EVAPOTRANSPIRATION [1]

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ABSTRACT

The existing methods of calculating evapotranspiration are discussed. No method based on monthly temperature can be expected to give reliable results for different regions. The temperature of the air lags behind solar radiation at moderate and high latitudes and, therefore, the amount of energy available for evaporation in the spring is quite different from that available in the autumn, if periods with the same temperature are compared. For the Netherlands, for instance, this difference amounts to nearly 300 % between the two months of March and November, with average temperatures of 5.0° and 5.4° C respectively. Further discrepancies occur in the case of cold or warm ocean currents, and low temperature owing to elevation.

The energy available for evapotranspiration can be calculated by applying the theory of heat exchange (including radiation) to a wet body of the same shape as the vegetation cover. Actual evapotranspiration will be smaller than the calculated maximum value by a factor depending on the physiology (not the morphology) of the plants.

1 Introduction

The practical importance of evaporation and transpiration of water, from the soil and by vegetation respectively, is well recognized in agriculture. A great deal of experimental data has been published on this subject. Several methods have been proposed for the calculation of evapotranspiration, which is directly related to the amount of water necessary for a crop. Some of these methods are based on an empirical correlation with such climatological data as, for instance, monthly air temperature or saturation deficit (Thornthwaite (1948), Leeper (1950)). A correlation between evaporation from a tank and temperature shifted in phase was given by Prescott (1943). In other methods the energy available for evaporation and for transpiration is calculated from the incident radiation, advective heat and flow of heat from the soil (Penman (1948), Albrecht (1950)). An empirical reduction factor is introduced to cover possible biological regulation of transpiration by the plant. In these articles particular attention is given to "Potential Evapotranspiration" (*P.E.*), which may be defined as the amount of water evaporated under optimal conditions of soil moisture and vegetation.

In a recent article Albrecht (1950) has given a survey of some (mainly German) contributions to methods of calculating evapotranspiration.

There can be no doubt that the methods based on available energy are in principle correct, since the latent heat of evaporation must be supplied durig the evaporation and transpiration processes. Mean monthly temperature is, in itself, not a measure of available energy. Yet formulae based on a correlation with temperature are in widespread use by agronomists. Since the temperature of a region is largely influenced by radiation and advective heat it may at first sight seem not unreasonable to expect that a formula based on local temperature could be so adapted to experimental data that evapotranspiration might be calculated from it.

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In the Anglo-American literature of the subject, the most used formula of this type is that of THORNTHWAITE (1948). It has previously been compared with PENMAN's calculation method, which is representative of the methods based on available energy (VAN WIJK et al., 1953). This comparison is discussed extensively in the present article, and the use of a correlation with temperature for calculating evaporation is especially investigated. From their study the present authors have arrived at the conclusion that the possibility of finding a correlation of this type which has sufficient general validity under varying meteorological conditions must be ruled out.

2 SOME GENERAL REMARKS ON THORNTHWAITE'S METHOD

A brief description of THORNTHWAITE's method of obtaining $P.E.$ from temperature data will now be given, and some features of the method will be discussed. For a full account of it the reader is referred to THORNTHWAITE's own paper (1948).

In THORNTHWAITE's nomogram the monthly values of the so called "unadjusted" potential evapotranspiration in cm, $P.E.^*$, are plotted on a logarithmic scale against the logarithm of mean monthly temperature in °C, which is denoted by ϑ. The term $P.E.^*$ stands for potential evapotranspiration reduced to a standard month of 30 days, with a day length of 12 hrs [2]).

In this nomogram the relation between $P.E.^*$ and ϑ is represented by a straight line, passing through the points with coordinates $P.E.^* = 13.5$, $\vartheta = 26.5$ and $P.E.^* = 1.6$, $\vartheta = I/10$, where the heat index I is defined as $I = \Sigma\ (\vartheta/5)^{1.514}$, the summation being extended over the twelve months of the year.

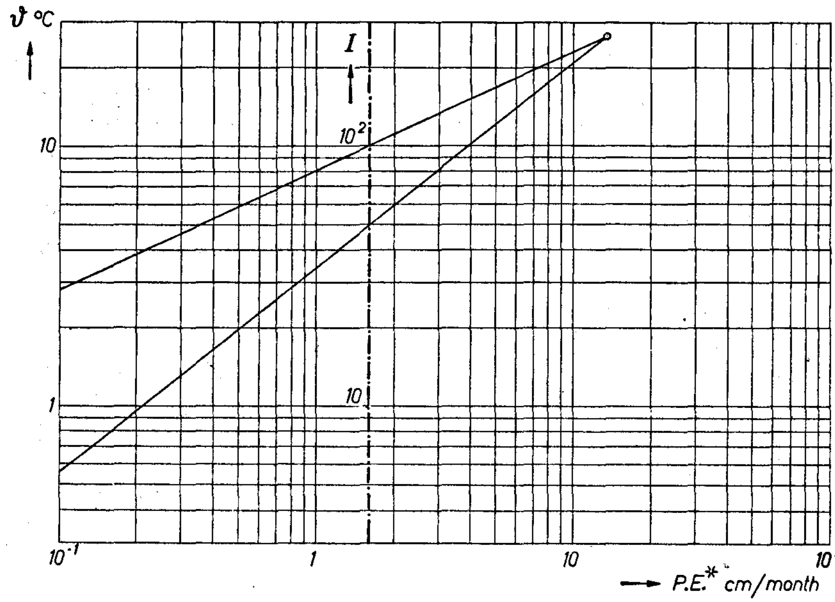

FIG. 1. THORNTHWAITE's nomogram of the relation between mean monthly temperature, ϑ, and unadjusted potential evapotranspiration $P.E.^*$ (= evapotranspiration for a standard month with 30 days of 12 hours each).

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2) Thus, for example, in the month of January, at a latitude of 40° N, the average day length is 9.7 h and $P.E./P.E.^* = (31/30) \times (9.7/12) = 0.84$. According to THORNTHWAITE, the same reduction factors must be used for latitudes above 50° as for 50°. The reason for this must probably be sought in the comparatively rapid decrease of solar altitude with increase of latitude for latitudes above 50°, so that a further increase in day length does not result in a further increase of available energy.

This empirical relationship is based on the evaporation data obtained for a number of catchment areas in the U.S.A., in latitudes ranging from 29° N to 43° N.

In Fig. 1 the nomogram is represented and two lines are drawn with I values of 100 and 50 respectively. At a given temperature the value of *P.E.** is largest for the station with the lower I value. This station has the lower value of $\Sigma\ \vartheta^{1.514}$ and thus it will, in general, be situated at a higher latitude than the station with the higher I value. Now, if the temperature is the same at both stations, the intensity of solar radiation will generally be higher at the station with the lower I value, and this explains the difference in evaporation between the two places.

In connection with the convergence of the lines at high temperatures, it is interesting to note that the monthly values of the incoming solar radiation per unit of surface, if the atmosphere were transparent (these values were computed for different latitudes by Angot, cf. Brunt (1944)), show a relatively small variation with changes of latitude in the months from May to August inclusive, for the northern hemisphere. From these values, and from the values of average cloudiness for zones of 10° latitude given by Conrad (1936), we find (using Eq. 3 in the appendix) that, for latitudes from 0° to 60° N, the average monthly values of solar radiation reaching the earth's surface range from about 9,000 cal/cm² in August for the zone of 50° to 60° latitude to about 15,000 cal/cm² in July for the zone of 30° to 40° latitude. The corresponding average monthly temperatures in these two cases are 14° C and 26° C respectively.

3 Penman's theory

For a full description of Penman's method of obtaining *P.E.* from meteorological observations, viz., temperature, humidity, cloudiness and wind velocity, reference should be made to Penman's original paper (1948). The notation used in that paper has been adopted here. In order to calculate evaporation from a water surface, E_0, Penman has evolved the following equation:

$$E_0 = \frac{\Delta H_0 + \gamma E_a}{\Delta + \gamma} \tag{1}$$

For the meaning of these symbols the reader is referred to the list at the end of this article. Evaporation from a soil with vegetation, E_T, is obtained by multiplying E_0 by an empirical reduction factor. The experimental values of this factor were deduced from experiments at Rothamsted with a number of cylinders, some filled with water and some filled with soil carrying a cover of short grass. The ratio E_T/E_0 appeared to range from 0.6 during the months from November to February inclusive, to 0.8 from May to August inclusive, whilst it was 0.7 in the remaining months.

In a later paper Penman and Schofield (1950) discussed the influence of diffusion resistance in the stomata, and Penman (1952) showed that, on the basis of these considerations, the value of *P.E.* can be found from the following equation:

$$E_T' = \frac{\Delta H_T + \gamma E_a}{\Delta + \gamma / SD}, \tag{2}$$

where S arises from the influence of diffusion resistance in the stomata if open and D arises from the influence of the closing of the stomata during the night. Recently Bange (1953) has shown that the evaporation from leaves of Zebrina pendula at 21–25° C could indeed be quantitatively explained by this diffusion resistance. Furthermore, H_T is computed by taking the reflection coefficient of a vegetation surface equal to 0.20, instead of the value 0.05 adopted for a water surface.

Our values of E_T have, in most cases, been calculated from Eq. 2, since it was felt that the empirical factors of E_T/E_0 obtained for S.E. England might not apply under widely different climatic conditions. These values are indicated as E_T', whereas the values of *P.E.* obtained by multiplying E_0 by Penman's empirical factors are indicated as E_T. In general, E_T appeared to be somewhat larger than E_T' [3]).

An example of the calculations is given in the Appendix.

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3) Penman (1952) gives the following reasons for this discrepancy: (1) a greater aerodynamic roughness of a natural vegetation cover as compared with a water surface or a cover of short grass, (2) increased ventilation caused by air movement inside the vege-

4 Evaporation at different latitudes

To arrive at a comparison of the results obtained, on the one hand, with Thornthwaite's formula and, on the other, with Penman's theory, two modes of procedure were adopted. Firstly, the values of *P.E.* in a number of latitudes were compared with the corresponding values of E_T and E_0, all values being computed for average conditions on the circle of latitude under consideration. Secondly, a comparison was made in respect of a number of selected stations.

The results of the first procedure for latitudes 20° N, 40° N and 55° N are shown in Table 1 and in Fig. 2, while a discussion of the basic material and some technical details are presented in the Appendix, together with a complete example of the calculations.

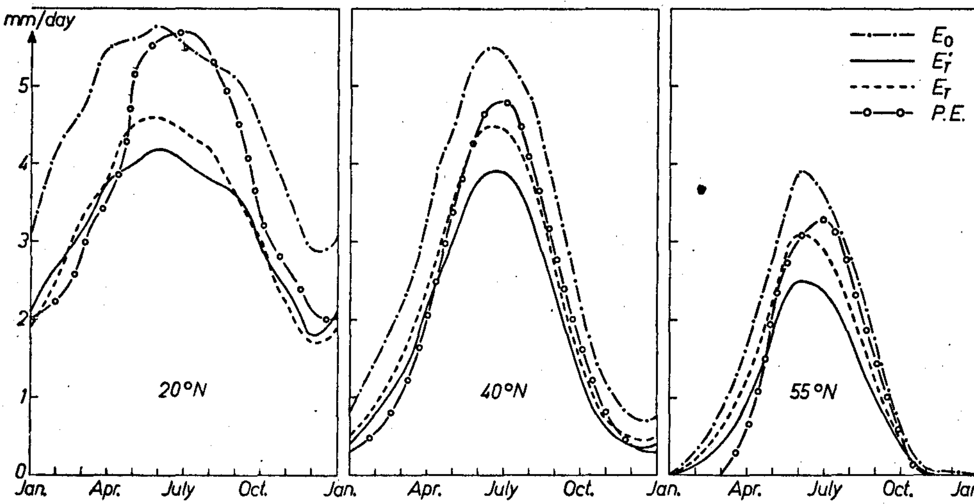

Fig. 2. Potential evapotranspiration for different latitudes after Thornthwaite (*P.E.*), and after Penman (E_0 for a water surface, E_T or E_T' for a land surface; see the note on p. 107 for the differences between E_T and E_T').

Table 1. Yearly potential evapotranspiration according to Thornthwaite (*P.E.*) and Penman (E_T, E_T', E_0) for different circles of latitude.

Latitude	*P.E.*	E_T'	E_T	E_0
20° N	138	117	121	169
40° N	80	71	82	110
55° N	42	35	43	56

It can be seen from the table that the agreement between the yearly values of *P.E.* and E_T is reasonable. The differences are greater if monthly values are considered, especially in the case of low monthly temperatures. In most cases E_T is higher than *P.E.* in winter, whereas the reverse is true during the summer months.

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tation, (3) the fact that the reflection coefficient is smaller than 0.2, (4) evaporation of intercepted rainwater.

The first two factors would cause an increase in E_a, the third would cause an increase in H_T. The present authors have made no attempt to incorporate these features in their calculations.

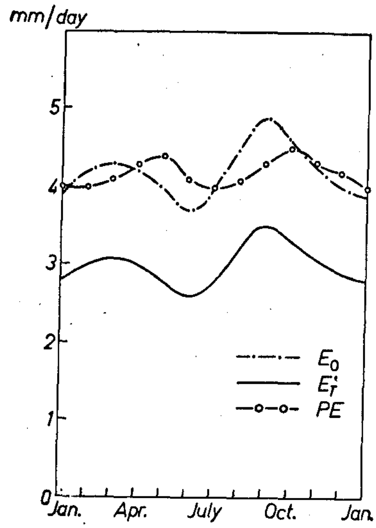

FIG. 3. Potential evapotranspiration for Djakarta (6° 11′S, 106° 50′E, elevation 8 m), after THORNTHWAITE (*P.E.*), and after PENMAN (E_0 for a water surface, E_T' for a land surface).

Near the equator the temperature shows a semi-annual periodicity but the amplitude is rather small, say about 0.5° C, the average temperature being about 26.0° C. From THORNTHWAITE's formula the average monthly evaporation is computed at 13.4 cm. About the same value is found for E_0, as can be seen from the graph for Djakarta (Fig. 3). The values of E_T' are 25 to 30 % lower; the reflection coefficient of 0.20 used in the calculation of E_T' will, however, probably be too high at very low latitudes (see note on p. 107).

5 COMPARISON OF *P.E.* AND E_T FOR SELECTED STATIONS

We will now discuss some results obtained by the second method referred to in the introductory paragraph of the preceding section. If we plot the monthly values of E_T in THORNTHWAITE's nomogram, after dividing them by the corresponding reduction factors and thus reducing them to "unadjusted" values, we find, in general, that these points can be grouped on two different lines, the first line containing the values from January to June and the second line containing the values from July to February.

Thus, at each temperature two values of *P.E.* are obtained by PENMAN's method, *and this is also true of any other method based on available energy.* It is due to the fact that air temperature lags behind solar radiation. Thus, for example, in the spring a lower temperature occurs with the same intensity of solar radiation than in the autumn. Since the available energy is to a large extent supplied by solar radiation two different temperatures are associated with approximately the same *P.E.* The difference is relatively small near the maximum value of the average monthly temperature when *P.E.* is large. But during winter a relatively large difference in E_T at the same temperature occurs. The correlation of *P.E.* with temperature must therefore break down when applied to individual months, although it may give a fair estimate of cumulative *P.E.* for the growing season as a whole, owing to the fact that *P.E.* is maximum in summer, when the temperatures corresponding to the same value of solar radiation most nearly approach each other.

To illustrate this feature, we have computed values of E_T for two stations that could be considered to be under about the same meteorological conditions as two of the examples presented in THORNTHWAITE's paper. The stations are Tampa (Florida) and Cincinnati (Ohio), and the corresponding examples are Kissimmee River Basin and Lysimeter Y 101 B, Coshocton. The results are represented in figures 4a and b; the experimental points given by THORNTHWAITE are also included; part of the scatter shown by them may possibly be explained by the aforementioned phase shift.

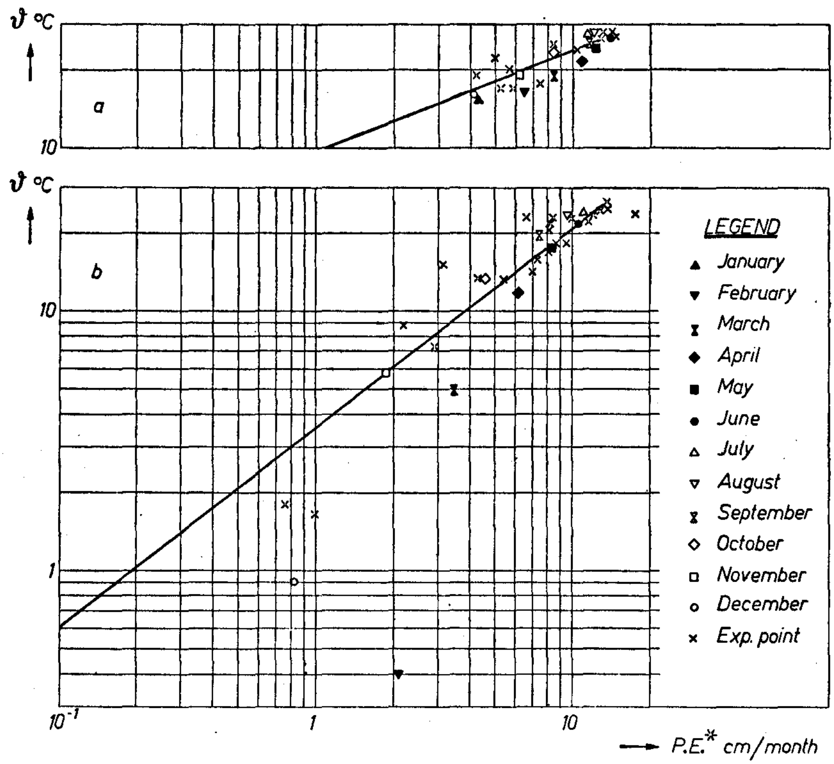

FIG. 4. Unadjusted potential evapotranspiration (*P.E.**) against mean monthly temperature, ϑ, after THORNTHWAITE (straight line) and after PENMAN (cf. legend) for: a) Tampa (Florida; 27°57′N, 82°27′W, elevation 11 m); b) Cincinnati (Ohio; 39°6′N, 84°30′E, elevation 191 m). Note the systematic difference between the values for the first half (black signs) and for the second half (open signs) of the year. Crosses are experimental points from: a) Kissimee River Basin (Florida) and b) Lysimeter Y 101 B, Coshocton (Ohio).

As an illustration of the variation in *P.E.* in different years results of calculations for De Bilt, Holland, are given in Fig. 5. The points of E_T were calculated for a "normal" year (average, 1911—1951) and for the years 1949 and 1951. Both these last years had winter temperatures above average, but 1949 had a warm summer whereas in 1951 the summer temperatures were below average.

Large differences between THORNTHWAITE's and PENMAN's values are found even in the case of average conditions in places where the average monthly temperatures are strongly affected by special influences, such as high altitude, or the presence of warm and cold ocean currents. A number of examples are presented in Figures 6a to d. The special influences are mentioned briefly in the captions to these figures.

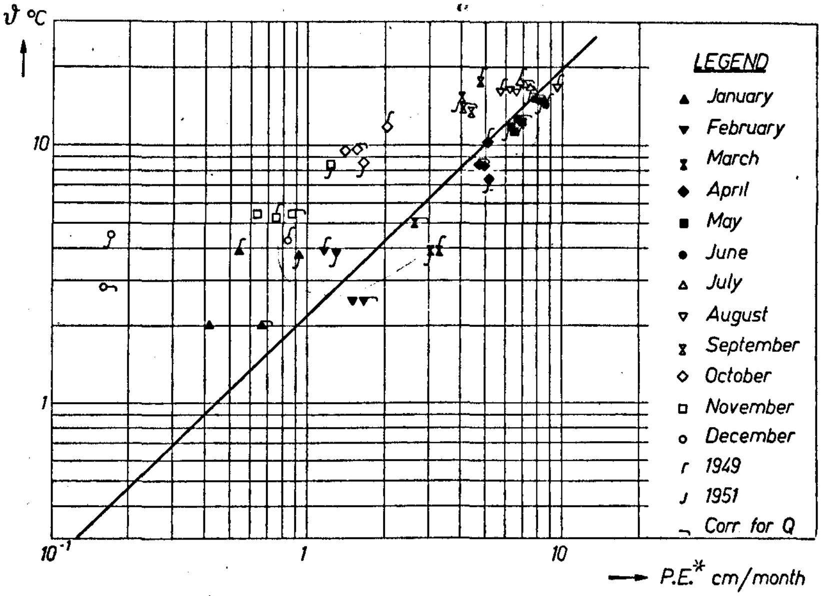

FIG. 5. Unadjusted potential evapotranspiration ($P.E.^*$) against mean monthly temperature (ϑ), after THORNTHWAITE (straight line), and after PENMAN for a "normal year" and two actual years (see legend) at De Bilt. The signs for the "normal" values have no affix; the signs modified by a ⌐ indicate normal values corrected for the heat flux in the soil (Q). Note the systematic difference between the values for the first half of the year (black signs), and for the second half (open signs).

6 DISCUSSION

The data presented above show that it is theoretically not permissible to use temperature as an indication of the energy available for evaporation. Cumulative *P.E.* may be calculated by a formula in which temperature is considered the important factor if the formula is empirically adjusted to average conditions over the wide region, or to the special local conditions of the small region, to which it is applied. In both cases such a formula will have no general applicability. Moreover, it cannot be used to calculate *P.E.* for the separate months. Thus, for example, at De Bilt about the same temperature occurs during November (5.4° C) as during March (5.0° C). The average intensity of solar radiation in these two months is, however, 67 cal/cm² per day and 195 cal/cm² per day respectively, and Fig. 6 shows that the $P.E.^*$ culculated from the energy balance is nearly four times as high in the latter month as it is in the former. Accordingly, any formula in which monthly temperature is used gives too high a value of evapotranspiration in winter if it gives approximately correct results for the spring [4]).

[4]) The scattering of the measurements which were used by THORNTHWAITE to arrive at his empirical formula is large, and of the same order of magnitude as the differences between the two branches of the curve *P.E.* against temperature which have been calculated by PENMAN's method (cf. Figs. 4 and 5). It would be interesting to see whether the points would fit the two branches better if the data to which the measurements corresponded were taken into consideration. These data are, however, not included in THORNTHWAITE's publication.

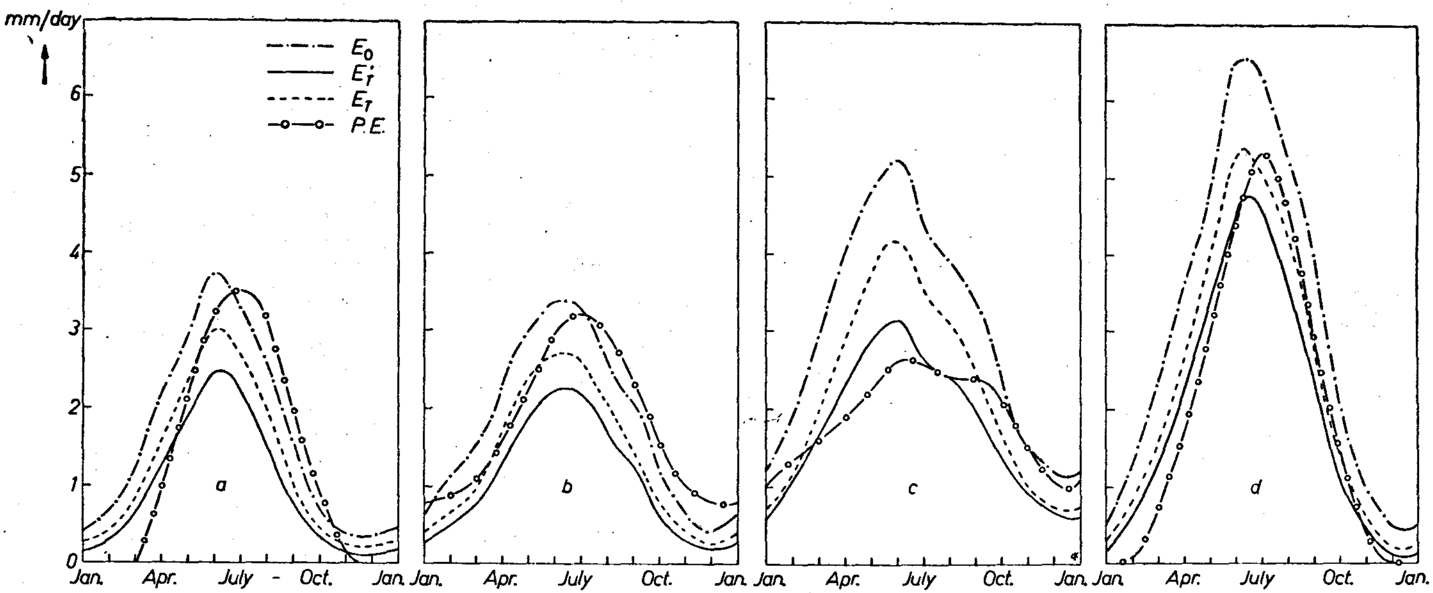

FIG. 6. Potential evapotranspiration after THORNTHWAITE (*P.E.*) and PENMAN (E_0 , E_T' , E_T') for : a) Trondheim (Norway ; 63°26′N, 10°25′E, elevation 40 m) ; b) Valentia (Ireland ; 51°56′N, 10°15′W, elevation 9 m). In both cases *P.E.* is rather large in comparison with E_T. due to the presence of the warm Gulf Stream ; c) San Francisco (U.S.A. ; 37°48′N, 122°26′W, elevation 47 m). *P.E.* is rather small in comparison with E_T , due to the cold California Current ; d) Grand Junction (U.S.A. ; 39°N, 108°34′W, elevation 1403 m). The ratio $P.E./E_T$ is comparatively small (cf. Fig. 2), due to high elevation. *)

*) The yearly values of E_0 and E_T' previously reported for this station (VAN WIJK et al. (1953)) are incorrect. The correct values are : 115 cm and 86 cm respectively ; the yearly value of *P.E.* = 74 cm.

A further fundamental difficulty arises if the temperature of a region is influenced by advective air. A calculation shows that the air temperature at a height of two metres above ground changes rapidly in the first few hundred metres after the air has passed a boundary at which the temperature of the surface of the soil changes, but, that then the temperature becomes practically adjusted to the new surface conditions, and varies only very slowly with distance. The flow of energy between the surface of the soil and the advective air then becomes small compared with normal daily radiation intensity, and thus evaporation is not markedly increased in the case of warm advective air, although the temperature of the air at a height of 2 metres, and that of the soil surface, may be appreciably higher than when advective air is absent. Accordingly, a calculation of *P.E.* based on temperature and empirically corrected for latitude must necessarily yield erroneous results if the temperatures are largely influenced by advective air. Too high a *P.E.* will then be calulated if the air is warm, as is the case in the British Isles and in the Netherlands, and too low a value will be calculated in the case of advective cold air.

Direct measurement of the transpiration of plants under well-defined conditions also shows that the instantaneous value of transpiration depends on the available energy (cf. Briggs and Shantz (1916), and De Vries and Van Duin (1953)).

Since the daily radiation flux is shifted in phase compared with the temperature as well, equal temperatures are again obtained for two different intensities of radiation. Daily temperature can, therefore, not be used as an indication of available energy, and similar discrepancies are to be expected as with monthly temperature.

It is interesting to note that the actual value of the shift in phase of the yearly temperature wave compared with the intensity of solar radiation is in itself an indication that the latter, and not the temperature, is the important factor in evapotranspiration, as has already been shown in a preceding article (Van Wijk et al. (1953)).

The value of the phase shift of the daily temperature wave is such that more precise discussion is necessary to enable a definite conclusion to be arrived at.

The net amount of energy available for evaporation and for the heating of soil and air (H_0 or H_T) is, in most cases, in phase with solar radiation. The term E_a occurring in the numerators of the expressions for E_0 and $E_T{}'$ (eqs. 1 and 2 respectively) is, however, generally in phase with the temperature, since the saturation deficit increases with increasing δ. Evaporation is therefore somewhat retarded in comparison with solar radiation, but the effect is small at the summer maximum, since here γE_a is much smaller than ΔH_0 or ΔH_T ; this retardation is most pronounced in the case of E_T , because $H_0 > H_T$. In the months around the winter minimum of solar radiation γE_a is sometimes larger than ΔH_0 or ΔH_T. The latter quantities can even be negative, whereas E_0 and E_T may remain positive, in which case the energy for evaporation is supplied by the soil and air.

Accordingly, in the authors' opinion it will be impossible to evolve a general method or formula for calculating evapotranspiration which is based on temperature and disregards available energy.

7 Actual evapotranspiration

The *P.E.* calculated in the preceding sections applies to a closed, level cover of vegetation of considerable horizontal extent. Since reflectivity for solar radiation, and emissivity for long-wave radiation, are about the same for different closed vegetations, almost the same energy is available per unit of surface for evaporation and for the heating of soil and air, irrespective of the plant cover. A considerable fraction of this energy is consumed to provide the latent heat of evaporation; it is 60 % or more in the tropics, and on bright summer days in moderate latitudes, provided the wind velocity is low in both cases. Therefore *P.E.* cannot increase appreciably in the case of a higher wind velocity, and the influence of wind on transpiration does not need to be known to a high degree of accuracy. Actual evapotranspiration may of course be smaller, owing to regulation by the plant and lack of water.

The situation is, however, quite different as regards a nonhomogeneous vegetation. If the plants are scattered in groups they may intercept more solar radiation and receive extra energy from the air flowing past them [5]). For the same reason projecting plants or shrubs may transpire far more than the underlying plants. This is not a property of their length itself, but due to the fact that they receive extra energy. An extreme case is presented by a group of trees or shrubs in desert country. The actual transpiration may exceed the *P.E.* by several times when *P.E.* is calculated by the methods discussed above. PENMAN has already pointed out the importance of the shape of the vegetation cover, and reported a transpiration for maize amounting to about twice that of a closed grass cover if the transpiration of the maize is calculated only for the overgrown surface (PENMAN (1948), discussing experiments at Pusa).

The present authors are of the opinion that the most promising approach to a reliable method of calculating actual evapotranspiration of a tract of vegetation of a given shape is to consider actual transpiration, E, as a product of two factors, i.e.:

$$E = A.B.$$

The factor B is the evaporation of a body with a wet surface, and of a shape similar to that of the evaporating parts of the plants or plant cover, and receiving the same energy. This factor can be approximately calculated for an arbitrary shape by applying the laws of heat exchange to the atmospheric conditions under consideration.

The factor A is a reduction factor the value of which is determined by the physiology of the plant, e.g., the diffusion resistance in the stomata. The factor A will thus depend on the temperature, and perhaps on other climatological factors. Moreover, the tension of the water in the soil, and the treatment given to the plants, such as cutting of leaves, will have an influence. The maximum value of A under the climatological conditions prevailing when water is abundant would correspond to a maximum value of E, and this is the potential evapotranspiration for a vegetation cover of the shape under consideration.

APPENDIX

In this appendix some additional information is given concerning the calculations, and the sources of the meteorological data which served as a basis for these calculations are

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[5]) Advective heat will also affect the evaporation values obtained from an evaporation tank, especially where evapotranspiration from the surroundings is much less than the potential evapotranspiration. In this case the tank values may be appreciably higher than the *P.E.* calculated by the foregoing methods. The values from evaporation tanks in Australia reported by PRESCOTT (1943) constitute an example.

referred to. The complete calculations for average conditions on the circle of latitude of 40° N are given as an example.

If negative values of the monthly temperature occur, the terms $\vartheta^{1.514}$ in the expression $I = \Sigma\, \vartheta^{1.514}$ in THORNTHWAITE's formula assume a complex value. THORNTHWAITE gives no explicit directions as to how this case should be dealt with, but from examples presented in the Interim Reports of the Laboratory of Climatology at Seabrook, and from some values given by Miss SANDERSON[6] (1950) it was concluded that, for months when $\vartheta < 0$ both the contribution of this month in the expression for I and the value of $P.E.$ are assumed to be equal to zero.

In the calculations according to PENMAN the intensities of the net solar radiation received by the earth's surface, R_C, and of the net long-wave radiation emitted by the earth R_B, were calculated with the help of the semi-empirical formulae given by PENMAN (1948) (see the list at the end of this paper for the meaning of the symbols and for the units used).

$$(3) \qquad R_C = (1-r)\, R_A \quad (0.18 + 0.55\ n/N),$$

$$(4) \qquad R_B = \sigma\, T_a^{\,4}\ (0.56-0.092 \sqrt{e_d}\,)\ (1-0.9\ m).$$

For the reflection coëfficient of short wave radiation, r, the values 0.05 and 0.20 were substituted in calculating E_0 and $E_T{}'$ respectively.

The values of S and D are calculated from the following equations (PENMAN and SCHOFIELD (1950)):

$$(5) \qquad S = L_a\ /(L_a\ + L_s\ ),$$

with $L_a = 0.65\ (1 + 0.54\ u_2)$

$$(6) \qquad D = \frac{N}{24} + \frac{T_a\,(\text{max.}) - T_a\,(\text{min.})}{2\pi\ (\overline{T}_a - \overline{T}_d\,)}\ \sin \frac{N\pi}{24}$$

PENMAN gives 0.16 cm as a probable value of L_s, and this value has been adopted in all our calculations. For the ratio $\left\{T_a\ (\text{max.}) - T_a\ (\text{min.})\right\} /(\,\overline{T}_a - \overline{T}_d\,)$ the value 2 has been substituted. This value provides a reasonable estimate, and possible deviations from it will have no significant effect on the resulting values of $E_T{}'$.

In PENMAN's calculations the heat flux in the soil, Q, is ignored. In principle the quantity H_T can be easily corrected for this heat flux. To show that our conclusions are not affected by such a correction, we have applied it to calculations for a "normal" year at De Bilt, using the following formula for Q:

$$Q = A \sqrt{\lambda C \omega}\ \ \sin\,(\omega t + \pi/4),$$

while $A \sin \omega t$ represents the temperature wave. With $A = 7.9°$ C, $\lambda = 346$ cal/cm/day/°C and $C = 0.5$ cal/cm³/°C the amplitude of Q becomes 13.6 cal/cm²/day. The results are shown in fig. 5.

In the calculations for different latitudes the yearly variation of temperature was also considered to be sinusoidal. The average temperatures and the amplitudes were taken from the article by CONRAD (1936) in the "Handbuch der Klimatologie" Vol I, Part B. The following expressions for the temperatures were obtained:

$$20°\ \text{N}:\ \vartheta = 25.0 + 4.0\ \sin \omega t$$
$$40°\ \text{N}:\ \vartheta = 14.5 + 9.5\ \sin \omega t$$
$$55°\ \text{N}:\ \vartheta = 2.4 + 13.3\ \sin \omega t$$

The averages of cloudiness and relative humidity in the calculations for these latitudes were also taken from CONRAD (l.c.). No zonal averages of the wind velocity at 2 m (u_2) could be found in the literature of the subject. From available data the average of u_2 was estimated as 2.5 m/sec for 20° N and 40° N, and 3.0 m/sec for 55° N. In view of the uncertainty inherent in PENMAN's empirical expression for the factor denoting the influence of wind velocity, and in view of the approximate character of our computations, the yearly variation of u_2 was left out of consideration.

The meteorological data for the different stations were also taken from the "Handbuch der Klimatologie" (mainly from the article of WARD et al. (1938), Vol II, Part J), except for the data pertaining to De Bilt, which were derived from the monthly reports of the "Koninklijk Nederlands Meteorologisch Instituut". The data on solar radiation for Djakarta were published by DEE and REESINCK (1951).

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[6]) The values of measured evapotranspiration in this article also indicate that this quantity is in phase with insolation.

Example of calculations.

Calculations according to THORNTHWAITE *(40° N).*

	J	F	M	A	M	J	J	A	S	O	N	D	Y
ϑ °C	5.0	6.3	9.75	14.5	19.25	22.7	24.0	22.7	19.25	14.5	9.75	6.3	14.5
$(\vartheta/5)^{1.514}$	1.00	1.42	2.75	5.01	7.70	9.88	10.75	9.88	7.70	5.01	2.75	1.42	65.28
PE^*	1.1	1.5	3.0	5.4	8.4	10.7	11.7	10.7	8.4	5.4	3.0	1.5	—
PE (cm/month)	0.9	1.25	3.1	6.0	10.4	13.4	14.9	12.6	8.7	5.2	2.5	1.2	80.2
PE (mm/day)	0.3	0.45	1.0	2.0	3.4	4.5	4.8	4.1	2.9	1.7	0.8	0.4	—

Calculations according to PENMAN *(40° N).*

	J	F	M	A	M	J	J	A	S	O	N	D
ϑ °C	5.0	6.3	9.75	14.5	19.25	22.7	24.0	22.7	19.25	14.5	9.75	6.3
R_A (cal/cm²/day)	358	535	663	845	930	1000	943	841	719	525	396	320
$n/N = 1-m$	0.45	0.47	0.47	0.47	0.48	0.50	0.52	0.56	0.56	0.51	0.49	0.45
$R_C^{(O)}$ $(r = 0.05)$	146	222	276	352	393	433	418	390	333	229	170	130
$R_C^{(I)}$ $(r = 0.20)$	123	187	232	297	331	365	352	328	280	193	143	109
σT^4 (cal/cm²/day)	704	716	750	805	864	907	923	907	864	805	750	716

relative humidity	*0.75*	*0.76*	*0.77*	*0.76*	*0.75*	*0.72*	*0.69*	*0.68*	*0.69*	*0.72*	*0.74*	*0.75*
e_d (mm Hg)	4.9	5.5	7.0	9.4	12.6	14.9	15.4	14.1	11.6	8.9	6.7	5.4
R_B (cal/cm²/day)	125	129	123	116	106	102	104	116	128	128	130	124
$H_O = R_C^{(O)} - R_B$	21	93	153	236	287	331	314	274	205	101	40	6
$H_T = R_C^{(T)} - R_B$	–2	58	109	181	225	263	248	212	152	65	13	–15
L (cal/g)	593	592	591	588	585	584	583	584	585	588	591	592
Δ (mm Hg/°C)	0.46	0.49	0.61	0.80	1.05	1.26	1.34	1,26	1.05	0.80	0.61	0.49
$\Delta H_O/0.1\ L$ (mm/day)	0.16	0.77	1.58	3.21	5.15	7.15	7.22	5.91	3.66	1.38	0.41	0.05
$\Delta H_T/0.1\ L$ (mm/day)	–0.016	0.48	1.12	2.46	4.04	5.68	5.70	4.58	2.73	0.88	0.13	–0.12
γE_a ($u_2 = 2.5$ m/sec)	0.64	0.68	0.84	1.20	1.68	2.32	2.76	2.64	2.08	1.40	0.96	0.72
S ($u_2 = 2.5$ m/sec)	0.63	0.63	0.63	0.63	0.63	0.63	0.63	0.63	0.63	0.63	0.63	0.63
N (h)	9.5	10.5	11.7	13.1	14.2	14.7	14.5	13.6	12.3	11.0	9.8	9.2
D	0.70	0.75	0.81	0.87	0.90	0.92	0.91	0.88	0.84	0.78	0.71	0.68
$\Delta H_O/0.1\ L + \gamma E_a$	0.80	1.45	2.42	4.41	6.83	9.47	9.98	8.55	5.74	2.78	1.37	0.77
$\Delta H_T/0.1\ L + \gamma E_a$	0.62	1.16	1.96	3.66	5.72	8.00	8.46	7.22	4.81	2.28	1.09	0.60
$\Delta + \gamma$	0.95	0.98	1.10	1.29	1.54	1.75	1.83	1.75	1.54	1.29	1.10	0.98
$\Delta + \gamma/SD$	1.57	1.52	1.57	1.69	1.91	2.10	2.19	2.13	1.97	1.79	1.70	1.63
E_O (mm/day)	0.84	1.5	2.2	3.4	4.4	5.4	5.4	4.9	3.7	2.2	1.2	0.79
E_T (mm/day)	0.40	0.76	1.2	2.2	3.0	3.8	3.9	3.4	2.4	1.3	0.64	0.37

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List of symbols and units

Symbol		Unit
A	— amplitude of annual variation of surface temperature . . .	°C
C	— volumetric heat capacity of soil	cal/cm³/°C
D	— factor denoting influence of length of day (eq. 6)	—
e_a	— saturation vapour pressure at air temperature	mm Hg
e_d	— saturation vapour pressure at dew point	mm Hg
E	— actual transpiration	mm/day
E_0	— evaporation from water surface	mm/day
E_a	— auxiliary quantity $= 0.35\ (1 + 0.54\ u_2)\ (e_a - e_d)$	mm/day
E_T, E_T'	— potential evapotranspiration, according to PENMAN's theory *)	mm/day
H_0	— net gain in radiation energy per unit of water surface . .	cal/cm²/day
H_T	— net gain in radiation energy per unit of land surface . .	cal/cm²/day
I	— heat index, after THORNTHWAITE $= \Sigma\ (\vartheta/5)^{1.514}$	—
L	— latent heat of water vapour	cal/g
L_a	— effective diffusion length in air	cm
L_s	— effective diffusion length in plant	cm
m	— fraction of sky covered by cloud	—
n	— duration of bright sunshine	h
N	— length of day	h
$P.E.$	— potential evapotranspiration, usually according to THORNTHWAITE's formula	mm/day or cm/month
$P.E.^*$	— unadjusted $P.E.$, according to THORNTHWAITE	cm/month
Q	— heat flux into the soil	cal/cm²/day
r	— reflection coëfficient for solar radiation	—
R_A	— ANGOT's value of short-wave radiation flux	cal/cm²/day
R_B	— net long-wave radiation flux at the earth's surface . . .	cal/cm²/day
R_C	— net short-wave radiation flux at the earth's surface . . .	cal/cm²/day
S	— factor denoting influence of diffusion resistance (eq. 5) . .	—
t	— time	h, day, month
T_a	— absolute air temperature at screen level †)	°K
T_d	— absolute dew point temperature at screen level †)	°K
u_2	— wind velocity at height of 2 m	m/sec
γ	— psychrometric constant $= 0.49$	mm Hg/°C
Δ	— de/dT	mm Hg/°C
ϑ	— mean monthly air temperature	°C
λ	— thermal conductivity of soil	cal/cm/day/°C
σ	— STEFAN-BOLTZMANN constant	cal/cm²/day/°K⁴
ω	— circle frequency	day⁻¹, month⁻¹

*) E_T' is found from eq. 2; E_T is found by multiplying E_0 by PENMAN's empirical reduction factors.

†) A bar above the symbol denotes an average value.