COMMISSIE VOOR HYDROLOGISCH ONDERZOEK TNO

COMMITTEE FOR HYDROLOGICAL RESEARCH TNO

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EVAPORATION IN RELATION TO HYDROLOGY

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EVAPORATION IN RELATION TO HYDROLOGY

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EVAPORATION UNDER PRACTICAL CONDITIONS

L. WARTENA

1. INTRODUCTION

In meteorology some topics enjoy particular interest from other disciplines. The interest of hydrologists for the evaporation of water is a case in point, understandably and rightly so.

In fact, over the last few decades, hydrology has clearly done more than using meteorological knowledge thus offered or otherwise available. Hydrology has distinctly stimulated the development of evaporation research. That has evidently influenced the very character of this research. Today's papers will focus on hydrological application. It is gratifying to note that a special day can be devoted to hydrological aspects of evaporation science. It is an indication that, over the last twenty years or so, the stimulus has worked well. The previous technical meeting on evaporation was held in 1958; it covered three days. That meeting spent a considerable part of the time in describing the physical processes underlying evaporation.

2. PROBLEM FORMULATION

Today's emphasis will be on the empirical cum practical approach. In 1958, the physical models had hardly ever been tested in terms of hydrological problems. Generally speaking, there were no basic data enabling the application of models, many requisite types of measuring were too complicated, and the pertinent instruments were not suitable for long use in a catchment area, under trying weather conditions and without supervision. Attention has meanwhile been paid to all these points, and results have been forthcoming. Knowledge has deepened. More robust sensors have been developed, and nowadays data can in situ be processed successfully. The expectable came true: The process of evaporation was found to be more complex than first thought, and measuring turned out to be more difficult. In other words, the user of meteorologic knowledge still had to exercise some patience.

3. USERS OF EVAPORATION DATA

Incidentally, the very questions from users are rather divergent and, therefore, do not always emphasize identical aspects.

Meteorologists themselves use evaporation data. They are principally interested in the changes that will be caused in terms of atmospheric behaviour due to the vapour input. This interalia means that the requirements of accuracy for evaporation shall be the same as those for sensible heat, and that advective transports shall not be ignored. The time-scale involved may run from a few hours to a few days. The length-scale is usually a few thousand kilometers. On the other hand, a phytopathologist usually handles timescales of one to a few hours, or sometimes days. He is generally interested in advection as relevant to distances of some meters at the borders of soil parcels. He is moreover particularly interested in the differences, higher up and lower down, of a plant or crop externally moistened by dew or rain.

An "all weather" farmer, relying on sprinkling for his crop, may well be interested in the pF of its root-zone. He want to know evaporation values for his soil plots, and is less interested in boundary effects or divergency of evaporation. But he does want to know whether evaporation is highly affected by environmental factors. Neither is he interested in sensible heat transmission. That, if for instance, there is a much higher or a much lower crop, all sorts of advective influences play a role. The time-scale such a farmer will use in his thinking tends to cover a few days, perhaps a few weeks.

The wishes of the hydrologist are different again. Research workers trying to solve rather fundamental problems cannot deal with all sub-problems simultaneously. Therefore: a little more patience, please. However, a hydrologist cannot keep waiting indefinitely. He really needs evaporation figures. Therefore, he will tend to gamble, if necessary, using a fair amount of physical sense in the process. Just the same, there are ways that lead more quickly to usable results; I mean the empirical techniques. The more these empirical models contain some basic physical elements, the smaller the hazard of unpleasant surprise.

4. CENTRAL THEME OF THE MEETING

Meteorologists thus interested in hydrology, and hydrologists comparably familiar with meteorology, have evolved, along the route I just sketched, methods for calculation and measurement that are useful in present practice. This, I suggest, is today's central theme; it is to be considered by De Bruin, a meteorologist with hydrologic affinities that is known in this group of scientists, and by Stricker and Van Bakel, two hydrologists oriented in terms of meteorology.

Application research is by definition much needed, as the three central presentations will show.

I briefly mentioned the physical backgrounds of empirical formulae. Fundamental research on evaporation is continued, and that is a good thing for al its users. Essential shortcomings of empirical models are liable to show up after years of application. The culprit is often found to be a narrowness of physical base. That is the price eventually to be paid for accepting an empirical approach. In the event, hopes will have to be set on renewed joint thinking. Actually, one should be more wary in this respect now that twenty years ago. At the time, most of the evaporation research activities were performed to benefit hydrology and quite a number of hydrologists were actively engaged. The complexity of inherent problems, then experienced, led to specialisation. Internal turbulent boundary layers, roughness transitions, self preserving flows, coherent turbulence structures, all those are some of the fields now under investigation through evaporation research that corrects or diverts its boundaries.

There is a danger at present that, in the event of shortcomings shown up, the connection with meteorologists engaged in evaporation research is missed. And, also, that one then tries to amend the empirical formulae without making any essential correction. In other words, with only little change of attaining limited success.

Before embarking on the items of practical approach, which I specified earlier, Mr. Keyman will deal with the physical backgrounds of evaporation. He will restrict himself to the fundamental science used in the operational methods featuring later in today's programme.

THEORETICAL BACKGROUND OF SOME METHODS FOR THE DETERMINATION OF EVAPORATION

J.Q. KEIJMAN

SUMMARY

The theoretical foundations of the methods for the determination of evaporation are balance equations and transport equations. Of the balance equations only the energy balance of the earth's surface is treated. The transport equations are based on the similarity theory of Monin-Obukhov. The concept of a diffusion resistance is introduced in the transport equations.

A powerfull method for the determination of evaporation is the combination of the energy balance and the transport equations. The combination method was introduced by Penman. The method has been extended by a.o. Monteith, Rijtema and Thom-Oliver. In many cases the necessary input data for this method are lacking. Then one can use simplified versions of which those of Priestley-Taylor and Makkink are well-known examples. Evaporation from forests requires a special approach owing to the great aerodynamic roughness of a forest and the large fraction of intercepted precipitation.

1. INTRODUCTION

The purpose of this lecture is to give a survey of the equations which will be used in the subsequent lectures. The theoretical foundations of the methods for the determination of evaporation are twofold: on the one hand one has balance equations and on the other hand transport equations. With regard to balance equations we will confine ourselves to the energy balance of the earth's surface. The water balance will not be considered. Subsequently the equation for the energy balance of the earth's surface, transport equations and combination methods will be treated.

2. THE ENERGY BALANCE OF THE EARTH'S SURFACE

The energy budget of the earth's surface can be expressed by

$$Q^* = H + \lambda ET + G \tag{1}$$

where	Q^*	is the net radiation,	$(W.m^{-2})$
	H	the sensible heat flux,	$(W.m^{-2})$
	ET	the evapotranspiration,	$(kg.m^{-2}.s^{-1})$
	λ	the latent heat of vaporisation of water,	$(J.kg^{-1})$
	λET	the latent heat flux,	$(W.m^{-2})$
	G	the heat flux into the soil.	$(W.m^{-2})$

Fluxes towards the surface are given a positive value.

The net radiation, also called radiation balance, is made up of the following components:

$$Q^* = K^* + L^* = K^{\downarrow} - K^{\uparrow} + L^{\downarrow} - L^{\uparrow} \qquad (2)$$

Here K means short wave and L long wave electromagnetic radiation. The arrow $\downarrow$ denotes incoming and the arrow $\uparrow$ outgoing radiation. The short wave radiation is emitted by the sun. It is confined to the wavelength interval from 0.15 to 4.0 μm. The long wave radiation is emitted by the earth's surface and certain components of the atmosphere as cloud particles and the gases water vapour and carbon dioxide. This radiation has a wave length between 4.0 and 50 μm. The quantity $K^{\downarrow}$ is called global radiation. About 45% of it consists of visible light with a wave length between 0.40-0.74 μm. $K^{\uparrow}$ is the fraction of $K^{\downarrow}$ which is reflected at the earth's surface, thus

$$K^{\uparrow} = rK^{\downarrow} \qquad (3)$$

where r is the reflection coefficient of the surface. Then the net short wave radiation can be written as

$$K^* = K^{\downarrow}(1-r) \qquad (4)$$

The long wave radiation emitted by the earth's surface can be expressed by

$$L^{\uparrow} = \epsilon \sigma T_o^4 \qquad (5)$$

where ϵ is the emissivity of the surface, σ the Stefan-Boltzmann constant (5.67×10^{-8} $W.m^{-2}$ $.K^{-4}$) and T_o the temperature of the surface (K). Many natural surfaces can be consisdered black surfaces, which implies that $\epsilon = 1$. The downward long wave radiation, $L^{\downarrow}$, emitted by the atmosphere, is usually smaller than $L^{\uparrow}$. The difference of these two radiation fluxes is called the net long wave radiation:

$$L^* = L^{\downarrow} - L^{\uparrow} \qquad (6)$$

An example of the daily course of the components of net radiation is given in fig. 1. Average values are given in table 1.

The figure shows that the variation of the long wave components is very small. The component $L^{\downarrow}$ decreases slightly as a consequence of decreasing cloudiness. The component $L^{\uparrow}$ has a maximum in the middle of the day because surface temperature is then maximal. The variation of $L^{\downarrow}$ and $L^{\uparrow}$ is so small because they are mainly a function of the absolute temperature. The erratic course of the global radiation $K^{\downarrow}$ is due to variations in cloudiness. As the reflection coefficient r is in good approximation a constant during the day, $K^{\uparrow}$ shows the same course as $K^{\downarrow}$.

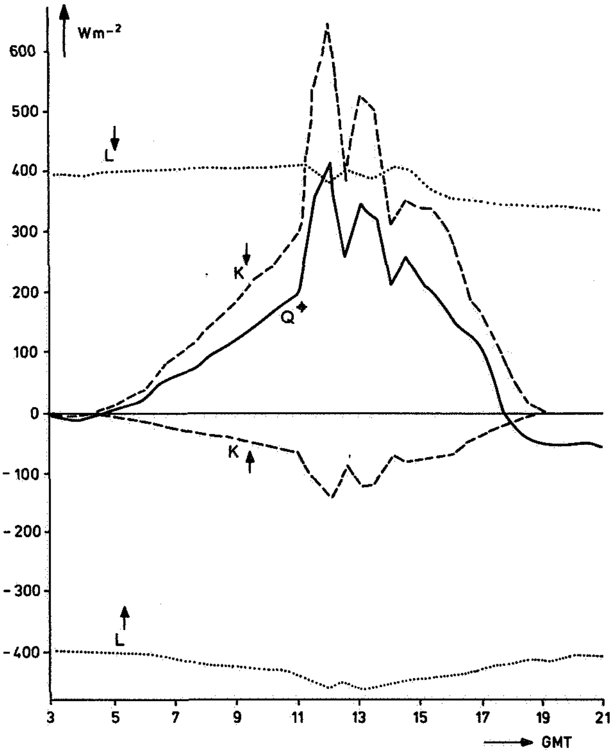

Fig. 1 Example of the diurnal variation of incoming and outgoing short wave ($K^{\downarrow}$, $K^{\uparrow}$) and long wave ($L^{\uparrow}$, $L^{\downarrow}$) radiation. Q* is the resulting net radiation

Table 1 Day-time and daily averages of the components of the net radiation above a grass cover measurred 14 Aug. 1980 at Cabauw. Day-time is period with $K^{\downarrow} > 0$.

	day-time averages	daily averages
$K^{\downarrow}$	214 $W.m^{-2}$	138 $W.m^{-2}$
$K^{\uparrow}$	49 $W.m^{-2}$	30 $W.m^{-2}$
$L^{\downarrow}$	386 $W.m^{-2}$	378 $W.m^{-2}$
$L^{\uparrow}$	422 $W.m^{-2}$	413 $W.m^{-2}$
Q*	131 $W.m^{-2}$	74 $W.m^{-2}$

The net radiation can be measured directly with a net radiometer or net pyrradiometer. It has a sensor consisting of a small blackened plate equipped with a thermopile in such a way that one set of junctions is in contact with the upper face while the other set is attached to the lower face. When the plate is aligned parallel to the surface, the temperature difference between the two faces is proportional to difference of the incoming, $K^{\downarrow}$ + $L^{\downarrow}$, and the outgoing, $K^{\uparrow}$ + $L^{\uparrow}$, radiation fluxes. So Q* is proportional to the thermopile output. This factor of proportionality depends however on the wind speed over the plate. To overcome this effect the plate is either forcefully ventilated at a constant rate or protected by hemispheric domes of polyethylene. This material is virtually transparent to radiation with wave lengths in the band of 0.3 to 100 μm.

If no direct measurements of Q* are available, it can be estimated from standard weather data. These data are sunshine duration or cloud cover, temperature and humidity of the air at screen height (see e.g. De Bruin and Kohsiek, 1979; Oke, 1978).

The soil heat flux G is usually measured with a heat flux plate similar to that used to measure net radiation. The plate is placed horizontally in the soil near the surface. The heat flux through the plate is approximately equal to the flux in the soil at the same depth. So the output of the thermopile is proportional to the heat flux in the soil. The factor of proportionality has to be determined experimentally.

With Q* and G measured as described above, one can arrive at the evaporation in a number of ways which will be treated in the subsequent sections. First the transport equations for the fluxes of sensible and latent heat will be reviewed.

3. TRANSPORT EQUATIONS

The transport of sensible heat and water vapour can be calculated from the vertical gradients of wind speed, temperature and water vapour. These gradients are related to the corresponding fluxes by the so-called flux-profile relations which are based on the similarity theory of Monin-Obukhov (1954):

$$\frac{u(z)}{u_*} = \frac{1}{k}\{\ln(\frac{z}{z_{om}}) - \psi_m(\frac{z}{L_s}) + \psi_m(\frac{z_{om}}{L_s})\} \quad (7)$$

$$\frac{T(z)-T_o}{T_*} = \frac{1}{k}\{\ln(\frac{z}{z_{oh}}) - \psi_h(\frac{z}{L_s}) + \psi_h(\frac{z_{oh}}{L_s})\} \quad (8)$$

$$\frac{e(z)-e_o}{e_*} = \frac{1}{k}\{\ln(\frac{z}{z_{ov}}) - \psi_v(\frac{z}{L_s}) + \psi_v(\frac{z_{ov}}{L_s})\} \quad (9)$$

where u(z), T(z) and e(z) are wind speed ($m.s^{-1}$), temperature (K) and vapour pressure (mbar) respectively, k the Von Kármán constant, z_{om}, z_{oh} and z_{ov} the roughness lengths for momentum, sensible heat and water vapour transport respectively (m), L_s the Obukhov length (m), ψ_m, ψ_h and ψ_v stability functions. The quantities u_*, T_*, e_* and L_s are defined by:

$$\left.\begin{aligned}
u_* &= \sqrt{\tau/\rho} \\
T_* &= -\frac{H}{\rho c_p u_*} \\
e_* &= -\frac{\gamma\lambda E}{\rho c_p u_*} \\
L_s &= -\frac{T}{g}\frac{\rho c_p}{k}\frac{u_*^3}{(H + 0.61 c_p T E)}
\end{aligned}\right\} \qquad (10)$$

where	τ	is the surface stress,	$(N.m^{-2})$
	ρ	the density of the air,	$(kg.m^{-3})$
	c_p	the specific heat of the air,	$(J.kg^{-1}.K^{-1})$
	T	the absolute temperature,	(K)
	g	the acceleration due to gravity, and	$(m.s.^{-2})$
	γ	approximately a constant.	(0.66 mbar K^{-1} at sea level)

About the roughness length z_{om} much information is available. For a vegetated surface it can be estimated with $z_{om} = 0.13$ h where h is the vegetation height (Monteith, 1973; Brutsaert, 1975). About z_{oh} and z_{ov} much less is known and in many cases one simply assumes $z_{oh} = z_{ov} = z_{om}$. This assumption gives comparatively small errors in combination formulas (Thom, 1972).

The stability functions ψ_m and ψ_h have been determined by experiment (Businger et al., 1971; Paulson, 1970; Dyer, 1974). See also Wieringa (1980) for a revaluation of the results of Businger et al. (1971). Generally one assumes $\psi_v = \psi_h$. This will also be done in this lecture.

If the surface is covered by vegetation, the zero plane displacement d has to be introduced and z-d must be taken as height of measurement instead of z. In practice for a wide range of crops the value of d is approximately given by d = 0.7 h (Monteith, 1973; Brutsaert, 1975).

With (7), (8) and (10) the flux of sensible heat can be determined if the temperature at the heights z_1 and z_2 and the wind speed at the height z_3 are measured. Then one has

$$H = -\rho c_p u_* \{T(z_2) - T(z_1)\} \{\ln(\frac{z_2}{z_1}) - \psi_h(\frac{z_2}{L_s}) + \psi(\frac{z_1}{L_s})\}^{-1} \qquad (11)$$

$$u_* = k u(z_3) \{\ln(\frac{z_3}{z_{om}}) - \psi_m(\frac{z_3}{L_s}) + \psi_m(\frac{z_{om}}{L_s})\}^{-1} \qquad (12)$$

By iteration H and u can be determined from (11) and (12). Similarily to (11) the flux of latent heat is given by

$$\lambda ET = -\frac{\rho c_p k u_*}{\gamma} \{e(z_2) - e(z_1)\} \{\ln(\frac{z_2}{z_1}) - \psi_h(\frac{z_2}{L_s}) + \psi_h(\frac{z_1}{L_s})\}^{-1} \tag{13}$$

From (11) and (13) it follows that the Bowen ratio β, regardless of stability, is given by

$$\beta = \frac{H}{\lambda ET} = \gamma \frac{T(z_2)-T(z_1)}{e(z_2)-e(z_1)} \tag{14}$$

With Q^* and G measured as described before and β from (14), one has for the flux of latent heat

$$\lambda ET = \frac{Q^* - G}{1+\beta} \tag{15}$$

This method for the determination of λET is usually called the energy balance/Bowen ratio method. It is a reliable method and often used as a reference.

In fig. 2 an example is given of the energy balance of a grass cover. There was no shortage of water in the soil. We see that λET is much greater than H. The soil heat flux G is very small compared with Q^*. It is often neglected or estimated as a small fraction of Q^*. Fig. 3 refers to a situation with a shortage of soil water. Now H is during a part of

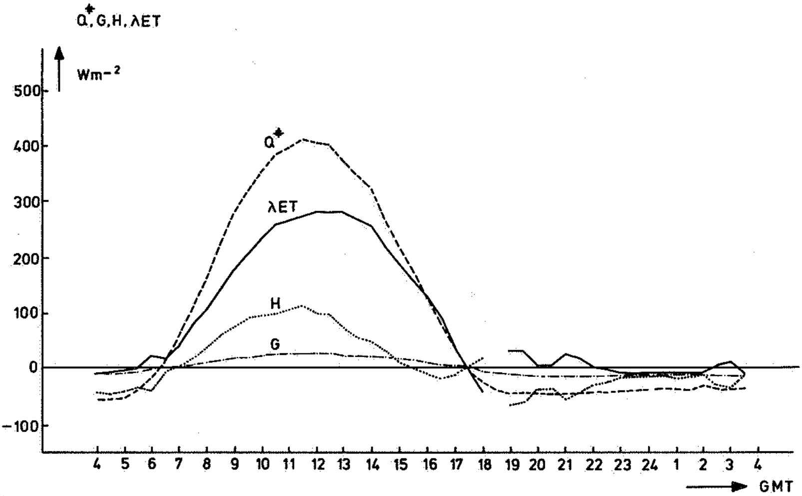

Fig. 2 Example of the diurnal variation of the 4 terms of the energy balance of a grass cover well supplied with water.
Q^* is the net radiation,
λET and H the fluxes of latent and sensible heat and
G the soil heat flux (from De Bruin and Kohsiek, 1977).

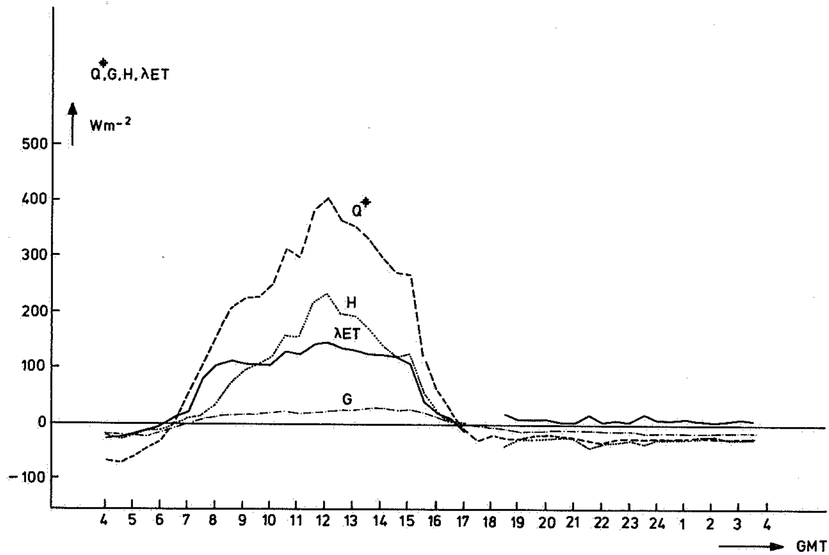

Fig. 3 The same as fig. 2, but the soil drying out.

the day greater than λET. The evaporation reaches early in the morning a certain value which cannot be exceeded because of the restricted water supply to the roots.

The concept of transport resistance is often introduced in combination formulas. One can write the equations for H and λET as

$$H = \rho c_p \frac{T_o - T(z)}{r_{ah}} \tag{16}$$

$$\lambda ET = \frac{\rho c_p}{\gamma} \frac{e_o - e(z)}{r_{av}} \tag{17}$$

where r_{ah} and r_{av} are the resistance to the transport of heat and water vapour in the air layer between the surface and the height of measurement z, $(s.m^{-1})$

T_o the surface temperature (K)

and

e_o the vapour pressure at the surface. (mbar)

Comparing (16) and (17) with (8) and (9) and assuming $z_{oh} = z_{ov} = z_{om}$, gives

$$r_{ah} = r_{av} = \{\ln(\frac{z}{z_{om}}) - \psi_h(\frac{z}{L_s}) + \psi_h(\frac{z_{om}}{L_s})\}/ku_* \tag{18}$$

The influence of (in)stability on the transports is small for wind speeds greater than $3\ m.s^{-1}$ (De Bruin and Kohsiek, 1979). Then it follows from (18) and (7), assuming $\psi_m = \psi_h = 0$,

$$r_{ah} = r_{av} = \{\frac{1}{k} \ln(\frac{z}{z_{om}})\}^2/u(z) \tag{19}$$

The errors introduced by the assumption $z_{oh} = z_{ov} = z_{om}$ are not well known at this moment. The quantity z_{om} is for a rough surface probably larger than z_{oh}. If one introduces the quantity B^{-1} (Owen and Thomson, 1963) defined by $\ln(z_{om}/z_{oh}) = k\,B^{-1}$, then (19) can be rewritten as

$$r_{ah} = r_{av} = \frac{\ln(\frac{z}{z_{om}})\ \{\ln(\frac{z}{z_{om}}) + k\,B^{-1}\}}{k^2 u(z)} \tag{20}$$

According to Thom (1972) $B^{-1} = 6.27\ u_*^{1/3}$ (u_* in $m.s^{-1}$) for many types of vegetated surfaces. This gives an increase in r_h in many cases of 30-40% compared with (19). The magnitude of B^{-1} is rather uncertain at this moment.

The vapour pressure at the surface, e_o, is practically always unknown. But one may assume that in the stomatal cavities of the plant leaves the vapour pressure has the saturation value at the surface temperature T_o: $e_s(T_o)$. We can now introduce a canopy resistance r_c defined by

$$\lambda ET = \frac{\rho c_p}{\gamma} \frac{e_s(T_o) - e_o}{r_c} \tag{21}$$

Eliminating e_o from (17) and (21) gives

$$\lambda ET = \frac{\rho c_p}{\gamma} \frac{e_s(T_o) - e(z)}{r_c + r_h} \tag{22}$$

4. COMBINATION METHODS

By combining the energy balance equation (1) with the transport equations (16) and (22) one can derive an expression for the flux of latent heat in which the surface temperature does not occur (Monteith, 1965; Rijtema, 1965):

$$\lambda ET = \frac{s(Q^* - G) + \rho c_p\,[e_s\{T(z)\} - e(z)]/r_h}{s + \gamma(1 + r_c/r_h)} \tag{23}$$

This method is based on the work of Penman (1948). Combination equations as (23) can only be used if the vapour pressure in or at the surface has the saturation value of the surface temperature. Eq. (23) shows clearly the factors on which evaporation in nature depends: on the one hand on the available energy, Q^*-G, and on the other on the dryness of the air, $e_s\{T(z)\}-e(z)$, and the turbulent exchange which is expressed by r_h.

If a vegetation completely covers the ground and soil moisture is not a limiting factor, then we speak of the potential evaporation of that vegetation. As the quantities Q^*-G, r_c and r_h depend on the vegetation, is it clear that potential evaporation may differ from one vegetation to another. It is also possible to refer the concept of potential evaporation to a specified surface e.g. a grass cover of a certain height. See for this problem De Bruin (1981).

Eq. (23) can also be applied if soil moisture limits the evaporation. In that case r_c depends i.a. on the water status of the vegetation e.g. leaf water pressure (Van Bakel, 1981). If the vegetation is not completely closed and radiation penetrates to the soil surface, evaporation from the soil surface has to be taken into account (Van Bakel, 1981).

The conventional Penman equation can be derived from (23) by setting $G = r_c = 0$ and introducing a wind function f(u) defined by

$$f(u) = \frac{\rho c_p}{\gamma} \frac{1}{r_h}$$

Then we have

$$\lambda ET = \frac{s}{s+\gamma} Q^* + \frac{\gamma}{s+\gamma} \lambda ET_a \tag{24}$$

where s is the slope of the saturation vapour pressure curve at air temperature and λET_a is defined by $ET_a = \frac{f(u)}{\lambda} \{e_s(T_2)-e_2\}$. The first term at the right hand side of (24) is called the radiation term and the second term the aerodynamic term. ET_a is called the aerodynamic evaporation or the drying power of the air. Penman used the wind function

$$f(u) = 7{,}4\,(1 + 0.54\,u_2)\,W.m^{-2}.mbar^{-1} \tag{25}$$

With this choice of f(u) the transport resistance is

$$r_{ah} = \frac{\rho c_p}{\gamma} \frac{1}{7{,}4(1+0.54\,u_2)} \simeq \frac{250}{1+0.54\,u_2}\ s.m^{-1} \tag{26}$$

Comparing (26) at higher wind speeds with (19) shows that Penman used in fact a very small roughness length namely $z_{op} = 1.37 \times 10^{-3}$ m (Thom and Oliver, 1977). Natural surfaces have usually a much larger roughness length. Therefore Thom and Oliver proposed to introduce into (25) a factor defined by

$$m = \left\{ \frac{\ln\left(\frac{z}{z_{op}}\right)}{\ln\left(\frac{z}{z_{om}}\right)} \right\}^2 \tag{27}$$

where z_{om} is the roughness length of the surface under consideration. Instead of (25) and (26) we then have

$$f(u) = 7.4\,m\,(1 + 0.54\,u_2)\,W.m^{-2}.mbar^{-1} \tag{25a}$$

$$r_h = \frac{250}{m}\,\frac{1}{1 + 0.54\,u_2}\,s.m^{-1} \tag{26a}$$

These simple expressions also take into account the influence of stability at low wind speeds. For rural areas in Southern England Thom and Oliver estimated $m \simeq 2½$ and

$n = \frac{r_c}{r_h} \simeq 1.4$. Instead of (24) we then have

$$\lambda ET = \frac{s}{s + 2.4\,\gamma}\,Q^* + \frac{2.5\,\gamma}{s + 2.4\,\gamma}\,\lambda ET_a \tag{24a}$$

It turns out that the evaporation calculated with (24a) is often nearly equal to the evaporation calculated with (24). The effects of the introduction of m and n are often approximately selfcancelling. On the other hand the ratio of the aerodynamic term to the radiation term increases by the factor m. Therefore the approximate equality of (24) and (24a) does not apply to all circumstances. In winter Q* is very small but the aerodynamic term of (24a) cannot be neglected. For a application of (24a) with a slightly different value of m, see Stricker (1981) and De Bruin (1981).

In case of a very rough surface as a forest, (24a) can be simplified, since then $n \gg 1$ and (24a) becomes in good approximation

$$\lambda ET = \frac{\rho c_p}{\gamma}\,\frac{e_s\{T(z)\} - e(z)}{r_c} \tag{24b}$$

The application of (24b) is complicated by the fact that forests intercept a fairly large proportion of the precipitation. During the time the intercepted water evaporates, we have $r_s \simeq 0$. This means that we have to use in (24b) an effective value of r_s which decreases when the contribution of interception to total evaporation increases. Thom and Oliver (1977) deduced the relation $r_c = r_{cd}\,(1 - \frac{I}{E})$ where r_{cd} is the resistance of the dry surface and I the interception. Substitution of this expression in (24b) leads to

$$ET = \frac{\rho c_p}{\lambda\gamma} \; \frac{e_s\{T(z)\} - e(z)}{r_{cd}} + I \tag{24c}$$

Evaporation from extremely rough surfaces depends primarily on saturation deficit and interception.

An important simplification of (24) with a wide range of applicability has been introduced by Priestley and Taylor (1972). They considered first the case that the atmosphere is saturated with water vapour in the lowest layer. Then from (14) we have $\beta = \gamma/s$ and from (1) $\lambda ET = s/(s+\gamma)\,(Q^* - G)$. Then they added an experimental constant α to account for the fact that the atmosphere is usually not saturated, even in the case of large areas well supplied with water.

$$\lambda ET = \alpha \frac{s}{s+\gamma} (Q^* - G) \tag{25}$$

They found $\alpha = 1.26$. This value has been confirmed by a number of other investigators. (Ferguson and Den Hartog, 1975; Stewart and Rouse, 1976, 1977; Davies and Allen, 1973; Mukammal and Neumann, 1977). For applications of (25) in The Netherlands see De Bruin and Keijman (1979), De Bruin (1981) and Stricker (1981).

A further simplification of (24) has been proposed by Makkink (1957, 1960). As net radiation Q^* is highly correlated with global radiation $K^{\downarrow}$, eq. (25) can be approximated by

$$\lambda ET = c_1 \frac{s}{s+\gamma} K^{\downarrow} + c_2 \tag{26}$$

Values of c_1 and c_2 are given by De Bruin (1981).

From an experimental point of view it is easier to determine H from a transport equation than λE. Furthermore in saturated or nearly saturated circumstances H is small compared to λET. Therefore it is often feasible to determine the evaporation from

$$\lambda ET = Q^* - G - H \tag{27}$$

Soer (1980) used this approach and calculated H with (16) taking into account the influence of stability on r_h. The surface temperature was remotely sensed by infrared line scanning (IRLS). Another possibility is to measure air temperature at two or more heights and use eqs. (12) and (13). This method has been used by Stricker and Brutsaert (1978) for a small rural catchment.

The latter method avoids the difficulties about the quantity z_{oh} but a high instrumental accuracy is required for measuring vertical temperature differences in the air. Moreover, this is a point measurement in contrast to the measurement of surface temperature by IRLS which surveys an area.

5. CONCLUDING REMARKS

Although the theory presented in the preceding sections is based on sound physical principles, it still has serious limitations. One of these limitations is the fact that the theory is one dimensional. This means in practice that the necessary measurements have to be taken on a reasonably homogeneous field of sufficient dimensions. The surrounding fields may have different properties of soil and vegetation. So different values of the evaporation can be expected on these fields.

Another difficulty is presented by a vegetation layer. This layer is represented in the theory by a few parameters such as roughness length, zero plane displacement and surface diffusion resistance. In reality the exchange processes in a vegetation layer are too complicated to be accounted for by a few parameters. There exist of course models for these processes but they require detailed information about the vegetation which is seldom available (see Goudriaan, 1977; Legg and Monteith, 1975).

Notwithstanding these limitations and others satisfactory results can be obtained with the theory presented, as will be shown by the subsequent authors.

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THE DETERMINATION OF (REFERENCE CROP) EVAPOTRANSPIRATION FROM ROUTINE WEATHER DATA

H.A.R. DE BRUIN

SUMMARY

The first part of this paper deals with the determination of reference crop evapotranspiration, ET_o, (= the potential evapotranspiration of a short grass cover). The formulas for ET_o of Penman, Thom-Oliver, Priestley-Taylor and Makkink are considered and a comparison is made. In the second part ET_o is compared with the actual evapotranspiration ET. Finally, some measuring techniques for ET are discussed which, probably, can be applied in the future on a routine base.

1. INTRODUCTION

This paper deals with the problem of the determination of evaporative water losses from vegetative surfaces. A first analyses leads to the conclusion that, besides on weather elements, evapotranspiration depends on many other factors such as type and stage of development of vegetation, soil type and soil moisture, groundwater table etc. So, at first sight it seems to be impossible to evaluate evaporation from routine weather data only; or, as an old proverb from China says: "One can not applaud with one hand".

However, the hydrologist in the field needs evaporation data and mostly he has only routine weather data at his disposal and some rough information on soil and canopy. Under such circumstances a special type of researchers comes foreward who simply ignore the theoretical difficulties and construct straightforward ad hoc models, which often yields satisfactory results. In this paper such approaches will be treated with respect to the evapo(transpi)ration problem. In the first part the use of simple routine weather data for the determination of crop water requirements is considered. For this the so-called reference crop evapotranspiration (ET_o) is introduced. This quantity depends on weather elements only. Several methods for the evaluation of ET_o will be discussed.

In the second part of this paper ET_o will be compared with the actual evapotranspiration of grass as measured at Cabauw in The Netherlands. Finally the question will be touched if it is possible to get more information on actual evaporation when a small number of additional observations is made on routine weather stations.

2. THE ROUTINE WEATHER DATA

Almost every country in the world is a member of the World Meteorological Organization of the United Nations. This organization demands of its members, among other things, to establish in their territories a network of *synoptic* and *climatological* stations. In the *Technical Regulations* (WMO) rules are given for the meteorological quantities to be measured, methods of observation, minimum accuracy and time of observation. In this way the hydrologist has the following elements relevant for evaporation at his disposal:

- cloud amount,
- temperature at screen height,
- humidity at screen height,
- wind speed at 10 m.

Some countries, among which The Netherlands, have also established stations at which duration of sunshine and global radiation are observed. For evaporation studies we need net radiation (Keijman, 1981) which is not measured directly at a weather station. Fortunately, this quantity can be estimated from sunshine duration (or cloud cover), temperature and humidity. When available use van be made of global radiation data for these estimates.

3. REFERENCE CROP EVAPOTRANSPIRATION AND CROP WATER REQUIREMENTS

An important example of how standard weather data can be used is the determination of crop water requirements. Often, for this, the concept of "potential evapotranspiration" is introduced to account for the influence of weather on the transpiration rate of crops. Unfortunately there is confusion about the interpretation of this term. Mostly potential evapotranspiration is defined as the maximum transpiration under the given weather conditions, but there is no unanimity about the crop that is considered. One group of authors refers to the actual vegetation of interest, whereas others consider potential evapotransipiration the maximum transpiration rate from a hypothetical reference crop (which mostly is a short grass cover). To avoid this confusion, in this paper, the term potential evapotranspiration will not be used, but, the terminology of the FAO-researchers Doorenbos and Pruitt (1977) will be followed. These authors introduce the following quantities:

– *The crop water requirements* (ET_{crop}), which are defined as the depth of water needed to meet the water loss through evapotranspiration of a disease-free crop, growing in large fields under non-restricting soil conditions including soil water and fertility and achieving full production potential under the given growing environment.

– *The reference crop evapotranspiration* (ET_o), which is defined as the rate of evapotranspiration from an extensive surface of 8 to 15 cm tall, green grass cover of uniform height, actively growing, completely shading the ground and not short of water.

It is obvious that ET_{crop} depends, besides on weather, also on crop factors, such as crop height, reflectivity and shading percentage of the ground. As noted before in literature both ET_{crop} and ET_o are sometimes called "potential evapotranspiration" e.g. Van Bakel (1981) who takes $ET_{pot} = ET_{crop}$.

The introduction of ET_{crop} and ET_o is only useful when they can be applied in practice. Fortunately, it appears that the following assumptions can be made which are rather realistic in a wide range of climatological conditions.

1) The reference crop evapotranspiration (ET_o) is a function of weather elements only and is independent of plant and soil.
2) ET_o and ET_{crop} are related as follows:

$$ET_{crop} = k_c\, ET_o \tag{1}$$

where k_c is a crop factor.

3) The best results are obtained when ET_o is evaluated with the well-known Penman equation in a somewhat modified form, notably

$$ET_o = \frac{s(Q^* - G)/\lambda + \gamma\, f(u)\, (e_s(T_a) - e_a)}{s + \gamma} \tag{2}$$

where	Q^*	is the net radiation of the grass surface,	($W.m^{-2}$)
	s	the slope of the saturation water vapour pressure – temperature curve at air temperature,	($mbar.K^{-1}$)
	γ	the psychometric constant,	($mbar.K^{-1}$)
	λ	the latent heat of vaporization of water,	($J.kg^{-1}$)
	$e_s(T_a)$	the saturation vapour pressure at air temperature T_a at screen height,	(mbar)
	e_a	the actual vapour pressure at screen height,	(mbar)

$$f(u) = \frac{3.7 + 4.0\, u_2}{\lambda}{}^{☆} \quad (kg.m^{-2}.s^{-1}.mbar^{-1}) \tag{3}$$

u_2	the wind speed at 2 m,
G	the soil heat flux, and
ET_o	is expressed in $kg.m^{-2}.s^{-1}$

Note: ☆ This is the windfunction proposed by Penman and used by Kramer (1957). Strictly speaking Doorenbos and Pruitt adopted a somewhat different form of f(u).

The windfunction f(u) refers to a hypothetical water surface. This is important to note, since a grass cover has, through its greater roughness length, aerodynamic properties which leads to a higher wind function (see e.g. Keijman, 1981). Mostly the soil heat flux can be ignored.

Strictly speaking, Doorenbos and Pruitt (1977) introduced a multiplication factor in the right hand side of eq. (2). However, for the common conditions in The Netherlands this factor can be taken equal to 1.

Experimental confirmation of the above assumptions is very difficult to give, expecially, when one askes for verifications that will convince physicists. The evidence given by Doorenbos and Pruitt (1977) is certainly too meagre to be convincing for a physicist, however, since we are dealing with a practical problem, in my opinion, the experimental evidence that can be found in literature supports the assumptions made sufficiently in order to adopt them as very useful working hypotheses for practical applications.

It will be clear that the treatment of the crop factor k_c falls outside the scope of this paper. For this, the reader is referred to Doorenbos and Pruitt (1977).

It is noted that there is an alternative approach based on the work of Rijtema (1965) and Monteith (1965) for the determination of ET_{crop}. This method is e.g. mentioned by Van Bakel (1981) and explained by Keijman (1981). The advantage of this technique is that it has a certain physical basis, however, from a practical point of view, it has the disadvantage that weather and plant-soil factors are not separated explicitly as is the case in the approach of Doorenbos and Pruitt (1977).

3.1. *The routine evaporation data published by the KNMI.*

Since 1956 the Royal Netherlands Meteorological Institute (KNMI) publishes in its monthly weather review the so-called "evaporation from a free water surface" often denoted as "open water evaporation", E_o. Until 1-1-1971 monthly totals were presented of 12 stations, while after that date the evaporation amounts of 10-days periods are published. (Strictly speaking, the time intervals are not exactly 10 days; the evaporation totals are given for the first and the second 1- days of a month and of the remaining period. Thus, the latter has a variable length between 8 and 11 days.)

The determination of E_o is based on the work of Penman (1948, 1956); the calculations are carried out with eq. (2), however, with two differences (De Bruin, 1979; Buishand and Velds, 1980).

a) The actual net radiation Q^* is replaced by the net radiation of a hypothetical water surface.
b) Instead of 24-hourly averages of temperature and humidity the means of these quantities over day-time are used for the computations.

It appears that for the period April-September E_o, thus determined, is related to ET_o by the simple expression (Wesseling, 1977):

$$ET_o = 0.8\ E_o \tag{4}$$

The E_o data are published by the KNMI in its monthly weather review with a retardation of about 1,5 month. For many purposes this is too late. Therefore, the KNMI, started in 1980 with a daily "fast" evaporation data provision. Every day between 1 April till 1 November the E_o values of 5 stations of the previous day are given in the morning radio weather bulletin. The evaluation method used for these E_o values differs slightly from eq. (2), since at the time of computation no bright sunshine observations are available (De Bruin and Lablans, 1980). It is very likely that these data provision will be continued in the future.

3.2. *Long term records of E_o and rainfall*

Kramer (1956) publishes monthly values of E_o of 12 Netherlands stations for 1931-

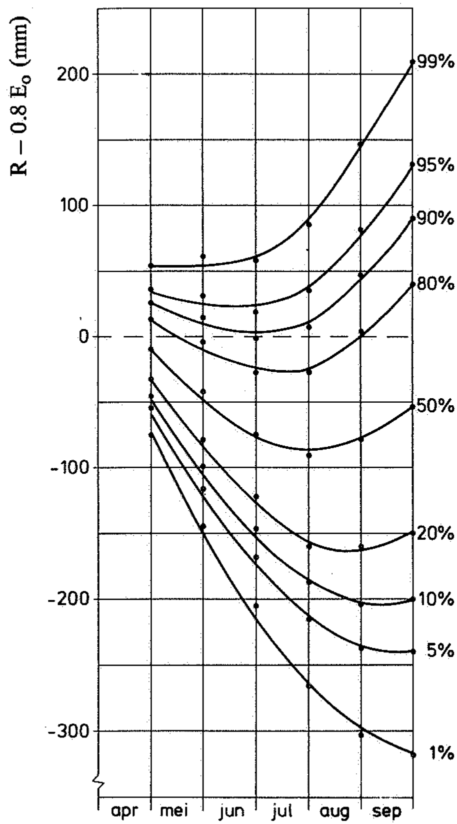

Fig. 1 Isolines for cumulative frequencies of potential precipitation surplus (R – 0.8 E_o) of De Bilt, computed from 1 April. (after Buishand and Velds, 1980)

1951. This data set has been extended by De Bruin (1979) to 1911-1975, while he added the corresponding rainfall totals (R). On the other hand the number of stations is diminished to 6, since it appeared that the quality of the basic weather observations (if present) was too poor of the other stations for 1911-1950. Moreover De Bruin (1979) presents cumulative frequency distributions of E_o, R and R -0.8 E_o (= R $-$ ET_o) for the growing-season. These are of practical importance (e.g. for the assessment of the crop water requirements) for hydrologists and agricultural researchers. A graphical representation of a set of these frequency distributions for De Bilt is given by Buishand and Velds (1980). An example is shown here in fig. 1, which refers to (R $-$ ET_o), what can be called "potential rainfall-surplus". It is seen e.g. that over the period April to August inclusive there is a probability of 1 percent that R $-$ ET_o is less or equal to about 300 mm. The six stations for which this type of information is available are

Den Helder
De Bilt
Winterswijk
Oudenbosch
Gemert en
"Avereest/Den Hulst/Wijster/Witteveen/Dedemsvaart".

3.3. *Alternative methods for the determination of ET_o*

In literature several methods can be found for the determination of ET_o which possibly can serve as an attractive alternative for eq. (2). The most simple expression is given by Makkink (1960), which can be written as:

$$\lambda ET_o = (1-r)\frac{s}{s+\gamma}\, c_1 K^{\downarrow} + c_2 \qquad (\mathrm{W.m^{-2}}) \tag{5}$$

where c_1 and c_2 are constants, $K^{\downarrow}$ is the incoming short-wave radiation (which is observed directly on a routine base at several stations in The Netherlands), and r is the reflectivity.

It is noted that $c_1 \simeq 0.9$ and c_2 is small, say 0.2 mm day^{-1}, further for grass $r \simeq 0.25$.

A second simplification of eq. (2) is given by Priestley and Taylor (1972). They arrived at

$$\lambda ET_o = 1.26\,\frac{s}{s+\gamma}\,(Q^* - G) \qquad (\mathrm{W.m^{-2}}) \tag{6}$$

Where G is the soil heat flux. Since eq. (6) refers to daily mean values, G can be ignored in eq. (6), since mostly $Q^* \gg G$. It is noted that this expression applies also to a water surface (see e.g. De Bruin and Keijman, 1979a).

Finally it is worthwhile to mention the expression of Thom and Oliver (1977) which is a variant of the Penman-Monteith-Rijtema combination equation. It reads

$$\lambda ET_o = \frac{s(Q^* - G) + m\gamma 7.4\,(1 + 0.54u_2) \cdot (e_s(T_a) - e_z)}{s + \gamma(1 + n)} \qquad (W.m^{-2}) \qquad (7)$$

where m and n are constants. The advantage of this approach is that it has a better theoretical bases than eq. (2), (see e.g. Keijman, 1981). The quantity m accounts for the greater roughness of a grass cover with respect to open water, while n describes the effect that in the case of grass the water vapour transport goes mainly through the plant's stomata resulting in an extra resistance which is not present in the case of open water.

3.4. *Comparison of the different methods*

In table 1 the different methods to determine ET_o are compared. This table contains some statistical features such as mean values, correlation coefficients and the least square estimate of the regression constant a from the regression equation $\hat{y} = ax$. Table 1 refers to daily values and is based on a set of micrometeorological data collected at Cabauw during the summer period of 1976 and 1977.

In the calculations of ET_o with the equations of Penman, Priestley and Taylor, and Thom and Oliver the measured values of $(Q^* - G)$ are used, while for the evaluation of ET_o with Makkink's formula the measured global radiation has been applied.

It was not possible to determine E_o for Cabauw, since the bright sunshine was not observed at this station. Therefore it was decided to compare E_o of De Bilt (ca 25 km north western of Cabauw) with ET_o, determined with eq. (2), of Cabauw.

The daily values of E_o are evaluated from the 10-day totals as published by the KNMI with a method proposed by De Bruin and Kohsiek (1977):

$$E_{oi} = K_i^{\downarrow} \frac{E_o^{10}}{K^{\downarrow 10}}. \qquad (8)$$

where the index i refers to the i-th day (i = 1, 2, ...) of a 10-day period, and the index "10" to the entire period of 10 days.

With respect to table 1 the following remarks can be made:

a) The interrelation between the different models for ET_o is high. The correlation coefficients for daily values (!) are all greater than 0.9.
b) When it is believed that Penman's formula [eq. (2)] gives the best results, the model of Makkink appears to be a very attractive alternative, since it has about the same skill, while it needs only the global radiation and air temperature as input.
c) In 1977 (a "normal" year) there is no significant difference between the models of Penman, Thom-Oliver and Priestley-Taylor; the concept $ET_o = 0.8\ E_o$ yields a somewhat ($\simeq 8\%$) greater value.

Table 1. Comparison of different methods for the determination of ET_o (mm day^{-1}). Data of Cabauw, except E_o.
PNM = Penman [eq. (2); ($Q^* - G$) measured],
PR–TA = Priestley-Taylor [eq. (6); ($Q^* - G$) measured],
MAK = Makkink [eq. (5), $(1-r)c_1 = 0.65$, $C_2 = 0$],
TH–OL = Thom-Oliver [eq. (7); m = 1.9, n = 1.2 (roughness length = 1 cm); ($Q^* - G$) measured],
E_o = open water evaporation of De Bilt (see text),
Slope = regression coefficient a from $\hat{y} = ax$.

a. 1976 (55 days in June-August)

x	y	$\bar{x}$	$\bar{y}$	Corr. Coef.	Slope
PR–TA	PNM	3.1	3.5	.91	1.12
MAK	PNM	3.5	3.5	.94	1.00
TH–OL	PNM	3.6	3.5	.97	. .96
0.8 E_o	PNM	3.5	3.5	.93	1.00

b. 1977 (119 days in May-September)

x	y	$\bar{x}$	$\bar{y}$	Corr. Coef.	Slope
PR–TA	PNM	2.2	2.3	.97	1.06
MAK	PNM	2.2	2.3	.93	1.02
TH–OL	PNM	2.2	2.3	.99	1.06
0.8 E_o	PNM	2.5	2.3	.91	0.92

d) In 1976 (a dry year) the Priestley-Taylor model differs from the other ones. There is evidence that under very dry conditions the latter over-estimate ET_o (see e.g. Brutsaert and Stricker, 1979) and that then, the Priestley and Taylor concept must be prefered. However, we must realize that under very dry conditions the common formula's for the estimation of net radiation are not verified, which reduces the application of the Priestley and Taylor concept, under very dry circumstances. Possibly, the method proposed by Doorenbos and Pruitt (1977) offers an alternative solution in this case. They introduce a correction factor in eq. (2) which depends, among other tings, on radiation and relative humidity. Anyhow, the reader should use the different ET_o estimates with caution when these are applied for crop water requirements calculations in very dry years. In such years an overestimation of, say, 15% can be obtained very easily.

e) There is a great resemblance between the results of eq. (2) and those of the Thom and Oliver equation. This is a typical example of how two errors can cancel each other. In e.q. (2) in the denominator the stomatal resistance is ignored, but at the same moment, in the numerator, a too low wind function is taken. Apparently, these effects counteract. So, it is rather fortuitous that eq. (2) yields such good results.

4. ACTUAL EVAPOTRANSPIRATION

As noted before, it is to be expected that the actual evapotranspiration (ET), will depend, apart on weather, on soil and plant factors, so it is unlikely that ET can be determined from weather elements only. Nevertheless, it is worthwhile to examine what is the "weight" of the weather influence. It will be clear that this "weight" is a function of plant and soil conditions, e.g. when there is no water in the soil the evapotranspiration will be zero, regardless the weather circumstances. In the next section a comparison will be made between ET and ET_o under Netherlands conditions. This will be done for an extreme dry year (1976) and a more or less normal year (1977).

4.1. *Comparison between ET and ET_o*

In table 2 a comparison is given between the actual evapotranspiration ET of a grassland and ET_o as evaluated with the models of Penman [eq. (2)] and Priestley-Taylor [eq. (6)] respectively. ET has been determined with the energy-balance method, using Bowen's ratio (De Bruin and Kohsiek, 1977). From table 2 it is seen that in the "normal" year 1977 the two ET_o estimates are only slightly greater than ET. As to be expected, in the extremely dry year 1976 ET is much less than ET_o; then Penman's equation (2) overestimates ET about 40%, whereas the Priestley-Taylor concept give an overestimation of, say, 25%.

Table 2. Comparison between the actual evapotranspiration ET ($mm.day^{-1}$) and
a) ET_o as determined with the models of Penman and Priestley-Taylor respectively, and
b) with the model for ET of Brutsaert and Stricker (BR–STR).
Data of Cabauw; daily values.

a. 1976 (55 days in June-August)

x	y	x	y	Corr. Coef.	Slope
ET_o^{PNM}	ET	3.5	2.4	.75	.68
ET_o^{PR-TA}	ET	3.1	2.4	.82	.77
ET^{BR-STR}	ET	2.7	2.4	.72	.86

b. 1977 (119 days in May-September)

x	y	x	y	Corr. Coef.	Slope
ET_o^{PNM}	ET	2.3	2.1	.89	.89
ET_o^{PR-TA}	ET	2.2	2.1	.87	.93
ET^{BR-STR}	ET	2.0	2.1	.78	.98

It is important to remark that the above results apply to the "summer" period May-September; further they are restricted to grass.

In wintertime net radiation will be small, while furthermore the canopy is often wet. This makes that then the second term of Penman's formula plays a dominant part, and a better description of the surface roughness is needed. Therefore in winter the expression of Thom and Oliver must be preferred. The implications of this for very rough wet surface is discussed by Keijman (1981).

4.2. *The advection-aridity approach*

In chapter 2 the assumption is made that ET_o is independent of the soil conditions. Under very dry circumstances this assumption is questionable. Now, the problem is that quantities such as ET_o are hypothetical, which implies that discussions on the behaviour of such quantities under rare conditions will be hypothetical themselves and are therefore rather vague and difficult to understand. Anyhow, if there is an interaction between ET_o and the soil conditions, it can be imagined that use of this interaction can be made for the determination of ET.

Because, at the moment, the exact interaction between ET_o and soil condition is unknown, the models based on this interaction must be seen primary as empirical.

Brutseart and Stricker (1979) constructed such a model and they tested it in The Netherlands. This model, which they called themselves "heuristic" reads:

$$ET = 2\,ET_o^{PR-TA} - ET_o^{PNM} \tag{9}$$

where ET_o^{PR-TA} and ET_o^{PNM} are evaluated with eqs. (6) and (2) respectively. For the very dry year 1976 the results of this approach are satisfactory, while also in 1977 the skill is rather good (see table 2). However, in normal years the random scatter is greater than that of e.g. the Penman model. This supports the conclusion that in very dry periods eq. (9) must be used, while otherwise e.g. Penman's equation must be applied. At the moment, however, no simple rules are available to decide whether it is dry or not.

5. WHAT CAN BE ACHIEVED WITH ADDITIONAL WEATHER OBSERVATIONS?

In this brief chapter we will mention some methods for the determination of actual evapotranspiration for which only a limited number of weather observations has to be done. The choice of these methods is rather arbitrary.

a. *The modified aerodynamic approach.*

Recently Stricker and Brutsaert (1978) used the so-called profile-energy-budget method for the determination of ET from grass. This method is discussed in this report by Keijman. The only quantity that has to be measured apart from the standard weather observations, is the vertical temperature gradient. De Bruin and Keijman (1979b) tested this method. They concluded that it has a good skill.

b. *The remote sensing technique.*
This method is similar to the previous one. The only difference is that the lowest level at which the temperature is measured is at the ground. The theoretical consequences are discussed by Keijman (1981). The surface temperature can be measured with a remote sensing (I-R) thermometer. It is certainly possible, that in the future this instrument can be used on a routine base.

c. *The temperature fluctuation method*
Finally I will treat a method which is based on the fact that the sensible heat flux H is an empirical function of the standard deviation of the temperature, see e.g. Businger

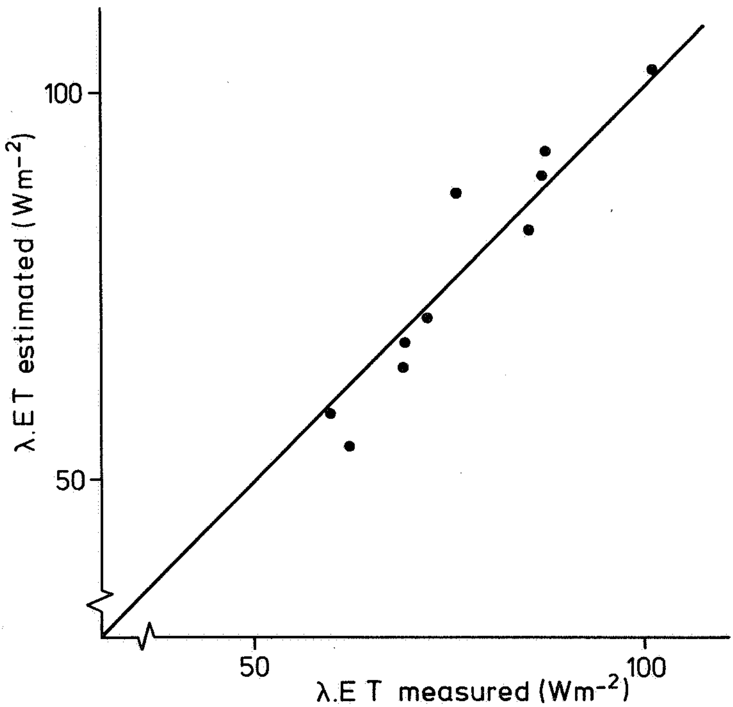

Fig. 2 Measured values of ET plotted against the computed ones using the temperature fluctuation method. (½ hourly values)

(1973). So, H can be measured with a fast-responding thermometer which is sampled, say, every second. This fast sampling is a problem at the moment, but, in the near future it will be very easy to do with a micro-processor.

After having determined H, ET can be obtained from the energy-balance equation.

At the moment the present author examines this approach. A (very, preliminary result is shown in fig. 2, where ET as measured with the energy-balance method, using Bowen's ratio, is compared with ET determined with the temperature-fluctuation method. It is seen that in this example the approach works splendid, but it is far too early to state that it is a reliable method applicable at a routine weather station.

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UNSATURATED ZONE AND EVAPOTRANSPIRATION

P.J.T. VAN BAKEL

SUMMARY

This paper mainly deals with the hydrological processes in the unsaturated zone. Firstly the most relevant theory concerning this zone is given. The saturated groundwater system and the atmospheric system constitute the boundaries of the unsaturated zone. The relationship between these three systems are discussed in short.

An important issue is the mathematical description of water uptake by plant roots. Only the most common methods to quantify this uptake are treated.

An overview of the various categories of methods available to evaluate effects of changes in soil water conditions on evapotranspiration is presented. The problems encountered in practice mostly relate to the translation of model results to field situations. The various aspects with respect to this problem are discussed. Finally conclusions and recommendations are made.

1. INTRODUCTION

In The Netherlands the demand for water by agriculture, industry and domestic water supply and also for nature is steadily increasing. This calls for proper water management schemes. In order to find the optimal distribution of water among the users – taking into account the effects on safety, environment and pollution – one needs a weighing of the conflicting interests.

Agriculture can be characterized by an intensive use of the soil, a need for good drainage conditions during winter and an additional water supply in summer. The latter is obtained from surface water and/or groundwater.

For the domestic and industrial water supply the main objective is a continuous delivery of sufficient water of good quality, also from surface/groundwater resources.

The main objective for managing nature is the creation of such environmental conditions that the diverse and mutual relationships are guaranteed. Often this implies conservation of the existing hydrological situation.

Proper water management of any region with competing interests implies an evaluation of the various measures that are available. The hydrological part of this evaluation can be made with the aid of, among others, mathematical-hydrological models. These models concern the entire system of surface water – groundwater – soil water – evapotranspiration.

The upper zone of the soil, i.e. the unsaturated zone, constitutes the medium between the atmosphere and the saturated groundwater system. This zone is very important for the biological, chemical and physical processes occurring in the soil – plant system.

Hydrological measures taken will affect the processes in the unsaturated zone.

In the following the soil-physical aspects of the unsaturated zone will be treated with respect to groundwater movement and evapotranspiration.

2. UNSATURATED ZONE

2.1. *General theory*

Water in soil moves from points where it has a high energy status to points where it has a lower one. The energy status of water is the water potential which consists of several components. Potentials are defined relative to the reference status of water at atmospheric pressure and elevation datum zero. In hydrology potential is usually expressed as energy per unit weight of soil water, with the dimension of length, i.e. cm and potential is then denoted as 'head'. When dealing only with the matric head, h_m, arising from local interacting forces between soil and water and gravitational head, z, arising from the gravitational force, total (hydraulic) head, H, can be expressed as:

$$H = h_m + z \text{ (cm)} \tag{1}$$

where the vertical coordinate z is considered positive in upward direction. Under conditions of atmopheric pressure and non-swelling soils the matric head can be denoted as pressure head.

For each soil there exists a relation between the pressure head, h (cm), and the soil water content, θ ($cm^3.cm^{-3}$), so:

$$\theta = f(h) \tag{2}$$

To describe the flow of water in soil systems, it is customary to use Darcy's law. For one dimensional vertical flow, the volumetric flux q ($cm^3.cm^{-2}.d^{-1}$) can be written as:

$$q = -K\frac{\delta H}{\delta z} \text{ (cm.d}^{-1}\text{)} \tag{3}$$

where K is the hydraulic conductivity ($cm.d^{-1}$)

For saturated (groundwater) flow the total soil pore space is available for water flow and the hydraulic conductivity is constant. With unsaturated flow, however, part of the pores are filled with air. Therefore K is not a constant but depends on the soil moisture content θ or because [$\theta = f(h)$] on the pressure head

$$K = f(\theta) \quad \text{or} \quad K = f(h) \tag{4}$$

Substitution of eq. (1) into eq. (3) yields:

$$q = -K(h)\left(\frac{\delta h}{\delta z} + 1\right) \tag{5}$$

In order to get a complete mathematical description we apply the continuity principle (Law of Conservation of Matter):

$$\frac{\delta \theta}{\delta t} = -\frac{\delta q}{\delta z} \quad (d^{-1}) \tag{6}$$

Substitution of eq. (5) in eq. (6) yields:

$$\frac{\delta \theta}{\delta t} = \frac{\delta}{\delta z}\left[K\left(\frac{\delta h}{\delta z} + 1\right)\right] \tag{7}$$

To avoid the problem of the two dependent variables θ and h, the derivative of θ with respect to h can be introduced, which is known as the differential soil water capacity C

$$C = \frac{d\theta}{dh} \quad (cm^{-1}) \tag{8}$$

Writing

$$\frac{\delta \theta}{\delta t} = \frac{d\theta}{dh}\frac{\delta h}{\delta t} \tag{9}$$

and substitution of eq. (8) in to eq. (7) yields the one-dimensional equation for water flow in heterogeneous soils

$$\frac{\delta h}{\delta t} = \frac{1}{C(h)} \frac{\delta}{\delta z}\left[K(h)\left(\frac{\delta h}{\delta z} + 1\right)\right] \tag{10}$$

2.2. *Soil physical properties*

The soil water characteristic, $\theta = f(h)$, can be represented as a “pF-curve”, according to $pF = {}^{10}\log(-h)$.

The pF-curves are usually determined by removing water from an initially wet soil sample (desorption). If one adds water to an initially drier sample (adsorption), the moisture content in the soil will be different at corresponding tensions (see fig. 1).

This phenomenon is referred to as hysteresis. The hysteresis problem is greatest in sandy soils at low suction. With the aid of numerical models it is possible to account for effects of hysteresis. Whether or not considering hysteresis depends mainly on the type of soil and if the soil is exposed to frequent wetting and drying. In practice hysteresis curves are not determined at the laboratory and are hardly applied in models. For more information about hysteresis, see Mualem and Dagan (1975) and Royer and Vachaud (1975).

The hydraulic conductivity curve can be represented in different ways. Examples of such expressions are given in fig. 2.

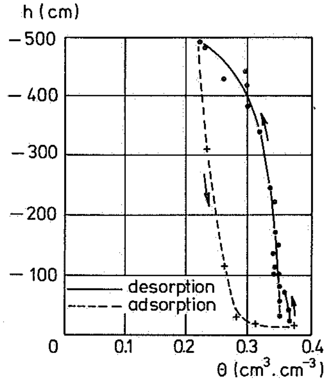

Fig. 1 Example of the hysteresis phenomenon as determined in the field. (after Royer and Vachaud, 1975)

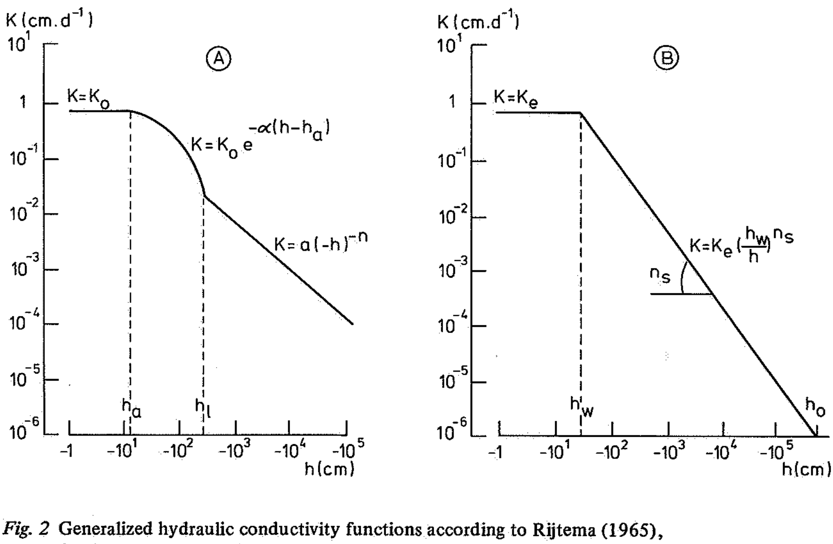

Fig. 2 Generalized hydraulic conductivity functions according to Rijtema (1965),
fig. 2A, where K_o = saturated hydraulic conductivity, h_a = pressure head at air-entry point,
h_l = some pressure limit, α, a and n are constants; and according to Bloemen (1979),
fig. 2B, where K_e = effective conductivity attained after rewetting, h_w = pressure head at which K_W is attained,
n_s = the slope of the curve and
h_o = pressure head at which the hydraulic conductivity is negligibly small.

As far as hysteresis in the K(h) curve is concerned, similar remarks as for the θ (h)-relation can be made. In fig. 2B an average scanning curve between drying and wetting was taken to account for this effect. In heavy clay soils swelling and shrinking during the year affects the K(h)-relationship, Bouma and De Laat (1980) have shown that reduced K(h)-curves should be used for a column as compared with the curve obtained for a ped by standard laboratory methods.

2.3. *Steady-state capillary rise*

In The Netherlands the water use of crops is effected by the upward flow from the relatively shallow groundwater table. This upward flow is usually termed capillary rise.

Integration of eq. (5) in steady-state conditions yields the relationship between flux, q, pressure head, $\bar{h}$, and vertical coordinate, z. For special types of K(h)-functions analytical solution of the integration is possible. Solution by numerical integration, however, is always possible, both for homogeneous and heterogeneous (layered) profiles. Therefore eq. (5) can be written in a finite difference notation as:

$$\Delta z = -\frac{1}{1 + q/K(h)}\Delta h \qquad (5a)$$

and applied to each layer separately.

2.4. *Non-stationary flow*

In case of a non-stationary flow, $\delta h/\delta t \neq 0$ and eq. (10) is valid. Solution of this equation for non-homegeneous soil profiles is only possible with numerical and analog models.

Writing eq. (10) in a finite difference from yields:

$$\frac{\Delta h}{\Delta t} = \frac{\Delta}{C(h)\,\Delta z}\,[K(h)\,(\frac{\Delta h}{\Delta z} + 1)] \qquad (10a)$$

The functional relationships C(h) and K(h) now must be 'averaged' over time and space. Vauclin et al. (1979) give a good description of the consequences of this averaging process. Because of the high non-linearity in the C(h) and K(h) functions, the maximum time step to be allowed during the computation is relatively small.

For a unique solution of h with respect to time and space initial and boundary conditions must be applied. As initial condition either the pressure head as a function of depth z must be given

$$h(z, t = 0) = h_0\,(z) \qquad (11)$$

or the moisture content as a function of the depth z

$$\theta(z, t = 0) = \theta_0(z) \tag{12}$$

must be applied.

Boundary conditions at $z = z_B$ i.e. at the top and the bottom of the unsaturated zone can be specified in three different ways:
– Dirichlet condition: the pressure head is specified as a function of time

$$h(z = z_B, t) = h_B(t) \tag{13}$$

– Neumann condition: the flux is specified as function of time

$$q(z = z_B, t) = q_B(t) \tag{14}$$

– Functional relationship (Cauchy condition): the flux is a function of the dependent variable h at $z = z_B$

$$q(z = z_B, t) = f(h_B, t) \tag{15}$$

Use of this type is only possible in iterative computations.

In general at the top and at the bottom of the unsaturated zone different types of boundary conditions can be used at the same time.

Even during the process of computation the type of boundary conditions may be changed.

For the unsaturated zone the boundaries are constituted by the soil surface and the phreatic surface. Through these boundaries relations can be established with the atmosphere and the saturated zone. In the next chapter the systems at the boundaries will be treated, as far as they affect the unsaturated zone and vice versa.

3. BOUNDARIES OF THE UNSATURATED ZONE

3.1. *Saturated groundwater system*

As "lower" boundary of the unsaturated zone the phreatic surface is acting. The connection with the saturated zone is usually established by the flux – groundwater table depth relationships which van be derived from some drainage formula.

In the more traditional approach of evaluating water management measures, changes in the water table depth were calculated without considering the actual respons of the unsaturated zone and a mostly unique relationship between mean groundwater table depth during the growing season and evapotranspiration/crop yield was taken (see fig. 3). Recently, with the use of numerical models, it is possible to consider both the saturated and unsaturated zone as one continuum (see Neumann et al., 1974).

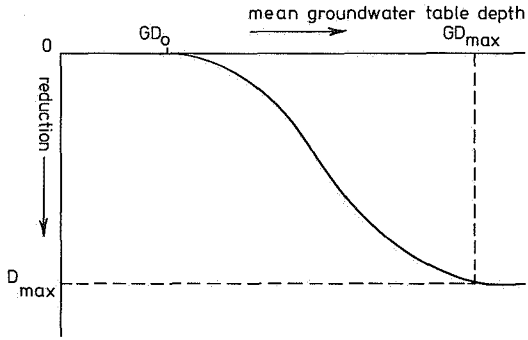

Fig. 3 Effect of groundwater table depth, GD, during the growing season on crop yield reduction. GD_o = lowest groundwater table depth where reduction is appr. zero; GD_{max} = highest groundwater table depth where reduction is appr. at it's maximum (after Grootentraast, see Van den Berg, 1979).

3.2. *Atmospheric system*

The atmosphere influences the unsaturated zone in two ways: at the soil surface through infiltration/evaporation and at the crop surface through transpiration. The soil can lose water to the atmosphere by evaporation or gain water by infiltration. While the maximum possible (potential) rate of evaporation from a given soil depends only on atmospheric conditions, the actual flux across the soil surface is limited by the ability of the porous medium to transmit water from below. Similarly if the potential rate of infiltration (e.g. the rain or irrigation intensity) exceeds the absorption capacity of the soil, part of the water will be lost by surface run-off. Here, again, the potential rate of infiltration is controlled by atmospheric (or other) external conditions, whereas the actual flux depends on the soil physical properties and on the soil water conditions.

The water uptake by roots, per unit area of soil, i.e. the transpiration is dependent on the conditions of the atmosphere, the kind, stage and condition of the crop and the soil water status.

Often the combined water loss from the crop – soil surface is considered as one term, i.e. evapotranspiration.

The maximum possible flux through the soil and crop surface can be defined as

$$ET_{pot} = E_{pot} + T_{pot} \; (kg.m^{-2}.s^{-1}) \tag{16}$$

where ET_{pot}, E_{pot} and T_{pot} are the potential evapotranspiration flux, soil evaporation flux and transpiration flux, respectively. In this way these fluxes constitute a Neumann type of boundary condition.

Potential evapotranspiration flux can be derived from a combination of the energy balance equation and the vapour transport equation (Rijtema, 1965; Monteith, 1965; Keijman, 1981).

The evapotranspiration flux of a wet crop surface ET_{wet}, can be written as:

$$ET_{wet} = \frac{s\frac{Q^*}{\lambda} + \gamma\, ET_a}{s + \gamma} \quad (kg.m^{-2}.s^{-1}) \tag{17}$$

where s is the slope of the saturation vapour pressure curve ($mbar.K^{-1}$), Q^* is net radiation flux ($W.m^{-2}$), λ is latent heat of vaporization ($J.kg^{-1}$), γ is psychrometer constant ($mbar.K^{-1}$) and ET_a is the aerodynamic evapotranspiration flux ($kg.m^{-2}.s^{-1}$).

The potential evapotranspiration flux, ET_{pot}, can now be defined as:

$$ET_{pot} = \frac{s + \gamma}{s + \gamma(1 + r_c/r_a)\, ET_{wet}} \quad (kg.m^{-2}.s^{-1}) \tag{18}$$

where, according to Rijtema (1965), r_a is the aerodynamic diffusion resistance and r_c the internal canopy resistance built up of a stomatal resistance dependent on light intensity, r_1, and a resistance dependent on the fraction of soil cover, r_{sc}

$$r_c = r_1 + r_{sc} \quad (s.m^{-1}) \tag{19}$$

Alternative expressions for estimating maximum possible evapotranspiration fluxes are the Priestley and Taylor (1972) equation and the Thom and Oliver (1977) equation.

Stricker (1981) compares the results of the calculations defined with the three expressions with water balance measurements. He also discusses the practical applicability of the various methods.

The potential evaporation of a cropped soil can be computed by neglecting the aerodynamic term and taking into account only that fraction of Q^* which reaches the soil surface (Ritchie, 1973).

$$E_{pot} = \frac{s}{(s + \gamma)}\, Q^* - 0.39I \quad (kg.m^{-2}.s^{-1}) \tag{20}$$

where I is the leaf area index. This index generally can be related to soil cover.

According to eq. (16), the potential transpiration flux now can be calculated as the potential evapotranspiration flux minus potential soil evaporation flux (minus evaporation flux of intercepted precipitation).

Neglecting storage in the crop, the potential or actual transpiration flux must be equal to the water uptake rate of the plant roots. In the next chapter this water uptake will be described mathematically. Also the way in which non-optimal soil moisture conditions influence the transpiration rate will be discussed.

4. WATER UPTAKE BY PLANT ROOTS

To describe water uptake by plant roots several approaches are available. We restrict ourselves to the most common methods.

(a) Analogous to Ohm's law the steady-state flow of liquid water through the soil – root system can be described as

$$T = \frac{h_{soil} - h_{root}}{R_{soil}} = \frac{h_{root} - h_{leaf}}{R_{plant}} \quad (cm.d^{-1}) \tag{21}$$

where h_{soil}, h_{root} and h_{leaf} are the pressure heads (cm) in the soil at the root surface and in the leaves respectively; R_{soil} and R_{plant} are the flow resistances (d) in soil and plant respectively, considered as liquid-phase resistances. Transpiration is expressed here as a volume flux, i.e. $cm^3.cm^{-2}.d^{-1}$. (Note that a mass flux of 1 kg $H_2O.m^{-2}.d^{-1}$ is equivalent with a volume flux of 1 $mm.d^{-1}$). Therefore R_{plant} does not include stomatal resistance. When the transpiration demand of the atmosphere on the plant system is too high or when the soil is rather dry, R_{soil} and R_{plant} influence h_{leaf} in such a way that transpiration is reduced by closure of the stomata. As it is still not possible to measure h_{root}, one generally prefers to use an equivalent form of eq. (22):

$$T = \frac{h_{soil} - h_{leaf}}{R_{soil} + R_{plant}} \quad (cm.d^{-1}) \tag{22}$$

Equation (22) can be applied to either the root system as a whole or to discrete horizontal layers. The soil resistance, R_{soil}, can be written as:

$$R_{soil} = b/K(h) \quad (d) \tag{23}$$

where b is an empirical root effectiveness function (cm) and K(h) is the hydraulic conductivity of the root zone. One of the major difficulties of eq. (23) is the determination of this root effectiveness function. Feddes and Rijtema (1972) derived from a number of crops that

$$b = 0.0013\, z_r^{-1} \tag{24}$$

where z_r is the rooting depth (cm). From calculations of Van Bakel (1979) with an unsaturated flow evapotranspiration model (De Laat, 1980), it appeared that the function b could be changed quite drastically without seriously affecting the transpiration rate.

From eq. (22) it can be derived that

$$h_{leaf} = h_{soil} - (R_{soil} + R_{plant})\, T \quad (cm) \tag{22a}$$

Actual evapotranspiration can now be described similar to eq. (18) as (Rijtema, 1965):

$$ET = \frac{s + \gamma(1 + r_c/r_a)}{s + \gamma(1 + r_{c'}/r_a} \; ET_{pot} \tag{25}$$

with $r_{c'} = r_c + r_h$ (26)

and the diffusion resistance r_h is depending on leaf water pressure head. So

$$r_h = f(h_{leaf}) \tag{27}$$

or, according to eq. (22a):

$$r_h = f[h_{soil} - (R_{soil} + R_{plant})T] \tag{27a}$$

Some empirical r_h-functions for different types of crops are listed by Van Bakel (1979). For crops with full soil cover $T \simeq ET$. It appears from eqs. (25) and (26) that ET depends on r_h. Because r_h depends in its turn on (E)T (eq. 27a), in principle ET can be solved iteratively.

(b) In the second approach, water uptake by the roots is represented by a volumetric sink term, which is added to the continuity equation (6):

$$\frac{\delta\theta}{\delta t} = - \frac{\delta q}{\delta z} - S \quad (d^{-1}) \tag{28}$$

where S represents the volume of water taken up by the roots per unit bulk volume of soil in unit time ($cm^3 . cm^{-3} . d^{-1}$).

Again, one faces the problem of determining the magnitude of the sink term. Feddes et al. (1978) considers the sink terms as a function of the soil water pressure head, h. By definition the integral of the sink term over the rooting depth, z_r, equals the actual transpiration rate

$$T = \int_{z=0}^{z=z_r} S(h)\, dz \quad (cm.d^{-1}) \tag{29}$$

For optimal soil moisture conditions, $T = T_{pot}$ and

$$S(h) = S_{max} = \frac{T_{pot}}{z_r} \quad (d^{-1}) \tag{30}$$

In non-optimal soil water conditions water uptake is reduced according to

$$S(h) = \alpha(h)\, S_{max}\, ; 0 \leq \alpha \leq 1 \tag{31}$$

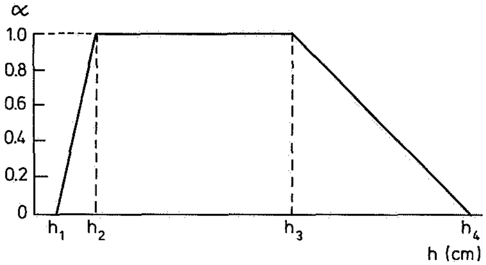

Fig. 4 Relation between sink term variable α, ($\alpha(h) = S(h)/S_{max}$) and soil water pressure head, h. h_1 = anaerobiosis point; h_2 = lowest value of h where $S = S_{max}$; h_3 = highest value of h where $S = S_{max}$; h_4 = wilting point. (after Feddes et al., 1978)

An example of the shape of $\alpha(h)$ is shown in fig. 4. Little is known about the anaerobasis point, (h_1, fig. 4), at which deficient aeration conditions exist. In any case, it will depend on type of crop, temperature and type of soil, duration of anaerobiotic situation.

For the h-range ($h_2 \rightarrow h_3$) where transpiration is potential, different values of h_2 and h_3 are used. Some authors (e.g. Feddes et al., 1978) apply a fixed range, while others (e.g. Yang and De Jong, 1972) found a varying range, depending on the evaporative demand of the atmosphere. For the wilting point (h_4) one usually takes a h-value of – 16,000 cm (i.e. pF = 4.2).

Incorporation of the sink term into eq. (10a) yields:

$$\frac{\Delta h}{\Delta t} = \frac{\Delta}{C(h)\,\Delta z}\left[K(h)\left(\frac{\Delta h}{\Delta z} + 1\right)\right] + \frac{S(h)}{C(h)} \qquad (10b)$$

The numerical solution of eq. (10b) is essentially the same as for eq. (10a).

5. INTEGRATED APPROACH TO EVALUATE EFFECTS OF CHANGES IN SOIL WATER CONDITIONS ON EVAPOTRANSPIRATION

To optimize different schemes of water management it is important to have methods available that can help the water manager in making his decisions. With such methods he may evaluate effects of water management upon evapotranspiration, crop production, water quality, composition of vegetation, etc.

Restricting ourselves to the effects of changes in soil water conditions on evapotranspiration, different approaches can be used. They range from methods that are based on semi-empirical relationships to fysical-mathematical models that describe the system in a very detailed way. In the following an overview of the various categories of methods that are nowadays in use or developed in The Netherlands will be presented.

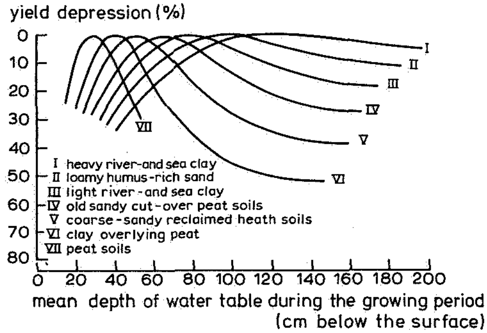

Fig. 5 The effect of the mean depth of the groundwater table during the growing season on final crop yield for seven groups of soils. (after Visser, 1958)

5.1. *Empirical methods*

This category of methods has the advantage that they are simple to apply, but the disadvantage that they are not applicable to other areas than for which they were developed.

Visser (1958) gives the yield depression for seven soil classes with various mean depths of the water table during the growing seasons (see fig. 5). Yield depressions at high water tables can be ascribed to lack of aeration of the soil, depressions at deeper water tables are due to shortage of water. Since both conditions may occur dependent on climatological circumstances, it is evident that the curves may undergo a horizontal shift to the right in wet years and a shift to the left in dry years. These curves are reflecting the behaviour of the unsaturated zone and its effect through evapotranspiration on crop production. Hence the unsaturated zone is used here as a 'black-box'.

It should be remarked that attempts have been made to adopt the curves to changed circumstances. Use of it is to be advised against, however.

5.2. *Parametric models*

This type of models is still empirical but make use of some properties of the unsaturated zone.

Grootentraast (see Van de Berg, 1979) determined the yield depression of grassland on sandy soils due to an artificial drawdown of the groundwater table according to the principle presented in fig. 3. For a number of years the relative evapotranspiration/yield is computed for conditions where no water table is present. This gives the maximum yield depression D_{max}. This point corresponds with a theoretical groundwater table depth GD_{max} where the capillary rise approximately is zero. Then the groundwater table depth GD_0 is determined at which a reduction in evapotranspiration/yield never will occur and the points GD_{max} and GD_o are connected by a smooth curve. With this method a number of properties as the soil moisture retention curve and the hydraulic conductivity implicitly are taken into account.

One of the most important parameters is the water storage ST of the unsaturated zone. Thornthwaite and Mather (1955) calculated the actual evapotranspiration ET according to

$$ET = -\frac{dST}{dt} \quad (cm.d^{-1}) \tag{32}$$

under the assumption that

$$ET = cST \quad (cm.d^{-1}) \tag{33}$$

and taking at $t = t_o$: $ST = ST_o$ and $ET = ET_{pot}$ it follows that

$$c = \frac{ET_{pot}}{ST_o} \tag{34}$$

so

$$ET = \frac{ST}{ST_o} ET_{pot} \quad (cm.d^{-1}) \tag{35}$$

Eq. (34) applies to (vegetated) soil with deep groundwater tables (no capillary rise).

A reservoir model that takes into account the influence of the groundwater table has been developed by a Working Group of ICW (1979).

For a unit area the water balance of the unsaturated zone over a certain period can be written as:

$$ST_t = ST_o + Q - ET \quad (cm) \tag{36}$$

where

ST_t, ST_o	is water storage of soil profile at t = t and t = o respectively,
Q	water supplied from outside the system (precipitation, sprinkling, seepage),
ET	actual evapotranspiration.

Evapotranspiration ET depends on ST as follows:

$$ET = cET_{pot} \tag{37}$$

where

$c = 1$	for	$0 \leqslant pF \leqslant 3.2$
$c = \dfrac{ST - ST_{4.2}}{ST_{3.2} - ST_{4.2}}$	for	$3.2 < pF < 4.2$
$c = 0$	for	$pF = 4.2$

The value of ST is derived as

$$ST = \frac{ST_o + ST_t}{2} \tag{38}$$

Because ST_t is not know beforehand, the procedure of calculation starts with a first estimation of ST_t, i.e. $ST_t = ST_o$. From eq. (39) ET is computed. This value is substituted in eq. (38) yielding a new value for ST_t. If this value differs significantly from the initial ST_t-value, the computing process is repeated. Otherwise the next timestep is taken.

In the above mentioned two methods the reduction factor is independent of the magnitude of (potential) evapotranspiration. One can imagine that at low potential evapotranspiration rates reduction in evapotranspiration occurs at lower values of available soil water than at high potential evapotranspiration rates. Therefore some authors (e.g. Federer, 1979) found a relation between the reduction factor, and the ratio of available soil water (or availability factor) and potential (evapo)transpiration. This type of relationship can also be obtained from experiments of Denmead and Shaw (1962) and Van Bavel (1967).

5.3. *Physical-mathematical models*

This group of models is based on fysical processes and mathematical relationships as described in the chapters 2-4.

a. Steady-state models

These models can be applied when calculating percolation profiles or steady state capillary rise (evaporation) from the groundwater table towards the root zone or the soil surface. Numerical integration of eq. (5a) yields the relationship between q, z and h. For examples and computer programs see Aliverti-Piuri and Wesseling (1979) and Bloemen (1980). An illustration of the effect of a disturbing coarse sand layer in a fine sandy soil on the capillary flux from the groundwater table towards the root zone is presented in fig. 6. From a computational point of view there is no problem in handling non-homogeneous soil profiles.

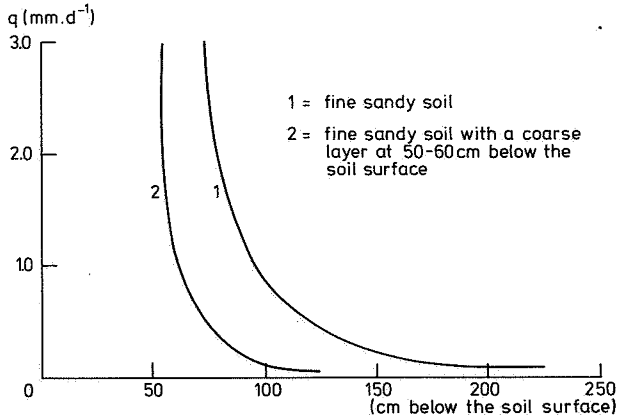

Fig. 6 Relation between groundwater table depth, GD, and capillary rise, q, in a homogeneous (1) and a heterogeneous profile (2). The pressure head at the lower boundary of the root zone is in both cases – 1000 cm. (after Mooy, personal communications)

b. Pseudo steady-state models

A non-stationary process of water flow can be approximated by a subsequent series of steady-state flow situations. The pressure head profiles h(z, q) can be converted into water content profiles θ(z, q) through the pF-curves. The θ(z, q) profiles can be integrated over a certain soil depth yielding water storage ST(q). Changes in water storage of the soil profile are found from the difference in two subsequent steady-state situations, $ST(q_2) - ST(q_1)$.

Rijtema (1971) used a pseudo-steady state approach to compute relative evapotranspiration during the growing season. He distinguished two zones in the soil profile: the root zone and the subsoil. In the root zone all water is taken to be transported through the roots, i.e. no vertical gradient over this zone exists. The subsoil is considered as a homogeneous single system. The model computes actual evapotranspiration over a certain period from the water balance of the root zone:

$$ET = P + Q + \Delta ST \quad (cm) \qquad (39)$$

where

- ET is evapotranspiration,
- P precipitation (infiltration),
- Q capillary rise from the subsoil,
- ΔST change in available soil water, $\Delta ST = ST(q_2) - ST(q_1)$.

It is assumed that the evapotranspiration rate is at its potential value when the tension in the root zone is smaller than a certain limit h_1. When this limit is exceeded, a reduction in evapotranspiration occurs.

De Laat (1980) developed a computer program for this model and extended it for periods with rainfall surplus and for heterogeneous subsoils. Rijtema (1971) assumed a fixed value of h_1 of $-$ 16,000 cm. De Laat can apply subsequently the eqs. (18), (22a), (27a) and (25) to calculate ET.

c. Transient models

The nonstationary process of water flow can be approximated by a numerical solution of eq. (10b). In fact this approach is the most simple one as it needs less restricting assumptions. The traditional disadvantage of a large amount of computing time inherent to this approach becomes less and less significant as recent trends in computer technology are directed towards faster computers.

Feddes et al. (1978) developed a computer program based on eq. (10b) with the sink term described according to eq. (30) and (31), for two- layered soil profiles. Theoretical

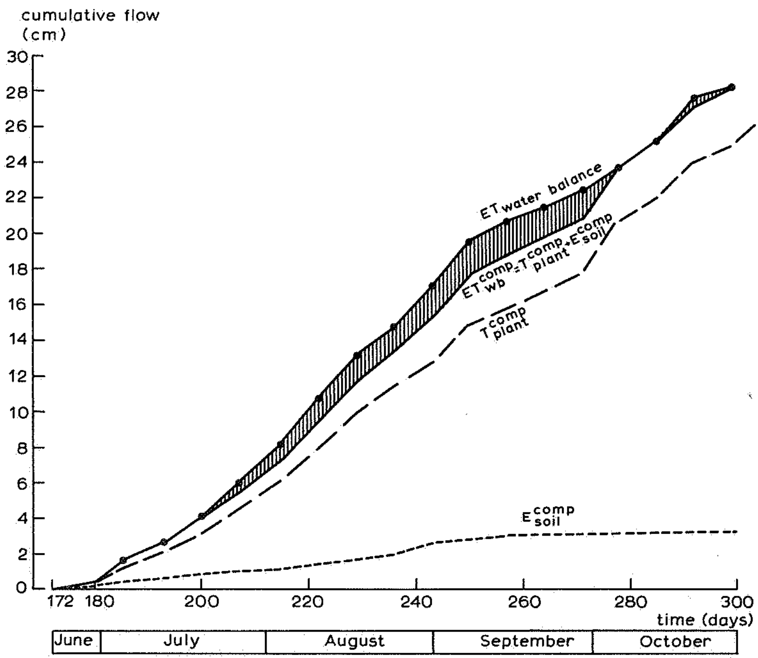

Fig. 7 Computed cumulative evapotranspiration, E^{comp}_{wb}, transpiration T^{comp}_{plant}, soil evaporation, E^{comp}_{soil} and measured cumulative evapotranspiration, $Et_{water\ balance}$. (after Feddes et al., 1978)

results predicted by the model were compared with a field experiment in which red cabbage was grown on a heavy clay soil in the presence of a water table. Water balance studies were performed with a specially designed non-weighable lysimeter.

In fig. 7 curves of cumulative flow are given: first the measured cumulative evapotranspiration ($ET_{water\ balance}$) as obtained from the lysimeter; secondly the cumulative transpiration T_{comp} as computed with the model by integration of the sink term over depth; thirdly the cumulative soil evaporation E_{soil}^{comp} derived from the computed terms of the water balance.

The above mentioned model applies to one-dimensional vertical flow. Other examples of such models are given by Van Keulen (1975), Van der Ploeg et al. (1978), Nimah and Hanks (1973a/b) and Childs and Hanks (1975). The latter take also into account the presence of salts.

Neuman et al. (1974) developed a two-dimensional finite element model for transient flow in saturated-unsaturated soils. An example of this approach can be found in Feddes et al. (1975), where flow takes place under cropped field conditions in a five-layered anisotropic soil, the boundaries of which include two ditches on the side and a pumped aquifer at the bottom.

6. CONCLUSIONS AND RECOMMENDATIONS

In the previous section a number of methods and models have been discussed. In practice there are difficulties in choosing the right method or model and translating the one-dimensional model results obtained at one point to larger areas. The complex reality has to be schematized into a proper way, i.e. one has to distinguish the soil-plant-atmosphere system in a limited number of characteristic situations. Important properties are:

- soil use as far as it influences evapotranspiration through its roughness, rooting depth, etc.;
- soil physical parameters as soil water characteristic, capillary conductivity and derived properties such as infiltration capacity, water availability for the plant and storage coefficient;
- topographical data such as altitude, slope, etc.

In the first place the choice of method or model depends on the objective of the study (communication between policy maker and model user). Secondly it depends on the availability of the above mentioned data. Reversely, collection of data depends on the type of model chosen. From experience it appears that most problems are encountered with regard to the following items:

- estimation of rooting depths, also as function of time. From sensitivity analyses (De Laat, 1980) it appeared that a good estimation of this depth is very important for the calculated evapotranspiration;
- influence of very wet conditions upon plant growth and transpiration, which is very poorly understood;

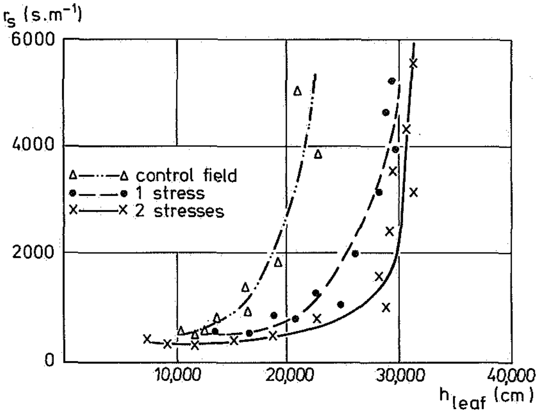

Fig. 8 Differences in relationship between stomatal diffusion resistance, r_s, and leaf water pressure head, h_{leaf}, due to differences in soil water stress history. (after Thomas et al., 1976)

- reaction and adaptation of an active growing plant system on changes in external circumstances. From experiments (Thomas et al., 1976) it can be deduced that the history of water stress can be very important for the subsequent reaction of crops on water shortages. This is clearly demonstrated in fig. 8. Of all the models discussed in section 5 none has the possibility to incorporate this phenomenon;
- soil loosening and compaction have distinct effects on soil physical properties (Boone et al., 1978). Mostly, however, they are taken as invariant with time;
- translation of the soil map into a map of soil physical characteristics. Bouma (1977) stated that much testing remains to be done to extrapolate data measured in a particular soil to unmeasured soils with identical classification elsewhere;
- spatial variability of soil physical properties. Nielsen et al. (1973) found from their investigation in California that even seemingly uniform soil areas manifest large variations in hydraulic conductivity values;
- fast and simple determination of the soil physical properties. A promising approach is for example the estimation of K(h)-relationships, based on soil texture and organic matter content of the soil, as developed by Bloemen (1979);
- field determination of the drainage respectively infiltration resistance of open water courses. Especially the attempts of Ernst (1978) to establish drainage formulas for areas with composite water course systems should be emphasized;
- in simulation models topograhical variation of the soil surface within the area of investigation, as for example analyzed by a Working Party (1968), is seldomly taken into account;
- representativity of meteorological data. This problem is discussed elsewhere in this paper;

- coupling of the meteorological system and the soil plant system. Micro-meteorological approaches are in fact bounded by the soil water pressure head in the root zone, while on the other hand the soil physical models are bounded by the atmospheric conditions.
  This is a drastic simplification of reality;
- accessibility of data. The present-day models need much data which should be computer accessible. Especially data on crops, soil physical and hydrological properties for larger areas that can be applied on a routine basis are scarce and difficult to obtain;
- choice of the method or the model that is best suitable to apply to a certain situation. A kind of a clearing house for models, that can present an overview of models that are available and that contains directions for the user, would be very useful. Maybe one could take advantage of the experience obtained in this respect in the USA (Bachmat et al., 1978);
- formulation of the problem in hydrological terms. There seems to be large discrepancies between policy maker, model user and model builder.

Looking over the present state of the scientific knowledge dealing with the unsaturated zone in relation to evapotranspiration, one may conclude that many problems remain to be solved. From the field hydrologist it is not to be expected that he keeps himself fully informed about all the methods and models available, including new developments.

A way out of this dilemma would be to let model specialists compute a great number of selected flow cases and let them derive from it some generalized relationships. This will help to select criteria for practical field situations. As an example of such an approach one could refer to the evaluation of water management on crop production for a number of soils, drain depths and weather conditions as carried out by Feddes and Van Wijk (1977) in the framework of the working group on Re-evaluation of Land Reclamation Projects (see also Working Group H.E.L.P., 1980)

Some knowledge of the physical-mathematical background of methods and models will always remain necessary for the field hydrologist, however. This paper may serve as a small contribution to that end.

7. ACKNOWLEDGEMENT

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METHODS OF ESTIMATING EVAPOTRANSPIRATION FROM METEOROLOGICAL DATA AND THEIR APPLICABILITY IN HYDROLOGY

J.N.M. STRICKER

SUMMARY

Next to rainfall evapotranspiration (ET) is the most important term in the waterbalance of catchment areas in temperate regions, such as The Netherlands.

Therefore it is of interest to develop and test methods for estimating ET. From March 1976 to December 1978, daily values of ET have been calculated by different methods. The data are from point-measurements and were collected in the experimental basin "Hupselse Beek".

The results of different methods, including the waterbalance, are compared, and some conclusions are drawn concerning their usefulness and applicability.

Furthermore, preliminary comment is given on interception of grass and the so-called rainfall-deficit, defined by (N-0.8 E_o).

1. INTRODUCTION

In the water balance, evapotranspiration (ET) is normally of the same order of magnitude as rainfall or run-off. Nevertheless the number of studies on rainfall and run-off far exceed the number on the process of evapotranspiration in hydrological literature. Greenwood (1979), for example ascertained: "that hydrologists have always found it hard to come to grips with transpiration". One may wonder why. Perhaps it is because of the background of many hydrologists that makes them unfamiliar with the subject. Or may be, the process of ET seems less attractive and less urgent to be studied. A third reason may be the complexity of the process of evapotranspiration. For example, measurements of run-off produce spatially integrated values. Rainfall measurements normally represent an estimate for an area. But for actual evapotranspiration the representativity is always open to question, because of the spatial variation in vegetation and soil moisture regime over a catchment.

Besides, careful measurements of ET need more equipment and daily management than rainfall or run-off measurements, for which many experimental and representative basins are well equipped, but are poorly equipped with meteorological instruments, even to determine potential ET. In my opinion this fact may be considered as a shortcoming in the program of the International Hydrological Decade (IHD): no time-series of actual ET from several experimental basins are available.

The situation now looks more promising. Meteorologists and hydrologists have come to accept that more testing is needed of existing equipment and approaches and more development of new tools for estimating actual and potential ET. The Final Report

(WMO, 1977) of the Technical Conference on the Assessment of Areal Evapotranspiration, held in Budapest, recommended more emphasis on such problems. A report on Operational Hydrology (WMO, 1975) stated, that ET functions in existing catchment models may be inadequate and need improvement by future research. It discussed the results of an intercomparison of catchment models and concluded that black-box and conceptual models have a weakness in the way of modelling rainfall to effective rainfall.

The previous papers indicate that progress has been made in the last years and will certainly continue.

This paper enlarges on some methods discussed earlier by De Bruin (1981) and which I have tested against data from the Hupselse Beek Experimental Catchment. Some conclusions are drawn about spatial representation of ET by different methods, their validity and restriction.

2. EXPERIMENTAL SITE AND AVAILABLE DATA

For a detailed description of the Hupselse Beek Experimental Catchment see the biennial reports of the Hupselse Beek Study Group (1971; 1972; 1974; 1976). Only basic information is given here. The catchment (fig. 1) is situated near Groenlo and Eibergen. It covers 6.50 km^2 and may be considered representative of sandy regions in The Netherlands. It is supposed to be watertight in the subsoil, and no deep seepage or infiltration interferes in the water balance. A complicating factor is the local presence of

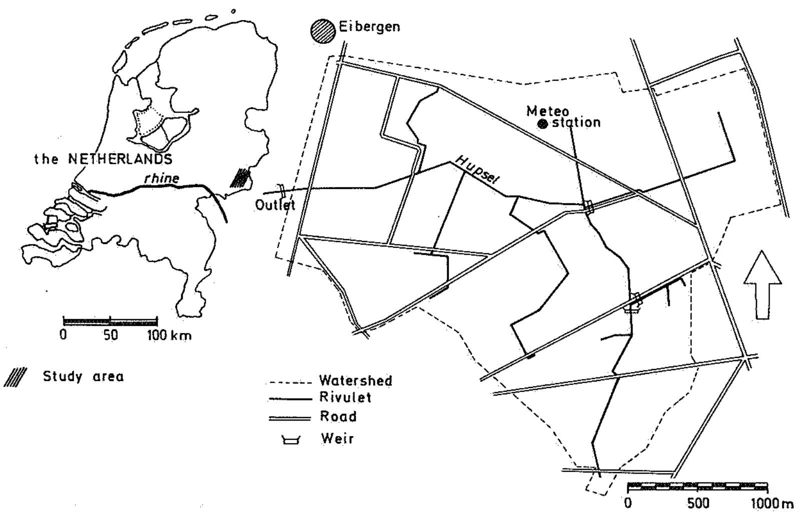

Fig. 1 Experimental catchment "Hupselse Beek"

tilt layers in the upper 1.50 m, which causes a high local resistance to vertical transport of water.

Land use is predominantly agricultural: 70% pasture, 20% arable with maize as the main crop, and 6% forest.

Over the period March 1976 to December 1978 daily rainfall, run-off, and actual and potential evapotranspiration were collected or calculated from catchment data. Table 1 summarizes the available records. Rainfall was also collected from other types of rain gauges (Warmerdam, 1981). A full discussion on the accuracy of the meteorological data was given by Stricker and Brutsaert (1978).

Data on relative humidity (RH) became erroneous from June 1977 and data from Winterswijk were taken. Careful comparison between RH from Hupsel and Winterswijk over the period March 1976 to the first half of 1977 provided a one-to-one relationship between the two sets of data for daily mean values above 60% RH. However below 60% RH, 10 items from the Winterswijk data needed some correction: for 5 days, RH had to be corrected upwards by 1 or 2% and for 5 days by 3% or more.

In 1978, temperature from the sensor at 3.00 m became unreliable because of insufficient shielding against radiation. So, application of the profile method was restricted to temperature at 0.10 and 1.50 m in that year.

Table 1 Available data for calculations over the period March 1976 to December 1978 (1036 days or 34 months)

Data	*Period*	*Interval*	*Instruments*
Rainfall	complete	15 min	Recover at groundlevel
Run-off	complete	15 min	HL-flume
Net radiation	± 950 days*[1]	60 min	C.S.I.R.O.
Rel. humidity	complete until*[2] 77-07-09	60 min	Hair hygrometer
Windspeed	± 1000 days*[3]	60 min	Thiess anemometer
Air temperature (three levels)	917 days complete*[4] at 2 or 3 levels; ± 80 days at one level available	60 min	Semiconductors
Soil temperature	± 1000 days*[5]	60 min	Semiconductors
Groundwater at Assink Met. Station	complete	15 min	
Soil moisture	nearly complete*[6] at six sites	bimonthly	NEA-probe

*[1] : Additional data from Wageningen, Met. Station of the Department of Physics and Meteorology.
*[2] : Additonal data from KNMI Met. Station "Winterswijk" (see text).
*[3] : Additional data from Lambrecht cupanemometer at "Hupselse Beek".
*[4] : Additonal data from KNMI Met. Station "Winterswijk".
*[5] : Additional data by interpolation.
*[6] : No definite judgement exists about the accuracy of the data and about their representation of the catchment.

3. CALCULATION OF POTENTIAL AND ACTUAL EVAPOTRANSPIRATION

3.1. *Potential evapotranspiration*

Several formulations of potential evapotranspiration have been applied to mean daily data over different periods of the year. The following have been used, with the period of the year given in brackets:

$$\lambda ET^{PEN} = \frac{s}{s+\gamma}(Q^* - G) + \frac{\gamma}{s+\gamma}(3.7 + 4.0\,\overline{u})\,\overline{\Delta e} \text{ (all days)} \qquad (1)$$

$$\lambda ET^{TH+O} = \frac{s}{s+\gamma(1+n)}(Q^* - G) + \frac{m \cdot \gamma}{s+\gamma(1+n)}(7.4 + 4.0\,\overline{u})\,\overline{\Delta e} \qquad \text{(all days) } (2)$$

$$\lambda ET^{PR+T} = \alpha\frac{s}{s+\gamma}(Q^* - G) \text{ (April, May, June, July, August, Sept.)} \qquad (3)$$

List of symbols:

λ	: latent heat of vaporization	(J/kg)
ET^{PEN}	: potential evapotranspiration above grass under measured conditions of net radiation	($kg/m^2.s$)
s	: change of saturated vapour pressure of air per °K	(mbar/K)
γ	: psychrometric constant: 0.66 at $\simeq$ 295°K	(mbar/K)
Q^*	: net radiation or net radiant flux density	(W/m^2)
G	: soil heat flux or heat flux density of ground	(W/m^2)
$\overline{u}$	: mean daily wind velocity	(m/s)
$\overline{\Delta e}$	: mean daily water vapour deficit in the air at 2 m height	(mbar)
ET^{TH+O}	: potential evapotranspiration of a dry canopy by the method of Thom and Oliver (1977)	($kg/m^2.s$)
n	: ratio of canopy resistance (r_c) to aerodynamic resistance (r_a). Here 1.2 has been adopted for grass. With eq. 2 for a wet canopy n becomes zero. $r_c \simeq 65$ s/m	
m	: ratio of the aerodynamic resistance as expressed by the Penman equation to the aerodynamic resistance over a grass surface. Here 1.9 has been adopted.	(1)
ET^{PR+T}	: potential evapotranspiration formula, as proposed by Priestley and Taylor (1972)	($kg/m^2.s$)
α	: empirical factor which has to be calibrated in areas of different climates. Here $\alpha = 1.28$.	(1)

Note: $86\,400 \times kg/m^2.s = 1\ kg/m^2.d$
For water, $1\ kg/m^2.d \triangleq 1$ mm/d

3.2. *Actual evapotranspiration*

From temperature measurements at heights of 0.10, 1.50 and 3.00 m, sensible heat flow into the air can be calculated by the aerodynamic profile method. This method was already applied to Hupsel data for the summer period of 1976 (Stricker and Brutsaert, 1978). From the daily sensible heat flow, the daily rate of evapotranspiration can be derived by means of the energy budget at ground level by the equation:

$$\lambda ET^{AC} = Q^* - G - H \tag{4}$$

λET^{AC}, Q^*, G and H are the actual rate of vapour loss, net radiation, soil heat flux and sensible heat flux respectively, all expressed in W/m^2. Spil heat flux was calculated by the calorimetric method (Kimball and Jackson, 1975).

For 917 days, daily λET^{AC} could be calculated, but values from the winter half-year were of limited value, because of the relatively low accuracy in the estimate of H. For 119 days, values were missing.

For the month June, July and August, actual evapotranspiration was also estimated by the advection-aridity method, as outlined by Brutsaert and Stricker (1979). The basic idea stems from Bouchet (1963). Two empirical formulations are suggested here:

$$\lambda ET^{AD1} = 2 * \lambda ET^{PR+T} - \lambda ET^{PEN} \quad \text{(June, July, August)} \tag{5}$$

and

$$\lambda ET^{AD2} = 2 * \lambda ET^{PR+T} - \lambda ET^{TH+O} \quad \text{(June, July, August)} \tag{6}$$

in which λET^{AD1} and λET^{AD2} are supposed to be estimates of daily actual evapotranspiration (ET), expressed in W/m^2.

No further comment will be given here on the different methods, but the reader is referred to the preceding papers (Keijman, 1981; De Bruin, 1981) or to the papers cited.

4. RESULTS AND DISCUSSION

4.1. *Water balance*

All relevant components of the water balance were measured independently at Hupsel. Thus ET can be estimated as a remainder by the simple equation

$$ET^W = P - D + \Delta B \tag{7}$$

Table 2 Monthly totals of daily net radiation, rainfall, runoff and evapotranspiration, estimated by different methods and expressed in mm/month.
Note: The symbol E in tables and fig. is identical to ET in the text.

Time	R_n	*Precipitation*	*Runoff*	E^{AC}		E^{PEN}	E^{TH+O}	E^{PR+T}	E^{AD1}	E^{AD2}
March '76	18.1	27.9	10.7	not av.:	36 (estim)	35.2	39.6	–	–	–
April	59.2	10.5	5.1	32.6	(3.7)*	52.6	53.6	36.3	–	–
May	96.6	55.6	2.1	63.0	(23.6)	77.4	77.0	68.4	–	–
June	115.5	22.0	0.5	73.6	(28.2)	102.1	107.1	90.9	79.7	74.7
July	100.4	53.3	0.1	64.1	(9.4)	105.8	115.3	89.0	72.2	62.7
August	87.3	18.2	0	56.9	(28.7)	87.9	94.5	72.5	57.1	50.7
September	36.0	54.4	0	34.4	(1.3)	38.3	40.6	30.2	–	–
October	5.8	34.4	0.1	18.7	(0)	18.9	22.9	–	–	–
November	– 9.5	63.1	1.2	3.7	(0)	4.3	8.5	–	–	–
December	– 21.3	44.0	4.3	2.0	(0)	2.0	6.0	–	–	–
Subtotals	488.1	383.4	24.1	385	(94.9)	524.5	565.1			
January '77	– 11.1	71.6	21.5	1.1	(0.2)	0.2	3.8	–	–	–
February	2.9	67.1	34.0	10.3	(0)	8.9	12.4	–	–	–
March	32.0	40.9	15.2	28.5	(6.4)	31.0	33.2	–	–	–
April	49.5	61.5	23.1	39.7	(22.7)	38.6	37.6	28.0	–	–
May	94.3	55.8	97	73.1	(46.2)	79.4	79.0	64.6	–	–
June	81.7	96.0	84	64.8	(21.2)	64.3	60.8	61.4	58.5	62.0
July	92.3	50.2	3.8	70.6	(0.)	80.5	78.8	74.5	68.5	70.2
August	60.8	124.1	18.0	49.0	(2.3)	52.3	51.7	48.7	45.1	45.7
September	34.6	3.6	8.5	36.5	(0)	38.8	40.7	30.7	–	–
October	9.4	41.4	3.3	17.4	(0)	19.8	24.7	–	–	–
November	– 11.8	128.5	45.4	20.7	(0)	18.3	25.5	–	–	–
December	– 13.4	60.4	38.2	9.3	(3.6)	6.3	11.0	–	–	–
Subtotals	421.2	801.1	229.1	421.0	(102.6)	437.8	459.2			

Table 2 (second part)

Time	R_n	*Precipi-tation*	*Runoff*	E^{AC}		E^{PEN}	E^{TH+O}	E^{PR+T}	E^{AD1}	E^{AD2}
January '78	– 12.9	57.7	47.6	9.8	(0)	9.5	15.8	–	–	–
February	– 3.3	18.0	18.5	7.4	(0.2)	10.3	15.3	–	–	–
March	20.3	101.7	54.7	25.4	(0)	26.9	31.4	–	–	–
April	61.5	18.1	13.4	39.5	(0)	48.9	48.6	35.5	–	–
May	83.1	47.8	10.5	56.1	(1.4)	62.9	61.5	55.1	–	–
June	88.9	75.5	3.6	75.6	(0.9)	78.2	76.4	70.4	62.6	64.4
July	91.5	65.3	13.3	75.9	(4.1)	72.2	70.3	67.7	63.2	65.1
August	64.4	57.4	2.5	62.1	(5.0)	59.4	57.3	55.3	51.2	53.3
September	32.1	88.7	2.1	39.5	(3.7)	36.8	38.6	27.8	–	–
October	9.1	12.4	7.4	17.2	(1.8)	16.1	19.2	–	–	–
November	– 8.8	20.4	3.0	17.2	(1.9)	11.8	16.8	–	–	–
December	– 17.1	110.3	49.7	10.5	(1.6)	5.2	11.2	–	–	–
Subtotals	408.8	673.3	226.3	436.2	(20)	438.2	462.4			

* Values between brackets have been taken from either E^{TH+O} or E^{AD2}

where P : precipitation (mm)

D : discharge (mm)

ΔB : increase in moisture storage over the considered period (mm)

ET^W : actual evapotranspiration from the water balance (mm)

Note : This ET has a different dimension from that in energy balance, eq. (4).

From 76-03-01 to 78-12-31 (34 months), P and D were 1858 mm and 480 mm respectively. Storage increased by 40 mm, accompanied by a rise of the watertable of about 40 cm, which represents the average of observations from 18 groundwaterwells. Equation 7 then yields 1338 mm for ET^W. Table 2 gives montly sums of ET, calculated from daily rates with equations 1-6.

Because of non-potential ET during the dry summer of 1976, one cannot simply take the sum of each column as a reliable and meaningful estimate of actual ET of the catchment, except for λET^{AC}.

If the catchment suffered from drought only in June, July and August 1976, potential ET could not be reached.

No moisture deficits would occur in all other months and several reliable estimates of actual (catchment) evapotranspiration could be made by taking monthly values from different columns of table 2.

Table 3 shows the results and explains the combinations selected. The various water balances gave deficits, ranging from -20 to 97 over nearly 3 years.

Several effects may be responsible for these relatively small deficits:

- errors in data recording. Particularly errors in net radiation would influence the water balance result;
- representativity of the meteorological site. The meteorological data are point measurements. Daily mean values of wind, temperature and relative humidity would not vary

Table 3 Summary of the water balance: March 1976 – December 1978 (ET = evapotranspiration)

Method for ET	*P (mm)*	*D(mm)*	*ET (mm)*	*Increase of storage (mm)*	*Deficit (mm)*	*Deficit as % of P*
Water balance	1858	480	1338	40	0	0
E^{AC}	id.	id.	1241	id.	97	5.2
E^{PEN}, E^{AD1}***)	id.	id.	1256	id.	82	4.4
E^{TH+O}, E^{AD2}***)	id.	id.	1323	id.	15	0.8
E^{PEN}, E^{AD1}*)	id.	id.	1314	id.	24	1.3
E^{TH+O}, E^{AD2}**)	id.	id.	1358	id.	–20	–1.1
				Mean	40	2.1

*) E^{AD1} : only in June, July and August 1976.

**) E^{AD2} : id.

***) E^{AD1} and E^{AD2} : applied in June, July and August of 1976, 1977 and 1978.

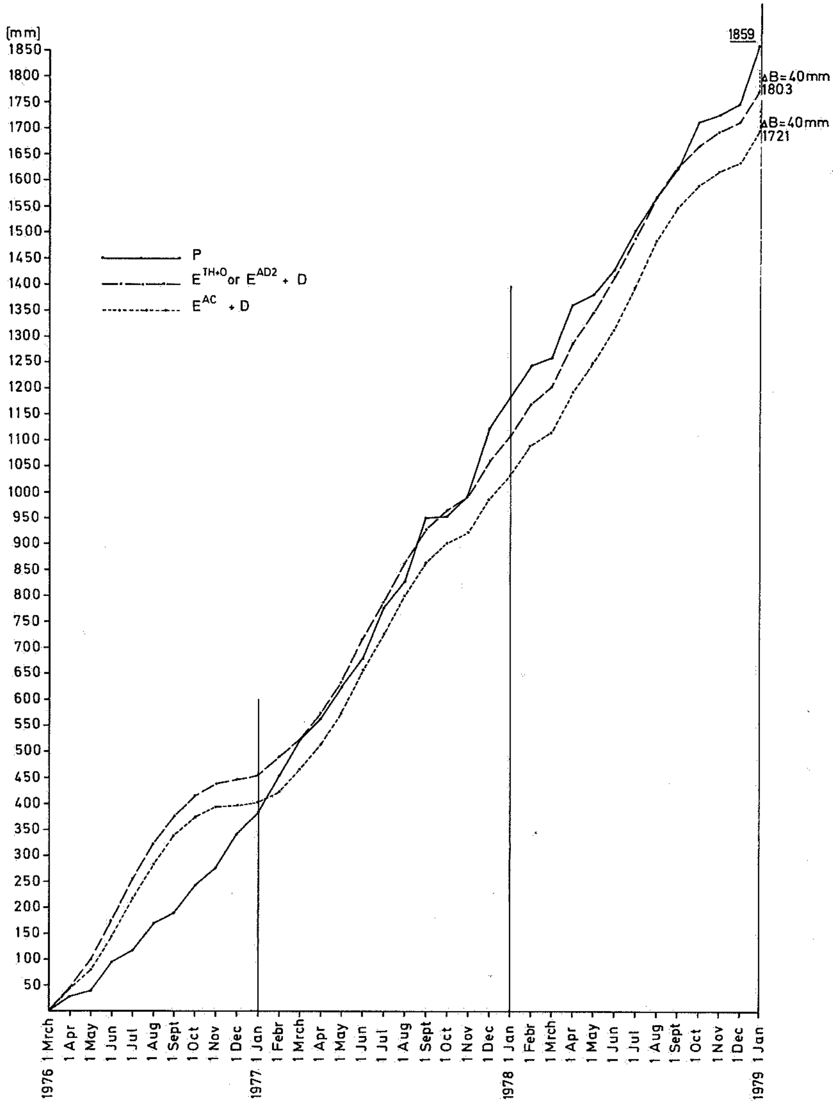

Fig. 2 Graphical representation of accumulated precipitation (P) and accumulated evapotranspiration (resp. E^{AC} and E^{TH+O} or E^{AD2}) plus discharge (D) against time. Period: March 1976 to December 1978.

significantly inside the catchment and would be fairly representative. But the temperature profiles, used to calculate λET^{AC}, would represent the more local conditions. Under moist soil conditions local differences in the rate of vapour loss will not be significant. But under moisture limiting circumstances it becomes relevant whether the profile data are collected at a representative site, reflecting average conditions in the catchment. So measuring sites have to be selected carefully and with some feeling;
- inaccuracy of the ET methods applied;
- some seepage may occur in westerly direction.

Figure 2 shows a graphical representation of accumulated P en (D + λET) against time for two methods to calculate evapotranspiration.

4.2. *Potential evapotranspiration by the methods of Penman and Thom & Oliver*

De Bruin (1981) showed that for the year 1977 and 1978 daily actual evapotranspiration, λET^{PEN} and λET^{TH+O} are in good agreement. Monthly, especially during summer, and annual values confirm that conclusion. λET^{TH+O} continuously exceeds λET^{PEN} during winter (October-March).

Thom and Oliver (1977) obtained a similar result for a catchment study in England and concluded, after careful analysis, that it is caused by the different structure of the ventilation term (second term) in the equations. They found a redistribution of ET of about 27 mm in favour of the winter period, if one applies the equation of Thom and Oliver instead of that of Penman. For Hupsel it amounts to about 5 mm. Nevertheless either method is appropriate to estimate potential evapotranspiration.

During the summer of 1976, soils became dry in the catchment. Particularly during June, July and August actual ET dropped below its potential level and results by the methods of Penman and of Thom & Oliver must be carefully interpreted. The equations are applied outside the boundary conditions for which they were formulated.

4.3. *Potential evapotranspiration by the method of Priestley & Taylor*

Basic to the use of an empirical relationship to estimate evapotranspiration is the need for local or regional calibration and definition of the conditions under which it can be applied. Priestley & Taylor (1972) defined their formula as an estimator of evaporation from saturated surfaces. But results from the literature (Rouse et al., 1977; Mc. Naughton et al., 1979; Mukammal et al., 1977; Davies and Allen, 1973; Priestley and Taylor, 1972) indicate that it estimates reasonably well potential ET of vegetation with non-limiting soil moisture and during the period of the year with relatively high net radiation.

It is therefore important to examine in which period of the year the Priestley and Taylor equation was applicable in Hupsel. The average ratio of ET^{PR+T} (α = 1.28) to ET^{TH+O} for 10-d intervals was calculated for 1976 (except June, July and August), 1977 and 1978 (figure 3). Seasonal influence is evident.

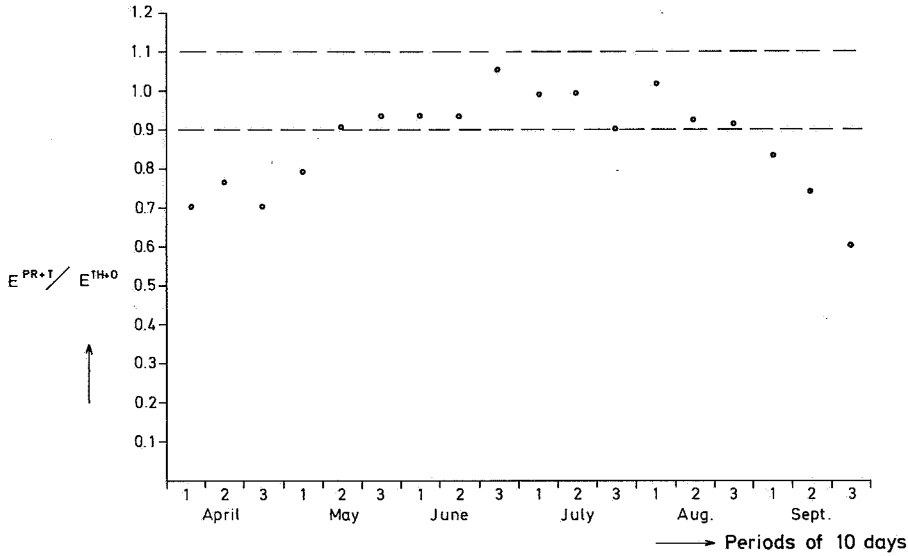

Fig. 3 Average ratio of E^{PR+T}/E^{TH+O} for periods of 10 days. Periods: April, May and Sept. over 1976, 1977 and 1978. June, July and Aug. over 1977 and 1978.

From the second half of May through August, the ratio is between 0.9 and 1.1. In spring, soil heat flux influences the ratio, as can be seen from the equations of Priestley & Taylor and Thom: the first term of λET^{TH+O}, in which λG appears, is about 50% of λET^{PR+T} and so λG has about twice as much influence on λET^{PR+T} as on λET^{TH+O}.

For α is 1.34 in eq. 3, the sums of daily evapotranspiration over June, July and August 1977, 1978 (potential conditions!) are comparable for eqs. 2 and 3. And to get comparable results of ET^{PEN} and ET^{PR+T} over the same period 1.38 would be necessary in eq. 3. This is fairly high for grassland. A value of 1.34 results from comparison of eq. 3 and actual evapotranspiration over the indicated period. This suggests that a value of α equal to 1.28 seems somewhat to low and with 1.34 for α ET^{PR+T} would be in better agreement with results from comparable methods.

4.4. *Use of the methods of Penman and Thom & Oliver or of Priestley & Taylor under moisture-limiting conditions*

For June, July and August 1976, there were noticeable differences between λET^{PEN} and λET^{TH+O} on one hand and λET^{PR+T} on the other hand. This fact was (partly?) caused of non-optimal moisture conditions in the soils in the region. During periods of moisture deficit, the conversion of available energy, Q*, into latent and sensible heat shifts in favour of the last component, so that at least the lower layer of the atmosphere becomes warmer and drier than with potential evapotranspiration. This change in the lower atmosphere enhances the ventilation term in eqs. 1 and 2. Thus, the second terms in λET^{PEN} and λET^{TH+O} are determined not only by climatic or large-scale atmospheric conditions, but also by *regional* moisture conditions. In fact, eqs. 1 and 2 estimate potential evapotranspiration under *actual* conditions of the lower, say 10 m, layer of air. That means for instance that if the micro climate is affected by irrigation at a regional scale, resulting in an increase in ET, estimates of λET^{PEN} and λET^{TH+O} will change too and will decrease.

Eq. 3 of Priestley and Taylor is considerably less sensitive to regional conditions of soil moisture and therefore, in my view provides a more realistic estimate of regional potential ET, however for the period June, July and August.

4.5. *Estimation of actual evapotranspiration by the advection-aridity method*

Because eq. 3 is a component of eqs. 5 and 6, the applicability of eqs. 5 and 6 is the same as for eq. 3:

Figs. 4 and 5 give daily ET^{AD1} and ET^{AD2} against ET^{AC}.

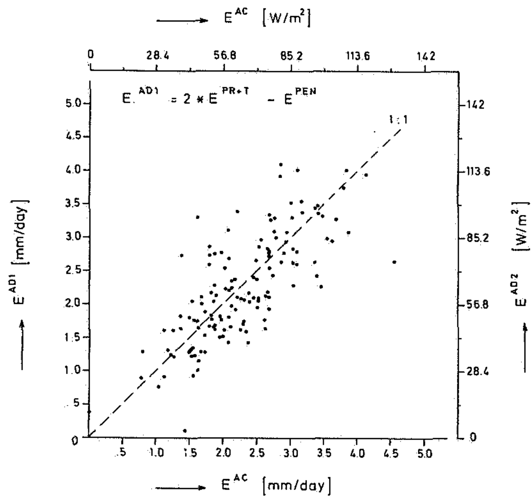

Fig. 4 E^{AC} against E^{AD1} for days with no or few precipitation ($P < 0.5$ mm/day). Period: June, July and Aug. of 1976, 1977, 1978.

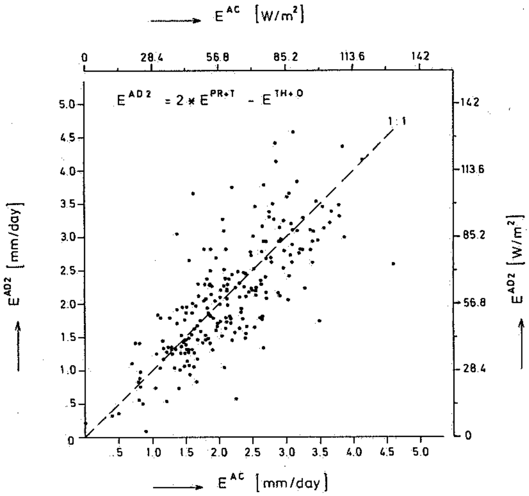

Fig. 5 E^{AC} against E^{AD2} for wet and dry days. Period: June, July, Aug. of 1976, 1977, 1978.

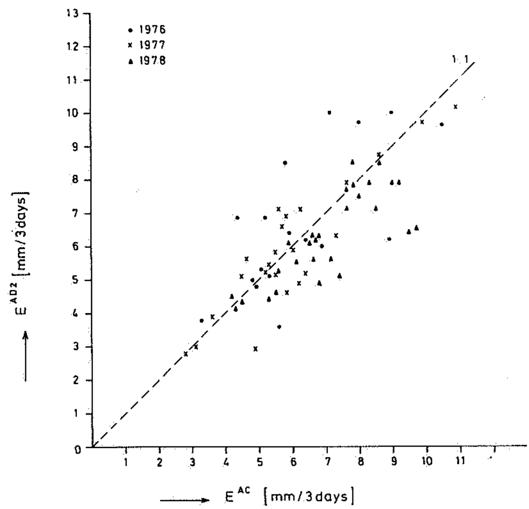

Fig. 6 E^{AC} against E^{AD2} as sums of 3 *consecutive* days. Period: June, July, Aug. of 1976, 1977, 1978. Total of 69 points.

Fig. 5 also includes values for wet days, which were inexplicable left out of fig. 4. Fig. 6 shows ET^{AD2} (preferred over ET^{AD1}) against ET^{AC} for sums of three consecutive days. In all figures, the 1:1 line was drawn. The method, applied to summer-periods of three years, of which one was dry (1976), proved encouraging, but needs more theoretical and practical evidence before it may become operational in future. In fact, the method was quantitatively based on the identical behaviour of eqs. 1, 2 and 3 under sufficiently moist conditions and on the different behaviour with limiting moisture conditions (section 4.4) of eqs. 1 and 2 on the one hand and eq. 3 on the other hand. Several other quantitative formulations for the Bouchet hypothesis are possible, but climatological calibration is always indispensable.

Although the method can only be used for a relatively short period of the year, it covers the period of possible moisture deficit under Dutch conditions.

4.6. *Interception*

Returning to the water balance, no evidence was found that interception played a marked role in evapotranspiration for pasture. Philips (1979) estimated the increase in evapotranspiration at 2 – 3 mm a month (with a canopy storage-capacity of 0.5 mm).

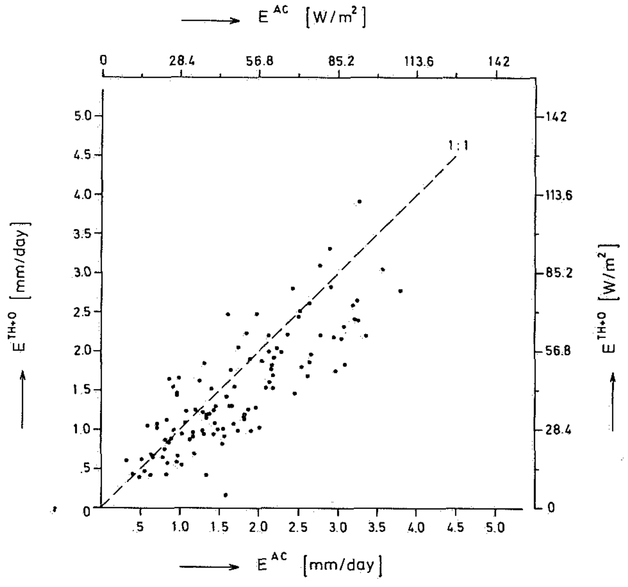

Fig. 7 E^{AC} against E^{TH+O} for wet days ($P \geq 05.$ mm). Period: Apr., May, June, July, Aug., Sept. 1977, 1978.

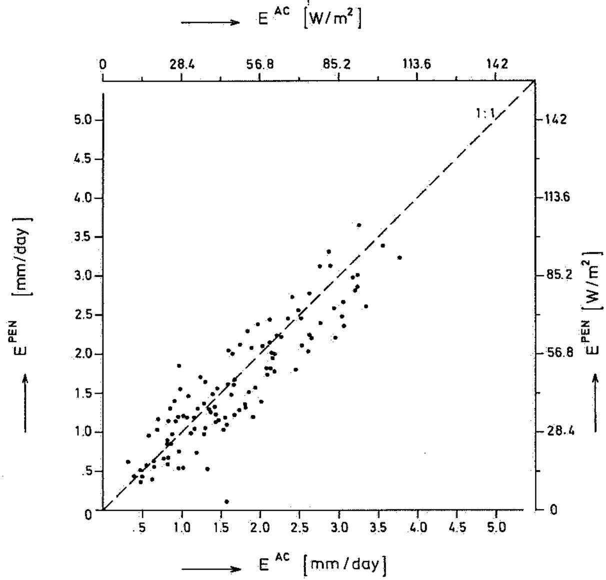

Fig. 8 E^{AC} against E^{PEN} for wet days. Period: as in figure 7.

He applied Rutter-type interception models (e.g. Gash, 1979) to hourly data of Hupsel over a winter period, making full use of eq. 2 for wet, dry and partially wet conditions. If we take 2.5 mm/month as an average value, interception over 34 months would increase ET by 85 mm; rather a different value to compensate several deficits in table 3. More evidence of the minor role of interception for pasture is given by figs. 7 and 8. The estimates of ET^{TH+O} (with n = 1.2) and ET^{PEN} are drawn against ET^{AC} for days with rainfall exceeding 0.5 mm/d. If interception plays any role, the 1:1 lines would not fit the data. Linear regression may yield a small deviation from the 1:1 line in fig. 7 in favour of ET^{AC}, but by no more than 0.1 mm/d on average or about 1 mm/month ($\simeq$ 10 wet days). The few, more sophisticated studies with lysimeters indicated, that interception would not markedly influence ET above grassland (McMillan and Burgy, 1960). However research in the last twenty years on interception of coniferous and hardwood forests has proved that interception is an important component of water loss from forests (e.g. Rutter, 1972, 1975, 1977; Pearce and Rowe, 1979).

Whether a forest lose more or less water than other vegetation is not easy to answer and depends highly on the amount, intensity and distribution of rainfall, on the type of forest and on the area covered.

Table 4 Comparison of $(P - E^{PEN})$ and $(P - E^{TH+O})$ of Hupsel against $(P - 0.8\ E_0)$ and $(P - 0.7\ E_0)$ of Winterswijk.
Period: March to September; years 1976, 1977, 1978

Year	Prec.* (Wint.)	Prec. (Hup.)	E^{PEN}	E^{TH+O}	E_0(Wint.)	$(P - E^{PEN})$	$(P - E^{TH+O})$	$(P - 0.8\ E_0)$	$(P - 0.7\ E_0)$
1976	187	187	461	487	653	–274	–300	–336	–270
1977	430	429	346	341	514	+ 83	88	+ 19	+ 70
1978	359	366	348	345	488	+ 18	21	– 32	+ 17

* Rain gauge: 2 dm^2 at 40 cm above ground level.

4.7. *Rainfall excess: Period March to August*

In a report of CHO-TNO (Van der Heide, 1977, figs. 12 and 13), openwater evaporation, E_0, for the year 1976 and the average E_0 are given for several meteorological stations in The Netherlands. For Winterswijk E_0 amounted to 653 mm in 1976 against an average of 544 mm.

In fig. 13 of the same report, potential rainfall excess was calculated over a six-month period from 1 March.

Rainfall excess was defined by the equation: $(P - 0.8 \times E_0)$ (mm).

$$E_0 = \frac{s}{s+\gamma}(R_n^{water} - \lambda G) + \frac{\gamma}{s+\gamma}\,\frac{3.7 + 4.0\,\bar{u}}{\lambda}\,\overline{\Delta e}$$

λG is normally neglected and R_n, $\bar{T}$, $\bar{u}$ and $\overline{\Delta e}$ are mean values over daylight hours (see Kramer, 1957).

Potential ET of grass is often expressed by $0.8 \times E_0$.

The average for Winterswijk is –61 mm, but in 1976 $(P - 0.8 \times E_o)$ decreased to –344 mm.

Rainfall excesses between Hupsel and Winterswijk were compared for sixmonth period of 1976, 1977 and 1978.

Rainfall excess in Hupsel was defined by $(P-ET^{PEN})$ and $P-ET^{TH+O})$ and rainfall excess at Winterswijk by $(P - 0.8 \times E_0)$, as usual, and by $(P - 0.7 \times E_0)$. The results are given *in table 4*. The calculation $(0.8 \times E_o)$ overestimated potential ET of grass, whereas $(P - 0.7 \times E_0)$ gave similar results to those from Hupsel. Thus, a factor of 0.7 seems a better estimate of potential ET for Winterswijk.

5. CONCLUSIONS AND RECOMMENDATIONS

– The small deficits in the different water balances of table 3 indicate that the methods of calculating actual and potential evapotranspiration are satisfactory. All methods were based on meteorological observations and not linked to soil moisture storage. The profile-energy budget method showed the highest deficit, perhaps because of inaccuracies in the method itself or in available profile data, especially during winter, or because of local variation in actual evapotranspiration, for which this method is more sensitive than the other methods.

– If there is no regional moisture deficit, eqs. 1 to 3 produce reliable estimates of real potential evapotranspiration. However eq. 3 can only be used within a limited period, from about the second half of May until September for Hupsel. Open-water evaporation, E_0, with a factor 0.8, gave too high an estimate of potential ET for grass and a better estimate seems to be E_0 with factor 0.7, whose wider application still needs to be tested.

– If actual evapotranspiration drops below potential during drought, eqs. 1 to 3 cannot be used to estimate actual ET. The advection-aridity method then seems to provide a fair estimate of actual ET. The method is attractive, because it needs routinely measured meteorological data. Climatological calibration is indispensable, as earlier indicated.

– In hydrology, wide use is made of Penman-type equations and reduction functions based on moisture storage to estimate actual evapotranspiration. As shown for pasture, a Penman-type equation may correctly be replaced by the Priestley & Taylor equation in summer. The advantage is that less data are needed. But a more important reason to recommend replacement is that during *drought* the Priestley & Taylor formula produces a more realistic estimate of regional potential evapotranspiration than a Penman-type equation. However a value of $\alpha = 1.28$ in the Priestley & Taylor-expression may be somewhat too low.

– In the Hupselse Beek Experimental Catchment, all components of the water balance were determined independently. Daily values (and for rainfall and discharge at much smaller intervals) of a period of about three years are available now and the period will be extended. Observations of 80 wells are also available and land-use is surveyed annually. In my view it makes the catchment very attractive for testing of operational hydrological models, such as numerical groundwater models and conceptual rainfall-run-off models. But it could also be attractive for multidisciplinary research on water management and the environment.

6. ACKNOWLEDGEMENT

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EVAPOTRANSPIRATION: ADVANCE AND FUTURE OUTLOOK

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SUMMARY

In this paper a review is given of the developments in evapotranspiration research in The Netherlands since the last Evaporation Symposium of the Committee for Hydrological Research, TNO (CHO-TNO, 1960). The emphasis has shifted from empirical methods (such as Penman's equation) to more physically justified models and later on to physiologically controlled processes (equations in which a canopy resistance is taken into account).

The problems in evapotranspiration research that still exist today are discussed and a future outlook is given on the need of evapotranspiration data and the way the research is expected to go.

1. INTRODUCTION

The principal objective of evapotranspiration research is to find methods for calculating the loss of water under varying conditions of climate, soil and vegetation.

Various groups need evapotranspiration data in their work, for example agriculturists, horticulturists, drinking-watersuppliers, hydrologists, polderboards, environmentalists. Besides these the predictive numerical models for medium-range weather forecasts (2-10 days) require that the effects of the planetary boundary-layer, such as fluxes of heat and water vapour shall be taken into account. For all these users it is interesting to see if, and if so how, the evapotranspiration research has evolved since the last Evaporation symposium of the Committee for Hydrological Research, TNO, (CHO-TNO, 1960) some 20 years ago.

We will start by looking to the state of evaporation research in 1958-1960 and try to figure out the developments in different subjects of evaporation during the last 20 years. Doing so we will arrive at the state of art at this moment with the problems yet unsolved. At last we will have a future outlook on the need of evapotranspiration data and on the way the research is expected to go.

2. STATE OF THE ART OF EVAPORATION 1958-1960

Many articles in the proceedings of the Evaporation symposium (CHO-TNO, 1960) concentrated on the computation method by Penman. The calculation of the evaporation from a free water surface according to Penman was simplified by using tables, graphs and nomograms.

Besides the Penman-equation, the formula of Thornthwaite-Holzman has been used for an atmosphere in neutral equilibrium. Makkink pointed out the lineair relation between open-water evaporation E_o and measured global radiation and Rijtema compared calculated values of the potential evaporation E_p after Penman, Makkink, Turc and Haude with measured evapotranspiration values from lysimeters and evaporation pans. He concluded that it is possible to calculate E_p according to Penman, Makkink and Turc from meteorological data with the same accuracy as is obtainable with lysimeters and evaporation pans.

The eddy-correlation technique was developed only very recently at that time. The use of lysimeters was still in full swing; further experimental methods were the water balance, the energy balance and the aerodynamical method (vertical transport of water vapour). The Rottegatspolder acted as an experimental catchment area for evapotranspiration research.

Research has been done on the uncertainty in the evaporation of a free water surface computed according to Penman's method and on the influence of a lake's depth on the validity of the Penman formula and the assumption that the soil heat flux $G = 0$.

Makkink reported about the influence of advective heat on the actual evapotranspiration from any vegetation. More detailed studies were wanted at that time on vegetation factors for the relation between the actual evapotranspiration and E_p or E_o. Furthermore there was a need for cumulative frequency distributions of precipitation surplus in the growing season.

3. DEVELOPMENT DURING THE LAST 20 YEARS

Briefly it can be said that the emphasis in evapotranspiration changed from the early simple empirical methods to methods which more closely represent the physical and biological processes. The last years signaled a shift in evapotranspiration research from physically controlled processes to evapotranspiration as a physiologically controlled process, such as the study of the stomatal resistance.

3.1. *Development in theory and formula's*

Measurements of the energy budget at ground level were continued in the Rottegatspolder until 1969. Results of the energy balance were compared with those obtained by the water balance and the aerodynamical method. In 1975 extensive evaporation research started at the KNMI-ground in Cabauw with the energy-balance method and eddy-correlation measurements. The aerodynamical method or profile method was developed further by De Bruin and Kohsiek (1979).

As already stated by Keijman (1981) energy balance equations can be combined with transport equations leading to so-called combination formula's.

The oldest combination formula, the model of Penman has been discussed in this symposium by De Bruin (1981) and Keijman (1981). The concept for potential evaporation

$E_p = f. E_o$ is an approximation method with a strongly empirical character, due to the empirical constant in the windterm and f. Rijtema (1965) stated: "due to the development of the crop and to a possible lack of water, the evapotranspiration has under many conditions no direct relation with the evaporation from a free water surface. It is there fore necessary to take all the factors governing the real evaporation into account". Rijtema's thesis gives an analysis of the most important factors determining the actual evapotranspiration. These are:

- the transport of water vapour from the air layers close to the evaporating surface to higher layers;
- the amount of energy available for the vapourization of water;
- the aperture of the stomata in connection with the diffusion of water vapour through them;
- the rate of supply of water to the evaporating surface.

Each of these factors can act as a limiting factor for the evapotranspiration. On account of this Rijtema derived an equation in which the factors determining the actual evapotranspiration are taken into account. So the canopy resistance term, found already in Monteith's model was expanded by Rijtema and later by Feddes to include terms for stomatal resistance, resistance dependent on the availability of soil moisture and on liquid flow in the plant, and resistance dependent on the fraction of soil cover. Ziemer (1979) called the Rijtema – type model a substantial improvement upon the Penman – Monteith estimate.

Various other combination-formula's have been set forth such as Stricker – Brutsaert (1978), who combined the energy-balance method with the profile method.

The usefullness of the Priestley – Taylor formula was tested by De Bruin and Keijman (1979) on Lake Flevo, who found a diurnal and seasonal variation in the constant α (mean value 1.26).

The advection-aridity method of Brutsaert and Stricker (1979) has been applied to the Hupselse Beek catchment.

Recently the planetary boundary-layer theory has been applicated to evaporation models; this requires additional research into the functional form of the similarity functions for sensible heat and bulk water vapour transfer, under various conditions of atmospheric stability (Brutsaert and Chan, 1978).

3.2. *Instrumentation*

The search for a simple economical, reliable evaporation pan continued in the first years of the period. Van Wijk and De Wilde (1962) concluded that it was impossible to use fixed pan coefficients for vegetation under different climatic and exposure conditions. The WMO Technical Conference on Assessment of Areal Evaporation in Budapest (1977) discouraged the further development of evaporation pans.

The research with lysimeters yielded a large number of data for setting up water balances for various vegetations, types of soil and climates. The lysimeter research has contributed highly in the development of relations which can be applied in hydrological research; for example at Castricum the interception on trees has been studied.

Experiments in catchment areas have been done in Tielerwaard-West and in the Leerinkbeek area, later Hupselse Beek area, with the purpose of studying the effects of reclamation works and the testing of instruments. The profile method, the Bowen ratio and the water balance have been compared to assess the representativity of point measurements with routinely determined data for areal evapotranspiration.

In 1969 research started in the Sleen catchment on mechanisms determining the relation between the components of the water balance and research on the relation between evapotranspiration and soil moisture availability.

Great strides have been made in the development of remote sensing techniques for measuring reflectance and surface temperature. As the surface temperature of the leaves depends on the amount of water available to evaporation, remote sensing can be used to determine the water stress of vegetation. Furthermore it is possible to calculate actual evapotranspiration rates from the remotely sensed surface temperatures with the aid of the energy-budget equation, the aerodynamical profile equations, incorporating soil-plant-water relations (Soer, 1980). Keijman and De Bruin (1979), however, pointed to large discrepancies that can occur between the surface temperature measured by Infra Red Line Scanning-techniques and the theoretical temperature.

Remote sensing measurements have been made from low laying platforms (at Cabauw, Barnes and Heijmann), from aeroplanes and from satellites (NOAA-TIROS-N, Tellus-project centered around EXPLORER-A bearing sensors for Heat Capacity Mapping Mission).

Besides all this, improvements have been made in measuring the different components of evapotranspiration models such as neutron moisture-probes for soil moisture, fast-response sensors for wind, temperature and water vapour and of course the enormous developments in registration and data collection by the advancement of the computer.

3.3. *Lake evaporation*

In 1967 evaporation measurements have been made at Lake Flevo where water balance, energy balance, aerodynamical- and Penman models have been compared.

For the evaporation from lakes the water-surface temperature T_w has to be known. Keijman (1974) published a method to calculate E from isothermal lakes, without determining T_w. Furthermore he showed that the evaporation of a lake can be estimated rather well from simple meteorological data of a station alee.

De Jong and Keijman (1971) treated the evaporation of an estuary where a periodical tide run has to be taken into account. Verhagen (1977) published an article on non-isothermal lakes with a sudden temperature jump and Sweers (1976) connected the wind function in Penman's formula to the area of the lake.

3.4. *Soil water evaporation*

The complexity of the evaporation of water from the soil has been found from laboratory and lysimeter experiments. As the soil dries energy availability becomes less important and the rate of soil water conduction becomes more important. Precipitation and irrigation have direct and indirect effects on soil temperature and hence on evaporation.

Van Bavel and Hillel (1976) stated that the transition from the energy-limiting phase to the soil-limiting phase is not due to changes in albedo, but to the hydraulic properties of the soil and to a reduction of the relative humidity at the surface to values less than one. They proposed an extension of the Penman equation by incorporating terms related to the hydraulic and thermal properties of the soil profile.

3.5. *Evapotranspiration of vegetation*

In this case the soil-plant-atmosphere system, of which no one part operates independently of the other, has to be considered. When evaporative demand exceeds the ability of the roots to supply the necessary water, some species close their stomata. The role of stomata in the regulation of transpiration has been widely studied. Stomatal resistance has been measured by Stigter (1975) who used a closed diffusion porometer as the most suitable device for field measurements on separate leaves and by Kohsiek (1979) who used a perspex chamber in open or closed condition with circulating air to measure the bulk stomatal resistance of a grass surface. In the last twenty years understanding of the factors on which stomatal resistance depends has increased.

According to Rijtema the aerodynamical resistance in the Penman-Monteith's equation can be related to the inverse of a roughness function. This function is the product of vegetation height and a dimensionless function of wind speed.

Feddes et al. (1976) proposed a model that takes into account a root extraction term, depending on potential evapotranspiration, soil moisture content and the depth of the roots.

Soer (1977) tested the TERGRA-model, describing the terms of the energy balance of a vegetable surface, on grassland (Cabauw) and wheat (Flevoland).

3.6. *Interception*

Relatively little research has been done on the evaporation of forests and interception in The Netherlands. The research concentrated on 4 lysimeters at Castricum of which one has a foliage-tree vegetation and one a coniferous vegetation. Rijtema and Ryhiner (1968) found that interception gives a big increase in evapotranspiration of coniferous forests. In the Hupselse Beek catchment it has been found that interception did not influence evapotranspiration in a noticeable way for grassland (Stricker, 1981).

3.7. *Advection*

The influence of advection of sensible heat from a relatively dry area to a more moist area is a problem which has long plagued attempts to evaluate evapotranspiration. The energy used for evapotranspiration can exceed the energy supplied by net radiation by a factor of two. Under advective conditions the Bowen-ratio underestimates evapotranspiration. For the lysimeters at Castricum negative advection with moist sea breezes can reduce the evapotranspiration. With this phenomenon we arrive at one of the problems in evapotranspiration research which still exists today.

4. PROBLEMS WITH EVAPORATION RESARCH TODAY

As stated in the last sentence of the preceding paragraph, one of the problems still existing is the influence of regional and local advection upon calculated evapotranspiration. Another point which is still under investigation is the relation between actual and potential evaporation.

Ziemer (1979) put it this way: "We have advanced a great deal in understanding the physical and biological controls on evapotranspiration. The ability to apply models to field situations is less succesful, particularly in forested areas and in other areas where data are lacking. We are still unable to predict the effect of timber cutting, wildfire, changes in species composition, or other cultural activities on watershed water balances. In many wildland areas, we are unable to measure adequately even areal precipitation – let alone the rather detailed meteorological data required to calculate evapotranspiration with the Penman-Monteith equations. Our ability to calculate accurately evapotranspiration within that cover condition between bare soil and full cover is still weak, particularly as to areal water loss from scattered vegetation of different species and sizes".

The uncertainty in the evaporation of a free water surface and the evapotranspiration of a wet vegetation calculated with the Penman-equation is about 20 percent, as can be seen from tables 1 and 2, which have been taken from De Bruin and Kohsiek (1979).

There are still problems in incorporating evapotranspiration into integrated models for watermanagement: It is impossible to verify the evapotranspiration calculated with an integrated groundwater current model with the measured evapotranspiration for a whole catchment area and for short periods (10 days). This has two causes:

- the methods measuring evapotranspiration accurately to this only for point measurements. Application on a bigger scale (more measuring points) is too costly;
- the methods which are suited for estimating areal evapotranspiration (remote sensing) are not reliable enough.

The same problem arose from the Hupselse Beek project. Is it possible with the collected data to form a water balance, for what period and how reliable? What is the value of such a balance for the evapotranspiration research?

Table 1. Error analysis of the Penman-equation applied to a free water surface (10-day or monthly totals).

Error source	Systematical error	Random error
neglecting the soil heat flux G	spring: +20% autumn: −20% year: 0%	– – –
wrong estimation of net radiation Q*	summer: ±10%*) autumn: +20% year: +20%	10%
wrong wind function	year: 5%	–
random errors from input data and reading errors	– – –	summer: 10% winter: 20%

) If Q less than 100 Wm^{-2} then +10%
if Q* greater than 140 Wm^{-2} then −10%

Table 2. Error analysis for the evapotranspiration of a wet vegetation.

Error source	Systematical error	Random error
neglecting the soil heat flux G	summer: 3% year: 0%	– –
wrong estimation of net radiation Q*	summer: 10% autumn: 10% spring: 10%	5% 10%
wrong wind function	?	10%
vapour pressure deficit e_s-e	–	5%
Total	10–13% + ?	20–25%

5. FUTURE OUTLOOK

The data mostly wanted are reliable areal evapotranspiration data. It is still questionable whether remote sensing techniques from satellites will give an exact answer to this. It is to be expected that in the future remote sensing techniques will be more reliable than nowadays. A drawback of using satellite data is, however, that polar orbiting satellites pass only once a day. So the diurnal variation in evapotranspiration can only be estimated by mathematical models that are verified with satellite data. Moreover remote sensing nowadays gives only relative surface temperatures, so that on the other hand satellite data have to be calibrated with additional measurements on the spot.

It can be expected that in the future the research on catchment areas will go on, to improve our knowledge of evapotranspiration as a function of atmospherical and soil physical circumstances.

The WMO should promote the conduct of a number of large-scale intensively-instrumented experiments, such as the present GARP land-phase experiment in France, in which large heterogeneous areas are broken down into smaller separate homogeneous sub-areas, in each of which evaporation is measured by suitably accurate techniques.

Some improvements can be expected in measuring meteorological data for evapotranspiration calculation, in data-handling and data archiving; humidity measurements by means of ultra violet Lyman-α-techniques; wind measurements with sonic- or thrust anemometers and temperature measurements with fast responding thermistors, thermocouples or Pt-wire in behalf of the eddy-correlation method. Movable masts for eddy-correlation measurements with microprocessors will help to extend point measurements to areal means. At last the temperature fluctuation method that has been developed recently can be mentioned (De Bruin, 1981).

The accuracy of the radiosonde data in the lowest 2 km is expected to improve which can be a stimulus for the regional evapotranspiration model of Brutsaert and Mawdsley (1976) which uses rawinsonde data.

Some research will be done on the reliability and comparability of lysimeters.

For the routinely determination of evaporation the WMO Technical Conference on Assessment of Areal Evaporation (1977) recommended informally net radiation networks using Funk-type instruments. Whereever possible these should be supplemented by hourly measurements of solar radiation, air temperature, humidity and wind. At key stations additional direct measurements of vertical profiles of air temperature and humidity should be taken for energy balance or Bowen-ratio models.

It is possible that in the future more measurements will be made for calculating evaporation according to the Penman- or profile method. Maybe that more measurements will be made of the vertical profile of soil moisture and tension and more work will be devoted to interception on arable crops.

Due to the computer, integrated mathematical-hydrological models are expected to be developed which can take into account the influence of human activities and changes in land use on the evapotranspiration.

Furthermore the computer has the advantage that lots of data can be handled and computations can be done on a real-time basis. This means that the user may have the disposal of evaporation data at short notice, in every kind of output he wants (total values, mean values, frequency distributions). In the future the evaporation data might be combined with precipitation data so that real-time groundwater levels can be calculated, dependent on the type of soil.

Although evapotranspiration is essentially a physical process it is not to be expected that in the future models will be developed that can take into account all atmospheric, crop and soil properties influencing the process. As the complexity of the models increases the data requirements to drive the equations often make the model useless for routine applications. This means that also in the future the calculation of evapotranspiration will be a compromise between empirical and totally physically justified models.

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