# Studies of the natural evaporation and energy balance

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ABSTRACT

The natural evaporation from agricultural land has been evaluated on an hourly basis continuously for 29 months. The site is situated near Kristianstad in southern Sweden (56° N). The method employed is a "bulk aerodynamical method" with stability dependent coefficient. The key parameters are: wind speed at 12 m above ground, temperature and humidity at 1.5 and 12 m. The measurements are shown to be representative of areas extending about 1 kilometre upwind. The accuracy of individual hourly values is not very high according to inadequacy of the humidity profile equipment, but the values are progressively more significant as averaging time increases.

Examples of series of hourly, daily and monthly values are presented and discussed in terms of the influence of different weather elements.

Monthly values of the "potential evapotranspiration" has been derived with the Thorntwaite and Penman formulae. It is found that the actual evaporation during spring and early summer in this area is in the mean only half that obtained with the Penman formula but almost equal to the Thorntwaite values. During late summer and autumn the Penman values are about 1.3 times the actual, but the Thorntwaite values are about double the actual. During the late autumn and winter period the actual evaporation is considerably greater than the values obtained with the formulae.

A rough seasonal energy balance of the surface is obtained. It is found that in this area nearly equal amounts of the net radiation goes to evaporation and to heating of the atmosphere during spring and early summer. During late summer progressively more and more of the net radiation energy goes to evaporation. During the winter large amounts of heat is supplied by the atmosphere to account for local radiational loss and also for evaporation.

## Introduction

This is the third paper in a series of reports on an experimental evaporation study at Ugerup, situated on the Kristianstad plane in southern Sweden (56° N). In the first contribution to the series (Högström, 1966*a*) the results of 45 eddy vapour flux measurements were presented together with a description of the instrumentation used. In the second report (Högström 1966*b*) the eddy flux measurements formed the basis of an analysis of the influence of stability on the exchange coefficient for water vapour. The findings of that analysis will now be made use of, together with profile measurements, to evaluate the natural evaporation continuously on an hourly basis over a 29 month period.

## Evaluation of the evaporation

### *The method*

The term "bulk aerodynamical method" refers to the use of the following formula for determining evaporation:

$$E = B \cdot \bar{u} \cdot \Delta e \qquad (1)$$

In this equation $\bar{u}$ is the mean wind speed at a fixed level above ground and $\Delta e$ the difference in water vapour between a lower level $z_1$ and a higher level $z_2$. In most applications $B$ is supposed to be a constant at a site where the roughness of the ground is constant. This is however only true if the levels of measurement (of wind and vapour pressure) are quite close to the ground, so that the influence of stability can be neglected. In Ugerup the levels of measurement were: 1.5 and 12 m for the water vapour and 12 m for the wind speed. Supposing height-constant flux, $B$ is obtained from the following formula (see Högström, 196*b*), if $e$ is in millibars, $\bar{u} = \bar{u}_{12}$ in metres per second and $E$ in $\mu g/\mathrm{cm}^2 \cdot s$:

$$B = 76 \frac{1}{\bar{u}_{12}} \frac{1}{\int_{1.5}^{12} (dz/K_W(z))} \qquad (2)$$

where $K_W$ is the exchange coefficient for water vapour. As shown in Högström (1966*b*) the quantity $B$ is a function only of the stability of the layer below 12 m (and of the roughness length $z_0$). Explicit expressions for the relation between $B$ and stability was also given. The results found are in good agreement with the predictions of the *Monin-Obukhov* theory and with recent findings. The stability parameter used in the abovementioned analysis was

$$\Phi = \frac{\Delta\theta_m}{u_m^2}$$

with $\Delta\theta_m = \frac{1}{3}[(\theta_{12} + \theta_{6.4} + \theta_{3.4}) - (\theta_{1.8} + \theta_{0.95} + \theta_{0.5})]$
and $u_m = \frac{1}{6}(u_{12} + u_{6.4} + u_{3.4} + u_{1.8} + u_{0.95} + u_{0.5})$
where $\theta$ means potential temperature and the indices give the height in metres. For the evaluation of $B$ on a running basis a more convenient stability parameter was used:

$$\psi = \frac{\Delta T - 0.1}{\bar{u}_{12}^2} \qquad (3)$$

where $\Delta T$ is the difference in temperature (C°) between the 1.5 m level and the 12 m level. Based on the results of the analysis presented in Högström (1966*b*) a nomogram was constructed for the evaluation of the quanity $B\bar{u}_{12}$ from measurements of $\Delta T$ and $\bar{u}_{12}$. The application of eq. (1) with a stability dependent $B$ may be termed a "modified bulk aerodynamical method".

*Practical procedure*

To evaluate the evaporation with the aid of eq. (1) and the nomogram described above the following quantities must be abstracted from records: $\bar{u}_{12}$, $\Delta e$ and $\Delta T$.

The wind speed at 12 m, $\bar{u}_{12}$ was taken from the recordings of an anemometer of the type designed and used by the Swedish meteorological and hydrological institute. The specimen used in Ugerup was checked against the recordings of a Fuess anemometer (used for the profile measurements) which in turn had been carefully calibrated in a wind tunnel.

Temperature was measured with the aid of ventilated resistance thermometers, and $\Delta T$ could be abstracted to the nearest 0.1°C from records on a Simens Kompensograph (paper speed 5 cm/hour) of the temperature at 1.5 m and 12 m.

Relative humidity was measured with hair hygrometers with "Pernix-hair" and recorded on the potentiometer recorder. To evaluate the difference in vapour pressure $\Delta e$ between the 1.5- and 12-m levels a special evaluation device was used. This device was also used in evaluating the eddy flux measurements and has been described in detail earlier (Högström, 1966*a*). In short the principle of the instrument is as follows: The original record is transported with constant speed over a plane board. With the aid of two indices, which can be moved independent of each other, two operators can follow the temperature and the relative humidity curve respectively. Each index is meachnically coupled to a revolving drum on the mantle of which is fixed a fine metal thread. This metal thread slides against a linear electrical resistance. The metal threads are so arranged on the respective mantles that the voltage outputs from the resulting potentiometer bridges are proportional to the saturation vapour pressure at the temperature in question and to the relative humidity. These two signals are then connected in a simple bridge to give a signal proportional to the actual vapour pressure. This signal is recorded on a separate potentiometer recorder. The original record must be traversed twice to give the vapour pressure at 1.5 m and 12 m and so form the basis for the evaluation of $\Delta e$.

In the way described above hourly mean values were evaluated for the period March 1961–September 1963 (except July and August 1963).

*Accuracy and representativeness*

Evaluating the sources of error to the determination of $E$ from eq. (1) shows that there is *one* major contribution, and this comes from the determination of $\Delta e$. This error has been discussed in Högström (1966*a*). The result is that, due to instrumental inaccuray, the standard error in the individual hourly $\Delta e$-values is of the same order of magnitude as the $\Delta e$-values themselves. Fig. 1 gives an idea of the implication of this inaccuracy on the corresponding hourly $E$-values. The piles are values obtained directly with the eddy flux method. The continuous curve is formed as a connection of hourly mean values derived with the "modified bulk aerodynamical method". The correlation between the flux values and the "bulk values" is 0.88 (strictly speaking: the correlation between the logarithms of the two quantities). Not only the data of Fig. 1 were used in the correlation computation, but all the 45 eddy flux experiments. On the other hand a vast amount of

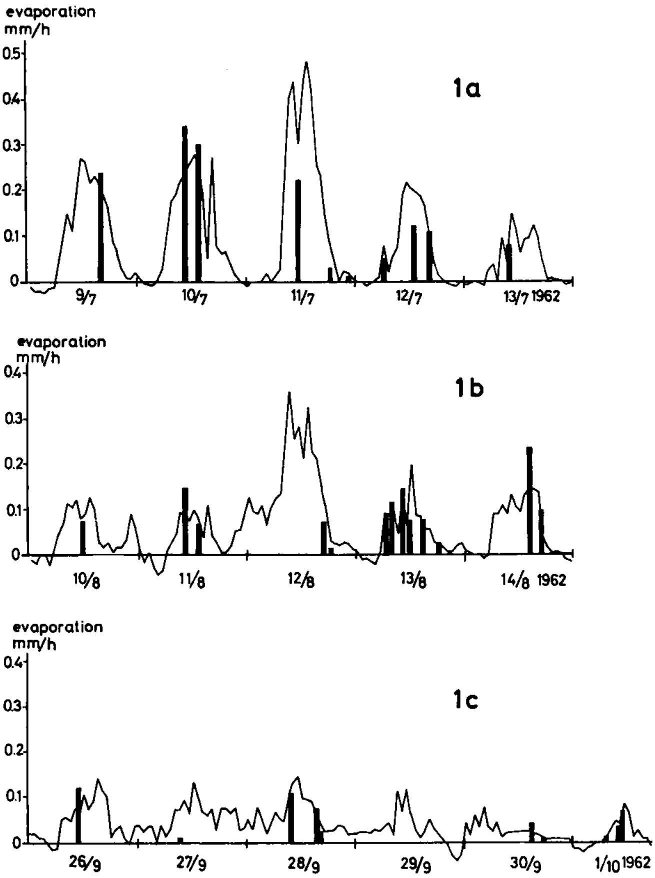


*Fig. 1 a–c.* Hourly evaporation during three separate periods. The piles are values obtained direct with the eddy flux method. The continuous curves connect successive hourly values of evaporation obtained with "the modified bulk aerodynamical method".

checks against Assmann phychrometer secure that systematic errors are avoided. The daily means of $E$ can then be regarded as the sum of 24 values with random errors. The standard error in the individual 24-hour values is then something of the order of 10–15% of the value. The inaccuracy of the monthly means (which are formed as sums of the daily values) is less than 5%. This view is highly supported by an independent analysis of the water balance of the test area, which is to be published soon.

In the applications of the results obtained it is of great interest to know how large area the measurements represent. This question will now be discussed in some detail.

In Ugerup there is a more or less continuous series of measurements of $\bar{u}_{12}$, $\Delta e$ and $\Delta T$ for the period March 1961–September 1963. During this period 45 eddy flux experiments were carried through. The duration of each experiment was of the order of an hour. The standard error of the individual $E$-values thus obtained is estimated in Högström (1966*a*) to be of the order of $\pm 7\%$. The concurrent measurements of $\bar{u}_{12}$, $\Delta e$ and $\Delta T$ were used to form the quantity $B$ (with the aid of eq. (1)) and the bulk stability parameter $\psi$ (eq. (3)). The relation between $B$ and $\psi$ is then analysed, and the result is found to be in good agreement with theoretical predictions and with other experimental findings. The next stage is to exploit this relation between $B$ and $\psi$ to evaluate $E$, with the aid of eq. (1) and the measurements of $\bar{u}_{12}$, $\Delta e$ and $\Delta T$, for those parts of the period March 1961–Sept. 1963 from which there are no flux measurements (that is all hours of the period except 45). It is evident then that the areal representativeness of these determinations must be the same as that of the flux measurements. An estimate of this areal representativeness can be obtained in the following way: Let the $x$-axis of a rectangular coordinate system point in the upwind direction from the point of observation and the $z$-axis vertically. Denote the observed eddy flux by $F_z$ and the absolute humidity by $\chi$ (massunits/volume). Then from continuity reasons:

$$F_z = E_0 + \int_0^{z_1} \frac{\partial \chi}{\partial x} \bar{u}(z)\, dz \tag{4}$$

with $E_0$ = the evaporation at the spot of observation and $z_1$ = the level of observation. Disregard for the moment the variation of the evaporation with the cross wind direction so that the evaporation can be considered as a function of $x$ only, and suppose further that it changes step-wise from one value to another (implying a possibility for the evaporation to vary from one field to another). The local water vapour concentration $\chi$ is made up of contributions from the area up-wind. Write the contribution from the evaporation at distance $x$ as

$$\delta\chi = E(x) \cdot f(x)\, \delta x \tag{5}$$

where $E(x)$ is the local evaporation at distance $x$ up-wind. But according to our assumptions above $E$ varies discontinuously with $x$. Then the derivative $\partial\chi/\partial x$ at the point of observation becomes:

$$\frac{\partial \chi}{\partial x} = -E_0 f(x_0) + E_1[f(x_0) - f(x_1)] + E_2[f(x_1) - f(x_2)] + \cdots \tag{6}$$

if the change from $E_0$ to $E_1$ takes place at the distance $x_0$, the change from $E_1$ to $E_2$ at $x_1$ etc.

The present author (Högström, 1964) has used the following formula to describe the concentration field from a ground level continuous point source:

$$\chi = \frac{Q}{2\sqrt{\pi}\, I(x) \sigma_z \sigma_y \bar{u}_{36}} \exp\left[-y^2/2\sigma_y^2 - z^2/2\sigma_z^2\right] \tag{7}$$

with

$$I(x) = \int_0^\infty \frac{\bar{u}(z)}{\bar{u}_{36}} \frac{1}{\sqrt{2}\,\sigma_z} \exp\left[-z^2/2\sigma_z^2\right] dz \tag{8}$$

$Q$ being the source strength, $\sigma_y$ and $\sigma_z$ the horizontal and vertical standard deviation of the concentration field at distance $x$ and $\bar{u}_{36}$ the mean wind speed at 36 m. If we make the substitution $Q = E\delta_x \delta_y$ in eq. (7) and integrate over all $y$ we obtain the function $f(x)$ of eq. (5):

$$f(x) = \frac{1}{\sqrt{2}\, I(x) \sigma_z \bar{u}_{36}} \exp\left[-z^2/2\sigma_z^2\right] \tag{9}$$

Substitution of $f(x)$ from eq. (9) into eq. (6) and further in eq. (4) gives:

$$F_z = E_0[1 - g(x_0)] + E_1[g(x_0) - g(x_1)] + \cdots \tag{10}$$

with

$$g(x) = \frac{1}{I(x)} \int_0^{z_1} \frac{\bar{u}(z)}{\bar{u}_{36} \sigma_z} \frac{1}{\sqrt{2}} \exp\left[-z^2/2\sigma_z^2\right] dz \tag{11}$$

Eq. (11) can be evaluated numerically if $\sigma_z = \sigma_z(x)$ and $\bar{u} = \bar{u}(z)$ are known. We perform the computations for neutral stratification and for $z_0 = 4$ cm. The procedure for determining $\sigma_z$ is described in Högström (1964). $z_1$, the level of the flux measurements, is equal to about 4 m. The result of the evaluation can be expressed approximately as

$$g = e^{-1.4 \cdot 10^{-3} x} \tag{12}$$

It is then easy to determine the contributions to the measured flux $F_z$ (eq. (10)) from areas situated at varying distance from the point of measurement. The result of this analysis can be summarized in the following way: Equal relative contributions to the measured flux arise from the following areas: 0–200 m, 200–500 m, 500–1000 m and the area behind 1000 m, all in the up-wind direction from the point of measurement. These figures are true for neutral conditions. In the case of unstable stratification the corresponding figures are somewhat smaller and in stable stratification greater. The assumption of cross wind homogeneity in the evaporative field, that underlies the above analysis, is of course an oversimplification. It does however not change the result materially, because the horizontal angle that affects the local flux is narrow.

## Evaporation results

*Hourly data*

As related above hourly values of the evaporation were calculated on a continuous basis. Fig. 1 gives three examples of this. The curves are from periods with simultaneous flux measurements (the piles). As discussed above the errors of individual hourly values are great, according to bad accuracy in the determination of the humidity profile (cf. p. 66). In spite of this there is a fairly good agreement between the eddy flux piles and the continuous curve, as for instance on the 13th of August 1962 when there were six flux measurements on the same day. Thus it seems justifiable to regard the general characteristics of the curves of Fig. 1 as correct. It might then be of some interest to notice the weather relevant to the different curves of Fig. 1: The period 9th of July to 11th of July 1962 is a typical low wind fine weather mid-summer situation (wind 0.5–4 m/s, scattered fine weather cumulus during day hours). From the evening on the 11th of July and to the end of this period the sky was covered mostly with strato-cumulus. The August-period began with high wind (7–9 m/s) and mostly covered with strato-cumulus. During the 12th the wind speed dropped to 5 m/s and the sky cleared up. The 13th was characterized by high wind and more clouds again. During the 14th the wind decreased but the sky did not clear up until the afternoon. The period September 26th to October 1 st 1963 was characterized by high wind (6–8 m/s) and cumulus clouds in the day.

Mean curves for the daily run of the evaporation during the different months of the year were constructed on the basis of the entire 29-month material. Fig. 2 shows the curves for January, April, July and October.

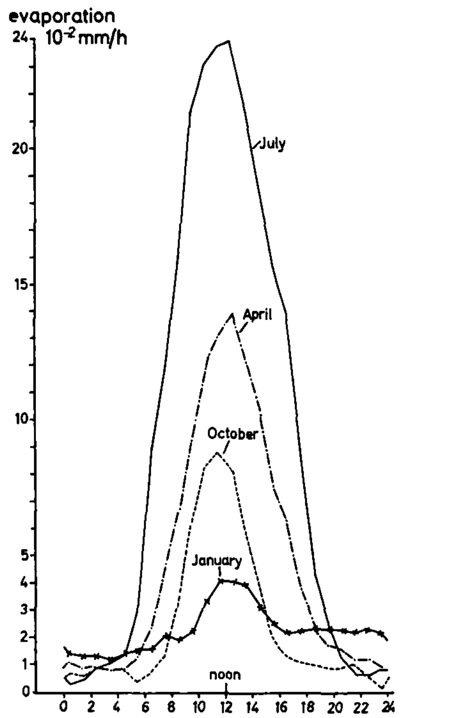


*Fig. 2.* The daily course of evaporation during January, April, July and October. The curves are constructed on the basis of data from 29 months.

*Daily data*

As discussed on p. 67 the standard error of individual 24-hour means is something between 10 and 15% of the value itself. Fig. 3 gives examples of the evaporation during two summer periods with different weather characteristics. The piles in the figures give the precipitation. The period in Fig. 3*a*, May–June 1963, was a pronounced high pressure situation with high insolation and insignificant precipitation. The water content of the soil was high enough to maintain high evaporation into the comparatively dry atmosphere. Observe however the rise of evaporation after the rain on the 2nd of June. Fig. 3*b* shows a period characterized by cyclonic weather with abundant showers, diminished insolation and high humidity of the air. The daily evaporation does not reach very high values, and the curve is subject to rapid oscillations.

Fig. 4 shows two mid-winter periods. The dashed curves give the course of the daily mean temperature. In the first period, 4*a* January 1962 the temperature was between +2°C and +5°C for a long period, and the corresponding evaporation was high. During the other period, December 1962–January 1963 the temperature was about 10°C lower and the evaporation very small.

*Monthly data*

Fig. 5 gives the month to month variation of the evaporation for the whole of the period April 1961 to September 1963. It also gives the corresponding precipitation. Some of the details

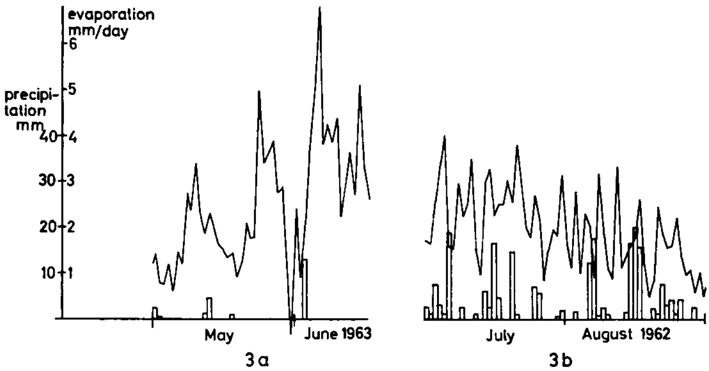


*Fig. 3 a–b.* Examples of the day to day variation in the evaporation during the summer half year. The piles give the daily precipitation.

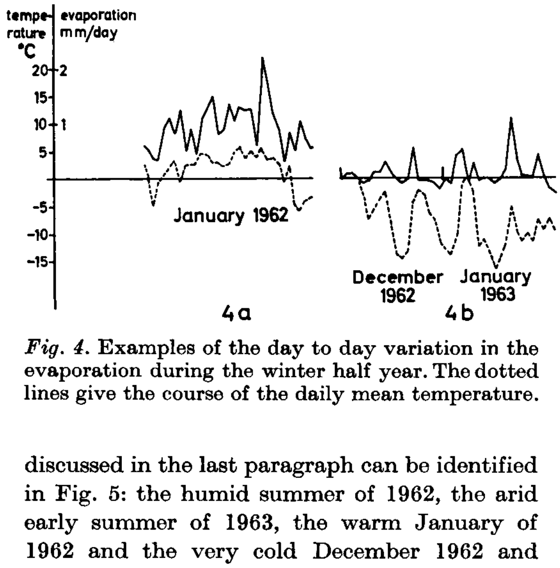


*Fig. 4.* Examples of the day to day variation in the evaporation during the winter half year. The dotted lines give the course of the daily mean temperature.

discussed in the last paragraph can be identified in Fig. 5: the humid summer of 1962, the arid early summer of 1963, the warm January of 1962 and the very cold December 1962 and January 1963.

## Comparisons with other methods

Monthly means for the so called potential evapotranspiration has been derived with the aid of Thorntwaite's and Penman's formulae. Further, monthly mean values were obtained from measurements with the WMO-tank.

### *Thorntwaite's method*

This method (Thorntwaite, 1948) is a purely empirical one, based on observations of normal and actual monthly mean temperature. The formula gives non-zero values only for monthly mean temperatures above the freezing point. Because of this the months November–March were treated separately.

For each month the potential evapotranspiration, $E_p$ was determined and the quotient $E_p/E$ was formed ($E$ being the actual evaporation, determined as described in the first section (p. 66). Fig. 6 gives the result for the months April–October. The points of this diagram form a remarkably consistent picture, and it is easy to draw a curve which fits all but a few points well. It is of course difficult to judge about the generality of Fig. 6, but there are climatological reasons to expect a picture like that obtained. The Thortwaite formula takes only one meteorological factor into account: temperature. But during late summer and autumn the humidity of the air in the observational area is much higher than it is during spring and early summer. Thus the actual evaporation is in the mean only half the computed "potential evapotranspiration" during September and October, in spite of the fact that the ground for most of the time is thoroughly wet. A similar result has been obtained by Aslyng (1961) from evapotranspirometer measurements near Copenhagen.

For the months November–March another representation is applied. In Fig. 7 the actual monthly evaporation is plotted against monthly mean temperature. The potential evapotranspiration according to Thorntwaite is represented as a straight line. A line of best fit has also been drawn. The observed winter evaporation values may seem surprisingly high. Very few actual determinations of Swedish winter evaporation do however exist. For January 1957

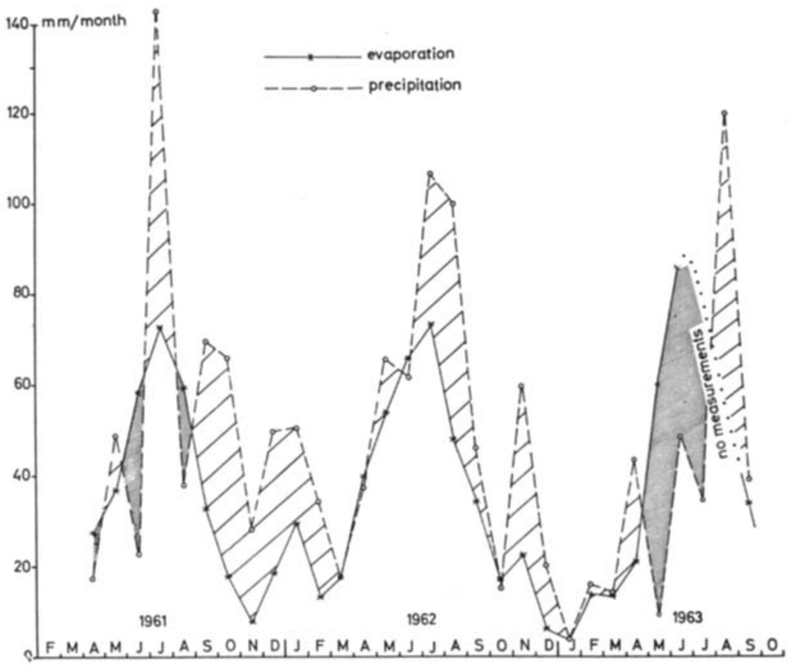


*Fig. 5.* The monthly evaporation for the whole of the period April 1961–September 1963. The dashed line gives the precipitation during the same time.

Nyberg (1965) reports a monthly evaporation of 20 mm as a mean for southern Sweden. The result was obtained with the aerological method.

*Penman's method*

Penman (1948) bases his method on the equation of energy balance of the surface. His assumptions can be summarized as follows:

1. The evaporation is supposed to take place from a hypothetical water surface.
2. The heat storage term is negelected.
3. The exchange coefficient is of the form $a(1+bu_2)$, where $a$ and $b$ are constant and $u_2$ the wind speed at the 2 m level. The potential evapotranspiration is to be computed from the following formula:

$$E_T = \frac{2DR + E_a}{2D+1} \qquad (13)$$

where

$$D = \frac{de_s}{dt}$$

$R$ = net radiation

$E_a = a(1+bu_2)(e_s - e)$, $e$ being the actual and $e_s$ the saturation vapour pressure at screen level.

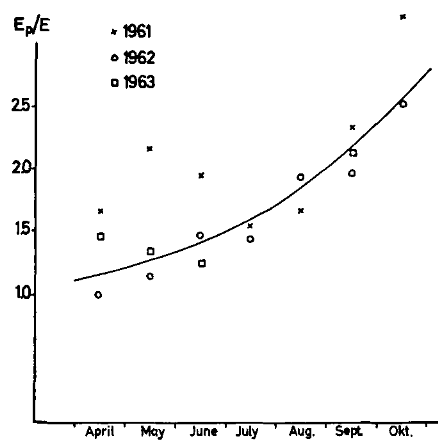


*Fig. 6.* The quotient between the "potential evapotranspiration" $E_p$, determined with Thorntwaite's method and the actual evaporation, $E$ for the different spring and summer months of the study.

In our computations monthly mean values of the different parameters have been used. All quantities except the radiation could be obtained from the measurements at the Ugerup site. The net radiation was estimated in the

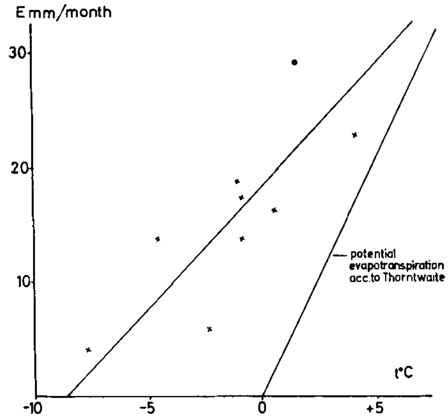


*Fig. 7.* Actual monthly evaporation for the months November–March plotted against monthly mean temperature. The line obtained with Thorntwaite's formula and a regression line are drawn.

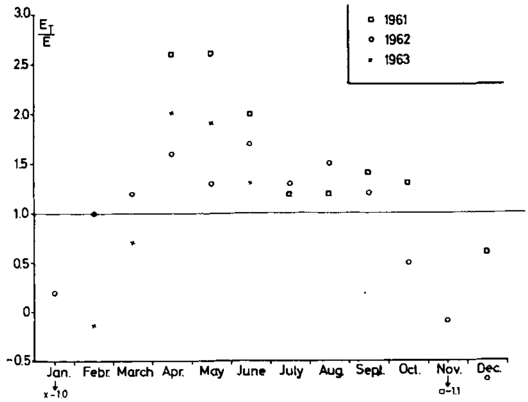


*Fig. 8.* The quotient between the "potential evapotranspiration", $E_T$, determined with Penman's method and the actual evaporation, $E$ for the different months of the period under investigation.

following way: *The incoming radiation,* $R_i$ was taken from records at Svalöv (situated some 60 km west Ugerup but having approximately the same amount and seasonal distribution of precipitation and cloudiness as Ugerup). *The reflected radiation,* $R_r$, was computed from this, by supposing the albedo to be 0.70 for snow covered ground and 0.20 otherwise. The *long wave radiation,* $R_L$, was derived from the following formula (Penman, 1948):

$$R_L = \sigma T_a^4 (0.56 - 0.092 \sqrt{e_a}) \left(0.10 + \frac{0.90\, n}{N-1}\right)$$

where $R_L$ is the net long wave radiation in cal cm$^{-2}$ hour$^{-1}$, $\sigma T_a^4$ is the Stefan-Boltzman relation at mean temperature $T_a$ in °K and $e_a$ is the vapour pressure in mm Hg, $n$ is the actual and $N-1$ the maximum possible duration of bright sunshine, when $N$ means hours between sunrise and sunset. The error in this determination of the net radiation as $R = R_i - R_r - R_L$ is estimated to be of the order of $\pm 10\%$.

$E_T$ was calculated from eq. (13) for every one of the 29 months of the period March 1961–September 1963 for which there was a determination of the actual evaporation $E$. The quotient $E_T/E$ was formed and the result displayed in Fig. 8. The picture shows three distinct features:

During the spring and early summer months (April, May, June) the quotient is in most cases high.

During late summer and autumn (July, August, September) the quotient is rather constant $\approx 1.3$ and with small scatter. During the late autumn and winter months the quotient is less than one and even negative for several months.

The high spring values of $E_T/E$ is probably due to lack of water (assumption 1 on p. 71 is violated), the normal precipitation sum for April, May and June being only about 100 mm, which is to be compared with the sum 230 mm for the period of July, August and September. The winter behaviour of the quotient $E_T/E$ is not so easy to explain. It seems not very probable that erroneous determination of "the actual evaporation", $E$ is the cause of the strange behaviour. As explained on p. 66 the main source of error to $E$ lies in the determination of the vertical water vapour gradient, and because of frequent checks systematic errors are improbable—the "winter points" in Fig. 8 are twelve in number and come from two separate winters. It seems thus as if Penman's formula gives quite erroneous results for winter conditions in this area. We will try to examine the possible reasons for this. There is almost always enough water available during the winter half year in this area, so assumption 1 on p. 66 is justified. During this time of the year most of the evaporation takes place when the air is in a neutral state, so the assumption 3 on the exchange coefficient cannot violate the whole result. The reason for the failure is probably a combination of two things: the neglect of

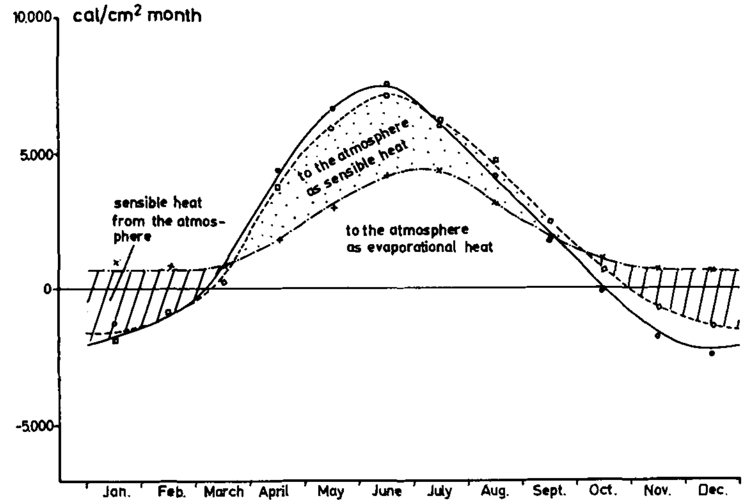


*Fig. 9.* Seasonal surface energy balance based on data from the period March 1961–September 1963. The full line curve gives the net radiation, the dashed line the net radiation minus the estimated heat flux into the ground. The point-dashed curve is the evaporational energy. The area between the dashed and the point-dashed curve gives the flux of sensible heat into the atmosphere.

the heat storage term and the use of too long an averaging time for the computations (a month). During the winter half year there are rapid air mass changes in the experimental area. When warm and wet air invades the area, the rate of evaporation is low, but the ground warms temporarily. Part of this heat is consumed by evaporation, when afterwards colder and drier air replaces the warm and humid air (see also the section on energy balance below).

*The WMO class A evaporation pan*

The metal pan is circular in shape, with a diameter 120.65 cm and with a 25.4 cm high rim. Measurements were not performed during the winter months. The values obtained are rather scattered. The mean of the quotient $E_{pan}/E$ for the 17 months is 1.37. For April, May and June this quotient is 1.54 and for July, August and September 1.20.

## Energy balance of the surface

The equation for the energy balance of the surface:

$$R - H - EL - G = 0 \tag{14}$$

where $R$ is the net radiation, $H$ the sensible heat flux, EL the evaporation (in energy units) and $G$ the heat flux into the ground.

The energy balance was studied on a monthly basis only. $R$ was determined as described in the last section. $G$ which is in most cases small compared to the other terms was taken from Aslyng's (1961) data from the Copenhagen area (mean values for the different months were applied). $H$ was obtained from eq. (14) as the difference between the other terms.

The result of the computations based on data from the period March 1961–Sept. 1963 is given in Fig. 9. The following features of the figure are worthwile mentioning: At about the vernal equinox in the mean the atmosphere begins to heat up locally. During most of April, May and June the incoming radiation is shared, so that about an equal amount goes to evaporation and to heating of the atmosphere. But during July and August progressively more and more of the radiational heat goes into evaporation. At about the autumn equinox the evaporational energy equals the net radiation. But as the ground is warm, there is still some net heating up of the atmosphere. From the middle of October to the beginning of March the radiation balance is negative, and large amounts of heat is imported with warm air masses and abstracted locally. This import is even greater than required to account for the radiational loss, and so a certain evaporation takes place. At the winter solstice, for instance, the evaporational energy takes about 25 % of the available energy.

Results similar to those of Fig. 9 in many respects was obtained by Aslyng (1961). He measured the evaporation with the aid of Thorntwaite evapotranspirometers. His result differs from ours mainly in the way that his winter values are very small compared to ours. As we have discussed in detail earlier it is difficult to present reasons why our winter values should be too great in general.

## Conclusions

A "bulk aerodynamical method" with stability dependent coefficient has been applied to evaluate the natural evaporation on an hourly basis for a 29 month period. The method is shown to give results which are representative of areas extending several hundred metres upwind. The measured parameters are: temperature and humidity at 1.5 and 12 m and wind speed at 12 m. The accuracy of the results obtained grows considerably with the averaging time used, because the systematical errors have been eliminated successfully. This is shown quite independently in a water balance study to be published: almost complete balance is obtained between precipitation, evaporation and run off. The main source of error to the individual hourly values of the computed evaporation comes from the determination of the vertical water vapour gradient.

Examples of series of hourly values, daily values and monthly values have been presented. It has been shown how these data contain detailed information on the influence of different weather elements on the local evaporation. No systematic analysis of the complicated relations has been performed, but it is clear from what has been presented that in the area under investigation air mass qualities, such as water vapour content, stratification etc. exercise an inluence on the local evaporation, which is at least equally important as the direct influence of radiation. This comes out particularly evident in a rough energy balance study which is presented. This study shows that during spring and early summer only about half of the available net radiation energy goes to evaporation (and almost an equal amount to heating the atmosphere), but in late summer most of the net radiation energy goes to evaporation. Similar results have been obtained by Aslyng near Copenhagen, which indicate that the observed features are of regional and not of purely local origin.

It is evident that the method employed in this study can be an excellent tool for studying "evolutionary aspects of energy transfer" (see Preistley, 1959). As recent information (see Högström, 1966*b*) indicates that $K_H = K_W$ over the entire range of stabilities also the heat flux can be evaluated from the same set of measurements as are required for the heat flux determinations. Of course more accurate instruments for recording the water vapour gradient is required, and for practical reasons automatic data logging is preferable.

The 29 monthly mean values of actual evaporation obtained with the method were compared with values of "potential evapotranspiration" obtained in three different ways: values computed with the aid of Thorntwaite's method and Penman's method and from measurements with the WMO evaporation pan. All three methods give pronounced seasonal trends for the quotient $E_p/E$ ($E$ = actual evaporation), but these are not in the same sense for the different methods. Thus the Thorntwaite formula indicates a progressive *rise* of the quotient from a value just above 1.0 in April to over 2.0 in September and October. The Penman formula on the other hand gives high (but scattered) values in the spring and early summer, but values consistently around 1.3 in July, August and September. The result is interpreted as follows: The Thorntwaite formula result reflects the fact that the computation does not take into account the seasonal change in air mass properties. The Penman formula however does. The high spring values of the quotient $E_p/E$ evidently reflects the lack of water during these months. The WMO-evaporation pan gave values roughly in the same sense as Penman's formula.

For the late autumn and winter months much higher monthly evaporation values are observed than obtained with the Penman and Thorntwaite formulae. The observed values are shown to be correlated to the monthly mean temperature to a certain degree. In spite of this correlation of the evaporation to *mean* monthly temperature it is evident that most of the winter evaporation takes place during particularly suited periods. Probably the ground heat storage which is left out of consideration in the derivation of Penman's formula accounts for the failure of this during the winter months.

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## ИЗУЧЕНИЕ ЕСТЕСТВЕННОГО ИСПАРЕНИЯ И БАЛАНС ЭНЕРГИИ

Было определено естественное испарение от обработанной для нужд сельского хозяйства земли на основе ежечасных измерений, продолжавшихся в течение 29 месяцев. Исследуемый участок находится вблизи Кристианстада в Южной Швеции (56° с. ш.). Используемый метод является «полным аэродинамическим методом» с коэффициентом, зависящим от устойчивости. Основными параметрами являются скорость ветра на высоте 12 м над землей, температура и влажность на высотах 1,5 и 2 м. Показано, что измерения являются репрезентативными для областей, простирающихся на расстояния до 1 км против направления ветра. Точность индивидуальных часовых величин не очень высока вследствие неадэкватности оборудования для измерения профиля ветра, однако, эти величины становятся статистически все более значимыми по мере увеличения времени осреднения. Приводятся примеры рядов часовых, суточных и месячных величин и обсуждается влияние на них различных элементов погоды.

С помощью формул Торнтуэйта и Пенмана выводятся месячные величины «потенциальной эвапотранспирации. Найдено, что реальное испарение весной и в начале лета в этой местности равняется в среднем лишь половине того, что получается при использовании формулы Пенмана, но почти совпадает с величинами, даваемыми формулой Торнтуэйта. Для конца лета и осени величины, даваемые формулой Пенмана, примерно в 1,3 раза больше реальных, но формула Торнтуэйта здесь дает завышение примерно в 2 раза. В течение поздней осени и зимы испарение в действительности постоянно больше величин, получаемых с помощью упомянутых формул.

Получен приближенный сезонный баланс энергии для рассматриваемой поверхности. Найдено, что в данной местности приблизительно половина приходящей радиации тратится на испарение и нагрев атмосферы в течение весны и начала лета. В конце лета все большая часть энергии приходящей радиации идет на испарение. Зимой большое количество тепла поступает из атмосферы на возмещение радиационных потерь и затрат энергии на испарение.