NORSK POLARINSTITUTT
SKRIFTER NR. 148

DEN NORSKE ANTARKTISEKSPEDISJONEN, 1956-60
SCIENTIFIC RESULTS NO. 10

TORGNY E. VINJE

# The turbulent transfer over an Antarctic ice shelf


NORSK POLARINSTITUTT
OSLO 1969

DET KONGELIGE DEPARTEMENT FOR INDUSTRI OG HÅNDVERK

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# Contents

## Abstract

Temperature and wind profiles obtained from a c. 40 m high tower on the ice shelf at Norway Station are discussed. Great dissimilarities are found in the profiles in the upper part of the tower, where a maximum in the wind speed at about 30 m occurs quite frequently. This suggests also that the shearing stress may decrease considerably within the lowest three decameters over the ice shelf. The shape of the temperature profile is found to change markedly as the stability increases. And the shape of both temperature and wind speed profile is found to be different for easterly contra westerly wind directions for otherwise similar conditions. For isothermal conditions a power profile represents the wind speed observations better than a logarithmic profile.

A vortex theory on the turbulent transfer is proposed. The theory suggests that the Kármán constant may be determined from the power profile exponent and further that it decreases with increasing stability. At the average height of the center of the surface vortices, it is found that the drag coefficient is determined by the profile exponent alone.

Besides offering a possible explanation of an observed zigzag profile near the surface, the turbulent properties calculated according to the vortex theory are in fair accordance with other calculations. The mean vertical extension of the surface vortices is, for small wind speeds, found to be of the same magnitude as the mean amplitude of the unevennesses on the surface (20 cm).

It is found that the onset of drifting snow causes a relative reduction in the drag of the surface on the stream. The calculated drag coefficient shows a maximum value when the profile exponent is about 0.25–0.30. And this feature is in accordance with a corresponding variation of the temperature gradients.

## Introduction

A knowledge of the wind and temperature profiles is of the greatest importance for the understanding of the turbulent transfer processes in the surface layer. In order to study these phenomena a tower was built at Norway Station, 70°30′S, 02°32′W, during the autumn of 1957 and observations were collected in the period 7.III, 1957 to 18.XII, 1959.

Norway Station is situated on an extensive ice shelf about 35 km south of the barrier. The closest ice rise starts at about eight km from the station in a north-westerly direction. In all the other directions the monotony of the ice shelf is unbroken over great distances. We had two predominant wind directions, a primary from south-east and a secondary from south-west. The distance to the nearest change in the topography in these directions was about 150 and 25 km respectively, and the influence of neighbouring obstacles on the wind and temperature profiles at Norway Station is thus in all probability non-existent.

To reduce the influence from the tower on the measurements, the measuring devices were mounted on a bar about 1.5 m outside the tower on its southern side.

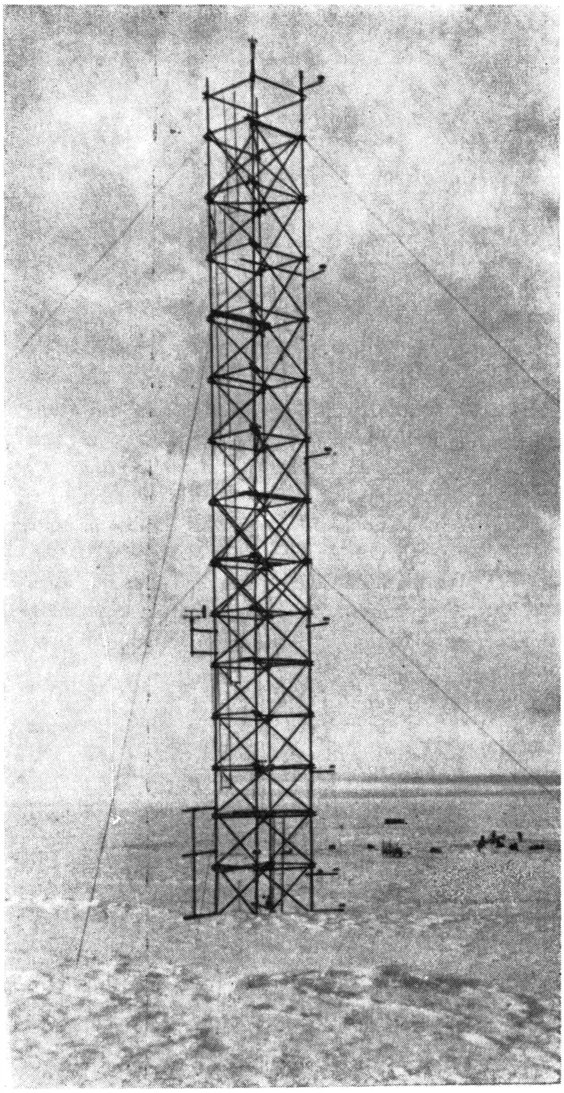

Fig. 1. *The meteorological tower seen towards NW.*

## Instruments

THERMOMETERS

The temperature was measured with platinum resistance thermometers which were connected with two Siemens multi-color point-printers and a series of accumulators producing a voltage of 24 V. Each color printer had six circuits and the thermometers were connected every other minute. During the mounting of the thermometers adjustments had to be made because of the possible difference in resistance of the different circuits. This was accomplished by varying a special resistance in each circuit, so that the deflection was zero when the current was switched on. These corrections were correct within $\pm 0.1$°C.

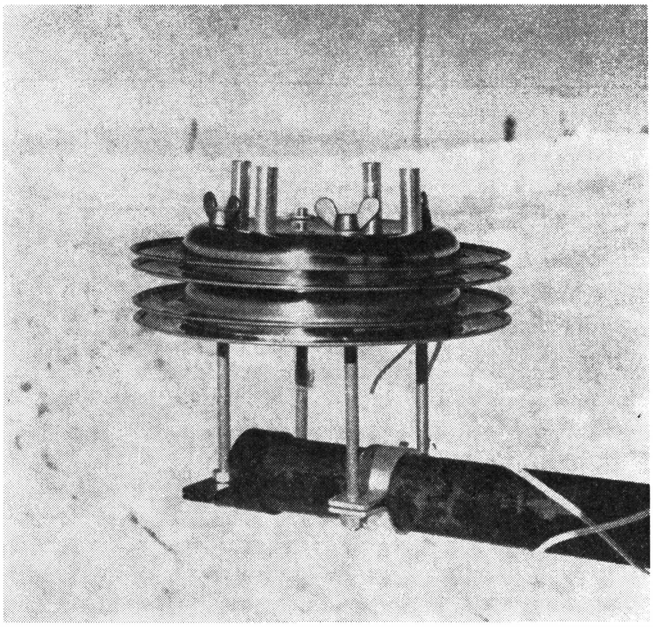

Fig. 2. *The radiation shield used for the thermometers.*

SOURCES OF ERROR

The most important source of error was assumed to be the radiation. To reduce this effect, the resistance thermometers were mounted within polished radiation shields made of stainless steel and formed as plates (Fig. 2). The radiation shields were designed by Mr. V. HISDAL at Norsk Polarinstitutt.

It was expected that during small natural ventilation, heated air could gather between the plates and to avoid this the upper plates had been furnished with four pipes as shown in the figure. The shielding system turned out to be very effective except for high insolation during complete calm weather.

Table 1 below is based on 57 readings of the temperature in the tower at screen level for isothermal conditions. The difference between this reading and the aspirated psykrometer temperature has been taken as the radiation error ($\Delta T$). The global radiation, G, is given in ly min$^{-1}$.

Table 1

| Wind speed intervals (kts) | Mean wind speed (m sec$^{-1}$) | $\Delta T/G$ | Number of observations |
|---|---|---|---|
| calm | 0 | 12.1 | 3 |
| 3– 5 | 2.1 | 0.17 | 9 |
| 6–10 | 4.1 | –0.005 | 20 |
| 11–31 | 9.1 | –0.072 | 25 |

The negative proportion of ($\Delta T/G$) for higher wind speeds must be caused by the fact that the thermometers are correct within $\pm 0.1°C$, thus $\Delta T$ gets a small negative value instead of a small positive one or zero. As $G = 1$ ly min$^{-1}$ is about the average maximum value of the global radiation during the summer season, the table gives information of the magnitude of the maximum radiation error. From the above it can be seen that the registrations are greatly affected when there is a high insolation during calm weather. This happened, however, only about 5 times during the summer seasons of the period March 1957–Dec. 1959, and lasted between 5 to 15 minutes only.

It turned out that the most important error was caused by the fact that the switches in the recorders were liable to corrosion, which could cause a change in the contact resistance. This effect sometimes caused small displacements of the registration curves of a few tenths of a degree. Moreover, the two recorders we had in operation were claimed by the makers to show a correct temperature with an accuracy of $\pm 0.7°C$. A relative correction had therefore also to be made. The absolute correction is of no importance for the discussion in this paper, and has therefore not been considered. The necessary relative corrections have been made according to the following: for isothermal conditions, i.e. generally for high wind speeds and overcast weather, we have read the recorded temperature at the different levels. The deviation of the different recordings from the mean of the readings has then been applied as a correction. This was made 3–4 times every month. A typical example of the relative corrections applied is given in Table 2 and refers to the year 1958. The corrections are given in a tenth of a degree for the different heights. We note the marked change in the corrections at about 15.IV, which occurred in connection with the cleaning of the switches.

Table 2

| Period | Temperature corrections (°C × 10) | | | | | | | | Remarks |
|---|---|---|---|---|---|---|---|---|---|
| 1–31.I | − 3 | +2 | 0 | 0 | −2 | | | | |
| 1–28.II | − 4 | +2 | −1 | | | | | | |
| 1–11.III | − 7 | −2 | +3 | +1 | | | | | |
| 12–31.III | − 7 | −2 | +3 | − 1 | −1 | +1 | | | Tower extended |
| 1– 9.IV | − 8 | −5 | 0 | +5 | +1 | +1 | +1 | +4 | |
| 10–15.IV | − 5 | −1 | −1 | +2 | +1 | +3 | +1 | +1 | |
| 16–30.IV | − 7 | 0 | 0 | 0 | 0 | 4 | 0 | 0 | Switches cleaned |
| 1–31.V | −10 | 0 | 0 | 0 | 0 | 3 | 0 | 0 | |
| 1.VI–31.VII | − 3 | 0 | 0 | 0 | 0 | 5 | 0 | 0 | |
| 1–30,VIII | − 1 | 0 | 0 | 0 | 0 | 5 | 0 | 0 | |
| 1.IX–31.X | − 2 | 0 | 0 | 0 | 0 | 4 | 0 | 0 | |
| 1.XI–31.XII | − 2 | 0 | 0 | 0 | 0 | 6 | 0 | 0 | |
| Level No. | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | |

## ANEMOMETERS

The anemometers were of Norwegian design and had been constructed at the Meteorological Institute, Oslo. They were specially designed to stand high wind speeds and were of the same type as those used at Maudheim. They have been described by Langlo (1952).

The anemometers were connected with a so-called wind gradient recorder,

Table 3

*Calibration values of the MI anemometers, m sec⁻¹.*

I: Calibration before the expedition at NTH 1956.
II: —»— —»— Lista 1956.
III: —»— after —»— RID 1960.

| Number of contacts | MI 217 I | MI 217 II | MI 217 III | MI 218 I | MI 218 II | MI 218 III | MI 219 I | MI 219 II | MI 219 III | MI 223 I | MI 223 II | MI 223 III | MI 523 I | MI 523 II | MI 523 III | MI 524 I | MI 524 II | MI 524 III | Mean |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 0 | 1.0 | 1.0 | 1.0 | 1.0 | 1.0 | 0.7 | 1.3 | 1.0 | 1.0 | 1.0 | 1.0 | 0.5 | 1.1 | 0.7 | 1.1 | 1.2 | 0.9 | 0.9 | 1.0 |
| 100 | 7.5 | 7.5 | 7.0 | 7.6 | 7.3 | 7.0 | 7.7 | 7.2 | 7.0 | 7.5 | 7.2 | 6.5 | 7.6 | 7.2 | 7.2 | 7.7 | 7.3 | 6.9 | 7.3 |
| 200 | 14.2 | 14.0 | 13.1 | 14.2 | 13.4 | 13.5 | 14.2 | 13.5 | 13.0 | 14.1 | 13.4 | 12.6 | 14.1 | 13.7 | 13.3 | 14.1 | 13.7 | 13.0 | 13.6 |
| 300 | 20.8 | 20.5 | 19.1 | 20.8 | 19.7 | 19.7 | 20.7 | 19.6 | 19.1 | 20.7 | 19.5 | 18.7 | 20.7 | 20.2 | 19.4 | 20.6 | 20.1 | 19.1 | 19.9 |
| 400 | 27.4 | 27.1 | 25.2 | 27.4 | 25.9 | 26.1 | 27.1 | 25.8 | 25.2 | 27.2 | 25.7 | 24.7 | 27.2 | 26.8 | 25.5 | 27.0 | 26.6 | 25.1 | 26.3 |

also built at the Meteorological Institute. The anemometers made a contact per 250 or 500 m wind way, and the number of these contacts were summed and printed by the wind recorder, per 10 minutes until 4.VII, 1958, and after that per 20 minutes. The recording device was a prototype and was liable to some technical faults for short periods. The series of wind observations are therefore not so complete as is the case with the temperature observations.

In each circuit the recorder was equipped with a control lamp which made a signal for every contact produced by the anemometers. Due to the dense drifting snow near the surface and because of the hoar frost formation, the contact system in the lowest anemometer was sometimes affected. This resulted in multiple signals on the control lamp. The registrations from the affected height could not be corrected and had to be abandoned. However, in spite of the troubles mentioned, we have obtained wind registrations for some quite long periods which enable us to get a fairly reliable impression of the vertical wind profile under different weather conditions.

## CALIBRATION

We had ten anemometers at Norway Station. Before the expedition (1956) they were compared at two places: in a wind tunnel at the Norwegian Technical University (NTH) in Trondheim and over level country at Lista in the south-western part of Norway. Considering the comparisons made at NTH in 1956, it was found that for a given number of contacts the corresponding wind speeds deviate from the mean by less than one per cent at about 30 m sec$^{-1}$, and by less than six percent at about 5 m sec$^{-1}$ for all ten anemometers. During the expedition, four anemometers ceased to function and, in 1960, the remaining six were re-calibrated in a wind tunnel at the Norwegian Research Institute of the Defence Department (RID). Table 3 gives the calibration values obtained by reading the calibration curves at suitable intervals. It can be seen that the comparison made at Lista (1956) and that made at RID (1960) generally shows a somewhat lower wind speed compared with the values obtained at NTH (1956) for a given number of contacts. It is supposed that these differences are due to dissimilarities in the mounting during the different calibrations, and we have therefore used the mean of all the calibrations for our anemometers, viz.: $U = 1.0 + 0.38\,n$ (m $sec^{-1}$), where n is the number of contacts per 10 minutes. As can be seen the anemometers, due to inertia, do not register wind speed less than about 1 m $sec^{-1}$.

### ALTITUDE OF THE INSTRUMENTS

The lowest instruments had to be moved now and then due to the fairly high accumulation of snow which was about 120 cm per year according to LUNDE (1961, p. 12). The higher instruments, however, had fixed positions before and after a rearrangement in March 1958 when the tower was extended from about 24 to about 37 metres. The accumulation was read on a stake at the tower. The change of the height of the instruments above the surface could be quite great during a storm with snow fall. During other periods with relatively small accumulation, new heights were noted when the accumulation had increased about 10 cm.

The accumulation read on the reference stake at the tower depends upon the causal position of the stake in the pattern of the ondulations on the surface. From two series of measurements, c. 21 months apart, on 36 stakes covering an area of 37 km² near Norway Station, LUNDE (1961, p. 22) found a standard deviation in the accumulation of 10 cm with respect to what we might call the zero plane level. Later on, when considering groups of profiles, the observation heights will refer to this zero plane level. In Table 4 is given an example of the mountings of the instruments at 3.V, 1958.

Table 4

| | Metres | | | | | | |
|---|---|---|---|---|---|---|---|
| Height of anemometers | 0.45 | 3.31 | 7.60 | 12.58 | 19.44 | 28.12 | 37.40 |
| —»— thermometers | 0.45 | 3.31 | 7.60 | 12.58 | 23.56 | 28.12 | 37.40 |

## Some features of the inversion

### CHARACTERISTIC EXAMPLES OF THE REGISTRATIONS

Figs. 3–9 are based on temperature and wind registrations at different levels in the tower. The net radiation at about 1.5 m above the surface is also given. The temperatures, as well as the net radiation, were registered every other minute while the wind speed was given by a 10 or 20 minute average.

Fig. 3 illustrates the close relation between the net radiation and the formation,

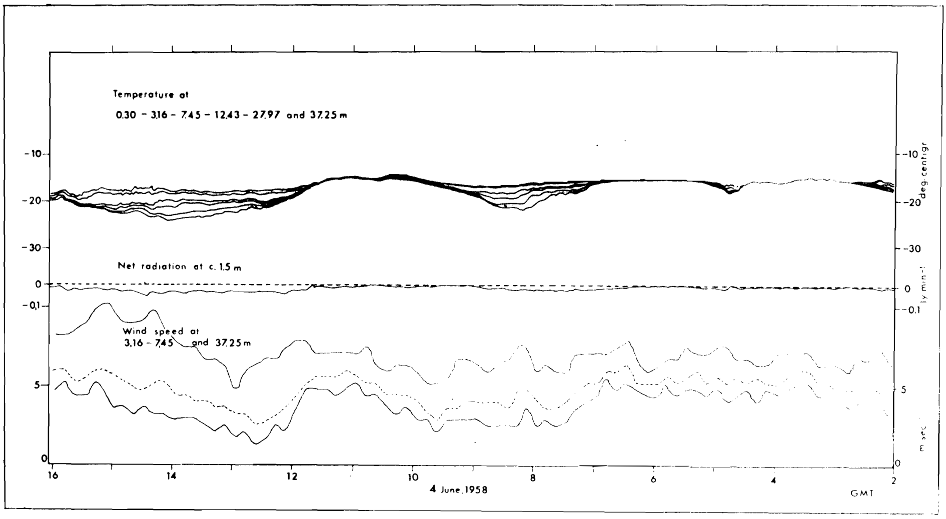


Fig. 3. *Registrations of temperature, net radiation, and wind speed.*

maintenance, and disappearance of the inversion. The figure indicates also quite clearly the time lag with height of the occurrence of the minimum temperature. The magnitude of the temperature variation with time varies considerably with height, and for the short-period inversions given in the figure no near stationary state is arrived.

Fig. 4 may indicate that the near stationary state is achieved about two or three hours after the building of the inversion has started. It is remarked that the temperature at the top of the tower shows a tendency to increase during the first hours. This feature was observed several times and is therefore in all probability not accidental. It can be seen that the wind speed at about 7 m very often exceeds that at about 38 m for the time interval in question.

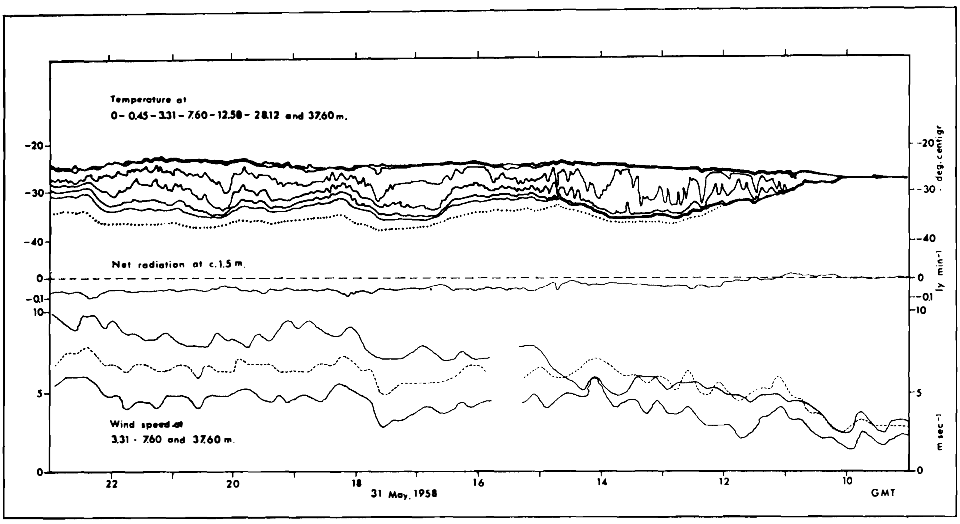


Fig. 4. *Registrations of temperature, net radiation, and wind speed.*

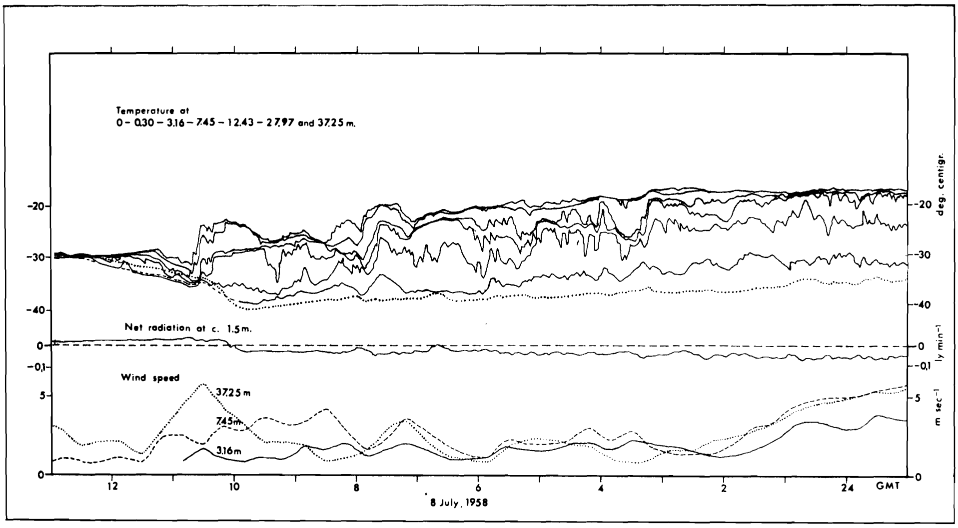


Fig. 5. *Registrations of temperature, net radiation, and wind speed.*

As for the temperature fluctuations, it is remarked that for the first 3—4 hours after the inversion has appeared, great variations may go on at certain heights with no or small effect on the temperature variations at the adjacent heights. Later in the period, at about 1400 GMT, it seems that variations go on simultaneously at several heights. In some cases similar falls or rises in the temperature can be traced from near the surface and up to about 12 m while the temperature above, about 28 m, is relatively constant.

Fig. 5 shows how an inversion is destroyed when the net radiation becomes positive. As the wind speed here is about half as high as in Fig. 3, the extinguishing of the vertical temperature differences goes on less rapidly. We note the close connection between the variation of the surface temperature and the net radiation. For about three hours the stratification near the surface is highly superadiabatic, while an inversion of considerable magnitude exists at higher altitudes. It is seen that in the last half of the period temperature variations of considerable magnitude occur more or less simultaneously at all heights. Except for the three last hours of the period, the wind speed at 37 m is generally less than that at 7 m and occasionally also lower than that at 3 m. This reveals the existence of a great dissimilarity in the form of the temperature and wind profile in the upper part of the tower.

Fig. 6 can serve as an demonstration of how the temperature stratification changes with wind speed while the net radiation shows no detectable systematic variation. We see the change of the stability towards the end of the period in connection with a considerable decrease of the wind speed at all heights. The vertical temperature gradient above 7 m decreases while below this height it is seen to increase considerably. It is also noted that as the stability near the surface increases, the wind speed at the top of the tower decreases and becomes, at the end of the period, less than that at 3.3 m. It can further be seen that the temperature variations with a period, ranging from 5 to 15 minutes, occurs at all heights

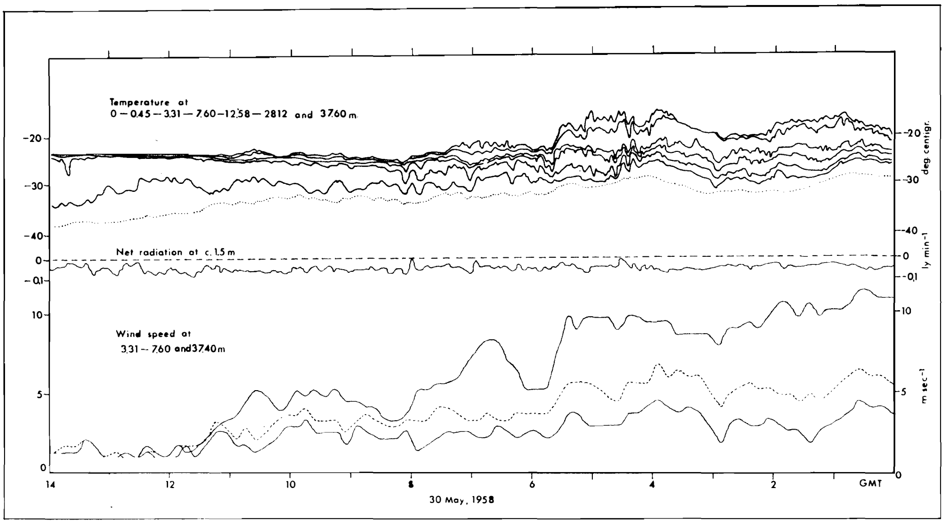


Fig. 6. *Registrations of temperature, net radiation, and wind speed.*

in the first half part of the period and sometimes simultaneously at all heights. The variations at the surface are relatively damped.

Fig. 7 reveals that pronounced variations of the wind speed at the top of the tower may go on with no corresponding variation at lower levels. At 0245 GMT there occurs a maximum in the wind speed within the inversion which prevails about three hours. This feature is accompanied with an increase in the difference between the temperature gradients in the upper and lower part of the tower. The temperature variations at the different heights are more or less of the same kind as found in the previous graphs. After about 1000 GMT, the wind gradient at the lower levels increases considerably with a relatively far smaller increase in the temperature gradient.

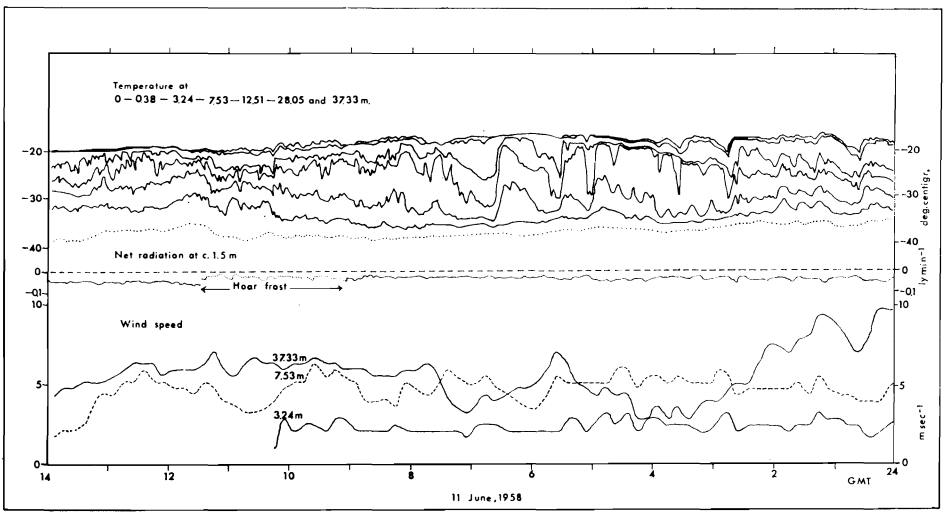


Fig. 7. *Registrations of temperature, net radiation, and wind speed.*

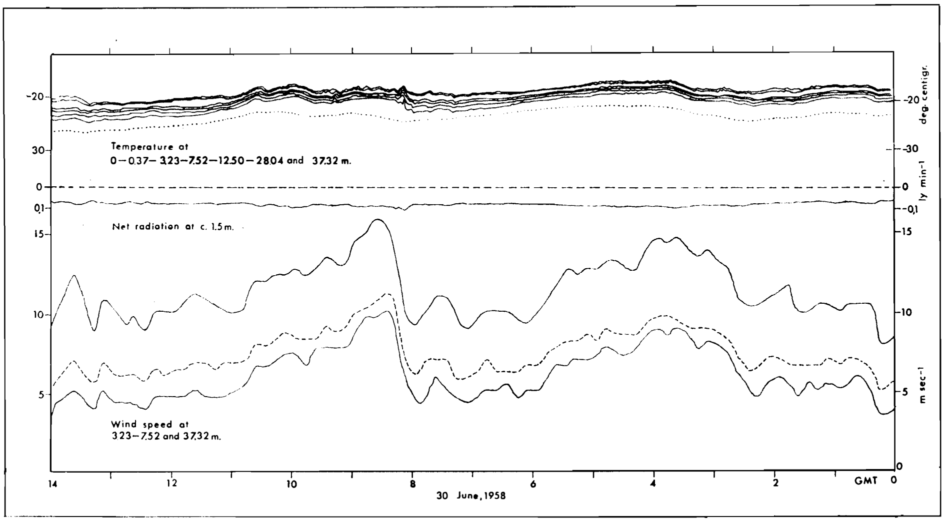


Fig. 8. *Registrations of temperature, net radiation, and wind speed.*

Fig. 8 represents conditions with the highest net radiation loss of the graphs represented here, and also the highest average wind speed at all three levels. As a high net radiation loss would tend to cause a higher stability under otherwise similar conditions, it is obvious that it here is the wind speed which has a dominating influence on the vertical stability. Considering the vertical stratification before and after 0800 GMT, it is also seen that in spite of the great wind speed increase, from about 6.5 to 11.5 m sec$^{-1}$ at 7.5 m, there occurs only a small variation in the vertical temperature gradient.

When making a comparison between the average vertical wind and temperature differences for the period represented by Fig. 8 and the first ten hours of the period represented by Fig. 7, we obtain the results shown in Table 5.

Table 5

| Height interval m | Fig. 8 | | Fig. 7 | |
|---|---|---|---|---|
| | $\overline{\Delta T}$ | $\overline{\Delta U}$ | $\overline{\Delta T}$ | $\overline{\Delta U}$ |
| 3.3– 7.5 | 0.6 | 1.2 | 6.2 | 2.5 |
| 7.5–37.3 | 1.6 | 3.8 | 6.7 | 1.1 |

Table 5 shows that when the temperature difference in the lowest interval increases ten-fold (from 0.6 to 6.2), the corresponding wind difference increases two-fold, i.e. a tendency which is in accordance with the concept of similarity in the profiles. When considering the upper interval we see, however, that when the temperature difference increases four-fold, the corresponding wind difference is *decreased* more than three-fold. This comparison, together with features found in the other graphs, thus indicate that a similar tendency in the variation in wind and temperature gradients does not hold for the full height of the tower.

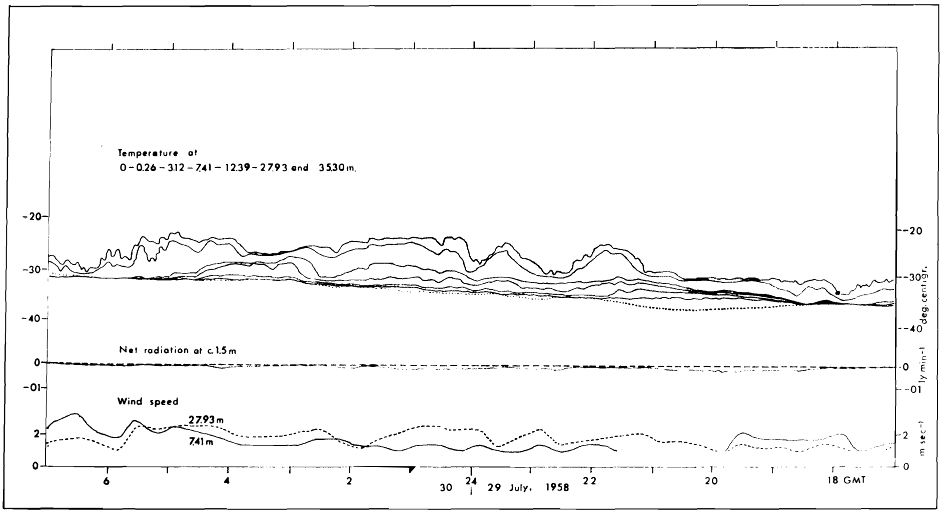


Fig. 9. *Registrations of temperature, net radiation, and wind speed.*

Fig. 9 shows a peculiar stratification with near isothermal conditions near the surface, which for periods extends to at least 12 m, while there is a marked inversion above. The net radiation loss is small, and for periods with the greatest vertical extension of the near isothermy, it is zero. In this period mist was reported, and it is possible that the concentration of water vapour at lower levels, together with the low wind speed may be of importance for this special stratification. The temperature distribution mentioned above was rarely observed.

## THE VARIATION OF THE TEMPERATURE PROFILE WITH STABILITY

The temperature profiles have been read every three hours and the necessary corrections have been applied. The desired data have been calculated with the aid of an electronic computor, and from linear interpolation we have calculated the temperatures at certain even heights adjacent to the actual heights of observation. From April 1958 the snow surface temperature was measured with a resistance thermometer, which was about 10 cm long. The thermometer was kept horizontal and half buried in the snow. Except for the middle of the summer, during fine weather conditions, there was always a good contact between the thermometer and the snow, and the given surface temperature should then represent the temperature of the uppermost millimetres of the snow. During windy weather with accumulation the thermometer was buried, and it had to be replaced. The temperature of the surface was under such conditions considered equal to that of the air which generally was isothermal.

To show the variation of the temperature profile with respect to the vertical stability, we have grouped the profiles with respect to intervals of the mean temperature gradient between 1 and 4 metres. We consider for this purpose the period April—September 1958 for which we have the best vertical distribution of the thermometers together with the greatest variation in the temperature gradient.

Table 6

*Temperatures given in negative °C at different heights for different magnitude of the temperature gradient (G) between 1 and 4 m.*

| Interval of G °C/m | Metres above the surface | | | | | | | Number of profiles | Profile No. |
|---|---|---|---|---|---|---|---|---|---|
| | 0 | 0.5 | 4 | 8 | 12 | 20 | 32 | | |
| 0.2>G>0 | 24.7 | 24.0 | 23.5 | 23.0 | 22.7 | 21.9 | 21.0 | 233 | 1 |
| 0.4>G>0.2 | 29.0 | 27.5 | 26.4 | 25.6 | 24.8 | 23.6 | 22.4 | 172 | 2 |
| 0.6>G>0.4 | 32.1 | 30.3 | 28.3 | 26.5 | 25.2 | 23.7 | 22.2 | 88 | 3 |
| 0.8>G>0.6 | 33.4 | 31.0 | 28.2 | 26.5 | 25.7 | 24.7 | 23.4 | 74 | 4 |
| 1.0>G>0.8 | 34.7 | 32.5 | 29.1 | 26.6 | 25.7 | 24.6 | 23.7 | 63 | 5 |
| 1.5>G>1.0 | 35.9 | 33.1 | 28.6 | 25.7 | 25.1 | 24.0 | 23.1 | 99 | 6 |
| 2.0>G>1.5 | 35.9 | 32.9 | 26.5 | 23.8 | 23.2 | 22.2 | 21.5 | 50 | 7 |
| G>2.0 | 39.2 | 36.2 | 27.4 | 25.2 | 24.7 | 24.1 | 23.5 | 25 | 8 |

For the mentioned months we have 1964 profiles out of which 804, or 41%, reveal a positive gradient. The result of the grouping of the profiles showing an inversion is represented in Table 6.

The actual heights of the thermometers during the period in question is about 0.0, 0.5, 3.2, 7.5, 12.5, 23.0, and 35.0 m. It can be seen from Table 6 that for increasing stability, the decrease in the temperature at the surface is considerable, in contrast to the decrease at the top of the tower. An increase in the stability near the surface occurs generally in connection with the existence of a net radiation loss together with a decrease in the wind speed. The feature mentioned is illustrated in Fig. 6.

A visual impression of Table 6 is represented in Fig. 10 which reveals that

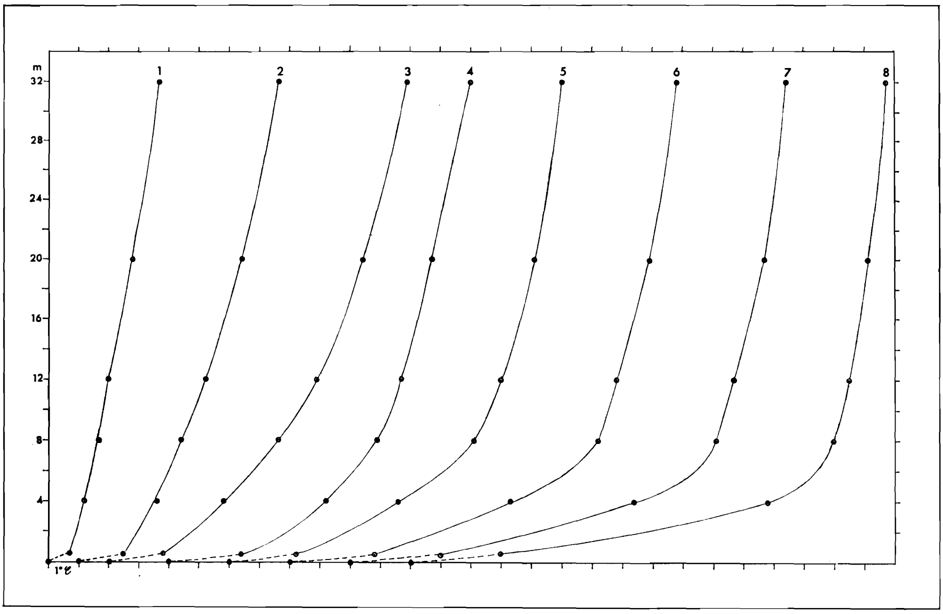


Fig. 10. *The shape of the temperature profile for different stabilities between one and four metres for the period April—September 1958. The reference numbers at the top of the profiles correspond to those given in Table 6.*

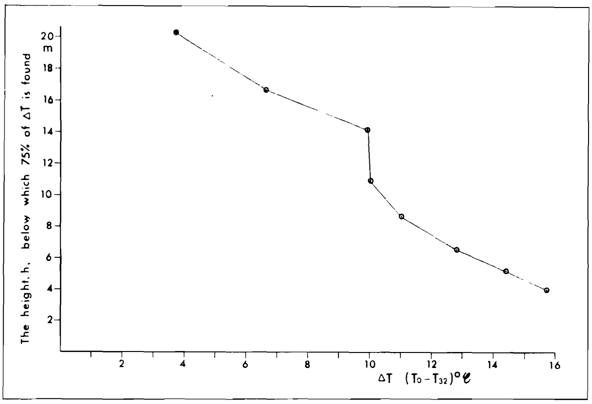


Fig. 11. *The relation between the magnitude of the inversion in the tower and the height below which 75% of this magnitude is found.*

there occurs a marked change in the form of the profile for increasing stability. For the first three profiles it can be seen that there is a corresponding variation of the temperature gradient at all heights, whereas for the rest of the profiles the temperature gradient in the upper 2/3 of the tower decreases as G increases. *This feature reveals that if all the observations are going to be fitted to a mathematical function of a special type, we should keep to observations within the lowest half decametre.* Below this level there apparently occurs no great change in the form of the profile for increasing stability.

It can be seen from the figure that the major temperature difference is found within a shorter and shorter distance from the surface for increasing stability. This feature is illustrated in Fig. 11, where we have given the height above the surface below which 75% of the total temperature difference in the tower is found. The figure shows that the height (h) falls markedly for increasing stability. Or in other words that the vertical extension of a dominant influence of the colder surface is reduced markedly as the stability increases.

## WIND SPEED AND MAGNITUDE OF INVERSION

Figs. 6 and 8 (p. 13, 14) give a clear illustration of how the temperature profile may change as the wind speed varies. To investigate this relationship closer we have considered 567 profiles of temperature and wind speed for the period 20.V to 17.VII, 1958. (The argument for this selection is given on p. 40.) Fig. 12 shows the temperature difference between the lowest and the highest thermometer (at about 0.4 and 37.0 m) with respect to the wind speed at the height of 10 m for inversion conditions. The wind speed is, as mentioned previously, based on a 10 or 20 min. average while the temperature is based on momentary values.

Fig. 12 shows how the maximum temperature difference is reduced as the wind speed increases up to a certain limit, about 11 m $sec^{-1}$. Above this approximate limit the influence of an increase in the wind speed on the magnitude of the inversion is drastically reduced. This feature is in fair accordance with the observations made by LILJEQUIST (1957, p. 205) at Maudheim in a 10 m high mast. In the considered height interval we get, however, marked surface inversions for wind speeds lower than about 11 m $sec^{-1}$, while LILJEQUIST gets this

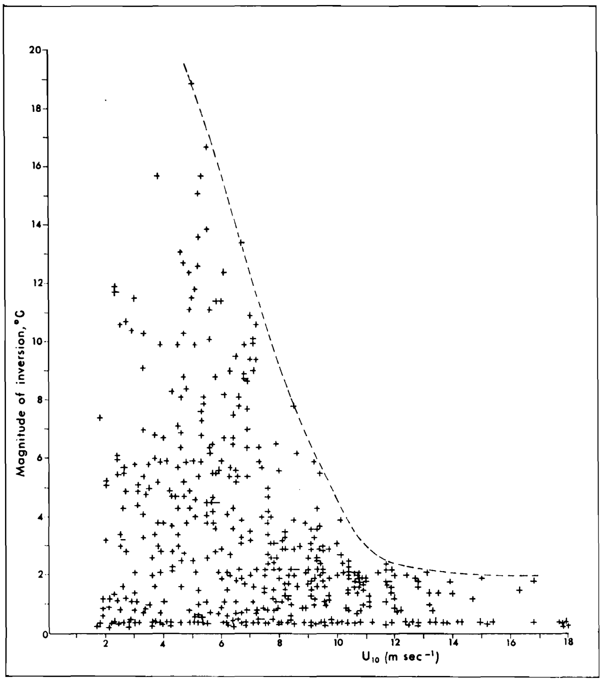


Fig. 12. *Observed temperature differences between c. 0.4 and 37 m and the corresponding wind speed at the 10 m level.*

limit at 8—9 m sec$^{-1}$. It may be noted that this observed limit corresponds fairly well to the wind speed for which drifting snow generally starts. Thus from BUDD et al. (RUBIN 1966) it can be found that the density of the snow drift at Byrd Station, for example at 12.5 cm, on an average increases from 5 to 30 gr m$^{-3}$ as the wind speed ($U_{10}$) increases from 12 to 18 m sec$^{-1}$. A possible damping influence of the drifting snow upon the vertical motions will be considered on p. 43 when discussing the turbulent displacements.

The temperature differences represent, as mentioned, momentary values. Due to the great fluctuations in the temperature, especially for lower wind speeds (cf. Figs. 3—9) the maximum magnitude is not necessarily a persistent one. It is therefore of interest to see how the temperature gradients at different levels vary with the wind speed for average conditions with a maximum magnitude of the inversion. The 567 profiles have been grouped within the following intervals of the wind speed at 10 m (m sec$^{-1}$): 1.0—4.9, 5.0—9.9, and so on. Within these groups the profiles have been further divided into groups within intervals of 0.010 ly min$^{-1}$ of the net radiation loss. (The short-wave radiation is zero, as we have polar night for the period considered.) Among the different wind groups we have selected those with the greatest corresponding temperature difference between the lowest and highest thermometer. The result is represented in Table 7 and Fig. 13. The approximate temperature gradients at the chosen levels have been found by fitting the temperature observations to a power profile with the aid of the method of least squares.

When the wind speed decreases from about 11 to 7 m sec$^{-1}$, the figure shows a maximum increase in the temperature gradient at all levels. For a further

Table 7

*The approximate temperature gradient (°C/m) at different levels with maximum magnitude of the inversion for different wind speeds (m sec$^{-1}$) at 10 m level.*

| Height m | Wind speed | | | |
|---|---|---|---|---|
| | 3.9 | 6.5 | 11.1 | 16.7 |
| 1 | 0.77 | 0.70 | 0.15 | 0.10 |
| 2 | 0.52 | 0.49 | 0.12 | 0.08 |
| 4 | 0.36 | 0.35 | 0.09 | 0.07 |
| 8 | 0.24 | 0.35 | 0.07 | 0.06 |
| 16 | 0.17 | 0.18 | 0.05 | 0.05 |
| 32 | 1.11 | 0.12 | 0.04 | 0.04 |
| | 15 | 35 | 9 | 1 |
| | Number of profiles | | | |

decrease in the wind speed, the temperature gradient in the lowest 4 metres still increases, though far less, while the temperature gradients at higher levels *decreases*. This feature is in accordance with the features revealed in Fig. 10 and demonstrates a reduction in the vertical of the efficiencey of turbulent transfer as the stability increases or as the wind speed decreases below 6—8 m sec$^{-1}$.

THE TEMPERATURE PROFILE FOR DIFFERENT WIND DIRECTIONS

It was sometimes observed during inversional conditions that when the wind direction changed radically, the temperature profiles underwent a great change, while the wind speed and cloud cover were more or less the same. To review the relation between the wind direction and the shape of the temperature profile

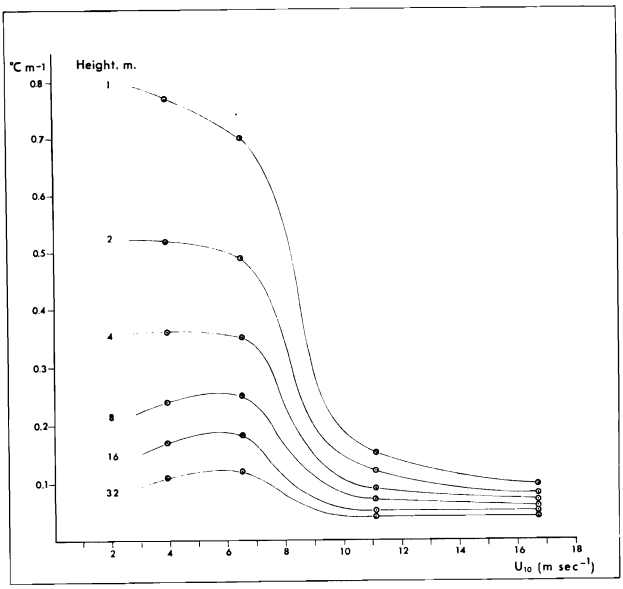


Fig. 13. *The maximum average temperature gradient at different levels with respect to the wind speed at the 10 m level.*

we have considered the measurements from 1 m up to 24 m for the months May, June, July and August 1957, 58 and 59. We have found the mean profile for conditions with small cloud amounts, viz.: 0, 1, 2, and 3 octas and for conditions with cloudy weather with the sky covered with more than 4 octas. The cases where the cloud cover is not observable, mainly due to drifting snow, have been omitted. Within each cloud cover group, the profiles have been further grouped with respect to the following wind speed intervals: 0.5–5.2 and 5.7–10.4 m sec$^{-1}$ at the 10 m level, and also with respect to intervals in the wind direction, viz.: 01–09, 10–18, 19–27, and 28–36 deca degrees. The result is reproduced in Fig. 14.

From Fig. 14 it can be seen that the magnitude of the inversion, as well as the temperature gradients in the lowest decametre show a great variability with respect to wind direction for a given wind speed.

When searching for a physical explanation for these features, the question arises if the net radiation loss indirectly is dependent upon the wind direction (for instance due to a difference in the humidity content of the air masses), so that this eventually may cause the features described above. To investigate this possibility we have considered the hourly mean net radiation loss from the surface near the term hours at which the total cloud cover is less than 3 octas. We have considered the period 20.V–14.VII, 1958 when the sun was below the horizon. The result is given in Table 8.

Table 8

*Net radiation, $Q$ ly min$^{-1}$, for different wind directions and wind speeds, $U_{10}$ m sec$^{-1}$. The letter n gives the number of observations. The total cloud cover is less than 3 octas.*

| Wind speed interval | Wind direction with | | | | | |
|---|---|---|---|---|---|---|
| | easterly component | | | westerly component | | |
| | $\bar{Q}$ | $\bar{U}_{10}$ | n | $\bar{Q}$ | $\bar{U}_{10}$ | n |
| $0.5 < U_{10} \leqslant 5.2$ | −0.036 | 3.2 | 32 | −0.037 | 3.5 | 40 |
| $5.7 \leqslant U_{10} \leqslant 10.4$ | −0.057 | 8.0 | 26 | −0.053 | 7.3 | 28 |

It has been found (see e.g. Vinje 1964, p. 20) that the net radiation loss from the surface increases with the wind speed for otherwise similar conditions, and the differences as can be observed in Table 8 is in accordance with the respective wind speed difference. There is thus no detectable indirect dependency between the net radiation loss and the wind direction. The difference between the profiles may thus, at least partly, be ascribed to advective effects. For the lower wind speeds, when the turbulence is relatively small, this effect may be illustrated by the marked difference between the profiles in Fig. 14 with advection of colder air from the south-easterly direction and that with advection of warmer air from the north-easterly direction. A general feature is that the greatest magnitude of the inversion is found for the less frequent wind directions, i.e. for those with a westerly component. A similar feature is also found at Little America (HOINKES 1967, p. 63). This will be discussed further when the wind profiles are considered (p. 25).

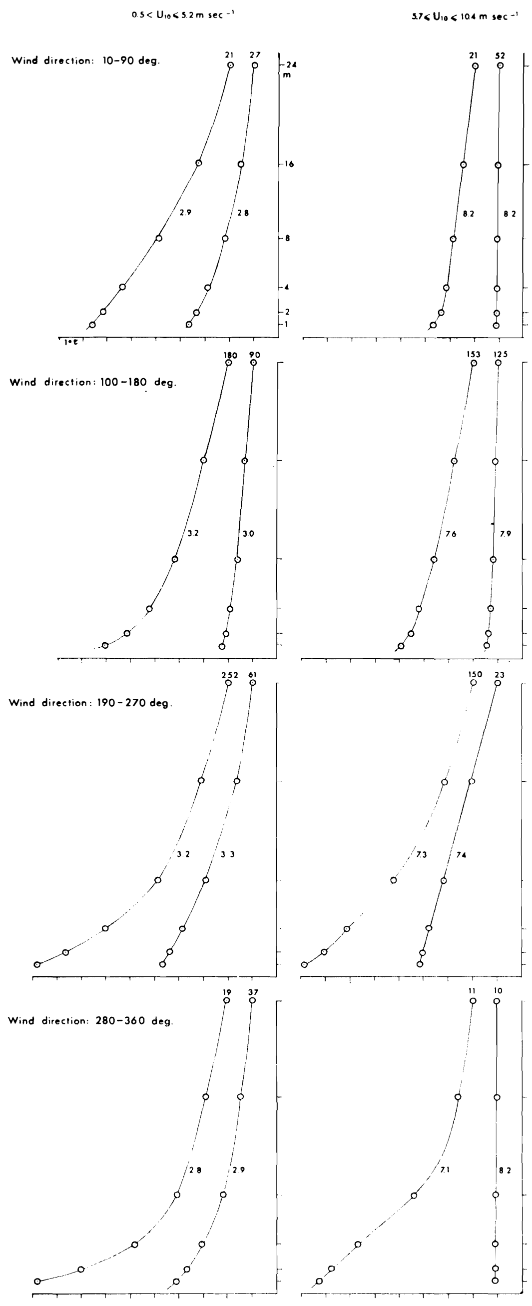


Fig. 14. *The temperature profile for different wind directions within different wind speed intervals Of the pairs of profiles placed together, the profile to the left represents conditions with cloud amount less than 4 octas while that to the right represents conditions with cloud amount more than 4 octas. At the 10 m level is given the corresponding wind speed, and at the top is given the number of profiles within each group.*

## THE FREQUENCY DISTRIBUTION OF THE TEMPERATURE GRADIENT FOR DIFFERENT HEIGHT INTERVALS

To obtain information of the frequency distribution of the temperature gradient at different height intervals during the coldest season, we have considered the period 1.IV–30.IX, 1958. (Heights of thermometers are noted on p. 10.) The result is given in Table 9.

Table 9

*Numbers of different temperature gradients (G, °C/m) occurring within different height intervals (m). The total number of observations is 1464.*

| G | Height intervals | | | |
|---|---|---|---|---|
| | 1–2 | 4–8 | 8–12 | 12–16 |
| $G \leqslant 1$ | 1264 | 1392 | 1446 | 1464 |
| $1 < G \leqslant 2$ | 146 | 65 | 18 | |
| $2 < G \leqslant 3$ | 43 | 7 | | |
| $3 < G \leqslant 4$ | 9 | | | |
| $4 < G \leqslant 5$ | 2 | | | |

It can be seen from Table 9 that the frequency of temperature gradients above 1 °C/m decreases fairly rapidly with height, and also that the "non occurrence level" decreases fairly rapidly with increasing temperature gradient. Thus gradients greater than 1 °C/m do not occur above 12 m, gradients greater than 2 °C/m do not occur above 8 m, and gradients greater than 3 °C/m are found in the lowest couple of metres only. MONIN and OBUKHOV (1954, p. 27, in the trans.) find, according to their theory based on the concept of similarity of temperature and wind profiles, that the temperature profile approaches a linear one for increasing stability. Then $z/L \gg 1$, where L is the height below which the sheering stress varies very little with height. From Table 9 it follows that an eventual linearity in the temperature profile must be sought below 2 m for the highest stabilities over the ice shelf. According to the inequality quoted above, it follows that the height L becomes considerably less than 2 m for the highest stabilities. This is in accordance with the results obtained by MONIN and OBUKHOV in the quoted paper. Over the smoothest surface considered they found $L = 0.7$ m for the most stable condition.

## THE MONTHLY MEAN MAGNITUDE AND HEIGHT OF INVERSION

From the radiosonde data covering the period 1.VI, 1957–31.XII, 1959 we have calculated the monthly mean height of the inversion and the corresponding mean difference between the temperature at screen level and at the top of the inversion. The result is given in Fig. 15.

As can be seen from the figure, there is a marked annual variation in the magnitude of the inversion for 1200 GMT, and it is of special interest to note that the

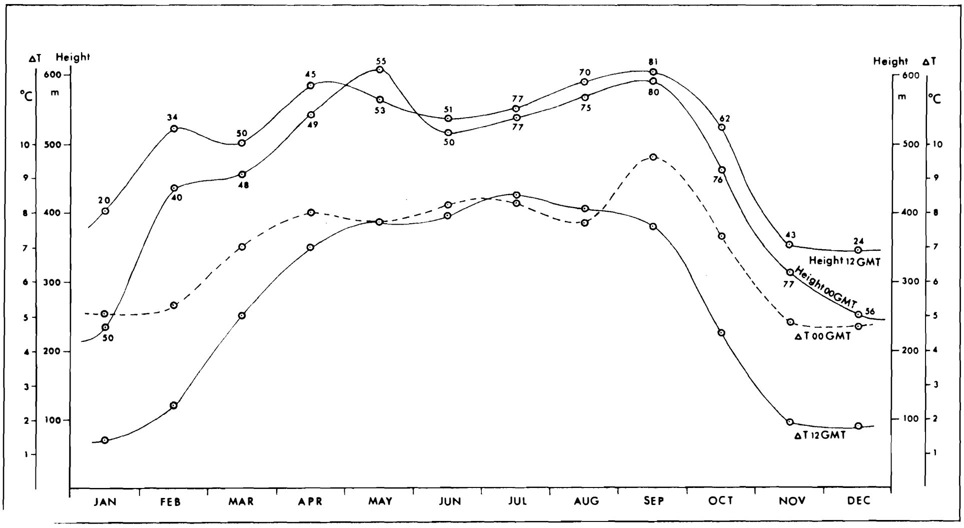


Fig. 15. *The monthly mean magnitude and height of the inversion for 0000 and 1200 GMT as obtained from radiosonde ascents. The number of inversions for each month is given at the upper two curves.*

inversion, as a rule, does not disappear completely during the middle of the day in the warmest season over the ice shelf at Norway Station.

The figure reveals further that the inversion, on an average, has a vertical extension of 5–600 metres during the coldest season and 2–400 metres during the height of the summer. It can also be seen that the height of the inversion during the warmest season is greater in the middle of the day than at night. For the coldest season there is no such marked diurnal variation as would be expected. For the months November, December, and January it is seen that the magnitude of the inversion, on an average, is between about 1 and 5°C. The corresponding range for Maudheim is 1–3°C (Liljequist 1957, p. 279). The difference in this range between the two stations might be due to the more continental climate existing at Norway Station (the distances from the barrier is 6 and 35 km, respectively, for the two stations). For the months May, June, July, August, and September the magnitude of the inversion at Norway Station is between about 8 and 10°C which is about the same as found at Maudheim. This similarity might be expected as the continentality of the climates at the two places should be more equal during the winter season, when the sea is ice-covered.

## The wind profile

### THE INFLUENCE OF STABILITY

The shape of the wind profile over an Antarctic snow field has been studied by e. g. Liljequist (1957), Rusin (1961), and Dalrymple et al. (1963). They considered measurements within the first decametre. As the tower at Norway Station was considerably higher, it is of interest to give some examples of the

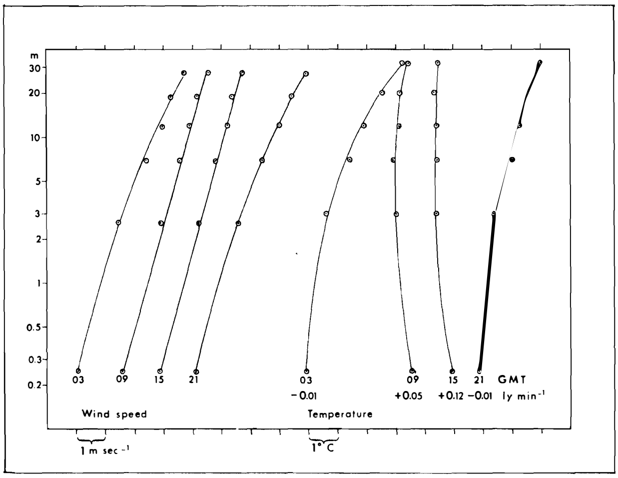


Fig. 16. *Illustration of the effect of the temperature stratification upon the shape of the wind profile.*

profile observed here. This in order to show that there is a general accordance in the measurements and, further, to show some special features of the profile shape which can be observed in the lowest 30—40 metres above the ice shelf.

To show the influence of stability on the profile we have chosen a summer month. For January 1959 we have a continuous registration of both wind speed and temperature. In Fig. 16 is given the monthly average profiles of the wind speed and temperature for the hours 0300, 0900, 1500, and 2100 GMT on a single logarithmic paper.

The figure reveals the well-known feature for stable stratification, that the wind profile is concave to the wind axis when the height is given in a logarithmic scale. For conditions with a net radiation gain (values given in Fig. 16), the wind speed observations can be represented by a quasi-linear equation between the wind speed and the logarithm of the height. At other places (see e.g. LAIKHTMAN 1961, p. 42, in the trans.) it is found, for a net radiation gain at the surface, that the wind profile is convex to the wind axis. This feature cannot be traced in Fig. 16, and the reason for this may be that the inversion, on an average for this month, is not completely destroyed at higher levels during the day over the ice shelf. This is indicated by the temperature profiles given in Fig. 16 and also when considering the full height of the inversion from the radiosonde data (p. 22). Thus this fact suggests that the theories based on similarity in the profiles of temperature and wind speed, e. g. LAIKHTMAN (1961) and MONIN and OBUKHOV (1954), are not valid for the full height of the tower at Norway Station at day time. This is moreover in accordance with the fact that these theories are restricted to stationary conditions, and this supposition is not fulfilled for the profiles shown in Fig. 16, and especially for the profiles at 0900 and 1500 GMT, when we have the greatest systematic increase or decrease, respectively, of the net radiation gain at the surface. A dissimilarity of the profiles for the day time is also found from a study of measurements in a 32 m high tower in Dallas, USA, made by O'BRIEN (1965, p. 2277).

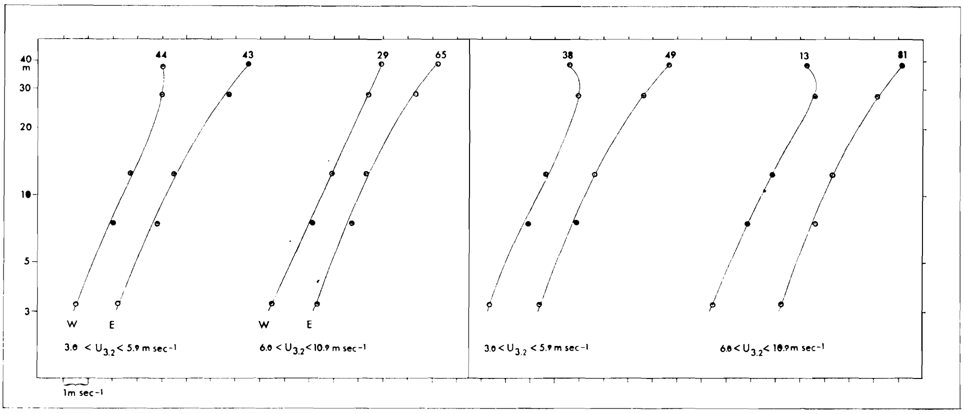


Fig. 17. *To the left: the wind speed profile for wind directions with a westerly component (W) and with an easterly component (E). To the right: average form of the profiles without and with a maximum at about 30 m. The number of profiles in each group is given at the top. The profiles represent conditions with a net radiation loss greater than about 0.03 ly min$^{-1}$.*

## SPECIAL PROFILES

As can be seen from the wind registrations reproduced on pp. 11—15, the wind speed does not always increase monotonically with height. Sometimes a maximum exists for longer or shorter periods below the top of the tower. At times this phenomenon was so pronounced that it could be observed by looking at the anemometers. Two types of maximums could be observed: 1) occurring mainly at about 30 m, and 2) a less frequent one, occurring at a lower level together with a marked minimum above. The shape of the profile for conditions with a maximum of type 1) shall be discussed first.

We consider for this purpose the 567 selected profiles for the period 20.V—15.VII, 1958. For the mentioned period the profiles of type 2) have been omitted. The lowest anemometer, at about half a metre above the surface, showed at this time a far too high wind speed which in all probability was due to multiple contacts (cf. p. 9) and the observations from this level have therefore been disregarded. Wind profiles for the following two intervals have been considered:

$$3.0 < U_{3.2m} < 5.9 \text{ and } 6.0 < U_{3.2m} < 10.9 \text{ m sec}^{-1}$$

for conditions when the net radiation loss is greater than about 0.03 ly min$^{-1}$. This corresponds roughly to conditions when the cloud cover is less than 3 octas (cf. Table 8, p. 20). Previously (p. 19) we have investigated the variation of the shape of the temperature profile with wind direction for similar wind intervals and cloud amounts. It was found there that the shape of the temperature profile is different for easterly and westerly winds for a given condition, and it is now of interest to see if also the shape of the wind speed profile shows a similar dependency. It will therefore here be distinquished between wind directions with an easterly and with a westerly component. The result of the grouping is represented on the left hand side in Fig. 17. The average form of the "abnormal" profile, when disregarding the direction, is shown on the right hand side of the figure together with the "normal" ones. The height is given in a logarithmic scale.

The figure shows that there is a marked difference in the shape of the wind profile for the two wind directions considered. The difference is, however, less pronounced for the higher wind speeds.

From the total number of profiles (783) from the period mentioned, where the wind direction has been observed, it is found that the profile with a maximum or a constant wind speed in the upper part of the tower, occurs both for easterly and westerly winds. However, the percentage occurrence of the mentioned profile is 68 and 19%, respectively, for the lower wind speeds, and 34 and 5%, respectively, for the higher wind speeds. It can accordingly be said that there is a marked tendency for the "abnormal" profile to occur for wind directions with a westerly component.

The corresponding change in the wind shear is of interest, and is shown in Table 10.

Table 10

*The variation of wind shear with height for different types of the wind profile represented in Fig. 17. The net radiation loss is$>0.03$ ly min$^{-1}$.*

| Profile | No. of profiles | Height in m | | | | | |
|---|---|---|---|---|---|---|---|
| | | 3.23 | 7.52 | 12.50 | 28.03 | 37.32 | |
| with max. near 30 m | 51 | 5.10 | 6.65 | 7.39 | 8.88 | 8.53 | Speed, m sec$^{-1}$ |
| | | 0.36 | 0.15 | 0.096 | -0.038 | | Shear, sec$^{-1}$ |
| without max. | 130 | 6.75 | 8.19 | 8.19 | 10.78 | 11.79 | Speed |
| | | 0.33 | 0.14 | 0.12 | 0.11 | | Shear |

It can bee seen from Table 10 that the profile with a pronounced maximum at c. 30 m, i.e. the profile which is most frequent for westerly winds, reveals a somewhat greater shear near the surface. The difference in shear should, moreover, have been increased if the wind speeds for the two different profiles had been more equal. When considering the temperature profiles (p. 19) a corresponding result was found. The temperature gradient within the first decametre was greatest for winds with a westerly component. And due to the known similarity in the variation of gradients of wind speed and temperature near the surface this result becomes reasonable.

In Table 11 is given a review of the frequency of the special profile for different wind speeds regardless of direction. We have considered all the profiles (797) obtained during the period 20.V–15.VII, 1958.

As can be seen from Table 11, the frequency of the special profile in question decreases with increasing wind speed and a marked drop occurs when the wind speed exceeds c. 11 m sec$^{-1}$. And a comparison of the variation of this frequency with Fig. 12, p. 18, indicates again that the occurrence of a maximum wind speed at about 30 m is related to the magnitude of the inversion.

Table 11

| Wind speed at 3.2 m (m sec$^{-1}$) | Number of profiles | Number of profiles with max. at c. 30 m | Percentage frequency |
|---|---|---|---|
| $U < 3.0$ | 216 | 122 | 56 |
| $3.0 \leqslant U < 5.9$ | 251 | 113 | 45 |
| $6.0 \leqslant U < 10.9$ | 223 | 45 | 20 |
| $11.0 \leqslant U$ | 107 | 5 | 5 |
| Total | 797 | 285 | 36 |

A wind profile with a maximum at about 30 m is observed by DEFANT for downslope winds over a steep hill (SUTTON 1933, p. 270). TAUBER (1960, p. 52) gives also some profiles which show a maximum relatively close to the surface for katabatic out-flow near Mirny. He also finds that the station (MS—4), 13.6 km from the coast near Mirny, is outside the limits of the influence of katabatic winds. According to this result we should not expect to find influence of katabatic winds at Norway Station due to the great distance to the slope of the rising continent, which is 25 and 150 km towards SW and SE, respectively. If the maximum in the wind speed at c. 30 m is caused by katabatic effects, it might possibly be due to a decreasing height of the surface inversion towards north over the flat ice shelf. However, as will be seen below, this special profile may occur also with a northwesterly wind direction and this indicates that katabatic effects not necessarily are responsible for the special wind maximum.

In Fig. 18 is demonstrated how the wind speed difference between the upper two levels, about 28 and 37 m, changed when the wind backed from a north-easterly to a north-westerly direction on 2.VI, 1958.

Up to the time when the wind shift took place, the wind speed at the 10 m level had been about 10 m sec$^{-1}$ for about 15 hours and there was a moderate drifting snow and a snow fall. After the wind shift, the wind speed decreased to less than

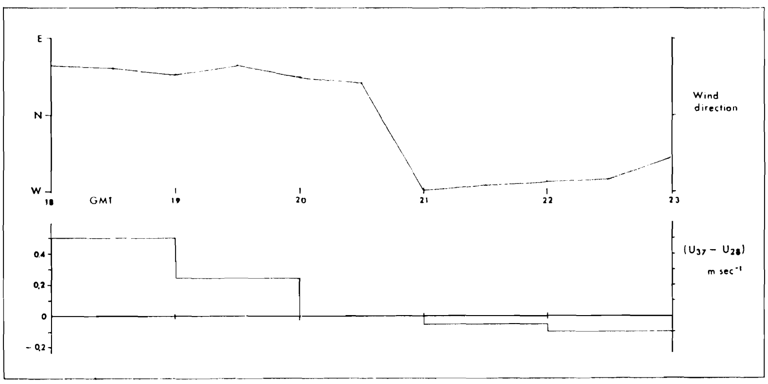


Fig. 18. *The variation of the wind conditions in the upper 9 m of the tower in connection with a weakening and backing of the wind speed. Near isothermal stratification.*

5 m sec$^{-1}$. The net radiation was zero for the whole period. It is possible that the features revealed in Fig. 18 are due to special wind distributions in or near a frontal zone. If this is so, "frontal effects" upon the wind profile may be active under inversional conditions as well, according to the previous discussion.

The wind profile from the radiosonde data at Norway Station and Maudheim reveals, according to HISDAL (unpublished data), that a maximum in the wind speed occurs near the top of the inversion. Similar features have also been found over the Great Planes in the United States. (See e.g. BLACKADAR 1957, p. 283, and BONNER 1968, p. 833.) These maxima are at the latter place denoted as the southerly low level jet. The relation between a maximum wind speed and the top of the inversion may suggest that secondary maxima could be developed at lower heights as well, due to marked difference in the temperature gradient within neighbouring height intervals. This may also be supported by the observed indication mentioned above, of a relation between the frequency of a maximum wind speed at about 30 m and the magnitude of the inversion, and also by the fact that the special profile is far more frequent for westerly winds for which we have the greatest magnitude of the inversion. The development in question is also clearly demonstrated in Fig. 7, p. 13, from the observations obtained between 0000 and 0500 GMT: when a maximum in the wind speed comes into existence below 37 m the temperature gradient in the upper part of the tower decreases while that in the lower part of the tower increases.

As seen above, a pronounced part of the wind profiles, especially for lower wind speeds, assumes a zero value of the wind shear between 20 and 30 m above the surface. Assuming the shearing stress to be directly proportional to the wind shear, this means that the vertical transfer of momentum becomes zero at this height quite often.

As mentioned on p. 25, we sometimes observed a maximum in the wind speed at a lower level together with a pronounced minimum above. In Fig. 19 is demonstrated how the wind profile changed when this special profile appeared.

As can be seen from the figure, the maximum as well as the minimum are very marked, and the difference in the extreme wind speeds may be as high as 3–4 m

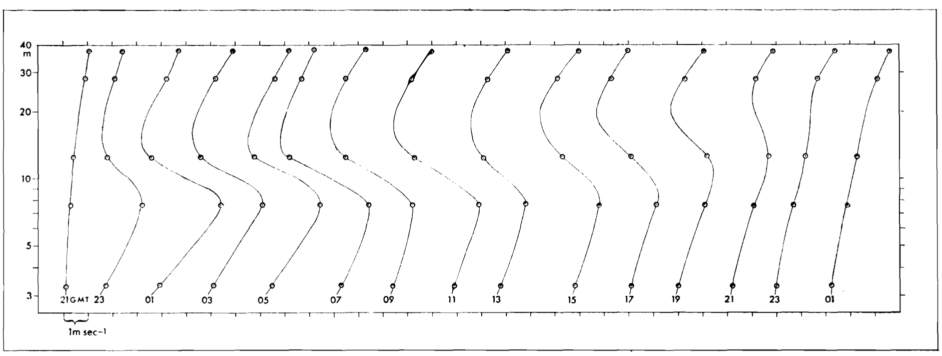


Fig. 19. *Illustration of a special wind profile which was observed rather seldom at Norway Station. In the case shown, it occurred in connection with an advection of considerably warmer air from East-South-East during the evening of 23.V, 1958.*

$sec^{-1}$ for some of the profiles. We remark also the very great increase in the wind shear at lower levels when the special profile appears. For the conditions in question there occurred an advection of considerably warmer air. Thus at 1700 GMT 23.V we had below −20°C at all levels in the tower, while at 2100 GMT the same evening the temperature had increased to about −10°C at all levels. At the same time moderate fog appeared. At about midnight the fog disappeared and a light to moderate snow fall started with some shorter intermissions during the next day. During 24.V the mean net radiation loss was 0.016 ly $min^{-1}$, and the magnitude of the inversion in the tower did not exceed 2°C. The wind speed varied between c. 8 and 11 m $sec^{-1}$ at the 10 m level, and the direction varied between 090 and 130 deg.

For the period 20.V–15.VII, 1958, it is found that the profile in question occurred with a frequency of 4.6%; however, 2.4% is found during the period represented in Fig. 19.

We note the similarity of the form in the lower part of the two special profile types considered in this section. And it might therefore be that the last considered profile represents a special case when the maximum wind speed, which for the former wind profile was found at about 30 m, now is observed at an extra low level. It is also noted that the form of the latter profile is in qualitative accordance with the form given by the equations applied by e. g. SVERDRUP (1918, p. 181).

As a conclusion from the measurements considered above, it follows that the normal friction profile, with a monotonically increase of the wind speed with height, does not as a rule exist within the height interval of about 40 m over the ice shelf. The deviation can at times be considerable and, when the aim is to consider the normal friction profile, it is necessary with a critical selection when the observations include heights above the first half decameter.

## THE PROFILES FOR ISOTHERMAL CONDITIONS

Of special interest is the wind profile for an isothermal stratification. Such conditions should prevail when the net radiation is zero. For the 567 profiles from the period 20.V–15.VII, 1958, the mean net radiation for the preceding hour has been noted, and the average wind profile when the net radiation is zero has been found for the following groups of the wind speed at the 10 m level: 1.0–4.9, 5.0–9.9, 10.0–14.9 m $sec^{-1}$, and so on. The resulting profiles are represented on a single logarithmic paper, Fig. 20.

By the method of least squares the profiles have been fitted to a power profile of the form

$$U = U_1 \, z^{\varepsilon}$$

where $U_1$ is the wind speed at the one meter level. This value is given in the figure.

For conditions with no or slight drifting snow, i.e. for $U_{10} < \sim 10$ m $sec^{-1}$, the exponent assumes values between c. 0.13 and 0.14 (Table 12). A similar value, $\varepsilon = 0.13$, is found by e.g. SCRASE (1930) from observations between c. 3 and 13 m over level downland for very small temperature gradients, and FROST (1947,

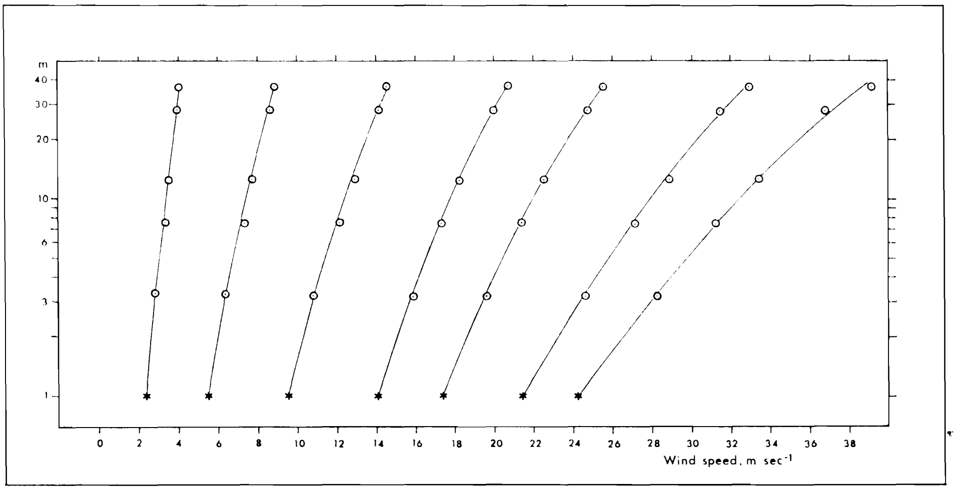


Fig. 20. *The wind speed profile for different wind speeds under near isothermal conditions. The net radiation is zero.*

p. 14) concludes that the value of the exponent for adiabatic conditions is 0.149 for the great height interval 1.2—300 m. This result is of particular interest because it indicates virtually the same profile as the well-known "seventh-root law" ($\varepsilon = 1/7$) of variation of the velocity with height for flow of liquids in pipes.

Table 12

*The variation of the profile exponent with wind speed for different height intervals.*

| | Drifting snow | | | | | | |
|---|---|---|---|---|---|---|---|
| | no or slight to moderate | | | moderate to heavy | | | |
| $U_{10}$ (m sec$^{-1}$) | 3.4 | 7.4 | 12.4 | 17.8 | 22.0 | 27.9 | 32.3 |
| (3.3–37.0 m) | 0.142 | 0.132 | 0.119 | 0.106 | 0.107 | 0.117 | 0.129 |
| (3.3–12.5 m) | 0.136 | 0.142 | 0.132 | 0.099 | 0.100 | 0.115 | 0.120 |

If we consider the height interval between 3.3 and 12.5 m (third line in Table 12), it is seen that the respective exponents are somewhat changed. This change may be due to instrumental errors or due to the influence of irregular wind profiles with a maximum occurring at about 30 m, considered in the previous section. The latter influence may specially be felt for the lower wind speeds according to the increasing frequency of the "abnormal" profile for decreasing wind speed (p. 27).

From Table 12 it can be seen that when the intensity of the drifting snow increases from slight to moderate, the profile exponent is notably decreased. This decrease means that the effect of the drifting snow is to reduce the variation of the wind speed with height. Our measurements indicate in other words that the effect of the drifting snow is to reduce the roughness of the surface. This means accordingly that the drag of the surface on the stream should be relatively reduced when snow drift occurs. And this result will therefore be further discussed in connection with the calculation of the shearing stress (p. 47).

It is seen from Fig. 20 that the logarithmic profile is not valid for isothermal conditions for higher wind speeds. It is further seen that for the lower wind speeds as well as for the lower heights a distinction between a logarithmic and a power profile is very difficult. For near neutral conditions at Antarctic stations LILJEQUIST (1957, p. 195) and RUSIN (1961, p. 311, in the trans.) found that the wind profiles within the lowest 10 m may be approximated with a logarithmic function.

## On the turbulent transfer

### A VORTEX THEORY

If an eddy or vortex of the form indicated in Fig. 21 is developed where there exist vertical gradients in the wind speed and temperature, we should expect to find a vertical zigzag profile, as sketched in the figure. The form of the vortex indicated to the left in the figure is similar to that proposed and observed in fluid by MALLOCK (1917, p. 14). ROSENHEAD (1932) investigates the variation with time of a surface of discontinuity. Initially the surface of separation is in the form of a sine-curve of small amplitude. He finds that the effect of the wind shear is to produce vortices of a form which is similar to that indicated to the left in Fig. 21. The assumed movements of the elements with respect to the surface, as indicated to the right in the figure, are in accordance with the tentative proposition made by LUMLEY and PANOFSKY (1964, p. 210). From the study of turbulence by statistical methods, they found some indications that, with sufficient wind, eddies tend to be of the corksrew variety with axis parallel to the mean wind. There is also an accordance between the model and the results obtained by PRIESTLEY (1959, p. 73). He considered the influence of buoyancy forces and found from a mathematical treatment that when an element is displaced vertically, it will first transcend its final equilibrium and subsequently executes damped harmonic oscillations.

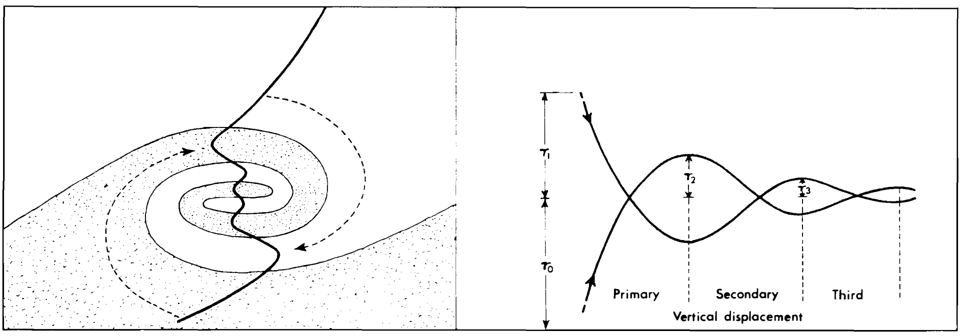


Fig. 21. *A tentative model of a friction vortex. The left hand side indicates how the displacements are related to the moving air. The zigzag line indicates the assumed temperature or wind speed distribution. The right hand side of the figure indicates how the displacements are related to the surface.* ($\varkappa = r_3 : r_2 = r_2 : r_1 = 0.4$).

When the air is streaming along the snow surface, it is reasonable to assume that the formation of friction vortices occurs behind the top of the relatively uniform unevenesses. Due to this uniformity, it is expected that the vertical extension of the vortices is of a predominant magnitude. If so, the repeated production of vortices should be observable and cause a systematizing influence upon the average profile forms near the guiding surface. In accordance with this view, we have tentatively explained the repeatedly observed zigzag profile of the temperature in the lowest 20—30 cm above the surface at Norway Station by the effect of passing vortices of the form sketched in Fig. 21 (VINJE 1967a). Similar zigzag profiles have also been found by CAISLEY et al. (1963) over a glacier in northern Sweden. They state in their preliminary paper that there is some similarity in the profiles for the different elements measured, particularly in those of the wind speed and temperature. Further, the mentioned profile can be found from temperature observations over the inland ice of Greenland (AMBACH 1964, p. 277). And in 1968 the present author found that a zigzag profile was prevalent below c. 10 cm over a relatively flat, melting snow surface in Spitsbergen. (Unpublished data.)

When the water vapour content is great, the radiative flux divergence becomes of importance. The special temperature profiles observed by e.g. LAKE (1956), is by GAEVSKAYA et al. (1962) ascribed to the effect of the radiation from the water vapour. For the low temperature in question during the winter months at Norway Station this effect is, however, supposed to be negligible.

In a previous paper (VINJE 1967a) based on the measurements made at Norway Station, the geometrical picture of the surface friction vortex has been discussed. An approximate expression for the net vertical transfer of heat was derived, and the estimated transfer of heat was of a correct order of magnitude. It was found that the net vertical heat flux (q) for stable conditions consists of alternating downward and upward directed fluxes which decrease in a geometrical manner as the elements are approaching the final equilibrium height. Thus

$$q = q_1 (1 - \varkappa^2 + \varkappa^4 \ldots),$$

where $q_1$, $q_1\varkappa^2$ etc. are the fluxes due to the primary, secondary, etc. vertical movements. The factor $\varkappa$ expresses the reduction of the amplitude of the oscillating elements (cf. Fig. 21) and also the successive reduction in the temperature difference between the upward and downward moving elements. Its value was from the observations mentioned found to be about 0.4 and the geometrical series above therefore converges very quickly.

To test the vortex theory in connection with our profile measurements, we shall refrain here from expressing the vertical velocity according to a special interpretation of the movements. We shall instead tentatively assume that the mean vertical velocity for the $m^{th}$ oscillation of a disturbed element (Wm) is proportional to the product of the frequency of the formation of vortices (n) and the respective average vertical displacement of the upward and downward moving elements ($\Delta$Zm). Thus

$$W_m = C\, n\, (\Delta Z_m)$$

where C is the factor of proportionality. (As we here will consider average conditions only, the bar indicating averaging is omitted.) Near the surface where the gradients are relatively great the effect of the adiabatic temperature changes can be neglected and, according to VINJE (1967a), the modified expression given above for the mean vertical velocity yields the following approximation for the net vertical transfer of heat:

$$q = \rho\, c_p\, 2\, C\, k\, n\, r^2 \frac{\delta T}{\delta z} \qquad 1$$

where $\rho$ and $c_p$ have their usual meaning and 2r is the mean vertical extension of the vortices. According to the geometrical picture, r is the average net vertical displacement of an element which is displaced. It was found in the quoted paper that $k = (1+\varkappa)/(1+\varkappa^2) = 1.21$, where $\varkappa$, as mentioned above, is the average proportion between the amplitudes of two consecutive oscillations of an upward and a downward moving element. In other words, $\varkappa$ may be considered as a damping factor of the amplitude of the oscillating element. It can be seen that for a variation of $\varkappa$ of as much as 50%, the factor k varies less than 5%, and it has therefore been considered as constant for the stable conditions considered in this paper.

According to the model vortex, there is similarity between the vertical transfer of heat and the vertical transfer of momentum. This should be the case as long as the effect of the adiabatic temperature changes are negligible. Then the mentioned similarity should hold near the surface where the gradients generally are great and where the vertical displacements are minimal. According to this, we may write for the net vertical transfer of momentum, viz.:

$$\tau = \rho\, 2\, C\, kn\, r^2 \frac{\delta U}{\delta z} \qquad 2$$

where $(\delta U/\delta z)$ is the vertical wind shear.

BRUNT (1952, p. 216) gives an expression for the Reynolds number (R) for the motions in the atmosphere, viz.:

$$f(R) = \frac{r^2}{\upsilon}\, \frac{\delta U}{\delta z}$$

where $\upsilon$ is the kinematic viscosity and where we have replaced the mixing length (l) with the net vertical displacement of a disturbed element (r). When we consider the layer where the friction forces are dominant, i.e. near the surface, it seems reasonable to assume that the vortices formed in the air stream should be approximately geometrically similar for a given condition. This assumed approximate similarity involves that f(R) is approximately constant with height and we may tentatively write that

$$\rho\, r^2 \frac{\delta U}{\delta z} = M \approx \text{constant near the surface.} \qquad 3$$

When following the air stream, it is seen that M may be interpreted as the moment of momentum per unit cube of air moving at a distance of r from the center of a vortex. It follows also from the assumed similarity that $\varkappa$, and accordingly also k in eq. 2) above, are constant with height for a given condition.

On condition that M is constant with height for a given stability, it follows from eq. 2) that $\tau$ is proportional with the frequency n. It is reasonable to assume that the turbulent impulses originating at the surface are reduced in number with height for a given condition. This suggests accordingly a reduction with height of the hampering effect of the surface on the air stream. A decrease of the shearing stress with height is in accordance with theoretical considerations (BRUNT 1952, p. 245) and it is also indicated from direct measurements in the atmosphere (DEACON 1955). The often assumed constancy of $\tau$ with height in the surface layer may be questioned, as it involves that the friction forces vanishes.

When now assuming that M is constant with height, equation 2) can be written as

$$\rho r^2 \frac{\delta U}{\delta z} = \frac{\tau_0}{2 C k n_0} = M \qquad 4$$

where the index refers to the mean height of the centers of the surface vortices, *down to which height the present theory is supposed to be valid.*

It is an observed fact that the wind profile for diabatic conditions can be described approximately by a power of the height. It is seen that to meet this demand, and further to obtain the logarithmic profile for adiabatic conditions, r may be expressed by a power of the height, thus,

$$r = a z^K$$

where $K = 0.5$ for adiabatic conditions. If the vertical extension of the displaced elements is small compared with the net vertical displacement, $r_0$, we may write as an approximation for the height $z = r_0$

$$r_0 = a r_0^K$$

and thus

$$r = r_0^{1-K} z^K \qquad 5$$

This expression is inserted in equation 4), and assuming that the wind speed at the zero plane level can be neglected, we obtain from an integration between the zero plane level and the height z that

$$U = \frac{\tau_0 r_0^{-(1+\varepsilon)}}{\varepsilon \rho 2 C k n_0} z^{\varepsilon} = U_1 z^{\varepsilon} = U_0 \left(\frac{z}{r_0}\right)^{\varepsilon} \qquad 6$$

where $U_1$ is the wind speed at the one metre level and $\varepsilon = 1 - 2K$. When $K = 0.5$ the integration of 4) gives a logarithmic wind profile. When $\varepsilon \rightarrow 1$ it is seen that the profile form approaches a linear one. As mentioned on p. 22, this is in accordance with what is suggested by the MONIN/OBUKHOV theory and in agreement with their observations. When considering the temperature profiles represented in Fig. 10, p. 16, it is seen that a power profile may be found within the first half decametre only, for the highest stabilities. Then, when considering the observations below the first c. five metres, it is seen that the above mentioned approach to a linear profile is clearly demonstrated from our observations.

*a. The Kármán constant.* – The Kármán constant is an important factor in the different profile equations and as its square appears in the expression for the shearing stress, its exact value is of great importance. There exists, however, no generally adopted value, neither for isothermal nor stable conditions. In the much applied theory of MONIN and OBUKHOV (1954), based on dimensional analysis, it is assumed that the Kármán constant is independent of stability, whereas from the similar theories of LAIKHTMAN (1961) and DEACON (1953), it follows from the work by OGNEVA (1955, p. 46) that the Kármán constant increases with increasing stability.

The relation between a characteristic length (l) of the turbulent motions and the Kármán constant ($k_0$) was given by KÁRMÁN (1930, p. 64), viz.:

$$l = k_0 \left[(\delta U / \delta z) / (\delta^2 U / \delta z^2)\right]$$

He obtained this equation by assuming an identical pattern in the turbulent flow. This is also assumed in the present theory, and the above equation should be compared with that which can be obtained from the derivation of eq. 4) with respect to height. It follows, when substituting for r and $(\delta r/\delta z)$ given by eq. 5), that

$$\frac{z}{2} = -K\left[(\delta U / \delta z) / (\delta^2 U / \delta z^2)\right]$$

This equation results from the assumption of similarity of the vortices below the height where the friction forces are of dominant importance for the shape of the wind profile. No assumption has been made as for the variation of the shearing stress with height. The Kármán equation refers to the heights above the surface where the effect of this variation can be neglected. The relationship has often been applied for the whole lower boundary layer under the assumption that $l = k_0 z$ for adiabatic conditions. With the additional assumption that $\tau$ is constant with height, the well-known logarithmic wind profile is obtained. It is, however, known that the shearing stress decreases with height (see e.g. BRUNT (1952, p. 245)) and THORADE (ibid., p. 247) suggests that when putting $l = k_0 z$, this compensates the neglection of the variation of $\tau$ with height thus that the logarithmic formula is correct within the limits of experimental error. According to this, the two equations above should be compared for adiabatic conditions. Then $l = k_0 z$ and $K = 0.5$, and it is seen that the equations given above become identical. This comparison thus suggests that $K$ might be identified as the Kármán constant. This constant is often assumed to have a value of 0.40 in the wind tunnel where the stratification is isothermal. However, LETTAU (1961) has suggested a revision to 0.428 as a result of his re-examination of earlier data on flow in rough pipes. As can be seen from Table 13, p. 41, the parameter $K$ is close to 0.43 for isothermal conditions when the drifting snow is small or nil ($U_{10} < \sim 10$ m sec$^{-1}$). The identification mentioned above is thus strongly supported.

It is also seen from Table 13 that $K$ decreases with increasing stability from about 0.43 for isothermal conditions to about 0.30 for the most stable conditions considered. This trend is also similar to that of the Kármán constant according to the calculations made by SHEPPARD (1947, p. 219). On condition that $k_0 = 0.4$ for isothermal conditions, he obtained the following results from measurements of wind profiles and gustiness performed by BEST in the lowest 2 metres above the surface:

| | | | |
|---|---|---|---|
| Temp.grad. °C/m | −1.67 | 0.0 | +0.56 |
| Kármán constant | 0.61 | 0.40 | 0.22 |

It is also seen from this table that the Kármán constant for adiabatic conditions should be found between 0.4 and 0.6. For similar conditions it follows from the present theory that $K = 0.5$. The suggestion above that $K$ may be identified as the Kármán constant is thus also supported when considering reduction with respect to increasing stability.

This trend might possibly also be supported by the following. KÁRMÁN (1930, p. 64) states that $k_0$ for one thing is dependent upon the mechanism of the turbulent fluctuations, and LETTAU (1964, p. 455) arrives at an expression for the Kármán constant in terms of physical-statistical fluctuating characteristics. As mentioned above (p. 32), the damping factor ($\varkappa$) of the oscillations have been estimated to be about 0.4 for a certain stable condition. The damping factor is of great importance for the form of the vortices and should thus be of great importance for the form of the profile as well. The similarity between $\varkappa$, $K$ and $k_0$ therefore immediately suggests a possible identification and a possible physical interpretation of the Kármán constant. We have also that $\varkappa$ should decrease with increasing stability, or in other words, that the amplitude of the oscillations is reduced more rapidly when the stability increases. When a total damping is approached, i. e. $\varkappa \to 0$, the possible identification mentioned would mean that also $K \to 0$. This involves an approach to a linear profile, which in turn indicates that the flow becomes laminar (SUTTON 1953, p. 37). And this should be expected when the damping of the the turbulent motions becomes total.

According to the positive indications above, we shall here make the assumption that $K$ equals the Kármán constant. On this condition it follows then that the Kármán constant may be determined by the profile exponent ($\varepsilon$) as $\varepsilon = 1 - 2K$.

It follows also, when referring to the height $z = r_0$, that

$$(\delta U / \delta z) r_0 = U_0^* / K r_0 \qquad 7$$

where $U_0^*$ is the friction velocity. The underlying well-known semi-empirical equation is generally supposed to hold close to the surface for different stabilities.

By definition

$$\tau_0 = \rho (U_0^*)^2$$

and from a combination of 2) and 7) we obtain for the height $z = r_0$

$$\tau_0 = \rho \left( \frac{2 C k n_0 r_0}{K} \right)^2 \qquad 8$$

From equation 6) and 8) then follows that

$$r_0 = \left( \frac{K^2 \varepsilon U_1}{2 C k n_0} \right)^{\frac{1}{2K}} \qquad 9$$

*b. The frequency of passing vortices.* – The vortex theory is based on the hypothesis that vortices of the form indicated in Fig. 21 are formed in the air, and that they pass a fixed place with the frequency n. It is obvious that this frequency must depend upon the wind speed, and we will below derive a tentative form of this relationship.

The number of oscillations in the wind speed and in the temperature observed at a certain height depends upon the sensibility of the measuring device. However, on the condition that the oscillations originally are caused by the passing vortices, we may tentatively suppose that the frequency (n) is proportional to the total number of oscillations observed with a certain device. To find the form of the relationship between n and the wind speed, we may then consider the measurements of the oscillations in the wind speed made by GIBLETT et al. (1932) over level country obtained by a Dines speed recorder. From the ultra quick run traces reproduced in the mentioned paper, the total number of oscillations in the wind speed have been counted, mainly over an interval of 10 minutes. All the observations refer to near neutral conditions, and we have considered those observations only with a wind direction for which there should be no influence of greater neighbouring obstacles. The result is given in Fig. 22.

The figure suggests that n may be expressed by a power function of the wind speed. LUMLEY and PANOFSKY (1964, p. 210) mention that there is some indication that there is a lower wind speed limit below which the eddies of the corkscrew variety may not form. However, supposing that this eventual wind speed limit is very small compared with actual wind speeds, we may apply a relation of the form

$$n = b\, U^c$$

There are most observations at the height of 15 m, and with the aid of the method of least squares, these observations have been fitted to a curve of the above mentioned form. It is found that $c = 0.56$ and that the coefficient of correlation is 0.96.

With a simple device we made some measurements at Norway Station of the frequency of the oscillation in the temperature at a height of 90 cm above the shelf ice. In a previous paper (VINJE 1967b, p. 85) these measurements have been

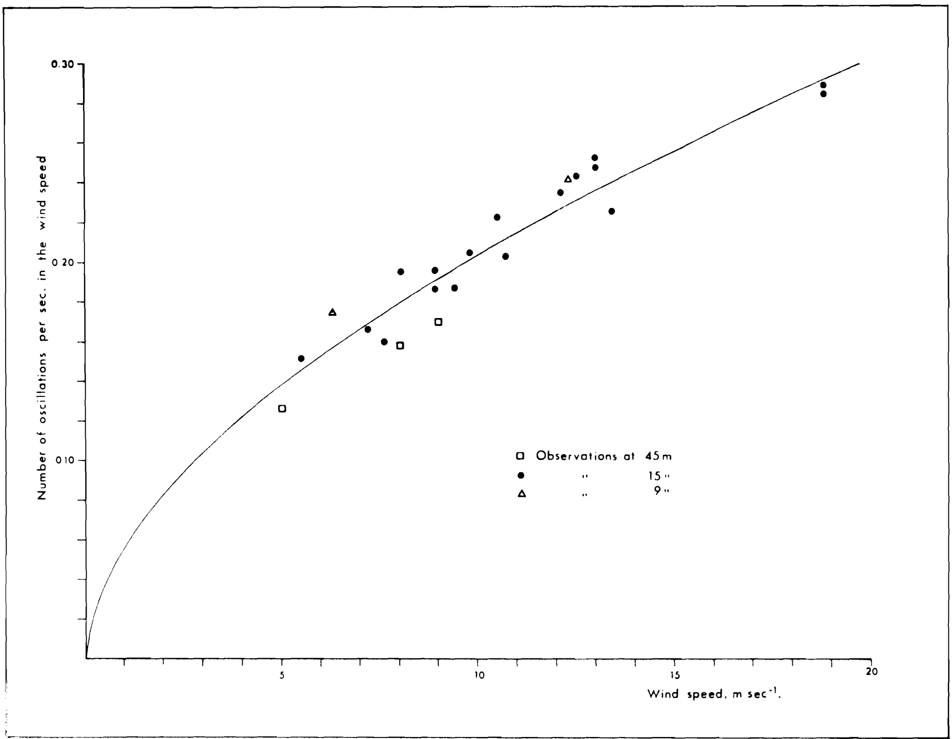


Fig. 22. *Number of oscillations in the wind speed with respect to the corresponding average wind speed, as can be evaluated from* GIBLETT *et al.'s registrations over level country.*

compared with measurements of the frequency of oscillations in wind speed and temperature made at other places at different heights. The comparison indicates that the exponent c is independent of height, and this appears also reasonable from the geometry.[1]

From eq. 6) it can be seen that $U_0 = r_0^{\varepsilon}/U_1$ and, according to the above hypothesis, we then may write, considering the height $z = r_0$, that

$$n_0 = b_0\, U_0^{0.56} = b_0\, (r_0^{\varepsilon}\, U_1)^{0.56}$$

Inserting this expression in eq. 9), we get for the average net vertical displacement of the elements in the surface vortices

$$r_0 = \left(\frac{K^2\, \varepsilon\, U_1^{0.44}}{2C\, k\, b_0}\right)^{\frac{1}{1-0.44\,\varepsilon}} \qquad 10$$

The net vertical displacement is thus given by profile constants and a product of unknown constants ($Cb_0$) which will be determined below.

[1] From the comparison mentioned a linear relationship was originally proposed. However, from the observations considered it is difficult to make a distinction between the two forms of relationship in question as they are very similar in the wind speed range considered. We have here kept to the exponential form as it is more convenient for the mathematical treatment of power profiles.

*c. Equations for the calculation of turbulent properties.* – As mentioned above, we made some measurements of the temperature profile near the surface at Norway Station, and from these measurements it was found that $r_0 \sim 0.15$ m for conditions with a wind speed of 6.0 m $sec^{-1}$ at the 10 m level and a net radiation loss from the surface of 0.050 ly $min^{-1}$. The profile observed will be more closely considered on p. 46, where it can be seen that it corresponds fairly closely to the conditions represented by profile 21 (Table 13, p. 41). The latter profile is characterized by the profile exponent $\varepsilon = 1 - 2K = 0.323$ and $U_1 = 3.2$ m $sec^{-1}$. Then when putting $r_0 = 0.15$ m for this profile, we obtain from eq. 10) that

$$2Ckb_0 = 0.32,$$

where $k = 1.2$ (cfr. p. 33). And we finally get an expression for $r_0$ which contains profile constants only, viz.:

$$r_0 = \left[3.2\, K^2 \varepsilon\, U_1^{0.44}\right]^{\frac{1}{1 - 0.44\varepsilon}} \qquad 11$$

When determining the product of the unknown constants in this way, it obviously involves an idealizing of the theory. Inaccuracies in the approximations and assumptions made to obtain the expression for $r_0$ are by this determination corrected for, thus, the theory fits on a single point, namely that $r_0$ is correct for a given condition.

Having determined $r_0$ from equation 11), $U_0$ is calculated from eq. 6), and we can determine the vertical transfer of momentum. From eq. 6) $(\delta U/\delta z)$ can be found, and when inserting this in eq. 7), it follows for the height $z = r_0$ that

$$\tau_0 = \rho\,(K\varepsilon\, U_0)^2 \qquad 12$$

This square equation is analogous to that suggested by Taylor (1916, p. 196) from an argument based on the measurements made by Stanton of fluid flowing through a pipe. It is remarked that in the above equation the drag coefficient $(K\varepsilon)^2$ is determined by the profile exponent alone.

For isothermal conditions with no snow drift we find that $\varepsilon \approx 0.14$, that is, nearly the same value as found for fluid ($\varepsilon = 1/7$). As $\varepsilon = 1 - 2K$, we get the drag coefficient equal to 0.0036 for the mentioned conditions. It is of interest to note that from the above mentioned measurements made by Stanton, the drag coefficient for the highest Reynolds numbers considered was found to be 0.004 when referring to a height "near" the surface.

Due to similarity in the mechanism of the vertical transfer of heat and momentum, we may expect that the profiles of temperature and wind speed are similar near the surface. Then, keeping to the measurements of the temperature at the lowest couple of metres, we may write

$$T = (T_1 - T_{00})\, z^{\varepsilon} + T_{00}$$

where $\varepsilon$ is found from the wind profile measurements made at "safe" distances from the surface vortices, and $T_{00}$ and $T_1$ are the absolute temperatures at the zero plane level and at the one m level, respectively. From the latter equation the temperature gradient is calculated, and from equations 1) and 2), p. 33, is then obtained for the net transfer of heat at the height $z=r_0$:

$$q_0 = c_p \left(\frac{T_1 - T_{00}}{U_1}\right) \tau_0 \qquad 13$$

Below we will make some calculations of $r_0$, $q_0$, and $\tau_0$ from the tentative equations derived above.

## TEST OF THE VORTEX THEORY

*a. The material applied.* – For the period 20.V–15.VII, 1958, we have selected 567 profiles of wind speed and temperature from the measurements on the tower at Norway Station. For the period mentioned the following profiles have been left out: those where the wind speed at the lowest level was indeterminable, i.e. below about 1 m sec$^{-1}$, (cf. p. 10), those where registration from one or more of the regular heights is missing, those where a maximum in the wind speed is observed at lower levels (cf. p. 28), and those where the net radiation is positive, which in most cases involves great dissimilarities in the temperature and wind profiles (cf. p. 12). Further we have left out the wind speed measurements at the lowest height (about 0.3 m) because these registrations revealed an unreasonably high wind speed. The reason for this may be due to multiple contacts (cf. p. 9). In the previous discussion it was found that a considerable part of the profiles revealed a pronounced maximum in the wind speed at approximately 30 m. We have therefore omitted the observations from the upper levels. The wind speed profiles will accordingly be determined from observations at three heights only (c. 3.2, 7.5, and 12.5 m). However, as can be seen from Table 12, p. 30, this reduction in the number of observation heights affect the profile exponent to a small extent.

We had no continuous registration of the wind speed at 10 m, and we have therefore calculated this wind speed for each of the 567 profiles by fitting the observations to a log+linear curve.[1] The profiles have then been grouped with respect to the following intervals of the wind speed at the 10 m level: 1.0–4.9, 5.0–9.9 and so on for every 4.9 m sec$^{-1}$ increase in the wind speed. The profiles within the mentioned groups have further been grouped with respect to intervals of 0.010 ly min$^{-1}$ in the average net radiation for the hour before the profiles were read. The conditions when the net radiation is zero will be specially considered. With the aid of the least square method the measurements have been fitted to a power function of the height. From these calculations the profile exponent and the wind speed at the one metre level have been determined. A condition for similarity in the exchange mechanism, and accordingly also in the profile

[1] The reason for this form of interpolation was that originally we intended to apply the MONIN/OBUKHOV theory. However, as it turned out that the registrations of the wind speed at the lowest level (c. 0.3 m) were dubious, this had to be abandoned. Registrations of wind speed at lower levels are necessary for the application of the mentioned theory.

Table 13

*Some quantities calculated from the profiles according to the vortex theory.*

$K$: The Kármán constant $=(1-\varepsilon)/2$, $\varepsilon$ is the wind profile exponent
$r_0$: net vertical displacement in the surface vortices
$U_0$: wind speed at height $z=r_0$
$U_0^*$ : friction velocity at height $z=r_0$
$(T_1-T_{00})$: air temperature difference between surface and one metre above
$q_0$: turbulent heat flux at height $z=r_0$
Q: net radiation at the surface

| Profile | $U_{10}$ m sec$^{-1}$ | $K$ | $r_0$ cm | $U_0$ m sec$^{-1}$ | $U_0^*$ cm sec$^{-1}$ | $(T_1-T_{00})$ °C | $q_0 x 10^3$ ly min$^{-1}$ | $-Qx 10^3$ ly min$^{-1}$ | Number of profiles |
|---|---|---|---|---|---|---|---|---|---|
| 4 | 3.3 | 0.432 | 10 | 1.8 | 10 | 0.0 | 0 | 0 | 26 |
| 5 | 3.0 | 0.384 | 11 | 1.1 | 9 | 1.0 | 10 | 7 | 13 |
| 6 | 3.6 | 0.367 | 11 | 1.1 | 11 | 1.2 | 13 | 16 | 12 |
| 7 | 3.5† | 0.312 | 10 | 1.1 | 8 | 3.4 | 24 | 26 | 14 |
| 8 | 4.1 | 0.274 | 7 | 0.4 | 5 | 1.8 | 7 | 36 | 15 |
| 9 | 3.9 | 0.302 | 9 | 0.6 | 8 | 1.7 | 12 | 44 | 23 |
| 10 | 3.7 | 0.292 | 8 | 0.5 | 6 | 1.6 | 8 | 57 | 14 |
| 11 | 4.1 | 0.277 | 7 | 0.5 | 6 | 1.3 | 5 | 65 | 4 |
| 16 | 7.5 | 0.429 | 15 | 4.1 | 25 | 0.0 | 0 | 0 | 50 |
| 17 | 7.0 | 0.422 | 15 | 3.6 | 24 | 0.4 | 9 | 7 | 26 |
| 18 | 8.2 | 0.420 | 16 | 4.2 | 28 | 0.6 | 16 | 17 | 19 |
| 19 | 7.1 | 0.385 | 16 | 2.8 | 24 | 1.1 | 30 | 25 | 24 |
| 20 | 7.1 | 0.377 | 16 | 2.6 | 24 | 1.3 | 36 | 35 | 29 |
| 21 | 6.8 | 0.338 | 15 | 1.7 | 19 | 1.6 | 35 | 45 | 35 |
| 22 | 7.4 | 0.380 | 16 | 2.7 | 25 | 1.7 | 49 | 54 | 30 |
| 23 | 7.9 | 0.380 | 17 | 3.0 | 27 | 1.3 | 41 | 66 | 34 |
| 24 | 8.8 | 0.388 | 18 | 3.6 | 31 | 0.9 | 33 | 75 | 23 |
| 28 | 12.5 | 0.434 | 18 | 7.4 | 42 | 0.0 | 0 | 0 | 32 |
| 29 | 12.3 | 0.446 | 16 | 7.9 | 38 | 0.2 | 8 | 7 | 14 |
| 30 | 12.1 | 0.448 | 15 | 7.9 | 37 | 0.1 | 4 | 15 | 11 |
| 31 | 12.3 | 0.433 | 18 | 7.2 | 42 | 0.7 | 29 | 26 | 5 |
| 32 | 11.6 | 0.435 | 17 | 6.9 | 39 | 0.7 | 25 | 34 | 6 |
| 33 | 12.6 | 0.430 | 18 | 7.2 | 43 | 0.7 | 30 | 48 | 7 |
| 34 | 11.6 | 0.425 | 18 | 6.4 | 41 | 1.3 | 53 | 56 | 12 |
| 35 | 11.3 | 0.422 | 18 | 6.0 | 40 | 1.4 | 56 | 64 | 9 |
| 36 | 11.6 | 0.426 | 18 | 6.4 | 40 | 1.6 | 65 | 79 | 20 |
| 40 | 17.8 | 0.450 | 18 | 12.0 | 53 | 0.0 | 0 | 0 | 33 |
| 52 | 22.0 | 0.450 | 20 | 14.9 | 67 | 0.0 | 0 | 0 | 16 |
| 64 | 28.2 | 0.442 | 24 | 18.4 | 93 | 0.0 | 0 | 0 | 7 |
| 65 | 32.4 | 0.440 | 27 | 21.0 | 111 | 0.0 | 0 | 0 | 4 |

† Extrapolated value. Due to an indicated maximum in the wind speed at about 7.5 m the profile constants have been determined from observations at the two lower heights only.

forms is, as mentioned p. 33, that the adiabatic temperature changes are negligible. When determining $(T_1-T_{00})$ with the aid of the wind profile exponent, we have therefore considered the temperature measurements at the two lowest levels only, c. 0.3 and 3.2 m, where the vertical displacements are relatively small. The values of $r_0$, $q_0$ and $U_0^*$ have then been calculated from equation 11), 12), and 13). The mentioned figures are noted in Table 13.

*b. The net vertical displacement.* – As for the net vertical displacement of an element in the surface vortices, it is seen from Table 13 that it assumes a value of about 10 cm for the lowest wind speeds and that it increases to 27 cm for the highest wind speed considered. It is remembered that the net vertical displacement equals half the vertical extension of a vortex, and LUNDE's measurements, referred to on p. 10, indicate that the average amplitude of the unevenesses on the surface is about 20 cm. Our calculations thus indicate that the average vertical extension of the surface vortices is about the same as the average amplitude of the unevenesses on the surface for the lower wind speeds. This seems reasonable as the air near the surface should show a tendency to follow the lower boundary and the vortices should show a tendency to form behind the relatively uniform unevenesses on the snow surface.

The variation of the net vertical displacement in the surface vortices with stability can be calculated from eq. 11). The corresponding values of $r_0$ and $\varepsilon$ are given in Table 14 below for $U_1 = 1$ m sec$^{-1}$.

Table 14

*The variation of the net vertical displacement ($r_0$) with the stability when $U_1 = 1$ m sec$^{-1}$.*

| $\varepsilon = 1-2\varkappa$ | 0.15 | 0.20 | 0.25 | 0.30 | 0.35 | 0.40 |
|---|---|---|---|---|---|---|
| $r_0$, cm | 7.0 | 7.9 | 8.3 | 8.2 | 7.7 | 7.0 |

It is seen that $r_0$ obtains a maximum value when $0.25 < \varepsilon < 0.30$. We denote this critical value with $\varepsilon_\chi$. The results obtained above refer to conditions with a given wind speed, whilst actually an inversion is generally formed in connection with a clearing up, a decrease in the wind speed, and an increase in the stability. Under such conditions the increase of $r_0$ with stability, indicated by Table 14 for $\varepsilon < \varepsilon_\chi$, should be reduced, whereas the fall of $r_0$ when $\varepsilon > \varepsilon_\chi$ should be increased. This apparent critical value of $\varepsilon$ for turbulent motions will be further discussed on p. 49 in connection with the calculation of the drag coefficient.

It follows from eq. 11) that $r_0 \to 0$ when $\varkappa \to 0$. This involves that the profile becomes linear which in turn indicates that the flow becomes laminar (SUTTON 1953, p. 37). This result thus gives a qualitative support for the variation of $r_0$ with stability.

In Fig. 23a some examples of the variation of r with height have been given. Profiles 16, 28, and 65 refer to isothermal conditions, and for these conditions it is seen that the characteristic length increases considerably with increasing wind speed for a given height. In the mixing length theory the corresponding characteristic length has been supposed to be independent of the wind speed for a given height for near neutral conditions. It may be argued that an increase of the vertical displacements with increasing wind speed is more plausible than a constancy.

Profile 21 refers to relatively stable conditions. From a comparison of this profile with profile 16 it can be seen that the effect of stability above the surface

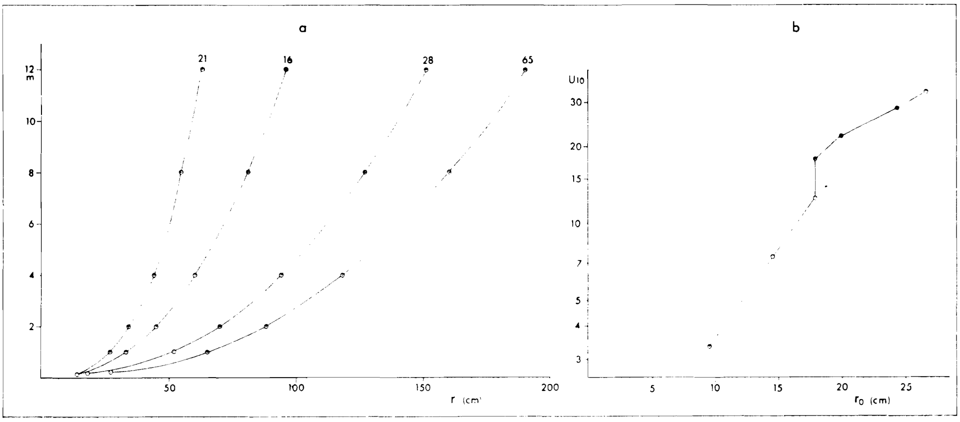


Fig. 23. *To the left: the variation of the net vertical displacement with respect to height. Reference numbers at the top refer to those given in Table 13. To the right: the variation of the net vertical displacement in the surface vortices with respect to log* $U_{10}$. *Near isothermal conditions.*

vortices is to reduce the net vertical displacements. The damping influence of the buoyancy forces on the net vertical displacements should decrease with height due to the corresponding reduction in the temperature gradient. The calculated, increasing reduction with height of r, when comparing the two mentioned profiles, might therefore reflect an increase in the reduction with height of the friction forces as the stability increases.

In Fig. 23b is given the calculated values of $r_0$ with respect to log $U_{10}$ for isothermal conditions. The figure reveals some interesting features. It can be seen, for example, that for wind speeds between about 12 and 18 m sec$^{-1}$ the net vertical displacement is constant near the surface. This constancy should be compared with the features revealed by Fig. 12, p. 18. It can there be seen that for wind speeds between about 11 and 17 m sec$^{-1}$ the observed maximum magnitude of the inversion is very nearly constant, in great contrast to the variation of this magnitude for lower wind speeds. The reduced importance of the wind speed upon the magnitude of the inversion together with the near constancy of the vertical displacements might reflect a damping effect of the drifting snow which in the considered wind speed interval increases considerably. As mentioned p. 18, BUDD et al. find that the density of the drifting snow, for example at 12.5 cm, on an average increases from 5 to 30 gr m$^{-3}$ when the wind speed increases from 12 to 18 m sec$^{-1}$. They find also that for a further increase in the wind speed the corresponding increase in the density of the drifting snow is markedly less. Then if the drifting snow has a damping influence upon the vertical displacements, the variation of $r_0$ with wind speed, as revealed by Fig. 23b, is in accordance with the corresponding variation in the density of the drifting snow.

It may be concluded from the above that the vertical displacement varies in a reasonable manner for different weather conditions. This, together with the fact that special features of the vertical displacement corresponds to other independent observations, lends a qualitative support for the vortex theory.

*c. The roughness length.* – According to the vortex theory it is $r_0$ which is the characteristic dimension of the turbulent motions near the surface, and it should also be remembered that the theory does not hold below the level $z = r_0$. However, to compare with other measurements we shall make some calculations of the roughness length, $z_{00}$, which is found from the logarithmic wind profile near the surface when assuming a constant shearing stress with height. We have

$$U = \frac{U_0^*}{K} \ln z + U_1$$

and

$$U = \frac{U_0^*}{K} \ln z + U_0 - \frac{U_0^*}{K} \ln r_0$$

when intergrating ($\delta U/\delta z = U_0^*/Kz$) between 1 and z cm and between $r_0$ and z cm respectively. The roughness length is then determined when putting

$$-\frac{U_0^*}{K} \ln z_{00} = U_1 = U_0 - \frac{U_0^*}{K} \ln r_0$$

From the vortex theory it follows that $U_0^* = K \varepsilon U_0$ (eq. 12), and we obtain then:

$$z_{00} = r_0 e^{-\frac{1}{\varepsilon}} \qquad 14$$

With the aid of this equation, $z_{00}$ has been determined for different wind speeds for isothermal conditions. The result is represented in Table 15, together with other values of the roughness length obtained from measurements in Antarctica. When making this comparison we have to bear in mind that $z_{00}$ is sensible to small variations in the profile exponent. From eq. 14) it thus can be seen that if $\varepsilon$ has an error of say 5%, $z_{00}$ will have an error of 40% for isothermal conditions (when $\varepsilon \approx 0.14$).

Table 15

*The roughness length (mm) for different wind speeds (m sec⁻¹).*

| Wind speed interval | No or slight to moderate snow drift | | | Moderate to heavy drifting snow | | | |
|---|---|---|---|---|---|---|---|
| | 0–5 | 5–10 | 10–15 | 15–20 | 20–25 | 25–30 | >30 |
| Norway Station | 0.06 | 0.12 | 0.09 | 0.01 | 0.01 | 0.04 | 0.06 |
| Maudheim (LILJEQUIST 1957) | 0.30 | 0.14 | 0.10 | 0.28 | 0.66 | 1.15 | 2.40 |
| Mirny (RUSIN 1961) | 0.37 | 0.13 | 0.07 | 0.01 | 0.17* | | |
| Bird Station (RUBIN 1966) | | | 0.13 | 0.21 | 0.27 | | |
| Amundsen-Scott (DALRYMPLE et al. 1963) | | 0.14 | | | | | |

* For $U > 20$ m sec$^{-1}$.

As can be seen, the low values of $z_{00}$ found at Norway Station are consistent with other measurements in Antarctica. For increasing wind speed there appears an increasing difference between the Norway Station values and those obtained at the other stations. For the determination of the roughness length at the other stations, apart from the South Pole Station, the logarithmic wind profile has been applied which was consistent with the measurements within the lowest 10 metres. From measurements in the considerably higher tower at Norway Station it is found that a power profile fits the observations better, and especially in the case of the higher wind speeds. The increasing discrepancy of the roughness length for increasing wind speed is therefore, in all probability, due to the difference in shape of the profiles applied.

Some values of the roughness length for different stabilities, as calculated from eq. 14), are given in Table 16.

Table 16

*The variation of $z_{00}$ with stability. Wind speed $U_1 = 1$ m sec$^{-1}$.*

| $\kappa$ | 0.425 | 0.400 | 0.375 | 0.350 | 0.325 | 0.300 |
|---|---|---|---|---|---|---|
| $z_{00}$, cm | 0.01 | 0.05 | 0.15 | 0.29 | 0.44 | 0.57 |

Consequently we find that the roughness length may increase considerably with increasing stability. SHEPPARD (1947, p. 219) considered the measurements made by BEST of wind profiles and gustiness in the lowest two metres over a shortgras surface. When assuming the Kármán constant equal to 0.40 for isothermal conditions, he obtained the following values:

| $z_{00}$, cm | 0.006 | 0.28 | 0.80 |
|---|---|---|---|
| $\kappa$ | 0.61 | 0.40 | 0.22 |

At the Amundsen–Scott Station, DALRYMPLE et al. (1963, p. 41) found an indication of a slight increase in the roughness length for extreme stabilities.

However, as for an eventual variation of $z_{00}$ with stability the different calculations and views are conflicting. Thus, according to the theory of LAIKHTMAN which is based on dimensional analysis, OGNEVA (1955, p. 34) found that $z_{00}$ decreases considerably with increasing stability, and in the MONIN/OBUKHOV theory, also based on dimensional analysis, the assumption is made that the roughness length is independent of the stability for a given surface. The discrepancies as revealed by the discussion above involve, however, not necessarily discrepancies in the outcome of the different theories. An eventual variation, or constancy, of the roughness length may be "counteracted" by the variation of other parameters which enter in the different theories.

In the literature is also found conflicting views as for an interpretation of the roughness length. LAIKHTMAN (1961, p. 52 in the trans.), for example, suggests that the roughness length might be taken as a characteristic dimension of the vertical motions at the height $z = z_{00}$ above the zero plane level. He mentions also that the decrease of $z_{00}$ with increasing stability, which follows from his theory, is unexpected and that this effect may be caused by an inadmissible extrapolation of the power profile. The latter view is in accordance with the considerations advanced in the present paper and it may be further supported by the following. As mentioned above (p. 10), the standard deviation of the unevenesses on the ice shelf is at Norway Station found to be 10 cm. This means that the zero plane level which intersects the unevenesses, partly must be found at the height of 10 cm above the snow surface when we consider the wind profile representative for a larger area. From measurements at different places in Antarctica the roughness length is found to be a fraction of a millimetre and it seems highly unrealistic to assume that the vertical motions at the height of 10 cm above the snow surface should be of such a small magnitude. It is also an observed fact that there occurs movements close to the snow surface below the top of the unevenesses. When considering the profile very close to the snow surface, the corresponding zero plane level will generally be tilting and the dimension of the local unevenesses responsible for its position above the snow surface will be of a considerable less order. This reasoning suggests accordingly a cut-off in the spectra of the dominant turbulent motions near the surface which are of importance for the profile form representative for a larger area.

*d. The heat transfer.* – It follows from the vortex theory that the turbulent transfer mechanism for momentum and heat should be equal near the surface. This involves similarity of the profiles at lower levels, and some measurements which support the existence of this similarity are shown in Table 17. With the aid of thermocouples, some measurements of the temperature profile between the surface and 89 cm were performed at Norway Station (VINJE 1964, p. 48). The average profile, based on 16 readings, was as follows:

Table 17

| | | | | | | |
|---|---|---|---|---|---|---|
| Height in cm | 0 | 1 | 2 | 3 | 4 | 6 |
| Temp. –°C | 28.62 | 28.01 | 27.86 | 27.76 | 27.74 | 27.45 |
| Height in cm | 9 | 14 | 19 | 29 | 44 | 89 |
| Temp. –°C | 27.54 | 27.46 | 27.63 | 27.30 | 27.04 | 26.80 |

The average wind speed at 10 m was 6.0 m $sec^{-1}$ and the average net radiation loss was 0.050 ly $min^{-1}$. The temperatures given reveal that a zigzag distribution is prevalent near the surface, and it was this feature which suggested the passing of vortices.

It was found in the paper mentioned (p. 36) that the heat exchange between air and snow may be executed by the ventilation of air in the upper layer of the permeable snow. This heat exchange presupposes a temperature difference between the surface and the adjacent air, and the temperature of the snow at the surface is therefore not included when fitting the above observations to a power function of the height. With the aid of the method of least squares, we have determined the value of the profile exponent for which the standard error of estimate is at its minimum. This minimum is obtained when choosing $\varepsilon = 0.311$ with the corresponding value of $(T_1 - T_{00}) = 1.53$. The calculated temperature of the air at the surface becomes $-28.30$°C, and as can be seen this is about 0.3°C higher than the temperature measured at the snow surface, and this difference is in accordance with the above mentioned presupposition.

As $\varepsilon = 1 - 2K$, it follows that $K = 0.345$ from the temperature measurements near the surface. For similar conditions, represented by profile 21, Table 13 p. 41, it can be seen from the wind profile measurements in the first 12 m, that $K = 0.338$. For profile 21 it can also be seen that $(T_1 - T_{00}) = 1.6$, which can be compared with the value 1.53 obtained from the profile measurements below one metre. This close agreement thus supports the view that the profiles are similar near the surface.

For stationary conditions an expression has been found for the vertical transfer of heat, viz.:

$$q = -0.54\,(U_{10})^{0.18}\,Q\ (\text{ly min}^{-1}),$$

where Q is the net radiation (ly min$^{-1}$) and $U_{10}$ is the wind speed (m sec$^{-1}$) at the 10 m level (VINJE 1964, p. 40). When making a comparison of the heat transfer, as calculated according to the vortex theory (eq. 13 p. 40) and that calculated according to the equation above, it is of importance to be sure that the different profiles in Table 13 represent stationary conditions. Under stationary conditions, for a given wind speed at a fixed level, the magnitude of the inversion (indicated by $(T_1 - T_{00})$ in Table 13), should increase with increasing net radiation loss from the surface. As seen from the table, this is only partly true for the two lowest wind speed intervals. And this indicates that profiles 8, 9, 10, 11, 23, and 24 do not represent stationary conditions. On p. 11 it was seen that stationary conditions are approached within a couple of hours after a clearing up. At least some of the profiles noted here may therefore refer to these time intervals when a relatively higher net radiation loss generally is observed (see e.g. VINJE 1964, p. 20). For such conditions the vertical turbulent transfer of heat in all probability is not fully developed, and this is in accordance with the corresponding, relatively low values of $q_0$ in Table 14. Then we consider the vertical transfer of heat for those conditions which in all probability represents approximate stationary conditions. The comparison of the different calculations is represented in Fig. 24.

As can be seen, there is a fair agreement and the validity of the vortex theory is thus quantitatively supported for stable stationary conditions.

*e. The friction velocity for isothermal conditions.* — In Fig. 25 our calculations of the friction velocity have been compared with values obtained from the logarithmic wind profile at other stations in Antarctica.

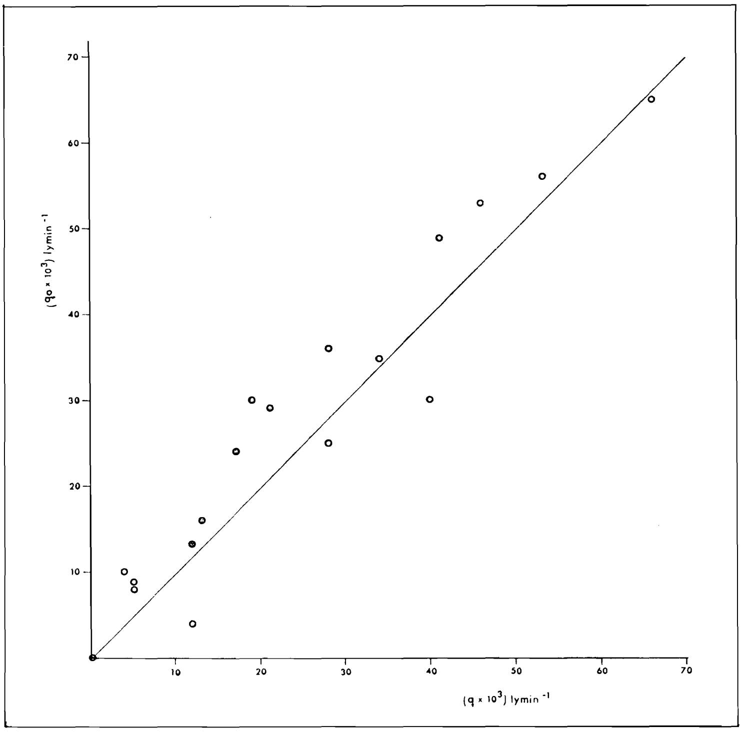


Fig. 24. *The calculated transfer of heat according to the vortex theory (ordinate) compared with the transfer of heat, as obtained from the consideration of the heat change in the upper layer of the snow under near stationary conditions. The straight line is that of complete accordance.*

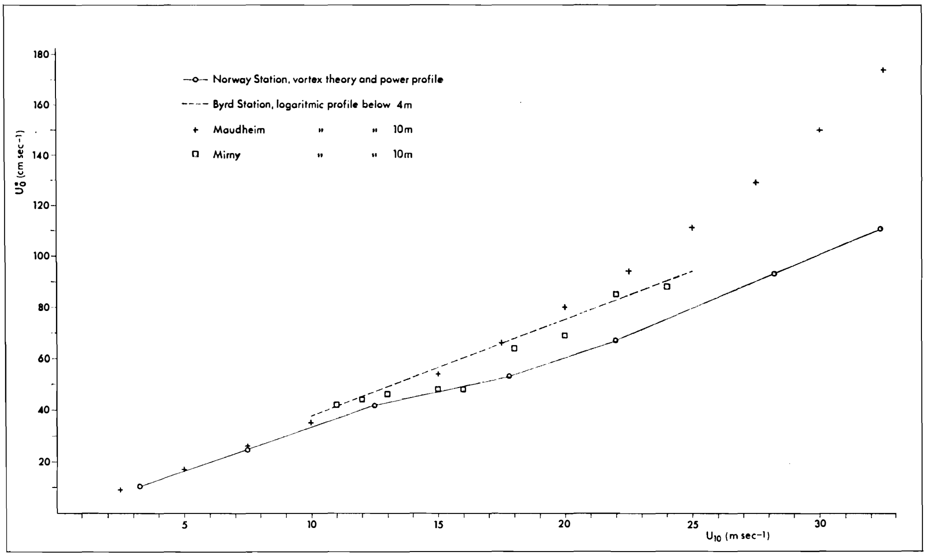


Fig. 25. *The relation between the friction velocity and the wind speed for near isothermal conditions, as obtained at different stations in Antarctica.*

The values for Maudheim have been given by LILJEQUIST (1957, p. 200), those for Mirny by RUSIN (1961, p. 75 in the trans.), and those for Byrd Station are given by BUDD et al. (in RUBIN 1966, p. 88). As can be seen from the figure, the values from Norway Station are generally lower than those given for the other stations, and the discrepancy is seen to increase with increasing wind speed. At the other stations the value of 0.40 has been applied for the Kármán constant, while at Norway Station a 7–12% higher value has been used. The effect of this is that if the same constant had been applied in both cases, the mentioned discrepancy would have been increased.

From the measurements performed at Byrd Station by BUDD et al. (in RUBIN 1966) it is found that the density of the snow drift, for example at 12.5 cm, on an average increases from 5 to 30 gr $m^{-3}$ when the wind speed ($U_{10}$) increases from 12 to 18 m $sec^{-1}$. As can be seen from Fig. 25, this marked variation in the density of the snow drift corresponds to a marked change in the relationship between the friction velocity and $U_{10}$, both for Mirny and Norway Station. The mentioned change may indicate that the effect of the drifting snow is to reduce the relative frictional drag of the variable surface on the air + snow stream. This reduction reflects the notable decrease in the wind profile exponent, as $U_{10}$ increases from about 12 to 18 m $sec^{-1}$ (cf. Table 12, p. 30). A decrease in this exponent means that the wind speed increases relatively less with height, i.e. that the frictional drag of the surface is, after the onset of drifting snow, relatively reduced. This result is in contrast with the arguments based on the observed increase of the roughness length with wind speed as calculated from logarithmic profiles. Thus LILJEQUIST (1957, p. 196) suggests that the snow drift may increase the friction against the surface, while BUDD et al. (in RUBIN 1966, p. 88) suggest instead that the increase in the roughness length is caused by genuine change in the surface roughness as the wind speed increases. At Mirny, RUSIN found no marked increase of the roughness length with increasing wind speed (RUSIN 1961, p. 74 in the trans.). And he argues (ibid., p. 73) that the surface roughness will only vary with the increase in the wind speed and drifting snow if the nature of the surface changes. In our opinion, the nature of the surface is changed when the snow drift increases. And, for a given wind speed, a surface with rigid obstacles should cause a greater resistance to the wind than a surface where the obstacles are indulgent to the friction forces. Our observations may also be supported by the observed fact that the roughness length decreases with increasing wind speed over a field of long grass (DEACON 1953, p. 17).

*f. The drag coefficient.* – It is of interest to give a survey of the variation with stability and wind speed of the turbulent transfer over a snow field, as suggested by the vortex theory. For this purpose we shall consider the drag coefficient at the one metre level ($C_1$) defined by

$$\tau_0 = \rho \, C_1 \, (U_1)^2$$

It follows from eq. 13) p. 40 that the turbulent transfer of heat then is given by

$$q_0 = \rho \, c_p \, C_1 \, (T_1 - T_{00}) \, U_1$$

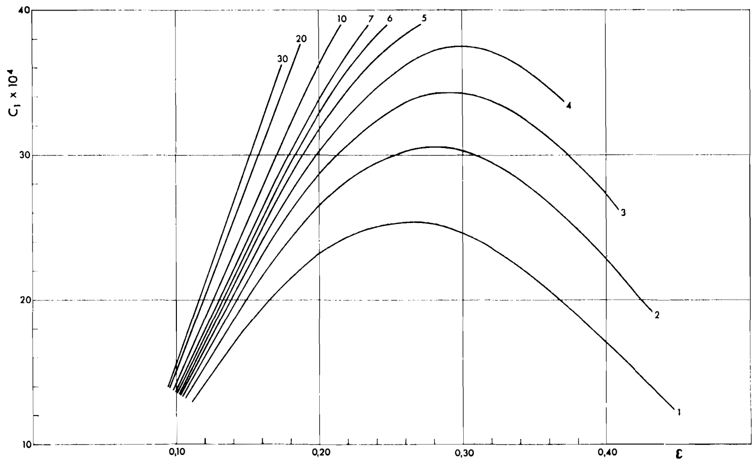


Fig. 26. *The relationship between the drag coefficient* ($C_1$) *and the profile exponent for different wind speeds* ($m\ sec^{-1}$) *at the one metre level.*

From equations 6), 11), and 12) (p. 34–39), we obtain that

$$(C_1)^{\frac{1}{2}} = K \varepsilon\, r_0^{\varepsilon} = K \varepsilon (3.2\, K^2\, \varepsilon\, U_1^{0.44})^{\frac{\varepsilon}{1-0.44\varepsilon}} \qquad 15$$

From this relationship, the drag coefficient has been calculated for different wind speeds and values of the profile exponent $\varepsilon = 1 - 2K$. The result is represented in Fig. 26.

Considering isothermal conditions, when ε is found to be about 0.13–0.14, it is seen that the variation of $C_1$ is relatively small. For the wind speeds $2 < U_1 < 10$ m $sec^{-1}$ it can for example be seen that $0.0018 < C_1 < 0.0022$. When determining the drag coefficient referring to the 10 m level ($C_{10}$) to compare with other measurements made in Antarctica, it is noted that $(U_1/U_{10})^2 \sim 0.5$ for isothermal conditions. Then $0.0009 < C_{10} < 0.0011$ over the ice shelf at Norway Station. At Maudheim, Liljequist (1957, p. 200) found that $C_{10} = (0.0024)/2 = 0.0012$ for a comparable wind speed range. And $C_{10} = 0.0012$ was also obtained by Dalrymple et al. (1963, p. 46) at the Amundsen–Scott Station for the conditions in question. There is thus a fair accordance between the results obtained from the vortex theory and other calculations. The increasing difference with decreasing wind speed is probably caused by the variation of $r_0$ with wind speed. At the other stations the corresponding characteristic dimension, the mixing length, has been assumed independent of wind speed.

As for the variation of the drag coefficient with stability, it can be seen from Fig. 26 that $C_1$ assumes a pronounced maximum for a critical value of ε, which lies between 0.25 and 0.30. From eq. 15) it is seen that $C_1$ is dependent upon $r_0$. And from Table 14, p. 42, it can be seen that there is a similarity in the variation of $C_1$ and $r_0$ with stability, and the above mentioned critical value of ε was there denoted by $\varepsilon_\chi$. The calculations suggest then that when ε increases beyond $\varepsilon_\chi$, the damping influence of the buoyancy forces on $r_0$ may be observed. A complete surpression of the turbulence is approached when $K \rightarrow 0$, i.e. when

$\varepsilon \to 1$. As known this corresponds to an assymptotical approach to linear profiles. The mentioned approach is clearly demonstrated by the observations of the temperature profile at Norway Station represented in Fig. 10, p. 16, when considering the lowest half decametre. This figure suggests, however, that a complete linearity in the profile can hardly come into existence over the ice shelf. This is in accordance with the result obtained by DALRYMPLE et al. (1963, p. 41), that if the requirement for truly laminar flow is assumed to be a strictly linear profile, a laminar flow never occurred at the South Pole.

From Fig. 10 it can also be seen that there is a marked difference in the shape of the temperature profiles numbered 1–3, which represent smaller stabilities, compared with those numbered 4–8, which represent greater stabilities. The corresponding, approximate critical value of $\varepsilon$ is from profile 4 found to be about 0.30 when considering the observations below 10 m. This critical value of $\varepsilon$ is seen to be similar to the critical value ($\varepsilon_\chi$) suggested from the calculations represented in Fig. 26. When considering Fig. 10, we then have that an increase in the gradients in the upper 2/3 of the tower is observed as $\varepsilon$ increases or decreases against $\varepsilon_\chi$. An increase of the gradients at these heights indicates that the influence of the colder surface increases. This observational fact therefore suggests that the turbulent transfer of colder air from the surface increases as $\varepsilon \to \varepsilon_\chi$. And this general feature is in accordance with the variation of the turbulent transfer with stability as suggested by the vortex theory.

## Summary

Some characteristic examples of the recordings of temperature and wind speed from the 37 m high tower at Norway Station are discussed. According to these graphs, it can be concluded that similarity of the temperature and wind speed profiles does not exist for the full height of the tower. There occurs sometimes for longer or shorter periods a marked maximum in the wind speed within the inversion.

When grouping the temperature profiles with respect to the temperature gradient between one and four metres, it is found that a marked change in the form of the profile occurs when the gradient becomes greater than about 0.5°C m$^{-1}$. For an increase in the mentioned gradient beyond this value, the stability above about 7 m decreases. The variation of the temperature profile form with increasing stability indicates that when the profiles for the whole range of stabilities are going to be fitted to a mathematical function of a special kind, one should keep to observations within the lowest half decametre only.

The maximum magnitude of the inversion in the tower is found to decrease rapidly as the wind speed at the 10 m level increases up to about 11 m sec$^{-1}$. For a further increase in the wind speed up to about 17 m sec$^{-1}$, there is a contrasting slow reduction in the maximum possible magnitude of the inversion. For higher wind speeds no inversions were registered.

It is observed that the inversion was deeper for westerly than for easterly winds. An analogous influence upon the wind speed profile was also observed.

It is found from radiosonde data that the inversion does not as a rule disappear

completely during the middle of the day in the summer season. This phenomenon is also suggested by the measurements taken from the tower.

The effect of stable stratification upon the wind profiles is in accordance with observations at other places. However, around noon in the middle of the summer the profiles are logarithmic and not concave towards the log z axis, as have been observed at other places. This may be due to the effect of a residual inversion in the upper part of the tower.

A considerable number of the wind profiles reveal a marked maximum at about 30 m above the surface, indicating a considerable reduction in the shearing stress with height. These profiles occurred far more frequently for westerly than for easterly winds. Their frequency decreases with increasing wind speed. A very special profile with a pronounced maximum and minimum within the lowest two decametres was observed on a few occasions.

The wind profile for isothermal conditions, referring to conditions with a zero net radiation, was found to be represented best by a power of the height. For conditions with no or slight snow drift, the power is 0.136–0.142, i.e. nearly the same as found for liquids flowing through pipes (0.142).

A tentative vortex theory for the turbulent transfer is advanced. The theory was suggested by the repeated observations at different places of a prevailing zigzag profile near the surface. The theory suggests that the Kármán constant may be determined by the wind profile, exponent ($\varepsilon$), viz.: $\varepsilon = 1 - 2K$. At the height of the center of the surface vortices ($z = r_0$) the drag coefficient is found to be a function of the profile exponent alone, viz.:

$$\tau_0 = \rho \, (K\varepsilon)^2 \, U_0^2$$

A characteristic dimension of the motions, the vertical extension of the vortices, is found to increase with the wind speed. And near the surface it is for lower wind speeds found to be about the same as the average amplitude of the unevenesses on the surface (20 cm).

A special expression for the roughness length is derived, and the calculated parameter is consistent with other measurements. For isothermal conditions the roughness length is found to be about 0.1 mm when there is no or slight drifting snow and about 0.01 mm when there occurs moderate to heavy drifting snow. These calculations thus suggest that a surface with rigid obstacles is less smooth than a surface where the obstacles are indulgent to the friction forces. The roughness length is found to increase considerably with increasing stability.

The calculated transfer of turbulent heat is found to be in good agreement with other calculations based on the heat budget of the surface. The friction velocity, on the other hand, is found to be somewhat less than the values determined from logarithmic profiles.

The drag coefficient for isothermal conditions is for small to moderate wind speeds found to be in fair agreement with other calculations. An increase is found for increasing wind speed, which probably is due to the corresponding increase of the net vertical displacements. A pronounced variation with stability is found, suggesting a maximum when the profile exponent is about 0.25–0.30. A similar critical value of the exponent is also indicated by the observed temperature profiles.

## Резюме

Обсуждаются температурные и ветровые профили, полученные на основе наблюдений, произведенных на башне высотой около 40 м., воздвигнутой на шельфовом леднике около станции Norway Station в Антарктиде. Они показывают большие различия в верхней части башни, где довольно часто повторяется максимум ветровой скорости приблизительно в 30 метрах над уровнем ледникового покрова. Это указывает на то, что напряжение сдвига может значительно уменьшаться в пределах трех нижних декаметров над шельфовым ледником. Устанавлено, что форма температурного профиля заметно изменяется по мере увеличивания термической устойчивости, и что формы как температурного профиля, так и профиля скорости верта меняются в зависимости от того, имеет ли ветер восточное или западное направление, в иначе одинаковых условиях. В изотермических условиях наблюдения за скоростью ветра могут быть лучше изображены силовым профилем, чем логарифмическим.

Предлагается вихревая теория турбулентного переноса. Эта теория, между прочим, сводится к тому, что постоянная Кармана может быть определена экспонентом силового профиля и, дальше, что она уменьшается при увеличиванием термической устойчивости. На средней высоте центра поземных вихрей, коэффициент трения ветра о землю определяется, по указанной теории, только профильным экспонентом.

Турбулентные свойства, вычисленные по данной теории, которая, к тому же, представляет возможное объяснение отмеченного близ земли зигзагообразного профиля, находятся в хорошем соответствии с другими вычислениями. При малых ветровых скоростях, среднее вертикальное протяжение поземных вихрей — той же величины, как средная амплитуда шероховатости поверхности (20 см).

Дальше выявлено, что начало поземка создает относительное уменьшение трения поверхности о воздушный поток. Вычисленный коэффициент трения имеет максимальное значение при профильным экспоненте, равным примерно 0,25—0,30. Эта же особенность согласуется с соответствующей вариацией температурных градиентов.



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