THE EDDY CONDUCTIVITY OF THE AIR OVER A SMOOTH SNOW FIELD

By *H. U. SVERDRUP*

(Manuscript received November 28th 1935)

1. Introduction.

The observations which will be dealt with here were taken on Isachsens's Plateau on West-Spitsbergen, lat. 79° 09′.0 N, long. 12° 56′.5 E Gr. at an altitude of 870 m above sea level. The party consisted of Professor H. W:son Ahlmann, Stockholm's Högskola, meteorologist H. Olsson, Stockholm, cand. mag. J. Knudsen, Bergen, and the author.

Professor Ahlmann, assisted by Mr. Olsson, undertook glaciological studies, partly on the Plateau and partly on the Fourteenth of July Glacier which leads down to Cross Bay on the western side of the Plateau. In connection with his studies Professor Ahlmann desired information as to the relative importance of the factors with influence the ablation of the snow fields and the glacier, viz., the radiation and the transfer of heat from the atmosphere. A study of the exchange of heat between the snow surface and the air is, independent of the application of the results to glaciological problems, of considerable interest to meteorology, being a special investigation within the wide field which deals with the exchange of heat between the surface of the earth and the air. The conditions for an investigation of this kind are especially favourable when undertaken over a snow field at a time when the surface is melting, since the temperature of the surface then remains constant at zero degree, and since the total amount of heat which the surface receives can be determined by measurement of the ablation of the snow.

The ablation was recorded by means of an apparatus constructed by O. Devik. Professor Ahlmann, who attended to the instrument, has published the results (Ahlmann, 1935). H. Olsson supervised the Robitsch actinograph which recorded the total incoming radiation from the sun and the sky, and undertook comprehensive measurements of the intensity of the radiation from the sun by means of Michelson's actinometer. He also measured the albedo of the snow and examined the penetration of radiation into the snow. By means of these data it is possible to compute the amount of heat which the surface receives because of processes of radiation, and, by subtracting this from the total amount as shown by the ablation, to find the amount which is received from the atmosphere.

In order to find the relation between the latter amount and wind velocity, air temperature and humidity, systematic meteorological observations were carried out by Mr. Knudsen and myself. We divided the day in watches since observations were to be taken every hour. Control observations and intercomparisons could be undertaken in the day-time when both of us were at work. Mr. Knudsen was an able mechanic, and it is mainly due to his skill that we succeeded in carrying out our programme, in spite of several mishaps which otherwise would have been serious.

The variation with elevation of the meteorological elements is of special importance, and observations were, therefore, made at three or four different levels. The observations are, on the whole, well suited for a study of the conduction of heat and water vapour in the layer directly above the ground, and these processes will be made the subject of the greater part of the following discussion.

It is, however, desirable to apply the results in general, but this can be done only if the problem of exchange within the layer of air next to the ground is discussed on a broad basis. Then the discussion cannot be confined to a consideration of exchange of heat and water vapour, but the exchange of momentum must be taken into account. The exchange of momentum is governed by the *eddy convectivity,* but the exchange of water vapour is governed by the *eddy conductivity*, which may differ from the eddy convectivity, and the exchange of heat depends upon the combined effect of eddy conductivity and radiative diffusivity. Furthermore, both convectivity and conductivity vary with increasing distance from the boundary surface and depend upon the wind velocity, the roughness of the surface and the stability of the stratification.

In order to solve the problem of exchange in the vicinity of the ground it is, therefore, necessary to answer the following questions:

1. Are eddy convectivity and eddy conductivity identical in the layer next to the surface? If not, is it possible to determine any relation between the two coefficients?
2. To what extent is the temperature distribution near the surface dependent upon eddy conductivity and to what extent upon radiative processes?
3. How do eddy convectivity and eddy conductivity depend upon the roughness of the surface and the stability of the air above the surface?
4. How do eddy convectivity and eddy conductivity vary with increasing distance from the boundary surface?

When dealing with these questions it is of great advantage to make use, as far as possible, of the results of laboratory experiments and of the theoretical considerations which are based on these results. Before turning to the observations from Spitsbergen we shall, therefore, give a brief review of the laboratory results and of the application of these results to meteorological problems.

2. Eddy Convectivity near a Boundary Surface.

A. Laboratory Results and Theoretical Considerations.

Smooth surfaces. The greater number of laboratory experiments have dealt with flow through pipes, but in a few cases the flow of air over a plane surface has been studied. The experiments have been undertaken in a wind tunnel where the air was in laminar motion before reaching the plate, which was suspended parallel to the direction of flow. Under these conditions a turbulent layer is formed along the plate, if the velocity of the laminar flow is sufficiently great. The thickness of this layer increases along the plate according to a law which can be written (Handbuch der Experimentalphysik IV, Teil 2):

$$\delta = 0.37\, x \left(\frac{\nu}{Ux}\right)^{\frac{1}{5}} = 0.37\, x\, R_x^{-\frac{1}{5}} \tag{1}$$

where x is the distance along the plate from the foremost rim, U the velocity of the laminar flow outside of the boundary layer, ν the kinematic coefficient of viscosity of the air and R_x the Reynold's number at the distance x. With $x = 40$ *cm* and $U = 20$ *m/sec.*, we obtain $\delta = 11$ *mm*. Thus, the turbulent layer is very thin.

Within the turbulent layer the average velocity increases with increasing distance from the plate according to a simple power law:

$$u = U\left(\frac{z}{\delta}\right)^{\frac{1}{n}} \tag{2}$$

where z is the vertical distance from the plate and where n is equal to 7 if the surface of the plate is smooth and the Reynold's number smaller than about 50 000. The value of n increases with the Reynold's number when this surpasses about 50 000, but is smaller if the surface is rough, other conditions being alike.

Within the turbulent boundary layer the tangential stress:

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

is supposed to be constant. Here η is the *eddy convectivity* according to Richardson's notation or the "Austauschgröße" according to W. Schmidt.

Resistance measurements have shown that in the case of a smooth surface:

$$\tau = 0.045 \frac{\varrho}{2} U^2 \left(\frac{\nu}{\delta U}\right)^{\frac{1}{4}},$$

where ϱ represents the density. By means of (2) and (3) we find:

$$\eta = 0.078\, n \varrho \frac{\delta^2}{x} U \left(\frac{z}{\delta}\right)^{\frac{n-1}{n}} = 0.0107\, n \varrho x R_x^{-\frac{2}{5}} U \left(\frac{z}{\delta}\right)^{\frac{n-1}{n}}$$

$$= 0.029\, R_x^{-\frac{1}{5}} n \varrho \left(\frac{\delta}{z}\right)^{\frac{2}{n}} u_z z.$$

The last form, in which we have introduced the velocity u_z at the level z (where $z < \delta$) instead of U, is of interest when the laboratory results are to be applied to meteorological conditions. Then the surface must be considered as having infinite extension;

the thickness, δ, of the turbulent boundary layer, loses its meaning, but the motion must, nevertheless, be characterized by a definite Reynold's number. In the formula for η we must, therefore, replace δ by another characteristic length which we may call h, and, supposing that the velocity distribution can be represented by a power law:

$$u = u_1 \left(\frac{z}{z_1}\right)^{\frac{1}{n}}, \tag{4}$$

we obtain:

$$\eta = C n \varrho \left(\frac{h}{z}\right)^{\frac{2}{n}} u_z z = C n \varrho \left(\frac{h}{z_1}\right)^{\frac{2}{n}} u_1 z_1 \left(\frac{z}{z_1}\right)^{\frac{n-1}{n}}, \tag{5}$$

where C is a dimensionless factor. This is the law which η should be expected to obey when dealing with meteorological problems, if analogy exists with the results from the laboratory, and if the surface could be considered smooth.

Influence of roughness. We have, until now, dealt with smooth surfaces, but, in nature, we encounter always rough surfaces and it is, therefore, necessary to consider the modifications which arise if the surface is rough. Conditions over rough surfaces have also been studied in the laboratory and Prandtl has suggested two different applications of the laboratory results to meteorological conditions. In his first paper dealing with meteorological problems (1924) he assumes that the variation of velocity with height is given by the power law (4), and that at the surface the stress is proportional to the square of the velocity at a distance h which characterizes the roughness of the surface:

$$\tau = C \varrho u_h^2 = C \varrho \left(\frac{h}{z}\right)^{\frac{2}{n}} u_z^2 . \tag{6}$$

Experiments by Hopf and Fromm have verified these assumptions, and from these experiments, which gave $n = 6.4$, the value $C = 0.0089$ was derived. From (3), (4) and (6) follows:

$$\eta = C n \varrho \left(\frac{h}{z}\right)^{\frac{2}{n}} u_z z. \tag{7}$$

This formula is exactly of the same type as formula (5), but now C appears as a universal, dimensionless constant and h has received a definite meaning.

Later investigations have, however, shown that the older measurements and a great number of new measurements all are brought in agreement by introducing a logarithmic law for the velocity increase instead of a power law (Handbuch der Experimentalphysik Bd. IV. 1 Teil). In a later paper (1932) Prandtl, therefore, starts with the expression

$$\eta = \varrho l^2 \frac{du}{dz}, \tag{8}$$

where l is "the mixing length." He avails himself of v. Kármán's theoretical result which has been verified by observations:

$$l = k_0 (z + z_0), \tag{9}$$

where $k_0 = 0.38$ and where $z + z_0$ is written instead of z since the mixing length has a definite value in the immediate vicinity of a rough surface. This value depends upon the roughness, and a relation, therefore, exists between the parameter z_0 and the linear dimension which characterizes the roughness. Calling this dimension h one has approximately $z_0 = \frac{1}{30} h$.

From (8), (9) and (3) follows:

$$\frac{du}{dz} = \frac{1}{k_0 (z + z_0)} \sqrt{\frac{\tau}{\varrho}}, \tag{10}$$

$$u = \frac{1}{k_0} \sqrt{\frac{\tau}{\varrho}} \ln \frac{z + z_0}{z_0} = c \log \frac{z + z_0}{z_0} \tag{11}$$

and

$$\eta = \varrho k_0 (z + z_0) \sqrt{\frac{\tau}{\varrho}} = \varrho k_0^2 (z + z_0) \frac{u_1}{\ln \frac{z_1 + z_0}{z_0}} \tag{12}$$

These are the forms in which the equations are written by Rossby and Montgomery (1935). Thus, the variation of velocity with height is represented by a logarithmic law instead of a power law. Prandtl remarks:

"Die Geschwindigkeitsverteilung nach Formel (11) stimmt weitgehend mit den bisher üblichen Potenzformeln überein, der Logarithmus ist ja bekanntermaßen der Grenzwert einer sehr kleinen positiven Potenz. Die Potenzformeln sind von unserem Standpunkt aus als Interpolationsformeln zu werten. Die Tatsache, daß man für verschiedene Verhältnisse verschiedene Potenzen $\left(\text{Exponent } \frac{1}{5} \text{ bis } \frac{1}{10}\right)$ bekommt, erklärt sich zwanglos durch verschiedene Größe von z_0, sowie auch durch thermische Einflüsse."

By means of (11) and (12) the influence of the roughness upon the eddy convectivity and the velocity distribution is shown. The logarithmic formula has the decided advantage that the influence of the roughness upon the velocity distribution appears in an explicit manner, whereas the roughness parameter does not appear in the equation for the velocity distribution if the latter is represented by a power law. One might, therefore, expect that the logarithmic law would have a wider application than the power law, but this is true only in case of indifferent equilibrium, since the velocity distribution in the vicinity of a boundary surface is modified by the stability.

Influence of stability. Richardson and Prandtl (1932) have shown that the influence of stability must depend upon the dimensionless ratio

$$-\frac{\frac{g}{\varrho}\frac{d\varrho}{dz}}{\left(\frac{du}{dz}\right)^2}$$

which, when we are dealing with the air, can be written:

$$\frac{\frac{g}{T}\frac{d\theta}{dz}}{\left(\frac{du}{dz}\right)^2},$$

where T is the absolute and θ is the potential temperature.

Rossby and Montgomery (1935) have, from a similar point of view, studied the influence of stability on the velocity distribution and the eddy convectivity, but their results do not apply to conditions in the immediate vicinity of a boundary surface and it is, therefore, necessary to consider these conditions more closely. When doing this we shall follow the lines of reasoning in Rossby's and Montgomery's paper.

The authors have first examined the distribution of velocity in the case of indifferent equilibrium, and have, with Prandtl, used v. Kármán's expression for the mixing length l. When turning to the effect of stability they write:

"Let us now consider the effect of stability in the surface layer, where the motion is simple in character and the stress is approximately independent of elevation. It is well known that a stable stratification tends to dampen the turbulent vertical movements of the air. At a certain elevation z above the ground we assume the mixing length to have the value l_s, whereas its value in the absence of stability would be

$$l = k_0(z + z_0). \tag{13}$$

"If the prevailing rate of shear is

$$C_s = \frac{du_s}{dz},$$

it follows that the turbulent kinetic energy per unit mass, $\frac{1}{2}(\overline{u'^2} + \overline{v'^2} + \overline{w'^2})$, must be proportional to $l_s^2\, C_s^2$. If the rate of shear were the same, but the stability were zero, the turbulent kinetic energy would be proportional $l^2\, C_s^2$. We assume that the difference now occurs in the form of potential energy, the turbulent elements at every instance having a density differing from the one prevailing in the surroundings. This potential energy must be proportional to (g/T) $(d\theta/dz)\, l_s^2$, where θ represents the potential temperature and g the acceleration of gravity. Thus:

$$l^2\, C_s^2 = l_s^2\, C_s^2 + \beta \frac{g}{T}\frac{d\theta}{dz} l_s^2, \tag{14}$$

where β is a proportionality factor as yet undetermined. It follows from the above equation that

$$l_s = \frac{l}{\sqrt{1 + \frac{\varkappa^2}{C_s^2}}}, \quad \varkappa^2 = \beta \frac{g}{T}\frac{d\theta}{dz}. \tag{15}$$

"Now let us compare a homogeneous and a stratified medium moving under the influence of the same stress. The rates of shear will vary and the two values will be designatad C_s and C. Assuming the expression

$$\frac{\tau}{\varrho} = l^2\left(\frac{du}{dz}\right)^2 \tag{16}$$

to be valid also in the case of a stratified medium, we obtain:

$$\sqrt{\frac{\tau}{\varrho}} = l_s\, C_s = l\, C. \tag{17}$$

Since the equation for l, (13), is independent of the rate of shear it is permissible to combine (15) with (17). The result is:

$$C_s = C\sqrt{1 + \frac{\varkappa^2}{C_s^2}}$$

"Solving this equation one obtains:

$$\frac{C_s}{C} = \sqrt{\frac{1}{2} + \frac{1}{2}\sqrt{1 + \frac{4\varkappa^2}{C^2}}} = v, \tag{18}$$

From this last equation combined with

$$C = \frac{d\,u_a}{d\,z} = \frac{1}{k_0(z+z_0)}\sqrt{\frac{\tau}{\varrho}}$$

it follows that

$$(19)\quad \frac{d\,u_s}{d\,z} = \frac{1}{k_0(z+z_0)}\sqrt{\frac{\tau}{\varrho}}\sqrt{\frac{1}{2}+\frac{1}{2}\sqrt{1+\frac{4\varkappa^2(z+z_0)^2}{\left(\frac{1}{k_0}\sqrt{\frac{\tau}{\varrho}}\right)^2}}}.\text{"}$$

In the last two equations u_s means the velocity in the stable case and u_a the velocity in the adiabatic case. Rossby and Montgomery write W_s and W_{ad} and use the letter u for the term which we have called v.

In their following discussion, Rossby and Montgomery introduce a constant lapse rate, since they assume $\varkappa$ to be constant. The assumption may be justified when treating conditions at some distance from the surface, but it is, undoubtedly, incorrect when dealing with the very lowest layer.

We shall instead assume that near the boundary surface the variation of temperature is similar to the variation af velocity, or if

$$(20)\quad \frac{d\,u_s}{d\,z} = m f(z), \text{ then } \beta\frac{g}{T}\frac{d\,\theta}{d\,z} = n f(z).$$

In order to justify this assumption we observe, that under stationary conditions

$$(21)\quad \tau = \eta_s\frac{d\,u_s}{d\,z} = \text{const. and } Q = c_p A_s \frac{d\,\theta}{d\,z} = \text{const.}$$

Here Q represents the amount of heat which is transported towards the surface, c_p represents the specific heat and A_s the eddy conductivity under stable conditions. Equation (20) is valid, if $\eta_s = A_s$ and it will later on be shown (p. 47) that this is correct.

Introducing (20) in (19) we obtain

$$(22)\quad \left(\frac{d\,u_s}{d\,z}\right)^3 - \left(\frac{1}{k_0(z+z_0)}\sqrt{\frac{\tau}{\varrho}}\right)^2\left(\frac{d\,u_s}{d\,z}+\frac{n}{m}\right) = 0$$

Thus, $d\,u_s/d\,z$ is determined by means of an equation of the third order and it is not possible to derive an exact analytical expression for u_s. We note, however, that if the term $4\varkappa^2/C^2$ is very great, we obtain

$$(23)\quad C_s = \frac{d\,u_s}{d\,z} \sim (z+z_0)^{-\frac{2}{3}}.$$

The condition: $4\varkappa^2/C^2$ very great, is fulfilled if the stability is great and the velocity small. Thus, we have that at great stability C_s is proportional to $(z+z_0)^{-\frac{2}{3}}$, but at indifferent equilibrium it is proportional to $(z+z_0)^{-1}$. As a suitable interpolation formula we can, therefore, introduce:

$$(24)\quad C_s = \frac{d\,u_s}{d\,z} \sim (z+z_0)^{\frac{1-n}{n}}, \text{ where } 3 < n < \infty .$$

This is in agreement with results from observations since these have shown that the interpolation formula (4), as a rule, gives a good approximation.

It follows:

$$(25)\quad \frac{d\,\theta}{d\,z} \sim (z+z_0)^{\frac{1-n}{n}}, \text{ where } 3 < n < \infty.$$

Consequently we assume:

$$(26)\quad \theta_z - \theta_0 = \triangle\,\theta_1\ (z+z_0)^{\frac{1}{n}}$$

where $\triangle\,\theta_1$ has the dimensions ${}^oC\,cm^{-\frac{1}{n}}$. It will be shown (p. 32) that (26) gives a good approximation to observed conditions.

Now we have:

$$(27)\quad \varkappa^2 = \beta\frac{g}{T}\frac{\triangle\,\theta_1}{n}(z+z_0)^{\frac{1-n}{n}} = \alpha^2(z+z_0)^{\frac{1-n}{n}}.$$

With this value we obtain:

$$(28)\quad \frac{C_s}{C} = \sqrt{\frac{1}{2}+\frac{1}{2}\sqrt{1+\frac{4\,\alpha^2\,(z+z_0)^{\frac{n+1}{n}}}{\left(\frac{1}{k_0}\sqrt{\frac{\tau}{\varrho}}\right)^2}}} = v$$

or

$$\frac{d\,u_s}{d\,v} = \frac{n}{n+1}\frac{2}{k_0}\sqrt{\frac{\tau}{\varrho}}\left(2+\frac{1}{v^2-1}\right).$$

Integration between the levels z_2 and z_1 gives:

$$u_{s_2} - u_{s_1} = \frac{n}{n+1}\frac{2}{k_0}\sqrt{\frac{\tau}{\varrho}}\left[2\,(v_2-v_1) - \frac{1}{2}\ln\frac{v_2+1}{v_2-1}\frac{v_1-1}{v_1+1}\right].$$

This formula differs from the corresponding formula by Rossby and Montgomery by the factor 2 and the different meaning of v.

Following the procedure of Rossby and Montgomery we remark that at $z=0$ we have, because of the smallness of z_0:

$$v = 1 + \frac{1}{2}\frac{\alpha^2\,z_0^{\frac{n+1}{n}}}{\left(\frac{1}{k_0}\sqrt{\frac{\tau}{\varrho}}\right)^2},$$

which differs only slightly from *1*. If we integrate between the ground and the level z we obtain, therefore:

$$u_s = \frac{n}{n+1}\sqrt{\frac{\tau}{\varrho}}\left[2(v-1) - \frac{1}{2}\ln\frac{v+1}{v-1} + \ln\frac{\frac{2}{k_0}\sqrt{\frac{\tau}{\varrho}}}{\alpha z_0^{\frac{n+1}{2n}}}\right],$$

or, since

$$\frac{\frac{1}{k_0}\sqrt{\frac{\tau}{\varrho}}}{\alpha} = \frac{(z+z_0)^{\frac{n+1}{2n}}}{v\sqrt{v^2-1}} \quad \text{and} \quad u_a = \frac{1}{k_0}\sqrt{\frac{\tau}{\varrho}}\ln\frac{z+z_0}{z_0},$$

we finally get:

$$u_s = u_a + \frac{n}{n+1}\frac{2}{k_0}\sqrt{\frac{\tau}{\varrho}}\left[2(v-1) - \ln\frac{v(v+1)}{2}\right]. \tag{29}$$

With $n=1$ we obtain Rossby's and Montgomery's solution. The character of the solution is more clearly seen by examining the cases when $4\varkappa^2/C^2$ is very small or very great.

If $4\varkappa^2/C^2$ is very small we obtain:

$$\frac{d u_s}{dz} = \frac{1}{k_0(z+z_0)}\sqrt{\frac{\tau}{\varrho}}\left[1 + \frac{1}{2}\frac{\alpha^2(z+z_0)^{\frac{n+1}{n}}}{\left(\frac{1}{k_0}\sqrt{\frac{\tau}{\varrho}}\right)^2}\right]$$

or

$$u_s = u_a + \frac{n}{2(n+1)}\frac{\alpha^2}{\frac{1}{k_0}\sqrt{\frac{\tau}{\varrho}}}(z+z_0)^{\frac{n+1}{n}}. \tag{30}$$

The eddy convectivity is:

$$\eta_s = \frac{\eta_a}{v} = \varrho k_0 (z+z_0)\sqrt{\frac{\tau}{\varrho}} - \varrho k_0^3 \frac{\alpha^2}{2\sqrt{\frac{\tau}{\varrho}}}(z+z_0)^{\frac{2n+1}{n}}. \tag{31}$$

These solutions are not valid when $z=0$ since the formula (26) must be regarded as an interpolation formula. Furthermore, the equations are valid only to a level below which

$$\frac{4\alpha^2(z+z_0)^{\frac{n+1}{n}}}{\left(\frac{1}{k_0}\sqrt{\frac{\tau}{\varrho}}\right)^2} = 4\beta\frac{g}{T}\left(\frac{\ln\frac{z_1+z_0}{z_0}}{u_{a_1}}\right)^2\frac{\triangle\,\theta_1}{n}(z+z_0)^{\frac{n+1}{n}}$$

is a small quantity. Below this level the variation of the velocity deviates but little from the logarithmic law.

If the term $4\varkappa^2/C^2$ is great we obtain, on the other hand:

$$\frac{d u_s}{dz} \rightarrow \sqrt{\frac{\alpha}{k_0}\sqrt{\frac{\tau}{\varrho}}}(z+z_0)^{\frac{1-3n}{4n}},$$

which gives:

$$u_{s_2} - u_{s_1} \rightarrow \sqrt{\frac{\alpha}{k_0}\sqrt{\frac{\tau}{\varrho}}}\left(z_2^{\frac{1+n}{4n}} - z_1^{\frac{1+n}{4n}}\right). \tag{32}$$

This means that at greater elevation the variation of the velocity within a given interval of altitude is represented by a power law, supposing the stability to be sufficiently great:

$$u_s = u_{s_1}\left(\frac{z}{z_1}\right)^{\frac{1+n}{4n}}. \tag{33}$$

If we claim that the variation of velocity is similar to the variation of temperature we must have

$$\frac{1}{n} = \frac{1+n}{4n}$$

which, in agreement with (23), gives

$$n = 3 \quad \text{and} \quad u_s = u_{s_1}\left(\frac{z}{z_1}\right)^{\frac{1}{3}}.$$

Then the eddy conductivity is:

$$\eta_s = \eta_{s_1}\left(\frac{z}{z_1}\right)^{\frac{2}{3}}. \tag{34}$$

On the assumption of a constant lapse rate Rossby and Montgomery also found $u_s \rightarrow u_a$ when $z \rightarrow 0$, whereas at greater altitudes they found

$$u_s \rightarrow u_{s_1}\left(\frac{z}{z_1}\right)^{\frac{1}{2}}, \quad \eta_s \rightarrow \eta_{s_1}\left(\frac{z}{z_1}\right)^{\frac{1}{2}},$$

but our formulae (33) and (34) ought to give much better approximations to the actual conditions near a boundary surface.

In general the eddy convectivity is:

$$\eta_s = \frac{\varrho k_0 (z+z_0)\sqrt{\frac{\tau}{\varrho}}}{\sqrt{\frac{1}{2} + \frac{1}{2}\sqrt{1 + \frac{4\alpha^2(z+z_0)^{\frac{n+1}{n}}}{\left(\frac{1}{k_0}\sqrt{\frac{\tau}{\varrho}}\right)^2}}}}. \tag{35}$$

Thus, η_s is no longer a linear function of elevation but decreases more slowly. When the stability is very great and the frictional stress small the formula (34) gives the best approximation.

Above, only the effect of a stable stratification has been discussed. A corresponding discussion of the influence of an unstable stratification has not yet been undertaken, but we will assume that under unstable conditions (11) and (12) are valid in the vicinity of the bundary surface.

In the preceding discussion the introductien of the mixing length is of fundamental importance. Sutton (1934) has treated the problem of exchange by turbulence from a different point of view. Following G. I. Taylor's procedure in the paper on "Diffusion by Continuous Movements", Sutton introduces the correlation coefficient r_ξ between the vertical motion $w'(t)$ associated with a mass of fluid at the time t and the vertical motion $w'(t+\xi)$ associated with the same mass at the time $(t+\xi)$. Furthermore, he defines as usual $\eta = \varrho \overline{w' l}$. Taylor has shown that:

$$\overline{w' l} = w'^2 \int_0^{t_0} r_\xi \, d\xi ,$$

where t_0 is the time taken by the eddy to transverse the vertical distance l. Introducing

$$r_\xi = \left(\frac{\nu}{\nu + w'^2 \xi}\right)^p ,$$

where ν is the kinematic viscosity and p a constant between 0 and 1, and, making use of Prandtl's and v. Kármán's equation

$$\left|\overline{w'}\right| = l \left|\frac{d\bar{u}}{dz}\right| ,$$

and of the fact that the eddy velocities are distributed according to Maxwell's law, (Hesselberg and Bjørkdal), Sutton obtains:

$$\overline{w' l} = \frac{\nu^p}{1-p}\left[\frac{1}{2}\pi l^2 \left|\frac{d\bar{u}}{dz}\right|\right]^{1-p} .$$

The essentially new in this result is the definition of r_ξ which is based on physical considerations and makes possible an extended use of Taylor's definition of l, the "mixing length," by means of the correlation coefficient r_ξ.

In (36) Sutton introduces v. Kármán's expression:

$$l = k_0 \frac{\left|\frac{d\bar{u}}{dz}\right|}{\left|\frac{d^2\bar{u}}{dz^2}\right|} ,$$

which is derived on the assumption that the eddy velocities at different points in the region are dynamically similar. Here k_0 is a dimensionless constant the value of which is approximately 0.4. (Rossby and Montgomery use the value 0.38) Sutton finally obtains:

$$(37) \quad \eta = \frac{(0.251)^{1-p}}{1-p} \varrho \nu^p \left\{\left(\left|\frac{du}{dz}\right|\right)^3 \left(\left|\frac{d^2u}{dz^2}\right|\right)^{-2}\right\}^{1-p} .$$

Since in the lowest layer the tangential stress is constant, Sutton obtains, on the assumption that $u = 0$ when $z = 0$:

$$u = u_1 \left(\frac{z}{z_1}\right)^{\frac{p}{2-p}} = u_1 \left(\frac{z}{z_1}\right)^{\frac{1}{n}}$$

in agreement with observations.

Laboratory experiments have, as previously mentioned, given $n = 7$. If we especially consider conditions over a plane surface where

$$u = U\left(\frac{z}{\delta}\right)^{\frac{1}{7}} ,$$

introduce this value in (37) and consider (1) we obtain from Sutton's formula

$$\eta = 0.065\, x R_x^{-\frac{2}{5}}\, U \left(\frac{z}{\delta}\right)^{\frac{6}{7}} ,$$

which differs from the formula derived from laboratory experiments only by the numerical constant (0.065 instead of 0.075, see p. 6). In general, Sutton's formula can be written:

$$(38) \qquad \eta = \left(0.251 \frac{n}{(n-1)^2}\right)^{\frac{n-1}{n+1}} \frac{n+1}{n-1} \nu^{\frac{2}{n+1}} \varrho (u_1 z_1)^{\frac{n-1}{n+1}} \left(\frac{z}{z_1}\right)^{\frac{n-1}{n}} .$$

When deriving this formula, the only assumption made as to n is that n must be greater than unity.

Sutton points out that n may depend upon the roughness of the surface and upon the stability, but considers n independent of u_1. It appears, however, more promising to attempt a separation of the effects of roughness and stability as has been done in the preceding discussion, and we shall, therefore, not make use of Sutton's formulae.

It must finally be mentioned that Köhler (1933), in agreement with Sakakibara, has introduced a complex value of $\eta : \eta = \eta_1 + i\eta_2 = \eta' e^{i\mu}$. Köhler arrives at the peculiar result that the character of the velocity distribution is essentially different when μ increases or decreases by 2π although, according to his definition, η is identical in the two cases. He has, however, treated his original equation in such a manner that he obtains the velocity near the ground represented by a multi-valued function, which must be without physical significance. In a second paper (1935) he has explained some of his assumptions more fully, but he has not entered upon the essential question: the validity of the many valued solution. His suggestions as to the character of the turbulence near the ground need not, therefore, be considered.

The result of this brief review can be summarized as follows:

According to experiences in the laboratories and to theoretical considerations it is probable that the increase of wind velocity near the surfaces is a logarithmic function of elevation if adiabatic conditions prevail. In this function a parameter enters which characterizes the roughness of the surface.

When the stratification is stable the rate of increase of the velocity is greater than under adiabatic conditions. Within any given interval of altitude the variation with height can be represented approximately by a power law, but the exponent probably increases with increasing distance from the surface. Thus, the deviation from the velocity distribution under adiabatic conditions increases with increasing elevation and at a given elevation it increases with increasing stability, other conditions being alike.

The eddy convectivity is under adiabatic conditions a linear function of the distance from the ground and, at a given elevation, it is a linear function of the wind velocity. In case of stable stratification it increases more slowly with elevation and is, at a fixed level, a more complicated function of the wind velocity.

We have now to examine if these conclusions are supported by observations.

B. The Eddy Convectivity of the Air near the Ground.

Variation with elevation. From the results of the preceding discussion it is evident that an examination of the eddy convectivity near the ground can be undertaken in two stages:

1. Investigation of the law according to which the eddy convectivity varies with elevation;

2. Determination of numerical values of the eddy convectivity at a fixed level under varying conditions.

The first investigation is reduced to an examination of the variation of the velocity with elevation, since at very low levels the frictional stress must be supposed to be constant, for which reason the eddy convectivity is inversely proportional to the rate of shear.

The variation of velocity with elevation has been studied in so many cases that it is impossible to quote the entire literature dealing with this subject. In general, it has been found that the variation can be represented either by a power law of the type (4) or by a logarithmic law, which different authors have given in slightly different forms. The influence of the roughness of the surface upon the velocity variation has, however, not been examined, but the influence of the lapse rate has been studied by several authors. The last-named examination has, however, had a purely empirical character. The first attempt on a rational treatment of all available data in the light of the latest theoretical results, was made quite recently by Rossby and Montgomery (1935).

Rossby and Montgomery commence with the equations by Prandtl and v. Karmán (see p. 7) and on the assumption that these are valid under adiabatic conditions, they derive the logarithmic law (11). They show that a great number of the available observations are in agreement with this simple law, and obtain values of the roughness parameter, z_0, which are in good agreement with the character of the surface above which the wind velocity has been measured.

They found the following values:

Over a smooth lawn (observations by Hellmann) .... $z_0 = 0.54$ cm.
Over open grass land (observations published by Shaw) .... $z_0 = 3.2$ »
Over moderately smooth sea (observations by Wüst) .... $z_0 = 3.9$ »

Taylor's determination of the frictional stress at an altitude of 30 metres above Salisbury Plain affords an additional check. Taylor found (1916):

$$\tau = \varrho \gamma^2 u_{30}^2$$

where the values for γ^2 range between 22×10^{-4} and 32×10^{-4}. From (3), (11) and (12) one obtains:

$$\gamma^2 \frac{k_0^2}{\ln \frac{3000}{z_0}}$$

which with $z_0 = 0.5$ cm gives $\gamma^2 = 19 \times 10^{-4}$ and with $z_0 = 3.2$ cm gives $\gamma^2 = 31 \times 10^{-4}$. Thus, the values of z_0 which were found over grass land give very nearly the range in γ^2 which was determined by Taylor.

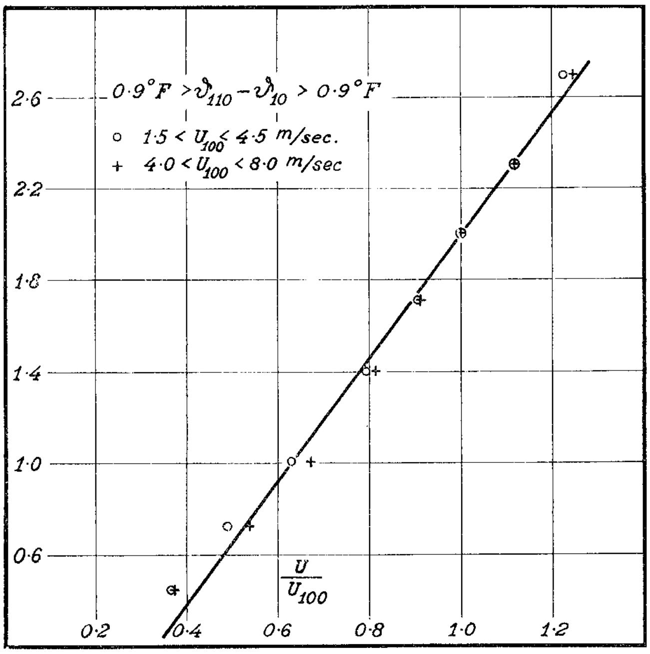


Fig. 1. Relative velocity as function of log z according to Best's observations.

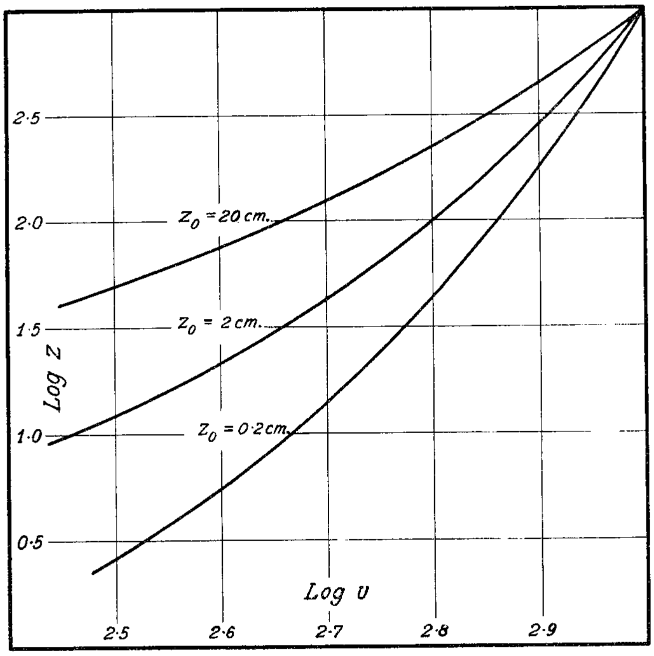


Fig. 2. Logarithms of the velocity plotted against log z assuming a velocity of 10 m/sec. at 10 m and different values of the roughness parameter.

Very detailed observations below 5 metres have recently been published by Best (1935). The observations were made at the levels 2.5, 5, 10, 25, 50, 100, 200 and 506 cm above a level cricket field, where for a distance of at least 10 m upwind the height of the grass was only 0.5 to 1 cm while beyond this distance the grass was still short. The temperature gradient between 10 cm and 110 cm was measured simultaneously. Best shows that with no temperature difference between these two points the wind velocity is practically a linear function of log (z—1), where (z—1) represents the distance over grass approximately 1 cm high. If the velocity at 1 m lies between 1.5 and 4.0 m/sec. Best obtains:

$$u_z = u_{100}\,[0.345 \log (z-1) + 0.30],$$

but the observations can also be represented by an equation of the type (11).

In fig. 1 two sets of observations are plotted and it is seen that all values fall near the straight full-drawn line, which give an average roughness parameter, $z_0 = 0.22$ cm. The very lowest observations show the greatest deviation from the straight line. Considering the difficulty of accurate measurements in the immediate vicinity of the ground, where the velocity gradient is very great, it is perhaps permissible to attribute less weight to the lowest observations. Doing this, the other data are best represented by a line which gives $z_0 = 0.14$ cm. Over a close-cropped and very smooth lawn the roughness parameter should, therefore, have a value between 0.14 cm and 0.22 cm.

Other authors have generally found that the variation of velocity with elevation is better represented by a power law but this is not surprising, considering that a power law gives a good approximation if no observations are available from the very lowest level, and that no attention has been paid to the stability. Before turning to the effect of stability we shall examine the first point somewhat closer.

Best has pointed out that if the velocity variation actually is represented by a logarithmic law, and if one attempts to use a power law within small intervals of elevation, one obtains values of n which increase with elevation. In order to illustrate this feature the following computation has been undertaken: The vertical distribution of velocity has been computed, assuming the velocity at 10 m to be 10 m/sec. and the roughness parameter to have the values 0.2 cm, 2.0 cm and 20.0 cm respectively. In fig. 2 the logarithms of the velocities are plotted against log z. We should obtain straight lines if the power law (4) were valid but we obtain curved lines. If we still wish to represent the velocity by

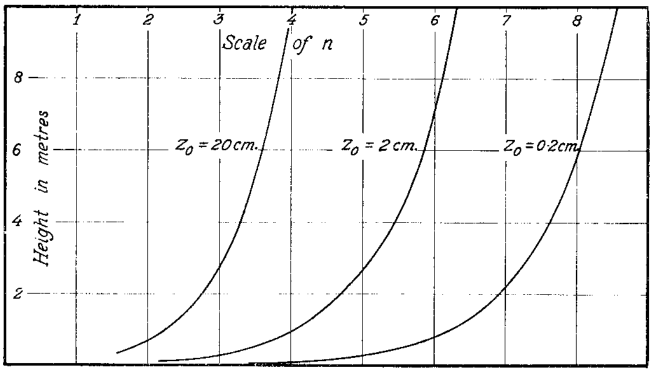


Fig. 3. The value of n as function of height over surfaces of different roughness.

a power law, we have to consider n as a function of z. From the shape of the curve it is evident that n increases with elevation. The value at any level is determined by the conditions that the tangents of the curves:

$$u = u_1 \left(\frac{z}{z_1}\right)^{\frac{1}{n}} \text{and } u = \frac{1}{k_0}\sqrt{\frac{\tau}{\varrho}}\,\ln\frac{z+z}{z_0}$$

shall coincide. This condition gives:

$$n = \frac{z+z_0}{z}\ln\frac{z+z_0}{z_0}.$$

In fig. 3 n is represented as a function of elevation up to 10 metres for $z_0 = 0.2$ cm, 2.0 cm and 20 cm. From this figure it is evident that observations which are taken above a level of 1 to 2 metres can be approximately represented by a power law with a constant value of n, since the variation of n with altitude is slow above these levels. The approximation is the better the smaller the interval is within which the observations are taken. It is, however, clear that such a formula has the character of an interpolation formula only, and must not be used for extrapolation. Further, it is evident that within a given interval of altitude, the value of n decreases with the roughness of the surface.

These features explain, to some extent, the different values of n which have been found by different investigators. Attention is especially drawn to the values of n which were derived by W. Schmidt (1929) from his observations over ground of varying character. His observations at Hommelsheim were undertaken above a stubble field after wheat and above a field of turnips where the upper parts of the plants where at a level of 40 to 50 cm. The numerical results are not given but some mean velocities are entered on a graph. According to this, the mean values apply to the levels 25 cm, 78 cm and 130 cm. In the case of the turnip field the heights are reckoned from the general top level of the plants. Schmidt shows that the variation of the velocity with height, approximately follows a power law and finds in the case of the turnip field $n = 2.4$ and in that of the stubble field $n = 4.2$. Regarding the former value, he remarks that this is surprisingly low and that probably no other investigation has given such a low value. In view of our preceding discussion, the different n-values can, however, be explained if the widely different roughness of the ground is taken into account. Schmidt's values are, actually, represented almost as well by logarithmic laws which in the case of the stubble field give z_0 about 0.8 cm and in that of the turnip field give z_0 between 3 and 4 cm. The former value is somewhat greater than the value $z_0 = 0.5$ cm which, according to Rossby and Montgomery, applies to a smooth lawn and the latter is nearly equal to the value which, according to the same authors, is found over grass land. The temperature distribution is not known, but, at these low levels, this should not alter the results materially. Thus, all available evidence supports the conclusion arrived at by Rossby and Montgomery that under adiabatic conditions the variation of velocity with height follows a logarithmic law.

Where stability is concerned, Rossby and Montgomery find that at some distance from the surface the difference between the velocities at two levels, z_2 and z_1 shall approach the value $c\,(z_2^{\frac{1}{2}} - z_1^{\frac{1}{2}})$. They have compiled a considerable amount of evidence which supports this result, but very few of the observations which they use are taken within the very lowest interval of altitude in which we are, at present, especially interested. Within this interval the logarithmic law is probably valid up to a small height, and, at somewhat greater height, the difference between the velocities at two levels should, according to our results, approach the value $c'(z_2^{\frac{1}{3}} - z_1^{\frac{1}{3}})$.

From Best's observations it is evident that the velocity distribution below 0.5 metre is little influenced by stability. Between 0.5 and 2 metres the velocity distribution can be represented by a power law, and the value of n decreases with increasing stability, but the smallest value of n is 5.25.

With regard to instability, Best's data indicate that the influence upon the velocity distribution below 1 metre is very small but above 1 metre the velocity increases more slowly than in the adiabatic case.

From the preceding results as to the variation of wind velocity with height, we conclude that the eddy convectivity is, under adiabatic conditions, directly proportional to the elevation. Under unstable conditions it increases more slowly and with stable conditions more rapidly. Within any given interval of altitude it can be represented by an interpolation formula of the type $\eta = \eta_1 \left(\frac{z}{z_1}\right)^q$ where q is somewhat smaller than 1.

Under adiabatic conditions the eddy convectivity is at any given level a linear function of the velocity but at non-adiabatic conditions the relationship between eddy convectivity and velocity is complicated and varies from one level to another. The conclusions are in agreement with the theoretical results in the preceding chapter.

Numerical values of the eddy convectivity. We have now found the laws according to which the eddy convectivity must be expected to vary with elevation directly above the ground and it remains to be examined whether the conclusions are supported by direct determinations of the eddy convectivity at different levels or at different wind velocities at a given level.

The eddy convectivity can be determined by different methods:

1. By a determination of the frictional stress in the lowest layer and a simultaneous determination of the rate of shear in the same layer.

2. By a computation based on the hydrodynamic equations and observations of the velocity distribution up to the altitude of the gradient wind.

3. By a study of the rapid fluctuations of the wind at a fixed level.

The frictional stress was as already mentioned determined by Taylor over Salisbury Plain. He found $\tau = \varrho \gamma^2 u_{30}^2$, where u_{30} is the wind velocity at an altitude of 30 metres and where γ^2 ranges from 22×10^{-4} to 32×10^{-4}. Rossby and Montgomery have shown that this result is in agreement with their conclusions. Their formulae lead in this case to the equation:

$$\eta_a = \varrho \gamma_a^2 \ln \frac{a + z_0}{z_0} (a + z_0) u_a \tag{39}$$

where $a = 30$ metres. Thus, we must know z_0 in in order to compute η, or we must know the rate of shear. In this case we can avail ourselves of the fact that, according to Scrase (1930) the velocity distribution over Salisbury Plain can be represented by the power law, putting $n = 7.7$. The equation:

$$\tau = \eta \frac{du}{dz} = \varrho \gamma_a^2 u_a^2$$

then gives

$$\eta_a = n \varrho \gamma_a^2 a u_a = k' u_a, \tag{40}$$

where k' ranges from 0.065 to 0.093. With $u_a = 10$ m/sec. we obtain $\eta_a = 65$ to 93 g/cm.sec. and these values are in good agreement with values which are determined by other methods.

From (39) and (40) we obtain

$$\frac{a + z_0}{a} \ln \frac{a + z_0}{z_0} = n, \tag{41}$$

which, with $a = 30$ metres and $n = 7.7$, gives $z_0 = 1.4$ cm. Rossby and Montgomery obtain, by a different method, that z_0 ranges between 0.5 cm and 3.2 cm. All these values of the roughness parameter are reasonable and the value $0.065\, u_a < \eta_a < 0.093\, u_a$ appears, therefore, to be approximately correct.

This result has been confirmed by a direct computation of the frictional stress, which was carried out by Scrase (1930) by means of observations of the eddy velocities at 19 metres. The frictional stress in the direction of the wind can, according to Reynolds, be written $\tau = -\varrho \overline{u' w'}$ where u' and w are the eddy velocities in the direction of the wind and in vertical direction. At a wind velocity of 730 cm/sec. at 19 metres Scrase found $\tau = 3.62$, and with $n = 7.7$, he obtained $\eta = 70$ g/cm. sec.

This value is of the order of magnitude which should be expected from Taylor's results, but is about twice too high. A similar computation by means of observations at a height of 1.5 metres gave a value of τ which was about one quarter of the value at 19 metres, or about one half of the value which shoud be expected from Taylor's result. The observations at the two levels were not simultaneous, for which reason the values may deviate, but both values agree with Taylor's results as to order of magnitude.

The second method for determining the eddy convectivity has mostly been used for computation of average values within greater intervals of height on the assumption that η can be considered independent of altitude. The results of such computations are of no interest in this connection, and only the results of a few computations on the assumption of a variable eddy convectivity will be considered. Rossby and Montgomery have already shown that Mildner's (1932) values of η, which were derived from 28 pilot balloon runs made during one October day, are in very good agreement with their theory. At a wind velocity of 10 m/sec. at 30 metres their equation gives $\eta = 63$, or a value which lies close to those over Salisbury Plain. Similar investigations by the author (1933) were based on single ascents and gave values which near the ground were of the same order of magnitude as Mildner's, and increased with height, but they are too uncertain to allow any further conclusions.

The theory of the third method, determination of η by means of the rapid fluctuations of the wind, has been developed by Hesselberg and Ertel. Hesselberg (1929) found:

$$\eta = \varrho \frac{\pi}{8} P (w'_{z,m})^2 = \varrho \frac{P}{8k},$$

where P is the "period length" of the eddy motion, $w'_{z,m}$ the mean vertical eddy velocity and k a constant which characterizes the distribution of the eddy velocities in case this distribution is in agreement with Maxwell's law. By means of this formula Hesselberg computed the eddy convectivity in a horizontal direction at Lindenberg and obtained $\eta_h = 1.0\, u^2$ at the altitude of the anemometer. According to our preceding conclusion η is proportional to u, under adiabatic conditions and increases more rapidly than u if stability prevails. Hesselberg's result may, therefore, be in approximate agreement with our conclusions, if the observations which he has treated were taken under stable conditions. It may be mentioned that Nomitsu (1935) has shown theoretically that η must be proportional to u^2, but on the assumption that η is independent of elevation. This is not true for which reason the theoretical result is without any value.

Ertel (1930) has derived an expression which is more suited for numerical computation:

$$\eta = \varrho_m \frac{S_m(u)\, S_m\left(\frac{\partial u}{\partial t}\right)}{\left(\frac{\partial u}{\partial z}\right)_m^2},$$

where

$$S_m(u) = \pm \sqrt{\frac{\Sigma_i (u_i - u_m)^2}{n}}, \quad S_m\left(\frac{\partial u}{\partial t}\right) = \pm \frac{1}{\triangle t} \sqrt{\frac{\Sigma_i (u_{i+1} - u_i)^2}{n-1}}.$$

Here u_i is the instantaneous velocity and $\triangle t$ the time interval between two observations, which must be of the order of 1 second, and n is the number of observations.

This formula has, to my knowledge, been used only by Lettau (1934), who has treated a set of observations which covered a time interval of 3 minutes. Lettau found $\eta = 2.8$ at an altitude of 95 cm and an average wind velocity of 150 cm/sec. This value is very high, since, if we assume the relation (12) to be correct, we obtain $z_0 = 38$ cm which is an unreasonable value. With $\eta = 0.7$ we would have obtained $z_0 = 2.7$ cm.

From this review it is evident that our knowledge of the eddy convectivity directly above the ground is poor in spite of the interest which has been taken in the study of turbulence near the ground. The results are contradictory, some are in accord with laboratory results and theoretical considerations based on these, but others are in obvious disagreement.

3. Eddy Conductivity near a Boundary Surface.

A. Laboratory Results and Theoretical Considerations.

Smooth Surface. The exchange of heat and water vapour between a surface and the air has been examined in the laboratory and the problem of exchange by eddy conductivity has been treated theoretically by Prandtl, v. Kármán and Latzko.

Let us consider a plate which is placed in a non-turbulent flow. A turbulent boundary layer forms along the plate (see p. 6), and, if the temperature of the air differs from the temperature of the plate, the flux of heat towards a square unit of the plate at the distance x from the rim is

$$Q_x = c_p A \frac{\partial \vartheta}{\partial z} \tag{42}$$

where c_p is the specific heat of the air and A is the eddy conductivity. This flux is constant when stationary conditions are established. Prandtl has shown that, if stability is of no importance, an exact analogy exists between the exchange of momentum and the exchange of heat if the two non-dimensionals, the Reynolds' number and the Peclét's number, are identical. When dealing with air this condition is so nearly fulfilled that, if the effect of stability is disregarded, we can put $A = \eta$. With our previous value of η above a smooth surface (p. 6) we obtain under stationary conditions:

$$\vartheta_z - \vartheta_0 = (\vartheta_L - \vartheta_0)\left(\frac{z}{\delta}\right)^{\frac{1}{n}},$$

where ϑ_z is the temperature at the distance z, ϑ_L the temperature of the air outside the turbulent boundary layer and δ the thickness of the boundary layer. Further, we obtain:

$$Q_x = 0.029\, c_p \varrho R_x^{-\frac{1}{5}}\, U(\vartheta_L - \vartheta_0)$$
$$= 0.029\, c_p \varrho \left(\frac{\nu}{x}\right)^{\frac{1}{5}} U^{\frac{4}{5}} (\vartheta_L - \vartheta_0).$$

Observations by Nusselt, Jürges and more recently by Éliás have confirmed these conclusions. They found that at great velocities of the air flow, the exchange of heat was proportional to $U^{0.8}$. Éliás (1929) has, furthermore, shown that the variation of temperature with increasing distance from the plate is similar to the variation of velocity. He also found the total loss of heat from a surface of unit breadth and of the length x, reckoned from the foremost rim, proportional to $x^{0.89}$, whereas the theory claims proportionality to $x^{0.8}$. During these experiments the plate was placed vertically and, therefore, any influence of stability was eliminated.

At low velocities, when the above considerations are not valid, Jürges found Q proportional to U, but Langmuir found generally that Q was proportional to $\sqrt{U + 0.3}$.

In the case of evaporation or condensation similar expressions are obtained. Under stationary conditions we must have

$$F = A\frac{\partial f}{\partial z} = \text{const.}, \tag{43}$$

where F is the amount of water which is transported away from the surface or towards the surface, and where f is the specific humidity of the air. The equation expresses that at any level the same amount of water vapour passes through a unit cross-section. Assuming A to vary with altitude as η it follows:

$$(f_z - f_0) = (f_L - f_0)\left(\frac{z}{\delta}\right)^{\frac{1}{n}},$$

or, that within the boundary layer the variation of specific humidity is similar to the variation of wind and temperature. Evaporation takes place if $f_L < f_0$ and condensation if $f_L > f_0$. If $f_L < f_0$ and $A = \eta$, the amount of water, which evaporates from a square unit at distance x from the rim of the plate, is:

$$F_x = 0.029\, \varrho R_x^{-\frac{1}{5}}\, U(f_L - f_0)$$
$$= 0.029\, \varrho \left(\frac{\nu}{x}\right)^{\frac{1}{5}} U^{\frac{4}{5}} (f_L - f_0).$$

The evaporation from a rectangular area of length x downwind and breadth dy becomes:

$$dF = 0.029\, \varrho \nu^{\frac{1}{5}} x^{\frac{4}{5}} U^{\frac{4}{5}} (f_L - f_0).$$

The evaporation from a surface is, therefore, proportional to $(f_L - f_0)$ and to $U^{0.8}$, and in agreement with this Himus found it proportional to $u_m^{0.77}$, where u_m is the mean velocity over a cross-section. Sutton (1934) has shown that measurements by Hine give a similar result.

Influence of roughness. The influence of roughness has not been examined in the laboratory, but if we disregard the influence of stability and assume $A = \eta$, we obviously obtain by means of (12) and (42):

$$\theta_z - \theta_0 = K \ln \frac{z + z_0}{z_0}$$

and

$$Q = c_p \varrho k_0^2 \frac{u_a}{\ln \dfrac{a + z_0}{z_0}} \frac{\theta_b - \theta_0}{\ln \dfrac{b + z_0}{z_0}}, \tag{44}$$

where θ means the potential temperature, which must be introduced if we consider exchange in a vertical direction, and where a and b are the levels at which velocity and temperature are measured.

The vertical distribution and the exchange of water vapour are similarly given by:

$$f_z - f_0 = K' \ln \frac{z + z_0}{z_0}$$

and

$$F = \varrho k_0^2 \frac{u_a}{\ln \dfrac{a + z_0}{z_0}} \frac{f_b - f_0}{\ln \dfrac{b + z_0}{z_0}}$$
$$= \frac{0.623}{p} \varrho k_0^2 \frac{u_a}{\ln \dfrac{a + z_0}{z_0}} \frac{e_b - e_0}{\ln \dfrac{b + z_0}{z_0}} \tag{45}$$

In the last term p represents the pressure of the air and e the pressure of the water vapour.

If we make use of the interpolation formula: $u = u_1 z_1^{-\frac{1}{n}} z^{\frac{1}{n}}$ and of Prandl's expression (7) we obtain:

$$\theta_z - \theta_0 = N\left(\frac{z}{z_1}\right)^{\frac{1}{n}},$$

$$Q = C \varrho c_p h^{\frac{2}{n}} (a b)^{-\frac{1}{n}} u_a (\theta_b - \theta_0) \tag{46}$$

and similar expressions for $f_z - f_0$ and F.

In many cases the existence of a laminar boundary layer is presumed when dealing with problems of exchange. It should, however, be stated that if we use the logarithmic formula or consider that the power law has the character of an interpolation formula, the introduction of a laminar boundary layer appears to be unnecessary. From the character of the logarithmic law it is evident that the eddy conductivity has a definite value at the level $z=0$ or that turbulence is effective at the very surface. If a boundary layer exists it must have a semi-laminar character, and this appears to be quite in agreement with the conception by Brunt (1934 p. 262). He considers the boundary layer as a time-mean phenomenon:

"The turbulence which may prevail at some distance from the boundary will from time to time break through the layer, carrying away portions of the fluid which instantaneously constitute it, but as soon as the individual eddy has removed a portion of the boundary layer normal processes will tend to build it up again."

Here we would add that, if a semi-laminar boundary layer exists, θ_0 differs from the temperature of the surface, and e_0 is not identical with the maximum vapour pressure corresponding to the surface temperature.

Influence of stability. We have already shown that the influence of stability upon eddy convectivity is negligible in the immediate vicinity of a boundary surface but increases with increasing distance from the surface. We must assume that the same applies to the eddy conductivity, but now the questions arise, whether eddy convectivity and eddy conductivity can be considered identical at the boundary surface, and, supposing this to be the case, whether they remain identical at heights where both are influenced by stability.

The first question cannot be answered definitely, but in the laboratory A and η were found practically to be equal in a homogeneous fluid and there are no reasons why the same should not be correct in the lowest layer of the atmosphere. The effect of stability has been discussed by Taylor (1931) who obtained:

$$\frac{A}{\eta} \leq \left| \frac{\left(\frac{du}{dz}\right)^2}{\frac{g}{\varrho}\frac{d\varrho}{dz}} \right|. \tag{47}$$

He has verified the correctness of this result by means of oceanographic observations. The mean values of A/η were 0.064 and 0.17 when the mean values of the right hand term were 0.118 and 0.27. The result has not been verified when dealing with the atmosphere, but we may apply Taylor's criterion to the conditions in the immediate vicinity of the ground in order to find the heights up to which we may expect A and η to be identical. If we assume that the variations of velocity and potential temperature are given by logarithmic laws and consider that in the atmosphere:

$$-\frac{g}{\varrho}\frac{d\varrho}{dz} = \frac{g}{T}\frac{d\theta}{dz}.$$

we obtain:

$$\frac{A}{\eta} \leq \frac{T}{g}\frac{u_a^2}{\triangle \theta_a}\frac{1}{(z+z_0)\ln\frac{a+z_0}{z_0}}$$

where u_a is the velocity at the level a cm and $\triangle \theta_a$ is the temperature difference between this level and the level $z=0$. If we now make use of the result that A is equal to η in a homogeneous fluid and introduce the reasonable assumption that A cannot be greater than η in the case of stability, we obtain, neglecting z_0 besides z, that

$$Z = \frac{T}{g}\frac{u_a^2}{\triangle \theta_a}\frac{1}{\ln\frac{a}{z_0}}$$

represents the maximum height up to which A and η can be identical. With $a=200$ cm and $\triangle \theta = 1°$ we obtain the following values of Z in metres:

u_{200} cm/sec.	z_0, cm		
	0.2	1.0	5.0
100	4.0	5.0	7.6
200	16.0	20.8	29.0
400	64	83	116
800	256	333	465

Above these heights A is no doubt smaller than η, and it is possible that A is smaller than η at heights which are somewhat lower than those tabulated. It is, in any case, evident, supposing Taylor's criterion to be correct, that A must be smaller than η at a short distance from the ground, if the wind velocity is very small and a considerable inversion exists, but in a strong wind A and η are probably identical up to considerable altitudes.

The influence of instability has not been considered but it is a priori not probable that A differs from η in the case of instability. If such differences appear to occur they must, perhaps, be explained as an effect of processes of radiation.

Influence of processes of radiation. We have until now, treated the variation of temperature near the ground as if it were dependent upon the eddy conductivity only, and were independent of radiation. A treatment of the influence of radiation upon the vertical distribution of temperature near the ground is very difficult, and satisfactory theoretical results have not yet been obtained. Brunt (1934, p. 120) arrives at the result, that radiation is only of slight importance in the spread of heat upward to any considerable distance above the ground, but thinks that at very small heights the transfer of heat is mainly by radiation. These questions can probably be decided by means of suitable observations which permit comparisons between the apparent values of eddy convectivity and eddy conductivity. The questions will later on (p. 48) be discussed more thoroughly.

B. The Eddy Conductivity of the Air near the Ground.

Variation with elevation. Similarly, as with the investigation of the eddy convectivity, an examination of the eddy conductivity near the ground can be undertaken in two stages:

1. Investigation of the law according to which eddy conductivity varies with elevation.

2. Determination of the eddy conductivity at a fixed level under varying conditions.

An examination of the eddy conductivity should preferably be based on a study of the variation with height of the vapour content of the air since the vapour content is not influenced by processes of radiation. If stationary conditions prevailed, the law according to which the eddy conductivity varies with height, could be obtained from the variation of the vapour content (43, p. 17), but, unfortunately, few observations exist, which are suitable for this purpose. The only series of measurements at different levels, which I have found, is the one short series which was obtained by Wüst (1920). In fig. 4 his average values of the vapour pressure, which is nearly proportional to the specific humidity, have been plotted against *log z*. The values fall nearly on a straight line, indicating that in this case the eddy conductivity was, approximately a linear function of z. The temperature decreased with height, and the stratification was, therefore, instable.

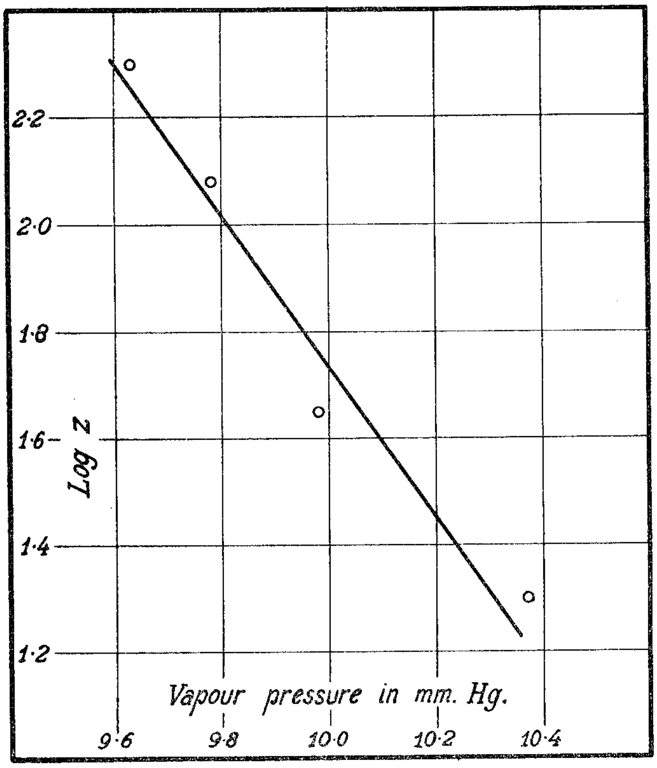


Fig. 4. Vapour pressure as function of log z according to Wüst's observations over the sea.

The temperature distribution has, on the other hand, been studied extensively, and from the observations, conclusions have been drawn as to the variation with height of the eddy conductivity assuming the influence of radiation to be small. Near the ground the temperature is, in general, a linear function of *log z* and A is, therefore, a linear function of z. R. Geiger (1927) has used very detailed observations below 1.5 metres and found an almost linear relation between the relative values of A and z and Marquardt (1932) has shown, that Wüst's observations over the sea also indicate a linear relationship. Best (1935) has shown that around midday and midnight the temperature up to an altitude of a few metres above Salisbury Plain is very nearly a linear function of *log z*, and Johnson's (1929) observations at levels below 2.8 metres give a similar result. Thus, A appears to be a nearly linear function of height, if conditions in the lowest one, or two metres only, are considered.

A study of the diurnal variation of temperature at very low levels may give additional information. Köhler (1932) has shown that Johnson's values of the diurnal variation of temperature at 1.2 m, 7.1 m and 17.1 m above Salisbury Plain can be well represented if the eddy conductivity is proportional to $z^{6/7}$ but Best concludes from his observations at 2.5 cm, 30 cm and 1.2 m together with observations at 7.1 m and 17.1 m that A is proportional to $z^{1.8}$. (Best has computed $K = A/\varrho$.) This result, which is in absolute disagreement with the fact that the temperature, as a rule, is a linear function of *log z*, is, however, derived on the assumption that A can be considered constant within the different intervals of altitude, 2.5 to 30 cm, and so on, and this procedure must lead to erroneous values. It can be shown that Best's data do not contradict that A is, approximately, a linear function of z.

Fjeldstad (1933) has shown that:

$$\frac{d}{d\,z}\left(\frac{A}{\varrho}R_n^2\,\frac{d\,\delta_n}{d\,z}\right)+n\,\sigma\,R_n^2=0\,,$$

where R_n represents the amplitude of the n^{th} term in a Fourier's series, δ_n the phase angle and, where $\sigma = 2\pi/T$ and T is the period length. By integration we obtain:

$$A_z = A_h\,\frac{\left|R_n^2\,\frac{d\,\delta_n}{d\,z}\right|_h}{\left|R_n^2\,\frac{d\,\delta_n}{d\,z}\right|_z} - n\,\sigma\,\varrho\int_h^z R_n^2\,d\,z\,.$$

Best has tabulated the values of R_1 and R_2, $-\delta_1$ and $-\delta_2$ for the months December, March and June. If we use the first term only, since this dominates, and put $h = 10$ cm and $\varrho = 1.25\ 10^{-3}$ we obtain, approximately:

December ..... $A = 0.09\,A_{10}\,z - 7\times10^{-6}\,(z-10)^{1.74}$
March .... $A = 0.11\,A_{10}\,z - 8\times10^{-6}\,(z-10)^{1.66}$
June .... $A = 0.13\,A_{10}\,z - 4\times10^{-6}\,(z-10)^{1.85}$

These equations are valid up to a height of about 5 metres. Evidently the character of the variation of A depends upon the value of A_{10}. If we assume this to be equal to η_{10} we have (see p. 7), since the wind observations gave $z_0 \approx 0.2$ cm:

$$A_{10} = \varrho\,k_0^2\,10.2\,\frac{u_{200}}{\ln 1000}$$

and with $u_{200} = 400$ cm/sec:

$$A_{10} \sim 0.1\,.$$

If this result is correct, it is evident that the first term quite dominates, and our equations for A are similar to (31) p. 10. If we examine the order of magnitude of the factor in the second term on the right-hand side of (31), we find, with $\beta = 11$ (see p. 45) that this factor lies between 10^{-6} and 10^{-5}, as in the above equations. The deviation from a linear relation between A and η may, therefore, indicate the influence of the stability in the night hours. It may be added that Köhler's investigation (1932) gave:

$$A_{10} = 0.13\,.$$

The available evidence, therefore, all points in the direction that in the lowest metres the eddy conductivity is nearly a linear function of height. The increase is, under stable conditions, probably somewhat slower, as was the case with the increase of the eddy convectivity. It is, furthermore, noteworthy that this result principally is based upon temperature observations. The similarity between the laws for the variation with height of eddy convectivity and eddy conductivity, therefore, indicates that processes of radiation are of subordinate importance and that turbulence is mainly responsible for the temperature distribution in the lowest layer.

Numerical values of the eddy conductivity. The eddy conductivity can be determined by several different methods:

1. By a study of the diurnal variation of temperature and specific humidity at different levels.
2. By a study of the rapid fluctuations of temperature and specific humidity at fixed levels.
3. By an examination of the change of temperature and vapour contents of air under non-stationary conditions.
4. By an examination of the exchange of heat and/or water vapour between the surface and the air under stationary conditions.

The first method was, as already mentioned, used by Köhler (1932) who found a value of A, which is of the same order of magnitude as the probable value of η. The second method has been used by Lettau (1934) who by means of Ertel's formula (see p. 16) and observations over a period of 4 minutes found $A=20.0$ at an altitude of 95 cm. A few minutes later, wind observations which were treated similarly gave $\eta=2.8$. The great discrepancy between these values cannot be explained by the small time difference between the measurements, but, considering the technical difficulties of the observations in question, it is not permissible, from this isolated result, to conclude that A can be greater than η.

The third method has been used by Taylor (1915) for his first determination of an average value of the eddy conductivity up to an altitude of several hundred metres, but average values are of small interest in our case. Köhler (1929) has, on the assumption that $A=A_1 z^{6/7}$, treated several series of detailed observations at Hallde which gave $A=0.0002 + 0.00071\, u_1^2$, where u_1 is the velocity at 1 cm. Marquardt (1932) has examined conditions over the Bodensee, but his numerical values have been computed on the basis of assumptions which cannot be accepted.

The exchange of heat and/or water vapour between the surface and the air under quasi-stationary conditions has been frequently examined but the efforts have been directed towards derivation of technical formulae. In the technique the heat and vapour exchange are expressed by equations of the type

$$Q=\varkappa\,(\vartheta-\vartheta_0)\,,\quad F=\gamma\,(e-e_0)\,, \tag{48}$$

where $\varkappa$ and γ are "coefficients of exchange", which are more or less complicated functions of several variables. When dealing with meteorological problems the exchange of heat and water vapour can be represented by similar equations and the coefficients of the exchange are functions of air temperature, pressure and wind velocity.

From our preceding equations (44) and (45) we obtain:

$$\varkappa=\frac{c_p\,\varrho\,k_0^2}{\ln\dfrac{a}{z_0}\,\ln\dfrac{b}{z_0}}\,u_a\,,\quad \gamma=\frac{0.623}{p}\,\frac{\varrho\,k_0^2}{\ln\dfrac{a}{z_0}\,\ln\dfrac{b}{z_0}}\,u_a \tag{49}$$

if velocity, temperature and vapour pressure vary according to logarithmic laws and if the velocity is measured at the level a and temperature and humidity at the level b, supposing a and b to be great compared to z_0.

If the vertical variations of the elements are represented by power laws we obtain, similarly:

$$\begin{aligned}\varkappa&=C\,c_p\,\varrho\,h^{\frac{2}{n}}\,(a\,b)^{-\frac{1}{n}}\,u_a\,,\\ \gamma&=\frac{0.623}{p}\,C\,\varrho\,h^{\frac{2}{n}}\,(a\,b)^{-\frac{1}{n}}\,u_a\,.\end{aligned} \tag{50}$$

Thus, the exchange coefficients should be proportional to the wind velocity and depend upon the levels of observation. If they were known with sufficient accuracy, it should be possible to check our preceding results. It should, however, be noted that in (48) ϑ_0 and e_0 represent temperature and vapour pressure of the air very near the surface, and may differ from the temperature of the surface and the corresponding vapour pressure owing to the existence of the semi-lamiar boundary layer.

The exchange of water vapour has been studied extensively in order to derive empirical formulae for the evaporation, but in some cases the evaporation has been found proportional to $\sqrt{u}$ and in other cases to u. The proposed formulae differ so much from each other that they give no foundation for a computation of the conductivity.

The exchange of heat has more recently been made the subject of investigation, and the coefficient of heat exchange has been computed, but the possible relation between this coefficient and the wind velocity has been studied in one case only.

Over a snow surface Ångström (1918) found $\varkappa=0.03$ at a wind velocity of 2.8 m/sec. The velocity was measured at 15 m and the air temperature at 0.6 m. The surface was colder than the air. Under similar conditions Falckenberg and Krügler (1932) found a much smaller value, $\varkappa=0.003$. In the desert, with the surface warmer than the air, Büttner (1934) found $\varkappa=0.016$ at wind velocities between 0.8 and 1.5 m/sec. and $\varkappa=0.021$ at wind velocities 2 to 2.5 m/sec. The velocity was measured at 0.4 m and the temperature at 1 m.

Devik (1932) assumed that $\varkappa$, according to Langmuir, was proportional to $\sqrt{u+0.3}$ and at velocities 4 to 6 m/sec., he found $\varkappa=0.0083\sqrt{u+0.3}$ over a water surface. The altitudes at which velocity and air temperature were measured are not stated. Over a snow surface he found $\varkappa=0.033\sqrt{u+0.3}$, but this value is considered uncertain.

None of these results can be used for computing the conductivity, but on a later occasion Ångström (1934) has undertaken a more detailed study of $\varkappa$ in connection with his discussion of the observations on the Swedish-Norwegian expedition to North-Eastland

(Spitsbergen) in 1931. He found, in agreement with the theoretical consideration, that $\varkappa$ was proportional to u: $\varkappa = 0.0067\ u$ (u in m/sec.). The wind velocity was measured at 3.2 m and the temperature at 1.9 m. The numerical value of the constant factor cannot, however, be accepted, since in his computation Ångström has neglected the processes of condensation and evaporation. A new determination of $\varkappa$ by means of the same observations has, therefore, been undertaken (Sverdrup, 1935 b).

Finally, it may be mentioned that over the pack-ice the author (1933) found:

$Q = (44 - 2.9\ C)\ 10^{-6}\ u$ g cal./cm² min. (u in cm/sec.)

and simultaneously the relation:

$$\theta_{30} - \theta_{4.5} = \frac{176 - 12.5\ C}{u}$$

existed. Here C represents the cloudiness on the scale 0 to 10. The velocity u was measured at 7 metres and θ_{30} and $\theta_{4.5}$ represent the potential temperatures at 30 and 4.5 metres. This result, which is obviously in disagreement with our theoretical considerations will be discussed later on.

This review of our present knowledge of the eddy convectivity and eddy conductivity near the ground shows that observations, which permit a further analysis of the processes of exchange, are very desirable. The observations in 1934 on Isachsen's Plateau, Spitsbergen, were planned with such analysis in view.

4. Meteorological Observations over the Snow Field on Isachsen's Plateau.

A. Wind.

Instruments and observations. The expedition had four anemometers, one electrically-recording contact anemometer, (Fuess No. 23142), two small cup-anemometers, (Fuess Nos. 4312 and 8685) and one hot-wire anemometer after Albrecht (Fuess No. 24377).

The recording contact anemometer was placed permanently at the top of a mast, which had been transported up to the plateau. The mast was made of two iron tubes screwed together; it stood on a foot of bamboo pieces and was held in position by means of six wire stays. The stays were fastened to long wooden pegs which were driven down into the snow, but these had often to be fastened anew since the snow melted rapidly. The anemometer was fastened to the top of the mast with the cups seven metres above the snow surface. A wind vane by means of which the direction of the wind was read off once an hour, was placed below the anemometer. An electrical contact was made for every 500 metres of wind way. These contacts were recorded by means of a chronograph in our living and working tent. The drum of the chronograph made one revolution in one hour.

One of the two small cup anemometers was placed on top of a pole with the cups at 2 metres above the snow surface, and the other on a second pole with the cups at 30 cm. Anemometer No. 8685 was used at 2 m from June 25, 17^h to July 22, 16^h and in this period anemometer No. 4312 was used at 30 cm. They were changed over on July 22 at 16^h and remained in their new positions until at 8^h on August 16, when the work was discontinued. The exchange was made in order to eliminate, if possible, a systematic error of the total mean values, due to errors in the constants of the two instruments. Both anemometers were read once an hour.

The hot-wire anemometer was only suitable for measuring very small velocities and was, therefore, used as close to the snow surface as possible. The instrument has four thin platinum wires mounted vertically on a ground plate, the latter being fastened to a handle in which the various resistances are placed and from which a cable leads to battery and galvanometer. When the anemometer was in use the handle was stuck so deep into the snow that the ground plate, that means the lower ends of the wires, was level with the surface. The middle of the wires was then about 3.3 cm above the surface, but owing to the rapid change of velocity near the surface it has been assumed that with this instrument we measured the wind velocity at an altitude of 4 cm above the snow. This height is, however, reckoned from the top of the small irregularities of the snow surface. These irregularities are a few cm high, and reckoned from the bottom of the irregularities, the velocity was probably measured at a height of about 6 cm. The hot-wire anemometer could be used in dry weather only. When it rained, or in the presence of fog, drops of water collected on the wires and made the readings worthless.

All anemometers except No. 4312 had been calibrated by the makers shortly before our departure and all were again calibrated after our return, and, in addition, a great number of inter-comparisons were undertaken in the field. Comparisons between the recording contact anemometer and one of the small hand-anemometers could be undertaken on two occasions only, namely on July 4, when the mast had to be taken down for minor repair, and on August 16, when the observations were discontinued. These comparisons were undertaken at wind velocities between 1.5 m/sec. and 6 m/sec. According to the calibration by the makers, the mean hourly wind velocity is obtained from the record by means of the formula:

Recording contact anemometer No. 23142:
$$v = 0.9 + 0.0258\,n,$$

where n is the number of contacts in one hour, but our comparisons in the field indicate that this formula was not valid at velocities smaller than 2 m/sec. When the number of contacts per hour was smaller than 25, the corresponding mean hourly velocity has, therefore, been derived by means of the following table:

n	v (m/sec.)	n	v (m/sec.)
25—23	2.1	13—12	1.6
22—21	2.0	11— 9	1.5
20—19	1.9	8— 6	1.4
18—17	1.8	5— 3	1.3
16—14	1.7	2— 1	1.2

This procedure is, however, somewhat doubtful, and the lowest velocities as recorded by the contact anemometer, are, therefore, uncertain.

The calibrations of the two hand anemometers gave the following results:

Hand anemometer No. 8685: $v = 0.6 + 0.0151\,n$
Hand anemometer No. 4312: $v = 0.5 + 9.0168\,n$

where n is the number of revolutions per minute.

On the plateau inter-comparisons between these two anemometers were undertaken on 19 days, as a rule, in periods lasting 5 minutes. When the velocities are computed by means of the above formulae they agree within 0.1 m/sec. in 7 cases, in 6, the velocity of No. 8685 is greater, and in another 6 it is less, than the velocity of No. 4312. The average difference, regardless of sign, is 0.1 m/sec., but taking the sign into account the difference is —0.03 m/sec. In the single cases the differences lie between 0.2 and —0.3 m/sec. and show no systematic variation with the velocity. The latter feature indicates that the differences are due to accidental variations of the constant term. Such variations may be caused by lack of lubrication, condensation of moisture, or similar circumstances which are difficult to control. The differences never remained systematic for longer periods and it has, therefore, been considered advisable to disregard them and always to compute the velocities by means of the above formulae. The hourly values which are obtained in this manner may be as much as 0.5 m/sec. wrong, and even the mean daily values may be 0.3 m/sec. in error. It is probable that these errors generally tend to make the wind velocities too small since the factors which exert an influence upon the constant term cause this to increase. Errors of this magnitude may have a considerable influence upon the differences between the observed wind velocities at 2 m and 0.3 m, and the ratio between these velocities, the influence being greatest at small velocities.

At small velocities our values are probably too great. The formulae are valid only if the cups are in continuous motion, but if the wind is very weak the cups may be at rest for long periods. Let us, as an extreme case, assume that the cups of anemometer No. 8685 made 300 revolutions in 10 minutes and that in the remaining 50 minutes the wind velocity was zero. Then the average velocity in one hour was 0.18 m/sec. (1.05 m/sec. in 10 seconds and 0.00 in 50 seconds). If we, however, assume that the cups turned continuously, we obtain an average velocity of 0.67 m/sec. The errors, which arise in this manner, are smaller than in this example but they tend, no doubt, to increase the small velocity.

More serious errors arose on some occasions because of the formation of frost on the instruments. The two hand anemometers at altitude 2 m and 0.3 m could, as a rule, be kept free from frost since they were read off and looked after once an hour and were easily accessible but the recording anemometer at an altitude of 7 m could not be cleaned and in some periods with temperature below freezing point and with fog it has recorded too low values. The cases in which the recorded velocities are doubtful owing to formation of frost are indicated in the tables.

In the case of the hot-wire anemometer the readings were made on a galvanometer and afterwards converted to wind velocity in m/sec. by means of a calibration curve. Calibrations which were undertaken by the makers before the departure and after the return gave both very similar curves, but the comparisons on the plateau gave deviating results. On the plateau inter-comparisons between the hot-wire

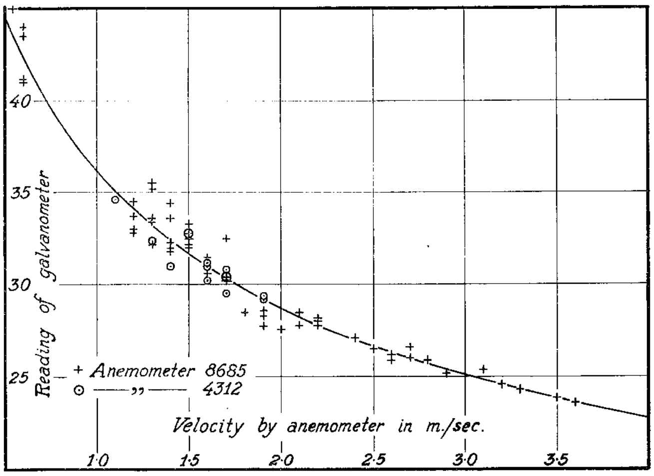


Fig. 5. Comparison between the hot-wire anemometer and cup anemometers, and resulting calibration curve for the hot wire anemometer.

anemometer and the hand anemometer were undertaken in the following manner. The hot-wire anemometer and one of the hand anemometers were exposed at the same altitude; the former being read at intervals of 10 seconds, the latter every minute. The mean values for one minute of the readings of the hot-wire instrument were compared with the corresponding one-minute value of the hand anemometer and in this manner 64 comparisons were obtained, 51 between the hot-wire anemometer and the hand anemometer No. 8685 and 13 between the hot-wire anemometer and the hand anemometer No. 4312. The results are shown in fig. 5 in which a curve has been drawn. This curve has the same shape as the calibration curve supplied by the makers but lies lower. It has not been possible to find a satisfactory explanation of the discrepancy between the results, but when converting our readings we have made use of the curve which was determined on the plateau by comparison with instruments which in the same period were used at other levels. We have, furthermore, extrapolated this curve to velocities lower than the observed ones, taking the shape of the other calibration curve into account. In this manner comparable observations are obtained from all levels.

Tabulation of the observations.

From the records at 7 metres, mean half-hourly values of the velocity were computed and by means of these a curve showing the velocity at 7 metres as a function of time was constructed for each day. Furthermore, hourly values, centred on the full hours, were computed.

From the readings of the anemometers at 2 and 0.3 metres the mean velocities in the intervals between two readings were computed. These intervals were not always exactly one hour because the different observations could not always be taken at the same minute. The mean velocities were entered on the graphs below the curves at 7 metres, and curves showing velocities at 2 and 0.3 metres were constructed, taking into account the shape of the more accurate curve for the upper level, except on days when the record at 7 metres evidently was erroneous owing to formation of ice. From these curves, mean hourly velocities, centred on the full hours, were read off.

Observations with the hot-wire anemometer were, when the weather permitted, taken for 10 to 15 minutes every hour. In these minutes five readings of the galvanometer were made. The galvanometer needle was, as a rule, moving to and fro. It was, therefore, watched for a fraction of a minute and the average position was recorded. From these average readings the mean value was computed and converted to m/sec. by means of the calibration curve. The mean values were entered on the same graphs as the other observations and, finally, a curve was drawn showing the velocity at an altitude of 4 cm as a function of time, taking the shape of the curves at the other levels into account. From the curve, mean hourly values were finally read off.

All hourly values are given in table I. The first line of the table gives the direction of the wind which was read off once every hour by means of the wind vane at the top of the mast.

B. Temperature and Humidity.

Instruments and Observations. Temperature and humidity were determined simultaneously by means of aspirated psychrometers. We had two psychrometers, one ordinary Assmann psychrometer, and one of similar pattern but supplied with reversing thermometers as suggested by Böhnecke (1933). The latter instrument had been acquired for measurements of temperature and humidity at different levels along the mast. By means of a simple arrangement it was hoisted up to the level at which the measurements were to be undertaken. As a rule, the ventilator was set in function immediately before hoisting the instrument up, although it was provided with a device which permitted us to release the ventilator when the instrument had reached its proper position, but we soon found that no error was introduced if the ventilator was running during the hoisting. When the instrument had been brought to its proper level, it was left undisturbed for some minutes in order to give the thermometers time to adjust themselves to the existing conditions. Four or five minutes were, as a rule, sufficient, but occasionally the observations had to be repeated several times before trustworthy readings were obtained. This happened especially when the air temperature was slighty above freezing point and the air was so dry that the wet bulb would show a negative temperature. In these circumstances the adjustment of the wet bulb took a very long time and occasionally the observations had to be repeated three or four times before satisfactory results were obtained. After a sufficient exposure the instrument was reversed by pulling a string, it was lowered in reversed position and in this position the temperatures of the two reversing thermometers were read off.

The instrument functioned, as a rule, in a very satisfactory manner, but some of the ordinary trouble met with when dealing with reversing thermometers, occurred. It happened for example that the mercury column would break off in a wrong place and then the thermometer had to be knocked or heated before it again functioned normally.

By means of the reversing psychrometer, temperature and humidity were observed once an hour at an altitude of 5 metres. The instrument was, as a rule, also used for measurement of the temperature at 1 meter as well, in order to obtain a check of the readings by this instrument and the readings of an ordinary Assmann psychrometer. The latter was used for measurement of the temperature and humidity at 1 meter and as close to the snow as possible. In order to undertake the last-mentioned observation the psychrometer was suspended vertically with the openings for the air currents only a few millimetres above the snow (3 to 5 mm). It must, however, be considered that the air, which is drawn in through the openings by the ventilator and passes the bulbs of the thermometers, is not drawn in horizontally, but is sucked down from higher levels. Nothing would be gained by placing the instrument in a horizontal position, because the air in that case would also be drawn from some higher level and the actual altitude to which the measured air temperature should be referred would have to be determined in that case as well.

In order to find this altitude, and also to examine the temperature distribution at the lowest metre in greater detail, several series of measurements were undertaken by means of a thermo-couple. When such measurements are to be undertaken the thermo-junction, which is to assume the temperature of the air, must be protected against radiation. After several attempts, Mr. Knudsen succeeded in constructing a protection which was sufficient when the radiation income was small and when the wind velocity was greater than about 1 m/sec. This instrument could be used at a distance of a few centimetres from the surface if the radiation income was small and the wind velocity at 2 metres was higher than about 2.5 m/sec., but at lower velocities the values near the surface were influenced by radiation, since there the ventilation of the instrument became unsatisfactory. Thus, the measurements could be undertaken only between 18^h and 6^h if the wind was not too weak, and, in order to obtain trustworthy results, when the temperature varied rapidly with height. These conditions were not often fulfilled simultaneously and, therefore, such series were obtained on four occasions only. The results of these series are:

7 series between July 22nd, 21^h and July 23rd, 3^h:

Average wind velocity at 2 m: 3.67 m/sec.

Heights in cm	100	50	20	5	x
Temp. by thermo-couple, °C	−1.84	−1.74	−1.67	−1.49	
Temp. by Assmann, °C	−1.87				−1.46

13 series between July 23rd, 18^h and July 24th, 6^h:

Average wind velocity at 2 m: 4.12 m/sec.

Heights in cm	100	50	20	5	x
Temp. by thermo-couple, °C	−1.44	−1.38	−1.33	−1.20	
Temp. by Assmann, °C	−1.45				−1.20

9 series between August 4th, 19^h and August 5th, 6^h:

Average wind velocity at 2 m: 2.22 m/sec.

Heights in cm	100	50	18	5	x
Temp. by thermo-couple, °C	3.25	2.84	2.46	1.94	
Temp. by Assmann, °C	3.36				1.88

5 series between August 13th, 18^h and 22^h:

Average wind velocity at 2 m: 2.58 m/sec.

Heights in cm	100	50	20	5	x
Temp. by thermo-couple, °C	1.54	1.42	1.34	1.12	
Temp. by Assmann, °C	1.56				0.92

The values are represented in fig. 6 where the observed temperatures are plotted against $\log z$. It is seen that the temperature distribution is nearly always a linear function of $\log z$ and by entering on the diagram the values observed by the psychrometer

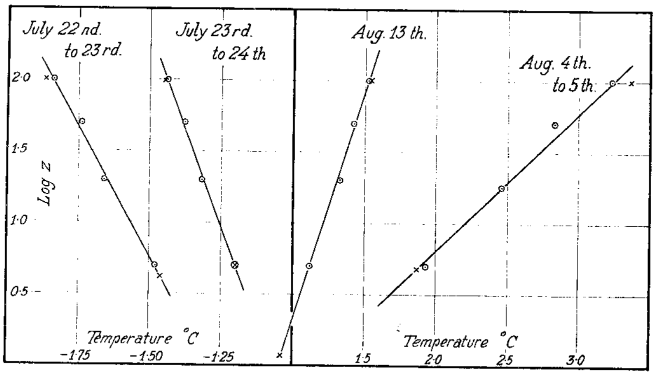


Fig. 6. Vertical variation of temperature near the ground according to observations with thermo-couples.

directly above the surface, one finds the following values of the height in cm: 4.2, 5.0, 1.2 and 4.8. As an average value we have adopted 4 cm. This result has been confirmed by a number of simultaneous measurements made between August 5th and 13th, using the psychrometer directly above the surface and the thermo-couple at 4 cm. In the hours between 18 and 6 the difference between the readings of the two instruments is a function of the wind velocity and approaches zero when the wind velocity at 2 metres surpasses 3 m/sec.

At low wind velocities the thermo-couple gives too high temperatures, evidently because the instrument is influenced by radiation if the ventilation is small. The fact that at 1 meter the corresponding differences are negligible, indicates that correct temperatures are obtained by the thermo-couple if the ventilation is sufficient.

The thermometers of the two psychrometers were frequently controlled. The zero points of the thermometers of the ordinary Assmann were determined on several occasions in melting snow. In the case of the reversing thermometers the correctness of the reading of the wet bulb could be ascertained by reversing the thermometer at a time when the water on the muslin was freezing, and the two thermometers could be compared on wet days when the humidity was 100 per cent because then both should give the same reading. The agreement between the two instruments was, as a rule, very good, but on some occasions rapid fluctuations of the temperature occurred, as evident from observations with the thermo-couple, and then the time difference between the readings of the two instruments explains existing discrepancies. In such cases the observations were often repeated in order to obtain several values distributed over a somewhat longer period. Uncertainties of the final values are, therefore, principally due to lack of simultaneity and lack of knowledge when the fluctuations of the temperature are involved.

The series of temperatures at the three altitudes are without any serious gaps, although one of the instruments was often out of function. We had not obtained any spare springs for the clockwork driving the ventilator and, owing to the constant use of the instrument, the wear on the spring was great. The springs accordingly broke several times but Mr. Knudsen succeeded in repairing them. After 20^h on August 4th all observations were, however, taken with the reversing psychrometer as on that day the clockwork of the ordinary Assmann was damaged beyond repair.

In the case of humidity, however, some large gaps occur, the reason being that at temperatures around freezing-point measurement of the humidity by means of a psychrometer is often difficult, as the adjustment to the correct temperature reading takes a long time if the water, which is applied to the muslin cover of the wet bulb, freezes. It happened that we used as much as one half hour in order to obtain a result which could be considered as correct. At first we did not pay enough attention to this circumstance and most gaps occurred then, but it has been possible to fill in a number of gaps by interpolation, taking into account trustworthy values at neighbouring levels.

During a short period when the Assmann psychrometer was under repair, the temperature at an altitude of 4 cm was measured by means of the thermo-couple. The relative humidity at 4 cm is, therefore, lacking for this period, but during the greater part of the time we had dense fog and relative humidity about 100 per cent at 1 meter, for which reason it has been assumed that the humidity was 100 per cent at 4 cm.

Tabulation of the results. From the observation of temperature and humidity with intervals of one hour, mean hourly values have been derived. Curves, representing the temperature and humidity as functions of time at each level, have been drawn. The courses of these curves were well determined when the elements varied slowly, but on days with rapid fluctuations the curves were uncertain. The latter drawback was to some extent reduced by the circumstance that in such cases several readings had often been taken at the same level, distributed over a period of about 15 minutes. From the curves the hourly mean values, centred on the full hour, were read off. These values are given in tables II and III. Values which have been obtained by interpolation are indicated especially only when it was necessary to interpolate several values in succession.

C. Clouds and Cloudiness, Fog and Precipitation.

The cloud forms, cloudiness and hydrometeors were recorded once an hour. The cloudiness and the hydrometeors are given in table IV. Our observations of the cloud forms are not published since they are of no special interest to the following discussion.

The amount of precipitation was measured twice a day, at 7^h and at 19^h, an ordinary rain gauge being used. Only rain fell in measurable quantities, the amounts of snow, which fell on a few occasions, proving too small to be measured. The measured amounts of precipitation are given in table V.

5. The Vertical Variation of Wind Velocity, Temperature and Vapour Pressure. Empirical Results.

A. Vertical Variation of Wind Velocity.

From the values in table II it is evident that, in most cases, we met stable conditions since the temperature, on an average, increased with elevation. An inspection of the table shows, however, that the stratification was instable on several days and that indifferent equilibrium occurred on some occasions. In view of the theoretical results, the rational procedure is to examine the variation of velocity with elevation at instable, indifferent, and stable stratification.

Let us first consider the conditions when the stratification is instable or indifferent. Omitting the cases in which the recorded velocity at 7 metres undoubtedly was in error owing to deposit of frost on the instrument, we find the values in table 1.

Cases with very small and with moderate wind velocities have been treated separately, especially since observations from 4 cm are so frequent at low velocities that averages for four levels could be computed. The relative values appear to reveal some difference between the conditions with weak and moderate wind, but it must be taken into account that the weak velocities are uncertain when measured with a cup-anemometer as was the case at all levels except at 4 cm. The relative error increases with decreasing

Table 1. *Average values of wind velocity and potential temperature at instable or indifferent equilibrium.*

Average velocity at 2 m and 30 cm		< 2 m/sec.		> 3 m/sec.	
Temperature difference, °C		<−0.2	−0.2 to 0.2	<−0.2	−0.2 to 0.2
Wind velocity in m/sec. at levels	700 cm	1.63	1.82	4.89	5.24
	200 cm	1.59	1.64	4.36	4.72
	30 cm	1.02	1.06	3.22	3.31
	4 cm	0.63	0.71		
Velocities relative to velocity at 200 cm	700 cm	1.03	1.11	1 12	1.11
	200 cm	1.00	1.00	1.00	1.00
	30 cm	0.64	0.65	0.73	0.70
	4 cm	0.40	0.43		
Potential temperatures °C at levels	500 cm	−2.54	−0.49	−1.85	−0.68
	100 cm	−2.47	−0.56	−1.81	−0.70
	4 cm	−1.75	−0.53	−1.22	−0.72
Number of hours		40	57	30	70

velocity and this feature explains perhaps that the ratio u_{30}/u_{200} is smaller with weak than with moderate velocities. The fact that the ratio u_{700}/u_{200} is very small under instable conditions and with weak wind may be due to the circumstance that we have not succeeded in eliminating all cases in which the anemometer at 7 metres was coated by frost.

In fig. 7 the relative values of the velocities are plotted against log z. It is seen that, in spite of the scattering, they group themselves fairly well around a straight line. Thus, the variation of velocity with elevation can be approximately represented by a logarithmic formula. If we introduce a formula of the type (11) we obtain:

$$z_0 = 0.23 \text{ cm}.$$

This value of the roughness parameter is nearly the same as the value which was derived from Best's observations over a smooth cricket field. The snow surface was actually very smooth and it is, therefore, probable that our value is approximately correct. Thus, our observations are in good agreement with the theoretical conclusions, if the conditions are instable or indifferent.

In order to use the most trustworthy observations for an examination of the influence of stability we have selected the data from hours with southerly wind and with positive temperatures both at 4 cm and 500 cm. Under these conditions errors due to formation of frost are avoided. We find 285 hours during which these conditions are fulfilled. We have

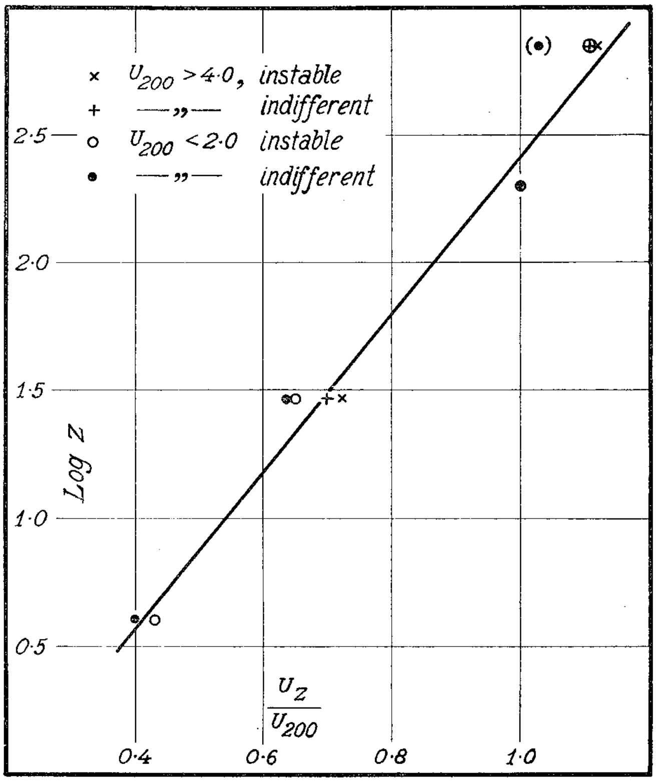


Fig. 7. Relative velocity as function of log z under instable or adiabatic conditions.

again treated separately the cases with weak and moderate velocity and have, furthermore, undertaken a grouping according to the temperature at 5 metres. The results are presented in table 2 which also contains the average potential temperatures and vapour pressures.

An examination of these data shows that under stable conditions the variation of velocity between 30 cm and 700 cm is, as a rule, represented better by a power law of the type (4) than by a logarithmic law. In figs. 8 and 9 the logarithms of the relative velocities are plotted against log z. At weak velocities, when the ratios are somewhat uncertain, the values fall on a straight line only when the stability is very great but none of the other values fall much outside of this line which gives $n = 4.3$. At moderate velocities the power law gives within each group a good approximation, but n decreases with increasing stability. Three lines are, therefore, entered in fig. 9 and these give $n = 6.3$, $n = 5.9$ and $n = 5.5$ respectively. We may also compute n directly from the values in table 2, using the method of least squares, and obtain

Group No.:	I	II	III	IV	V	VI
Computed values of n:	4.3	4.2	4.2	6.2	5.9	5.4

Table 2. *Average value of wind velocities, potential temperatures and vapour pressures on days with southerly wind and positive temperatures.*

Average velocity at 2 m and 30 cm	< 2 m/sec.			> 3 m/sec.		
Temperature at 5 m, °C	0° to 0.9°	1.0° to 2.4	> 2.4°	0° to 0.9°	0.9° to 1.9°	>1.9°
Wind velocity						
700 cm	1.85	2.26	1.97	5.55	5.49	5.80
in m/sec. 200 cm	1.51	1.62	1.46	4.77	4.54	4.65
at levels. 30 cm	0.90	1.07	0.92	3.39	3.22	3.24
Velocities						
relative 700 cm	1.22	1.39	1.35	1.16	1.21	1.25
to velocity 200 cm	1.00	1.00	1.00	1.00	1.00	1.00
at 200 cm 30 cm	0.60	0,66	0.63	0.71	0.71	0.70
Potential						
tempera- 500 cm	0.70	1.88	4.87	0.62	1.33	3.15
tures in 100 cm	0.62	1.40	3.70	0.54	1.07	2.53
°C 4 cm	0.42	0.84	1.92	0.34	0.76	1.61
at levels.						
Vapour						
pressures 500 cm	4.77	5.18	6.15	4.64	4.89	5.58
in mm Hg 100 cm	4.76	5.06	5.76	4.65	4.85	5.40
at levels. 4 cm	4 65	4.87	5.15	4.61	4.80	5.18
Number of hours	16	33	25	39	60	35
Group No.	I	II	III	IV	V	VI

Thus, we obtain that at weak velocities n is about 4, whereas at moderate velocities n decreases with increasing stability, but remains considerably higher than 4 under the conditions which are present.

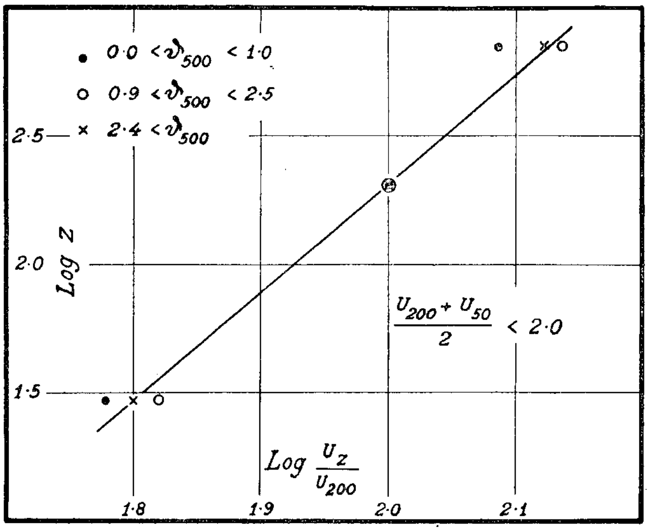


Fig. 8. Logarithm of relative velocity as function of log z under stable conditions. Weak velocity.

Table 3. *Average values of wind velocities, potential temperatures and vapour pressures on days with southerly wind and positive temperature.*

Average velocity at 2 m and 30 cm in m/sec.		< 2.0	2.1 to 2.95	3.0 to 3.95	> 4.0
Wind velocity in m/sec. at levels	700 cm	2.07	3.74	4.97	6.63
	200 cm	1.55	2.96	4.06	5.60
	30 cm	0.98	2.03	2.84	4.00
Velocities relative to velocity at 200 cm	700 cm	1.33	1.26	1.22	1.18
	200 cm	1.00	1.00	1.00	1.00
	30 cm	0.63	0.68	0.70	0.72
Potential temperatures in °C at levels	500 cm	2.68	2.22	1.67	1.48
	100 cm	2.02	1.65	1.33	1.25
	4 cm	1.12	1.00	0.87	0.84
Vapour pressures in mm Hg. at levels	500 cm	5.42	5.19	4.99	4.96
	100 cm	5.23	5.06	4.92	4.89
	4 cm	4.92	4.89	4.82	4.81
Number of hours		74	77	84	50
Values of n		4.2	5.2	5.6	6.2
Group No....................		VII	VIII	IX	X

It will be shown that this result is also in perfect agreement with the theoretical conclusions.

When the data which we use for an examination of the stability are grouped according to the velocity at 2 metres, we obtain the mean values in table 3 which also contains the average potential temperatures and vapour pressures.

The value of n increases regularly with increasing velocity and we obtain a similar result if

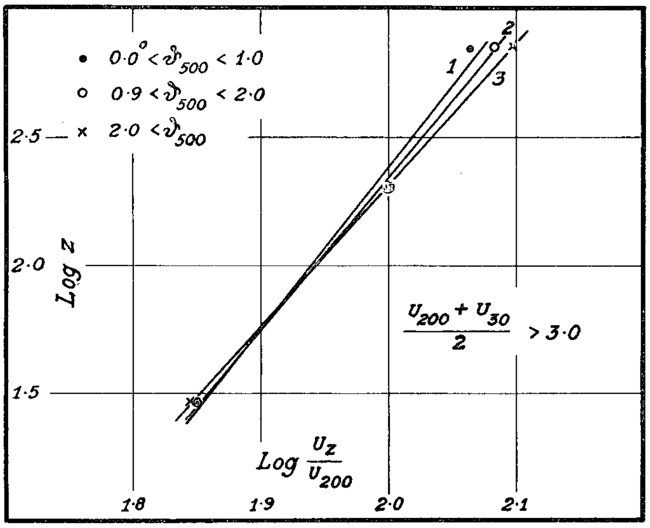


Fig. 9. Logarithm of relative velocity as function of log z under stable conditions. Moderate velocity.

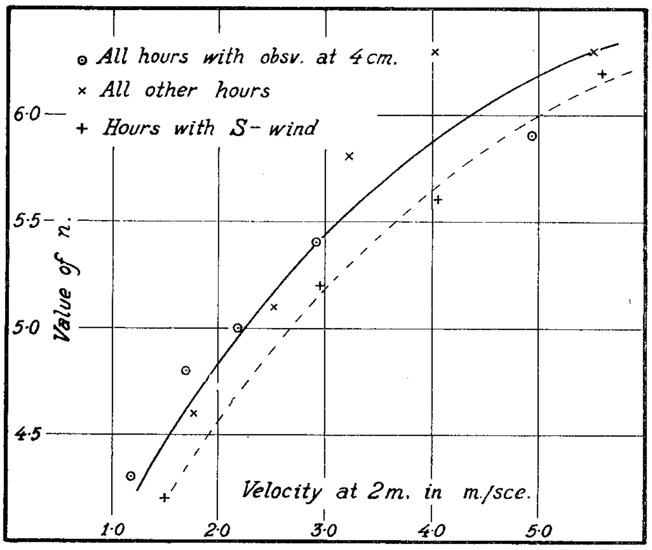


Fig. 10. Relation between the velocity at 2 m and the value of n.

this investigation is based upon the entire material and not upon a selected part of it (Sverdrup, 1935 a). In fig. 10, n is represented as a function of u_{200}. Values which have been derived from all observations and from observations with southerly wind only, are indicated separately. It will be shown that the increase of n with increasing velocity may be due to the influence of stability.

It should be noted that at high velocities and relatively small stability the logarithmic law gives better approximation than the power law. This is evident from fig. 11 in which the velocities in Groups IV and X are plotted against log. z. It is seen that within these groups the three points fall nearly on the same straight line.

B. Vertical Variation of Temperature.

Tables 2 and 3 contain the vertical distribution of the potential temperature under stable conditions and at different velocities, and in cases when the temperature of the surface was zero degree, If complete similarity existed between the variation of temperature, the temperature variation should in these cases be represented by the equation:

$$\theta = K \ln \frac{z + z_0}{z_0} \text{ or } \theta = \theta_1 \left(\frac{z}{z_1}\right)^{\frac{1}{n}}$$

where z_0 or n should have the values which were derived from the wind observations.

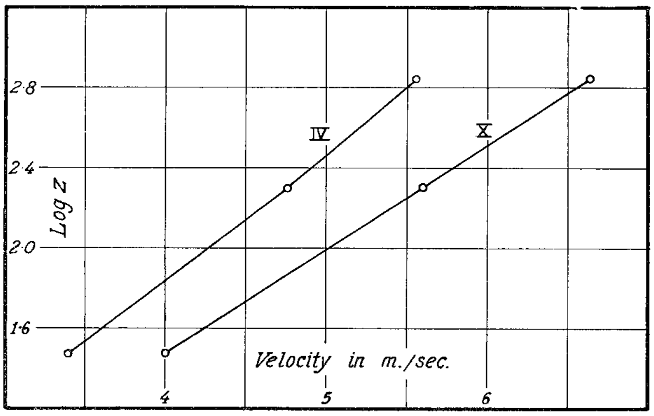


Fig. 11. Velocity as function of log z. High velocity and small stability.

We may expect the logarithmic law to be approximately valid if the wind velocity is high, or if the velocity is moderate and the stability small (Groups IV and X), as in these cases the wind variation can be represented by a logarithmic law. This expectation is fulfilled but we obtain the values $z_0 = 0.0072$ cm and $z_0 = 0.012$ cm, and these are less than one twentieth of the value derived from the wind observations. It was previously shown (fig. 6 p. 26) that the temperature variation approximately follows a logarithmic law up to an altitude of 1 metre. If we especially consider the observations with thermo-couple at the levels 50 cm, 18 cm and 5 cm on August 4th to 5th we find that they can be represented by means of a logarithmic law with $z_0 = 0.032$ cm. This value of z_0 is somewhat greater but is still far too small.

Thus we arrive at the result that z_0 is much smaller when dealing with temperature variation than when dealing with velocity variation and this conclusion is confirmed by the temperature observations above the sea which have been undertaken by Wüst (1920), who found:

Height in cm	0	20	50	120	200
Average temp. °C	15.27	14.32	14.03	13.88	13.80
Difference in potential temp.	0.00	—0.95	—1.24	—1.38	—1.45

The temperature at the level $z = 0$ represents the temperature of the sea surface. In the last line the differences in potential temperature between the surface and the various levels are given, and in fig. 12 these differences are plotted against log z. They group themselves fairly well around a straight line which gives $z_0 = 0.26$ cm. From simultaneous wind observations, Rossby and Montgomery found $z_0 = 4$ cm, or, in agreement with our results, a much greater value.

These results imply that A and η cannot be identical in the immediate vicinity of the boundary surface. If we assume that conditions are stationary and

$$\tau = \eta \frac{du}{dz} = \text{const. and } Q = c_p A \frac{d\theta}{dz} = \text{const.},$$

and that both velocity and temperature distributions are given by logarithmic laws, we obtain:

$$\eta = K_1 \ (z + z_0) \text{ and } A = K_2 \, (z + z_0').$$

Since z_0 and z_0' both are very small quantities we have $A = \eta$ at some distance from the surface if $K_1 = K_2$. Very near the surface we have:

$$\frac{A}{\eta} = \frac{z + z_0'}{z + z_0} \text{ or} : z \to 0 \ \frac{A}{\eta} \to \frac{z_0'}{z_0} \approx \frac{1}{20}.$$

The explanation of this result is probably that a semi-laminar boundary layer exists (see p. 18) within which the transfer of momentum and heat follows different rules. We may conceive that a small air-mass, which replaces one portion of the semi-laminar

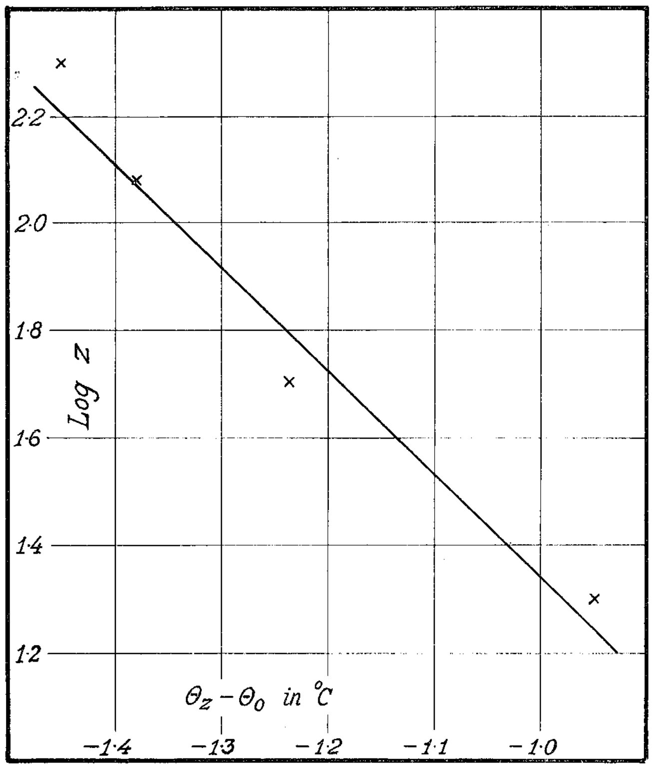


Fig. 12. Temperature differences as function of log z according to Wüst's observations over the sea.

boundary layer, comes completely to rest and, thus, in a very short time attains velocity zero, but that it does not immediately attain the temperature of the surface. This would imply that at the surface momentum is lost more rapidly than heat, for which reason the temperature gradient at the surface becomes relatively greater than the velocity gradient.

Instead of introducing a value of A which very near the surface is smaller than the corresponding value of η and approaches η rapidly with increasing height, it is better to introduce a value of A which is identical with η from a certain distance and upwards, and disregard conditions in the immediate vicinity of the surface, or within the semi-laminar boundary layer. We can do this by writing:

$$\theta - \theta_0 = K \log \frac{z+z_0}{z_0},$$

where z_0 is now identical with the value derived from wind observations and where θ_0 means the temperature of the air at $z=0$ which now is supposed to differ from the temperature of the surface. In Groups IV and X, dealt with above, we obtain $\theta_0 = 0.45^\circ$ and 0.17° respectively, instead of zero. These and other values will later on be used for a computation of the average thickness of the semi-laminar boundary layer.

In all other cases in which the temperature distribution cannot be represented by a logarithmic law, it can, with great approximation, be represented by a power law. If we determine n on the assumption that $\theta_0 = 0^\circ$, we find, however, other values of n than those derived from the wind observation. This is quite in agreement with the fact that we found other values of z_0. In that case we assumed the values of z_0 to be identical and found that the air temperature at $z=0$ differed from the temperature of the surface. Similarly we now introduce the values of n which are derived from the wind observation and obtain $\theta_0 > 0^\circ$. This actually implies that, looking away from the processes in the immediate vicinity of the surface, we assume $A=\eta$ below 7 metres even under stable conditions, an assumption which already has been justified by means of Taylor's criterion (p. 18). The correctness of this assumption is, furthermore, supported by the fact that the values of n, which are computed by means of the temperature distribution putting $\theta_0 = 0^\circ$, are greater than the corresponding values of n from the velocity distribution, If this were correct it would mean that A increased more rapidly with elevation than η, and since A and η must be practically identical at a short distance from the surface, it would mean $A > \eta$ at a greater elevation. This, probably, cannot be the case and, therefore, we are justified in proceeding in the manner which was described.

In order to explain that we obtain $\theta_0 > 0^\circ$ we must again consider the semi-laminar boundary layer and assume that within this layer the transfer of momentum and the transfer of heat take place according to different laws.

Writing $\theta = \theta_0 + \triangle \theta_1 z^{\frac{1}{n}}$, we obtain from the average values in tables Nos. 2 and 3:

Group No.	I	II	III	IV	V
θ_0:	0.26	0.36	0.63	0.11	0.31
$\triangle \theta_1$:	0.10	0.35	0.99	0.19	0.36
Group No:	VI	VII	VIII	IX	X
θ_0:	0.55	0.42	0.19	0.28	0.30
$\triangle \theta_1$:	0.83	0.52	0.61	0.46	0.44

It may be added that we can write $\theta_0 = \triangle \theta_1 (z+z_0)^{\frac{1}{n}}$, (see p. 9) and obtain $\theta_0' = \triangle \theta_1 z_0^{\frac{1}{n}}$. With $z_0 = 0.25$ cm and with the above values of $\triangle \theta_1$ and the values of n which are given on pages 28 and 29 we obtain:

Group No.	I	II	III	IV	V
θ_0':	0.07	0.26	0.71	0.16	0.28
Group No.	VI	VII	VIII	IX	X
θ_0':	0.64	0.37	0.47	0.36	0.37

These values do not deviate very much from the values of θ_0 and the vertical variation of temperature is, therefore, well represented by an equation of the type (26) p. 9.

We have until now considered the vertical distribution of temperature under stable conditions, but have used selected observations only (southerly wind and positive temperatures at all levels). The results are similar if we consider all other observations which are taken under stable conditions except in the situations in which we had a geostrophic wind with a maximum velocity below 7 metres. Such conditions were present on August 11th to 15th. On these days we find an unusual large increase of temperature between 1 m and 5 m which indicates that A (and η) decreased with height above a level of 1 to 2 metres.

When instable conditions prevail, the temperature decrease between 1 m and 5 m is, on the other hand, exceptionally small, as seen from the values in table 1, page 27. This may indicate that at short distance from the boundary surface the turbulence attains another character if the stratification is instable, but our observations are insufficient for a further discussion of this question. We are here mainly concerned with the conditions in the immediate vicinity of the boundary surface and there the variations of velocity and temperature are similar, as evident from a comparison between figs. 6 and 7. Both variations can be represented by logarithmic laws.

Our final conclusion is, therefore, that the vertical variation of temperature can be represented by a logarithmic law of the type (11) if (*i*) conditions are instable, but in that case the law is valid for a few metres only, or (*ii*) if conditions are stable but the wind velocity strong. If conditions are stable, we must, however, generally use interpolation formulae of the types:

$$\theta = \theta_0 + \triangle\,\theta_1\, z^{\frac{1}{n}} \quad \text{or} \quad \theta = \triangle\,\theta_1\,(z+z_0)^{\frac{1}{n}}$$

where θ_0 is a fictive air temperature at the level $z=0$.

C. Vertical Variation of the Vapour Pressure.

On days with southerly wind and positive temperatures the vertical variation of the wapour pressure is quite similar to the variation of temperature. The average values at 500 cm, 100 cm and 4 cm are given in tables 2 and 3.

Under the stated conditions the vapour pressure at the surface must be equal to the maximum pressure at zero degree, 4.58 mm, but if we attempt to represent the differences (e — 4.58) by means of a logarithmic law in the case of strong wind and small stability, or by means of power laws in all other cases, we obtain values of z_0 which are smaller than those derived from the variation of velocity, or values of n which are greater. If we introduce the values of z_0 or n which are obtained from the velocity distribution we obtain a vapour pressure at $z=0$ which is greater than 4.58. Introducing a logarithmic law when dealing with the values in group X, which were previously treated in a similar manner we obtain $e - 4.58 = 0.13$ mm. By means of a power law we obtain from the data in tables 2 and 3, writing:

$$e = e_0 - 4.58 + \triangle\, e_1\, z^{\frac{1}{n}}:$$

Group No.	I	II	III	IV	V
$e_0 - 4.58$:	0.03	0.16	0.14	0.01	0.14
$\triangle\, e_1$:	0.04	0.10	0.33	0.02	0.06
Group No.	VI	VII	VIII	IX	X
$e_0 - 4.58$:	0.33	0.12	0.11	0.11	0.11
$\triangle\, e_1$:	0.21	0.17	0.15	0.10	0.10

If we consider groups VII to X which contain all data, we obtain the average value $e_0 - 4.58 = 0.11$, or $e_0 = 4.69$ mm. The corresponding value of the temperature θ_0 is 0.31°, at which temperature the maximum vapour pressure is 4.68 mm. On an average the value of e_0 represents, therefore, the maximum vapour pressure at the computed temperature θ_0. This indicates that within the semi-laminar boundary layer the temperature and the vapour contents are transferred nearly according to the same law.

The question whether or not the vertical variation of vapour pressure is similar to the variation of temperature, is of considerable importance. In the above cases the variations are similar, but the agreement is of no value, since the air vas practically saturated with water vapour and since the vapour was transported towards the surface. Under these conditions the variation of the vapour contents is not independent of the variation of temperature.

In order to examine if similarity actually exists we have to study the cases in which the temperature increased and the vapour pressure decreased with height. Under these conditions the air is very dry, heat is conducted towards the surface, water vapour away from the surface, and the distribution of water vapour is independent of the distribution of temperature. Unfortunately, our determinations of the vapour pressure are less accurate in the cases in question (see p. 26), but omitting the most uncertain values we find the data in tables 4.

Mean values are computed for three velocity groups. In the middle group the temperature difference between 1 m and 5 m is very great, indicating that cases have been included with geostrophic wind. The vapour pressure shows, correspondingly, an abnormal decrease and there exists, therefore, agreement between the variation of temperature and vapour pressure. Turning to the two other groups, we find that the vertical distribution of temperature and humidity can be represented very closely by the equations

$$(51) \quad \theta = -0.74^\circ + 0.72\, z^{\frac{1}{5.0}} \quad e = 4.49 - 0.16\, z^{\frac{1}{5.0}}.$$

$$(52) \quad \theta = -1.19^\circ + 0.53\, z^{\frac{1}{5.9}} \quad e = 4.52 - 0.30\, z^{\frac{1}{5.9}}.$$

Thus, the distribution of temperature and vapour pressure are quite similar even when these distributions are independent of each other. Since the variation of vapour pressure is independent of processes of radiation it follows that the vertical variation of tem-

Table 4. *Average values of wind, potential temperatures and vapour pressure from hours with low relative humidity.*

Average velocity at 2 m and 30 cm in m/sec.		< 2.0	2.0 to 2.95	> 3.0
Wind velocity in m/sec. at levels	700 cm	1.54	2.73	5.53
	200 cm	1.36	2.25	4.85
	30 cm	0.81	1.46	3.23
Velocities relative to velocity at 200 cm	700 cm	1.13	1.21	1.14
	200 cm	1.10	1.00	1.00
	30 cm	0.60	0.64	0.67
Potential Temperature in °C at levels	500 cm	1.74	1.60	0.34
	100 cm	1.09	0.79	0.01
	4 cm	0.20	0.16	—0.54
Vapour pressure in mm Hg. at levels	500 cm	3.94	3.86	3.64
	100 cm	4.01	4.01	3.86
	4 cm	4.18	4.18	4.14
Number of hours		44	32	34
Value of n		5.0	5.1	5.9

perature within the lowest metres is also independent of processes of radiation and depends upon the turbulence only. We must add the reservation that conditions may be different when the vapour content is very great.

We encounter, however, a serious difficulty when we are to interpret the contents of the above equations (51) and (52). Since the temperature increases with height and the vapour content decreases, it is evident that heat is transported towards the surface but vapour is transported away from the surface. Evaporation takes place. It follows that within the semi-laminar boundary layer the temperature must increase with elevation but the vapour pressure must decrease. Assuming conditions to be similar to those which were found when the surface was melting, we obtain that the surface temperature must have been approximately —0.9° and —1.4°. At these temperatures the maximum pressure of the water vapour over ice, is, however, 4.25 mm and 4.08 mm respectively whereas we, from our equations, should have expected values greater than 4.49 mm and 4.52 mm. The discrepancy seems too great to be explained as a result of errors of observations.

This is the only case in which the observations do not fit into the system, which, otherwise, appears to be very satisfactory. The low temperature near the surface is in this case, no doubt, the effect of cooling due to evaporation; our observations show clearly a "psycrometer effect." There seems, however, to be no reason why the exchange in this case should follow other laws, and the discrepancy, perhaps, shows that our system is more or less incorrect. In view of the fact that this system, on the whole, is satisfactory, it is hoped that a more detailed study of similar conditions will lead to an explanation.

D. The Average Thickness of the Semi-Laminar Boundary Layer.

We can compute the average thickness of the semi-laminar boundary layer if we assume that outside of the boundary layer the eddy conductivity up to an altitude of 30 cm is a linear function of $(z+z_0)$ and can be represented by the formula (12) (p. 7):

$$A = \varrho k_0^2 (z+z_0) \frac{u_{30}}{\ln \frac{30}{z_0}},$$

where $k_0 = 0.38$. It will later on be shown that this assumption is approximately correct. With $\varrho = 1.17 \times 10^{-3}$ and $z_0 = 0.25$ cm we obtain.

$$A = 0.35 \times 10^{-4} u_{30} (z+z_0).$$

Furthermore, we assume that below 1 metre the temperature distribution can be represented by a logarithmic law (p. 17). From the observations in 100 cm and 4 cm we obtain

$$K = \frac{\theta_{100} - \theta_4}{\ln (100+z_0) - \ln (4+z_0)} = \frac{\theta_{100} - \theta_4}{3.16}$$

and

$$\theta_0 = \theta_{100} - \frac{\ln (100+z_0) - \ln z_0}{\ln (100+z_0) - \ln (4+z_0)} (\theta_{100} - \theta_4) =$$

$= \theta_{100} - 1.90\ (\theta_{100} - \theta_4)$, if $z_0 = 0.25$. Therefore

$$\frac{d\theta}{dz} = 0.317\ (\theta_{100} - \theta_4) \frac{1}{z+z_0}.$$

The transport of heat is

$$Q = c_p A \frac{d\theta}{dz} = 0.026 \times 10^{-4}\ u_{30}\ (\theta_{100} - \theta_4).$$

Within a laminar boundary layer we have, on the other hand,

$$Q = \lambda \frac{\triangle \theta}{\triangle z},$$

where λ is the heat conductivity of the air, $\triangle \theta$ the temperature difference between the surface and the upper limit of the boundary layer, and $\triangle z$ the thickness of the boundary layer. With $\lambda = 0.58\ 10^{-4}$ we obtain

$$\triangle\,\theta = 0.045\ u_{30}\ (\theta_{100} - \theta_4)\ \triangle z.$$

This temperature difference is equal to

$$\triangle\theta = \theta_{100} - 1.90\ (\theta_{100} - \theta_4) + 0.317\ (\theta_{100} - \theta_4)\ \ln \frac{\triangle z + z_0}{z_0}\ ,$$

if the logarithmich law is valid from the upper limit of the boundary layer and upwards. This condition expresses that we assume the temperature distribution to be continuous but the conductivity to be discontinuous at the upper limit of the boundary layer. The latter assumption cannot be correct but the computation ought, since we are dealing with a statistical quantity, to give an idea of the thickness of the semi-laminar layer.

When we consider the water vapour we obtain similarly:

$$\triangle\, e = 0.040\ \ (e_{100} - e_4)\ u_{30}\ \triangle z = (e_{100} - 4.58) - 1.90$$
$$(e_{100} - e_4) + 0.317\ (e_{100} - e_4)\ \ln \frac{\triangle z + z_0}{z_0}.$$

By means of these formulae and the mean values in table 3 we obtain:

Group No.	VII	VIII	IX	X
$\triangle z$ from temp.obsv. cm	0.10	0.08	0.09	0.07
$\triangle z$ » humidity obsv. cm	0.06	0.13	0.13	0.13

Thus, the average thickness of the semi-laminar boundary layer appears to be independent of the wind velocity. The values which are derived by means of the observations in Group VII are uncertain, since the velocity is very small, and if these are disregarded we obtain the mean values:

$\triangle z$ from temperature observations: 0.08 cm.
$\triangle z$ » humidity —»— 0.13 cm.

If this difference is real it may indicate that processes of radiation are of importance to the exchange of heat within the semi-laminar boundary layer. In our preceding equations we may replace λ by $\lambda + \lambda_r$, where λ_r represents a coefficient of radiative conductivity (see Brunt, 1934 p. 115). With $\lambda_r = 0.35 \times 10^{-4}$ we then obtain $\triangle z = 0.13$ cm from the temperature observations as well. This circumstance explains, perhaps, the discrepancy which was pointed out in the preceding chapter.

Wüst's observations which are represented in figs. 4 and 12 give similar results. It was pointed out that these observations also indicate the existence of a semi-laminar boundary layer, since we obtained $z'_0 = 0.26$ cm from the temperature observations, whereas Rossby and Montgomery found $z_0 = 4$ cm from the wind observations. If we compute $\triangle z$ as above, introducing $z_0 = 4$ cm, we obtain from Wüst's data:

$\triangle z$ from temperature observations: 0.03 cm.
$\triangle z$ » humidity » 0.09 cm.

It is, in the first place, noteworthy that the value derived from the humidity observations does not deviate much from our value from the plateau, although the roughness parameters are 4.0 and 0.25, respectively. Secondly, that the temperature observations again give a smaller value. If we introduce a coefficient of radiative conductivity, λ_r, we find now: $\lambda_r = 1.2 \times 10^{-4}$. This value is more than three times the value from the plateau, but on the plateau the vapour pressure at the surface was only 4.58 mm, whereas, during Wüst's observations, it was 12.89, and we must expect the radiative diffusivity to increase with increasing vapour pressure. Further studies of these questions are of the greatest importance to the problems dealing with exchange of heat and water vapour between the air and the sea.

The computations show, at all events, that the average thickness of the semi-laminar boundary layer must be very small and of the order of magnitude of 1 mm, as already pointed out by Brunt (1934 p. 263). Within short intervals of time a laminar layer of thickness of several millimetres may, therefore, exist, and laminar boundary layers of this thickness have been observed by Büttner (1934).

6. The Exchange of Heat and Water Vapour between the Snow Surface and the Air.

We have seen that, above the semi-laminar boundary layer, we can represent the temperature and the vapour pressure as functions of elevation by means of simple power laws. We can, therefore, compute the eddy conductivity if we can determine the amounts of heat and water vapour which are exchanged between the surface and the air. This exchange will now be considered.

The snow surface receives heat by incoming (short-wave) radiation, by conduction from the air if the air temperature increases with height, and by conduction from below if the temperature of the snow increases with depth. Furthermore, heat is received if the content of water vapour in the air increases with height and, therefore, water vapour is transported towards the surface where it condenses and the heat of condensation is set free.

The surface loses heat by outgoing (long-wave) radiation, by conduction to the air, if the air temperature decreases with height, and by conduction downwards, if the snow temperature decreases with depth.

The albedo of the snow will be called a and the radiation income I, the absorbed part of the incoming radiation is then $(1-a)\ I = \alpha I$. The heat lost by long-wave radiation will be called R and the total amount of heat which the surface receives because of radiative processes is, therefore, $(\alpha I - R)$. The amount of heat which is received from or lost to the air will be called Q_a, the amount of heat which is received or lost by conduction from below will be called Q_s. All amounts of heat will be given in gramme calories per square cm and per minute. The amount of water vapour which is transported towards the surface if condensation takes place, and away from the surface if evaporation takes place, will be called F.

The surplus of heat which the surface receives is used for melting and (or) evaporation. The total ablation of the surface, measured in cm of water per minute will be called H, the thickness of the layer which is evaporated will be called h and the thickness of the layer which is melted is, therefore, $(H-h)$. The amounts of heat used for melting and (or) evaporation are, therefore (in gramme calories per cm^2 and minute) $80\ (H-h)$ and (or) $680\ h$ since 80 gramme calories are needed for melting one gramme of snow and 680 gramme calories are needed for evaporating one gramme of snow.

Let us first consider the conditions when no evaporation takes place. In this case we have

$$80\ H = \alpha I - R + Q_a + Q_s + 600\ F, \tag{53}$$

where the terms Q_a and Q_s are positive if heat is conducted to the surface, and negative if the surface loses heat by the processes of conduction, and where the term $600\ F$ represents the amount of heat which is liberated by condensation on the surface of F gr of water vapour per minute.

If melting and evaporation take place we have:

$$80\ (H-h) + 680\ h = 80\ H + 600\ h = \alpha I - R + Q_a + Q_s.$$

But $h = F$, or equal to the amount of water vapour which is transported away from the surface. We obtain, therefore, in this case as well:

$$80\ H = \alpha I - R + Q_a + Q_s + 600\ F,$$

where F is negative since water vapour is transported away. Equation (53) is, therefore, valid, in general) if all terms are given their proper signs.

Since our observations were taken in a season when the temperature of the upper layers of the snow generally was at freezing point, except in short periods when a crust froze, we put $Q_s = 0$, but we must be aware that this condition is not always fulfilled within short intervals of time.

With $Q_s = 0$ we obtain:

$$Q_a + 600\ F = 80\ H - \alpha I + R. \tag{54}$$

In order to compute the amounts of heat and water vapour which are exchanged between the surface and the air we must know all quantities on the right hand side of (54).

Bi-hourly values of H have been published by Ahlmann (1935) and, for sake of completeness, the daily values of H are presented in table VI.

Daily sums of I will, later on, be published by H. Olsson, who has kindly placed his preliminary values at my disposal. These are communicated in table VII. Our equation contains, however, two unknowns, α and R and we must now attempt to determine these quantities.

The albedo $a = (1-\alpha)$ was measured by H. Olsson, who found values ranging between 54 % and 84 %. A definite relation existed between the character of the snow surface and the albedo; the latter was, as a rule, small when the surface was wet, larger when the surface was frozen, and largest on a few occasions when the surface was covered by newly fallen snow.

In order to examine the constancy of the albedo the preliminary values have been grouped according to the temperature at 4 cm, treating separately the cases in which the surface was wet or frozen. Cases with newly fallen snow were included in the latter group. The temperature at 4 cm was always positive when the surface was wet and with a frozen surface it was negative, except on days when the air was very dry. The average values are:

Surface wet	ϑ_4, °C :	0.35	0.76	1.36
	Albedo, %:	62	64	65
Surface frozen	ϑ_4, °C :	—1.63	—0.75	0.74
	Albedo, %:	76	72	73

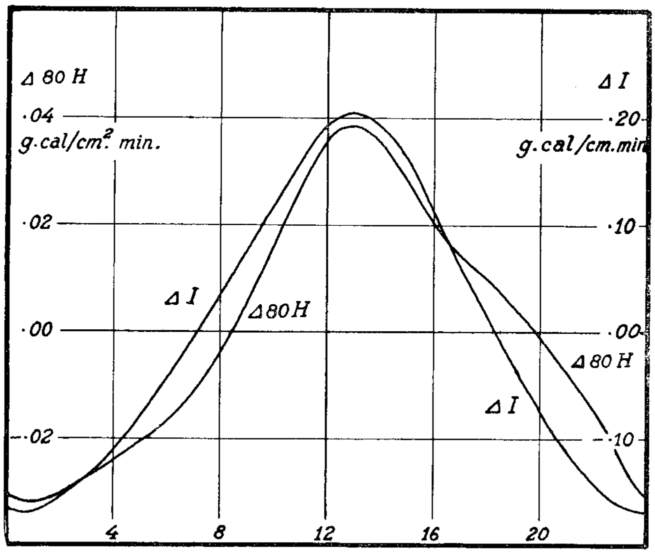


Fig. 13. Diurnal variation of ablation and radiation income on overcast days.

According to this compilation the mean values are: Surface wet: Albedo 64 %, surface frozen: Albedo 74 %.

By combining our different observations we can undertake an independent determination of the albedo and simultaneously determine a probable value of R as a function of the cloudiness. For this purpose we avail ourselves of the fact that on completely overcast days with a melting surface, the diurnal variation of temperature and humidity at a few metres above the surface is negligible and the same applies to the diurnal variation of the outgoing radiation. On such days the only elements which show a diurnal variation are the ablation H and the radiation income I. Indicating mean diurnal values by bars and the deviation from means by $\triangle$, we have:

$$80\,(\overline{H}+\triangle H) = a\,(\overline{I}+\triangle I) - \overline{R} + \overline{Q}_a + 600\,\overline{F}$$

and, since $\Sigma \triangle H$ and $\Sigma \triangle I$ by definition are zero:

$$80 \triangle H = a \triangle I \text{ or } 80 \Sigma |\triangle H| = a \Sigma |\triangle I|.$$

The radiation income was recorded by means of a Robitsch actinograph. From the records of this instrument mean diurnal values of I are obtained with sufficient accuracy, but hourly values are uncertain and will, not therefore, be published in the forthcoming paper by H. Olsson, who has, however, placed the values, which are read off from the records, at my disposal. Single values may be 10 to 20 per cent in error but the means from several days will be more correct and are, no doubt, sufficiently correct for our purpose.

The meteorological conditions, which must be satisfied if the above relations are valid, were present on 11 days but satisfactory records of the ablation were obtained on 7 days only: July 18, 19, 20, August 1, 2, 4, 9. For these days the mean values are: $80\,\overline{H}=0.077$, $\overline{I}=0.222$ and in 2-hourly intervals the deviations from the means are:

Interval of time	0^h-2^h	2^h-4^h	4^h-6^h	6^h-8^h	8^h-10^h	10^h-12^h	12^h-14^h	14^h-16^h	16^h-18^h	18^h-20^h	20^h-22^h	22^h-24^h
$80\,\Delta H$ (g cal./cm² min.)	—.031	—.026	—.021	—.011	.006	.027	.038	.029	.014	.005	—.008	—.024
ΔI (g cal./cm² min.) . . .	—.164	—.128	—.088	.007	.084	.159	.203	.164	.065	—.034	—.112	—.161

The values are smoothed by the formula

$$\frac{1}{4}(a+2\,b+c).$$

They are represented graphically in fig. 13 in which curves are drawn. It is seen that the heat used for ablation shows a diurnal variation which is quite similar to the diurnal variation of the radiation income except for a small phase difference. The maximum of the ablation comes somewhat later than the maximum of radiation and the whole ablation curve is somewhat displaced towards the afternoon. This displacement may be due to the circumstance that the ablatograph has not reacted instantaneously but the record has lagged somewhat behind the actual process. If this be true, the recorded amplitude of the ablation must be too small, and the same applies to the deviations from the mean values. By means of the deviations we obtain

$$\alpha = 1 - a = 0.18, \qquad a = 82 = 82\,\%.$$

This value of a is, no doubt, too low, owing to the lag of the ablatograph, and, therefore, we will, provisionally, adopt the value:

$$\alpha = 1 - a = 0.30, \qquad a = 0.70 = 70\,\%.$$

Even this value of α, which gives an albedo of 70 per cent, appears too low when compared with Olsson's measurements of the albedo, which gave 64 % on overcast days with a melting surface.

In order to determine R on overcast days we make use of a method which has been introduced by A. Ångström (1934) in his discussion of the results from The Swedish-Norwegian Expedition to North-East Land (Spitbergen) in 1931. Referring to results from lower latitudes Ångström puts $R = \beta I$, but this relation is evidently valid only when dealing with mean diurnal values. Anticipating subsequent results (p. 42) we, furthermore, introduce:

$$(55) \quad Q_a + 600\,F = \delta' \left((\vartheta_2 - \vartheta_0) + 2.27\,(e_2 - e_0) \right) u_3 = \delta' M,$$

where u_3 is the wind velocity at the altitude z_3, ϑ_2 and e_2 are temperature and vapour pressure at the altitude z_2 and ϑ_0 and e_0 represent temperature and vapour pressure at $z = 0$. Thus:

$$80\,H = (a - \beta)\,I + \delta' M = \gamma\,I + \delta' M.$$

The factor δ' in (55) is, as will be shown later on (p. 42), a function of n and at present it is, therefore, preferable to use days only on which n had nearly the same value. Futhermore, it is preferable to use days with strong wind, since the total ablation was great on such days. We find four days which satisfy these conditions: July 18, 19, 20 and August 2. On these days we have:

Day	July 18	July 19	July 20	August 2
I g cal./cm² min.	0.271	0.199	0.206	0.206
$80\,H$ g cal./cm² min. ..	0.069	0.075	0.053	0.130
Value of n (from wind observation)	6.2	5.8	6.4	5.6
ϑ_0, °C	0.28	0.30	0.19	0.52
e_0, mm	4.76	4.76	4.67	4.84

When computing M we have combined the observations of temperature and vapour pressure at 4 cm, 100 cm and 500 cm with the wind observations at 30 cm, 200 cm and 700 cm, respectively, or we have successively put (i) $z_2 = 4$ cm, $z_3 = 30$ cm (ii) $z_2 = 100$ cm, $z_3 = 200$ cm and (iii) $z_2 = 500$ cm, $z_3 = 700$ cm. The mean diurnal values of temperatures, vapour pressures and velocities are found in tables II and III. By means of the method of least squares we obtain:

From observations at	4 cm and 30 cm	100 cm and 200 cm	500 cm and 700 cm
Value of γ:	0.16	0.15	0.17

The mean value is $\gamma = 0.16$. The value of δ' is at present of no interest but will be dealt with later on. Since we have adopted $a = 0.30$ we obtain on these days, $\beta = 0.14$ or, as $\bar{I} = 0.227$, $R = 0.032$. This value is a probable one since fog prevailed on most of the days. In the presence of fog, therefore, we adopt: $R = 0.030$.

We can obtain a check on this result by looking up the hours when the temperature was nearly constant at zero degree from 4 cm to 5 m and the sky was completely overcast. We find 36 hours in which these conditions are fulfilled. The mean values are:

Height cm	Temperature °C	Vapour pressure, mm Hg
500	0.05	4.20
100	0.04	4.24
4	0.01	4.43

and

$80\,H = 0.052$ g cal./cm² min., $I = 0.432$ g cal./cm² min.

The temperature is so nearly constant that no heat is conducted towards the surface but water vapour is transported upwards. Anticipating subsequent results we can estimate $Q_a + 600\,F$ to -0.009 g cal./cm² min. The surface was partly wet, partly frozen in the 36 hours in question, and we introduce, provisionally, an average albedo of nearly 70 % in agreement with Olsson's results. We obtain:

$$0.052 = 0.30 \times 0.432 - R - 0.09$$
$$R = 0.068.$$

The fact that we now find a greater value of R is understandable, since fog was present on only 8 of the 36 hours in question, but the value $R = 0.068$ with a completely overcast sky appears too high and it seems, therefore, that we must use an albedo value which is somewhat greater than that which was observed by Olsson. If we were to use the observed value 64 %, we would, in the presence of fog, obtain $R = 0.045$ and this value is unreasonably great. It must be emphasized that the determination when fog is present must be considered trustworthy, since it is based on good observations. Furthermore, it must be considered that if an albedo of 64 % is introduced, it is necessary to assume that the ablatograph has recorded less than half of the actual diurnal variation and this assumption is hardly justified. We, therefore, adopt the values $a = 70$ % and $R = 0.030$ in the presence of fog and with a wet surface.

A similar investigation on clear days with a mostly frozen surface gives uncertain results since

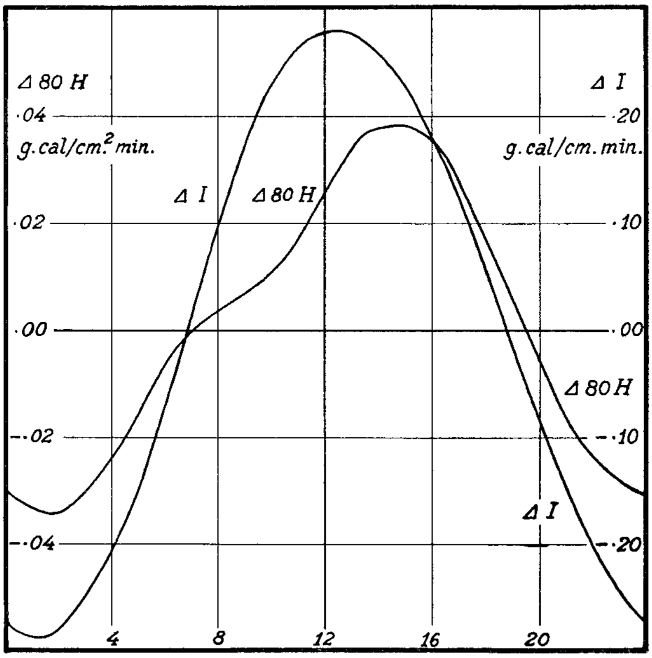


Fig. 14. Diurnal variation of ablation and radiation income on clear days.

the clear periods were few and short. We find only three 24-hourly periods with cloudiness less than 2, namely July 14, 18^h to July 16, 18^h and July 26, 6^h to July 27, 6^h. The mean values of $80\,H$ and I are 0.053 and 0.501, respectively. The smoothed values of $80 \triangle H$ and $\triangle I$ are:

Interval of time	0^h—2^h	2^h—4^h	4^h—6^h	6^h—8^h	8^h—10^h
$80\,\Delta H$ (g cal./cm² min.)	—.033	—.030	—.015	.000	.007
ΔI (g cal./cm² min.)...	—.286	—.249	—.146	.014	.178

	10^h—12^h	12^h—14^h	14^h—16^h	16^h—18^h
$80\,\Delta H$	.017	.034	.038	.027
ΔI	.264	.271	.227	.126

	18^h—20^h	20^h—22^h	22^h—24^h
$80\,\Delta H$	.005	—.017	—.028
ΔI	—.011	—.146	—.245

The values are represented in fig. 14 from which it is seen that on these days the ablation curve is displaced more towards the afternoon than in the preceding case. This circumstance and the fact that on two of the nights which are included a thin crust froze on the snow, tend to reduce the amplitude of the ablation and the hourly deviation from mean. By means of the deviations we obtain

$$\alpha = 0.12, \qquad a = 0.88 = 88\,\%.$$

The value of α is, in agreement with Olsson's result, lower than the corresponding value on overcast days but is, no doubt, considerably too high since the amplitude of the ablation was not recorded correctly. The displacement of the curve showing the ablation is even greater than in the preceding case. In view of this circumstance it seems reasonable to adopt the value 77 % which lies again somewhat above Olsson's result.

In order to obtain an idea of the magnitude of R on clear days with a frozen surface the clear hours ($C < 4$) have been selected at which the air temperature was constant up to 5 metres and about 0°. Only 14 such hours were found and the mean values were:

Height cm	Temperature °C	Vapour pressure, mm Hg
500	—0.18	4.21
100	—0.26	4.18
4	—0.31	4.36

$$80\,H = 0.042 \text{ g cal./cm}^2 \text{ min.}, \quad I = 0.753 \text{ g cal./cm}^2 \text{ min.}$$

Some heat is lost by evaporation since the contents of water vapour decreases with height. This amount may be estimated at 0.008 g cal./cm² min. and, therefore, we have:

$$0.042 = 0.23 \times 0.753 - R - 0.008,$$
$$R = 0.123,$$

at an average cloudiness of 2.9, but the value is uncertain since it is based on a small number of observations.

Our computation of R at cloudiness 10 must now be revised since we have introduced an average albedo of 73.5 %, or an average value of α equal 0.265. With this value we obtain at cloudiness $10 : R = 0.053$, but this value is still somewhat uncertain. Considering these circumstances and the results from other localities we put generally

$$R = 0.160\,(1 - 0.075\,C),$$

adding

$$R = 0.030, \text{ when fog is present.}$$

With this value of R and with the albedo equal to 77 % in case the surface is frozen, and equal to 70 % if the surface is wet, our different observations are brought into mutual agreement, but a discrepancy exists between the adopted values of the albedo and the observed ones. The difference is, however, not very great and amounts, as will be shown, on an average to less than 4 %. Two explanations of this discrepancy can be offered. It is, in the first place, known that the albedo changes with the altitude of the sun and is greater at low altitudes. Most measurements have been taken with a high sun and the observed values must, therefore, be somewhat above the 24-hourly mean value which enters into most of our computations. In the second place, it is possible that certain instrumental difficulties are met with when measuring the albedo. The energy spectrum of the reflected radiation, no doubt, differs from the energy spectrum of the incoming radiation and this circumstance may lead to some uncertainty in the values which are observed by means of any type of actinometer. In view of these possible sources of error it appears justified to use the somewhat higher albedo-values which are obtained by a co-ordination of various sets of observations.

Now we can compute $Q_a + 600\ F = 80\ H - (a\ I - R)$ for any given period. If this period comprises complete days only, we use the daily sums of the ablation and the radition income, as given in tables VI and VII and compute R from the average cloudiness, taking into account the number of hours with fog in the period concerned. If fractions of days are included, use is made of the hourly values of the ablation in table 1 in Ahlmann's publication (1935) and of the hourly values of I as communicated to the author by H. Olsson. The value of the albedo which is to be used depends upon the average character of the snow surface.

7. The Eddy Conductivity.

A. Computation of the Eddy Conductivity from the Observations on Isachsen's Plateau.

Knowing the value of $Q_a + 600\ F$ in a sufficiently long period we can compute the eddy conductivity A, since under stationary conditions:

$$Q_a = c_p A \frac{d\theta}{dz},\quad F = A \frac{df}{dz} = A \frac{0.623}{p} \frac{de}{dz}.$$

We avail ourselves of the empirical interpolation formulae:

$$\theta - \theta_0 = \triangle\theta_1 \left(\frac{z}{z_1}\right)^{\frac{1}{n}},\quad e - e_0 = \triangle e_1 \left(\frac{z}{z_1}\right)^{\frac{1}{n}}$$

from which follows: $A = A_1 \left(\frac{z}{z_1}\right)^{\frac{1}{n}}$.

For the sake of convenience we put $z_1 = 1$, but we must be aware that $\triangle\theta_1$, $\triangle e_1$, and A_1 then lose their proper dimensions. With $z_1 = 1$ we obtain

$$Q_a + 600\ F = A_1 \frac{c_p}{n} (\triangle\theta_1 + m \triangle e_1)\ 60, \tag{56}$$

where

$$m = \frac{600}{c_p} \cdot \frac{0.623}{p} = 2.27,$$

when the average value: $p = 690$ mm is introduced, and where the factor 60 enters since $(Q_a + 600\ F)$ is measured in gramme calories per square cm and per minute, whereas otherwise we use the C. G. S.-system. We have disregarded the fact that a very small amount of water vapour is condensed in the air if the air is saturated with water vapour at some distance from the surface. It can be shown, that at temperatures near 0^0 this amount is too small to be considered.

From (56) and (54) we obtain:

$$A_1 = \frac{(80\ H - a\ I - R)\ n}{c_p (\triangle\theta_1 + m \triangle e_1)} \cdot \frac{1}{60} \tag{57}$$

In order to compute A_1, we have selected suitable periods, but have made use only of the observation after July 1st since data from 4 cm are not available in June and since humidity observations at 5 m are also lacking for June except June 26th, when, however, the radiation measurements had not commenced. In the first place we have selected four warm and wet periods (I to IV in tables 5 and 6) in which the temperature was always positive and the humidity increased with height. Within each period the wind velocity has been approximately constant. Next we have combined all cold and dry periods in which the temperature near the surface always was so much below freezing point (V, tables 5 and 6) that the surface always remained frozen. Under these conditions no melting took place and the observed ablation was entirely due to evaporation. Similarly we have combined all periods which were definitely warm and dry (VI, tables 5 and 6). Finally, we have taken all intervals between the selected periods. These are of varying character and in them the wind velocity has often been variable.

The periods comprise 1016 of the 1112 available hours. The data from July 3rd and 4th have not been used since the observed humidities are doubtful and from July 25th no record of the ablation is available. Furthermore, a few hours which lie between typical periods have been omitted.

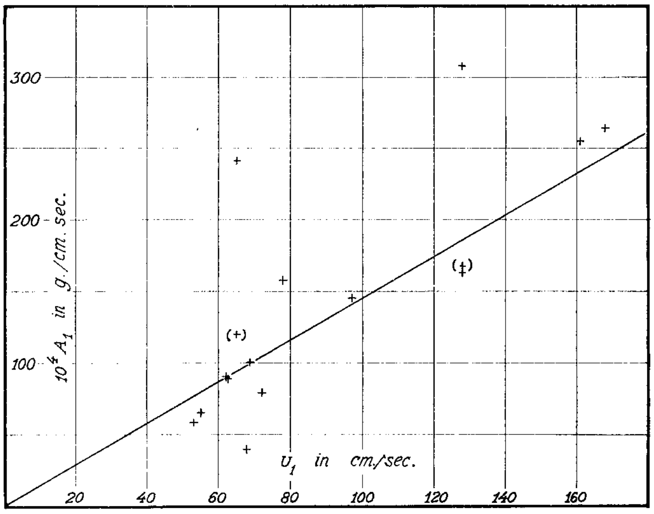


Fig. 15. Relation between the eddy conductivity at 1 cm, A_1, and the extrapolated value of the velocity at 1 cm, u_1.

Thus, the following 14 periods are considered:

Period No.	Length in hours	Character of period	Character of surface	Days included
I	72	Warm, wet	Wet	July 18, 1^h to July 20, 24^h
II	54	Warm, wet	Wet	Aug. 1, 13^h to Aug. 3, 18^h
III	42	Warm, wet	Wet	Aug. 3, 19^h to Aug. 5, 12^h
IV	96	Warm, wet	Wet	Aug. 7, 1^h to Aug. 10, 24^h
V	206	Cold, dry	Frozen	July 5, 1^h to July 8, 8^h July 9, 19^h to July 11, 24^h July 22, 1^h to July 24, 24^h
VI	140	Warm, dry	Partly frozen	July 15, 7^h to July 17, 2^h July 27, 11^h to July 31, 10^h
VII	48	Cold, dry	Partly frozen	July 1, 1^h to July 2, 24^h
VIII	34	Cold, dry	Mainly frozen	July 8, 9^h to July 9, 18^h
IX	72	Warm	Mainly wet	July 12, 1^h to July 14, 24^h
X	36	Warm, wet	Wet	July 17, 7^h to 24^h July 21, 1^h to 18^h
XI	34	Warm	Partly frozen	July 26, 1^h to July 27, 10^h
XII	26	Warm, wet	Wet	July 31, 11^h to Aug. 1, 12^h
XIII	36	Warm, wet	Wet	Aug. 5, 13^h to Aug. 6, 24^h
XIV	120	Warm	Mainly wet	Aug. 11, 1^h to Aug. 15, 24^h

Table 5 contains the mean values of wind velocity at three or four heights and potential temperature and humidity at three heights in these periods. Table 6 contains the observed values of I, the values of a which have been adopted on account of the character of the surface, the values of aI, the computed values of R and the observed values of 80 H, and the differences between the three last quantities. Furthermore, the table contains the values of n_1 u_1, $\triangle \theta_1$, and $\triangle e_1$ which have been derived from the data in the preceding table, and the values of A_1 which have been computed by means of formula (57). The last two columns show the observations which have been used when computing A_1.

In fig. 15 the values of A_1 are plotted against u_1. It is seen that A_1 is nearly a linear function of u_1 and that most points fall near the line $A_1 = 1.45 \times 10^{-4}\, u_1$.

Periods IX and X give both large values of A_1 but within both periods the temperature and humidity gradients were small and errors in the determinations of these gradients exercise, therefore, a considerable influence upon the result. The deviations are, however, too great to be explained entirely by such errors and it is more probable that the discrepancy, if real, is due to an error in the assumed albedo. If we introduce the average value on overcast days as derived from Olsson's observations, $a = 64\,^0/_0$ or $a = 0.36$ we obtain in period IX: $A_1 = 121$ and in period X: $A_1 = 167$. These values are added in fig. 15 and it is seen that they agree very well with the others.

In period XIV from Aug. 11 to Aug. 15 the computed value of A_1 is too small. In this case we may obtain agreement with the other values by introducing an albedo of 77 $^0/_0$ instead of 72 $^0/_0$ as assumed, but it seems improbable that the albedo in any period with a mainly wet surface has an average value as great as 77 $^0/_0$. There exist, on the other hand, reasons why a small value of A_1 may be found in the period in question. On these days the situation was in so far unusual, as the wind was northerly but the temperature was very high. Furthermore, the wind velocity was greatest at a few metres above the ground since at 7 metres it was less than at 2 metres. At 5 metres both temperature and humidity were, on the other hand, much higher than should be expected from the values at 1 meter. On a few occasions we could observe the cloud motion and then the clouds drifted from a southerly direction. Our camp was situated in a location where the plateau fell gently off towards the south and on several clear nights we had observed cold northerly wind, which, no doubt, represented a geostrophic flow from the highest part of the plateau. Considering these circumstances, it seems probable that on Aug. 11 to 15 the air at a low level above the plateau moved

Table 5. *Mean values of wind velocities, potential temperatures and vapour pressures in selected periods.*

Period No.	Length in hours	Wind velocity at altitude (cm)				Potential temperature at altitude (cm)			Pressure of water vapour at altitude (cm)		
		700	200	30	4	500	100	4	500	100	4
		m/sec.	m/sec.	m/sec.	m/sec.	°C	°C	°C	mm	mm	mm
I	72	5.00	4.08	2.96		1.20	1.02	0.74	4.89	4.87	4.80
II	54	4.95	3.91	2.71		2.97	2.31	1.40	5.47	5.26	5.04
III	42	2.61	1.97	1.31		3.74	2.81	1.60	5.72	5.49	5.09
IV	96	1.91	1.59	1.09		2.82	1.85	0.95	5.38	5.05	4.78
V	206	3.88	3.53	2.45		−1.82	−1.80	−1.64	3.64	3.72	4.00
VI........	140	2.35	2.11	1.43	0.97	0.97	0.16	−0.59	3.84	3.91	4.01
VII	48	2.22	2.01	1.29	0.88	−0.33	−0.43	−0.32	4.08	4.12	4.36
VIII......	34	2.13	2.08	1.40		−2.35	−2.24	−1.52	3.66	3.72	4.05
IX........	72	2.59	2.28	1.35		1.57	1.06	0.56	4.97	4.82	4.69
X	36	4.30	3.60	2.40		1.21	1.08	0.75	4.86	4.83	4.71
XI........	34	2.48	2.17	1.58		0.57	−0.16	−0.63	4.20	4.13	4.18
XII.......	26	2.06	1.49	1.05		1.75	1.32	0.62	4.95	4.84	4.79
XIII	36	1.42	1.69	1.18		5.70	3.13	1.56	6.65	5.55	5.04
XIV	120	2.04	1.85	1.29		2.21	1.26	0.56	4.74	4.61	4.57

Table 6. *The values which enter into the computation of A_1, and A_1 in selected periods.*

Period No.	I g cal. cm^{-2} min^{-1}	a	$a\,I$ g cal. cm^{-2} min^{-1}	R g cal. cm^{-2} min.$^{-1}$	$80\,H$ g cal. cm^{-2} min.$^{-1}$	$80\,H-a\,I+R$ g cal. cm^{-2} min.$^{-1}$	Value of n	u_1 cm sec.$^{-1}$	$\Delta\,\theta_1$	$\Delta\,e_1$	$10^4\,A_1$ g cm^{-1} sec.$^{-1}$	Observations used	
												Wind	Temperature and Humidity
I	.225	.30	.068	.030	.066	.028	6.0	168	0.30	0.06	264	All	All
II	.197	.30	.059	.033	.126	.100	6.0	161	1.01	0.28	255	All	All
III	.286	.30	.086	.040	.086	.040	4.6	62	0.85	0.25	90	All	All
IV.........	.219	.30	.066	.047	.043	.024	5.0	55	0.76	0.23	67	200, 30	100, 4
V	.418	.23	.096	.058	.006	−.032	5.2	128	−0.14	−0.25	163	200, 30	100, 4
VI.........	.453	.28	.127	.106	.038	.017	5.4	78	0.81	−0.09	158	200, 30, 4	100, 4
VII........	.400	.26	.104	.041	.056	−.007	5.2	72	0.00	−0.14	80	200, 30	All
VIII.......	.591	.24	.142	.072	.035	−.035	4.8	69	−0.56	−0.26	101	200, 30	100, 4
IX.........	.352	.29	.102	.053	.098	.049	4.6	65	0.40	0.11	242	All	All
X	.272	.30	.082	.032	.085	.035	5.3	128	0.24	0.08	308	All	All
XI.........	.423	.27	.114	.100	.043	.029	6.8	97	0.94	0.00	146	All	All
XII........	.277	.30	.083	.031	.064	.012	4.7	53	0.47	0.07	59	All	All
XIII	.325	.30	.097	.082	.076	.061	5.3	62	1.44	0.47	90	200, 30	100, 4
XIV	.250	.28	.070	.058	.020	.008	5.3	68	0.64	0.04	40	200, 30	100, 4

from the south but owing to the cooling effect of the snow a weak geostrophic flow in the opposite direction developed directly above the surface. Our low value of A_1, means then, that the turbulence was small within this gentle flow and this is probably a correct explanation.

If, in our preceding computation, we omit period XIV and assume an albedo of 64 % in periods IX and X, then we have introduced an average albedo in the periods in question of 70.5 %, whereas, according to Olsson's measurements, we should have used a value of 67 %. The difference is small, but it is, nevertheless, of interest to compute the values of A_1 by means of Olsson's average values for the albedo, retaining our value of R. This has been done, and even in this case A_1 is a nearly linear function of u_1 and equal to $1.3\times10^{-4}u_1$. The scattering of the values is, however, considerably greater and two negative values, which are meaningless, occur. In view of these circumstances and of the fact that the observed albedoes appear somewhat too low since they lead to unreasonably high values of R, we consider the result in fig. 15 as representing the best combination of the different observations.

We can check our result by means of a computation of a somewhat different character. On completely overcast days we found:

$$aI - R = 0.16\,I.$$

Therefore, if our result $A_1 = k u_1$ is correct, we obtain by means of our previous relations:

$$80\,H = 0.16\,I + \delta M t,$$

where t is the time interval in seconds and

$$\delta = \frac{k\,c_p}{n}(a\,b)^{-\frac{1}{n}},\quad M = ((\vartheta_a - \vartheta_0) + 2.27\,(e_a - e_0))\,u_b.$$

Here ϑ_a and e_a represent temperature and vapour pressure at the level $z = a$, ϑ_0 and e_0 temperature and vapour pressure at the level $z = 0$, and u_b is the wind velocity at the level $z = b$. Thus, M can be computed by combining observations at different levels, and on overcast days we can compute k if we know H, I and n. If our equations are correct we must obtain the same value of k regardless of how we combine our observations when computing M and regardless of value of u_1.

We find 10 overcast days from which observations are available which make possible computation of k. The kalues of H and I and the mean temperatures, vapour pressures and wind velocities on these days are found in the tables of results. Table 7 contains the values of $(80\,H - 0.16\,I)$, ϑ_0, e_0, n and Mt as computed by means of different combinations. Further, the table contains the values of k which have been computed by means of the different combinations, the mean values of k for each day and the values of u_1.

It is seen that on each day we find nearly the same value of k, regardless of the combination of observations. On different days the values of k are, on the whole, in good agreement, except that the values on August 1st appear to be too high. It is possible that the albedo was unusually small on this day. In view of the otherwise good agreement, the value from August 1st has been disregarded in the two last columns of the table.

From these last columns it is seen that no relation exists between k and u_1. On five days with u_1 greater than 100 cm/sec. and $\bar{u}_1 = 158$ cm/sec. we obtain $k = 1.57 \times 10^{-4}$, and on four days with u_1 less than 100 cm/sec. and $\bar{u}_1 = 40$ cm/sec we obtain $k = 1.50 \times 10^{-4}$. This examination, therefere, confirms the conclusion that k is proportional to u_1 and that the numerical value of the factor of proportionality is about 1.5×10^{-4}. A preliminary investigation by means of data from the same days gave $k = 1.4 \times 10^{-4}$ (Sverdrup, 1935 a). As the most probable value we adopt the value which was derived by means of the greater part of all observations: $k = 1.45 \times 10^{-4}$. Thus, we finally obtain the simple formula:

$$A = 1.45 \times 10^{-4}\,u_1\,z^{\frac{n-1}{n}}.$$

It must again be emphasized that this formula is an interpolation formula, and that u_1 is a fictive quantity. Further, that under stable conditions, A is not a linear function of the velocity at some distance from the surface since n increases with u_1. This

Table 7. *Values of k on overcast days.*

Date	$80\,H - 0.16\,I$ g. cal. cm^{-2} day^{-1}	ϑ_0 °C	e_0 mm	10^{-6} Mt from obsv. at 4 cm and 30 cm	100 cm and 200 cm	500 cm and 700 cm	Value of n	10^4 k from obsv. at 4 cm and 30 cm	100 cm and 200 cm	500 cm and 700 cm	10^4 k Mean	u_1
July 18th ...	38	0.28	4.76	1.88	4.06	6.51	6.2	1.1	1.2	1.2	1.17	185
» 19th ...	62	0,30	4.76	1.67	3.62	6.22	5.8	2.0	2.2	2.0	2.07	150
» 20th ...	29	0.19	4.67	1.33	3.26	4.46	6 4	1.2	1.1	1.2	1.17	182
Aug. 1st	89	0.15	4.72	1.65	4.06	8.28	4.9	3 0	3.4	2.8		
» 2nd	141	0.52	4.84	5.44	13.06	21.35	5.6	1.4	1.5	1.5	1.47	170
» 3rd	116	0.13	4.68	3.14	9.25	17.00	4.7	2.0	2.0	1.9	1.97	105
» 4th	58	0.28	4.79	2.12	5.69	9.86	5.0	1.5	1 6	1,6	1.57	69
» 5th	68	0.00	4.58	2.81	9.55	16.45	4.7	1.3	1.2	1 2	1.23	49
» 8th	32	0.19	4.58	0.82	3.75	6.05	3.7	2.2	1.9	1.7	1.93	27
» 9th	12	0.05	4.62	0.55	1.51	3.13	4.4	1.2	1.4	1.2	1.27	33

implies that a low velocity (n small) A increases slowly with elevation but u increases rapidly, whereas at high velocity, conditions are reversed. At a given level we have approximately $A \sim u^m$ where m is greater than 1 and increases with increasing distance from the surface.

We have now to compare our empirical result with the results from other investigations and with our theoretical conclusions.

B. Comparison with Results from other Investigations.

A comparison with results from other investigations ist best based upon the formula (48):

$$Q_a = \varkappa (\vartheta_a - \vartheta_0) = \delta\, u_b (\vartheta_a - \vartheta_0)$$

$$\text{where } \delta = \frac{k\, c_p}{n} (a\, b)^{-\frac{1}{n}}.$$

Here we have introduced a "coefficient of exchange" (see p. 21) which is proportional to the wind velocity and a function of the altitudes at which the measurements are taken. Such exchange coefficients have, as pointed out on p. 21, been determined by several authors. The order of magnitude agrees, in general, with our results but the values are in most cases not so detailed that a complete comparison is possible. A detailed examination of $\varkappa$ has been undertaken only by Ångström (1934) in connection with his discussion of the observations which were taken at Sveanor on the Swedish-Norwegian Expedition to North-East Land (Spitsbergen) in 1931. Ångström found, in agreement with our results, that $\varkappa$ was a linear function of u: $\varkappa = 1.12 \times 10^{-6} u$ (u and Q in C. G. S.-units), but this numerical value must be revised since Ångström in his computation disregarded the processes of condensation and evaporation. A new determination of $\varkappa$ has, therefore, been undertaken (Sverdrup 1935 b) and this gave $\varkappa = 1.0 \times 10^{-6} u$. The temperatures were measured at an altitude of 1.9 metres and the wind velocities at an altitude of 3.1 metres. When we introduce these values of a and b in (48) and assume $n = 5.6$ we obtain $k = 1.7 \times 10^{-4}$. This value is probably somewhat too low since, when undertaking the computation, it was assumed that the air temperature at $z = 0$ was equal to zero, whereas, according to our results, it was higher, owing to the existence of the semi-laminar boundary layer. Considering this circumstance, the value $k = 1.9 \times 10^{-4}$ is probably more correct. The agreement with our result from the plateau is satisfactory since k, as will be shown presently, increases with increasing roughness and since the roughness, no doubt, was greater where the observations of 1931 were taken.

Over the pack-ice the author found (Sverdrup 1933) with a clear sky:

$$(58) \qquad Q = 7.3 \times 10^{-7}\, u_7, \quad \theta_{30} - \theta_{4.5} = \frac{176}{u_7}$$

(C. G. S.-units).

Here θ_{30} and $\theta_{4.5}$ represent the potential temperatures at 30 metres and 4.5 metres and u_7 is the wind velocity at 7 metres. The relations appear peculiar but they must be considered as rough approximations, and as such they are not in contradiction to our present results. From the above equations we can compute our coefficient k on the assumption that a similar relation exists between wind velocity and n as on the plateau. We would first state that we have:

$$\theta_{4.5} - \theta_0 = \frac{\theta_{30} - \theta_{4.5}}{3000^{\frac{1}{n}} - 450^{\frac{1}{n}}}\, 450^{\frac{1}{n}}.$$

When we write $Q = \delta\, u_7 (\theta_{4.5} - \theta_0)$, we have $\delta\, (\theta_{4.5} - \theta_6) = 7.3 \times 10^{-7}$. Thus we can compute the value of δ which corresponds to any value of n. We put:

u_7 in cm/sec.	200	400	600	800
and the corresponding n .	4.2	5.3	6.0	6.3
and obtain $(\theta_{4.5} - \theta_0)$, °C. .	1.53	1.05	0.79	0.62
$10^6\, \delta$	0.48	0.69	0.92	1.17
and with $a = 450$ cm.				
$b = 700$, we obtain $10^4 k$.	1.4	1.6	1.9	2.2

Thus k appears to increase with increasing velocity, but this result may be due to the approximate character of our equations. If we had found Q proportional to $u_7^{0.7}$ instead of proportional to u_7 we would have obtained k independent of u. A probable value of k is in this case $k = 1.8 \times 10^{-4}$ with $n = 5.6$. This value, which is nearly the same as the value from Sveanor, is, however, probably too low.

From (58) and (42) follows the approximate relation $A = 0.5\, u_7^2$. Our present results would give $A \sim u_7^{1.7}$ and the discrepancy is, therefore, not so very great. The reason is that, owing to the great influence of stability at low velocities, we obtain very small values of A at some distance from the surface, but at greater velocities the value of A is less influenced by the stability and, therefore, much greater.

It is possible that this influence of the stability explains why Hesselberg (1929), from observations at Lindenberg, found $\eta_h = 1.0\,u^2$, but Köhler's result from Hallde (1932): $A_1 = a + b\,u_1^2$ cannot be explained. The latter result needs confirmation, especially since our result $A_1 \sim u_1$ is in excellent agreement with theoretical conclusions.

8. Comparison between Empirical and Theoretical Results.

A. The Influence of Stability on the Vertical Variation of Wind Velocity. Computation of Rossby's Constant β.

We can now undertake a complete comparison between the theoretical and the empirical results. When deriving our theoretical equations we supposed that the vertical variation of temperature with sufficient accuracy could be represented by means of a power law of the type (26). Our observations have shown that this assumption is correct, and, knowing the vertical variation of wind velocity under stable conditions, we can, therefore, compute Rossby's numerical constant β (see p. 8) which determines the influence of stability. If different computations give similar values, we must conclude that the introduction of β is correct and that β has a definite physical significance. When β has been determined we can compute the eddy convectivity under stable conditions η_s. A comparison between the theoretical values of η_s and our empirial values of A makes possible a crucial test of the theory.

Before computing β we must determine the roughness parameter z_0 as exactly as possible. From the observations at indifferent or instable equilibrium we found $z_0 = 0.23$ cm, but this value is somewhat uncertain since the data from which it was derived show too wide a dispersion (see fig. 7 p. 28). We can now avail ourselves of the circumstance that the logarithmic law is very nearly valid up to 7 metres if the velocity is high and the stability is relatively small, since we have seen (fig. 11 p. 30) that the values from Groups IV and X can be represented by means of logarithmic laws. For a computation of z_0 we use the values from the two lowest levels only, since the logarithmic law gives best approximation in the lowest metres. The mean velocities at 2 m and 0.3 m are 5.185 and 3.695. Introducing the preliminary value $z_0 = 0.2$, we obtain:

$$\frac{\log 200.2 - \log 30.2}{\log 30.2 - \log z_0} = \frac{5.185 - 1.695}{3.695} = 0.404,$$

which gives $z_0 = 0.28$ cm. If we use the observations from all three levels we obtain $z_0 = 0.24$ cm. As a probable mean value we adopt:

$$z_0 = 0.25 \text{ cm.}$$

In order to compute β we undertake a transformation of equation (29) p. 10 and write:

$$f(v) = \left[2(v-1) - \ln\frac{v(v+1)}{2}\right] = \frac{n+1}{n}\frac{1}{2}\ln\frac{z_1+z_2}{z_0}\left[\frac{u_s}{u_a} - \frac{\ln\dfrac{z+z_0}{z_0}}{\ln\dfrac{z_1+z_0}{z_0}}\right], \tag{59}$$

where u_a is the velocity which would be found at the level z_1 under adiabatic conditions and u_s is the velocity at the level z under stable conditions. The relation between v, β and the temperature variation is given by equation (28):

$$v^2(v^2-1) = \beta\frac{\dfrac{g}{T}\dfrac{d\theta}{dz}}{\left(\dfrac{du_a}{dz}\right)^2} = \beta\frac{g}{T}\frac{\triangle\theta_1}{n}\left(\frac{\ln\dfrac{z_1+z_0}{z_0}}{u_a}\right)^2 (z+z_0)^{\frac{n+1}{n}}. \tag{60}$$

As a preliminary value Rossby and Montgomery found $\beta = 40$, but the validity of their equations is doubtful and the observations which were at their disposal were not well suited for the purpose.

In order to apply our formulae we must know u_a at one level. It was shown that with a moderate or a strong velocity the logarithmic law is valid up to a limited height even in case of stability. We may in first approximation assume that the logarithmic law is sufficiently correct up to an altitude of 30 cm if the velocity at this level is greater than 2 m/sec. In our preceding equation, therefore, we put $z_1 = 30$ cm and $u_a = u_{30}$. Furthermore, we introduce $z_0 = 0.25$ cm. On this assumption we compute v by means of (59) from the wind observations at 7 m and 2 m, using the values of n which were derived from these wind observations, and by means of v we compute β from (60) using the temperature observations at 5 m and 1 m. From the data in groups VIII and IX we obtain the mean value $\beta = 10.5$. With this value of β we compute by means of (29) p. 10 the differences $u_s - u_a$ at the level 30 cm. On the right hand side of (29) we replace, as a first approximation the frictional strees by the observed velocity at 30 cm (11, p. 7). We obtain:

Group:	VII	VIII	IX
$(u_s = u_a)$ in cm/sec.:	6	2	1
u_a in cm/sec.:	92	201	283

The last line contains the corrected velocities which would have been observed at 30 cm under adiabatic conditions. The corrections, $(u_s - u_a)$, to the observed velocities are so small that it is unnecessary to introduce the corrected velocities on the right hand side of (29) and undertake a second approximation.

With the above values of u_a we can now undertake the final computation of β by means of (59) and (60). We have not used the data in Group X since in this group the deviation from the logarithmic law is so small that results are uncertain. The results from the other groups are:

Group:	VII	VIII	IX
β from velocities at 7 m	10.8	10.9	10.8
β » » » 2 »	11.8	9.5	13.3

The values are in excellent agreement and we are justified in concluding that β has a physical significance. The mean value of β is 11.2, but giving the observation at 7 m a somewhat greater weight, we adopt

$$\beta = 11.0.$$

As a check we compute by means of (29) the velocities at 7 m and 2 m in the Groups II, III, V and VI, introducing $\beta = 11.0$. The factor $\frac{1}{k_0}\sqrt{\frac{\tau}{\varrho}}$ is determined by means of the observed velocity at 30 cm, using the method of approach which was introduced when computing u_a at this level. We obtain:

Group:		II	III	V	VI
Velocity at 7 m in m/sec.	Computed	2.15	2.03	5.49	5.77
	Observed	2.26	1.97	5.49	5.80
Velocity at 2 m in m/sec.	Computed	1.62	1.45	4.53	4.63
	Observed	1.62	1.46	4.54	4.65

The agreement between the computed and observed velocities is very good, especially at high velocity. This means that by using our theoretical equation we can compute the vertical distribution of velocity if we know the velocity at one level near the ground, the roughness parameter, and the variation of temperature with height. In this computation two numerical constants enter, $k_0 = 0.38$ and $\beta = 11.0$.

Our value of β is much smaller than the preliminary value $\beta = 40$ which Rossby and Montgomery found. Our result is derived on a more correst basis and by means of observations which allow a more accurate computation.

The significance of the exponent n in the power law can now be explained more clearly. It is quite evident that the value of n which is derived from observations within a given interval of altitude depends upen the roughness parameter and upon the stability, as pointed out by Prandtl (1932). At a given locality where the roughness parameter, z_0, is known, n depends principally upon the stability but since the effect of a given stability is much greater at low wind velocity we must find an apparent relation between n and the wind velocity; with the same value of the temperature gradient, n must increase with increasing velocity. This feature, perhaps, explains the fact that our observations, on an average, show a well defined increase of n with increasing velocity, as during our observations the stratification was, on an average, stable.

B. The Relation between Eddy Conductivity and Eddy Convectivity.

Before we undertake a comparison between the empirical value of A_s and the theoretical value of η_s we shall again examine the relation between these coefficients. Introducing power laws, Taylor's criterion (47) p. 18 can be written:

$$\frac{A_s}{\eta_s} \leq \frac{T}{g} \frac{u_1^2}{n \triangle \theta_1} Z^{\frac{1-n}{n}}$$

where u_1 and $\triangle \theta_1$ have the same meaning as previously. The maximum altitude up to which A_s and η_s can be identical is, therefore, obtained from the equation:

$$Z^{\frac{n-1}{n}} = \frac{T}{g} \frac{u_1^2}{n \triangle \theta_1}.$$

In the groups which have been dealt with in the preceding chapter we have:

Group:	II	III	V	VI
n	4.2	4.2	5.9	5.4
u_1 cm/sec.	48	41	182	173
$\triangle\theta_1$	0.35	0.99	0.36	0.83
Z in metres	28	5	234	100

Group:	VII	VIII	IX
n	4.2	5.2	5.6
u_1 cm/sec.	44	105	156
$\triangle\theta_1$	0.52	0.61	0.46
Z in metres	12	48	121

We conclude that, looking away from conditions in the semi-laminar boundary layer, A_s and η_s are identical below 7 metres except in Group III (small velocity and very great stability) where A_s probably is smaller than η_s above the 5 metres level. If our theoretical results are correct the empirical values of A_s must, therefore, agree with the theoretical values of η_s.

C. Comparison between the Empirical and the Theoretical Values of the Eddy Conductivity.

We shall first consider our empirical result

$$A = k\,u_1\,z^{\frac{n}{n-1}}. \tag{61}$$

Here k is not a dimensionless constant and u_1 has not the dimensions of a velocity. In our original equation $u_z = u_1\,z_1^{-\frac{1}{n}}\,z^{\frac{1}{n}}$ we put $z_1 = 1$ (p. 39) and wrote $u_z = u_1\,z^{\frac{1}{n}}$ but this means that u_1 has the dimensions cm$^{\frac{n-1}{n}}$ sec.$^{-1}$. Since A has the dimensions g cm^{-1}sec.$^{-1}$ it follows that k has the dimensions g cm^{-3}cm$^{\frac{2}{n}}$. Thus, k is proportional to the density ϱ and to $h^{\frac{2}{n}}$ where h is a characteristic length. The factor of proportionality is a pure number which we may write Cn:

$$A = C\,n\,\varrho\,h^{\frac{2}{n}}\,u_1\,z^{\frac{n-1}{n}} = C\,n\left(\frac{h}{z}\right)^{\frac{2}{n}} u_z\,z\,. \tag{62}$$

This is exactly the form which Prandtl in 1924 suggested for the eddy convectivity over a rough surface (see p. 7). According to Prandtl C is equal to 0.0089. If this value has the character of a universal constant and if our coefficient k is really independent of u, we must obtain the same value of h, regardless of the value of n. With $C = 0.0089$ and $\varrho = 1.17\times10^{-3}$ (temperature 0° and pressure 690 mm) and with $k = 1.45\times10^{-4}$ we obtain:

n	4.0	4.5	5.0	5.5	6.0	6.5
h, cm	12.1	12.6	12.8	12.8	12.4	11.8

The variation of these values of h is insignificant. On an average we obtain $h = 12.5$ cm and since we found $z_0 = 0.25$ cm, we obtain $z_0 = {}^1/_{50}\,h$ whereas Prandtl as a preliminary value introduces $z_0 = {}^1/_{30}\,h$. An explanation of the discrepancy between our result and Prandtl's cannot be offered, but it must be pointed out that the average height of the undulations of the snow surface was a few cm only. The value $h = 30\,z_0 = 7.5$ cm is, no doubt, too high and $h = 12.5$ cm is far too high.

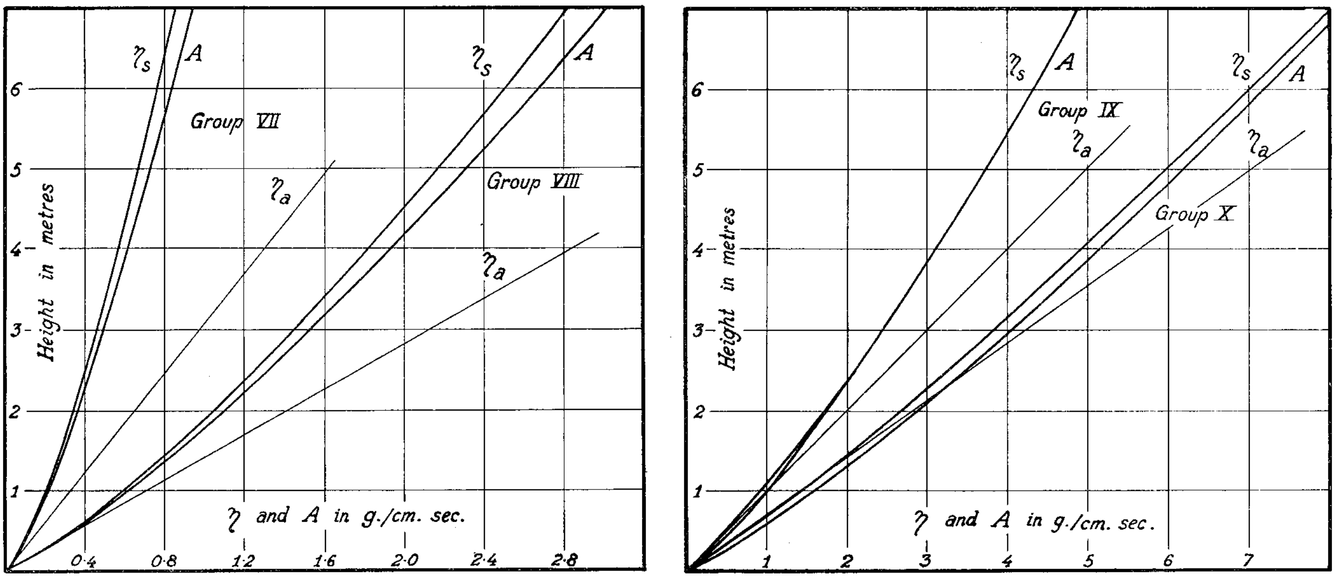
At Sveanor we found $k = 1.9\times10^{-4}$. With $n = 5.6$ and $\varrho = 1.29\times10^{-3}$ we obtain $h = 21$ cm. It is very probable that the roughness parameter had a somewhat greater value at Sveanor since there the surroundings were free from snow and were hilly.

From the observations in the pack-ice we found k about 1.8×10^{-4} and with $n = 5.6$ and $\varrho = 1.35\times10^{-3}$ we obtain $h = 17$. This value is probably too low, since the roughness of the pack-ice is much greater than the roughness of a smooth snow field. The agreement is in this case not very good, but it must be considered that the value of k from the pack-ice is very uncertain.

From our preceding considerations it is, as already mentioned, evident that formula (61) has the character of an interpolation formula and is not valid to the boundary surface, but is applicable only within the interval of altitude from which the meteorological observations are obtained. Furthermore, it has been emphasized that u_1 is a fictive velocity which is found by an extrapolation and does not represent the actual velocity at 1 cm. The formula is satisfactory for all practical purposes but from the theoretical standpoint the formula is deficient since it is not valid at the surface, and especially since the influence of stability does not appear in an explicit manner.

Our theoretical expression for the eddy convectivity in case of stability (equation (35) p. 10) is satisfactory:

$$\eta_s = \frac{\varrho\,k_0^{\,2}\,(z+z_0)}{\sqrt{\frac{1}{2}+\frac{1}{2}\sqrt{1+4\,\psi}}}\;\frac{u_a}{\ln\frac{z_1+z_0}{z_0}} \tag{63}$$



16 a 16 b

Fig. 16 a og b. Computed values of the eddy convectivity under stable conditions, η_s, and empirical values of the eddy conductivity A_s. The straight lines show the corresponding eddy convectivity at adiabatic conditions.

where:

$$(64) \qquad \psi = \beta \frac{g}{T} \frac{\triangle \theta_1}{n} \left(\frac{\ln \frac{z_1 + z_0}{z_0}}{u_a} \right)^2 (z + z_0)^{\frac{n+1}{n}}.$$

We observe that ψ is dimensionless since $\triangle \theta_1$, which was defined by $\theta_z - \theta_0 = \triangle \theta_1 z^{\frac{1}{n}}$ has the dimension $°C\, cm^{-\frac{1}{n}}$. Further, that u_a represents the velocity, which would be found at the level z_1 under adiabatic conditions.

If our theoretical conclusions are correct and if especially eddy convectivity and eddy conductivity are identical in the lowest layer, we must now obtain $A_s = \eta_s$, and a comparison of the results of these two computations makes possible a final test of the theory.

We have computed A_s and η_s by means of the data in Groups VII, VIII, IX and X. Within these groups we have, putting $z_1 = 30$ cm.

Group No.	VII	VIII	IX	X
Value of n	4.2	5.2	5.6	6.2
$\lvert u_1 \rvert$ in cm/sec.	44	105	156	232
u_a at 30 cm in cm/sec.	92	201	283	400
$\triangle \theta_1$	0.52	0.61	0.46	0.44

When computing A we use the constant $k = 1.45 \times 10^{-4}$ and when computing η_s we introduce $\varrho = 1.17 \times 10^{-3}$, $z_0 = 0.25$ cm and use the numerical constants $k_0 = 0.38$ and $\beta = 11.0$. We note that k, z_0, k_0 and β are not known with greater accuracy than 5 to 10 %.

The corresponding values of η_s and A_s are given in table 8 and represented graphically in fig. 16. In the figure, straight lines are entered showing the vertical variation of η_a, the corresponding eddy convectivity at adiabatic conditions. The agreement between A_s and η_s is excellent since the differences in all cases lie within the limits which are determined by the possible errors of the factors which enter into the computation. The remarkable similarity of the

Table 8. *Theoretical values of the eddy convectivity under stable conditions, η_s, and the corresponding empirical values of the eddy conductivity, A_s, both in $g.cm.^{-1}sec.^{-1}$.*

Height in cm	Group							
	VII		VIII		IX		X	
	η_s	A_s	η_s	A_s	η_s	A_s	η_s	A_s
100	0.20	0.21	0.59	0.63	0 92	0.99	1.36	1.60
200	0.34	0.36	1.05	1.10	1.71	1.75	2.62	2.86
300	0.46	0.49	1.45	1.53	2.43	2.45	3.80	4.02
400	0.57	0.61	1.82	1.92	3.10	3.12	4.91	5.12
500	0.67	0.73	2.17	2.31	3.72	3.74	6.00	6.19
600	0.76	0.84	2.50	2.67	4.32	4.33	7.01	7.19
700	0.85	0·94	2.82	3.02	4.89	4.91	8.01	8.18

empirical values of A_s with the theoretical values of η_s, therefore, demonstrates that our observations lead to results which are in the best possible agreement with those from the laboratories and the theoretical considerations based on them.

In view of this fact it seems at present unnecessary to enter upon results as to the relation between velocity and eddy convectivity, which apparently contradict our conclusions, since these results are derived from observations which are far less complete than those which have been discussed here.

We can now answer the four questions which were put in the introduction (p. 6); but must make the reservation that the answers apply to conditions over a relatively smooth surface, say a snow surface, open grass land or the surface of the sea.

Answer to question 1: Eddy convectivity and eddy conductivity are identical near the surface, except within the semi-laminar boundary layer.

Answer to question 2: Near the surface, but above the semi-laminar boundary layer, the temperature distribution depends only upon the processes of conduction (the turbulence) and is independent of processes of radiation. This is perhaps correct only when the vapour pressure is small, say less than 10 mm.

Answer to question 3: The relation between the eddy convectivity and the roughness of the surface can be expressed by means of a parameter which has the dimension of a length and characterizes the roughness of the surface. The relation between the eddy convectivity and the stability depends upon a dimensionless ratio between certain functions which characterize the stability and the state of motion and upon a numerical constant, the value of which has been determined.

Answer to question 4. In case of instability or indifferent equilibrium the eddy convectivity increases linearly with increasing distance from the surface. Under stable conditions the eddy convectivity very near the surface has the same value as in the case of indifferent equilibrium, other conditions being alike, but it increases more slowly with elevation. At some distance from the surface it is, therefore, smaller than in the case of indifferent equilibrium. Approximately, it can be represented by a power law.

To these answers we add: At stable stratification the eddy conductivity above a relatively smooth surface can be computed by means of a theoretical formula (63) if the roughness parameter of the surface is known, and if the vertical variations of wind velocity and temperature are know with such great accuracy that these variations can be represented by power laws. Two numerical constants enter into this computation, one, which determines the length of mixing, and was introduced by v. Kármán who determined the value by means of results from laboratory experiments, and one, which determines the influence of stability, introduced by Rossby and Montgomery, who determined a preliminary value by means of meteorological observations. A more accurate value of the latter constant has been determined by means of our observations. Thus, complete analogy exists between the character of the turbulence in the vicinity of a relatively smooth surface in nature, and the character of the turbulence near a surface according to laboratory observations, in spite of the widely different scale of turbulence in the two cases.

LITERATURE

1935. Ahlmann, H. W:son: Ablation Measurements at the Headquarters on Isachsen's Plateau. *Geogr. Annaler* **17.** Stockholm 1935.

1935. Best, A. C.: Transfer of Heat and Momentum in the Lowest Layers of the Atmosphere. *Geophysical Memoirs No. 65* (**7**, *No. 8*). London 1935.

1933. Böhnecke, G.: Ein Assmannsches Aspirationspsychrometer mit Umkippthermometern. *Ann d. Hydr. u. mar. Met.* **61.** Berlin 1933.

1934. Brunt, D.: Physical and Dynamical Meteorology. Cambridge 1934.

1934. Büttner, K.: Die Wärmeübertragung durch Leitung und Konvektion usw. *Veröff. des Pr. Met. Inst. Abhandlungen*, **10,** *Nr. 5.* Berlin 1934.

1932. Devik, O.: Thermische und dynamische Bedingungen der Eisbildung in Wasserläufen. *Geof. Publ.* **9.** *Oslo 1932.*

1929. Éliás, F.: Die Wärmeübertragung einer geheizten Platte an strömende Luft. *Zeitschr. für angew. Math. u. Mech.* **9** und **10.** Berlin 1929 und 1930.

1930. Ertel, H.: Eine Methode zur Berechnung des Austauschkoeffizienten aus den Feinregistrierungen der turbulenten Schwankungen. *Gerlands Beitr. zur Geoph.* **25.** Leipzig 1930.

1933. Fjeldstad, J. E.: Wärmeleitung im Meere. *Geof. Publ.* **10.** Oslo 1933.

1927. Geiger, R.: Das Klima der bodennahen Luftschicht, Braunschweig 1927.

1931. Handbuch der Experimentalphysik, **4,** 1. Teil. Leipzig 1931.

1932. Handbuch der Experimentalphysik, **4,** 2. Teil. Leipzig 1932.

1929. Hesselberg, Th.: Ein neuer Ausdruck für den Austauschkoeffizienten. *Ann. d. Hydr. u. mar. Met.* **57.** Berlin 1929.

1929. Johnson, N. K.: A Study of the Vertical Gradient of Temperature in the Atmosphere near the Ground. *Geophysical Memoirs No. 46.* (**5**, *No. 6*). London 1929.

1929. Köhler, H.: Über den Austausch zwischen Unterlage und Luft. *Gerlands Beitr. zur Geoph.* **24.** Leipzig 1929.

1932. Köhler, H.: Ein kurzes Studium des Austausches auf Grund des Potenzgesetzes. *Beitr. zur Physik d. fr. Atm.* **49.** Leipzig 1932.

1933. Köhler, H.: Meteorologische Turbulenzuntersuchungen 1, *Kungl. Svenska Vetenskapsakademiens Handl. Tredje Serien,* **13,** No. 1. Stockholm 1933.

1935. Köhler, H.: Meteorologische Turbulenzuntersuchungen 1 B, *Lantbrukshögskolans Annaler,* **2.** Uppsala 1935.

1932. Krügler, Fr.: Über den Anteil des Massenaustausches am nächtlichen Wärmehaushalt der Erdoberfläche. *Met. Zeitschr.* **49.** Braunschweig 1932.

1934. Lettau, H.: Turbulente Schwankungen von Wind und Temperature in der bodennahen Luftschicht als Austauschproblem. *Ann. d. Hydr. u. mar. Met.* **62.** Berlin 1934.

1932. Marquardt, R.: Untersuchungen des Wärme- und Wasserdampfsaustausches über dem Bodensee. *Gerlands Beitr. zur Geoph.* **36.** Leipzig 1932.

1932. Mildner, P.: Über die Reibung einer speziellen Luftmasse in der untersten Schichte der Atmosphäre. *Beitr. z. Physik d. fr. Atm.* **19.** Leipzig 1932.

1935. Nomitsu, T.: On the Wind in the Lower Atmosphere, I. *Japanese Journal of Astronomy and Geophysics,* **12.** No. 2. Tokio 1935.

1924. Prandtl, L. und Tollmien, W.: Die Windverteilung über dem Erboden, errechnet aus den Gesetzen der Rohrströmungen. *Zeitschr. für Geophysik* **1.** Braunschweig 1924.

1932. Prandtl, L.: Meteorologische Anwendungen der Strömungslehre. *Beitr. zur Physik d. fr. Atm.* **19.** Leipzig 1932.

1935. Rossby, C. G. and Montgomery, R. B.: The Layer of Frictional Influence in Wind and Ocean Currents. *Papers in Physical Oceanography and Meteorology published by Massachusetts Inst. of Techn. and Woods Hole Oceanogr. Inst.* **3.** *No. 3.* Cambridge Mass. 1935.

1929. Schmidt, W.: Die Struktur des Windes. *Sitzungsber. d. Akad. d. Wiss. in Wien, Mathem.-naturw. Abteilung II a,* **138.** *Heft. 3 und 4.* Wien 1929.

1930. Scrase, F. J.: Some Characteristics of Eddy Motion in the Atmosphere. *Geophysical Memoirs No. 52.* (**6.** *No. 2*). London 1930.

1934. Sutton, O. G.: Wind Structure and Evaporation in a Turbulent Atmosphere. *Proceedings of the Royal Society of London, Series A,* **146.** London 1934.

1933. Sverdrup, H. U.: Meteorology, Part I, Discussion. *The Norw. North Polar Exp. with the "Maud", 1918—1925. Scientific Results,* **2.** Bergen 1933.

1935 a. Sverdrup, H. U.: Varmeutvekslingen mellem en sneflate og luften. (Mit Zusammenfassung in Deutsch). *Ber. fra Chr. Michelsens Inst.* **5.** Bergen 1935.

1935 b. Sverdrup, H. U.: The Ablation on Isachsen's Plateau etc. *Geogr. Annaler.* **17.** Stockholm 1935.

1915. Taylor, G. I.: Eddy Motion in the Atmosphere. *Phil. Trans. Roy. Soc. A,* **215.** London 1915.

1916. Taylor, G. I.: Skin Friction of the Wind on the Earth's Surface. *Proceedings Royal Soc. of London, Series A* **92.** London 1916.

1931. Taylor, G. I.: Internal Waves and Turbulence in a Fluid of Variable Density. *Rapports et Procés-Verbaux Conseil Perm. Intern. pour l'Exploration de la Mer,* **76.** Copenhagen 1931.

1920. Wüst, G.: Die Verdunstung auf dem Meere. *Veröff. des Inst. für Meereskunde, Reihe A, Heft 6.* Berlin 1920.

1918. Ångström, A.: Radiation and Temperature of Snow etc. *Arkiv för Math., Astr, och Fysik,* **13.** *No. 21.* Stockholm 1918.

1934. Ångström, A.: On the Dependence of Ablation on Air Temperature, Radiation and Wind. *Geogr. Annaler,* **15.** Stockholm 1934.

RESULTS OF OBSERVATIONS

Table I

contains the true direction of the wind and the mean hourly values of the wind velocity in metres per second at three or four levels. The mean values are centred on the full hours. Velocities which are uncertain, especially owing to formation of frost on the anemometer are enclosed in brackets. The symbol $<$ indicates that the velocity was too weak to be recorded by the anemometer. On days with incomplete observations the mean diurnal values are placed in brackets.

Tables II and III

contain mean hourly values of temperatures and vapour pressures at three levels, centred on the full hours. Values which have been derived by interpolation are enclosed in brackets, if the interval concerned is longer than 2 hours.

Table IV

contains the cloudiness on scale 0 to 10 and the hydrometeors at the hour of observation. The following symbols are used:

• Rain	Frozen drizzle	= Mist (thin, wet fog)
Snow	Rain showers	Dew
Sleet	Snow showers	Hour frost
Drizzle	≡ Fog	Glaced frost

Table V

contains the precipitation in mm water between 7^h and 19^h and between 19^h and 7^h.

Table VI

contains the diurnal values of tne ablation of the snow in cm water.

Table VII

contains the diurnal values of the total radiation income in gramme calories per square centimetre.

Table I. *Wind. Mean Hourly Values of Wind Direction, and of Velocity in Metres per Second.*

Date	Height cm	Hour 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
June	Direction	E	E	N	N	NE	E	E	ESE	SE	SE	SSE	SE	S	ENE	ENE	ENE	NE	NE	NNE	N	N	N	N	N	
26	700	3.0	2.5	1.4	1.1	1.5	2.5	2.6	2.6	3.2	2.0	1.1	1.3	1.2	1.3	1.1	1.2	1.0	1.4	2.7	3.1	3.3	3.9	3.9	3.9	2.20
	200	2.1	1.7	1.0	0.7	1.3	1.9	1.9	2.4	2.8	1.5	0.8	1.2	0.9	1.1	0.8	1.0	0.9	1.2	2.2	2.2	2.6	3.0	3.2	3.1	1.73
	30	1.1	1.1	0.5	0.4	0.5	0.8	0.8	1.4	1.8	1.0	0.4	0.4	0.4	0.6	0.4	0.5	0.4	0.6	1.4	1.4	1.6	1.4	1.9	1.7	0.96
27	Direction	N	N	N	N	N	N	N	N	N	N	N	N	N	N	N	NNW	NNW	N	NNW	N	N	N	N	E	
	700	3.6	3.3	3.3	3.4	3.3	3.3	3.3	3.0	3.2	3.3	3.0	3.1	3.3	3.4	3.6	4.0	4.5	4.4	4.4	4.4	4.5	4.6	4.3	4.0	3.69
	200	2.8	2.4	2.7	2.6	2.5	2.6	2.5	2.4	2.5	2.6	2.4	2.5	2.7	2.9	3.1	3.5	4.0	3.7	3.6	3.7	3.6	3.5	3.2	2.9	2.95
	30	1.6	1.5	1.4	1.5	1.6	1.6	1.4	1.3	1.6	1.6	1.4	1.5	1.7	1.7	2.0	2.3	2.5	2.6	2.6	2.6	2.6	2.4	2.2	2.0	1.88
28	Direction	N	N	N	N	NNE	NNW	N	N	N	N	NNE	N	N	NNW	NNW	NNW	N	NNW	NNE	N	NNW	NNW	N	NNW	
	700	2.8	3.9	3.9	4.4	4.9	5.2	5.7	4.9	4.2	4.3	3.5	4.2	4.2	4.0	4.6	4.6	4.2	3.9	3.6	3.8	3.5	3.7	3.8	3.0	4.12
	200	2.2	3.1	3.5	3.8	4.2	4.5	5.0	4.2	3.6	4.0	3.1	3.7	3.7	3.4	3.8	4.1	4.0	3.5	3.3	3.5	3.0	3.1	3.3	2.6	3.59
	30	1.4	2.4	2.6	2.8	3.1	3.4	3.8	3.1	2.6	3 0	2.4	2.7	2.9	2.4	2.8	3.1	3.1	2.6	2.4	2 6	2.1	2.2	2.4	1.8	2.65
29	Direction	N	N	N	N	N	N	N	N	N	NNE	NNE	NNE	NNE	NNE	NNE	NNE	NNE	NNE	NNE	NNE	NNE	NNE	NE	ENE	
	700	3.0	2.8	2.6	2.6	2.2	2.0	1.4	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	1.3	1.2	(1.30)
	200	2.3	2.2	1.9	1.9	1.6	1.9	1.0	0.9	1.0	0.8	0.7	0.7	0.7	0.7	<0.6	<0.6	0.6	0.8	0.9	1.1	1.4	1.3	1.1	1.1	(1.14)
	30	1.4	1.3	1.1	1.1	1.0	1.1	0.5	0.5	0.6	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	0.5	(0.55)
30	Direction	ENE	NNE	NNW	NW	NNW	NNW	NNW	N	N	N	N	N	N	N	N	N	NNW	N	N	NNW	NNW	N	NNW	N	
	700	1.3	1.3	1.2	<1.2	<1.2	<1.2	1.8	3.0	2.8	2.4	2.6	2.6	2.6	2.8	1.7	0.8	1.6	2.9	2.7	2.7	1.5	1.3	2.2	2.1	(1.93)
	200	1.5	1.5	0.9	0.6	1.0	1.0	1.7	2.8	2.4	2.0	2.0	2.2	2.3	2.4	1.4	0.5	1.6	2.7	2.5	2.4	1.2	0.9	2.0	2.0	1.73
	30	0.5	0.7	0.5	0.4	0.7	0.5	1.1	2.0	1.7	1.4	1.3	1.5	1.6	1.6	0.9	0.4	1.0	1.9	2.0	1.7	0.9	0.5	1.5	1.3	1.15
	4	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.5	1.0	1.1	1.2	0.8	0.5	1.1	1.0	-
July	Direction	N	NNW	N	N	NNW	NNE	N	N	N	N	N	NNW	N	N	N	NW	N	N	NNW	NNW	NNW	N	N	N	
1	700	1.9	1.4	1.1	1.3	1.4	1.4	1.8	2.4	2.4	2.1	1.8	2.0	1.9	1.8	1.4	2.8	2.4	2.2	2.2	1.8	2.4	2.2	1.9	1.7	1.90
	200	1.7	1.2	0.9	1.0	1.3	1.6	1.9	2.0	2.1	1.9	1.7	1.7	1.6	1.7	1.2	2.5	2.2	2.1	1.9	1.8	2.2	2.2	1.9	1.6	1.75
	30	1.0	0.8	0.6	0.5	0.7	1.0	1.2	1.4	1.5	1.2	1.1	1.2	0.9	1.2	0.8	1.7	1.5	1.4	1.2	1.1	1.4	1.4	1.1	1.0	1.12
	4	0.8	0.5	0.4	0.4	0.6	0.6	0.6	0.8	0.9	0.8	0.8	1.0	0.8	0.9	0.5	0.9	0.9	0.9	1.0	1.0	1.1	1.2	1.0	0.8	0.80
2	Direction	N	N	N	N	NNW	NNW	NNW	SW	SW	WSW	WNW	WNW	WNW	WSW	SW	SW	SW	W	W	W	WNW	WNW	W	W	
	700	1.2	1.1	1.3	1.2	1.1	1.1	1.2	1.7	2.6	3.4	3.7	3.0	2.2	1.6	1.4	2.3	3.9	4.4	4.0	4.0	3.7	3.8	3.7	3.5	2.55
	200	1.1	0.7	1.0	0.9	0.9	1.0	0.9	1.4	2.2	3.2	3.4	2.5	1.9	1.5	1.3	2.2	3.4	4.1	3.9	3.7	3.4	3.5	3.5	3.2	2.28
	30	0.6	0.4	0.6	0.6	0.5	0.5	0.5	0.9	1.5	2.1	2.3	1.6	1.2	0.9	0.6	1.3	2.2	2.9	2.5	2.4	2.3	2.2	2.3	2.1	1.46
	4	0.5	0.3	0.6	0.5	0.2	0.2	0.3	0.4	0.7	1.1	1.5	1.3	0.9	0.4	0.4	1.0	1.6	1.9	1.8	1.7	1.6	1.5	1.5	1.5	0.97
3	Direction	WSW	W	SW	SW	SSW	WSW	W	WSW	WSW	SW	SW	S	SSW	SSW	S	SSW	SSW	S	S	S	S	S	S	S	
	700	3.1	2.7	2.5	4.1	3.0	3.6	4.1	4.6	4.7	4.0	4.5	4.4	5.3	5.8	6.4	6.4	6.7	6.6	6.1	6.1	6.5	6.2	5.5	5.2	4.92
	200	2.9	2.6	1.9	3.3	2.6	3.2	3.8	4.0	4.4	3.7	4.1	3.8	4.5	4.9	5.6	5.5	6.0	5.7	5.2	5.3	5.5	5.0	4.4	4.3	4.26
	30	1.8	1.4	1.0	2.1	1.6	2.1	2.4	2.4	2.7	2.4	2.7	2.4	3.0	3.3	3.7	3.5	3.6	3.4	3.4	3.6	3.6	3.1	2.7	2.9	2.70
	4	1.5	0.9	0.8	-	-	-	-	-	-	-	-	-	-	-	-	1.9	2.3	2.5	2.4	2.3	2.2	1.9	1.6	1.8	-
4	Direction	SSE	S	S	S	S	S	S	S	S	S	SSW	SSW	SSW	SSW	SSW	SSW	S	SSE	SSE	S	S	SSW	SSE	S	
	700	5.5	6.0	6.4	6.2	5.6	6.8	7.1	6.8	6.6	6.5	5.8	6.0	5.7	5.2	5.3	5.8	6.1	5.6	4.8	5.0	5.2	4.6	4.4	(3.8)	5.70
	200	4.6	5.0	5.2	4.7	4.8	6.2	6.4	6.2	6.0	5.8	5.3	5.8	5.4	4.8	5.1	5.5	5.8	5.2	4.6	4.8	4.8	4.4	4.2	3.8	5.18
	30	3.2	3.5	3.6	3.3	3.2	4.0	4.2	3.9	3.6	3.4	3.4	3.6	3.2	2.9	3.0	3.2	3.2	3.0	2.4	2.5	2.6	2.6	2.7	2.5	3.20
	4	-	-	-	-	-	-	-	-	2.5	2.3	2.2	2.3	2.2	2.2	2.2	2.2	2.3	2.4	1.9	1.8	2.0	-	-	-	-

Table I. (Continued.)

Date	Height cm	Hour 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
July 5	Direction	SSE	S	S	S	S	S	S	S	S	S	S	S	S	S	SSW	SW	SW	WSW	SW	SW	SW	W	N	N	
	700	(3.4)	(3.3)	(3 1)	(2.8)	(3.0)	(3.1)	(2.7)	(2.5)	(2.6)	(2.5)	(2.0)	(1.6)	(1.1)	(1.0)	1.5	2.8	1.6	1.3	1.1	1.4	1.4	1.1	1.1	1.2	(2.05)
	200	3.4	3.2	3.3	2.9	3.0	3.1	2.7	2.4	2.6	2.3	1.9	1.4	0.7	0.8	1.2	2.6	1.5	1.3	1.2	1.2	1.0	1.0	1.1	1.3	1.96
	30	2.3	2.1	2.1	1.9	1.7	1.7	1.8	1.6	1.7	1.6	1.2	1.0	0.4	0.5	0.8	1.7	0.8	0.9	0.7	0.8	0.6	0.5	0.5	0.4	1.22
	4	-	-	-	-	-	-	-	-	-	-	-	-	0.3	0.4	0.5	1.0	0.4	0.4	0.5	0.5	0.3	0.2	0.3	0.5	
6	Direction	NW	S	S	S	SSE	S	S	S	S	S	S	S	S	SSE	SSE	SSE	SSE	SSE	SSE	SSE	SSE	SE	SSE	S	
	700	1.2	1.2	(2.5)	(3.9)	(5.3)	(5.8)	(5.5)	(5.8)	6.5	6.9	7.0	7.0	7.0	7.4	8.7	1.8	6.7	6.0	6.3	5.6	4.3	4.4	4.6	3.5	(5.47)
	200	0.8	1.1	2.6	4.0	5.3	5.5	4.9	5.3	5.8	6.4	6.6	6.6	6.5	6.9	8.1	7.4	6.3	5.7	5.7	4.8	4.0	4.3	4.3	3.0	5.08
	30	0.4	0.8	1.7	2.6	3.5	3.7	3.0	3.4	3.7	4.0	4.3	4.3	4.3	4.7	5.4	5.0	4.1	3.8	3.8	3.3	2.6	2.9	2.9	1.9	3.34
	4	0.5	0.4	0.9	-	-	-	-	-	-	-	2.1	2.3	2.6	3.3	3.9	3.2	2.5	2.5	2.6	2.4	2.1	2.1	1.9	1.3	
7	Direction	SSE	SSE	SSE	SSW	SSW	S	S	S	S	S	S	S	S	S	S	S	S	S	SSW	SSE	SSE	SSE	SSE	SSE	
	700	3.6	4.0	3.6	4.0	5.1	5.0	4.5	4.8	4.7	6.0	6.3	5.7	6.1	5.6	5.0	4.6	3.6	3.4	3.5	2.7	3.2	3.6	4.3	5.2	4.50
	200	3.1	3.9	3.5	3.1	4.8	4.6	4.1	4.2	4.5	5.2	5.5	5.2	5.4	4.9	4.4	4.3	3.2	2.8	2.9	2.4	2.6	2.9	3.3	4.2	3.96
	30	1.9	2.6	2.6	2.1	3.4	3.2	2.8	2.9	3.1	3.5	3.5	3.3	3.6	3.2	2.7	2.7	2.1	1.7	1.8	1.3	1.5	1.8	1.9	2.7	2.58
	4	1.6	2.3	2.1	1.7	2.3	2.5	2.5	2.1	2.0	2.3	2.2	2.3	2.6	2.4	2.0	1.6	1.0	1.0	1.0	0.8	1.1	1.2	1.4	1.8	1.82
8	Direction	SSE	SSE	SSE	SSE	SSE	S	S	S	S	S	S	S	S	S	SSE	SSE	SE	ESE	S	SE	E	E	SE	S	
	700	5.0	4.5	3.5	4.3	3.5	3.1	2.7	2.0	1.3	1.3	2.4	2.4	1.7	2.5	1.8	1.2	1.6	1.6	1.2	1.3	1.3	1.3	1.2	1.2	2.25
	200	4.4	3.9	3.3	3.8	3.4	2.9	2.5	2.1	1.6	1 4	2.3	2.4	1.8	2.4	1.9	1.5	2.0	1.6	1.3	1.3	1.1	1.0	1.1	1.1	2.17
	30	2.9	2.5	2.1	2.4	2.1	1.8	1.6	1.4	1.0	0.6	1.4	1.6	1.0	1.6	1.2	0.9	1.3	1.0	0.9	0.8	0.7	0.7	0.7	0.8	1.38
	4	2.1	1.9	1.3	1.5	1.6	1.4	1.2	0.9	0.7	0.4	0.8	1.1	0.4	0.9	0.8	0.4	0.5	0.6	0.6	0.5	0.6	0.6	0.5	0.3	0.90
9	Direction	NE	N	N	NNE	NNE	NNE	N	N	NNE	N	N	NNW	N	N	N	NNW	NNW	NNW	N	N	NNW	NNW	NNW	N	
	700	(1.2)	(1.7)	(2.4)	(2.9)	(3.0)	(2.8)	(1.3)	(3.0)	(2.8)	(2.9)	(2.9)	(2.8)	2.6	2.7	2.7	3.0	2.9	3.2	3.3	2.9	2.2	2.4	2.7	2.2	(2.60)
	200	1.4	1.8	2.4	2.7	2.7	2.4	1.3	3.0	2.9	2.8	2.9	2.6	2.5	2.7	2.8	2.7	2.8	2.7	3.0	2.6	2.2	2.0	2.1	2.1	2.46
	30	1.1	1.4	1.8	1.9	2.0	1.7	1.1	2.0	2.0	1.9	1.9	1.8	1.8	1.7	1.7	1.7	1.8	2.0	2.0	1.7	1.5	1.4	1.3	1.4	1.69
	4	0.5	0.7	0.9	1.0	-	-	-	-	-	-	0.9	0.8	0.7	0.8	1.1	1.4	1.5	1.5	1.5	1.3	1.0	1.0	1.0	1.0	
10	Direction	NE	E	SE	SE	E	SSW	S	S	S	SSW	S	SW	SSW	SSW	SSW	SW	SSW	SW	SSW	S	S	SSE	SE	SSW	
	700	2.8	2.9	2.0	1.3	1.4	1.3	1.4	1.8	2.8	2.8	2.8	2.7	2.7	2.6	2.0	1.5	1.3	1.6	2.4	3.1	2.4	1.5	2.2	2.7	2.17
	200	2.8	2.9	2.0	1.4	1.5	1.6	1.5	1.8	3.0	2.7	2.7	2.6	2.6	2.5	2.0	1.4	1.1	1.7	2.5	3.1	2.6	1.7	2.4	2.7	2.20
	30	1.8	2.2	1.3	0.8	1.0	1.1	0.9	1.1	1.9	1.7	1.6	1.6	1.6	1.5	1.2	0.9	0.5	0.9	1.4	1.9	1.6	1.0	1.5	1.6	1.36
	4	1.3	1.8	1.0	0.7	0.7	0.6	0.6	0.9	1.1	1.1	1.1	1.1	1.0	0.9	0.7	0.5	0.4	0.6	1.0	1.4	1.3	0.9	1.0	1.1	0.95
11	Direction	SSE	S	SSW	S	SSW	S	S	S	S	S	S	S	S	S	S	SSW	S	SSW	S	S	S	S	S	S	
	700	2.6	2.6	2.5	2.4	2.2	2.0	2.3	3.3	3.7	3.9	4.0	4.3	(4.5)	(4.0)	(3.6)	(3.8)	(4.6)	(5.0)	(5.2)	(5.1)	(4.3)	(4.4)	(3.7)	(3.8)	(3.66)
	200	2.7	2.5	2.5	2 4	2.2	2.1	2.4	3.3	3.7	3.9	3.9	4.1	4.3	3.9	3.9	3.8	4.1	4.5	4.8	4.5	4.0	4.2	3.7	3.6	3.54
	30	1.6	1.6	1.6	1.5	1.4	1.3	1.5	2.1	2.4	2.4	2.4	2.6	2.7	2.6	2.6	2.5	2.8	3.1	3.3	3.1	2.8	2.9	2.4	2.3	2.31
	4	1.2	1.1	1.0	0.8	0.7	0.7	0 8	1.1	1.5	1.8	2.1	2.1	-	-	-	-	-	-	-	-	-	-	-	-	
12	Direction	SSW	SSW	SSW	SSW	SSW	S	S	SSE	S	SSW	S	SSE	S	S	S	S	S	SSE	S	SSE	S	S	SE	SSE	
	700	(3.4)	(3.4)	(4.0)	(3.7)	(3.6)	(3.6)	4.1	5.0	4.6	4.1	4.6	4.6	4.3	3.7	3.8	3.6	3.4	3.3	3.7	3.2	3.3	3.1	2.1	1.6	(3.66)
	200	3.2	3.4	3.5	3.6	3.3	3.1	3.7	4.7	4.2	3.8	4.2	4.2	3.9	3.3	3.3	3.0	3.0	2.9	2.8	2.7	2.6	2.5	1.8	1.3	3.25
	30	2.1	2.2	2.4	2.5	2.3	2.1	2.6	3.1	2.8	2.5	2.7	2.7	2.5	2.1	2.1	2.0	1.9	1.8	1.6	1.8	1.7	1.7	1.0	0.6	2.12

Table I. (Continued.)

Date	Height cm	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
		Hour																								
July 13	Direction	S	SSE	S	SSW	SE	SE	SE	SE	NE	NNE	NNE	WSW	NNW	NNW	NNW	NNW	NNW	NNW	NNW	NNW	NNW	NNW	NNW	W	
	700	2.1	2.0	1.5	1.4	<1.2	<1.2	<1.2	1.3	<1.2	<1.2	1.5	1.8	<1.2	<1.2	<1.2	1.2	1.6	1.9	1.6	2.1	1.7	1.4	1.0	1.6	(1.34)
	200	1.7	1.5	1.1	1.0	0.9	1.2	1.2	1.3	1.0	0.8	1.4	1.2	1.0	0.5	0.8	1.0	1.4	1.5	1.2	1.7	1.6	1.4	0.9	1.1	1.18
	30	0.8	0.7	0.5	0.6	<0.5	<0.5	<0.5	<0.5	0.6	0.5	0.6	0.6	<0.5	<0.5	<0.5	<0.5	0.6	0.5	0.5	0.8	0.8	0.6	0.4	0.6	(0.50)
	4	-	-	0.4	0.2	0.3	-	-	-	-	-	-	-	0.2	0.1	0.1	0.2	0.4	0.5	0.3	0.6	0.7	0.6	0.6	0.5	
14	Direction	WSW	SSW	SSW	SW	SW	SW	SW	SW	S	SSW	SSW	WSW	WSW	SSW	NE	N	N	NNE	NNW	NNW	N	NNE	NNE	N	
	700	4.1	3.9	3.1	5.2	5.7	5.4	5.4	4.7	3.4	1.9	2.5	2.7	2.7	1.9	1.4	2.0	1.5	1.5	1.4	1.5	1.1	1.2	1.3	1.4	2.79
	200	3.4	3.1	2.3	3.5	4.4	4.1	4.2	3.6	2.6	1.9	2.2	2.5	2.1	1.8	1.1	2.0	1.8	1.6	1.6	1.8	1.2	1.3	1.5	2.0	2.40
	30	2.1	2.0	1.6	2.6	3.1	2.7	2.8	2.3	1.6	0.9	1.6	1.4	1.1	1.1	0.5	1.0	0.8	0.6	0.8	0.8	0.4	0.5	0.7	1.0	1.42
	4	1.4	1.3	0.7	1.3	-	-	-	-	-	-	-	-	0.4	0.7	0.3	0.5	0.5	0.5	0.4	0.4	0.4	0.4	0.6	0.9	
15	Direction	N	N	NNW	N	N	NNW	N	ENE	NE	NNE	NNW	NNE	E	NNE	WNW	N	NNE	N	N	NNE	N	N	NE	NE	
	700	2.2	4.1	5.0	4.3	3.1	3.1	1.2	1.5	2.1	2.6	3.0	2.7	2.7	2.2	1.2	1.1	2.2	1.9	1.5	1.4	2.2	3.4	2.9	1.6	2.47
	200	2.3	3.2	3.9	3.1	2.2	2.5	1.5	1.6	1.8	2.3	2.5	2.2	2.1	2.1	1.2	1.0	1.8	1.6	1.5	1.4	1.5	2.4	2.3	1.8	2.08
	30	1.3	1.9	2.5	2.1	1.2	1.6	0.5	0.5	0.8	1.2	1.4	1.1	1.1	1.2	0.6	0.4	1.0	1.0	0.7	0.6	0.8	1.3	1.2	0.8	1.08
	4	1.1	1.5	1.7	1.3	0.8	1.1	0.4	0.3	0.5	0.7	0.9	0 5	0.8	0.8	0.6	0.3	0.4	0.6	0.5	0.4	0.6	1.0	1.0	0.7	0.77
16	Direction	NNE	NNE	NNE	N	NNE	N	N	NNE	E	E	SE	SE	SSE	SSE	SE	N	NNW	NNW	NE	NE	N	N	N	NNE	
	700	1.6	1.4	1.2	1.6	2.0	1.7	1.3	1.9	1.2	<1.2	1.2	1.3	1.5	1.8	1.1	2.3	2.3	1.5	1.7	1.8	1.4	1.2	1.2	1.4	1.52
	200	1.9	1.8	1.9	2.0	1.9	1.4	1.1	1.6	1.2	0.9	1.1	1.2	1.3	1.8	0.8	2.0	1.8	1.2	1.5	1.7	1.5	1.4	1.6	1.6	1.51
	30	0.9	1.1	0.7	0.6	0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	0.7	1.1	0.5	1.1	1.1	0.5	0.5	0.7	0.6	0.6	0.5	0.4	(0.63)
	4	0.9	0.9	0.7	1.0	1.0	0.4	0.2	0.4	0.4	0.3	0.3	0.2	0.4	0.6	0.4	0.9	0.7	0.4	0.6	0.6	0.4	0.5	0.6	0.6	0.56
17	Direction	ENE	SSE	SSE	SE	SE	SE	NE	NE	SE	SE	SSE	SSE	SE	S	SSE	SE	S	S	S	S	S	SSE	SSE	SSE	
	700	1.5	1.5	1.4	<1.2	<1.2	<1.2	1.2	1.5	1.1	1.1	1.4	1.8	2.3	3.6	4.7	4.1	4.7	5.1	4.7	5.3	6.4	6.6	6.0	6.1	3.10
	200	1.5	1.3	1.1	0.8	0.5	0.8	1.1	1.1	1.1	1.6	1.3	1.3	2.0	2.8	4.0	3.4	4.0	4.2	3.8	4.4	5.6	5.8	5.0	5.0	2.65
	30	0.5	0.6	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	0.9	0.7	0.6	1.2	1.7	2.6	2.2	2.8	2.8	2.5	2.8	3.6	3.8	3.3	3.5	(1.59)
	4	0.5	0.4	0.4	0.3	0.2	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
18	Direction	S	S	S	S	S	S	S	S	SSE	SSE	SSE	S	S	S	S	S	S	SSE	S	S	S	S	SSW	S	
	700	6.5	6.7	6.6	6.3	5.4	5.3	5.6	5.3	6.9	6.7	5.3	5.3	5.4	5.2	5.0	5.0	5.3	4.4	5.0	5.0	4.0	3.2	2.3	2.9	5.19
	200	5.2	5.4	5.4	5.0	4.1	4.1	4.9	4.1	5.5	5.9	4.6	4.4	4.6	4.4	4.3	4.3	4.3	3.7	4.3	4.4	3.4	2.5	1.7	2.3	4.28
	30	3.8	3.9	3.9	3.6	3.0	2.9	3.4	2.9	4.0	4.3	3.3	3.2	3.3	3.1	3.0	3.2	3.1	2.5	3.9	3.2	2.4	1.6	1.1	1.4	3.08
19	Direction	S	S	S	S	S	S	S	SSE	SSE	S	S	SW	SSW	SSW	SSW	S	S	S	S	S	SSW	S	SW	SW	
	700	3.5	5.0	4.2	3.7	5.2	5.1	5.4	4.9	4.4	4.5	5.0	4.8	4.7	4.4	4.4	3.9	3.8	3.5	3.6	4.3	4.5	5.0	5.3	4.8	4.50
	200	2.8	4.0	3.3	2.8	4.3	4.1	4.2	3.9	3.7	3.8	4.1	3.9	3.7	3.5	3.4	3.3	3.2	3.0	2.8	3.1	3 6	3.9	4.2	4.0	3.61
	30	1.8	2.7	2.4	2.1	3.0	3.0	3.1	2.8	2.6	2.8	3.0	2.8	2.6	2.4	2.3	2.3	2.4	2.2	1.9	2.3	2.6	2.7	3.1	3.0	2.58
20	Direction	SSW	SSW	S	S	S	S	S	SSW	SSW	SSW	SSW	SSW	S	S	S	S	S	S	S	S	S	SSE	S	S	
	700	4.6	4.5	4.9	5.4	5.5	5.1	5.6	5.2	4.8	4.7	4.0	4.1	4.8	4.8	4.3	4.5	5.1	4.8	6.1	6.8	6.5	6.7	6.8	6.7	5.26
	200	3.7	3.6	3.9	4.6	4.7	4.4	4.6	4.2	4.1	4.0	3.4	3.4	3.8	4.1	3.7	3.6	3.8	3.7	5.0	5.5	5.4	5.8	5.8	5.5	4.35
	30	2.7	2.7	2.9	3.4	3.6	3.3	3.4	3.2	3.1	3.1	2.5	2.5	2.9	3.2	2.8	2.7	2.9	2.7	3.6	3.9	4.0	4.0	4.1	4.1	3.22
21	Direction	S	SSE	S	S	S	S	S	S	SSE	S	S	S	SSW	SSW	SSW	SW	SW	SW	SW	SW	WSW	WSW	W	W	
	700	7.0	7.8	6.4	6.6	6.4	5.2	4.7	4.7	4.0	4.5	4.3	3.3	3.6	3.8	3.8	3.0	3.5	4.1	4.4	3.9	4.0	3.2	3.6	3.0	4.53
	200	5.8	6.8	5.5	5.5	5.3	4.3	4.0	4.0	3.1	3.5	3.6	2.6	2.7	3.1	3.2	2.4	2.8	3.5	3.8	3.5	3.6	2.9	3.0	2.5	3.79
	30	4.2	4.9	4.0	3.9	3.8	3.0	2.6	2.4	2.2	2.4	2.4	1.9	1.8	2.1	2.2	1.7	2.0	2.7	2.7	2.7	2.8	2.1	2.2	2.0	2.70
	4	-	-	-	-	-	-	-	-	-	-	-	-	1.2	1.2	-	-	-	-	-	-	-	1.6	1.7	1.4	

Table I. (Continued.)

Date	Height cm	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
July 22	Direction	NW	N	NNE	NNE	NNE	N	NNE	NE	NE	NE	N	N	N	N	N	N	NW	NW	NNW	N	NNW	N	NNW	N	
	700	(2.4)	(2.9)	(1.8)	(2.3)	(3.2)	(3.6)	(3.2)	3.9	4.7	4.2	3.1	3.8	2.7	1.6	2.1	1.4	2.3	3.0	2.9	3.5	3.9	5.7	4.3	2.8	(3.14)
	200	2.3	2.8	1.7	2.0	3.3	3.8	3.1	3.5	4.3	3.6	2.7	3.2	2.2	1.6	1.7	1.3	1.8	2.4	2.3	3.1	3.3	4.8	3.5	2.4	2.78
	30	1.5	2.1	1.3	1.4	2.2	2.6	2.1	2.4	3.3	2.8	2.0	2.6	1.7	1.1	1.1	1.0	1.4	2.2	1.9	2.6	2.8	4.0	2.9	2.1	2.13
	4	1.0	1.5	1.2	1.2	1.3	1.5	1.3	1.7	2.4	2.1	1.7	1.9	1.3	0.9	0.9	0.8	0.9	-	-	-	-	-	-	-	
23	Direction	N	WNW	WNW	NW	NW	NW	NW	NW	WNW	NW	NW	WNW	WNW	WNW	WNW	WNW	W	WNW	WNW	WNW	WNW	W	WSW	SW	
	700	3.4	4.6	5.0	4.7	4.8	5.2	5.6	5.5	5.0	4.4	5.0	5.9	5.0	5.4	5.1	3.7	4.5	5.0	(4.7)	(4.8)	(4.9)	(5.5)	(6.5)	(4.9)	(4.96)
	200	2.9	4.1	4.7	3.8	4.0	4.3	4.5	4.6	4.1	4.1	4.5	5.3	4.7	4.9	4.7	3.2	3.8	4.5	4.2	4.1	4.4	5.1	5.9	4.5	4.37
	30	2.4	3.0	3.1	2.8	3.1	3.2	3.5	3.6	3.1	3.2	3.4	3.9	3.5	3.5	3.5	2.7	3.3	3.6	3.3	3.2	3.1	3.5	4.2	3.0	3.28
	4	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	2.8	-	-	
24	Direction	WSW	WSW	WSW	WSW	SW	SW	SW	SSW	SSW	SW	SSW	S	S	S	S	S	S	S	SSW	S	SSE	S	S	S	
	700	(4.2)	(4.3)	(4.2)	(4.5)	(4.4)	(4.7)	(4.7)	(4.4)	(4.4)	(4.1)	(3.9)	(5.1)	6.1	5.5	5.8	6.7	7.3	7.1	5.9	6.1	7.0	(6.3)	(6.6)	(7.2)	(5.44)
	200	3.8	3.6	3.5	3.4	3.4	3.1	3.2	3.7	4.3	4.0	4.3	4.8	5.5	4.8	5.2	6.0	6.7	6.7	5.5	5.1	6.3	5.1	5.4	6.0	4.72
	30	2.9	2.9	2.8	2.7	2.9	2.8	3.2	3.0	3.4	3.0	3.0	3.1	3.8	3.5	3.7	4.3	4.8	4.8	3.9	3.7	4.8	3.7	3.6	4.4	3.53
25	Direction	SSW	SSW	SSW	SSW	SSW	S	S	S	S	S	SSW	WSW	SSW	SW	SW	SW	SW	SW	S	S	SE	S	S	S	
	700	(6.7)	(6.9)	7.3	5.9	6.2	6.5	6.5	8.0	7.5	6.9	5.2	5.6	5.3	5.4	4.4	4.7	6.3	5.2	3.1	1.9	1.4	1.8	2.4	2.1	5.13
	200	6.4	6.8	6.2	5.2	5.2	5.6	5.6	6.8	6.4	6.1	4.5	4.6	4.5	4.4	3.6	3.8	5.3	4.3	2.7	1.7	1.3	1.5	2.0	1.7	4.42
	30	4.4	4.5	4.3	3.8	3.7	4.3	4.2	4.6	4.3	4.1	2.9	3.1	3.1	3.1	2.5	2.5	3.7	3.1	1.9	1.2	0.9	1.0	1.4	1.2	3.08
26	Direction	SSW	SW	WSW	W	W	W	NW	NW	NNW	N	N	N	NNE	NE	NNE	NNE	NNE	N	N	NNW	NNW	N	N	N	
	700	1.5	1.5	2.7	2.8	2.0	1.6	1.3	1.1	2.0	3.9	3.7	4.3	5.2	3.7	3.9	4.1	3.2	2.8	2.2	2.7	2.2	1.5	<1.2	<1.2	2.56
	200	1.3	1.5	1.9	2.4	1.8	1.5	1.2	1.2	2.1	3.3	3.2	3.5	4.1	3.1	3.3	3.1	2.8	2.1	2.0	2.2	1.9	1.6	1.3	1.3	2.24
	30	0.9	1.0	1.2	1.6	1.2	1.2	1.0	0.9	1.6	2.5	2.4	2.7	3.2	2.5	2.6	2.6	2.1	1.6	1.3	1.4	1.2	1.0	0.8	0.7	1.63
	4	-	-	-	-	0.7	0.4	0.5	0.5	0.5	1.4	1.3	1.5	1.9	1.2	1.5	1.7	1.3	0.9	0.7	0.9	0.9	0.6	0.5	0.5	
27	Direction	NNW	NNW	NNW	N	NNE	NNW	N	NNE	NE	N	N	NW	NNW	NNW	NW	NW	N	NNE	NNE	N	N	N	NNE	N	
	700	<1.2	<1.2	1.3	2.3	3.5	3.1	2.4	2.7	3.4	2.0	2.8	3.0	1.8	1.9	1.4	1.7	1.2	1.5	2.0	3.0	1.8	3.7	5.4	4.7	2.42
	200	1.4	1.3	1.5	2.3	2.6	2.4	2.1	2.2	2.9	1.4	2.2	2.4	1.7	1.7	1.5	1.5	1.0	1.4	2.1	1.9	1.7	3.1	4.4	4.2	2.12
	30	0.8	0.8	1.2	1.4	1.8	2.0	1.5	1.7	2.3	1.2	1.5	1.8	1.3	1.2	1.0	1.1	0.6	0.8	1.6	1.4	1.2	2.2	3.5	3.3	1.55
	4	0.4	0.3	0.5	1.0	1.3	1.2	1.1	1.0	1.6	1.0	0.9	1.1	0.7	0.7	0.6	0.7	0.5	0.8	1.2	1.3	0.8	1.7	2.9	2.2	1.06
28	Direction	N	N	NNW	N	N	N	NNE	N	N	NE	NE	NE	N	N	N	N	N	NNW	NW	NW	NW	NW	NW	NW	
	700	5.6	5.9	6.0	4.5	2.6	1.6	4.3	5.4	6.1	5.9	5.3	5.1	5.1	5.0	4.4	5.1	4.7	3.8	3.1	2.5	2.1	1.6	2.1	1.4	4.13
	200	4.5	4.9	4.9	3.1	2.0	1.6	3.6	5.1	5.2	4.8	4.6	4.5	4.6	4.2	4.0	4.5	4.1	3.4	2.8	2.1	1.8	1.4	2.1	1.2	3.54
	30	3.3	3.8	3.8	2.3	1.3	1.1	2.9	4.1	4.0	4.0	3.6	3.5	3.6	3.5	3.1	3.6	3.4	2.6	2.0	1.6	1.4	1.1	1.5	1.0	2.75
	4	2.7	3.1	2.5	1.3	0.8	0.5	1.9	3.2	2.9	2.6	2.5	2.5	2.5	2.4	1.9	2.6	2.6	1.9	1.6	1.5	1.3	0.7	0.8	0.4	1.95
29	Direction	NW	NW	NW	NNW	NNW	NNW	NW	NW	NNW	NW	NW	NW	NW	NW	WNW	WNW	WNW	WNW	NNE	NNE	NNW	WNW	WNW	WNW	
	700	1.3	1.3	1.4	1.2	1.5	2.0	2.0	2.3	1.6	1.2	1.1	1.4	1.7	1.8	2.8	3.6	4.3	3.4	1.4	1.5	1.6	4.4	4.4	3.8	2.21
	200	1.2	1.4	1.4	0.9	1.1	1.8	1.8	2.0	1.3	1.1	0.8	1.2	1.6	1.5	2.2	3.1	3.7	2.2	1.4	1.4	1.6	3.7	3.5	3.0	1.87
	30	0.9	0.9	0.9	0.7	0.9	1.4	1.4	1.5	0.9	0.8	0.7	0.9	1.0	0.9	1.4	2.1	2.4	1.5	0.9	1.0	0.8	2.5	2.4	2.1	1.29
	4	0.2	0.5	0.7	0.7	0.6	0.8	0.9	0.9	0.7	0.4	0.3	0.5	0.8	0.6	0.9	1.5	1.9	1.3	0.5	0.7	0.6	1.7	1.9	1.5	0.88
30	Direction	WNW	WNW	WNW	W	NW	WNW	NW	NW	NW	NW	NW	WNW	WNW	-	-	-	-	-	-	-	-	-	N	N	
	700	4.0	5.2	5.4	4.7	3.8	3.3	3.0	2.0	1.5	3.6	3.5	1.3	1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	(2.14)
	200	3.2	4.2	4.2	3.7	2.9	2.6	2.5	1.7	1.3	2.7	2.6	1.1	1.0	0.8	0.7	0.9	1.0	0.6	1.2	1.1	0.8	1.2	1.6	1.5	1.88
	30	2.3	2.9	2.9	2.5	2.1	1.8	1.5	1.3	1.0	2.0	1.5	0.7	0.7	0.6	0.6	0.9	0.9	0.6	0.8	0.8	0.5	0.7	1.0	1.0	1.32
	4	1.8	2.4	2.4	2.0	1.6	1.3	1.1	0.9	0.8	1.1	1.2	0.6	0.5	0.2	0.2	0.3	0.3	0.2	0.3	0.5	0.5	0.5	0.8	0.8	0.93

Table I. (Continued.)

Date	Height cm	Hour 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
July 31	Direction	N	N	N	N	N	N	N	N	N	N	N	N	SSE	S	S	S	S	-	-	-	-	W	-	SSW	
	700	<1.2	1.2	1.4	1.4	1.1	1.3	1.4	1.2	1.2	1.4	0.9	1.2	3.2	3.7	3.1	2.4	1.9	1.1	<1.2	<1.2	1.1	1.4	1.2	1.4	(1.52)
	200	1.5	1.4	1.7	1.5	1.3	1.3	1.2	0.9	1.1	1.3	0.8	0.8	2.8	3.2	2.6	1.9	1.6	0.7	0.7	0.8	0.7	0.9	1.0	0.8	1.35
	30	1.0	1.0	1.2	1.1	0.9	0.9	0.9	0.7	0.8	0.8	0.6	0.5	1.6	2.4	2.0	1.5	1.0	0.5	0.5	0.5	0.6	0.6	0.6	0.5	0.95
	4	0.7	0.8	1.0	0.8	0.6	0.6	0.7	0.5	0.3	0.2	0.2	0.3	1.2	1.4	0.9	-	-	-	-	-	-	-	-	-	-
Aug. 1	Direction	S	SSE	SSE	S	S	S	SSW	SSW	S	S	S	S	S	SSW	SSW	SSW	SSW	SSW	SSW	SSW	S	SSE	SSW	S	
	700	2.1	2.9	2.1	1.9	2.0	2.6	1.1	1.6	2.5	3.4	3.8	3.6	4.0	4.2	4.3	3.6	3.5	4.2	5.0	5.4	3.9	3.0	3.2	3.5	3.22
	200	1.4	1.7	1.3	1.1	1.4	1.4	0.6	0.9	1.7	2.4	2.9	2.7	3.1	3.0	3.3	2.9	2.7	3.2	3.8	4.0	2.8	2.3	2.7	2.9	2.34
	30	0.8	1.1	1.0	1.0	1.0	0.9	0.5	0.6	1.1	1.7	2.2	2.2	2.4	2.3	2.6	2.2	2.0	2.3	2.6	2.8	1.9	1.6	1.7	1.8	1.68
2	Direction	SSW	S	SSW	SSW	SW	SW	SSW	SSW	S	S	SSE	S	S	SSE	S	SSE	S	S	SSW	S	S	S	S	S	
	700	3.2	4.0	4.9	6.2	5.4	5.3	5.2	4.7	3.4	5.4	6.3	6.5	6.3	5.7	4.0	3.0	5.0	6.3	7.2	8.4	8.8	8.0	6.8	5.9	5.66
	200	2.1	3.0	3.7	4.7	4.1	4.0	3.7	3.3	2.9	4.5	5.4	5.5	5.4	4.8	3.3	2.7	4.3	5.2	5.9	7.0	7.6	6.6	5.5	4.5	4.57
	30	1.5	2.0	2.5	3.0	2.5	2.4	2.4	2.1	2.0	3.2	3.9	4.0	3.9	3.5	2.7	1.9	3.2	3.9	4.1	5.2	5.6	4.7	3.9	3.1	3.22
	4	-	-	-	-	-	-	-	-	-	-	2.4	2.5	2.4	1.9	1.2	0.7	1.6	2.6	-	-	-	-	-	-	-
3	Direction	S	S	SSE	S	S	S	S	S	SSW	SSW	SSW	SSW	SSW	SSW	SSW	SSW	SW	SW	SSW	S	S	S	S	S	
	700	6.7	6.8	5.1	5.2	4.1	3.9	3.7	4.2	3.9	3.0	3.7	5.3	4.8	4.1	4.1	4.0	6.3	4.5	4.5	3.6	2.7	3.2	2.3	1.8	4.23
	200	5.6	5.3	4.3	4.4	3.0	3.2	2.7	3.1	3.0	2.5	2.4	4.0	3.7	3.0	2.7	3.4	4.6	3.6	3.3	2.7	1.9	2.1	1.9	1.3	3.24
	30	3.7	3.6	3.2	2.8	2.2	2.1	1.8	2.1	2.0	1.6	1.6	2.6	2.4	1.9	1.7	2.2	3.0	2.3	2.1	1.9	1.3	1.4	1.2	0.9	2.15
4	Direction	SSE	SSE	S	S	S	SSE	SSE	S	S	SSE	SSE	SW	W	-	SW	SSW	S	SSE	S	SSE	S	S	S	S	
	700	1.5	2.5	3.6	3.6	1.8	2.4	3.4	2.9	2.6	2.8	1.4	2.1	2.8	1.2	2.8	4.0	2.0	2.3	3.6	3.1	1.5	1.7	3.0	3.5	2.59
	200	1.1	1.7	2.5	2.1	1.3	2.1	3.1	2.6	2.4	2.3	1.2	1.7	2.0	0.9	2.2	3.3	1.7	1.8	2.9	2.4	1.1	0.9	2.1	2.8	2.01
	30	0.9	1.1	1.8	1.4	0.9	1.4	2.4	2.0	1.9	1.7	0.7	1.0	1.3	0.6	1.4	1.9	1.2	1.1	1.8	1.5	0.8	0.7	1.3	1.9	1.36
	4	-	-	-	-	-	-	-	-	-	-	-	-	-	0.2	1.0	1.5	0.8	-	-	-	-	-	-	-	
5	Direction	SSE	SSE	S	S	S	S	SSW	S	SSW	SW	SW	SSW	SW	W	WNW	N	N	NNW	NNW	NNW	NNW	N	N	N	
	700	4.0	3.8	2.5	3.3	3.3	2.1	2.4	1.4	2.5	2.0	1.2	1.1	1.6	1.5	1.2	1.2	1.2	1.4	1.4	1.2	1.5	1.2	1.2	1.4	1.90
	200	2.9	2.8	1.6	2.1	1.7	1.2	1.4	1.3	2.1	1.8	1.1	1.3	1.6	1.6	1.7	1.2	1.4	1.4	1.3	1.7	1.5	1.3	1.7	1.5	1.63
	30	2.1	1.9	0.8	1.2	1.0	0.8	0.9	0.9	1.3	1.1	0.8	0.7	0.8	1.1	1.3	0.9	0.9	0.8	0.9	1.2	1.0	0.9	1.1	1.0	1.06
	4	-	-	-	-	-	-	-	-	-	0.7	0.4	0.2	0.2	0.5	0.6	0.6	0.6	-	-	-	-	-	-	-	
6	Direction	NNW	NNW	N	N	N	NW	W	-	-	NE	N	NNW	N	N	N	N	NE	NNE	NNE	NNW	NNE	N	N	N	
	700	1.2	<1.2	<1.2	1.2	1.7	1.2	1.6	1.8	<1.2	1.5	1.6	1.9	1.8	1.3	<1.2	1.4	2.1	2.2	1.9	1.4	1.9	1.7	1.3	1.4	(1.47)
	200	1.3	1.5	1.4	1.3	1.5	1.7	1.9	1.6	1.3	2.1	2.2	2.2	2.2	1.9	1.7	2.0	2.1	2.1	1.8	1.9	1.9	2.2	1.8	1.4	1.79
	30	0.9	0.9	0.9	1.0	1.2	1.3	1.1	0.9	0.9	1.4	1.5	1.6	1.6	1.5	1.4	1.3	1.5	1.5	1.3	1.4	1.3	1.6	1.3	1.3	1.28
	4	-	-	-	-	-	0.9	0.7	0.7	0.6	1.0	1.1	1.0	1.0	0.9	0.9	1.0	1.1	1.1	0.9	0.9	0.9	1.0	0.9	1.1	
7	Direction	NNE	NE	NE	NE	N	NE	NE	NE	N	E	NNE	NE	N	NNE	N	NE	NE	NE	NE	NE	NE	NE	ESE	-	
	700	1.6	1.8	2.0	1.5	2.3	2.3	1.4	1.4	1.8	2.4	1.7	2.2	2.1	1.6	1.7	3.2	3.0	1.4	1.9	1.5	1.4	1.3	1.6	1.5	1.86
	200	1.9	1.7	1.8	1.3	1.9	1.8	1.6	1.6	2.0	2.0	1.7	2.1	1.8	1.5	2.0	2.5	2.5	1.8	1.7	1.7	1.2	1.2	1.4	1.1	1.74
	30	1.4	1.1	1.0	0.8	1.1	1.2	1.1	1.1	1.3	1.2	1.2	1.4	1.1	1.1	1.3	1.6	1.9	1.1	1.2	1.1	0.9	0.9	1.0	0.7	1.16
	4	1.0	0.7	0.7	0.4	0.5	0.7	0.7	0.6	0.6	0.6	0.7	0.7	0.8	0.9	0.5	0.8	0.8	0.5	0.7	0.6	0.4	0.3	0.3	0.3	0.62
8	Direction	S	N	E	-	-	S	S	S	S	S	SW	ESE	ESE	E	E	E	E	NE	NNE	N	NNE	NNE	N	N	
	700	1.3	1.2	1.3	1.7	1.5	1.2	1.8	1.3	1.3	1.2	1.2	<1.2	<1.2	<1.2	<1.2	1.3	1.3	1.3	1.4	1.1	1.4	1.6	1.8	2.5	(1.33)
	200	1.1	1.0	0.8	1.2	1.1	1.0	1.3	1.0	1.0	0.9	0.8	1.1	1.4	1.4	1.4	1.3	1.2	1.3	1.2	1.0	1.4	1.2	1.5	2.0	1.20
	30	0.8	0.7	0.5	0.6	0.6	0.6	0.8	0.7	0.6	0.6	0.6	0.7	0.8	0.8	0.8	0.8	0.7	0.8	0.7	0.7	0.9	0.7	0.7	1.2	0.72
	4	0.5	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-

Table I. (Continued.)

Date	Height cm	Hour 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
Aug. 9	Direction	NNW	N	N	ENE	NNE	NE	NE	NE	NE	ENE	NE	NE	NE	ENE	-	S	S	S	S	SE	S	-	-	-	
	700	3.2	2.9	2.2	1.6	1.7	1.5	1.0	1.1	1.5	1.7	1.3	1.1	1.3	1.5	1.4	1.3	1.4	1.2	0.9	1.3	1.4	<1.2	<1.2	<1.2	(1.45)
	200	2.4	2.0	1.4	1.0	1.2	1.0	0.9	1.0	1.1	1.2	0.9	0.9	1.3	1.3	0.8	0.6	0.9	1.0	0.9	0.8	0.8	0.5	0.6	0.7	1.05
	30	1.6	1.5	1.1	0.8	1.0	0.7	0.6	0.8	0.8	0.8	0.7	0.6	0.9	0.8	0.7	0.5	0.6	0.6	<0.6	<0.6	<0.6	<0.6	<0.6	0.6	(0.72)
10	Direction	-	-	N	NE	N	N	NNE	N	N	N	N	N	N	NNW	N	NNW	NNW	NNE	NNE	NNE	NNE	N	N	N	
	700	<1.2	<1.2	1.3	3.0	3.2	3.2	3.6	3.6	3.9	4.0	3.5	3.3	3.2	3.3	3.3	3.5	2.8	2.1	2.7	2.7	3.1	3.8	3.5	3.9	3.00
	200	0.6	0.9	1.6	2.2	2.3	2.3	3.0	2.8	3.2	3.2	2.9	2.5	2.4	2.4	2.5	2.8	2.1	1.9	2.2	2.1	2.4	2.9	2.9	3.1	2.38
	30	0.6	0.7	1.2	1.8	1.7	1.8	2.1	2.1	2.4	2.4	2.2	1.9	1.8	1.8	2.0	2.1	1.6	1.4	1.4	1.4	1.7	2.0	2.1	2.2	1.77
	4	-	-	-	-	-	-	-	-	-	-	-	-	-	1.2	1.5	1.5	1.1	0.9	1.0	1.0	1.3	1.4	1.4	1.5	
11	Direction	N	NE	NNE	NNW	NNW	NNW	NNW	NNE	N	N	N	N	NW	N	NNW	NW	NW	NNE	N	NE	-	N	ESE	-	
	700	3.8	3.7	3.3	3.1	3.2	1.8	1.2	1.7	1.2	1.2	1.2	1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	(1.51)
	200	3.3	3.2	2.9	2.7	2.7	1.8	1.2	1.6	0.9	1.0	1.0	0.7	0.7	1.0	1.2	1.3	1.4	1.2	0.9	1.0	0.9	0.9	0.9	0.9	1.47
	30	2.5	2.5	2.2	2.0	1.9	1.2	0.9	1.0	0.6	0.8	0.7	0.6	0.6	0.6	0.7	0.8	0.8	0.7	0.6	0.6	0.6	0.7	0.7	0.6	1.04
	4	1.7	1.8	1.7	1.5	1.3	0.9	0.6	0.6	0.5	0.4	0.3	0.4	0.5	0.6	0.6	0.6	0.6	0.6	0.5	0.5	0.4	0.4	0.5	0.5	0.75
12	Direction	-	-	ENE	E	E	E	ESE	ESE	SSE	ESE	E	NNE	NNE	NNE	NNE	N	N	N	NNE	N	NNE	N	NE	NNW	
	700	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	<1.2	1.6	1.6	1.3	1.3	1.3	1.4	1.7	1.2	1.6	2.0	1.8	1.7	1.9	2.6	1.4	(1.32)
	200	1.2	1.2	0.8	1.2	1.2	1.1	0.5	0.8	1.2	1.3	1.3	1.2	1.4	1.8	1.7	1.6	1.3	1.3	1.9	1.7	1.6	1.7	2.3	1.5	1.37
	30	0.7	0.6	0.5	0.7	0.7	0.5	0.5	0.7	0.8	0.7	0.7	0.8	1.0	1.1	1.1	1.1	0.8	0.9	1.2	1.1	0.9	1.1	1.6	1.0	0.87
	4	0.5	0.4	0.5	0.5	0.4	0.3	0.2	0.4	0.6	0.6	0.5	0.4	0.7	0.9	1.0	1.0	0.7	0.9	1.0	0.8	0.9	0.9	1.2	0.8	0.67
13	Direction	N	NNE	ENE	NNE	NNW	NNW	NE	NNW	NNW	NNE	NNE	N	N	NNE	N	W	NNW	NNW	N	N	NNW	N	N	ENE	
	700	1.3	1.2	1.6	2.1	1.9	1.6	2.2	1.8	1.7	1.5	1.7	1.8	1.6	2.0	2.6	3.1	3.4	3.8	3.6	2.5	1.8	2.5	1.9	2.0	2.13
	200	1.4	1.6	1.7	2.1	2.0	1.7	2.1	1.9	1.4	1.5	1.9	1.9	1.7	2.1	2.2	2.4	3.0	3.0	2.9	2.3	2.1	2.6	2.4	2.3	2.09
	30	0.9	0.9	1.1	1.4	1.3	1.2	1.3	1.2	0.9	1.0	1.4	1.3	1.1	1.2	1.4	1.6	1.9	2.0	2.0	1.6	1.4	1.8	1.6	1.6	1.38
	4	0.7	0.8	1.0	1.1	0.9	0.9	1.0	0.8	0.6	0.7	0.9	1.0	0.8	0.9	1.1	1.2	1.4	1.6	1.6	1.0	1.2	1.5	1.5	1.6	1.08
14	Direction	N	N	N	NNW	N	N	N	N	NNW	N	N	N	N	N	N	N	N	ENE	NE	E	N	N	N	ENE	
	700	2.9	3.0	4.3	4.8	4.1	3.3	3.8	3.6	3.8	3.7	4.1	3.8	3.5	3.0	3.1	3.9	3.1	3.1	3.5	2.1	2.8	2.3	3.7	3.7	3.46
	200	2.7	2.9	3.3	3.8	3.2	2.6	3.0	2.8	3.0	3.0	3.2	3.2	2.8	2.5	2.4	3.1	2.7	2.7	2.8	1.6	2.1	2.0	3.1	3.1	2.82
	30	1.8	1.9	2.1	2.7	2.2	1.9	2.1	1.9	2.0	2.1	2.2	2.2	2.1	1.8	1.7	2.3	2.0	2.0	2.1	1.4	1.6	1.6	2.4	2.4	2.02
	4	1.6	1.6	1.8	2.3	2.1	1.6	1.7	1.9	1.9	1.8	1.8	1.7	1.7	1.1	0.9	1.6	1.6	1.5	1.2	0.5	1.0	0.9	1.8	2.0	1.57
15	Direction	ENE	ENE	NE	N	N	NNE	N	N	N	NE	ENE	-	WNW	WNW	WNW	-	NW	N	NE	-	-	-	N	NE	
	700	3.1	2.7	2.9	3.4	3.1	2.7	2.8	2.4	1.7	1.3	1.2	<1.2	<1.2	<1.2	<1.2	<1.2	1.4	1.3	1.2	1.2	1.2	1.5	1.6	2.3	(1.79)
	200	2.8	2.4	2.1	3.2	2.9	2.2	2.6	2.0	1.7	1.6	1.0	0.6	0.5	0.6	0.5	0.6	0.9	0.9	0.9	0.8	0.7	1.3	1.3	1.9	1.50
	30	2.1	1.8	1.6	2.4	2.2	1.8	2.0	1.5	1.3	1.1	0.7	0.6	0.5	0.4	0.5	0.6	0.8	0.7	0.7	0.5	0.6	1.0	1.0	1.6	1.17
	4	-	-	-	-	-	-	1.7	1.2	1.0	0.9	0.6	-	-	-	-	-	-	-	-	-	-	-	-	-	-

Table II. *Temperature. Mean Hourly Values of Temperature in Centigrades.*

Date	Height cm	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
June 26	500	0.1	-0.3	-0.1	1.4	1.5	0.5	1.0	1.7	2.8	2.2	1.6	1.4	1.3	1.5	2.4	1.5	1.9	1.1	0.4	0.5	0.2	-0.1	-0.1	0.6	1.04
	100	-0.8	-1.2	-0.6	0.0	0.6	0.0	0.4	1.5	1.9	1.4	1.2	0.9	1.1	0.8	0.5	0.6	0.8	0.5	0.1	-0.1	-0.6	-0.8	-1.1	-1.2	0.25
27	500	0.7	0.3	0.8	0.5	1.0	0.8	1.9	1.2	1.3	1.1	1.2	1.2	1.0	1.1	1.0	0.6	1.3	0.1	0.1	-0.3	-0.8	-1.6	-2.2	-2.6	0.40
	100	-1.0	-1.7	-1.3	-1.1	-0.8	-0.3	0.2	0.7	0.6	0.5	0.5	0.3	0.5	0.9	0.4	0.2	-0.1	-0.6	-1.0	-1.1	-1.3	-2.1	-2.8	-2.8	-0.55
28	500	-3.4	-3.9	-4.1	-3.9	-3.9	-2.0	-2.5	-2.8	-3.0	-2.6	-2.3	-2.4	-2.6	-0.9	-0.7	-1.5	-2.1	-3.2	-3.6	-3.9	-4.3	-4.4	-5.3	-5.4	-3.11
	100	-3.4	-4.1	-4.0	-4.1	-3.0	-2.5	-1.5	-2.3	-2.0	-1.7	-2.1	-1.0	-2.4	-1.7	-1.4	-3.1	-3.3	-3.5	-3.9	-4.3	-4.6	-4.6	-5.5	-5.5	-3.15
29	500	-5.1	-5.2	-4.0	-4.4	-4.0	-3.3	-2.4	-1.2	-1.1	-1.0	-0.6	-0.6	-0.6	-0.5	-0.4	-0.1	-0.7	-1.0	-0.8	-1.1	-1.0	-1.2	-2.0	-1.9	-1.88
	100	-5.7	-6.0	-5.5	-5.1	-4.4	-3.5	-2.7	-2.0	-1.5	-1.2	-0.6	-0.5	-0.6	-0.6	-0.4	-0.8	-1.3	-1.1	-1.3	-2.0	-2.6	-2.8	-2.9	-2.3	-2.39
30	500	-2.4	-1.5	-0.4	-0.3	-0.2	0.0	0.2	-0.1	-0.2	-0.5	-0.1	0.2	0.0	0.1	0.5	0.8	0.5	0.1	-0.1	-0.1	-0.4	-0.5	-0.4	-0.5	-0.22
	100	-3.2	-2.1	-0.8	-0.5	-0.2	0.0	-0.2	-0.5	-0.3	-0.9	-0.5	0.1	0.0	0.1	0.5	0.6	0.2	-0.1	-0.1	-0.2	-0.8	-0.8	-0.6	-0.6	-0.45
July 1	500	-0.5	-0.2	-0.6	-0.9	-0.6	-0.2	-0.1	-0.2	0.3	0.3	0.2	0.2	0.0	0.1	0.0	0.0	-0.1	-0.4	-0.4	-0.4	-0.5	-0.6	-0.6	-0.7	-0.25
	100	-0.4	-0.3	-0.7	-1.0	-0.5	-0.1	-0.1	-0.3	0.0	0.3	0.3	0.2	0.0	0.1	0.0	-0.1	-0.3	-0.4	-0.2	-0.3	-0.5	-0.6	-0.6	-0.6	-0.25
	4	-0.4	-0.3	-0.5	-0.6	-0.4	-0.2	-0.1	-0.2	-0.1	0.2	0.3	0.2	0.2	-0.2	-0.2	0.0	0.0	-0.1	-0.3	-0.3	-0.5	-0.5	-0.5	-0.6	-0.21
2	500	-0.6	-0.4	-0.5	-0.8	-0.8	-0.4	-0.1	-0.1	-0.2	-0.1	-0.1	-0.1	0.1	-0.1	-0.2	0.0	-0.1	-0.5	-0.8	-1.1	-1.5	-1.4	-1.2	-1.2	-0.51
	100	-0.6	-0.5	-0.6	-0.6	-0.6	-0.5	-0.4	-0.2	-0.2	0.0	0.1	0.2	0.0	-0.2	-0.2	-0.2	-0.2	-0.6	-0.8	-1.3	-1.5	-1.4	-1.2	-1.2	-0.53
	4	-0.6	-0.5	-0.5	-0.6	-0.5	-0.2	-0.1	-0.2	0.1	0.2	0.1	0.2	0.1	-0.1	-0.2	-0.2	-0.1	-0.3	-0.7	-1.1	-1.4	-1.3	-1.3	-1.2	-0.43
3	500	-1.4	-1.7	-1.5	-1.1	-0.9	-0.7	-0.8	-0.6	-0.9	-1.0	0.1	0.3	0.1	-0.2	-0.3	1.4	1.3	0.3	-0.2	0.0	0.9	-0.4	-2.0	-1.7	-0.46
	100	-1.3	-2.2	-1.9	-1.3	-1.2	-0.8	-0.6	-0.8	-1.1	-0.7	-0.1	-0.1	-0.1	-0.2	-0.4	0.4	0.0	-0.6	-0.7	-0.2	0.0	-1.3	-2.4	-2.1	-0.82
	4	-1.2	-2.1	-1.6	-1.2	-1.2	-0.7	-0.9	-0.8	-0.9	-0.6	-0.3	-0.3	-0.3	-0.4	-0.8	-0.9	-1.4	-1.1	-0.9	-0.9	-1.0	-2.0	-2.7	-2.1	-1.10
4	500	-0.6	-0.1	0.3	0.4	0.6	1.1	0.9	1.0	1.1	1.7	1.9	1.2	0.6	0.3	-0.5	-1.2	-1.3	-1.4	-1.6	-1.6	-1.8	-2.1	-2.3	-2.5	-0.25
	100	-0.6	-0.3	0.0	0.4	0.4	0.9	0.5	0.6	0.7	1.0	0.9	0.3	0.1	-0.1	-1.0	-1.4	-1.4	-1.5	-1.6	-1.6	-1.7	-1.9	-2.2	-2.4	-0.50
	4	-0.7	-0.1	0.0	0.2	0.2	0.2	0.1	-0.1	-0.1	-0.1	-0.3	-0.6	-0.5	-0.8	-1.0	-1.2	-1.5	-1.3	-1.4	-1.5	-1.7	-1.7	-2.0	-2.2	-0.75
5	500	-2.6	-2.7	-2.8	-3.1	-3.2	-3.2	-3.3	-3.3	-2.9	-3.6	-4.3	-3.8	-3.3	-3.1	-2.8	-2.4	-2.7	-3.1	-3.1	-3.6	-4.2	-3.7	-3.7	-3.8	-3.26
	100	-2.5	-2.6	-2.8	-2.9	-3.2	-3.2	-3.1	-3.2	-3.1	-3.6	-3.7	-3.4	-2.9	-2.5	-2.7	-2.4	-2.3	-2.9	-3.1	-3.6	-4.1	-4.1	-5.1	-5.5	-3.27
	4	-2.3	-2.2	-2.2	-1.5	-2.0	-2.5	-2.2	-2.0	-2.5	-2.2	-1.7	-1.6	-1.7	-1.5	-1.3	-1.4	-1.8	-2.2	-1.8	-3.1	-3.7	-4.6	-5.5	-6.5	-2.50
6	500	-4.5	-4.3	-3.6	-3.1	-2.9	-3.0	-2.7	-1.8	-0.6	0.0	0.3	1.0	0.5	0.1	0.3	0.3	0.2	-0.4	-0.7	-1.1	-1.0	-1.2	-1.4	-1.7	-1.30
	100	-5.4	-4.5	-3.7	-2.8	-2.7	-2.7	-2.4	-1.9	-0.4	0.4	0.6	0.9	0.3	0.6	0.5	0.2	0.0	-0.4	-0.7	-1.1	-1.1	-1.1	-1.3	-1.8	-1.27
	4	-6.3	-4.7	-3.6	-3.3	-2.9	-2.5	-2.3	-1.6	-1.1	-0.7	-0.1	-0.2	-0.1	-0.1	-0.3	-0.3	-0.1	-0.4	-0.5	-1.1	-1.2	-1.0	-1.2	-1.5	-1.55
7	500	-2.1	-2.1	-2.0	-2.4	-2.3	-2.1	-1.6	-1.4	-1.7	-1.1	-0.6	-1.0	-1.2	-0.9	-0.7	-0.8	-0.7	-0.6	-0.2	-0.9	-1.8	-1.7	-1.8	-1.1	-1.37
	100	-2.0	-1.8	-2.0	-2.5	-2.3	-2.0	-1.5	-1.6	-1.6	-0.9	-0.7	-0.9	-1.1	-0.8	-1.0	-0.9	-1.0	-1.1	-1.3	-2.2	-2.1	-2.2	-2.3	-1.2	-1.54
	4	-1.7	-1.7	-2.0	-2.1	-2.0	-2.0	-2.0	-1.9	-1.6	-1.5	-1.3	-1.3	-1.4	-1.3	-1.7	-1.6	-1.4	-1.7	-1.7	-1.4	-2.1	-3.0	-3.0	-2.1	-1.81
8	500	-0.7	-0.9	-1.9	-1.4	-0.9	-0.8	-0.4	-0.4	-0.5	-0.8	-0.5	-0.6	-1.0	-1.3	-1.8	-1.5	-1.5	-1.5	-1.8	-2.1	-2.2	-2.7	-2.9	-3.5	-1.40
	100	-0.6	-0.8	-2.0	-1.6	-1.1	-1.0	-0.6	-0.5	-0.7	-0.9	-0.8	-0.6	-1.0	-1.1	-1.6	-1.5	-1.6	-1.6	-1.7	-1.8	-2.0	-2.2	-2.8	-3.4	-1.40
	4	-1.0	-1.3	-1.8	-1.4	-1.0	-0.9	-0.6	-0.8	-0.5	-0.3	-0.6	-0.8	-0.6	-0.7	-0.8	-0.7	-0.8	-0.8	-1.0	-1.1	-1.5	-1.6	-2.0	-2.3	-1.04
9	500	-4.0	-3.7	-3.5	-3.2	-3.5	-3.5	-3.5	-3.1	-2.8	-2.7	-2.8	-2.9	-2.7	-2.5	-2.2	-2.6	-3.1	-3.1	-3.7	-3.9	-3.6	-4.1	-4.8	-4.3	-3.32
	100	-3.9	-3.5	-3.6	-3.2	-3.3	-3.1	-3.3	-2.8	-2.5	-2.5	-2.6	-2.4	-2.3	-2.2	-2.0	-2.4	-2.7	-3.0	-3.6	-3.9	-3.8	-4.1	-4.5	-4.6	-3.16
	4	-3.0	-2.7	-2.1	-2.1	-2.1	-2.3	-2.1	-2.0	-1.8	-1.6	-1.5	-1.6	-1.6	-1.6	-1.4	-1.7	-2.3	-2.0	-2.8	-3.7	-3.6	-3.9	-4.7	-5.1	-2.47

Table II. (Continued.)

Date	Height cm	Hour 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
July 10	500	−3.9	−3.6	−3.7	−3.8	−3.9	−4.0	−4.0	−3.1	−2.7	−2.5	−2.1	−2.0	−2.0	−2.1	−2.3	−2.1	−2.0	−2.0	−1.8	−2.0	−2.1	−2.2	−2.1	−2.2	−2.68
	100	−4.3	−3.7	−3.5	−3.6	−3.7	−3.8	−3.8	−3.3	−2.7	−2.6	−2.3	−1.9	−1.9	−2.2	−2.2	−1.9	−1.7	−1.6	−1.9	−2.0	−2.1	−2.1	−2.1	−2.2	−2.63
	4	−5.1	−3.8	−3.0	−2.9	−2.8	−2.7	−2.6	−2.8	−2.4	−2.4	−1.8	−1.5	−1.6	−1.2	−1.0	−0.9	−0.9	−1.1	−1.5	−1.8	−1.9	−1.8	−1.9	−2.1	−2.15
11	500	−2.4	−2.5	−2.8	−3.0	−3.3	−3.5	−3.4	−3.2	−2.8	−2.4	−2.5	−2.2	−1.5	−1.1	−1.1	−1.0	−1.0	−1.1	−1.5	−1.7	−1.6	−1.6	−1.2	−1.0	−2.06
	100	−2.3	−2.4	−2.7	−3.1	−3.4	−3.4	−3.2	−3.1	−2.8	−2.5	−2.2	−2.0	−1.4	−1.1	−1.1	−0.9	−0.9	−0.9	−1.3	−1.4	−1.4	−1.4	−1.0	−1.0	−1.95
	4	−2.1	−2.1	−2.3	−2.6	−2.8	−2.6	−2.5	−2.5	−2.5	−2.4	−2.2	−2.0	−1.5	−1.0	−1.0	−0.6	−0.5	−0.8	−1.1	−1.3	−1.4	−1.4	−1.0	−0.7	−1.70
12	500	−0.9	−0.6	−0.5	−0.5	−0.4	−0.2	0.0	0.1	0.1	0.2	0.5	0.8	1.0	1.1	1.0	1.0	0.9	0.6	0.7	0.7	0.6	0.6	0.6	0.6	0.33
	100	−0.9	−0.6	−0.4	−0.4	−0.3	−0.1	0.1	0.0	0.1	0.3	0.4	0.8	0.8	0.8	0.9	0.8	0.7	0.6	0.6	0.6	0.6	0.6	0.6	0.5	0.30
	4	−0.5	−0.3	−0.2	−0.3	−0.3	−0.2	0.0	0.0	0.0	0.1	0.2	0.4	0.5	0.5	0.5	0.5	0.4	0.4	0.4	0.4	0.4	0.4	0.4	0.4	0.17
13	500	0.8	0.7	0.7	0.8	0.7	0.8	1.1	1.3	1.4	1.6	1.9	1.8	2.0	2.1	1.8	2.3	2.5	2.4	2.0	1.8	0.9	1.6	1.6	2.7	1.55
	100	0.6	0.6	0.6	0.8	0.7	0.6	0.8	1.1	1.4	1.8	1.7	1.2	1.2	1.3	1.3	1.6	2.1	1.7	1.5	1.5	0.7	0.9	0.8	1.9	1.17
	4	0.4	0.4	0.4	0.4	0.3	0.4	0.5	0.7	0.7	0.8	0.8	0.6	0.8	0.9	0.6	0.9	1.0	0.9	0.8	0.9	0.5	0.4	0.5	0.8	0.64
14	500	4.3	3.1	1.9	2.6	2.4	2.5	2.9	3.3	2.7	2.0	2.0	1.6	2.2	2.0	0.6	1.3	2.5	3.7	3.4	3.8	3.0	2.8	2.6	4.5	2.65
	100	3.4	2.2	1.3	1.9	1.9	1.9	2.1	2.3	2.0	1.7	2.0	1.6	1.5	1.3	0.5	1.0	1.8	2.5	1.9	1.8	0.8	1.3	1.0	0.3	1.69
	4	1.9	1.4	1.1	1.1	1.1	1.2	1.4	1.6	1.5	1.1	0.9	0.9	1.2	0.9	0.4	0.5	0.7	0.9	0.9	0.4	0.1	0.1	0.0	−0.4	0.87
15	500	4.5	2.0	2.6	2.4	4.2	5.3	3.7	2.8	3.1	3.6	3.6	3.8	4.1	3.9	3.4	3.5	2.9	3.4	3.4	1.9	2.6	3.8	3.3	1.4	3.30
	100	0.6	1.0	2.0	2.0	1.2	2.2	2.1	1.6	1.7	1.9	2.4	2.7	2.7	2.0	2.1	3.0	2.5	1.5	0.9	0.3	0.9	1.2	1.0	−0.7	1.62
	4	−0.3	0.2	1.0	0.8	0.7	0.7	0.5	0.4	0.5	0.6	1.0	1.2	1.3	1.0	0.4	0.9	0.8	0.2	−0.1	−0.6	−0.5	−0.4	−0.8	−2.2	0.30
16	500	0.1	−0.3	0.3	1.0	1.4	1.5	−0.3	0.1	−0.1	−0.5	0.1	0.7	0.4	−0.2	−0.1	−0.2	−0.1	0.3	0.3	1.1	1.7	0.8	0.5	1.7	0.42
	100	−1.9	−2.6	−1.7	−1.4	−0.3	0.6	−0.8	−0.1	−0.2	−0.3	−0.2	0.5	0.4	−0.3	−0.6	−0.3	0.0	0.2	−0.4	−0.7	−0.9	−1.6	−0.7	0.0	−0.55
	4	−2.8	−3.1	−2.6	−1.7	−1.2	−0.1	−0.7	−0.3	−0.2	−0.1	−0.1	0.1	0.2	−0.3	−0.5	−0.3	−0.1	−0.1	−0.7	−1.2	−1.5	−2.4	−1.6	−0.3	−0.90
17	500	1.9	1.0	−0.1	−0.4	0.0	0.2	0.6	0.8	0.5	0.0	0.3	0.4	0.5	1.3	1.3	1.2	1.0	1.1	1.0	1.0	1.0	0.8	1.1	1.2	0.74
	100	0.8	0.9	−0.1	−0.1	0.0	0.3	0.4	0.8	0.5	0.1	0.4	0.5	0.6	1.1	1.2	1.0	0.8	0.9	0.9	0.9	0.9	1.0	1.1	1.0	0.65
	4	0.3	0.5	−0.2	−0.2	0.1	0.2	0.3	0.4	0.2	0.1	0.2	0.3	0.5	0.6	0.7	0.7	0.6	0.6	0.6	0.6	0.6	0.6	0.6	0.5	0.39
18	500	1.1	1.0	1.1	1.4	1.2	1.0	1.0	1.3	1.4	1.6	1.2	1.3	1.6	1.5	1.4	1.7	1.8	1.4	1.3	1.2	1.3	1.3	0.9	1.2	1.30
	100	0.9	0.9	1.0	1.1	1.0	0.8	1.0	1.2	1.3	1.1	0.9	1.1	1.4	1.3	1.2	1.3	1.3	1.1	1.1	0.9	1.0	1.0	0.9	1.1	1.08
	4	0.6	0.6	0.7	0.8	0.7	0.6	0.7	0.8	0.9	1.0	0.8	0.8	1.0	0.7	0.7	0.8	1.1	0.9	0.7	0.7	0.8	0.7	0.8	0.8	0.78
19	500	1.5	1.5	1.3	1.4	1.7	1.7	1.7	2.5	2.3	1.8	1.5	1.7	1.8	1.4	1.6	1.2	1.1	1.2	1.0	0.8	0.7	1.0	1.1	1.0	1.44
	100	1.2	1.2	1.1	1.2	1.3	1.4	1.3	1.9	1.7	1.2	1.2	1.3	1.1	0.9	1.2	1.0	0.9	0.8	0.7	0.7	0.6	0.8	0.9	0.8	1.10
	4	0.8	0.9	0.8	0.9	1.1	1.0	1.0	1.2	1.3	0.9	0.7	1.0	1.0	0.8	1.0	0.6	0.6	0.6	0.6	0.5	0.4	0.7	0.7	0.6	0.82
20	500	0.9	0.9	0.7	0.6	0.8	0.8	1.1	1.5	1.7	1.5	1.0	0.9	0.9	0.7	0.8	1.0	1.2	1.4	1.3	1.1	1.1	1.0	1.0	0.8	1.03
	100	0.9	0.8	0.6	0.5	0.6	0.5	0.9	1.2	1.3	1.2	0.7	0.7	0.9	0.7	0.8	0.8	1.0	1.2	1.1	1.0	0.9	0.9	0.8	0.7	0.86
	4	0.6	0.5	0.4	0.4	0.5	0.4	0.7	1.0	0.9	0.6	0.3	0.4	0.5	0.4	0.6	0.6	0.8	0.8	0.9	0.7	0.7	0.7	0.7	0.6	0.61
21	500	0.8	0.7	0.8	1.1	1.1	1.3	1.4	1.1	1.6	2.0	2.2	2.2	2.2	2.4	2.0	1.7	1.4	0.8	0.2	−0.3	−1.0	−1.6	−2.0	−2.1	0.83
	100	0.8	0.7	0.8	1.0	1.0	1.2	1.3	1.2	1.7	1.8	2.1	2.1	1.9	1.6	1.6	1.4	1.2	0.6	0.2	−0.2	−0.9	−1.5	−1.9	−2.1	0.73
	4	0.7	0.6	0.7	0.8	0.8	0.8	0.9	0.9	1.1	1.4	1.6	1.4	1.2	1.0	1.0	1.1	0.9	0.5	0.2	−0.1	−0.5	−1.1	−1.8	−2.1	0.50
22	500	−2.4	−3.7	−2.9	−2.1	−2.2	−2.3	−1.7	−1.2	−0.8	−0.5	−0.4	−0.1	−0.2	−0.1	−0.6	−0.5	−1.2	−2.0	−2.5	−2.4	−2.5	−2.0	−1.7	−1.8	−1.58
	100	−2.4	−3.3	−2.9	−2.6	−2.4	−2.0	−1.5	−1.0	−0.6	−0.4	−0.2	0.1	0.1	−0.3	−0.6	−0.5	−0.8	−1.7	−2.2	−2.2	−2.2	−2.0	−1.7	−1.6	−1.45
	4	−2.4	−3.1	−3.0	−2.1	−2.1	−1.9	−1.0	−0.6	−0.3	−0.2	−0.1	0.2	0.1	−0.1	−0.3	−0.2	−0.6	−1.1	−1.4	−1.6	−1.7	−1.8	−1.3	−1.0	−1.15

Table II. (Continued.)

Date	Height cm	Hour 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
July 23	500	-1.8	-1.9	-2.3	-1.9	-1.7	-1.9	-1.4	-1.5	-1.5	-1.7	-1.7	-1.6	-1.7	-1.4	-1.6	-1.7	-1.8	-1.8	-1.9	-2.2	-2.4	-1.6	-1.2	-1.1	-1.72
	100	-1.7	-1.7	-2.2	-2.1	-1.8	-1.8	-1.6	-1.5	-1.5	-1.6	-1.4	-1.5	-1.6	-1.5	-1.6	-1.6	-1.7	-1.7	-1.8	-2.2	-2.4	-1.5	-1.1	-0.9	-1.67
	4	-1.2	-1.3	-1.9	-1.3	-0.9	-1.0	-0.8	-0.5	-0.4	-0.3	-0.7	-0.8	-0.7	-0.7	-0.8	-0.7	-1.0	-1.1	-1.2	-1.7	-2.4	-1.7	-1.0	-0.8	-1.04
24	500	-0.8	-1.1	-1.4	-1.3	-1.5	-1.4	-1.3	-1.2	-0.9	-1.0	-1.1	-0.6	-0.5	-0.3	-0.6	-0.8	-0.6	-0.5	-0.7	-0.6	-0.5	-0.3	0.0	0.2	-0.78
	100	-0.7	-1.1	-1.3	-1.2	-1.4	-1.5	-1.4	-1.3	-1.1	-1.0	-1.4	-0.6	-0.5	-0.5	-0.6	-0.6	-0.6	-0.6	-0.7	-0.6	-0.4	-0.3	-0.1	0.1	-0.81
	4	-0.5	-0.9	-1.0	-0.9	-1.1	-1.0	-0.9	-0.7	-0.8	-0.8	-0.9	-0.4	-0.3	-0.2	-0.2	-0.6	-0.7	-0.8	-0.8	-0.7	-0.4	-0.4	-0.3	0.0	-0.64
25	500	0.2	0.2	0.2	0.0	0.6	0.2	0.0	0.9	1.0	1.1	1.1	0.9	0.9	0.8	0.5	0.8	0.7	0.8	1.0	0.8	0.9	0.6	0.6	0.5	0.64
	100	0.2	0.2	0.2	0.0	0.6	0.1	0.3	0.7	0.8	1.1	1.1	0.7	0.6	0.6	0.5	0.6	0.9	0.8	0.9	0.6	0.7	0.5	0.6	0.5	0.58
	4	0.0	0.0	0.0	0.0	0.2	0.0	0.1	0.4	0.6	0.8	0.7	0.5	0.5	0.5	0.4	0.4	0.5	0.5	0.5	0.4	0.5	0.5	0.5	0.3	0.37
26	500	0.7	0.8	0.9	0.7	0.4	0.1	-0.3	-0.8	-0.5	0.3	1.0	1.4	1.0	0.7	1.4	1.9	2.0	1.1	0.2	0.4	1.0	0.2	0.1	0.2	0.62
	100	0.7	0.8	0.9	0.7	0.3	0.2	-0.3	-0.7	-0.4	-0.1	0.2	1.1	0.4	0.2	0.9	1.2	1.3	0.7	0.0	-0.1	-0.6	-0.9	-1.5	-2.2	0.12
	4	0.4	0.6	0.6	0.5	0.3	0.2	-0.1	-0.4	-0.3	0.0	0.1	0.6	-0.2	-0.3	0.2	0.3	0.3	-0.3	-0.7	-0.8	-1.2	-1.5	-2.1	-2.8	-0.28
27	500	-0.4	-0.8	-1.1	-0.2	-0.3	-0.3	0.8	1.3	1.7	2.1	1.9	2.0	2.0	1.7	1.9	1.6	1.2	1.4	1.5	1.3	1.1	1.0	0.7	0.0	0.92
	100	-2.3	-2.5	-2.5	-2.4	-1.6	-0.9	-0.3	0.7	1.3	2.0	1.4	1.4	1.7	1.5	1.6	1.4	1.1	1.1	1.3	1.0	1.0	0.8	0.4	-0.2	0.29
	4	-3.0	-3.3	-3.1	-2.7	-1.9	-1.2	-0.5	0.1	0.4	0.4	0.3	0.6	0.7	0.6	0.6	0.4	0.4	0.7	0.7	0.6	0.7	0.5	-0.1	-0.7	-0.37
28	500	-0.3	-1.1	-1.5	-1.3	-0.5	-0.2	-1.9	-2.4	-2.0	-1.2	-1.2	-0.6	-0.6	-0.9	-0.6	-0.4	-0.4	-0.5	-0.3	-0.1	0.0	0.1	-0.1	-0.2	-0.76
	100	-0.6	-1.2	-1.5	-1.2	-1.2	-0.8	-1.7	-1.9	-1.6	-1.2	-1.1	-0.7	-0.8	-0.9	-0.7	-0.3	-0.4	-0.4	-0.2	-0.2	-0.1	0.1	-0.1	-0.1	-0.78
	4	-1.1	-1.6	-2.0	-1.9	-1.7	-1.4	-1.5	-1.6	-1.4	-1.2	-1.2	-1.2	-1.2	-1.2	-0.8	-0.3	-0.3	-0.4	-0.3	-0.2	-0.2	-0.1	-0.2	-0.2	-0.97
29	500	0.0	-0.7	-1.8	-1.1	-0.3	0.9	1.0	1.3	1.5	1.8	1.8	1.2	1.5	1.7	1.2	1.2	1.1	0.9	1.0	1.2	1.5	1.2	0.6	0.6	0.80
	100	-0.2	-2.4	-3.2	-1.2	-0.3	0.8	0.9	1.2	1.2	1.6	1.7	1.1	1.1	1.6	1.1	1.0	0.5	0.2	0.8	0.5	0.6	0.6	0.3	0.1	0.40
	4	-0.5	-2.8	-3.4	-1.3	-0.2	0.4	0.5	0.5	0.6	0.6	0.8	0.6	1.7	1.0	0.8	0.5	0.1	0.0	0.1	-0.3	-0.1	-0.1	-0.3	-0.3	-0.09
30	500	0.4	-0.3	-1.0	-0.9	0.4	0.8	0.6	0.8	1.1	1.2	1.6	1.6	1.5	1.6	1.0	1.2	1.4	1.0	0.8	1.0	1.0	1.0	0.9	1.1	0.82
	100	-0.1	-0.9	-1.1	-1.3	-0.6	-0.2	0.5	0.4	0.6	1.0	1.2	1.2	1.3	0.9	0.8	0.9	0.7	0.3	0.0	-0.8	0.1	0.0	-1.0	-1.4	0.10
	4	-0.5	-1.5	-1.6	-1.7	-1.4	-0.9	-0.4	-0.4	-0.5	-0.4	-0.2	0.1	0.4	0.3	-0.1	-0.4	-0.7	-1.4	-1.9	-2.1	-2.4	-2.8	-2.6	-2.3	-1.06
31	500	1.5	1.4	0.2	0.5	1.4	1.8	2.0	2.5	3.4	2.1	1.6	2.3	2.3	1.4	1.2	1.2	1.0	1.1	1.2	1.2	1.6	1.9	1.8	1.8	1.60
	100	-0.8	-1.1	-3.2	-2.1	-0.7	-1.5	-0.2	2.1	2.5	1.5	1.7	1.8	1.9	1.2	1.0	1.0	0.9	0.9	0.9	0.7	0.7	1.4	1.5	1.5	0.57
	4	-3.1	-4.2	-4.7	-3.6	-2.3	-2.9	-1.0	0.3	0.4	0.5	0.5	0.5	0.4	0.4	0.5	0.5	0.5	0.5	0.4	0.3	0.5	0.8	0.8	0.8	-0.55
Aug. 1	500	1.6	1.8	2.1	2.0	1.5	2.0	1.8	1.8	2.2	2.2	1.8	1.7	1.7	1.8	2.0	2.3	2.3	2.2	2.5	2.4	1.6	1.7	1.9	1.8	1.95
	100	1.0	1.2	1.4	1.3	1.2	1.0	1.5	1.6	1.1	1.4	1.5	1.2	1.4	1.5	1.3	1.3	1.7	1.6	1.8	1.9	1.4	1.1	1.4	1.4	1.38
	4	0.6	0.7	0.8	0.7	0.6	0.4	0.7	0.9	0.8	0.8	0.9	0.9	0.8	0.8	0.9	1.1	1.1	1.0	1.2	1.3	0.9	0.9	0.9	0.9	0.86
2	500	1.9	3.2	4.3	3.8	3.8	4.1	4.1	5.1	3.9	3.0	2.7	3.2	3.2	2.9	2.8	2.9	3.3	3.3	3.5	3.3	2.5	3.1	3.7	4.3	3.33
	100	1.6	2.4	3.7	3.1	3.0	2.9	3.3	3.8	3.0	2.5	2.6	2.9	2.9	2.6	2.6	2.6	2.8	2.7	2.7	2.6	2.4	2.6	2.9	3.8	2.84
	4	1.1	1.6	2.0	2.3	2.2	1.9	1.9	2.0	1.7	1.5	1.6	1.9	1.8	1.6	1.6	1.7	1.8	1.6	1.8	1.9	1.7	1.8	1.9	2.5	1.81
3	500	4.4	3.1	2.8	2.7	2.8	3.0	3.2	3.1	3.1	2.8	1.9	2.2	3.2	3.1	3.0	3.2	3.0	3.0	3.6	2.7	2.0	2.3	2.3	2.3	2.87
	100	3.3	2.6	2.2	2.4	2.1	2.5	2.1	2.1	1.8	1.7	1.3	1.4	2.2	1.8	1.8	2.2	2.3	2.6	2.6	2.3	1.5	1.4	1.4	1.9	2.06
	4	2.5	1.6	1.2	1.1	1.0	1.0	1.0	1.0	1.0	0.8	0.6	0.8	1.1	1.1	0.9	1.2	1.3	1.4	1.7	1.1	0.9	1.1	1.0	1.0	1.14
4	500	2.0	1.6	1.9	1.7	1.4	1.8	2.4	2.0	2.2	2.4	2.6	2.2	1.8	2.4	5.1	6.2	5.6	5.0	3.5	4.5	4.3	4.8	5.1	4.7	3.22
	100	1.8	1.3	1.4	1.2	1.1	1.6	2.1	1.7	1.7	2.2	1.9	1.5	1.6	1.9	3.2	3.7	4.4	3.4	2.2	3.0	3.6	4.2	3.9	3.8	2.43
	4	1.0	0.8	0.9	0.9	0.7	0.9	1.2	1.1	1.1	1.3	1.3	0.9	0.8	1.1	1.9	2.2	1.5	1.5	1.8	2.0	2.2	2.4	2.4	2.2	1.42

Table II. (Continued.)

Date	Height cm	Hour 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
Aug. 5	500	4.2	4.4	4.8	4.6	4.5	4.9	4.9	5.3	6.2	6.4	5.9	6.4	6.9	60	6.2	7.0	6.7	5.3	4.5	5.3	5.7	5.3	5.6	4.6	5.48
	100	3.4	3.2	3.4	3.2	3.4	3.8	4.6	4.7	4.5	5.1	4.7	5.2	4.8	34	2.8	4.3	3.2	3.0	3.6	3.7	3.5	3.4	3.3	2.6	3.79
	4	1.7	1.3	1.5	1.9	2.0	2.1	2.5	3.1	3.2	2.9	1.9	2.2	2.5	16	1.1	1.4	1.3	1.6	1.6	1.2	0.9	1.0	1.1	1.3	1.79
6	500	4.2	4.9	5.5	5.3	4.8	6.9	8.3	6.9	5.0	5.0	5 0	4.6	5.0	6.1	7.0	6.5	5.5	5.1	5.3	5.3	5.1	5.3	6.3	5.5	5.60
	100	3.2	2.2	2.4	2.8	1.8	4.1	7.1	5.8	2.6	2.6	2.8	3.5	2.6	2.4	3.8	2.7	2.5	2.5	2.3	2.1	2.9	2.1	2.5	1.3	2.94
	4	1.9	1.0	0.9	1.0	0.5	1.5	4.0	3.5	1.5	1.2	1.5	1.6	1.5	1.5	2.1	1.7	1.5	1.6	1.9	2.0	2.0	1.4	1.4	0.8	1.65
7	500	5.3	5.3	5.7	6.1	5.4	5.0	6.4	6.8	6.4	6.5	7.0	6.7	5.1	4.1	4.2	4.5	4.5	4.0	3 5	4.8	5.5	5.2	3.4	4.1	5.23
	100	2.3	3.7	3.9	3.7	4.4	4.0	3.9	3.8	3.0	2.5	2.6	3.0	3.0	2.7	3.0	3.4	3.6	3.4	3.0	3.3	3.8	3.6	2.7	3.1	3.26
	4	0.9	1 5	1.8	1.3	1.3	2.2	2.4	2.4	2.3	2.0	2.0	2.0	1.8	1.4	1.7	2.1	2.3	2.3	2.0	1.7	1.6	1.5	1.4	1.5	1.81
8	500	3.6	3.4	3.6	3.7	3.2	3.6	4.5	3.8	3.6	3.5	3.6	3.5	3.4	3.5	3.5	2.9	2 2	2.0	2 5	2.1	2.6	2.1	1.7	2.7	3.12
	100	2.9	2.7	3.0	2.8	2.7	3.0	3.6	3.3	2.8	2.4	2.3	2.5	1.9	1.4	1.7	1.6	1.2	0.8	1.0	1.1	1.3	1.1	1.4	1.4	2.08
	4	1.4	0.9	1.0	1.4	1.5	1.3	1.3	1.0	1.0	1.2	1.2	1.0	0.9	0.8	0.8	0.7	0.6	0.4	0.4	0.5	0.6	0.5	0.8	0.8	0.92
9	500	2.8	2.0	2.0	2.1	2.3	1.8	1.7	1.7	1.8	1.5	2.0	1.8	1.9	2.0	1.6	1.2	1.2	1.0	1.0	1.0	0.7	0.6	0.6	0.9	1.55
	100	1.4	1.3	1.6	1.3	1.6	1.3	1.2	1.1	1.6	1.3	1.5	1.2	0.9	1.2	1.3	1.2	1.1	0.8	0.7	0.6	0.5	0.6	0.5	0.4	1.09
	4	0.7	0.7	0.8	0.9	1.0	0.9	0.8	0.7	0.8	0.7	0.7	0.5	0.3	0 6	0.6	0.6	0.5	0.4	0.4	0.3	0.2	0.2	0.2	0.2	0.57
10	500	0.6	0.7	0.9	0.9	1.0	1.0	1.0	1.4	1.5	1.5	1.3	1.4	1.5	1.5	1.5	1.7	1.7	2.1	2.0	1.4	0.8	0.2	0.4	0.3	1.18
	100	0.4	0.4	0.8	0.8	0.9	0.9	1.0	1.1	1.1	1.3	1.3	1 2	1.3	1.4	1.5	1.7	1.6	1.6	1.5	0.7	0.4	−0.2	−0.1	−0.4	0.93
	4	0.2	0.2	0.5	0.5	0.5	0.6	0.7	0.7	0.7	0.9	0.9	0 8	0.8	0.8	0.9	1.1	1.2	1.2	1.0	0.0	−0.4	−0.6	−0.6	−0.8	0.49
11	500	0.2	0.6	0.4	−0.3	−0.7	−0.3	0.7	1.3	1.6	2.1	2.4	2.2	3.0	3.0	2.7	3.2	3.0	2.7	3.0	2.7	3.0	2.7	2.8	2.7	1.86
	100	−0.4	0.4	0.0	−1.0	−1.3	−0.7	−0 6	0.3	1.3	1.2	1.8	2.1	2.0	2.0	2.1	2.3	1.9	1.6	1.7	1.7	1.8	1.9	2.2	2.0	1.10
	4	−1.1	−0.7	−0 7	−1.6	−1.8	−1.1	−0.9	−0.3	0.2	0.5	0.6	0.8	0.9	1.0	1.2	1.5	1.1	0.7	0.6	0.7	0.6	0.6	0.6	0.5	0.16
12	500	2.9	2.8	2 5	2.1	1.9	2.1	2.0	2.2	2.4	2.6	2.2	2.7	2 6	2.6	2.0	2.5	3.1	2.3	2.4	2.7	3.1	2.5	3.0	4.2	2.56
	100	1.7	1.5	1.5	1.1	0.8	1.2	1.6	1.7	1.3	1.5	1.6	1.5	1.5	1.4	1.3	1.3	1.7	1.5	1.6	1.6	1.5	1.0	1.8	1.9	1.46
	4	0.4	0.5	0.5	0.6	0.6	0.5	0.5	0.6	0.7	0.7	0.7	0.7	0.7	0.7	0.7	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.7	0.68
13	500	3.2	3.2	3.1	2.4	2.5	2.6	2.7	3.2	2.5	2.5	3.3	3.1	3.1	3.1	2.5	3.0	4.4	4.3	3.7	4.9	4.1	2.2	3.2	3.0	3.16
	100	0.9	0.7	1.3	1.2	1.1	2.0	2.0	1.8	1.6	1.4	1.7	1.4	1.7	1.8	1.7	1.8	2.0	1.9	1.7	1.6	1.4	1.2	1.1	1.0	1.50
	4	0 6	0.4	0.4	0.1	0.0	0.6	1.1	1.1	1.0	0.8	0.8	0.8	0.8	0.8	1.1	1.2	1 2	1.3	1.2	0.9	0.7	0.5	0.3	0.2	0.75
14	500	3.5	4.6	3.9	3.1	3.1	2.4	2.7	3.6	3.3	3.8	3.7	3.3	3.1	3.3	3.6	2.7	1.3	1.6	1.5	1.3	0.8	1.0	0.8	0.7	2.61
	100	1.3	1.2	1.3	2.0	1.9	1.4	2.2	2.3	2.0	2.3	2.7	2.5	2.3	2.0	1.9	1.1	0.5	1.0	1.3	1.3	0.7	0.9	0.7	0.6	1.56
	4	0.6	0.2	0 5	1.3	1.2	1.3	1.4	1.4	1.2	1.4	1.7	1.4	0.6	0.5	0.5	0.6	0.4	0.6	0.7	0.7	0 5	0.6	0.6	0.5	0.85
15	500	0.6	0.5	0.4	0.1	0.0	0.1	0.3	0.8	1.1	1.2	1.0	0.5	0.8	0.8	0.6	0 8	0.7	0.6	0.6	0.7	0.6	0.5	0.9	0.4	0.61
	100	0.6	0.5	0.4	0.2	0.0	0.1	0.2	0.7	1.1	1.1	1.0	0.8	0.9	1.2	0.9	0.8	0.7	0.5	0.5	0.6	0.5	0.6	0.8	0.4	0.63
	4	0.4	0.3	0.2	0.1	0.1	0.1	0.3	0.3	0.7	0.6	0.4	0.4	0.5	0.4	0.4	0.4	0.5	0.4	0.3	0 3	0.3	0.2	0.6	0.1	0.35

Table III. *Vapour Pressure. Mean Hourly Values of Vapour Pressure in mm Hg.*

Date	Height cm	Hour 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
June 26	500	3.4	3.3	3.0	2.8	2.9	3.1	3.2	3.0	3.1	3.4	3.6	3.5	3.3	3.0	2.7	2.9	2.9	3.1	3.6	3.8	(3.9)	(3.9)	(4.0)	4.0	3.31
	100	3.5	3.3	3.1	3.0	3.3	3.4	3.4	3.5	3.5	3.4	3.5	3.6	3.6	3.8	3.9	4.0	3.9	3.8	3.9	4.1	4.1	4.0	3.8	3.7	3.63
27	500	4.2	4.2	4.1	-	-	-	-	-	-	3.8	4.0	4.1	4.3	4.2	4.1	-	-	-	4.0	-	-	-	-	-	-
	100	3.6	3.6	3.6	3.8	3.8	3.7	3.8	4.0	4.1	4.0	4.1	4.2	4.1	3.9	3.9	3.8	3.8	3.7	3.6	3.5	3.5	3.4	3.3	3.2	3.75
28	500	3.0	-	-	-	-	-	-	-	-	-	-	-	-	3.7	-	-	-	-	-	-	-	-	-	-	-
	100	3.2	3.3	3.4	3.5	3.6	3.7	3.7	3.7	3.9	3.8	3.7	3.7	3.6	3.7	3.8	3.7	3.6	3.5	3.4	3.3	3.2	3.1	3.0	2.9	3.50
29	500	-	-	-	-	-	-	-	-	-	-	-	3.0	3.2	3.4	3.4	3.4	3.3	3.1	3.0	2.8	2.7	2.7	2.9	2.9	-
	100	2.8	2.8	2.8	2.9	2.9	3.0	3.0	3.0	3.0	3.1	3.2	3.3	3.3	3.4	3.5	3.6	3.5	3.4	3.1	3.0	3.1	3.1	3.2	3.1	3.13
30	500	2.8	2.6	2.6	3.4	3.5	3.7	-	-	-	-	-	3.5	3.7	3.8	4.0	4.0	3.9	3.8	3.6	3.5	3.6	3.6	3.8	3.8	-
	100	3.2	3.3	3.3	3.5	3.6	3.7	3.7	3.6	3.6	3.8	3.9	3.9	3.9	3.9	4.0	4.0	4.0	3.9	3.7	3.8	4.0	3.9	3.9	3.9	3.75
July 1	500	3.7	3.8	3.8	3.8	3.8	3.7	3.8	4.0	3.9	3.8	3.9	4.0	3.9	4.0	4.1	4.2	4.3	4.2	4.2	4.3	4.3	4.3	4.2	4.2	4.01
	100	3.9	3.9	3.9	3.9	3.9	3.8	3.7	3.9	3.9	3.9	3.9	3.9	4.0	4.0	4.2	4.3	4.3	4.3	4.2	4.3	4.3	4.2	4.3	4.3	4.05
	4	4.2	4.1	4.1	4.3	4.3	4.2	4.2	4.3	4.4	4.4	4.2	4.2	4.2	4.4	4.5	4.4	4.4	4.5	4.5	4.5	4.4	4.4	4.4	4.4	4.33
2	500	4.2	4.1	4.1	4.2	4.2	4.2	4.3	4.3	(4.3)	(4.3)	(4.2)	4.0	4.3	4.3	4.3	4.3	4.4	4.3	4.2	3.8	3.7	3.8	3.9	4.0	4.15
	100	4.3	4.2	4.2	4.2	4.3	4.2	4.2	4.4	4.4	4.4	4.3	4.1	4.2	4.3	4.3	4.4	4.4	4.3	4.2	3.9	3.7	3.7	3.9	4.0	4.19
	4	4.4	4.4	4.4	4.4	4.4	4.4	4.4	4.5	4.5	4.6	4.5	4.5	4.5	4.5	4.5	4.5	4.5	4.4	4.4	4.2	4.1	4.0	4.1	4.2	4.39
3	500	4.0	4.0	4.1	4.1	4.2	4.3	4.2	4.0	4.0	4.0	4.1	4.1	4.1	4.0	3.4	2.8	3.1	3.8	3.9	3.8	3.7	3.6	3.8	4.1	3.88
	100	4.1	3.9	4.0	4.2	4.2	4.3	4.2	4.1	4.0	4.1	4.3	4.3	4.2	4.1	4.0	3.8	3.8	4.0	4.1	4.0	4.0	3.9	3.9	4.1	4.07
	4	4.0	3.8	4.0	4.2	4.3	4.4	4.3	4.3	4.3	4.4	4.6	4.6	4.5	4.4	4.3	4.2	4.2	4.1	4.2	4.3	4.1	3.9	3.8	4.0	4.22
4	500	4.4	4.4	-	-	-	-	-	-	-	-	3.6	3.4	3.4	-	-	-	3.7	3.8	3.8	3.9	3.8	3.8	3.7	3.7	-
	100	4.1	4.5	4.5	4.5	4.5	4.5	4.4	4.3	4.3	4.3	4.3	3.9	3.6	3.6	3.9	4.1	4.0	4.0	4.0	4.0	3.9	3.8	3.8	3.8	4.11
	4	4.3	4.5	-	-	-	-	-	-	-	-	-	4.4	4.4	4.3	4.2	4.2	4.1	4.1	4.2	4.1	4.1	4.0	3.9	3.9	-
5	500	3.7	3.6	3.6	3.4	3.4	3.5	3.5	3.5	3.4	3.3	3.3	3.4	3.6	3.6	3.7	3.6	3.5	3.5	3.5	3.4	3.3	3.2	3.3	3.3	3.46
	100	3.7	3.7	3.7	3.6	3.5	3.6	3.6	3.5	3.5	3.5	3.4	3.5	3.6	3.7	3.7	3.8	3.7	3.5	3.5	3.3	3.3	3.2	3.0	3.0	3.50
	4	3.8	3.8	3.8	3.9	3.8	3.8	3.8	3.8	3.8	3.9	4.0	(4.0)	(4.0)	(4.1)	4.1	4.1	3.9	3.8	3.7	3.6	3.5	3.2	3.0	2.9	3.75
6	500	3.2	3.2	3.2	3.3	3.5	3.6	3.8	3.9	3.9	3.8	3.8	3.9	3.9	3.7	3.6	3.7	3.8	3.7	3.7	3.7	3.8	3.7	3.7	3.8	3.66
	100	3.1	3.2	3.4	3.6	3.6	3.7	3.8	4.1	4.3	4.2	4.2	4.2	4.1	3.8	3.6	3.7	3.8	3.7	3.8	3.8	3.9	3.9	3.7	3.7	3.79
	4	3.0	3.4	3.5	3.6	3.7	3.7	3.9	4.2	4.3	4.4	4.5	4.5	4.6	4.5	4.4	4.3	4.4	4.3	4.3	4.2	4.2	4.3	4.2	4.1	4.10
7	500	3.7	3.6	3.6	3.5	3.5	3.4	3.4	3.5	3.5	3.5	3.6	3.6	3.6	3.6	3.5	3.4	3.5	3.5	3.4	3.5	3.6	3.5	3.7	3.9	3.55
	100	3.8	3.7	3.7	3.6	3.6	3.6	3.5	3.6	3.6	3.5	3.6	3.6	3.5	3.6	3.5	3.5	3.6	3.6	3.6	3.6	3.6	3.6	3.7	3.9	3.61
	4	4.1	4.1	3.9	3.9	3.9	3.9	3.8	3.8	3.9	3.9	4.0	4.0	4.1	4.1	4.0	4.0	4.0	3.9	3.9	3.8	3.7	3.6	3.7	3.9	3.91
8	500	4.0	4.0	3.9	3.9	3.8	3.6	3.4	3.6	3.7	3.6	3.6	3.7	3.8	3.8	3.9	3.9	3.9	3.9	3.9	3.7	3.6	3.6	3.6	3.5	3.75
	100	4.1	4.0	3.8	3.9	4.0	3.8	3.7	3.8	3.7	3.6	3.7	3.8	3.9	3.9	3.9	3.9	3.9	3.9	3.9	3.8	3.7	3.8	3.7	3.5	3.82
	4	4.1	4.0	4.0	4.1	4.3	4.3	4.2	4.2	4.3	4.3	4.2	4.3	4.4	4.3	4.3	4.2	4.2	4.3	4.3	4.2	4.1	4.1	4.0	3.8	4.19
9	500	3.4	3.4	(3.5)	(3.5)	(3.5)	(3.5)	(3.5)	(3.6)	(3.7)	3.8	3.7	3.7	3.7	3.7	3.7	3.7	3.6	3.4	3.3	3.2	3.2	3.1	3.1	3.2	(3.48)
	100	3.4	3.5	3.5	3.5	3.5	3.5	3.5	3.6	3.7	3.8	3.9	3.9	4.0	3.9	3.8	3.7	3.6	3.4	3.3	3.2	3.3	3.2	3.1	3.2	3.54
	4	3.7	3.8	3.9	3.8	3.7	3.7	3.9	4.0	4.0	4.0	4.1	4.1	4.1	4.1	4.1	4.0	3.8	3.8	3.6	3.4	3.4	3.3	3.1	3.1	3.77

Table III. (Continued.)

Date	Height cm	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
July 10	500	3.2	3.4	3.4	3.4	3.3	3.2	3.2	3.3	3.3	3.3	3.2	3.1	3.1	3.2	3.2	3.3	3.4	3.3	3.2	3.2	3.2	3.2	3.2	3.2	3.25
	100	3.2	3.4	3.4	3.4	3.4	3.3	3.3	3.4	3.5	3.4	3.3	3.3	3.3	3.2	3.3	3.4	3.4	3.4	3.3	3.2	3.3	3.4	3.3	3.3	3.34
	4	3.2	3.5	3.6	3.6	3.6	3.7	3.7	3.6	3.7	3.7	3.8	3.7	3.7	3.9	4.0	4.0	4.0	3.9	3.8	3.7	3.7	3.7	3.7	3.6	3.71
11	500	3.2	3.2	3.2	3.2	3.3	3.3	3.2	3.2	3.3	3.4	3.5	3.6	3.8	3.9	(4.0)	(4.0)	(4.0)	(4.0)	(4.0)	(3.9)	3.9	4.0	4.1	4.1	(3.63)
	100	3.3	3.3	3.3	3.3	3.3	3.4	3.4	3.3	3.4	3.5	3.6	3.7	4.9	4.0	4.2	4.3	4.3	4.3	4.2	4.1	4.1	4.2	4.3	4.3	3.79
	4	3.6	3.7	3.6	3.6	3.6	3.7	3.6	3.7	3.7	3.8	3.8	3.9	4.1	4.2	4.3	4.4	4.4	(4.4)	(4.3)	(4.3)	(4.3)	4.3	4.4	4.4	(4.00)
12	500	(4.2)	(4.3)	(4.4)	(4.4)	(4.5)	4.5	4.5	4.4	4.4	4.4	4.4	4.4	4.6	4.7	4.7	4.7	4.7	4.8	4.7	4.7	4.7	4.7	4.8	4.8	(4.56)
	100	4.4	4.4	4.5	4.4	4.5	4.5	4.4	4.4	4.5	4.4	4.4	4.5	4.5	4.7	4.7	4.7	4.7	4.8	4.8	4.7	4.7	4.8	4.8	4.8	4.58
	4	4.4	4.5	4.5	(4.5)	(4.5)	(4.5)	(4.5)	(4.4)	(4.4)	(4.4)	(4.4)	(4.4)	(4.5)	(4.6)	(4.7)	4.7	4.7	4.7	4.7	4.7	4.7	4.6	4.6	4.6	(4.55)
13	500	4.9	4.8	4.8	4.7	4.8	4.8	4.9	5.0	5.1	5.2	5.1	5.2	5.2	5.1	5.2	5.3	5.4	5.3	5.1	4.9	4.7	4.6	4.7	5.2	5.00
	100	4.8	4.8	4.8	4.8	4.7	4.7	4.8	4.9	5.0	5.1	5.1	5.0	5.0	4.9	5.0	5.2	5.2	5.1	5.0	4.9	4.7	4.6	4.6	5.0	4.90
	4	4.7	4.7	4.6	4.6	4.7	4.7	4.7	4.7	4.8	4.8	4.8	4.8	4.8	4.8	4.8	4.9	4.9	4.9	4.9	4.8	4.6	4.5	4.5	4.8	4.74
14	500	5.5	5.4	5.3	5.5	5.4	5.6	5.6	5.5	5.3	5.1	5.1	5.2	5.2	5.1	4.8	4.8	5.0	5.1	5.0	5.1	5.1	5.0	4.9	5.1	5.20
	100	5.3	5.2	5.1	5.2	5.2	5.2	5.3	5.3	5.2	5.2	5.2	5.1	5.1	4.9	4.7	4.8	4.8	4.7	4.7	4.7	4.6	4.6	4.5	4.4	4.96
	4	5.0	5.0	5.0	5.0	5.0	5.0	5.1	5.1	4.9	4.8	4.9	4.9	4.9	4.8	4.6	4.7	4.7	4.6	4.5	4.5	4.5	4.4	4.5	4.4	4.78
15	500	5.0	4.9	4.9	4.7	4.4	4.3	4.3	4.3	4.2	4.2	4.1	4.0	4.1	4.1	4.1	4.0	4.0	3.9	4.1	4.1	3.8	3.6	3.6	3.7	4.18
	100	4.4	4.6	4.7	4.7	4.6	4.4	4.4	4.4	4.4	4.6	4.7	4.4	4.2	4.3	4.3	4.2	4.2	4.2	4.1	4.1	3.9	3.8	3.8	3.6	4.29
	4	4.3	4.5	4.6	4.5	4.5	4.5	4.5	4.5	4.4	4.6	4.7	4.5	4.4	4.5	4.6	4.4	4.4	4.3	4.3	4.2	3.9	3.8	3.7	3.5	4.34
16	500	3.7	3.7	4.0	4.1	3.9	3.7	3.8	4.0	4.2	4.3	4.3	4.4	4.4	4.4	4.3	4.4	4.3	4.2	4.0	3.8	3.8	3.8	3.7	3.6	4.03
	100	3.5	3.4	3.6	3.7	3.8	3.8	4.0	4.2	4.3	4.2	4.3	4.4	4.4	4.4	4.3	4.3	4.3	4.1	3.9	3.8	3.6	3.5	3.8	4.1	3.99
	4	3.5	3.6	3.8	3.9	4.0	4.1	4.2	4.4	4.5	4.5	4.5	4.6	4.6	4.4	4.4	4.5	4.5	4.4	4.1	3.8	3.6	3.6	3.9	4.2	4.15
17	500	3.8	4.2	4.3	4.3	4.2	4.3	4.3	4.4	4.5	4.5	4.5	4.6	4.6	4.8	4.9	4.9	4.8	4.8	4.8	4.8	4.8	4.8	4.9	4.9	4.57
	100	4.2	4.3	4.3	4.3	4.4	4.4	4.4	4.6	4.6	4.5	4.5	4.6	4.6	4.8	4.9	4.9	4.8	4.8	4.8	4.8	4.8	4.8	4.9	4.9	4.62
	4	4.4	4.5	4.4	4.5	4.5	4.6	4.5	4.6	4.6	4.6	4.6	4.6	(4.6)	(4.7)	(4.8)	(4.8)	(4.8)	(4.8)	(4.8)	(4.8)	(4.8)	(4.8)	(4.8)	(4.8)	(4.65)
18	500	5.0	4.9	5.0	5.0	4.9	4.9	4.9	4.9	5.0	5.0	4.9	4.9	5.0	5.0	5.0	5.0	5.0	5.0	4.9	5.0	4.9	4.9	4.9	5.0	4.95
	100	4.9	4.9	4.9	4.9	4.8	4.8	4.8	4.9	4.9	4.9	4.9	5.0	5.0	4.9	4.9	4.9	4.9	4.9	4.9	4.8	4.8	4.8	4.9	5.0	4.89
	4	(4.8)	(4.8)	(4.8)	(4.8)	(4.8)	(4.8)	(4.8)	4.9	4.9	4.9	4.9	4.9	4.9	4.8	4.8	4.9	5.0	4.9	4.8	4.8	4.9	4.8	4.9	4.9	(4.85)
19	500	5.0	5.0	5.0	5.0	5.0	5.0	5.1	5.3	5.3	5.1	5.0	5.0	4.9	4.8	4.9	4.9	4.8	4.9	4.9	4.8	4.8	4.8	4.9	4.8	4.96
	100	5.0	5.0	5.0	5.0	5.0	5.0	5.0	5.2	5.1	5.0	4.9	4.9	4.8	4.8	4.9	4.8	4.8	4.9	4.8	4.8	4.8	4.9	4.9	4.9	4.92
	4	4.9	4.9	4.9	4.9	5.0	4.9	4.9	4.9	5.0	4.9	4.8	4.8	4.9	4.8	4.8	4.8	4.8	4.8	4.8	4.8	4.7	4.8	4.8	4.8	4.85
20	500	4.9	4.8	4.8	4.7	4.6	4.7	4.9	4.9	4.8	4.7	4.7	4.6	4.6	4.7	4.7	4.6	4.7	4.7	4.7	4.8	4.9	4.8	4.9	4.9	4.75
	100	4.9	4.9	4.8	4.8	4.7	4.7	4.8	4.9	4.9	4.9	4.8	4.7	4.7	(4.8)	(4.8)	(4.7)	(4.8)	(4.7)	(4.8)	(4.8)	(4.8)	(4.8)	(4.8)	(4.8)	(4.80)
	4	4.8	4.7	4.7	4.6	4.7	4.7	4.8	4.8	4.8	4.8	(4.7)	(4.6)	(4.6)	4.7	4.7	4.7	4.6	4.7	4.7	4.6	4.7	4.8	4.7	4.7	4.70
21	500	4.9	4.8	4.9	5.0	5.0	5.0	5.0	5.0	5.1	5.2	5.3	5.3	5.2	5.1	5.0	5.0	4.9	4.7	4.5	4.4	4.1	3.9	3.8	3.6	4.78
	100	4.8	4.8	4.9	4.8	4.8	4.9	4.9	5.0	5.1	5.1	5.2	5.1	5.1	5.0	4.9	5.0	4.9	4.7	4.5	4.4	4.2	4.0	3.8	3.6	4.73
	4	4.7	4.8	4.8	4.8	4.8	4.8	4.7	4.8	4.8	5.0	5.0	5.0	4.8	4.8	4.8	4.9	4.8	4.7	4.6	4.5	4.3	4.1	3.8	3.6	4.65
22	500	3.5	3.5	3.6	3.7	3.8	3.9	4.0	4.0	3.9	3.9	4.1	4.0	3.7	3.7	3.9	4.0	3.9	3.8	3.8	3.9	4.0	3.9	3.8	3.9	3.84
	100	3.5	3.5	3.5	3.6	3.7	3.8	4.0	4.1	4.1	4.2	4.1	4.0	3.8	3.7	3.9	4.0	4.0	3.8	3.8	3.9	4.0	3.9	3.9	3.9	3.86
	4	3.6	3.5	3.6	3.7	3.7	3.9	4.1	4.2	4.2	4.4	4.4	4.4	4.5	4.5	4.5	4.4	4.3	4.2	(4.1)	(4.1)	(4.0)	(4.0)	(4.2)	(4.3)	(4.12)

Table III. (Continued.)

Date	Height cm	Hour 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
July 23	500	3.9	3.8	3.5	3.5	3.5	3.6	3.5	3.5	3.6	3.6	3.7	3.8	3.9	4.0	4.0	4.0	3.9	3.9	3.8	3.7	3.6	3.8	4.1	4.2	3.77
	100	3.9	3.8	3.6	3.6	3.6	3.7	3.6	3.6	3.7	3.8	3.8	3.9	3.9	4.0	4.0	3.9	4.0	4.0	3.9	3.7	3.5	3.7	4.1	4.3	3.82
	4	(4.2)	(4.2)	(4.0)	(4.2)	(4.3)	(4.3)	(4.3)	(4.4)	(4.4)	(4.5)	(4.4)	(4.3)	(4.4)	(4.4)	(4.3)	(4.4)	(4.3)	(4.2)	-	-	-	-	-	-	-
24	500	4.2	4.1	4.1	4.1	4.0	4.0	4.0	3.9	3.8	3.8	3.7	3.8	3.8	3.8	3.9	3.9	3.8	3.8	3.9	4.1	4.2	4.3	4.3	4.4	3.99
	100	4.3	4.2	4.1	4.2	4.1	4.1	4.1	4.0	4.0	3.9	3.8	3.9	3.9	3.8	3.9	3.8	3.9	3.9	4.0	4.1	4.3	4.3	4.4	4.5	4.06
	4	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
25	500	4.5	4.5	4.5	4.6	4.6	4.5	4.6	4.7	4.7	4.6	4.6	4.7	4.7	4.7	4.7	4.6	4.7	4.7	4.8	4.8	4.9	4.8	4.8	4.8	4.67
	100	4.5	4.5	4.4	4.6	4.7	4.6	4.7	4.6	4.6	4.6	4.6	4.7	4.7	4.6	4.7	4.6	4.7	4.8	4.9	4.8	4.8	4.8	4.8	4.7	4.67
	4	(4.5)	(4.5)	(4.5)	(4.5)	(4.6)	(4.5)	(4.6)	(4.6)	(4.5)	(4.6)	(4.6)	4.6	4.6	4.6	4.6	4.6	4.7	4.7	4.7	4.7	4.7	4.6	4.6	4.6	(4.60)
26	500	4.8	4.8	4.7	4.6	4.6	4.5	4.4	4.3	4.4	4.5	4.4	4.3	4.3	4.3	4.3	4.3	4.2	4.2	4.2	4.0	3.9	3.9	3.9	3.9	4.32
	100	4.8	4.8	4.6	4.6	4.6	4.5	4.4	4.3	4.3	4.4	4.4	4.4	4.4	4.3	4.2	4.2	4.2	4.2	4.3	4.1	3.9	3.8	3.7	3.5	4.29
	4	4.7	4.7	4.6	4.6	4.6	4.5	4.5	4.4	4.4	4.4	4.4	(4.4)	(4.4)	(4.4)	(4.3)	(4.3)	4.3	4.4	4.3	4.0	3.9	3.7	3.6	3.5	(4.30)
27	500	3.7	3.7	3.7	3.9	3.9	3.9	4.1	4.1	4.1	4.1	4.1	4.1	4.2	4.2	4.2	4.2	4.2	4.1	4.1	4.1	4.0	4.2	4.4	4.3	4.07
	100	3.4	3.4	3.4	3.5	3.7	3.8	4.0	4.2	4.3	4.2	4.3	4.3	4.3	4.3	4.2	4.2	4.3	4.2	4.1	4.1	4.1	4.2	4.4	4.2	4.05
	4	3.4	3.4	3.3	3.4	3.7	3.9	4.1	4.5	4.5	4.5	4.5	4.5	4.5	4.5	4.5	4.5	4.4	4.3	4.3	4.3	4.3	4.3	4.2	4.1	4.16
28	500	4.2	4.1	3.9	3.8	3.8	3.8	3.7	3.8	3.7	3.6	3.4	3.5	3.8	3.9	4.0	4.2	4.2	4.2	4.2	4.2	4.0	3.8	4.0	4.1	3.91
	100	4.3	4.2	4.1	3.9	3.8	3.7	3.8	3.8	3.7	3.5	3.4	3.5	3.7	3.9	3.9	3.9	4.1	4.2	4.3	4.3	4.1	3.9	4.0	4.2	3.92
	4	(4.1)	(4.0)	(3.9)	3.7	3.7	3.8	3.9	3.9	3.8	3.6	3.6	3.7	3.8	3.9	3.9	3.7	4.0	4.0	4.0	4.0	4.1	4.1	4.1	4.3	3.90
29	500	3.9	3.8	3.8	3.9	4.0	4.0	4.1	4.0	3.9	3.8	3.9	3.9	3.9	3.9	4.0	4.0	3.8	3.7	3.5	3.2	3.0	3.1	3.2	3.2	3.73
	100	3.9	3.5	3.6	3.9	4.1	4.2	4.2	4.1	4.2	4.1	4.0	4.1	4.1	4.0	4.1	4.0	3.9	3.8	3.7	3.6	3.4	3.5	3.6	3.5	3.88
	4	3.9	3.2	3.6	4.1	4.2	4.3	4.3	4.3	4.3	4.3	4.3	4.4	4.4	4.3	4.4	4.3	4.2	4.0	4.0	4.0	3.8	3.7	3.6	3.6	4.06
30	500	3.3	3.2	3.2	3.3	3.3	3.4	3.5	3.6	3.6	3.5	3.4	3.4	3.6	3.7	3.8	3.7	3.6	3.5	3.5	3.3	3.3	3.2	3.3	3.2	3.43
	100	3.5	3.4	3.4	3.4	3.5	3.5	3.6	3.8	3.9	3.7	3.6	3.7	3.8	3.8	3.9	3.8	3.8	3.7	3.6	3.4	3.3	3.3	3.2	3.2	3.58
	4	3.5	3.4	3.4	3.4	3.5	3.6	3.8	3.9	4.0	4.1	4.0	3.9	3.9	3.9	4.0	3.9	3.8	3.6	3.6	3.5	3.3	3.2	3.3	3.3	3.66
31	500	3.2	3.2	3.4	3.6	3.6	3.7	3.7	3.6	3.7	4.1	4.1	3.9	4.1	4.4	4.6	4.8	4.9	5.0	4.9	5.0	5.1	5.2	5.2	5.2	4.26
	100	3.2	3.1	3.0	3.2	3.5	3.6	3.7	3.7	3.9	4.1	4.1	4.1	4.2	4.5	4.7	4.8	4.9	4.9	4.8	4.8	4.8	5.0	5.1	5.1	4.20
	4	3.0	2.8	2.8	3.2	3.4	3.4	3.9	4.3	4.4	4.4	4.5	4.5	4.6	4.6	4.7	4.7	4.8	4.8	4.7	4.7	4.7	4.8	4.8	4.9	4.22
Aug. 1	500	5.2	5.3	5.3	5.2	5.1	5.2	5.2	5.3	5.3	5.2	5.1	5.1	5.1	5.1	5.2	5.3	5.3	5.3	5.3	5.3	5.2	5.1	5.2	5.3	5.22
	100	5.0	5.0	5.1	5.0	5.0	5.0	5.1	5.1	5.0	5.0	4.9	4.9	5.0	4.9	5.0	5.0	5.1	5.1	5.2	5.2	5.1	5.0	5.1	5.1	5.04
	4	4.9	4.9	4.9	4.8	4.8	4.7	4.8	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	5.0	5.0	4.9	5.0	5.1	5.1	4.9	4.9	4.9	4.91
2	500	5.4	5.9	6.1	5.9	5.7	5.6	5.6	5.4	5.2	5.0	5.2	5.3	5.4	5.3	5.3	5.4	5.4	5.6	5.7	5.6	5.4	5.4	5.7	6.1	5.52
	100	5.2	5.5	5.8	5.6	5.3	5.2	5.3	5.2	5.0	5.0	5.1	5.1	5.1	5.3	5.3	5.3	5.2	5.2	5.3	5.3	5.4	5.5	5.7	5.8	5.32
	4	5.0	5.2	5.3	5.2	5.1	5.0	5.1	5.2	5.0	4.8	5.0	5.1	5.1	5.0	5.0	5.1	5.1	5.2	5.3	5.3	5.2	5.2	5.3	5.5	5.13
3	500	6.0	5.7	5.6	5.5	5.6	5.6	5.7	5.6	5.6	5.4	5.1	5.1	5.5	5.7	5.7	5.7	5.6	5.6	5.6	5.4	5.3	5.4	5.3	5.3	5.52
	100	5.7	5.5	5.4	5.4	5.4	5.4	5.3	5.3	5.2	5.2	5.0	5.0	5.2	5.2	5.2	5.3	5.3	5.4	5.4	5.3	5.2	5.1	5.1	5.2	5.28
	4	5.4	5.1	5.0	5.0	5.0	4.9	4.9	4.9	4.9	4.8	4.8	4.9	4.9	4.9	4.9	5.0	5.0	5.1	5.1	5.0	5.0	5.0	5.0	4.9	4.98
4	500	5.2	5.1	5.1	5.1	5.1	5.2	5.3	5.2	5.2	5.4	5.4	5.3	5.2	5.2	5.0	4.8	4.8	4.9	5.3	6.0	6.2	6.4	6.5	6.4	5.39
	100	5.2	5.1	5.1	5.0	5.0	5.2	5.3	5.2	5.2	5.3	5.2	5.1	5.1	5.1	4.9	4.9	5.1	5.1	5.1	5.6	5.9	6.0	6.0	5.9	5.28
	4	4.9	4.9	4.9	4.9	4.9	4.9	4.9	4.9	5.0	5.1	5.0	4.8	4.7	4.8	4.8	4.9	5.0	5.1	5.1	5.2	5.3	5.4	5.4	5.4	5.01

Table III. (Continued.)

Date	Height cm	Hour 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	Mean values
Aug. 5	500	6.2	6.3	6.4	6.3	6.3	6.4	6.5	6.6	6.9	6.9	6.9	7.2	7.1	6.9	7.0	7.3	7.1	6.5	6.4	6.6	6.7	6.6	6.6	6.4	6.67
	100	5.8	5.8	5.8	5.7	5.7	6.0	6.2	6.3	6.4	6.4	6.3	6.4	6.1	5.7	5.6	5.8	5.5	5.4	5.7	5.8	5.8	5.8	5.7	5.6	5.89
	4	5.2	5.0	5.1	5.3	5.3	5.2	5.4	5.6	5.5	5.3	5.3	5.4	5.3	5.1	5.0	4.9	4.9	5.0	5.0	4.9	4.9	4.8	4.9	5.0	5.14
6	500	6.3	6.6	6.6	6.5	6.8	7.4	7.6	7.1	6.5	6.5	6.5	6.4	6.6	6.9	6.7	6.4	6.3	6.4	6.5	6.5	6.2	6.3	6.5	6.3	6.60
	100	5.6	5.4	5.5	5.4	5.5	6.4	7.0	6.3	5.5	5.2	5.4	5.5	5.4	5.4	5.4	5.3	5.1	5.1	5.1	5.2	5.2	5.2	5.1	5.0	5.47
	4	5.1	4.9	4.8	4.8	4.9	5.6	5.6	5.6	5.2	5.1	5.1	5.1	5.1	5.1	5.0	4.9	5.0	5.2	5.3	5.1	5.0	4.8	4.7	4.6	5.07
7	500	6.1	5.9	5.9	6.0	5.9	5.9	6.2	6.4	6.5	6.4	6.5	6.3	5.9	5.8	5.7	5.6	5.7	5.6	5.5	5.8	6.0	5.7	5.4	5.5	5.92
	100	5.2	5.5	5.6	5.6	5.6	5.4	5.4	5.5	5.2	5.0	5.1	5.3	5.2	5.0	5.1	5.3	5.4	5.2	5.1	5.3	5.5	5.3	5.0	5.2	5.29
	4	4.6	4.7	4.7	4.6	4.7	4.8	4.8	4.8	5.0	5.1	5.0	4.9	4.9	4.8	4.9	5.0	5.2	5.4	5.1	4.8	4.7	4.6	4.7	4.9	4.86
8	500	5.6	5.6	5.7	5.7	5.7	5.9	6.0	5.8	5.7	5.8	5.8	5.8	5.8	5.8	5.7	5.5	5.3	5.3	5.4	5.4	5.4	5.3	5.3	5.4	5.61
	100	5.3	5.4	5.6	5.6	5.6	5.6	5.7	5.5	5.4	5.3	5.3	5.3	5.1	5.0	5.1	5.0	5.0	4.8	4.8	4.9	4.9	5.0	5.1	5.1	5.22
	4	4.9	4.8	4.9	5.0	5.0	4.9	4.8	4.7	4.8	4.9	4.9	4.9	4.9	4.9	4.9	4.8	4.8	4.7	4.7	4.7	4.7	4.8	4.8	4.9	4.84
9	500	5.4	5.3	5.3	5.3	5.3	5.3	5.1	5.0	5.1	5.1	5.2	5.2	5.2	5.2	5.0	5.0	4.9	4.9	4.8	4.8	4.7	4.8	4.8	4.8	5.06
	100	5.1	5.1	5.1	5.1	5.1	5.0	4.9	4.9	5.0	4.9	4.9	4.9	4.9	5.0	4.9	4.9	4.9	4.7	4.7	4.7	4.7	4.8	4.7	4.7	4.90
	4	4.8	4.8	4.8	4.9	4.9	4.8	4.8	4.8	4.8	4.8	4.8	4.7	4.7	4.7	4.7	4.8	4.8	4.8	4.8	4.7	4.7	4.7	4.7	4.7	4.77
10	500	4.8	4.8	4.8	4.8	4.9	4.9	4.9	5.0	5.0	5.0	5.0	5.0	4.9	5.0	5.0	5.1	5.1	5.2	5.1	5.0	4.8	4.6	4.6	4.6	4.91
	100	4.7	4.7	4.8	4.8	4.9	4.8	4.9	5.0	5.0	4.9	4.9	4.8	4.8	4.7	4.8	4.9	4.9	4.8	4.8	4.7	4.6	4.5	4.4	4.4	4.77
	4	4.7	4.7	4.7	4.8	4.8	4.7	4.7	4.7	4.7	4.7	4.7	4.8	4.8	4.8	4.8	4.7	4.8	4.9	4.8	4.4	4.2	4.3	4.3	4.2	4.65
11	500	4.5	4.6	4.5	4.3	4.1	4.1	4.1	4.1	4.0	3.9	3.7	3.6	3.8	4.1	4.5	4.9	5.0	5.1	5.2	5.2	5.3	5.4	5.4	5.5	4.54
	100	4.4	4.5	4.2	3.9	3.9	4.0	4.1	4.2	4.2	4.2	4.0	3.9	4.1	4.3	4.5	4.6	4.7	4.8	4.9	4.9	5.0	5.1	5.2	5.2	4.45
	4	4.1	4.2	4.1	3.9	3.9	4.0	4.2	4.4	4.4	4.4	4.3	4.2	4.2	4.3	4.5	4.7	4.7	4.7	4.7	4.7	4.7	4.7	4.7	4.7	4.39
12	500	5.5	5.5	5.4	5.2	5.2	5.2	5.3	5.3	5.3	5.4	5.3	5.6	5.5	5.4	5.3	5.3	5.2	5.0	5.0	4.9	5.0	4.9	4.6	4.5	5.20
	100	5.1	5.0	5.0	4.8	4.8	4.9	5.0	5.0	4.9	5.0	5.0	5.0	5.0	4.9	4.9	4.9	5.0	4.9	4.9	4.8	4.8	4.6	4.5	4.5	4.88
	4	4.7	4.7	4.7	4.7	4.7	4.8	4.8	4.8	4.8	4.7	4.7	4.7	4.7	4.7	4.7	4.8	4.8	4.7	4.8	5.0	4.8	4.7	4.6	4.6	4.74
13	500	4.5	4.3	4.3	4.6	4.7	4.6	4.5	4.4	4.4	4.3	4.4	4.4	4.4	4.5	4.7	4.6	4.4	4.5	4.5	4.4	4.4	4.4	4.3	4.3	4.45
	100	4.5	4.4	4.3	4.3	4.4	4.4	4.4	4.4	4.4	4.4	4.4	4.4	4.4	4.5	4.5	4.5	4.5	4.6	4.6	4.5	4.4	4.4	4.4	4.4	4.43
	4	4.5	4.4	4.3	4.2	4.4	4.6	4.5	4.5	4.5	4.5	4.5	4.6	4.6	4.5	4.5	4.6	4.6	4.7	4.6	4.5	4.4	4.5	4.4	4.4	4.49
14	500	4.1	4.1	4.4	4.8	4.8	4.6	4.8	5.0	5.1	5.1	5.1	5.2	5.2	5.1	5.2	5.2	5.0	4.9	4.8	4.7	4.7	4.7	4.8	4.8	4.84
	100	4.3	4.1	4.4	4.7	4.7	4.5	4.5	4.5	4.5	4.5	4.7	4.9	4.8	4.6	4.6	4.7	4.8	4.8	4.8	4.8	4.7	4.8	4.8	4.8	4.64
	4	4.3	4.3	4.5	4.4	4.4	4.4	4.4	4.6	4.7	4.6	4.7	4.7	4.7	4.7	4.7	4.6	4.7	4.6	4.6	4.6	4.7	4.8	4.8	4.8	4.60
15	500	4.8	4.7	4.7	4.6	4.6	4.7	4.7	4.6	4.6	4.6	4.7	4.7	4.8	4.8	4.8	4.7	4.7	4.7	4.6	4.7	4.7	4.7	4.6	4.4	4.68
	100	4.8	4.8	4.7	4.6	4.6	4.6	4.6	4.6	4.5	4.6	4.7	4.7	4.8	4.8	4.7	4.7	4.7	4.7	4.7	4.7	4.7	4.6	4.6	4.4	4.66
	4	4.8	4.7	4.7	4.6	4.6	4.7	4.7	4.6	4.6	4.7	4.7	4.7	4.7	4.7	4.7	4.7	4.6	4.6	4.7	4.7	4.7	4.7	4.5	4.4	4.66

Table IV. *Cloudiness and Hydrometeors.*

Date		Hour 1	2	3	4	5	6	7	8	9	10	11
June 26	Cloudiness	6	5	8	8	6	5	6	6	6	4	6
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
27	Cloudiness	1	2	1	2	1	3	2	3	2	2	2
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
28	Cloudiness	1	1≡°	1	1≡°	10≡°	1	1	2	1	1	1
	Hydrometeors ..	-	-	⌴°	⌴°	⌴°	⌴°	⌴°	⌴°	⌴°	⌴°	⌴°
29	Cloudiness	1	1	1	1	1	1	1	1	1	1	1
	Hydrometeors .	-	-	-	-	-	-	-	-	-	-	-
30	Cloudiness	7	9	10	9	10	10	10	10	9	9	9
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
July 1	Cloudiness	10	10	9	9	10	10	10	10	10	10	10
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
2	Cloudiness	10	10	10	10	10	10	10	10	10	10	10
	Hydrometeors ..	Ⱥ	Ⱥ	Ⱥ	Ⱥ	Ⱥ	✱°	-	-	-	✱▽°	-
3	Cloudiness	10≡°	3	10	7	10	10	9	9	8	10	10
	Hydrometeors ..	✱°	-	-	✱,°	✱°	✱°	✱°	✱°	✱°	✱°	✱°
4	Cloudiness	10≡	10≡^{2}	10≡^{2}	10≡^{2}	10≡^{2}	10≡^{2}	10≡	10≡	1	1	1
	Hydrometeors ..	,	-	-	Ⱥ	Ⱥ	-	-	-	-	-	-
5	Cloudiness	10≡°	10	10≡	10≡	10≡^{2}	10≡	10≡	10≡	10≡	10≡	10≡
	Hydrometeors .	⌴°Ⱥ	-	⌴°Ⱥ	⌴✱°	⌴	⌴✱°	⌴✱°	⌴Ⱥ	⌴Ⱥ	⌴Ⱥ	⌴
6	Cloudiness	3	7	10	10	10≡°	10≡°	10≡°	10≡°	10	9	7
	Hydrometeors ..	-	-	⌴°	⌴°	⌴,	⌴	⌴	⌴	-	-	-
7	Cloudiness	10	10	10	10	10	10	4	4	2	2	3
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
8	Cloudiness	8	8	8	9	9≡°	10	10	10	8	8	9
	Hydrometeors ..	-	-	-	-●	-	-	-	Ⱥ	Ⱥ°	-	Ⱥ°
9	Cloudiness	10≡	10≡^{2}	10≡^{2}	10≡	10≡	10≡°	10≡°	10≡°	1	1	7≡°
	Hydrometeors ..	⌴°Ⱥ°	⌴°Ⱥ°	⌴°Ⱥ°	-	-	-	-	-	-	-	-
10	Cloudiness	2	10	10	10	10	10	10	10	10	10	10
	Hydrometeors ..	-	-	-	✱°	Ⱥ°	-	-	-	-	-	-
11	Cloudiness	10	10	10	10	10	10	10	10	9	10	10
	Hydrometeors ..	-	-	-	-	-	-	Ⱥ°	-	-	-	-
12	Cloudiness	10≡^{2}	10≡	10≡	10≡	10≡	10≡	10≡	10≡	10≡	10≡	10≡
	Hydrometeors ..	Ⱥ∞	✱°∞	Ⱥ∞	,∞	,	,	°	,	,°	,°	—
13	Cloudiness	10≡^{2}	10≡^{2}	10≡^{2}	10≡^{2}	10≡^{2}	10≡^{2}	10≡^{2}	10	10	10≡	10≡
	Hydrometeors ..	⌒	⌒	⌒°	-	-	,°	,°	,°	,°	,	●
14	Cloudiness	4	10≡^{2}	10≡^{2}	10≡°	10≡^{2}	10≡^{2}	10≡^{2}	10≡	10≡^{2}	10≡^{2}	10≡
	Hydrometeors ..	-	⌒	⌒	⌒	⌒	⌒	,^{2}	,^{2}	,^{2}	,	,
15	Cloudiness	0	0	0	1	1	1	1	1	1	1	1
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
16	Cloudiness	1	1	1	1	1	1	2	2	2	1	1
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
17	Cloudiness	10	10	10	10	10	10	10	10	10	10	10
	Hydrometeors ..	-	●▽°	-	-	-	●°	●°	●°	●°	●°	●°

Table IV. *Cloudiness and Hydrometeors.*

Hour													Mean cloudiness
12	13	14	15	16	17	18	19	20	21	22	23	24	
8 -	9 -	9 -	6 -	9 -	8 -	3 -	2 -	2 -	2 -	1 -	1 -	1 -	5.3 -
1 -	1 -	1 -	1 -	1 -	1 -	1 -	1 -	1 -	1 -	1 -	1 -	1 -	1.4 -
1 -	1 -	2 -	2 -	2 -	5≡° ⌴°	8≡° -	2 -	2≡° ⌴°	1 -	1 -	1≡° ⌴°	1 -	2.1 -
1 -	1 -	1 -	1 -	1 -	1 -	1 -	3 -	2 -	3 -	6 -	9 -	7 -	2.0 -
9 -	9 -	9 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	9.5 -
10 -	10 -	10 -	10 -	9 -	10 -	10 -	10 -	10 -	10≡° -	10 -	10 -	10 -	9.9 -
10 ✱▽	9 ✱▽°	7 -	10≡ ✱°	10≡° -	10 ✱°	10 ₳	10 -	8 -	10 -	10 -	10 -	10 -	9.8 -
10≡° -	10≡° -	10≡° -	10 -	9 -	6 -	9 -	10 -	7 -	4 -	2 -	3≡° -	3 -	7.9 -
1 -	9 -	10 -	10 -	10 -	10 -	10≡° -	10≡° -	10 -	10· -	10≡ ₳	10≡² ⌴°₳	10≡ ⌴°₳	8.5 -
10≡ ⌴	10≡ ⌴	10≡ ⌴	3 -	4 -	2 -	1≡° -	1 -	1 -	3≡° -	2 -	1 -	3 -	6.7 -
6 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	9 -	8 -	9.1 -
1 -	6 -	7 -	7 -	4 -	4 -	1 -	2 -	9 -	7 -	1 -	1 -	3 -	5.3 -
3 -	3 -	10≡ -	10≡ -	9 -	10 -	10 -	10 -	10 -	10 -	10≡ -	10≡ -	10≡² ⌴°₳	8.8 -
7≡ ⌴	1 -	1 -	2 -	2 -	3 -	1 -	3 -	2 -	3 -	2 -	2 -	1 -	5.0 -
10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 ₳°	10 ₳°	10 -	10 -	9.7 -
10≡° ₳°	10≡ ₳	10≡° ❟	10≡° ❟²∾	10≡ ₳∾	10≡ ✱°∾	10≡° ✱°	10≡° ✱°	10≡° ✱°∾°	10 ₳∾	10≡ ₳∾	10≡² ₳∾	10≡ ₳∾	10.0 -
10≡ -	10≡ ❟	10≡ ❟	10≡ ❟²	10≡² ❟	10≡ ❟	10≡² ❟	10≡² ❟	10≡² ❟°	10≡² ❟°	10≡² ❟°	10≡² ⌒	10≡² ⌒	10.0 -
10≡² ⌒	10≡² -	10≡² -	10≡² -	10≡² ❟°	9 -	9≡ -	10≡ -	5≡ -	1 -	1 -	1 -	3 -	8.3 -
10≡² -	10≡ -	10≡ -	2≡ -	1 -	1 -	0 -	0 -	0 -	0 -	0 -	0 -	0 -	5.8 -
1 -	1 -	1 -	1 -	1 -	1 -	1 -	1 -	1 -	1 0	1 -	2 -	1 -	0.9 -
1 -	1 -	10≡° -	2 -	1 -	2 -	2 -	2 -	2 -	2 -	4 -	8 -	9 -	2.5 -
10 •°	10 •°	10 •°	10≡ ❟²	10≡ •	10≡ •	10≡ •°	10≡ ❟²	10≡ ❟²	10≡² •°	10≡² •°	10≡² •°	10≡ •°	10.0 -

Table IV. (Continued.)

Date		Hour 1	2	3	4	5	6	7	8	9	10	11
July 18	Cloudiness	10≡	10≡	10≡	10≡	10≡	10≡	10≡	10≡	10≡	10≡	10≡
	Hydrometeors ..	●°	●°	,²	,²	,²	,²	,²	,²	,²	,²	,
19	Cloudiness	10≡²	10≡²	10≡²	10≡²	10≡²	10≡²	10≡²	10≡°	10≡²	10≡²	10≡²
	Hydrometeors ..	,°	,°	,²	,²	,²	,	,°	-	,°	,	,
20	Cloudiness	10≡²	10≡	10≡²	10≡²	10≡²	10≡²	10≡°	10≡°	10≡	10≡	10≡
	Hydrometeors ..	,²	,	,²	,²	,²	,²	●°	●°	,	,²	,²
21	Cloudiness	10≡	10≡	10≡	10≡	10≡	10≡	10≡	10≡²	10≡²	10≡²	10≡²
	Hydrometeors ..	●	●	●	●	●	●	●	,°	-	-	-
22	Cloudiness	3	5≡°	2	9	8	10	10	10	10	10	9
	Hydrometeors ..	-	∾	-	⌴	⌴	-	-	-	-	-	-
23	Cloudiness	10	10	9	8≡°	10≡°	10≡°	10≡°	10≡°	10≡°	10≡°	10≡°
	Hydrometeors ..	Ⱥ°	⌴°	Ⱥ°	Ⱥ°	Ⱥ°	✱°	✱°	✱°	Ⱥ	Ⱥ	Ⱥ²
24	Cloudiness	10≡	10≡²	10≡	10≡²	10≡	10≡	10≡	10≡²	10≡²	10≡°	10≡°
	Hydrometeors ..	,∾	∾	∾	Ⱥ∾	Ⱥ°	,°∾	,°∾	,∾	∾	✱̇°	Ⱥ
25	Cloudiness	10≡°	10≡°	10	10≡	10≡	10≡	10≡²	10≡	10≡	10≡	10≡°
	Hydrometeors ..	Ⱥ∾	,∾	,°	,	,²	,²	,²	,²	✱̇°	✱̇°	✱̇°
26	Cloudiness	10≡°	9	6	9	9	6	10≡	4	2	3	4
	Hydrometeors ..	,°	-	-	-	,°	-	-	-	-	-	-
27	Cloudiness	1	1	2	2	2	2	2	2	2	2	2
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
28	Cloudiness	2	1	7≡	1	1	1	10≡	9≡°	8≡°	6	3
	Hydrometeors ..	-	-	⌴	-	-	-	⌴	,°	,°	-	-
29	Cloudiness	9	4	7	9	9	8	7	6	5	8	8
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
30	Cloudiness	7	4	2	2	2	3	6	3	3	2	3
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
31	Cloudiness	6	5	6	8	8	8	8	9	10	10	10
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
Aug. 1	Cloudiness	10≡	10≡²	10≡	10≡	10≡	10≡	10≡	10≡²	10≡²	10≡²	10≡²
	Hydrometeors ..	,°	,	,	,²	,	,	,°	,°	-	,	,
2	Cloudiness	10≡	10≡	10	10	10	10	10	10	10	10	10≡²
	Hydrometeors ..	●	●	●	●°	●°	,	,	-	-	-	-
3	Cloudiness	10≡²	10≡²	10≡²	10≡²	10≡²	10≡²	10≡²	10≡²	10≡²	10≡²	10≡
	Hydrometeors ..	⌒	⌒	⌒	⌒	⌒	⌒	⌒	,°	,	,²	,²
4	Cloudiness	10≡	10≡	10≡	10≡²	10≡	10≡	10≡	10≡	10≡²	10≡²	10≡
	Hydrometeors ..	,°	,°	,	,	,°	,°	,°	-	-	-	-
5	Cloudiness	10≡²	10≡²	10≡²	10≡²	10≡²	10≡²	10≡²	10≡²	4≡²	2	7≡°
	Hydrometeors ..	⌒	⌒	⌒	⌒	⌒	⌒	⌒	-	-	-	-
6	Cloudiness	10≡²	10≡²	10≡²	10≡²	1≡	1≡°	1≡°	10≡²	5≡²	5=°	6=°
	Hydrometeors ..	,°	,°	⌒	⌒	⌒	-	-	-	⌒	-	-
7	Cloudiness	2=°	4=°	7=°	7=°	7=°	8=°	8=°	6=°	8=°	8=°	8=°
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-

Table IV. (Continued.)

Hour													Mean cloudiness
12	13	14	15	16	17	18	19	20	21	22	23	24	
10≡ ,^{2}	10≡ ,^{2}	10≡^{2} ,°	10≡^{2} -	10≡^{2} ⌒	10≡^{2} ⌒	10≡^{2} ⌒	10≡^{2} ⌒	10≡^{2} ,°	10≡^{2} ●°	10≡^{2} ,	10≡^{2} ,°	10≡^{2} ⌒	10.0 -
10≡^{2} ,^{2}	10≡ ,	10≡ ,^{2}	10≡ ,^{2}	10≡ ,	10≡ ,	10≡^{2} ,	10≡ ,	10≡^{2} ,	10≡^{2} ,	10≡^{2} ,^{2}	10≡^{2} ,^{2}	10≡^{2} ,^{2}	10.0 -
10≡ ,°	10≡^{2} ,°	10≡^{2} ,	10≡^{2} -	10≡ -	10 -	10 -	10 ●°	10≡ ●°	10≡ ●	10≡ ●	10≡ ●	10≡ -	10.0 -
10≡^{2} -	10 -	10≡ -	10≡ -	10≡^{2} ⌒	10≡ ,	10≡ ⌒	10≡ ⌒	10≡ ⌒∽	9≡ ∽	6≡° -	3 -	3 -	9.2 -
9 -	10 -	10 -	10 -	9 -	10≡° -	9≡ ⊔°	10≡ ⊔°	10≡ ⊔°	10≡ ⊔°	5 -	9 -	10 ⊔°	8.6 -
10≡° Ⱥ^2	10≡ Ⱥ^2	10≡° Ⱥ°	10≡^{2} Ⱥ°	10≡^{2} -	10≡^{2} Ⱥ∽	10≡ ∽	9 ∽	5 -	7 -	10 -	10≡^{2} ,°∽	10≡ ,°∽	9.5 -
10≡° Ⱥ°	10≡° Ⱥ°	10≡° Ⱥ^2	10≡° Ⱥ^2	10≡° Ⱥ^2	10≡ Ⱥ^2	10≡° *°	10≡ *°	10≡° Ⱥ^2	10≡° ,^{2}∽	10≡° Ⱥ∽	10≡° Ⱥ∽	10≡° Ⱥ∽	10.0 -
10≡° ,	10≡ •*°	10≡ •*	10≡° •*	10≡° •*	10≡ •*	10 •*°	10≡ ,°	10≡ ,°	10≡ ,°	10≡ ,°	10≡ ,°	10≡ ,°	10.0 -
2 -	8≡ -	3 -	1 -	1 -	1 -	1 -	1 -	1 -	1 -	1 -	2 -	1 -	4.0 -
2 -	3 -	3 -	2 -	2 -	8 -	9 -	9 -	10 -	10 -	10 -	10 -	3 -	4.2 -
2 -	2 -	3 -	8 -	9 -	9 -	10 -	10 -	10 -	10 -	10 -	10≡° *°	10 -	6.3 -
9 -	9 -	9 -	9 -	9 -	8 -	9 -	9 -	10 -	10 -	9 -	9 -	9 -	8.2 -
2 -	3 -	5 -	3 -	1 -	1 -	2 -	2 -	6 -	6 -	7 -	8 -	8 -	3.8 -
10 -	10 -	10 •∇°	10 -	10≡ ⌒	10≡^{2} ,°	10≡^{2} ,°	10≡^{2} ,°	10≡ ,°	10≡ ,°	10≡^{2} ,	10≡ ,°	10≡ ,°	9.1 -
10≡^{2} ,	10≡^{2} ,^{2}	10≡ ,	10≡^{2} ,	10≡^{2} ,	10≡^{2} ,	10≡^{2} ,	10≡^{2} ,°	10≡^{2} ⌒	10≡ ,°	10≡ ,°	10≡ ,°	10≡ ●°	10.0 -
9≡ -	10≡^{2} -	10≡ -	10 -	10 -	10≡° -	10 -	10 •∇	10≡^{2} -	10≡^{2} -	10≡^{2} ⌒	10≡^{2} ⌒	10≡^{2} ⌒	10.0 -
10≡ ,	10≡ ,	10≡ ,	10≡ ,	10≡ ,	10≡^{2} ,	10≡^{2} ,°	10≡^{2} ,	10≡^{2} ,^{2}	10≡^{2} ,	10≡ ,	10≡^{2} ,	10≡ ,	10.0 -
9≡^{2} -	8≡^{2} -	6 -	8 -	10 -	10 -	10 •∇°	10 ●°	10≡^{2} ●	10≡^{2} •∇⌒	10≡^{2} ⌒	10≡^{2} ⌒	10≡^{2} ,°	9.6 -
1≡° -	10 -	8≡° -	9≡° -	10≡^{2} -	10≡^{2} ⌒	9≡^{2} ⌒	10≡^{2} ⌒	10≡^{2} ⌒	10≡^{2} ⌒	10≡^{2} ⌒	10≡^{2} ,°	10≡^{2} ,°	8.8 -
6≡ -	6≡ -	3≡° -	1 -	1≡° -	2≡° -	2≡° -	1≡° -	7≡° -	2≡° -	5≡°	3≡° -	3≡° -	4.6 -
5≡° -	7≡° -	8≡° -	8≡° -	8≡° -	8≡° -	8 -	9 -	10 -	10 -	10 -	10 -	10 -	7.7 -

Table IV. (Continued.)

Date		Hour 1	2	3	4	5	6	7	8	9	10	11
Aug. 8	Cloudiness	10≡	10≡	10≡	$10\equiv^{2}$	$10\equiv^{2}$	$10\equiv^{2}$	$10\equiv^{2}$	$10\equiv^{2}$	$10\equiv^{2}$	$10\equiv^{2}$	$10\equiv^{2}$
	Hydrometeors ..	-	-	●	-	$\overset{\bullet}{\nabla}^{\circ}$	$\overset{\bullet}{\nabla}^{\circ}$	$\bullet^{\circ}$	$\bullet^{\circ}$	-	$\bullet^{\circ}$	,
9	Cloudiness	$10\equiv^{\circ}$	$10\equiv^{\circ}$	10	$10\equiv^{\circ}$	10	10≡	10≡	$10\equiv^{2}$	$10\equiv^{2}$	10≡	10≡
	Hydrometeors ..	⌓	-	-	$,^{\circ}$	$,^{\circ}$	$,^{\circ}$	$\overset{\bullet}{\nabla}^{\circ}$	-	$,^{\circ}$	$\bullet^{\circ}$	$\bullet^{\circ}$
10	Cloudiness	10	10	10	10	10	10	10	10	10	10	10
	Hydrometeors ..	-	-	-	$\bullet^{\circ}$	$\bullet^{\circ}$	$\bullet^{\circ}$	-	$\overset{\bullet}{\nabla}^{\circ}$	-	-	$\bullet^{\circ}$
11	Cloudiness	7	6	4	5	5	5	4	4	3	4	8
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
12	Cloudiness	$10\equiv^{\circ}$	10	10	$10\equiv^{2}$	$10\equiv^{2}$	10	10	$10\equiv^{\circ}$	10	10	10
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
13	Cloudiness	8	9	8	8	8	9	9	9	9	9	8
	Hydrometeors ..	-	-	-	-	-	-	-	-	-	-	-
14	Cloudiness	8	6	7	3	3	3	2	2	2	2	4
	Hydrometeors ..	-	.	-	-	-	-	-	9	-	-	-
15	Cloudiness	10≡	$10\equiv^{2}$	$10\equiv^{\circ}$	10	10	10	10	10	10	10	$10\equiv^{\circ}$
	Hydrometeors ..	⌓	⌓	⌓	⌓	⌓	-	-	-	-	-	$,^{\circ}$

Table V. *Precipitation in mm.*

Date	Precipitation in mm		Date	Precipitation in mm		Date	Precipitation in mm		Date	Precipitation in mm		Date	Precipitation in mm		Date	Precipitation in mm	
	Hour 19—7	Hour 7—19		Hour 19—7	Hour 7—19		Hour 19—7	Hour 7—19		Hour 19—7	Hour 7—19		Hour 19—7	Hour 7—19		Hour 19—7	Hour 7—19
June			July			July			July			July			Aug.		
26	0.0	0.0	4	<0.1	0.0	13	<0.1	<0.1	22	0.0	0.0	31	0.0	<0.1	8	2.9	0.4
27	0.0	0.0	5	<0.1	<0.1	14	0.0	<0.1	23	<0.1	<0.1	Aug.			9	2.2	1.0
28	0.0	0.0	6	0.0	0.0	15	0.0	0.0	24	<0.1	<0.1	1	1.0	0.6	10	0.5	<0.1
29	0.0	0.0	7	0.0	0.0	16	0.0	0.0	25	0.9	1.8	2	9.8	0.0	11	0.0	0.0
30	0.0	0.0	8	0.0	<0.1	17	<0.1	3.2	26	0.8	0.0	3	1.6	0.4	12	0.0	0.0
July			9	<0.1	0.0	18	0.8	<0.1	27	0.0	0.0	4	0.9	<0.1	13	0.0	0.0
1	0.0	0.0	10	<0.1	<0.1	19	0.9	0.2	28	0.0	<0.1	5	<0.1	0.0	14	0.0	0.0
2	<0.1	<0.1	11	<0.1	<0.1	20	0.5	0.2	29	<0.1	0.0	6	0.2	0.0	15	0.0	0.3
3	<0.1	0.3	12	<0.1	<0.1	21	18.5	0.1	30	0.0	0.0	7	0.0	0.0			

Table VI. *Diurnal values of the total ablation in cm water.*

Date	H cm	Date	H cm	Date	H cm	Date	H cm	Date	H cm	Date	H cm	Date	H cm	Date	H cm
June		July		July		July		July		July		Aug.		Aug.	
26	12.5	2	9.0	9	6.5	16	8.5	23	0.0	30	1.5	5	18.0	12	4.5
27	4.5	3	7.5	10	0.0	17	12.5	24	0.0	31	8.5	6	11.5	13	2.5
28	7.5	4	5.0	11	1.0	18	12.5	25	-	Aug.		7	11.0	14	0.0
29	6.5	5	0.0	12	12.0	19	13.5	26	11.0	1	17.0	8	7.5	15	7.5
30	7.5	6	3.5	13	19.5	20	9.5	27	5.0	2	23.5	9	6.0	16	0.0
July		7	3.0	14	21.5	21	11.5	28	4.0	3	20.0	10	6.5		
1	11.0	8	4.0	15	11.5	22	0.0	29	9.5	4	15.0	11	3.5		

Table IV. (Continued.)

Hour													Mean cloudiness
12	13	14	15	16	17	18	19	20	21	22	23	24	
$10\equiv^{2}$ ⌒	$10\equiv^{2}$ $,^{0}$	$10\equiv^{2}$ $,^{0}$	$10\equiv^{2}$ $●^{0}$	$10\equiv^{2}$ $,^{0}$	$10\equiv^{2}$ ⌒	$10\equiv^{2}$ ⌒	$10\equiv^{2}$ ⌒ $\overset{●}{▽}^{0}$	$10\equiv$ $,^{2}$	$10\equiv^{0}$ ●	$10\equiv$ $,^{0}$	$10\equiv$, $\overset{●}{▽}^{0}$	$10\equiv^{0}$ $,^{0}$	10.0 -
$10\equiv$,	$10\equiv^{0}$ $●^{0}$	10 $,^{0}$	$10\equiv^{2}$ -	$10\equiv^{2}$ ⌒	$10\equiv^{2}$ -	$10\equiv^{2}$ ⌒	$10\equiv$ -	$10\equiv$ ⌒	$10\equiv^{2}$ $,^{0}$	$10\equiv$ $,^{0}$	$10\equiv$,	$10\equiv^{0}$ $,^{0}$	10.0 -
10 -	10 -	10 -	9 -	6 -	7 -	6 -	7 -	3 -	4 -	6 -	$8\equiv^{0}$ ⌒	7 -	8.5 -
10 -	10 -	10 -	9 -	8 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	7.6 -
10 -	9 -	9 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	9 -	9.9 -
8 -	8 -	8 -	9 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	10 -	9 -	9.0 -
3 -	4 -	6 -	7 -	5 -	7 -	8 -	9 -	9 -	10 -	10 -	$10\equiv^{0}$ -	$10\equiv$ ⌒	5.8 -
$10\equiv^{0}$ -	$10\equiv$ $,^{0}$	$10\equiv$ $,^{0}$	$10\equiv$,	$10\equiv^{0}$ $,^{0}$	$10\equiv^{0}$ $\overset{●}{✳}^{0}$	$10\equiv^{2}$ $,^{0}$	$10\equiv^{0}$ $,^{0}$	$10\equiv^{0}$ ⌒	$10\equiv^{0}$ ⌒	9 -	10 -	10 -	10.0 -

Table VII. *Diurnal values of the total radiation income in g. cal./cm² day.*

Date	I g. cal./ cm² day	Date	I g. cal./ cm² day	Date	I g. cal./ cm² day	Date	I g. cal./ cm² day	Date	I g. cal./ cm² day	Date	I g. cal./ cm² day	Date	I g. cal./ cm² day
June		July		July		July		July		Aug.		Aug.	
27	855	4	645	12	488	20	296	27	576	2	296	9	227
28	855	5	725	13	488	21	315	28	664	3	273	10	366
29	820	6	652	14	540	22	586	29	558	4	385	11	412
30	583	7	778	15	785	23	435	30	611	5	472	12	227
July		8	698	16	709	24	523	31	400	6	507	13	374
1	558	9	849	17	400	25	400	Aug.		7	507	14	488
2	593	10	655	18	390	26	664	1	296	8	174	15	296
3	715	11	523	19	287								