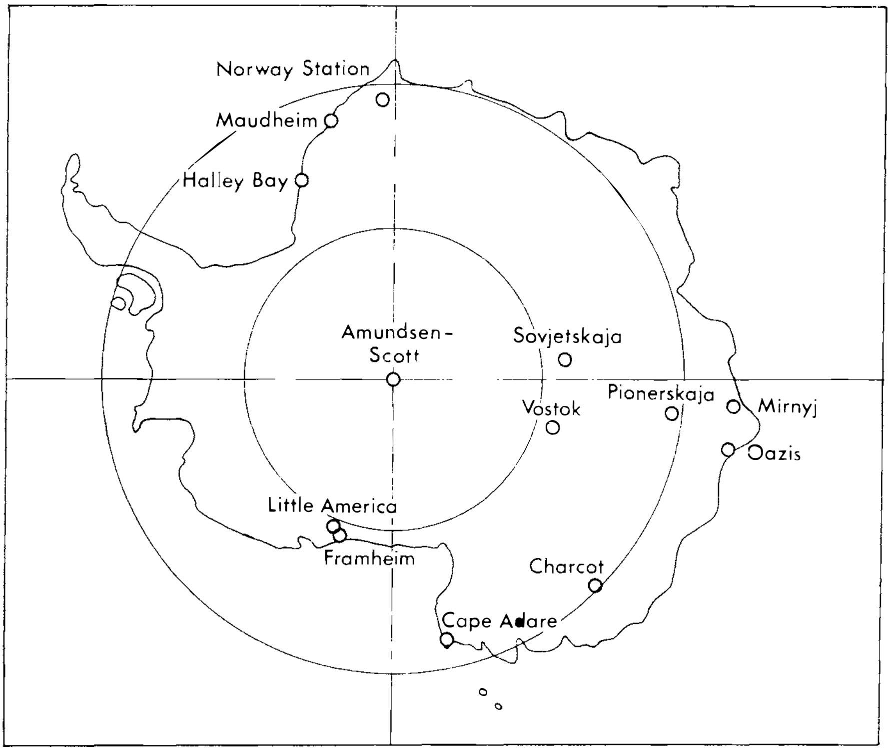

Fig. 4
Sites of Antarctic stations mentioned in this paper.

the foehn at Mirnyj is a prevalent phenomenon. In contrast we can mention that marked foehn effect at the surface at Norway Station was observed only once in three years. The phenomenon is described on page 43. The higher surface temperature and dryness of the air caused by the foehn will increase the radiation from the surface and decrease the counter-radiation from the atmosphere, resulting in a higher radiation loss. The mean relative humidity at Mirnyj is about 70 % while at Norway Station about 85 %. The annual mean air temperatures are about –10 and –17° C, respectively, thus indicating the influence of the foehn effect at the former station. The differences in the meteorological conditions at the two stations could be used as an explanation for the dissimilarity in the respective frequency distributions, and thus also for the tendency of bi-modality in the different graphs. If so, the mode to the left for the different graphs (see Fig. 3), indicated or pronounced, could represent observations when there is no dominant influence of the foehn, and the indicated mode to the right (which for Mirnyj is well pronounced) could represent observations when the foehn is predominating, at the surface or at higher levels.

The frequency distribution for Little America is, in the main, similar to that for Norway Station. This should be expected, as both stations lie on an extensive ice shelf near the coast with relatively distant mountains. The mean value of the radiation loss is 0.035 ly/min, i.e. 26 % lower than at Norway Station. It is reasonable to ascribe this difference to the combined effect of a lower radiation from the surface and less radiation from a colder atmosphere.

All weather conditions

Fig. 5 represents the frequency distribution of the long-wave net radiation in all weather situations when the sun is below the horizon. Hourly mean values for the synoptic hours have been used for Norway Station (1958–1959), hourly mean values for each hour have been used for Little America (1957) and the values for the Russian stations are momentary ones (1958).

There are some interesting features revealed in the figures. We have the marked difference in the shape of the distribution curves for the inland stations compared with those for the stations relatively near or at the coast. Compared with Fig. 3 it is further seen that the frequency distribution for clear weather for the two inland stations do not differ markedly from that representing all weather situations. This means that the change in the counter-radiation from the atmosphere will not be so marked for a clouding-over here, as it is at stations at lower altitudes where the thick clouds radiate as a black body. This "cloud effect" causes a radical change in the frequency distribution for all the coast stations and also for Pionerskaja which is situated at an altitude of 2700 m. The pronounced bimodal shape revealed on the graphs must mainly be due to the bi-modality in the frequency distribution of the cloud-amounts. For Maudheim HISDAL (1963) found that the frequency of the cloud-amounts 7 plus 8 octas is 55.6 % and for 0 plus 1 octa 26.4 %. The intermediate octas, 2-3-4-5 and 6, thus occur at a frequency of 18.1 % only, and it is for these latter cloud-amounts that we most frequently observe net radiation values corresponding to the depression area in the frequency distribution in Fig. 5.

Positive values of the net radiation means that the atmospheric counter-radiation is greater than that from the surface. This occurs when we, after a period with radiation loss, get a clouding-over and the temperature of the cloud-base is higher than that of the surface. It is interesting to see that it is Little America which has the highest frequency of positive values of the net radiation and also the highest single values, between +0.030 and +0.040 ly/min. This is, however, reasonable as the station lies far more south than, for instance, Norway Station, and therefore displays a lower surface temperature. As it moreover is a coast station, the advective clouds which reach this area should give rise to a greater contrast between cloud-base and surface temperature than is possible at Norway Station.

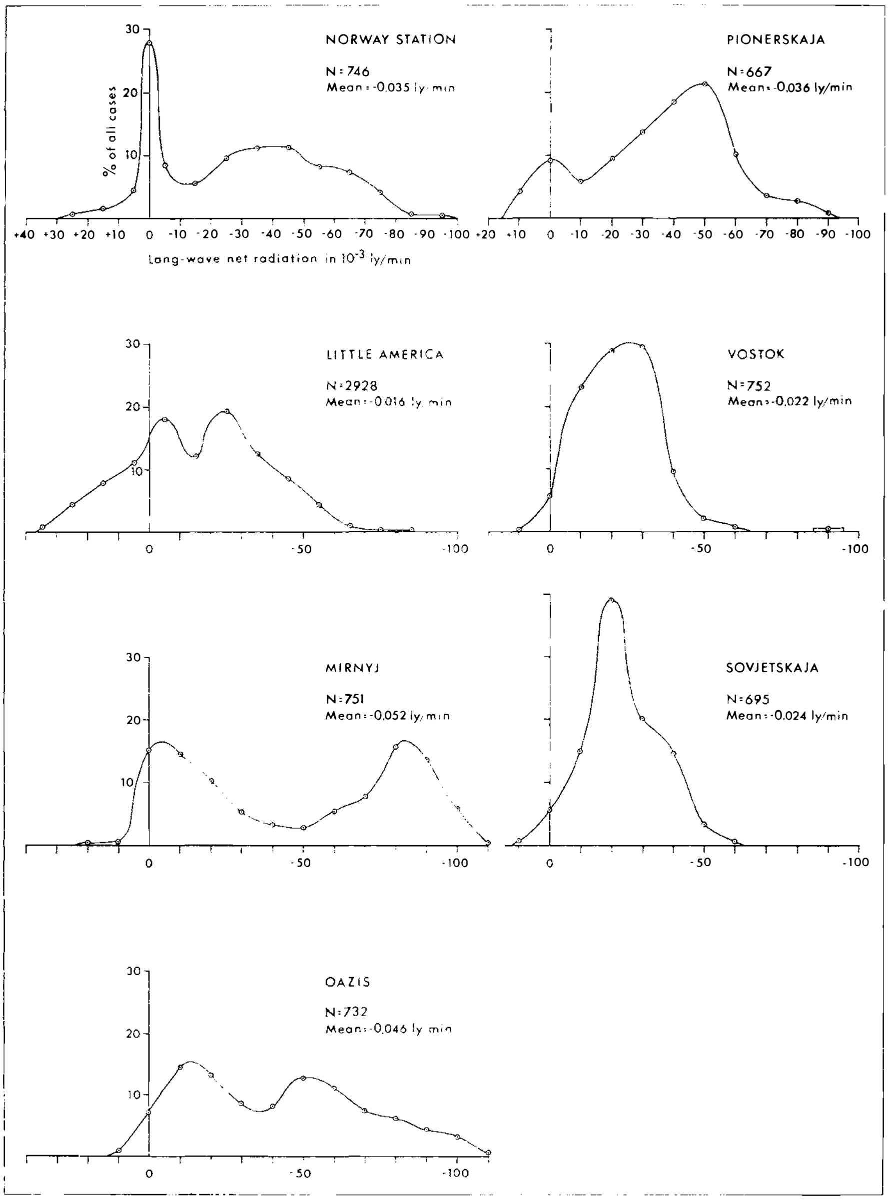

Fig. 5

Frequency distribution of the net radiation for all weather conditions. Curves have been drawn to make comparison between the graphs easier. N is the number of observations. Sun below the horizon.

Net radiation related to various meteorological elements

Cloudiness

One of the main deciding factors for the magnitude of the net radiation is the amount of low and medium clouds. Cirrus clouds are too thin and, moreover, due to their great height, they have a far lower temperature than the lower clouds. Therefore they have generally very little influence on the magnitude of the net radiation. Fig. 7 can serve as an illustration of the change in various meteorological elements for a clouding-over or for a clearing up.

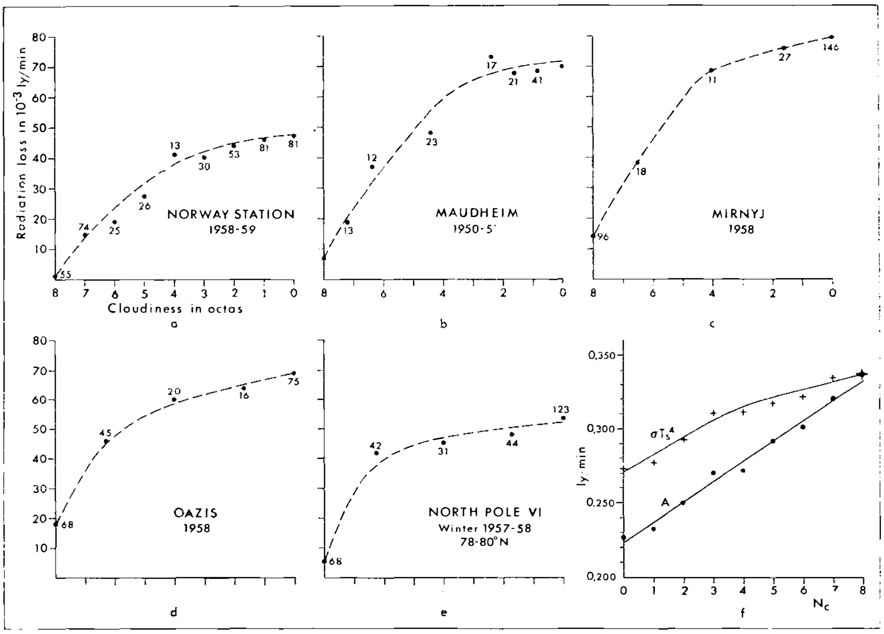

Fig. 6

Radiation loss and its components in relation to cloudiness.

In Fig. 6 we have represented the mean net radiation with respect to the different amounts of cloudiness for various stations in a polar climate. We have disregarded cases where cirrus clouds have been dominant, for reasons mentioned above. For a broken sky the cloud-amount will generally be increasing or decreasing. Overcast weather, however, can last for days with only small temperature variations due to advection. If we now, therefore, (for Norway Station) consider

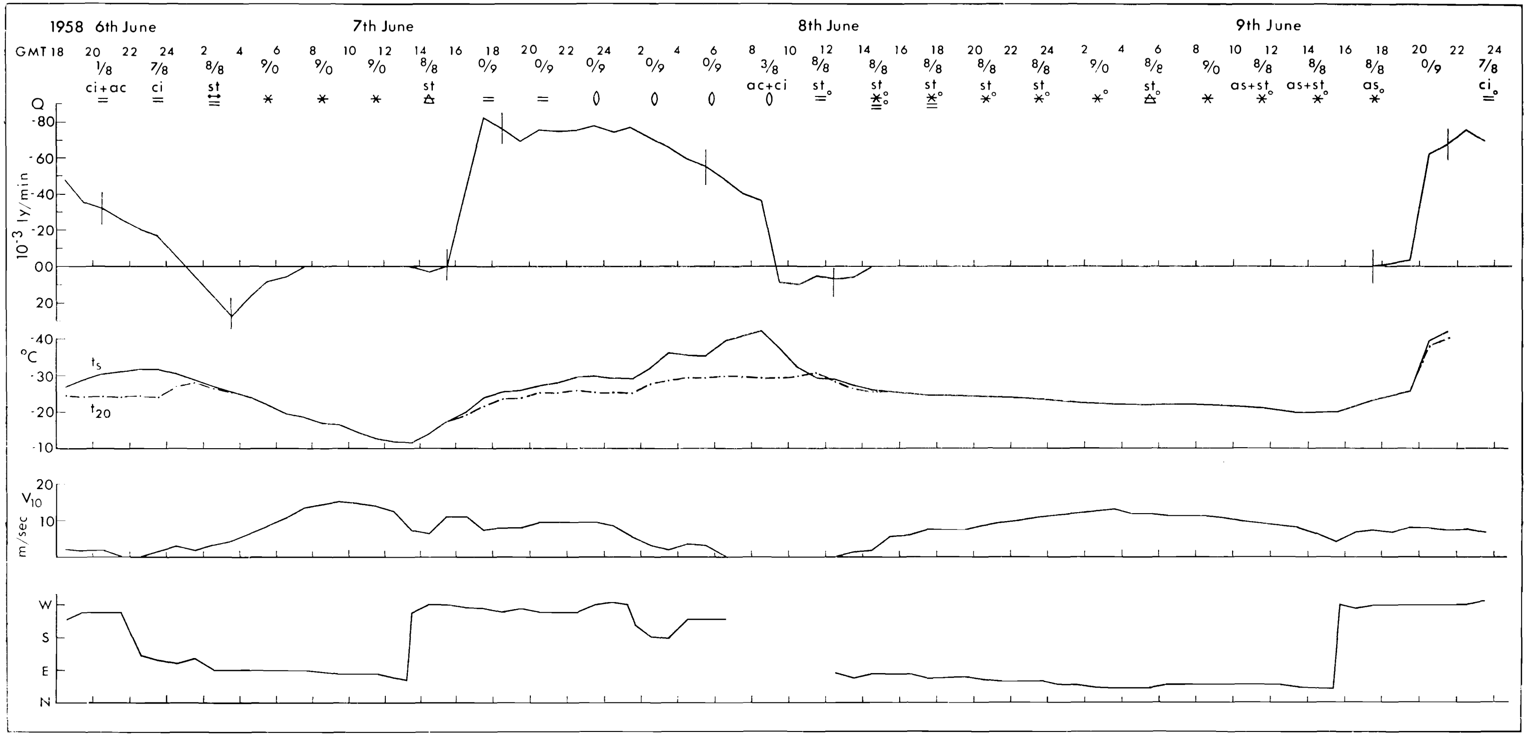

Fig. 7

The course of different meteorological elements for changing cloud conditions. Amount of clouds, the types and different meteorological phenomena are represented by international abbrevations. The curves give the variation of the net radiation, the surface temperature (solid line) and the temperature at c. 20 m (broken line). Further the wind speed at the 10 m level and the direction. Hourly mean values have been used. The vertical lines limit periods used in the computations below.

only the cases for $N = 8/8$ immediately after the sky has become overcast or just before it starts clearing up, we should get more homogeneity in the material as the cases used for all octas then should represent situations when we either have a clouding-over or a clearing-up. (Table 1.)

Table 1

Meteorological elements for different amounts of low or/and medium clouds.

Cloudiness (in octas)	Number of observations	Surface temperature (°C)	Net radiation (ly/min)	Atmospheric radiation (ly/min)	Wind speed at 10 m (m/sec)
8	55	-20.2	0.000	0.337	7.5
7	74	-20.6	-0.015	0.320	7.2
6	25	-23,3	-0.019	0.302	7.4
5	26	-23.9	-0.027	0.291	6.5
4	13	-25.1	-0.041	0.271	6.4
3	30	-25.5	-0.040	0.270	6.1
2	53	-28.7	-0.044	0.250	5.3
1	81	-32.1	-0.046	0.232	4.9
0	81	-33.0	-0.047	0.227	5.0

For Norway Station we have used hourly mean values of Q for different observed cloud-amounts (N_c) at every three synoptic hours during the polar night. The net radiation at Maudheim, given by Liljequist (1956), and at the Russian stations are based on momentary values. At the former station this value refers to the pyrgeometer level, while otherwise to the surface. All the graphs reveal curves of more or less the same shape. It is especially interesting to see the relative small decrease in the net radiation when the cloud-amount increases from 0 to about 4/8, compared with the 4 to 6 times greater decrease for greater cloud-amounts. In order to trace the reason for this curvature we have represented the mean temperature radiation from the surface (σT_s^4) and the mean measured atmospheric radiation (A) with respect to N_c in figure 6 f for Norway Station. We see that the relation between A and N_c is very near a linear one. The latter relation can by using the method of least squares be expressed by the following equation:

1) $$A = 0.223 + 0.0138\ N_c$$

where the maximum deviation of the group mean values from those obtained with this equation is less than ± 2.5 %.

Liljequist (1956) finds a more complex relation which can be compared with the trend in the observation series from Fairbanks made by Wexler (Liljequist 1956 p. 156). Other authors, Ångstrøm (1936) and Bolz (1949), have proposed or derived quadratic formulae.

However, the equation above should give a good representation of average conditions at Norway Station. From Fig. 6 it is seen that it must be the variation of the surface temperature which causes the curved relation between the net radiation and the cloud-amount, and it is probable that it is the decreasing heat support to the surface from the gradually destroyed inversion which is the original cause for this curvature. Considering the figure it is seen that when N_c goes from

0 to 4 octas, the increase in the atmospheric radiation is nearly balanced by the increase in the radiation from the surface, while for greater cloud-amounts σT_S^4 is decreasingly dominating mainly because of the abatement of the heat support from the weaker inversion. (See page 40.)

It can be shown that there is clear evidence that the change of A for mean conditions also hold for actual cases when there is a clearing-up or a clouding-over. We have found 14 cases during the polar night 1958 which are well defined in this respect. Cases where the cloud-observations might be doubtful or when we have fog or thick mist have been omitted. For the different 14 cases we have found the value of A both for clear sky and for 8/8 of low or medium clouds. The variation of A with the cloud-amount has been calculated, and we find 0.0130 ly/min per octa ± 22 %, for clouding-over and for clearing-up situations. The value is nearly the same as that previously found for mean conditions, namely 0.0138 ly/min per octa.

Temperature in clear weather

In Fig. 8 the hourly mean net radiation from surface for a clear sky is plotted against the hourly mean surface temperature. It is seen that there is a marked decrease in the radiation loss for decreasing surface temperature to begin with. However, below –35° C the surface temperature does not seem to have any influence upon the magnitude of the net radiation. This marked change in the relationship can not be due to a regular continuous cooling of the air and of the snow surface. It is assumed that we here see the effect of another phenomenon: The draining of the atmosphere which is disclosed by the formation of ice needles at lower temperatures, bringing about a certain change in the decline of the atmospheric radiation. This change has been demonstrated to the right in Fig. 8. In Fig. 9 is given the frequency distribution of ice needles observed when the sky is clear. The relative humidity with respect to ice is also given, and we note the increasing supersaturation and frequency of observed precipitation of ice needles as the temperature falls. The cases represented in the graphs are collected

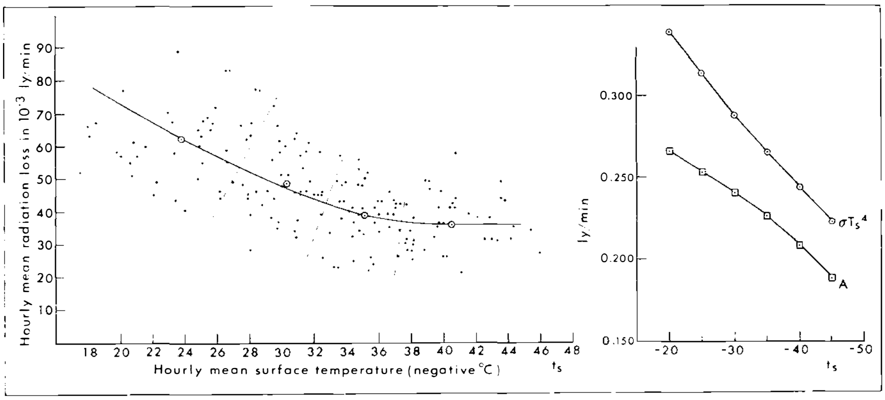

Fig. 8

Hourly mean radiation loss and its components in relation to the surface temperature.

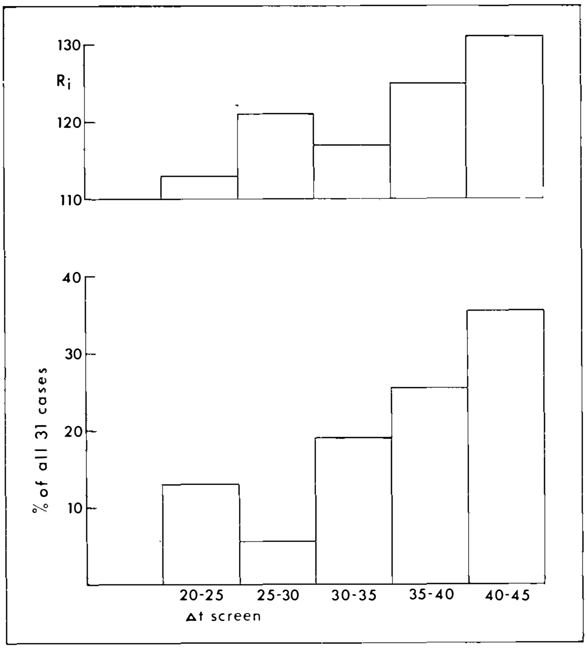

Fig. 9

Frequency distribution of the occurrence of precipitation of ice needles in clear weather in relation to the screen temperature. The upper graph shows the relative humidity with respect to ice.

from three years 1957, 1958 and 1959, and cloudy situations have been excluded to be sure that the hydrometeors were not precipitated from clouds.[1] The observations show clearly that the probability for the occurrence of ice needles, i.e. draining, increases rapidly with decreasing temperature. One could perhaps have expected to find more cases in a period of three years, however, the precipitation of ice needles is more a short-lived process which necessarily makes the number of observations smaller.

For a later use an equation for the relation between Q and t_s, representing average conditions, has been found. The groupings have been made as indicated by broken lines in Fig. 8, where each group contains an equal number of observations. The coefficients have been found by means of the method of least squares.

$$2) \qquad Q = -0.171 - 0.0063\, t_s - 0.000073\, t_s^2$$

Fig. 10 represents group means of the net radiation from the surface with respect to the instrument temperature for Norway Station, Maudheim (LILJEQUIST 1956) and for Little America (HOINKES 1963). All groupings are made within intervals of about 5° C of the instrument temperature, t_i.

Maudheim and Norway Station are sited so close to each other on ice shelves that the climatic conditions at and above the two places should be quite similar. In spite of this, the radiation loss given for Maudheim is considerably higher than that for Norway Station. The excess is in mean 27 %. The difference is in all

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[1] If we count all cases of observed ice needles, without snow-fall, we get 154 cases for the same period.

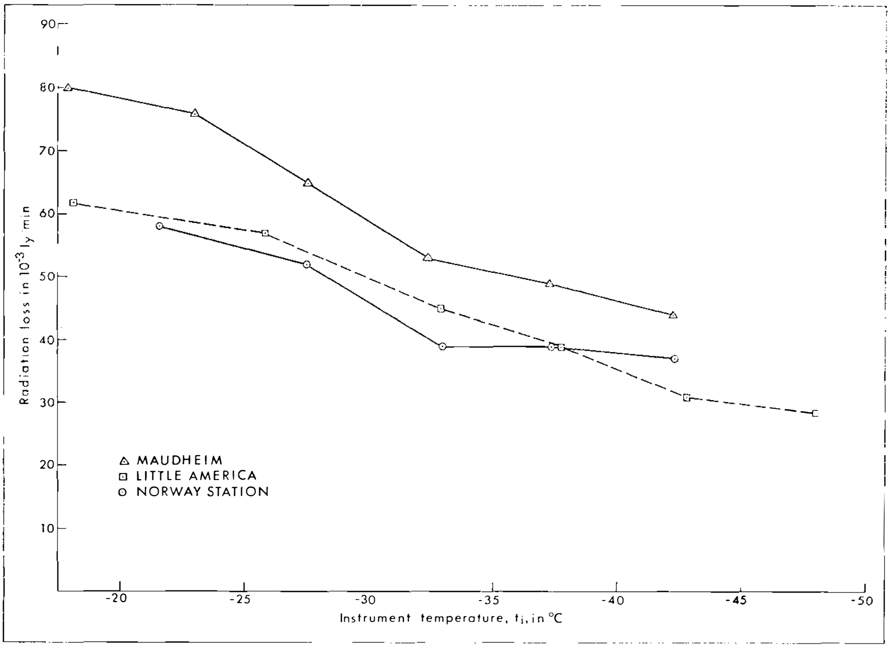

Fig. 10

Radiation loss in clear weather in relation to instrument temperature.

probability due to wind influence on the instrument as will be shown later.

At Little America and Norway Station the same kind of instrument and calibration factor have been used so the measurements can be compared. We are interested in the counter-radiation from the atmosphere during clear weather. There can easily be differences in the property and magnitude of the inversion at the two stations. To avoid this, we may consider the values with relatively high temperatures at which the wind speed probably will be so high that it diminishes the magnitude of the inversion considerably. Hence for temperatures above –30° C it is found that the radiation loss at Norway Station in mean is 5 % higher than at Little America. This means, according to the measurements, that for similar surface temperatures there is a very little difference, about 1 %, in the atmospheric counter-radiation at the two stations.

Wind speed in clear weather

Grouping Q within intervals of 2 m/sec in the wind speed we get, as shown in Fig. 11, group means which lie very nearly on a straight line. Applying the method of least squares, the equation for the line of best fit is found to be

$$3) \qquad Q = -0.030 - 0.0036\ V_{10}$$

A similar grouping has been done by LILJEQUIST for his clear weather measurements with the pyrgeometer. He found for wind speed less than 7 m/sec the linear relation

$$Q = -0.030 - 0.0053\ V_{10}$$

Above 7 m/sec he finds a non-linear relationship. We note that for calm weather, the mean values given by the two equations are exactly the same, like –0.030

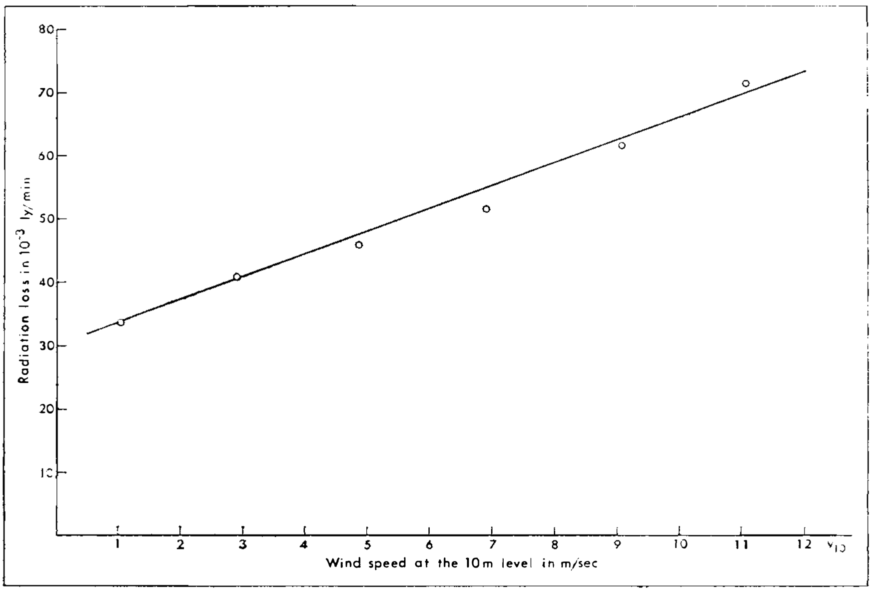

Fig. 11

Radiation loss in clear weather in relation to the wind speed at the 10 m level. Group means with respect to intervals of 2 m/sec in the wind speed.

ly/min. This indicates the similarity in the calibration factors applied. For $V_{10} =$ 7 m/sec the Maudheim observations of Q are 22 % higher than ours and for 12 m/sec, 32 % (which can be found from Liljequist's tables). The comparison thus reveals an increasing difference with increasing wind speed. It is not thought that the shielded Schulze radiometer is affected to any great extent by the wind. The discrepancy may therefore probably be ascribed to wind influence on the pyrgeometer measurements.

It is of interest to see if the clear weather observations previously dealt with represent equilibrium conditions. As a definition for equilibrium we have postulated that the temperature variation at the surface should be less than one degree per hour plus that no special disturbances take place at the instrument level or at the 20 m level. Applying this framing on the 172 clear-weather observations, we get 61 cases left. The observations have been grouped with respect to V_{10} as follows: 0 – 3 m/sec (18 cases), 3 – 7 m/sec (23 cases) and 7 – 12 m/sec (20 cases). By the method of the least squares we have found the coefficient in the linear relation which is

$$Q = -0.029 - 0.0037 \, V_{10}$$

This last equation can be compared with 3) page 19. The close agreement found should mean that the group means of all measurements for clear weather represent mean equilibrium conditions.

One can from a combination of the equations 2) and 3) above calculate the magnitude of the variation of the surface temperature with respect to the variation in the wind speed. By derivation we get

$$4) \qquad \frac{dt_S}{dV_{10}} = \frac{36}{63 + 1.4 \, t_S} = \delta$$

For a variation in the wind speed of one m/sec we get a change in the surface temperature of 1.1° at –20°, 1.4° at –25°, 2.0° at –30° C and 2.6° at –35° C. For lower values of the surface temperature, the calculated value of δ is in all probability not true, as the draining effect, mentioned on page 17, could partly account for the mean relation on which the equation above is built.

LILJEQUIST (1957) has studied the variation with time of the temperature at 0.5 m with respect to the wind speed at the 10 m level. Based on six days registrations in so to say clear weather he gets in mean $\delta = 1.7$. HISDAL (1960) found the same value for the screen level by studying the variation of the mean temperature deviation for situations with a clear sky during the winter half year. His result is based on the tentative assumption that the thermal advection effect for southerly and easterly winds (the two directions he studied for this purpose) is approximately equal but has opposite signs. SVERDRUP (1933) found $\delta = 3.8$ in the pack ice of the North Polar Sea. That he gets a greater value is reasonable as a wind increase over the more uneven surface should set up a relatively greater turbulence and thus cause a greater energy exchange than over the flat Antarctic ice shelf.

Clear sky atmospheric radiation in relation to various meteorological elements

Instrument temperature

To compare the atmospheric radiation with surface conditions in clear weather we have selected the hourly mean values of different surface elements for hours at which an ascent has been made (12 or 24 h) when the sun is below the horizon (or very nearly so, in December). We get 85 cases.

The counter-radiation from the atmosphere can also be treated as a black body radiation with an effective temperature T_a (Kelvin deg.). If the instrument temperature is T_i, we have

$$Q_1 = \sigma T_a^4 - \sigma T_i^4$$

where Q_1 is the net radiation from the upper cell. The instrument temperature and the atmospheric radiative temperature given in deg. centigrades are t_i and t_a, respectively. t_a is easily found from the equation above. In Fig. 12, t_i is plotted against t_a. The line running through the two group means, which are formed with respect to equal intervals in t_a, can be expressed by the simple equation

5) $$t_i = 11.6 + t_a$$

(We have omitted here a discussion of the correlation coefficient as the physical meaning of this quantity is obscure due to the automatic correlation involved.)

For a given atmospheric radiative temperature the maximum deviation of the instrument temperature, will, as seen from the figure, be about ±5° C, corresponding to a maximum deviation in the net radiation of about ±0.025 ly/min. This great deviation is in accordance with what roughly can be found from Fig. 11, when the wind speed varies within its natural range for clear weather. The cause for the considerable deviation can therefore probably be ascribed to a certain extent to the influence of the wind upon the surface temperature. The wind executes the heat transport from the inversion to the snow and can therefore, due to its variation, give rise to an alteration of the surface temperature, within a certain range (about ± 5° C) for a given atmospheric radiative temperature. It is previously shown (page 20) that average clear sky conditions represent average equilibrium conditions. The measurements considered here have been collected under the same condition as for those previously treated, so we can reckon that also equation 5 represents the mean equilibrium state. Therefore, in other words, *when equilibrium is obtained, the temperature near the surface (in the instrument height of 1.5—2m) will be, on an average, c. 12° C higher than the atmospheric radiative temperature for any value of the former between about – 10 and – 45° C.*

For the station "North Pole 6", drifting between 78 and 80° N in the Arctic Sea, is given the instrument temperature and the radiation loss at this level when

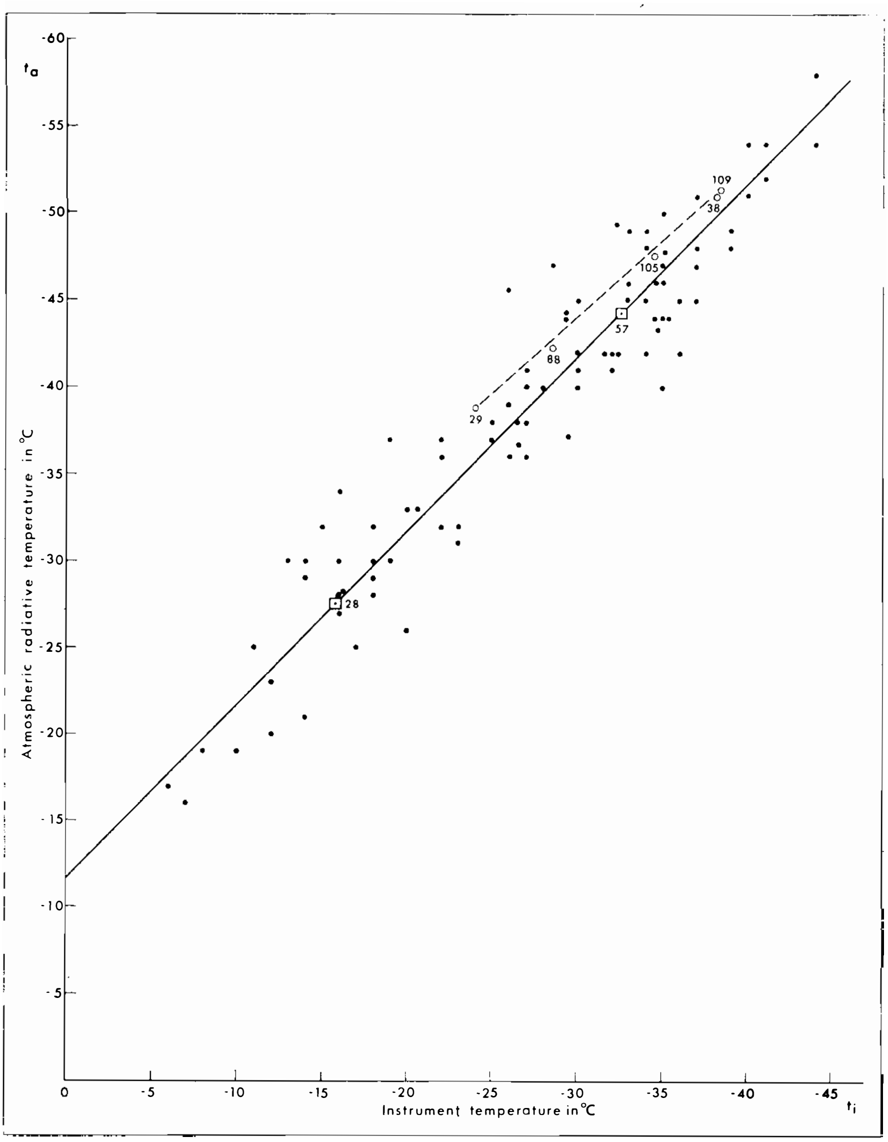

Fig. 12

Relation between the instrument temperature and the atmospheric radiative temperature in clear weather.

the sun is below the horizon (DRIACKOG 1961). Monthly mean values have been noted in Fig. 12. It is seen that the points are lying very nearly on a straight line, which, however, reveals a slightly less declination than the line for Norway Station. For a given mean instrument temperature the mean atmospheric radiative temperature is found to be from 3.0 to 1.2° C lower here than at Norway Station.

Monthly mean screen temperature

For Antarctic stations there have often been found a double (or multiple) minimum for the monthly mean temperature in the winter half year, besides the general pattern of "der kernlose Winter". WEXLER (1959) gives an attempted ex-

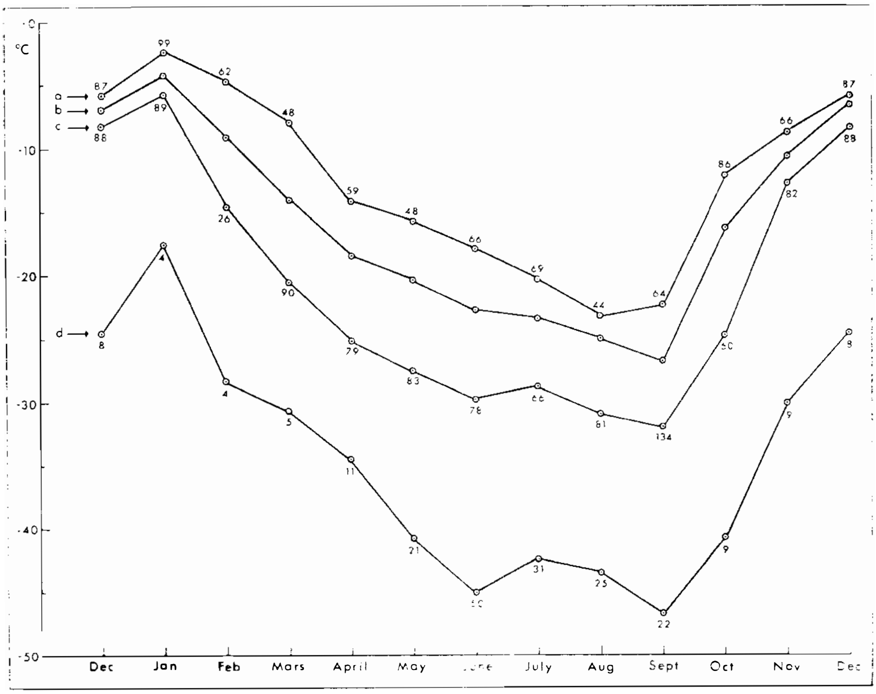

Fig. 13
Annual variation of different temperatures: a) for overcast weather, b) for all weather conditions, c) for clear weather, and d) gives the atmospheric clear weather radiative temperature.

planation, where he proposes that the double minimum can be explained by changes in the intensity of cyclonic activity. The following discussion supports the view that the phenomenon can be explained by subsidential effects as well.

In Fig. 13 we have represented the mean annual course of four different temperatures for 1958 and 1959. Each curve is marked with a letter. Curve a) represents the monthly mean temperature at screen level for 8/8 of low clouds and should indicate a monthly mean surface temperature for situations with advection of air from lower latitudes together with the indicated frequency of this phenomenon given by the number of observations for each month. It is seen that this temperature curve has *one* minimum. Curve b) represents the monthly mean temperature at screen level. This curve has a somewhat different shape, with a marked minimum in September and a more flat course in the preceeding winter months. Curve c) represents the monthly mean temperature at screen level for clear sky. This curve has a distinct double minimum, one in June and one in September. This indicates that the tendency of a double minimum in the monthly mean screen temperature is due to changes in processes taking place when the sky is clear. Curve d) represents the monthly mean atmospheric radiative temperature, t_a, in clear weather, and it is seen that this curve has a pronounced *double* minimum. The figure reveals that we have a warmer atmosphere in July and August than in June and September. It is therefore reasonable to believe, when comparing the latter curve with curve a), that this is due to heating caused by subsidence. The temperature t_a has a major influence upon the surface temperature in clear weather and thus also upon the monthly mean temperature.

From the upper air measurements at Maudheim, SCHUMACHER (1958 p. 36) arrived at the conclusion that heating due to subsidence "seems to play an important role for the determination of the temperature distribution during winter in the lower part of the polar atmosphere". HISDAL (1960) also proposes this effect to be important for the explanation of the flat winter minimum of the surface temperature of polar stations. Moreover, SCHUMACHER found that in the winter season the mean tropospheric temperature, especially at 850 mb, was considerably lower for wind directions with a northerly component than for wind directions with a southerly component.

Long-wave radiation and the Elsasser radiation chart

We will now compare our registrations of the atmospheric long-wave radiation with values evaluated on a theoretical basis, using the radiation chart described by ELSASSER (1942). Other radiation charts have also been constructed but none of them have yet been generally accepted as the "best one". We consider the 85 cases restricted to the hours when an ascent has been made. The grouping of the observations is made with respect to intervals of 10° C in the atmospheric radiative temperature. The respective means of the vertical distribution of the temperature can be seen in Fig. 14.

In Table 2 is given the mean values of the measured and evaluated atmospheric long-wave radiation flux in ly/min, received at the surface.

Table 2

Atmospheric long-wave radiation received at the surface in ly/min. Mean values.

Temperature at 70 mb	Measured	Evaluated from Elsasser chart	Difference	Number of observations
-42	0.335	0.283	0.052	8
-58	0.288	0.259	0.029	20
-75	0.242	0.215	0.027	31
-77	0.207	0.194	0.013	26

As seen, all the measured values are greater than the evaluated ones. As ELSASSER has considered radiation from water vapour and carbondioxide only, it seems reasonable to ascribe some of this difference to the radiation from the ozone in the 9.6 μ band which lies within the "window" of the water vapour (9–13.5 μ), where also CO_2 has an extremely weak absorptive power. It is seen that the previously mentioned difference becomes greater when the measured atmospheric radiation increases. We thus get an agreement between the increasing difference and the increasing radiative power of the ozone layer due to its increasing temperature. From the ozone soundings taken at Halley Bay, MAC DOWALL and SMITH (1962) found the vertical distribution of this gas. The stratospheric ozone layer begins at the tropopause and has its maximum concentration at 16–18 km, i. e. at about 70 mb. The total amount of the gas is roughly 0.270 cm (standard temperature and pressure) in winter, with a flat minimum, and 0.350 cm in the summer half year,

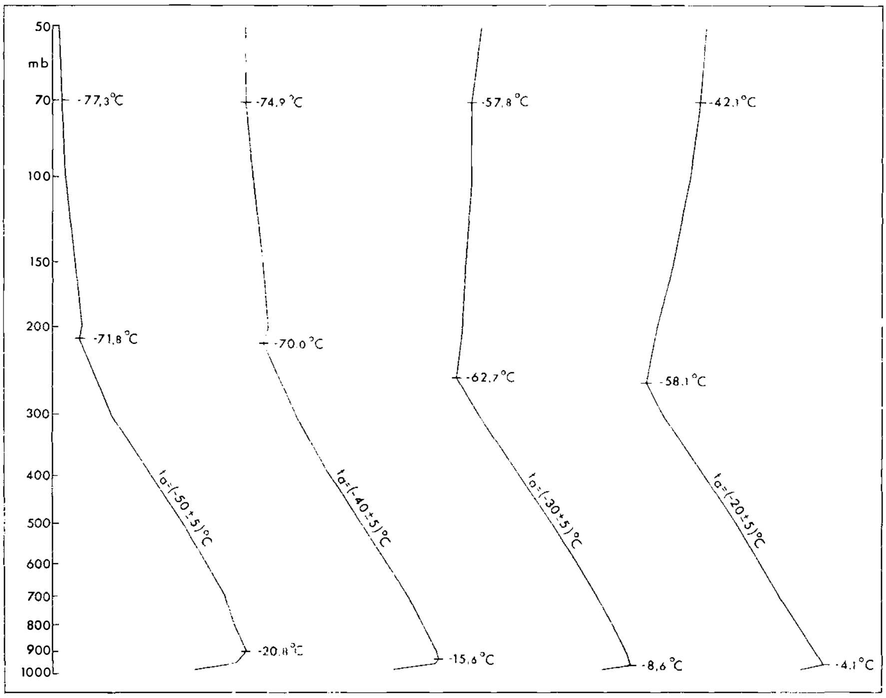

Fig. 14

The mean vertical temperature distribution within intervals of 10° C in the atmospheric radiative temperature.

with a sharp maximum in November. They found further that it is tropospheric values of meteorological elements which show the closest relationship to the amount of surface ozone. The best covariance is found between the monthly mean air temperature gradient in the first 1.1 km and the surface ozone concentration; the deeper the inversion the greater the amount of surface ozone.

PLASS (1956) has calculated the downward long-wave flux due to emission from the ozone in the 9.6 μ band. The total amount of the gas in his example is between 0.210 and 0.260 cm, which is lower than that found over the Antarctic coast. For a surface temperature of −10° C the downward fluxes due to long-wave radiation from the ozone is calculated to be about 0.015 ly/min.

It is supposed that the lower the surface concentration of the ozone the greater is the radiation which can reach the surface from the stratospheric ozone layer. In summer time, when the concentration at the surface is at its minimum (corresponding to the relatively small vertical temperature gradient) it is assumed that radiation from the layer in the stratosphere reaching the surface, can obtain its annual maximum, especially as the temperature at this height is considerably higher in summer than in winter. As the total amount of ozone in Antarctica is higher than in the examples used in PLASS' calculations, it is possible that the downward flux in this region at summertime could reach higher values than 0.015 ly/min. However, consideration of the radiation from the ozone alone does not give a satisfactory explanation of the considerable difference noted in Table 2. There are probably also other factors involved.

Besides some uncertainties about the absorption coefficient for the ozone gas, the correct values of the radiation fluxes from water vapour and carbondioxide are questionable. At first we have a simplifying assumption made by ELSASSER to overcome the difficulties with the carbondioxide and further there is not full agreement as to the pressure effect upon the absorptivity of the water vapour. For our calculations we have used the factor $(p/p_s)^k$ where $k = 1/2$. In a recent work by MANABE and MÖLLER (1961) k is considered as usually smaller than unity. In their computations they adopted $k = 0.6$, tentatively. On p. 507 they say: "Strictly speaking, k is not constant and depends upon amount of absorbing gas and pressure. When the amount of gas is small, k decreases significantly with decreasing amount of gas". Therefore, due to the small amounts of water vapour found over Antarctic Stations, it is probable that values of k less than 0.5 should be applied. This would give a greater value of the factor $(p/p_s)^k$ and consequently bring about an increase in the calculated flux.

From this it is not reasonable to expect full agreement between the measurements of the atmospheric long-wave radiation and those values calculated by the ELSASSER radiation chart. The radiation from the ozone should be taken into account, as well as the latter mentioned possibility for a smaller value of k than 0.5 for the very small amounts of water vapour in the Antarctic atmosphere.

Heat exchange between air and snow

Measuring devices

In June–July 1958 we made some measurements of the temperatures from about 3 cm above to about 36 cm below the snow-surface. The series of measurements were taken when the sun was below the horizon, so we will consequently be dealing with simplified radiation conditions.

We had 12, about 10 m long, thermo-couples which were connected with a Siemens millivoltmeter. The inactive cell, or the reference cell, was fastened to a mercury thermometer and was placed indoors in a well-insulated box to keep its temperature constant while performing the measurements. The twelve thermo-couples were connected with a switch, and the twelve deflections on the millivoltmeter could be read in less than one minute.

The end-couplings of the thermo-couples were made by twisting the respective pair of threads together and afterwards covering the junctions with white varnish paint.

All the thermo-couples were compared with each other by two series of measurements in which we had a difference in the temperature of the active cell and of the inactive cell of about 20° C and of about 40° C (marked with M in Fig. 15). We found that the maximum deviation from the mean of the response temperature for all thermo-couples was less than ± 0.3 %. Due to this the further calibrations were carried out for one thermo-couple only. (It can be mentioned that if the end-couplings were soldered the above-mentioned deviation became considerable.) The calibration curve is linear, as seen from Fig. 15, and its inclination is 0.612 deg. per scale unit.

Eleven thermo-couples were fastened to a wooden stake at fixed distances from the top, and one was fastened to a small plate of tin which would press the (uncovered) thermo-couple against the snow surface. We dug a hole in the snow by pressing a pipe into it and pulling out a cylinder of the material. The stake with the thermo-couples was pressed against the wall in the hole so that the protruding junctions were stuck into undisturbed layers. The space around the stake was filled with snow which was pressed hard together.

Meteorological conditions

To find the amount of heat transferred from air to snow and the conduction of heat from below to the surface, we have to look for simplified conditions during the polar night when we have no influences due to advection, precipitation,

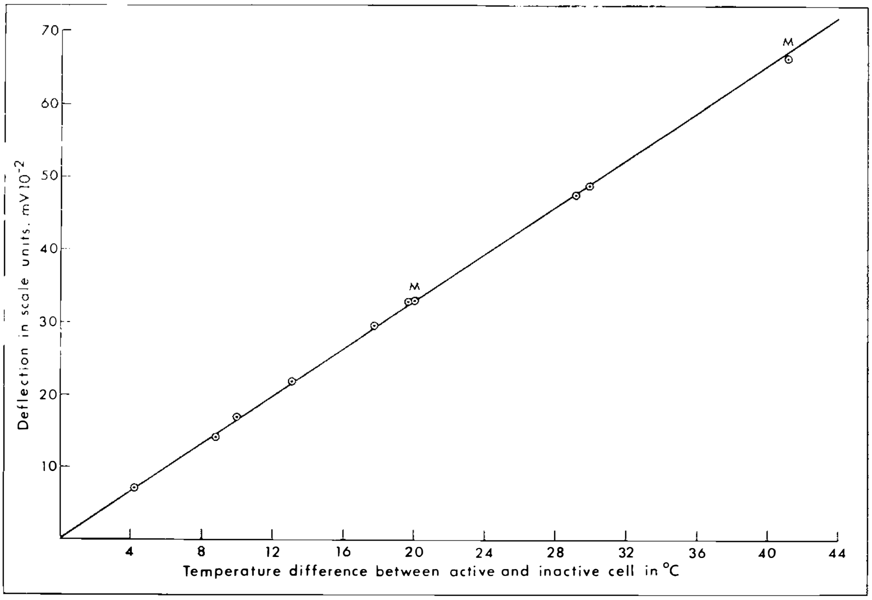

Fig. 15
Calibration curve for the thermo-couples.

drifting snow or air pressure variations. At the time the measurements were made, not all these conditions were fullfilled, so we have had to make some eliminations. Regarding the advection we have omitted three cases; two, where the variation of the surface temperature was about four times greater than the greatest variation (1.0°C/h) for the rest of the cases, and one, for which there was an abnormally great persistent fall in the temperature at the height of 20 m. Three cases have been left out due to slight precipitation and a similar number due to great pressure variations together with drifting snow. (A change in the air pressure will change the amount of air in the permeable snow.)

After these omissions there are 14 cases left with measurements of the temperature profiles at the beginning and at the end of certain time intervals, within which there should be negligible advective effects in the lowest 20 m of the air, no precipitation, no drifting snow and no great pressure variations. Four of the cases are for positive and ten for negative net radiation.

Temperature profiles

Fig. 16 shows the temperature profiles used for the evaluations. The graphs with letter designations k, l, m, and n are for situations with long-wave radiation gain at the surface, the others are for radiation loss situations. A prerequisite for the transport of heat from air to snow ,or vice versa, is a temperature difference in the exchange area, and we see that the surface temperature is generally an extreme one, and a considerable difference is found between the snow surface and the adjacent air (at 0.5 to 1 cm). This difference is for radiation loss situations (when the surface is coldest) about 0.5° C, and for radiation gain situations (when the surface is warmest) about 0.4° C. For the profiles in Fig. 16 n, the wind is so strong

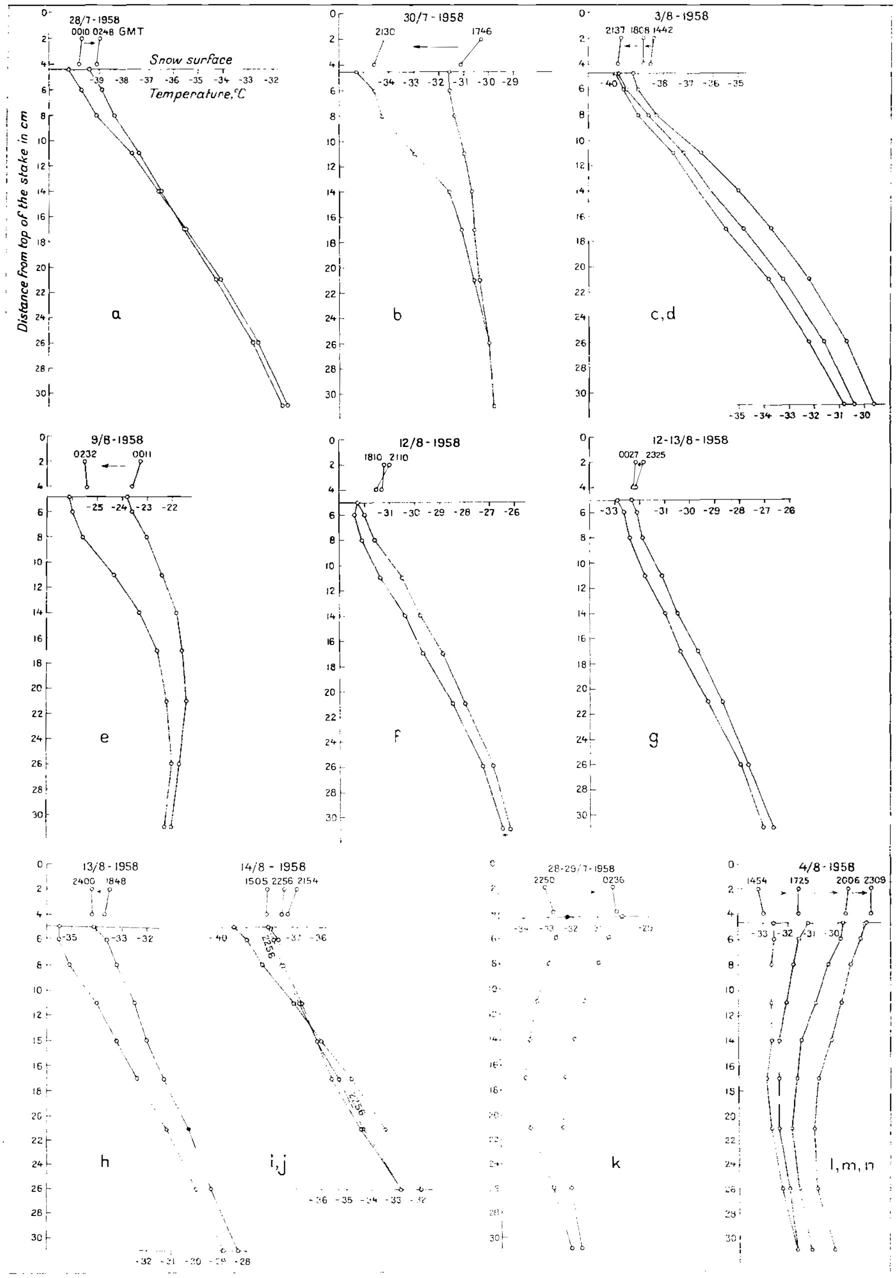

Fig. 16

Temperature profiles used for the calculation of the energy exchange between air and snow.

that turbulent transfer of heat from above probably dominates and thus causes the air temperature to be higher than that of the surface.

In the pack ice in the North Polar Sea, SVERDRUP (1933), found a 0.5° C higher temperature in the air at 80 cm than at the snow surface. The cloudiness was 0–3/10 which means that there must have been radiation loss. He found also that, as a rule, the snow surface was warmer than the air when the sky was overcast.

TAKAHASHI and SOMA *et al.* (1956, abstract) says that: "The lowest temperature is observed at a depth of about 0.7 cm. This feature of the profile is attributable to semi-transparency of snow layer, and is also demonstrated theoretically, assuming that radiation takes place in the snow layer down to a certain depth. In the case when heat transfer between snow and air is small due to weak turbulence of air, the surface of snow is kept at the lowest temperature". The density of the snow is in this case only 0.16 gr/cm³. As in our case the density is very nearly 3 times as high, it would be natural to expect radiation from the snow surface only. A closer examination (not reproduced here) of our data for the purpose of finding influences on the profiles due to net radiation from below the surface did not produce any significant results.

Fig. 17 represents the mean temperature profile for the cases when we have radiation loss. The figure reveals an interesting feature which is also more or less pronounced in the single measurements. We see that the temperature gradient *increases* with depth down to about 6 cm, from here on it decreases and at about 16 cm it becomes equal to that in the surface layer. If the snow temperature is controlled by the conduction alone, we should have got a *continuous decrease* of the temperature gradient with the depth. As this is not the case, we here, in all probability, can trace the effect of another heat transfering agent influencing the temperature in the upper layer of the snow. We have supposed that this additional effect is caused by a *frequentation of air in the snow.*[1] This will be the subject of a more detailed discussion below.

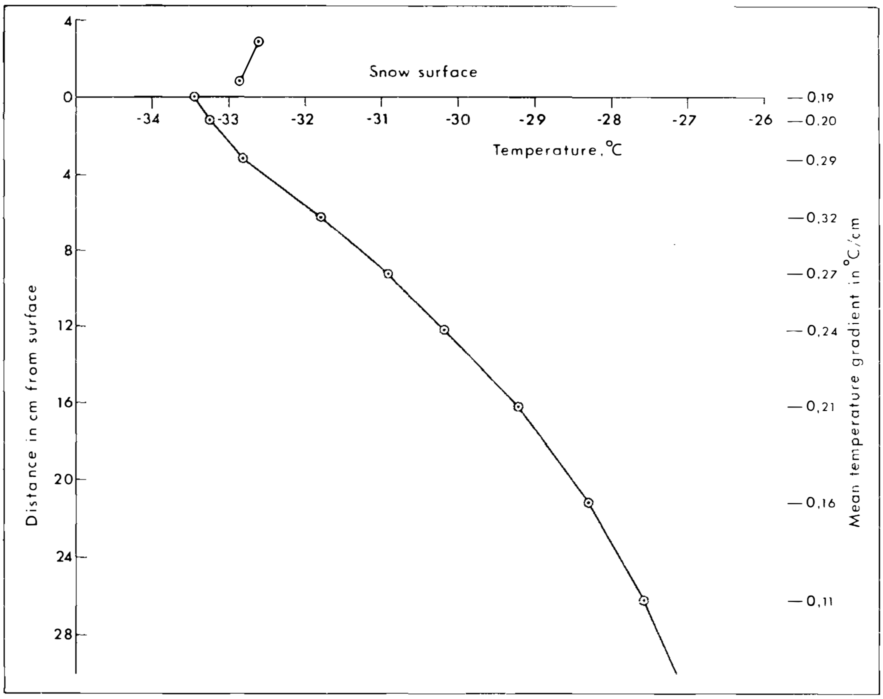

Fig. 17
The mean temperature profile for radiation loss conditions.

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[1] By air frequentation is meant an alternating in- and outflow of air.

Thermal conductivity of snow

For the calculations of the conductive transport of heat we must know the thermal conductivity, λ. This quantity is dependent upon the density of the snow, and from LUNDE (1961) we find that for the time in question the density with respect to depth in the upper layer is very nearly constant like 0.45 gr/cm^3, so for our purpose this value can be used for the whole column (0–36 cm).

As for the diffusivity of the snow, many data have been published. We will reproduce the table given by SCHYTT (1960, p. 166) for a mean density of 0.525 gr/cm^3 in the upper 10 m of the snow

Author	Diffusivity: cm^{2}sec^{-1}x10^5
M. S. VAN DUSEN	503
I. DEVAUX	785
M. JANSSON	597
H. ABELS	698
A. S. KONDRAT'EVA	866
Z. YOSIDA	445
A. WEGENER	722
V. SCHYTT	827
Mean	680

It is seen that the mean is very near the value given by ABELS. We have therefore used his expression for the thermal conductivity of snow, which is $\lambda = 0.408\ \rho^2$ cal/cm/min/deg. For the density $\rho = 0.45$ gr/cm^3 valid for our measurements we get

$$\lambda = 0.083 \text{ cal/cm/min/deg.}$$

This value is in accordance with our own measurements as revealed in the following calculations. SCHYTT mentions that his value of the thermal diffusivity is probably 25–30 % too high due to influences caused by the permeability of the snow with respect to air. In the preceeding section we found evidence of the influence of the air frequentation on the temperature profile in the upper layer of the snow. This influence is in the following section found to be traceable down to about 12 cm for the conditions under consideration. This must therefore be taken into account when evaluating the thermal conductivity. If we, for instance, consider the conductive processes alone to be responsible for the energy changes in the whole snow column, we find an extremely high value of λ, namely 0.266 cal/cm/min/deg, the sp. heat of ice taken to be 0.45 cal/gr/deg. If we, however, consider the column below 12 cm, we get the thermal conductivity of 0.081 cal/cm/min/deg, a value which is only about 2 % lower than that evaluated with the aid of ABELS' expression. Due to the short time intervals and the small quantities of energy involved in our calculations we have kept to ABELS' values.

Different energy fluxes

For the lower temperatures in question here, the radiative flux divergence in the surface layer is assumed to be of negligible importance, and the result of the following calculations will moreover show that this assumption is in all probability correct. For the simplified meteorological conditions under consideration we have to deal with three energy fluxes only. The long-wave net radiation from the surface (Q) and the transfer of sensible heat from the air (q) jointly cause the resulting net flux of heat from the snow, $\lambda \frac{\delta T}{\delta z}{}_{o}$ · Thus we have

$$\lambda\left(\frac{\delta T}{\delta z}\right)_{o} = Q + q \tag{6}$$

As Q and $\lambda\left(\frac{\delta T}{\delta z}\right)_{o}$ are known, the quantity q can easily be evaluated. The result is noted in Table 3.

Table 3

Heat loss from snow and air for different values of the net radiation

Letter ref. to Fig. 3	Net radiation Q ly/min	Heat loss from snow $\lambda\left(\frac{\delta T}{\delta z}\right)_{o}$ ly/min	Heat loss from air q ly/min	Heat left in snow from air q' ly/min	Clouds octas-type
a	-0.017	0.026	-0.009	-0.008	7 ci + as
b	-0.022	0.019	0.003	0.010	6 ci + as
c	-0.039	0.016	0.023	0.029	2 ci
d	-0.031	0.022	0.009	0.015	1 ci
e	-0.046	0.010	0.036	0.031	0
f	-0.038	0.008	0.030	0.021	1 ci
g	-0.031	0.018	0.013	0.012	4 ci + as
h	-0.030	0.021	0.009	0.016	1 ci
i	-0.020	0.037	-0.017	-0.012	3 ci
j	-0.019	0.025	-0.006	-0.001	2 ci
k	+0.020	-0.024	0.004	0.000	8 st
l	+0.020	-0.012	-0.008	0.003	8 st
m	+0.015	-0.017	0.002	-0.003	8 st
n	+0.007	-0.010	0.003	-0.008	8 st

We will now calculate the amount of heat transferred by the air frequentation in the snow. The result, besides being a control of the previously evaluated quantities, will in addition give information about the mechanism of the energy exchange between air and snow. We consider the mean change per min. of the heat content, ΔW_{0-d}, in a column with a square of one cm², extending from surface down to d cm. At the top of the column we have the loss $\lambda\left(\frac{\delta T}{\delta z}\right)_{o}$ and at the bottom the gain $\lambda\left(\frac{\delta T}{\delta z}\right)_{d}$ (when the temperature increases with depth). If the change in the heat content was due to conductive processes only, we would have

$$-\lambda\left(\frac{\delta T}{\delta z}\right)_{o} = \Delta W_{0-d} - \lambda\left(\frac{\delta T}{\delta z}\right)_{d}$$

As the quantity to the left is independent of d, the quantity to the right should also be the same and equal to the former for any value of d. We call the expression at the right hand side for $\sigma(0)$ and represent it with respect to the depth in Fig. 18. It is seen that $\sigma(0)$ differs from $\lambda\left(\frac{\delta T}{\delta z}\right)_0$ and further that it converges against a constant value at a certain depth, down to which, accordingly, another heat transferring agent must contribute to the change in the heat content of the column. This must in all probability be the frequentation of air in the permeable snow, and the magnitude of this effect, q', can be found from the equation

$$\Delta W_{0-d} + q' = \lambda\left(\frac{\delta T}{\delta z}\right)_d - \lambda\left(\frac{\delta T}{\delta z}\right)_0$$

where the left hand side should represent the energy change of the heat content in the column due to conductive fluxes only. The calculated values of q' are noted in Table 3.

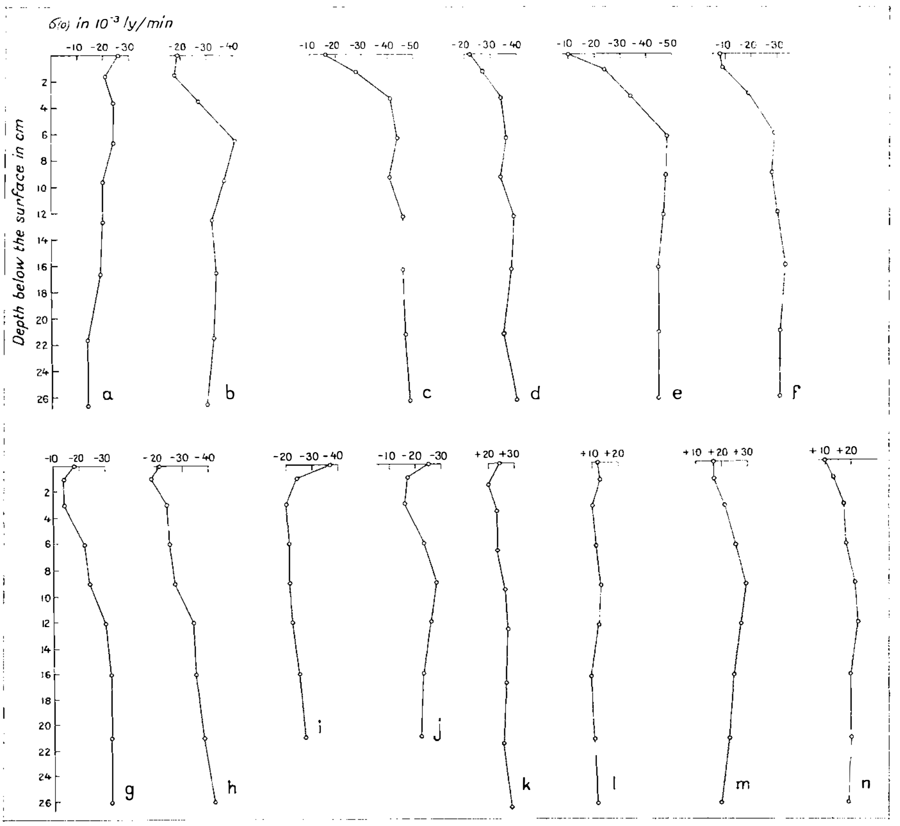

Fig. 18
The variation with depth of σ (0)

It is now of interest to compare these values with the previously evaluated heat transfer from air to snow. As the net heat loss from the snow is a common determinant for the two quantities, a direct comparison is not satisfactory. However, the relation between Q and $\sigma(0)$ can give an impartial determination of the relation between q and q'. It comes naturally, when correlating $\sigma(0)$ to Q, that $\sigma(0)$ is taken as the dependent variable. Then the regression equation simply becomes $\sigma(0) = Q$, (where the coefficient of correlation is 0.96). This result implies that $q = q'$, on an average. Our measurements thus suggest that *the heat exchange between air and snow is effected by the frequentation of air in the upper layer of the permeable snow.*

Discussion of the air frequentation in the snow

In Fig. 19 the mean variation with depth of $\sigma(0)$ is represented for the radiation loss situations. We have distinguished between whether the loss is greater or less than 0.022 ly/min as the variation of $\sigma(0)$ reveals a difference in the two cases. The probable reason for this will appear from a later discussion. It is seen that the penetration of air influences the flux in the snow down to about 12 cm for the greater radiation losses and down to about 6 cm for a radiation loss below 0.022 ly/min.[1]

We consider the radiation loss situations, which mainly represent clear weather situations (cf. Table 3). Adopting the mechanism of the heat exchange as suggested from our measurements, it is evident that the transfer of heat from the air to a high degree determines the magnitude of the loss from the snow. As the radiation loss must be covered by the loss from the air and the loss from the snow it is, thus, reasonable that there exists a relatively close dependency between q and Q, this is also fairly well demonstrated in Fig. 20. It is seen there that the relation is some-

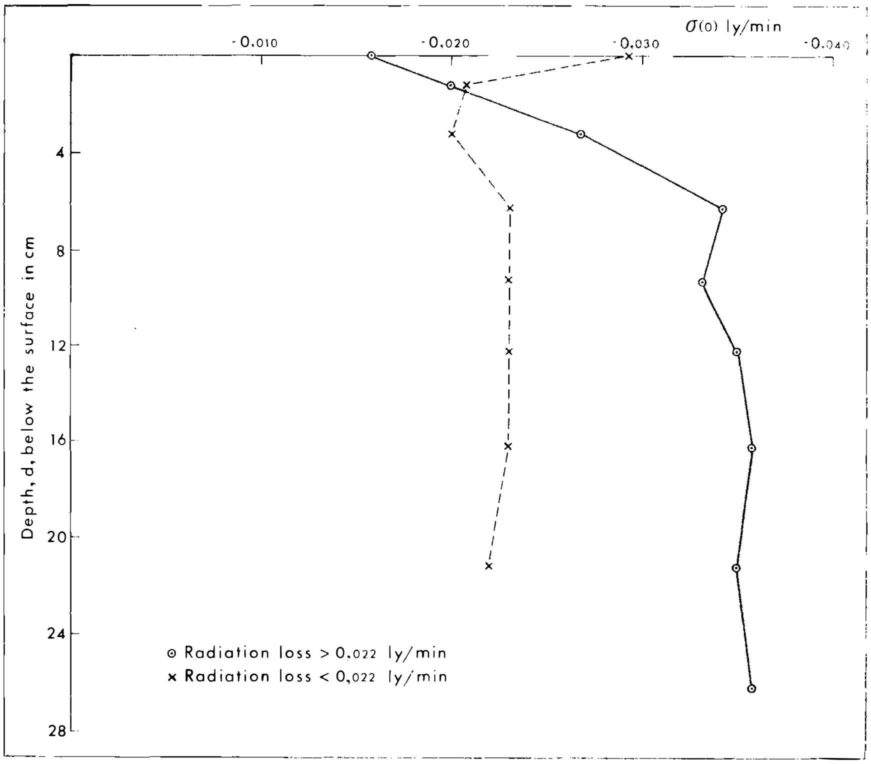

Fig. 19

The mean variation with depth of σ (0)

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[1] In all probability this effect can be traced down to greater depth for higher wind speeds or for great pressure variations.

what curved for smaller values of the radiation loss, with an approach to a linear dependency for increasing radiation loss. Due to this we have made a linear extrapolation as indicated by the broken line in the figure, and the corresponding value of q for a given Q is noted in Table 4.

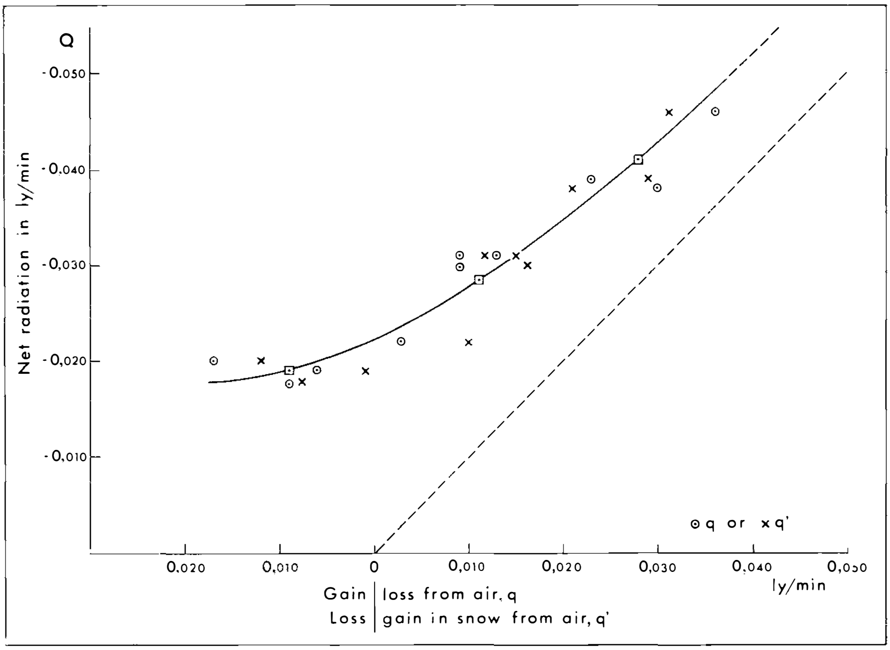

Fig. 20

The energy flux between air and snow in relation to the radiation loss.

Table 4

Corresponding values of the net radiation and the loss from the air in clear weather

Radiation loss ly/min	Energy loss from the air ly/min
0.022	0
0.025	0.005
0.030	0.014
0.035	0.021
0.040	0.027
0.050	0.039
0.060	0.050*
0.070	0.061*

* Extrapolated

With respect to intervals in the loss from the air, we have formed three group means which are marked with squares on Fig. 20. With the aid of the method of least squares the coefficients in an equation representing a curve going through these three points have been found.

$$Q = -0.022-0.45\ q-8.27\ q^2 \qquad 7)$$

This expression is valid for a radiation loss up to about 0.050 ly/min. An interesting result is immediately derived from the equation. When $Q = -0.022$ ly/min the loss from the air is zero, and consequently the whole radiation loss is covered by the loss from the snow alone. It is seen from Table 3 that the observations have been taken for mainly small cloud-amounts of cirrus and we can therefore compare the measurements treated here with the clear air observations dealt with in the previous chapters. There we saw that the radiation loss (out of 172 hourly means) was never found to be less than 0.021 ly/min. This lower limit can now be explained when one considers the equation above. We see that if the radiation loss becomes less than 0.022 ly/min, there will be a gain of energy in the air, the effective radiative temperature of the surface will then rise and consequently the net radiation. *It is therefore likely that the lower limit of the hourly mean radiation loss during clear sky and clear air conditions on the ice shelf is about 0.020 ly/min.* In this connection one notices the very abrupt fall in the occurrence of radiation loss values below c. 0.020 ly/min observed at Little America (see page 10), a station with surroundings comparable to those of Norway Station. For Little America the mist cases, are not separated so there may be a possibility that the lowest values could refer to such weather conditions.

In Fig. 21 we have indicated the expected course of the temperature variation in an air-parcel frequenting the snow for different values of the temperature gradient in the layer. It is seen that for radiation loss less than 0.022 ly/min the air is supposed to be warmed during its visit in the snow, when $Q = -0.022$ ly/min there is no change, when the loss is greater than 0.022 ly/min the air comes out colder and when $Q > 0$ the air comes out warmer. If the pattern indicated in Fig. 21 is accepted, it is easy to understand the course of $\sigma(0)$ as revealed in Fig. 19. The resulting energy transfer due to the frequentation of the air will either cause a heating or a cooling of the upper layer depending upon the magnitude of the temperature gradient in the snow.

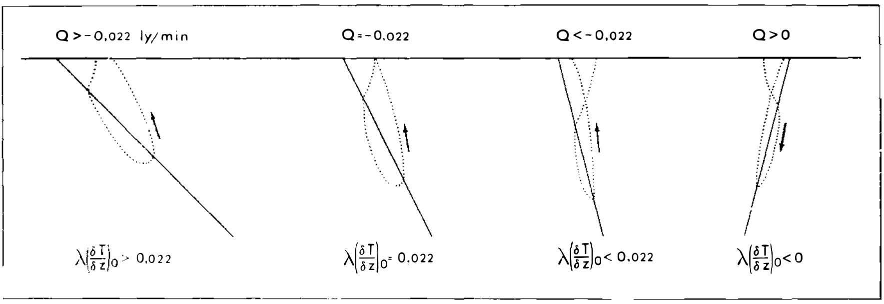

Fig. 21

The dotted curves indicate the assumed tendency of the temperature variation in an air–parcel which frequents the snow for different temperature gradients in the upper snow layer. The horizontal line can be considered as the snow surface and also as the temperature abscissa, positive to the right.

The heat loss from the air

For the hourly mean values of the radiation loss for clear sky conditions, dealt with previously (page 9), we have, with the aid of Table 4 above, found the corresponding values of the heat transfer from the air, q. Group means of q have been found with respect to intervals in the wind speed at the 10 m level and the results are given in Table 5 below.

Table 5

Heat loss from the air with respect to the wind speed during clear sky conditions

Number of obs.	Intervals m/sec	Heat loss from air ly/min	Wind speed m/sec
27	0.0- 2.0	0.018	0.9
41	2.1- 4.0	0.026	2.9
35	4.1- 6.0	0.034	4.8
26	6.1- 8.0	0.041	7.0
29	8.1-14.0	0.053	9.8

The group means of q are represented with respect to the group means of V_{10} in Fig. 22. It is seen that the mean relation is a linear one, and with the aid of the method of least squares we find

$$8) \qquad q = 0.015 + 0.0039\ V_{10}$$

Combining equation 3) page 19 and equation 8) above a *mean proportion between q and Q for the different wind speeds* can be found, viz: $p = -q/Q$. This mean relation nearly gives a straight line on a double logarithmic paper and we get

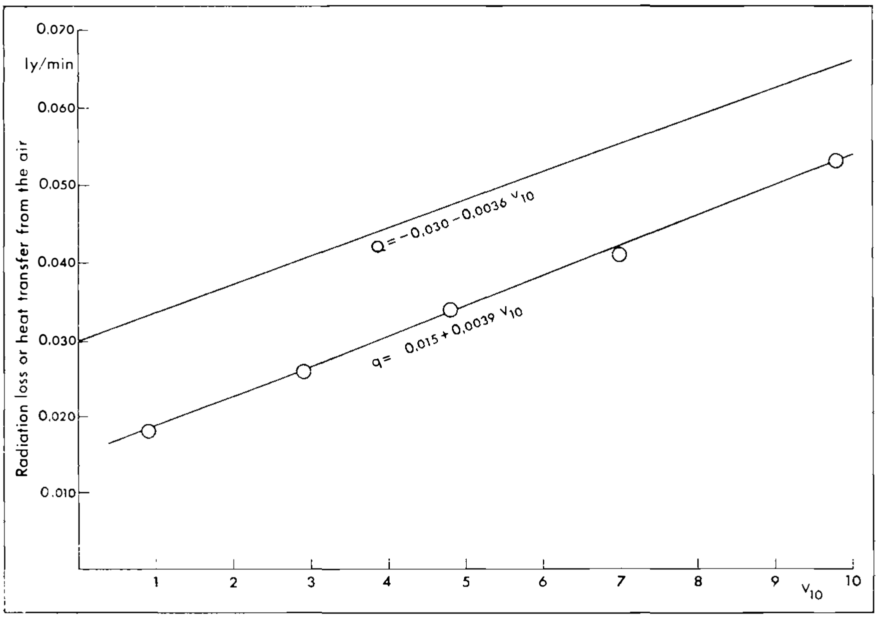

Fig. 22

The heat loss from the air in relation to the wind speed at the 10 m level.

$$9) \qquad p = 0.543\ (V_{10})^{0.176} \qquad (V_{10} \geqq 1\ \text{m/sec}).$$

The mean values of p calculated from this equation deviate less than about ± 0.5 % from those derived from the mean ratio –q/Q. The heat transfer from air to snow is then determined by the equation

10) $q = -p\,Q = -0.543\,(V_{10})^{0.176}\,Q \qquad (V_{10} \geqq 1 \text{ m/sec}).$

The calculation of q has been demonstrated in Table 6 below for the selected conditions represented in Table 1, page 16.

Table 6

Cloud cover in octas	Net radiation Q (ly/min)	Wind speed V_{10} (m/sec)	Heat transfer from the air q (ly/min)	p = – q/ Q
0	-0.047	5.0	0.034	0.72
1	-0.046	4.9	0.033	0.72
2	-0.044	5.3	0.032	0.73
3	-0.040	6.1	0.030	0.75
4	-0.041	6.4	0.031	0.75
5	-0.027	6.5	0.021	0.76
6	-0.019	7.4	0.015	0.77
7	-0.015	7.2	0.012	0.77
8	0.000	7.5		

We notice the small decrease of the heat transfer from the air for cloud covers up to 4/8, and the greater decrease for higher octas. This feature explains the curvature in the relation between the net radiation and the cloud cover as revealed by Fig. 6 page 14.

As a result of our measurements we get as seen in Table 6, that, on an average, 25 % of the radiation loss is covered by the heat loss from the snow. LUNDE (1964) gives the monthly change of the heat content in a column, extending from surface down to a depth, below which, the influence of the variation of the conditions at the surface are very small. For June and July, 1958 and 1959 at Norway Station, he gets in mean a loss which corresponds to about 17 % of the radiation loss. LUNDE'S values contains the effect of other phenomena, for instance advection and precipitation. However, as the radiation loss is a dominant heat flux, it is probable that the magnitude of the two percentages given are compareable.

The heat budget

Direct registration of the net radiation could not be accomplished due to the great difference in the sensitiveness of the two cells of the radiometer. The net radiation was therefore evaluated by calculating the different short and long-wave radiation fluxes separately. With the aid of two Moll–Gorczynski pyranometers we find the income, T, and reflected, R, short-wave radiation. The upper cell of the radiometer produces a deflection $u = u_s + u_l$, where the letters *s* and *l* refer to short and long-wave radiation. The calibration factor, k_s, of the radiometer is known, and therefore $u_s = T/k_s$. As u is measured, u_l is easily found and thereby the long-wave radiation balance for the upper cell. There is a close linear connection between the long-wave radiation loss from the upper cell and that from the surface. We have therefore found a factor of proportion, based on 150 daily sums of the two fluxes, and this has been applied in the radiation balance evaluations.

The equation of the energy budget for the snow surface is considered in the form

$$S = Q + r$$

where S is the energy change in the snow, $Q = Q_s + Q_l$, Q_s being the short-wave and Q_l the long-wave net radiation, and r represents the non-radiative heat flux necessary for balance requirements. The different components are given in Table 7. The values of S are given by Lunde (1964).

A short review for the year 1959 will be given: The total income short-wave radiation energy was 98100 ly of which 84.9 % was reflected from the surface. The annual long-wave radiation loss amounts to 15000 ly, with a monthly maximum of about 1900 ly in the summer, and a minimum of about 600 ly in the winter. The total radiation balance varies between –1000 ly/month and +1900 ly/month, while the energy change in the snow varies comparatively little, between c. –400 and +400 ly/month. This implies quite a large variation in the remainder, r, which assumes values between c. –1600 and +1100 ly/month and is consequently of the same order of magnitude as the radiation balance.

The relation between r and Q is consequently very close. The equation for the linear regression line is

11) $$r = 32 - 0.82\,Q \qquad \text{(ly/month)}.$$

Here r has been taken as the dependent variable when applying the method of least squares for the 19 monthly sums given in Table 7. It is found from the equation that the non-radiative heat flux, r, on an average amounts to 86 % of the net

Table 7

	1958							1959											
	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Jan.	Feb.	Mars	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Q_s	0	0	51	413	1084	2961	4240	3259	1721	1152	319	18	0	4	92	584	1738	2392	3518
Q_l	-939	-926	-703	-1024	-1034	-2237	-1964	-1331	-816	-1862	-927	-603	-845	-1048	-943	-1488	-1690	-1632	-1736
Q	-939	-926	-652	-611	50	724	2276	1928	905	-710	-608	-585	-845	-1044	-851	-904	48	760	1782
S	-114	-229	-99	-76	302	335	319	324	4	-210	71	-115	-404	93	-207	-142	45	278	413
r	825	697	553	535	252	-389	-1957	-1604	-901	500	679	470	441	1137	644	762	-3	-482	-1369
P	25	6	18	32	17	0	0	0	21	8	30	35	15	20	18	17	25	0	0
q	714	700	468	417	-57	-484	-1660	-1457	-711	514	461	450	592	827	642	676	-18	-560	-1333
s+a	86	-9	67	86	292	95	-351	-147	-211	-22	188	-15	-166	290	-16	69	-10	78	-36
ΔT	3.9	-7.2	9.5	5.1	9.5	-1.8	6.4	-3.4	-11.8	-4.3	16.6	7.5	-15.2	11.9	0.1	2.4	1.6	7.7	3.5
$\overline{p}$	0.76	0.76	0.72	0.68	-	0.67	0.71	0.75	0.79	0.72	0.76	0.77	0.70	0.79	0.75	0.75	-	0.74	0.75

radiation for the months with radiation loss (winter months), while for the months with a radiation gain (summer months) this flux amounts to 79 % of the net radiation. There is thus a better balance between the net radiation and the non-radiative heat fluxes for the winter months than for the summer months, and this result is in accordance with the observed relatively flat winter minimum of the temperature. (Cf. page 24.)

For Little America Hoinkes (1963) found that r amounts to 74 % of Q for the winter night, and Hanson and Rubin (1962) found 85 % for the winter half-year at the South Pole, while Lorius (1963) found 93.5 % for the winter period at Station Charcot. This percentage will probably vary somewhat from place to place because the most important of the non-radiative fluxes (the transport of sensible heat from the air) is dependent upon the windiness at the various places.

We will now form an estimate of the different components of the non-radiative heat flux for the eleven winter months. For the transfer of sensible heat an expression was found in the previous chapter, viz.:

$$q = -p\,Q = -0.543\,(V_{10})^{0.176}\,Q, \qquad (V_{10} \geq 1 \text{ m/sec}).$$

By means of this equation, we have calculated the daily values of q, applying the daily means of V_{10} and Q. The monthly sums have been noted in Table 7, where also the monthly averages of p are given. The latter factor is seen to vary between 0.68 and 0.79. It is further seen that the flux of sensible heat to the snow amounts to 89,2 % of the non-radiative heat flux to the surface. The rest, 10.8 %, is then covered by the heat flux due to sublimation, pure advection,[1] and precipitation. The latter flux, P, has been calculated as follows: For each month we have found the mean temperature, t_R, at which snow fall has been observed. Lunde (1961) gives the monthly accumulation, R, in grams per square centimeter. Then $P = cR\,(t_R - t_M)$, where c is the sp.heat of ice and t_M the monthly mean temperature at the surface. The monthly sums of P are given in Table 7 and it is seen that the total flux to the surface during the 11 winter months amounts to 224 ly. Then $q + P = 0.923\,r$, and the remaining 7.7 % (or 558 ly) of r is covered by the heat transfer to the snow caused by the sublimation, s, and pure advection, a. The monthly sums of (s+a) are given in Table 7. ($s + a = r - q - P$).

As for the heat flux due to sublimation, it has at most places in Antarctica during the winter half-year been found to be directed downwards. In some areas it is mainly directed upwards, and this has been ascribed to the drying effect of predominant foehn or dry catabatic winds. Foehn wind at Norway Station was rare, and the phenomenon was observed only once in three years. Its effect was very pronounced as regards its influence upon the direction of the flux of water vapour. The wind speed was around 14 m/sec from south-east, there was drifting snow and 7/8 of cirrus. Great variations occurred in the temperature and the relative humidity. When the former rose and the latter fell, the drifting snow *disappeared* and the snow surface developed a *glossy* appearance. As the wind speed was about the same (14 m/sec), we here in all probability saw the effect of direct evaporation. When the temperature fell and the relative humidity rose, the drifting snow started again. This sequence occurred repeatedly several times

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[1] By pure advection is meant the advection which occurs when Q = 0.

during the day. The phenomenon, well known from other places in Antarctica, is the cause of so pronouced disturbances in the temperature and in the relative humidity that the effect of the foehn wind can be discovered from the registrations, regardless of visual observation. These disturbances are, as mentioned above, very rare at Norway Station, and we may therefore expect that the net flux of latent heat for the different winter months is directed towards the surface, i. e. that the formation of hoar frost is predominant.

We denote as before the effect of the flux of latent heat by, s and the effect of pure advection by a. The monthly sum of the two effects is correlated with the respective monthly change in the temperature ΔT, the latter taken as the difference between the observed values on the first of the months at 12 h. ΔT is noted in Table 5. With the aid of the method of least squares the equation for the line of regression is found to be

$$12) \qquad \Delta T = -2.2 + 0.071\,(s+a) \qquad \text{(deg/month)}$$

$\triangle T$ is taken as the dependent variable and the coefficient of correlation becomes 0.90.

An approximate determination of s and a can be made if they can be found to be proportional to some known quantities. Like net radiation condensation is a process which *primarily* affects the snow surface. According to this it is reasonable to assume that *the total heat gain at the surface due to condensation is secondarily distributed to the air and snow with the same percentage share as is found for the radiation loss.* The *total* flux of latent heat should then become $L = s/(1-p)$. Assuming that there is a proportionality between the heat and the moisture transfer one might put $L = s/\,(1-p) = kq$, where k is a factor of proportion. Then

$$s = k\,(1-p)\,q.$$

It is seen that when $(s + a) = 0$ in equation 12) above, $\Delta T = -2.2°$ C/month. This average monthly temperature fall must be caused by the average monthly net effect of radiation (and precipitation) alone, viz.: 768 ly/month for the periods considered. This corresponds to a temperature change of $\Delta T_{Q+P} = 0.00286°$ C per ly. Due to the above emphasized assumption this figure should as well, in all probability, be valid for the effect on the temperature change, ΔT_L, of the total flux of condensation. The effect of pure advection, a, is in all likelihood proportional to the temperature change it causes, ΔT_a. Then

$$a = \varkappa\,\Delta T_a, \text{ where}$$

$$\Delta T_a = \Delta T - 0.00286\,(Q + P + kq) \text{ and then}$$

$$s + a = k\,(1-p)\,q + \varkappa\,[\Delta T - 0.00286\,(Q + P + kq)],\ \text{(ly/month)}.$$

To find the average values of k and $\varkappa$ we have summed up the figures given in Table 7 for the five months when $(s + a)$ is negative and for the six months when $(s + a)$ is positive. We get

$$-228 = 754k + (-23.1-8.3k)\varkappa$$

$$786 = 895k + (62.6-10.2k)\varkappa$$

from which we find

$$k = 0.066 \text{ and } \varkappa = 11.7 \text{ ly}/°\text{ C}.$$

The effects in the snow of condensation, and pure advection for the eleven months then become 109 and 449 ly, respectively. The *total* flux of heat due to condensation amounts to $L = kq = 426$ ly. The latter figure corresponds to a mean accumulation of 0.6 mm/month of water equivalent, when the latent heat of sublimation is 677 cal/gram.

As for the order of magnitude this result is in accordance with what is suggested from the analysis of the stake measurements at Norway Station made by LUNDE (1961). He finds that the net accumulation and ablation due to sublimation is probably less than 10 mm for the whole period of 33 months. The monthly average for a whole year derived from our calculations should be less than 0.6 mm as evaporation in all probability is predominant during the summer. Further, our estimate of the accumulation due to condensation can be compared with the calculations made by RUSIN (1959), based on the observations of humidity, air temperature and wind data. Extracting values from his table for stations which, with regard to the weather conditions, can be compared with Norway Station, there is found an accumulation for the winter months, for example, of 1.6 at Little America and 0.9 mm/month for Framheim and Cape Adare. LETTAU, DALRYMPLE and WOLLASTONE (1963) using measurements of the vertical gradient of temperature, vapour pressure and wind speed at Amundsen–Scott station, calculated the flux of latent heat under the assumption that there is similarity between heat and moisture transfer. The calculations gave a figure which on an average corresponds to an accumulation of about 0.5 mm/month of water equivalent for the winter half-year.

However, other estimations of the flux of latent heat greatly exceed our figure. We find that the *effect in the snow* of the heat flux due to sublimation is insignificant (10 ly/month) at Norway Station for the winter months. SCHYTT (1960), using LILJEQUIST's values for net radiation and flux of sensible heat, finds for the months March–Sept., at Maudheim, that the effect in the snow of the condensation is in mean 394 ly/month. HOINKES (1963), applying LILJEQUIST's formula for the transfer of sensible heat, arrives at 170 ly/month for the effect mentioned during the polar night at Little America. Further LETTAU, DALRYMPLE and WOLLASTONE (1963), using a method denoted as the "heat budget residual method", find 305 ly/month as an average for the winter half-year at Amundsen–Scott station.

As for the magnitude of the transfer of sensible heat from the air our calculations are fairly well in agreement with those obtained for the winter months at Maudheim. (LILJEQUIST (1957).) For June, July, and August we get between 470 and 830 ly/month and for the corresponding period Liljequist gets values between 750 and 810 ly/month. The methods applied at the two stations are different, they originate, however, from similar studies of the heat budget. Other calculations greatly exceed our values. Thus RUSIN (1960) estimated for Norway Station, among others, the parameters necessary for the application of the method of turbulent diffusion, and found for June, July, and August 1957 a transfer of heat ranging from 1300 to 1800 ly/month. The maximum value corresponds to a month with a comparatively high wind speed, while the other value (1300) corresponds to a mean wind speed which is about the same as in our calculations.

To complete the discussion of the annual heat budget for Norway Station it remains to consider the summer months, i. e. the months with a positive net radiation. It was previously found that the exchange of sensible heat between air and snow could be expressed by an equation (10), containing the net radiation, Q, and a function of the wind speed. The equation may be interpreted as follows: Q causes and maintains the temperature difference between the snow surface and the adjacent air, and the wind speed determines the intensity of the heat exchanging vertical motions. It is therefore probable that for the summer months we have an exchange of sensible heat by analogy to that for the winter months. However, as the snow surface (for positive net radiation) generally is warmer than the adjacent air, the flux must now be directed upwards.

According to a tentative calculation applying equation 10) for the daily evaluations it is found that the transfer of sensible heat from snow to air during the eight summer months, amounts to 6226 ly. This corresponds to 96.5 % of the non-radiative heat flux, r, and to 73.5 % of the net radiation. The average value of $p = -q/Q$ is thus 0.735 for the summer months which can be compared with 0.745 for the winter months. The rest of the non-radiative heat flux *from* the surface, 227 ly, should express the net effect of pure advection, sublimation and precipitation. The latter corresponds to a flux of 63 ly *to* the surface (see Table 7), so the effect of pure advection and sublimation alone should be a flux of $(s + a) = -290$ ly. When making the calculations for the winter months, we found that

$$\Delta T_{Q+P} + \Delta T_L = 0.00286\ (Q + P + s/(1-p)) \text{ and}$$
$$\Delta T_a = a/\varkappa = 0.0855\ a$$

As the sum of the total temperature change for the eight summer months is + 11.7° C, we get:

$$11.7 = 0.00286\ [8473 + 63 + s/\ (1-0.735)] + 0.0855\ a$$
$$\text{or} \quad -12.7 = 0.0108s + 0.0855a. \text{ Above we found}$$
$$-290 = s + a, \text{ and thus}$$
$$a = -129 \text{ ly and } s = -161 \text{ ly for the eight summer months.}$$

These tentative calculations accordingly indicate that evaporation is predominant during the summer months. By analogy with the calculations made for the winter months we now get that the total flux of heat due to the evaporation, L, is 608 ly, corresponding to a mean accumulation of about –1 mm/month of water. As $L = s/\ (1-p) = kq$, we also get that $k = 0.098$, which can be compared with $k = 0.066$, obtained for the winter months. A higher value of k for the summer months might possibly be ascribed to the higher water vapour content in the air for this season.

Combining the results of the calculations for the winter months with the tentative ones for the summer months, we get the net accumulation due to sublimation for a period of one year as –0.8 mm of water. This small value of the annual effect of the sublimation on the accumulation is in accordance with the insignificant value suggested from the analysis of the stake measurements at Norway Station. (Cf. above.)

It can be seen that a small variation in the average value of p will cause a great change in the annual accumulation due to sublimation. Thus, if the average of p for *the summer months* is 0.70 or 0.77 instead of 0.735 as found above, we get an annual net accumulation of –11 or +12 mm of water, respectively. As this result is in poor accordance with what is found probable from the stake measurements, it seems likely that the exchange of sensible heat between air and snow can be approximately determined, both summer and winter, by the equation

$$q = -0.543(V_{10})^{0.176}\ Q.$$

Table 8 shows the moving annual totals of different heat fluxes for the period 1-6-1958 to 31-12-1959.

Table 8

Q_s	15218	15218	15222	15263	15434	16088	15519	14797
Q_l	−14366	−14272	−14394	−14634	−15098	−15754	−15149	−14921
Q	852	946	828	629	336	334	370	−124
S	512	222	544	436	370	113	56	150
r	−340	−724	−284	−193	+34	-221	-314	+274

It is seen that there is a gain of heat in the snow for all the periods, and that for seven out of the eight running years this gain must be due to the positive annual radiation balance. LUNDE (1964) finds that the moving annual mean temperature in the snow between 2 and 10 metres increases steadily. According to the radiation measurements this should mainly be due to a predominance of the short-wave radiation balance in the period under consideration.

Temperature profiles in the air near the surface

With the same device as described on page 29, we made some measurements during the winter of 1958 of the temperature profile between the surface and 90 cm. Due to unexpected strange variations in the vertical temperature distributions, we thought that the measurements would give less reliable information and they were therefore abandoned. However, when making evaluations after the return from Antarctica it was found that the relatively few measurements revealed some common features which should be closer examined.

The different profiles, represented in Fig. 23, are based on the mean of two consecutive readings. In the figure are noted the wind speed at the 10 m level and at 0.43 m, $V_{0.43}$, besides the total cloudiness, cloud-type and the radiation loss. Two cases, in the figure designated with h and i, represent conditions when the radiation loss is about zero. These two cases reveal a stratification which is nearly isothermal. The deviation of the temperatures at the different heights, from their mean, is less than ±0.1° C. We therefore have a reasonable control of the similarity in the sensitiveness of the different thermo-couples, and that systematic errors have not arisen after the calibration.

All the profiles for the clear weather situations reveal a very great variation in the vertical stability below about 20 cm. Thus we find highly superadiabatic strata mingled with highly stable ones. From the figure it is seen that this feature is found for a wind speed as high as 5 m/sec at 43 cm.

In the period elapsed between the measurement of profile g and j, the weather had been overcast with low and medium clouds with a variable wind direction. The occurrence of the peculiar stratification after this period, and especially the repeated observations before, indicate that the special stratification is not caused by a temporary phenomenon but rather by one which is persistent from day to day. To show an example of the persistence over a shorter time, the two series for "profile a" have been represented in Fig. 24 left. It is seen that the two observed profiles are nearly identical. Marked similarities are also found in the other pairs of profiles; even if the temperature at the different heights varies a little the likeness in the profiles is pronounced, revealing a zigzag pattern with a minimum temperature above a maximum temperature. The graph to the right in Fig. 24 probably reveals the "short time" persistence best. The maximum and minimum temperatures, given here, refer to 30 seconds[1] variations at the different heights of the stake. This graph indicates that the profile, even when extreme

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[1] During this period more than two temperature oscillations should have occurred, (see p. 53).

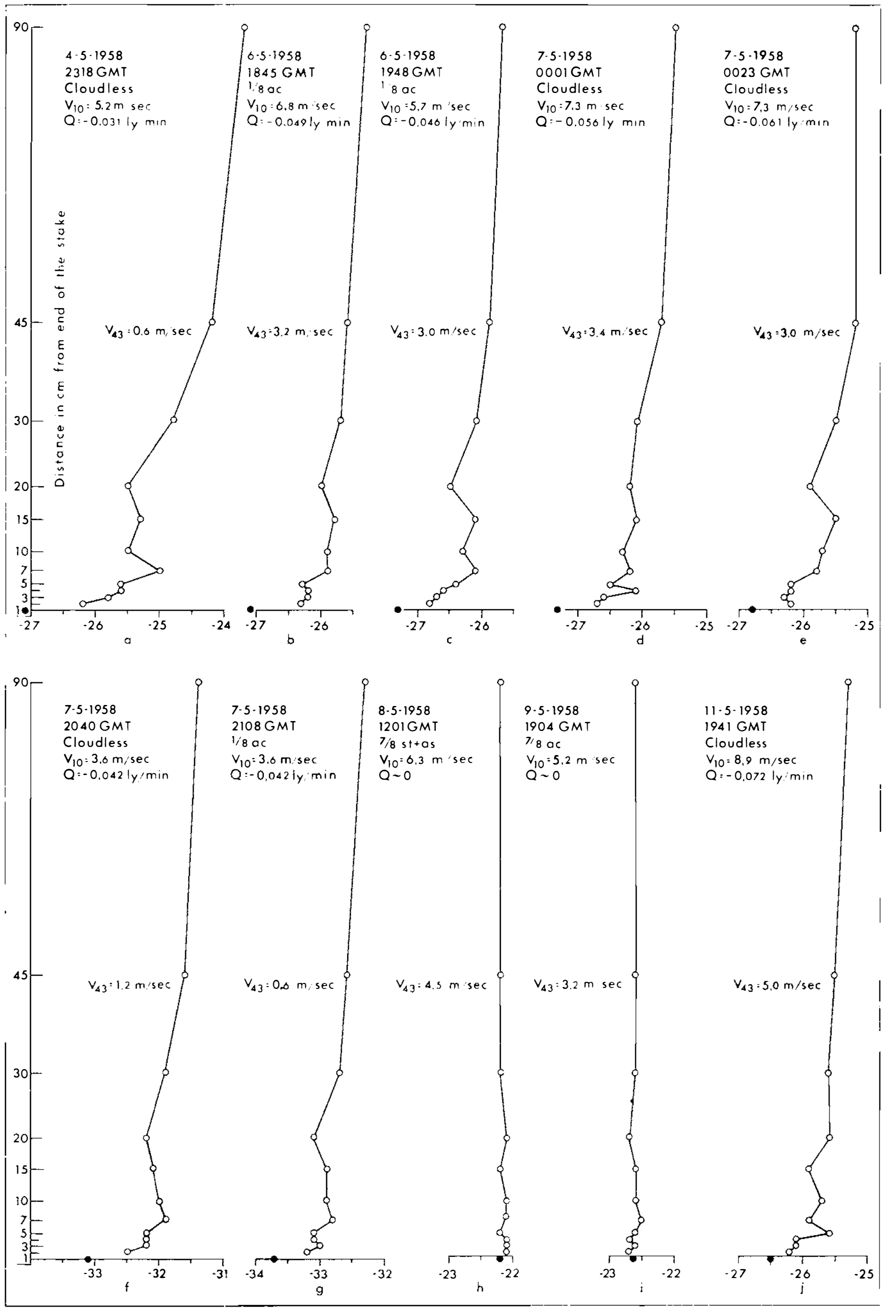

Fig. 23

Observed temperature profiles in the air adjacent to the snow when sun is below the horizon. Filled circles give the temperature at the snow surface.

temperatures are considered, does not deviate from the observed pattern common for the single measurements.

We return to Fig. 23 in order to give a closer description of the profiles (clear weather). The temperature difference between surface and the air one cm

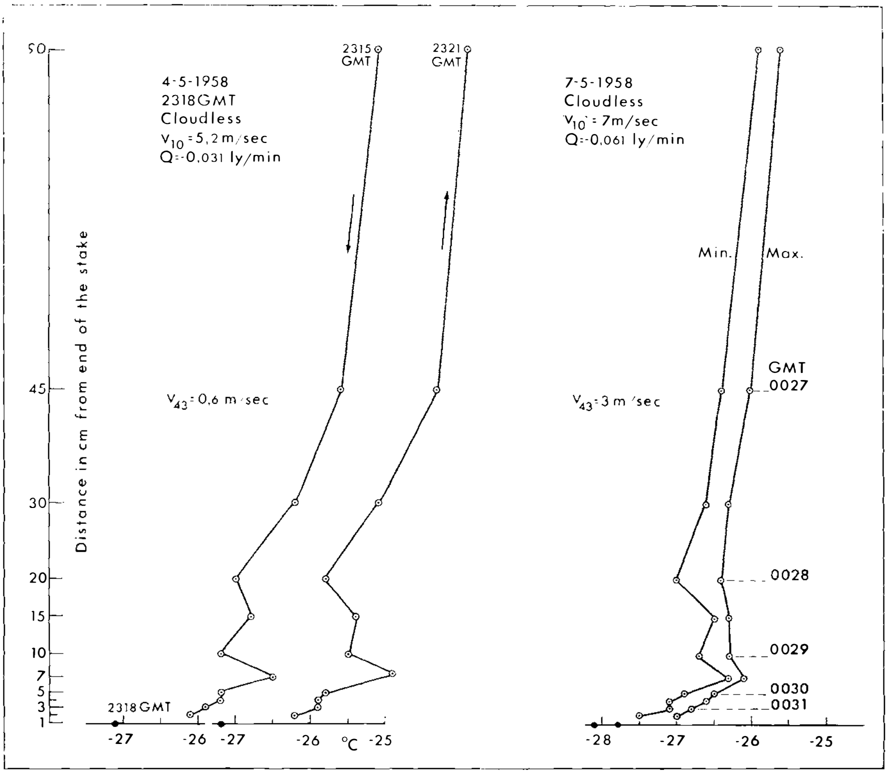

Fig. 24

To the left: Consecutive observations of the vertical temperature profile. To the right: Observation of the maximum and minimum temperature within periods of 30 seconds for different levels.

above is in mean 0.6° C. Generally, there is an increase in the temperature from near the surface up to about 6 cm. From here on and up to about 19 cm a mixing of superadiabatic and stable stratifications are observed. Above 19 cm the pattern seems to be in great contrast to that below. The mean profile, Fig. 25, based on the clear sky measurements gives the following mean temperature gradients:

Height intervals	Temperature gradients
cm	°C/m
1 - 6	+11,2
6 - 9	— 1.7
9 - 14	+ 1.6
14 - 19	— 3.4
19 - 44	+ 2.4
44 - 89	+ 0.5

It is seen that the zigzag pattern found in the single observations is retained when average temperature distribution is plotted, and this indicates that the observed profiles must be caused by a phenomenon which exercises a *systematic* effect on the vertical temperature distribution. For an explanation of the above effects we have to consider the possible influence of the radiative flux divergence.

The observed profiles differ materially from observations made in temperate

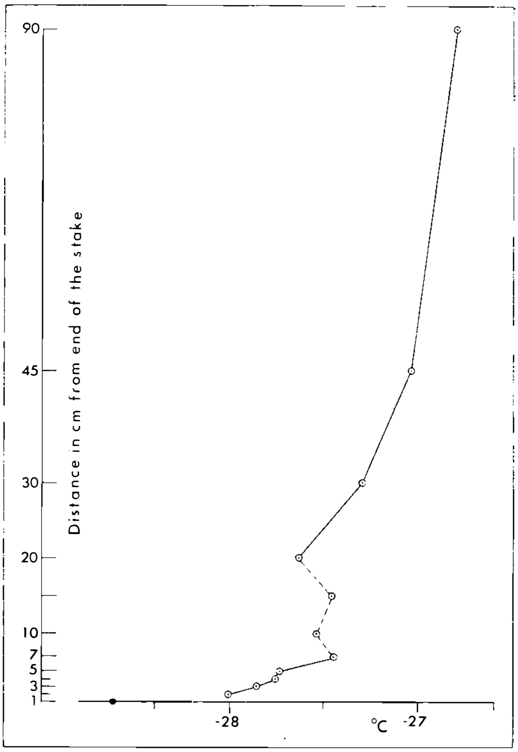

Fig. 25
Mean vertical temperature profile based on 16 readings in clear weather. Sun below the horizon.

climates where the radiative flux divergence could be of importance. Thus LAKE (1956) for instance, found the minimum temperature at heights between 4 and 15 cm above the surface of bare soil during very stable atmospheric conditions at night time. The maximum temperature was found at the surface, and the inversion started above the height at which the minimum temperature was found. GAEVSKAYA, KONDRATI'EV and YAKUSHEVSKAYA (1962) consider measurements from radiometers at two different levels and find (page 95) "that there is a great influence of radiative flux divergence on the thermal regime of the near-ground layer of the atmosphere", giving rise to temperature profiles such as those found for instance, by LAKE. The authors further point out that due to LAKE'S work and those of FLEAGLE, RASCHKE, NIILISH and MOLDAU there can be no doubt that the phenomenon is real and not caused by instrumental errors or advective processes. That the profiles found at Norway Station are in great contrast to these observations should indicate that the radiative flux divergence at the low temperatures in question (–25 to– 35° C) is of little or no significance. The special stratification here observed should therefore be due to the predominance of another phenomenon, which we have assumed to be the systematic movements probably caused by the frictional forces between air and snow. As the lowest temperature is found near the snow, cold air will repeatedly be forced upwards by the undulations on the surface, and as a compensation, warmer air from above will be forced downwards, bringing about what we might call, a frictional transport of heat. The ascending colder air should accordingly on an average cause a secondary temperature minimum at about 20 cm, and the descending warmer air, a secondary temperature maximum at about 6 cm (see Fig. 25). What kind of movement the displaced parcels go through, is of course not possible to say, even though the mean profile may indicate the temperature distribution in a vortex.

When accepting the causation outlined above, a plausible explanation can now be given for the occurrence of the absolute minimum temperature observed by others at some distance from the surface. It is obvious that momentum and moisture as well, may be transferred by the "frictional transport", so one should reasonably expect to find conformity in the vertical distribution for both temperature and wind speed. Furthermore, on condition that the surface is a source of water vapour, a zigzag pattern would probably be brought about in the vertical moisture distribution as well. One may thus expect growth of a moisture layer at some distance from the surface with a dryer layer below, and in accordance with the known effect of the radiation from a cloud layer, this moisture distribution might be favourable for the formation of a temperature profile as observed, for example, by LAKE.

It is likely that the air, with e.g. the secondary minimum temperature, has originated at a lower level, where the air shows a similar temperature. From an investigation of the clear sky profiles, represented in Fig. 23, one finds, accordingly, that the air with the secondary minimum temperature, at about 19 cm, should have originated at 5 ± 2 cm which gives a displacement of about 14 cm, and further that the air with the secondary maximum temperature, at about 6 cm, should have originated at 28 ± 2 cm revealing a displacement of about 22 cm. (In the profile j, the latter height cannot be determined due to isothermal stratification.) Our measurements thus indicate the remarkable feature that *the downward moving air primary is displaced over a longer vertical distance than is the upward moving air.*

While this paper was in print a preliminary note on the measurements made by CAISLEY, LISTER and MOLYNEUX (1963) was published. They have measured the vertical distribution of the temperature, humidity, and wind speed near the surface of a glacier in Northern Sweden in the summer of 1962. They found temperature profiles which are similar to ours, and they state that there is some similarity in the profiles of the different elements, particularly in those of wind speed and temperature.

Temperature oscillations near the snow surface

A closer examination of the deflection on the millivoltmeter (device described on page 29) revealed that the air temperature passed through small oscillations with a period of 4–12 sec. These oscillations were studied closer by placing a lens over the scale making the readings easier and more reliable. With a stop watch the time elapsing for five whole oscillations, mainly at c. 90 cm above surface, was measured in seconds. This was repeated five times so that the total number of oscillations was 25, on which a mean was based for each observation. The variation of the deflection was generally less than one scale unit, i. e. less than 0.6° C.

Some typical series of the measurements have been reproduced for different wind speeds (10 min. average at about 35 cm above the surface, $V_{0.35}$) in Table 9.

Table 9

$V_{0.35}$ 10 min. average	Time in seconds for five whole oscillations at c. 90 cm above surface	Mean period in seconds
0.6	40–57–69–50–62	11.2
3.0	47–46–31–41–45	8.4
5.0	36–34–28–50–28	7.0
7.0	27–44–29–40–44	7.4

There is a relatively great variation within each series of the number of seconds elapsed for five oscillations. This can probably, at least partly, be ascribed to the variation in the wind speed. We remember also that $V_{0.35}$ is a 10 min. average. Thus, as the series were completed within a considerably less period of time, the wind speed noted may not be representative for single measurements, but should, however, be expected to be so when considering the mean of a number of observations. As we do not possess complete wind data from the height of 35 cm, we have represented our measurements of the oscillations with respect to the wind speed at the 10 m level. (Fig. 26.) The figure reveals a marked relation between the two elements considered, thus, the higher the wind speed the greater the number of oscillations.

From Fig. 26 no (detectable) difference is revealed between clear sky observations, marked with dots, and cloudy to overcast observations, marked with crosses. However, the clear sky measurements will here be treated separately, as we intend later to compare them with measurements of other elements, taken under similar

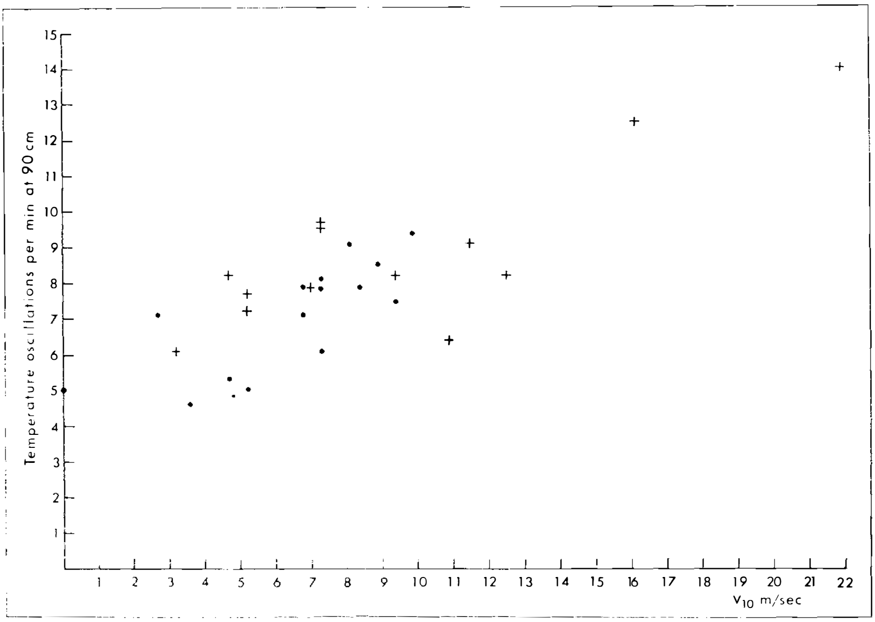

Fig. 26

Frequency of the temperature oscillations at c. 90 cm above surface in relation to the mean wind speed at 10 m. Crosses refer to cloudy to overcast conditions, while dots refer to clear weather.

conditions. The temperature oscillations at 90 cm, $N_{0.90}$, have been correlated to the wind speed at 35 cm, and also to that at the 10 m level, showing that the closest relationship exists between the two former elements. By the method of least squares, taking $N_{0.90}$ as dependent variable, we get

13) $$N_{0.90} = 5.2 + 0.60\ V_{0.35} \text{ and}$$

14) $$N_{0.90} = 4.7 + 0.37\ V_{10}$$

The correlation coefficients are 0.71 and 0.65 respectively. $N_{0.90}$ and $V_{0.35}$ are measured at two different levels, and we are interested in the value of these elements for one and the same height. To find such a relation one can use the simultaneous measurements of N at 1–4 cm and at 90 cm. We have eight observations which are represented in Fig. 27. Seven out of the eight cases reveal an increase of N with height. A closer examination of the reading series shows that in the divergent case an abnormal variation of the oscillations has taken place. If we therefore omit this observation, we get as a mean

15) $$\Delta N/\Delta Z = 1.72 \text{ osc/min/m}$$

assuming linear variation within the short height interval. Combining 15) with 13) we get

16) $N = 4.3 + 0.60\ V$, where N and V refer to the same height, 35 cm. If n is oscillations per seconds we have

17) $n = 0.072 + 0.010\ V = t^{-1}$, where t is the period in seconds.

It will now be assumed that the temperature oscillations measured are due to motions in the air with a wavy character and we will calculate $l = Vt$, which

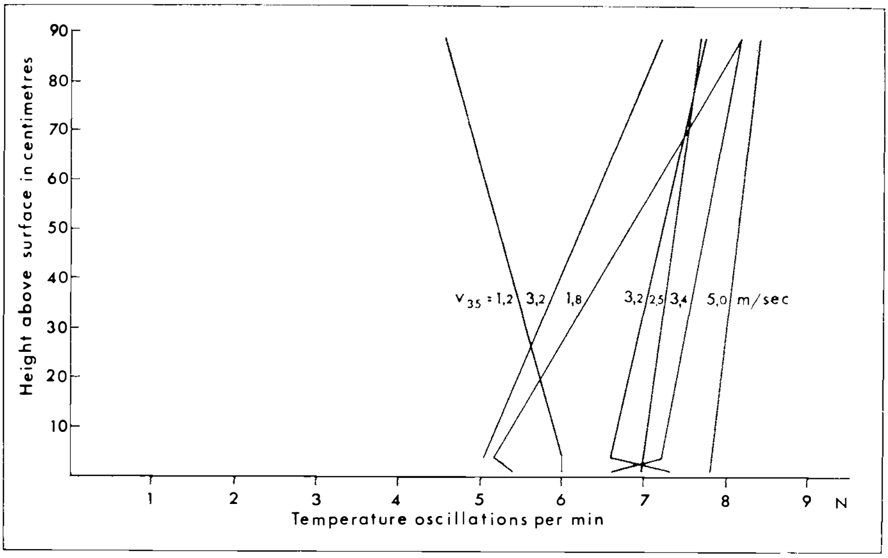

Fig. 27
The variation of the frequency of the temperature oscillations with height.

could be called the "apparent wave length". (If we have a wavy motion in the air, it is reasonable to expect that it has a certain propagation with respect to the air, so that the real wave length should differ from the apparent one.) l (in meters) has been calculated with the aid of equation 17), and the values are represented by the broken line in Fig. 28. Furthermore two single measurements with strong snow drift have been plotted in the figure to give additional information for comparison with other measurements. Barkow (1915) made registrations of the temperature oscillations for 16 days in January at Potsdam at a height of 32 meters. He calculated the "apparent wave length", the values of which are noted in the figure with crosses. We notice the very close agreement between his and our evaluations for wind speeds less than 8 m/sec. Barkow says that the tower swings considerably for strong wind. This influenced the registrations so that a special installation had to be made. The uncertainties caused by this, and moreover the fact that other observations (see below) indicate that his values for higher wind speeds (above c. 8 m/sec) are in excess, make the relevance of his measurements above this limit questionable. Barkow's measurements, for wind speed higher than 8 m/sec, have therefore been left out in the further discussion.

The close agreement (in the interval comparable) between Barkow's and our values of the "apparent wave length", for a given wind speed is suggestive for that l is a dimension which could be determined by the wind speed alone at different heights. For a closer investigation of this we have searched for measurements of temperature oscillations made for other intermediate heights over plain surfaces. This have not been possible to find. However, Giblett *et al.* (1932) in their extensive investigation upon the structure of the wind, made ultra-quick run speed records at Cardington over level country, and as there are several works which reveal a remarkable high degree of correlation between variation in the wind speed and in the temperature, Giblett's measurements can reasonably be

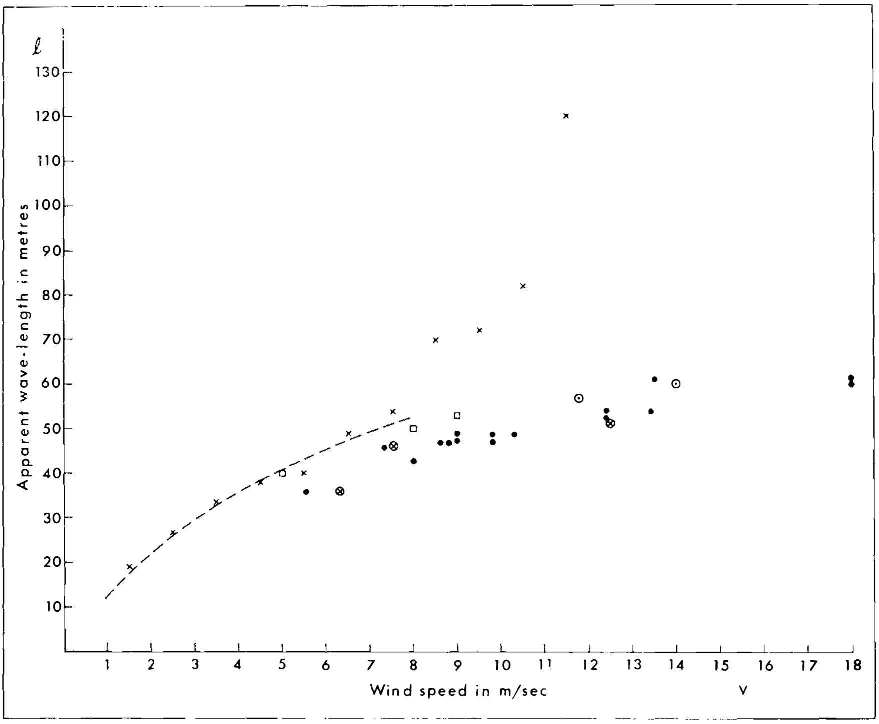

Fig. 28

Relation between the "apparent wave length" and the wind speed. – – – and ⊙ refer to 35 cm above surface and are based on our measurements of the temperature oscillations. × refer to Barkows calculations from his measurements of the temperature oscillations at 32 m. Further ⊗, ● and □ refer to 9, 15, and 45 m respectively, calculated from Giblett et al.'s registrations of the oscillations in the wind speed.

compared with ours. First we will make further references to the correlation just mentioned.

Priestly (1959, page 53) reproduces some "records obtained at Edithvale of simultaneous fluctuations of temperature and horizontal wind speed at 1.5 meters, taken on a clear day and the following clear night without change in wind speed. The remarkably high degree of correlation between the two elements of each pair of records must be due to the dominance of some connecting link – clearly, the vertical motion". Barkow finds that there is a clear tendency for simultaneous increase or decrease in both wind and temperature in the night time, when in all probability an inversion exists (January), and Scrase (1932) finds from his investigation with a bidirectional vane that there is a tendency for positive horizontal gusts to be associated with negative vertical gusts. This means that slower air goes upwards while faster moving air comes downward suggesting a tendency of simultaneously increase and decrease in temperature and wind speed. Brunt (1952 p. 218) mentions that Möller also has considered Scrase's results finding a coefficient of correlation of –0.8. Thus there should be a close interdependence between the oscillations in the temperature and in the wind speed.

From Giblett *et al.*'s ultra-quick run speed records we have considered all the cases where the oscillations in the wind speed could be distinctly determined.

(Sometimes the oscillations have been too frequent for a clear registration.) From 22 cases we have calculated the "apparent wave length", which is based on the whole time interval of the ultra-quick run records, this being mainly about 10 minutes. The result is plotted in Fig. 28, with different marks for the different heights. GIBLETT states that it is possible to see a very distinct effect of a large neighbouring obstacle on the eddying. The 22 cases,therefore, refer to occasions with a wind direction which excludes such an influencing effect. It is seen from the figure that there is no detectable systematic variation with height of the "apparent wave length" for the different heights: 9,15, and 45 m. It is further seen that there is a fairly good agreement between GIBLETT's, BARKOW's and ours measurements, which is revealed by the overlapping in Fig. 28.

It seems therefore legitimate to say that Fig. 28 indicates that in temperature and wind fields in the first 4–5 decameters over even surfaces there can exist oscillations with an "apparent wave length" between 15 and 60 meters, and that this wave length to a great extent could be determined by the wind speed at the different levels. It should be mentioned that the scale of this apparent pattern cannot be compared with the space scale for the micro-structure of the atmospheric turbulence which, according to OBUKHOV and YAGLOM (1959), for example, varies from a few centimeters to some meters.

The mean of our clear sky measurements, reproduced in Fig. 28, can be represented by the equation

$$18) \qquad l = 13\ V^{0.7} \qquad (1 \leqq V \leqq 8 \text{ m/sec})$$

and, according to the foregoing result, l and V should refer to the same height (below the first 4–5 decameters).

BEST (1937) finds that his figures on gustiness indicate that the flow of air may, at times, approach very closely to a streamline flow during inversions, and GIBLETT says that the small regular oscillations found during inversion conditions perhaps are due to wave motions. To what extent these waves are persistent in time and space is not possible to say, but that they can exist was seen twice at Norway Station. They were visualized, probably in a thin layer of mist. One could see a pattern of parallel, light bands travelling quickly and gradually vanishing. The first time no dimensions were estimated. The second time the meteorologist JARL TØNNESEN, seeing the waves running over head, estimated the wave length to be about 30 m. The author estimated from where he was standing the same dimension to be 5–10 m. However, he viewed the bands perpendicularly and they vanished before they reached the zenith. Due to the perspective the latter estimation is the less representative. LILJEQUIST (1963) mentions, page 100, in connection with the appearance of distant objects, that "a succession of wavelets could often be seen running along the upper rim of the objects, apparently in the direction of the wind".

By means of equation 18), an estimation will be made of the true wave length at the 10 m level, supposing that the period is determined by the buoyancy forces. At this height we have observations of the wind speed and of the vertical temperature gradient as mean of $(t_{20}-t_s) : 20 \sim (\partial t/\partial z)_{10}$ where t_{20} is the temperature at 20 meters and t_s at the surface. Values of $(\partial t/\partial z)_{10}$ have been found for intervals of 2 m/sec in the wind speed. The period, τ, (in seconds) for an oscillating particle due to buoyancy forces has been found with the aid of the formulae

$$\tau^2 = \frac{4\pi^2 T}{g\left(\frac{\delta t}{\delta z} - \lambda_a\right)}$$

where T is the absolute temperature, g the acceleration of gravity and λ_a the adiabatic temperature gradient. The true wave length, L, has been calculated according to the formula

$$19) \qquad L = \frac{V_{10}}{\frac{1}{t} - \frac{1}{\tau}}$$

which has been found from the following: Propagation of the wave in the air is $\upsilon = L/\tau$ and the velocity of the wave, at 10 m level, with respect to the surface is $V_x = V_{10} + \upsilon = L/t$. The calculated values have been given in Table 10.

Table 10

V_{10} m/sec	$(\delta t/\delta z)_{10}$ deg/m	τ sec	t sec	L m	υ m/sec
1	0.60	41	14.6	23	0.6
2	0.57	42	11.9	33	0.8
3	0.53	43	10.0	39	0.9
4	0.50	44	8.7	43	1.0
5	0.46	45	7.6	46	1.0
6	0.43	47	6.8	48	1.0
7	0.39	49	6.2	49	1.0
8	0.35	52	5.6	51	1.0

It is seen from the table that under the assumptions made we find wave lengths of the same order of magnitude as have been observed. Agreement with GIBLETT's measurements is moreover revealed since our calculated propagation with respect to the air is also very little for the higher wind speeds (about 1 m/sec). GIBLETT *et al.* (1932, page 46) find results which "agree well with the supposition that there was a wind pattern travelling with a velocity only very slightly greater than the mean wind at the height of the anemometers".

It is interesting in this connection to learn about D. W. S. LIMBERT's[1] measurements of the wave length of the undulations *perpendicular* to the prevailing wind direction on the ice shelf at Halley Bay in Antarctica. This wave length was in mean 44 m, and thus of the same order as that found in the air-flow over the ice shelf at Norway Station.

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[1] Private communication.

Summary

The frequency distribution for the net long-wave radiation has been considered for some coastal and interior stations in Antarctica. Along the coast values are found within the limits of +0.040 and -0.110 ly/min, when the sun is below the horizon. In the interior these limits are +0.010 and -0.060 ly/min. Except for the high plateau, the frequency distribution of the net radiation is bimodal, the one mode having a small negative value, and the other one having a value varying between -0.025 and -0.080 ly/min. Of the stations considered, Mirnyj has by far the highest radiation loss, which accords with the prevalence of the foehn wind at this place. Little America displays the highest positive values. This may be due to the higher temperature contrast between surface and advective clouds attainable at this coast station lying very far south. The single mode found in the frequency distribution for the stations on the interior plateau (3–4000 m a.s.l.) reveals the relatively little importance of a clouding-over at this height.

A close relation between the atmospheric counter radiation, A, and the amount of low or/and medium clouds, N_c, at Norway Station can be expressed by the equation

$$A = 0.223 + 0.0138\ N_c$$

A mean relation between the surface temperature and the radiation loss for cloudless sky is found to be non-linear. It is possible that this curvature can be caused by draining of the atmosphere due to the precipitation of ice needles. This occurs more frequently for falling temperature, especially below -35° C.

The net radiation is found to be dependent upon the wind speed. This is just another effect of the heating of the surface due to the transport of sensible heat caused by the wind. Comparisons made with Liljequist's measurements of the net radiation taken with an uncovered pyrgeometer reveal great differences. Calm weather observations, however, were on an average equal, indicating similarity in the calibration factors.

For average equilibrium conditions there is found a simple relation between the temperature of the radiometer, t_i, and the radiative temperature, t_a, of the atmosphere, viz:

$$t_i = 11.6 + t_a$$

This equation holds for any mean value of t_i between about -10 and -45° C.

Indication of subsidence during winter time are found. The clear atmosphere is found to have a higher radiative temperature in July and August than in June and September.

Evaluation of the atmospheric long-wave radiation from the Elsasser radiation chart (1942) reveals that our measured values are higher. It has been supposed that radiation from the ozone in the "window" of the water vapour may contribute to the difference.

Measurements performed during equilibrium conditions in the polar night indicate that the transfer of sensible heat to the snow is effected by air frequentation in the upper layer of the snow. The transfer of heat is calculated in two different ways and the agreement is fairly good. According to our measurements the flux of sensible heat can be expressed by the equation $q = -0.543 (V_{10})^{0.176} Q$, where Q is the net long-wave radiation and V_{10} the wind speed at the 10 m level.

Considering the heat budget of the surface, one finds that the total income of the short-wave radiation is 98100 ly/year for 1959, and that the annual mean albedo is 0.849. The long-wave radiation loss amounts to 15000 ly/year, reaching a monthly sum in the summer which is three times greater than the lowest monthly sum in the winter. The best balance between the net radiation and the non-radiative heat fluxes is found during the winter period, this is in accordance with the relatively flat winter minimum of the temperature. It is found probable that the exchange of sensible heat between air and snow, both summer and winter, can be expressed approximately by the equation given above. According to this the transfer of sensible heat to or from the surface amounts, on an average, to 73.5 and 74.5 % of the net radiation for summer and winter months, respectively. Estimation of the effect of sublimation indicates an insignificant annual accumulation, in accordance with what is suggested by the stake measurements. The positive balance of the snow during the period considered is found to be due to a predominance of short-wave radiation.

A zigzag pattern is observed in the temperature profile between 5 and 20 cm above the surface, and this in all probability discloses the existence of systematized vertical movements in this layer. A probable consequence of this is that a zigzag pattern should also be brought about in the vertical distribution of wind speed, and in humidity as well, if the surface is a source of water vapour.

Periods of 4–12 seconds have been found for the small-scale oscillations in the temperature near the surface. The period is dependent upon the wind speed, V, and assuming that the oscillation is due to wavy motions in the air, we find that the "apparent wave-length", can be expressed by the equation

$$l = 13\, V^{0.7} \qquad (1 \leqq V \leqq 8 \text{ m/sec}).$$

From comparison with other observations it is probable that the equation expresses the order of magnitude of the "apparent wave-length" in the first 4–5 decameters above the surface. Assuming that the motions are caused by buoyancy forces, we have calculated wave-lengths between 25 and 50 meters, and this is of the same order as what has been observed.

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