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J. C. Hooghart

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CONTENTS

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METEOROLOGICAL USE OF EVAPORATION DATA

C.J.E. Schuurmans

Data on evaporation to be used in agriculture, hydrology, forestry, etc. are usually supplied by meteorologists. Meteorologists themselves also use evaporation data. Air mass properties determining weather are strongly dependent on the input of water vapour from the surface. So for weather prediction purposes evaporation data, or rather methods to compute evaporation are needed.

This situation is not new. It has been observed already by Wartena in the Proceedings of Technical Meeting 38 (Wartena, 1981). New is the fact that at present operational weather prediction models indeed include computation of evaporation as an interaction between the atmosphere and the underlying surface. New is furthermore the emerging evidence that evaporation processes not only influence short time weather developments but also more long range and large scale changes of the atmospheric circulation. E.g. lack of soil moisture may cause persistence of drought producing circulation anomalies. Finally, evaporation as a component of the global hydrological cycle plays an extremely vital role in numerical simulation experiments of world climate.

These recent developments have made evaporation of major interest for use by meteorologists and physical climatologists. The former supplier has become a major user!

At this Technical Meeting, entitled Evaporation and weather, we do not fully enter the new fields of applications of evaporation. We rather start with the traditional role of meteorology to supply evaporation data for use in agriculture, hydrology, etc. Papers by De Bruin, Feddes and Lablans introduce the new practice adopted for computing evaporation data for the Netherlands on a daily basis. The new method is based on a simple formula, introduced by Makkink, to estimate evaporation from observed data on global radiation and air temperature only. It replaces the method based on the well-known Penman-formula. Evaporation data thus produced refer to the so called reference value of evaporation. These values, with due correction for different types of vegetation are suited for use in technical applications.

It seems logical that evaporation data as referred to above eventually will be produced as a by-product of limited area numerical weather prediction models. Present operational models are still too poorly resolving the small scale differences in evaporation. The present meeting however discusses the basic principles of such models. Papers by De Bruin on the physical aspects and by Holtslag on a specific model do give an impression of the potential capabilities of weather models in regard to the estimation of evaporation. The more far reaching developments related to evaporation, or rather hydrology in general, in large scale global circulation models (GCM's) are discussed in the paper by Shuttleworth. He introduces the term macrohydrology, defining first of all those activities which seek to improve the incorporation of hydrological processes including evaporation into GCM's. Improvements have to come from studies into the relation between small scale hydrological processes occurring in nature and their average process descriptions for large areas, used in GCM's. Such studies necessarily also include field experiments in areas of different terrain. In such experiments satellite observations may play an important role.

So on the one hand we still have the user of evaporation data on scales of tens of kilometers or less, while at the other hand evaporation process descriptions are needed as inputs for large scale global climate models on scales of the order of 300 x 300 km. Since these descriptions already are used in some operational weather prediction models (e.g. in the model of the European Center for Medium Range Weather Forecasts on which our 5-day weather forecasts are based) it is timely to present information on these developments at this Technical Meeting.

I expect that this meeting being called Evaporation and weather in a few years from now will have to be followed by one covering the subject from a climate modelling viewpoint. The emergence of global scale hydrology as Eagleson calls it (Eagleson, 1986) brings evaporation and other hydrological processes to the forefront of research and application as well.

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FROM PENMAN TO MAKKINK

H.A.R. de Bruin

1 INTRODUCTION

Since 1956 the Royal Netherlands Meteorological Institute (KNMI) publishes on a routine base an evaporation figure (E_o) which is evaluated with Penman's formula. From the first of April 1987 the KNMI has changed over from Penman's equation to the formula proposed by Makkink (1957). It is the objective of this paper to explain the background of this alteration.
In section 2 a brief review is given of the Penman formula: the physics on which it is founded is treated; its applicabilities and its limitations are discussed. It will be shown that, in spite of the fact that Penman's equation is based on a lot of physics, most of its practical applications are primarily empirical.
In the Netherlands, the majority of these applications concern the so-called *crop factor method*. Herein the evaporation figure E_o, published by the KNMI, is multiplied by a suitable crop factor to obtain an estimate of the evapotranspiration of an optimally growing crop, no short of water, under the prevailing weather conditions.

For theoretical and practical reasons in the past decades numerous modifications of Penman's formula have been proposed.
These concern for instance the estimation schemes for (net and global) radiation or the influence of the wind and stability of the air. Some of these alterations improved parts of the Penman equation from a

physical point of view. Unfortunately, introduction of these improvements did not always improve the skill of the (empirical) applications. Examples are known that after introduction of e.g. a "better" wind function the final result appears to be much worse than that of the original method. This was due to the fact that in the original approach errors were cancelling out, so that an "improvement" of a single part caused an inbalance and leads to worse results.
In course of time, the KNMI has changed the way E_o is evaluated for practical reasons (De Bruin, 1979; Lablans, 1987). Correction schemes were developed in order to avoid inhomogeneities in the E_o-series.

The result of the developments described above is that Penman's formula experienced a large number of changes in the last decades and that at this very moment tens of different versions of the formula exist. This causes a tremendous confusion.
This confusion is the main reason that it has been decided to stop the routine use of Penman's equation and to apply in the future Makkink's formula.
The reasons to choose the equation by Makkink are the following:

a. its behaviour is very similar to that of the Penman formula;
b. it is remarkably simple: it requires only air temperature and global radiation as input. Both can be measured directly and very accurately;
c. under dry conditions Makkink's formula appears to have even a beter performance.

These aspects will be discussed in section 3. A detailed comparison of the two methods will be presented.

It should be stressed that the new evaporation figures according to Makkink are meant only to be used in the crop factor method. For this purpose new crop factors have been determined. These will be presented by Feddes (1987). This author will show also the limited accuracy of the crop factor approach.

In section 3 it will be pointed out that Makkink's formula has also limitations.

2 THE PENMAN FORMULA

2.1 General

Penman (1948) combined the aerodynamic formulas for the vertical transfer of sensible heat and water vapour (Dalton's equation) with the surface energy balance equation.
He considered the case that the air at the surface is saturated, i.e. $e_o = e_s\,(T_o)$ and he approximated $e_s(T_o)$ by $e_s(T_a) + s\,(T_o - T_a)$. For the symbols see Appendix. In this way he derived his well-known formula that in our notation reads:

$$E = \frac{1}{\lambda}\;\frac{s(Q^{*}-G) + \rho c_p\,[e_s(T_a) - e_a]\,/\,r_a}{s+\gamma} \tag{1}$$

Note that in other publications often a wind function f(u) is used in stead of r_a (see Appendix).

Tacitly, Penman assumed that the surface is horizontally uniform, so that advection effects can be ignored.
Eq. (1) applies to both open water and wet land surface, but it is noted that the quantities Q*, G and r_a strongly depend on the surface properties.

The evaporation figure E_o, published by the KNMI, is evaluated with Eq. (1) taking G = 0 and using the semi-empirical relations for Q* and r_a described by e.g. De Bruin (1979). For further information see De Bruin (1979), Buishand and Velds (1980) and Lablans (1987).
Strictly speaking, Penman (1948) developed his formula to describe the water loss of the evaporation pan he used at his experimental site. Partly, he fitted constants of his equation to his pan data. In particular this concerns the "wind function", i.e. in our notation the dependence of the aerodynamic resistance r_a on the wind speed. So, the original Penman formula is based on a mixture of physical principles and empirical facts.

Moreover, Penman (1948) introduced an empirical method for the estimation of evapotranspiration from a well-watered short grass cover, being a version of the *crop factor* approach (see Feddes, 1987).

It is the merit of Penman that he was one of the first who recognized the significance of (net) radiation for the evaporation process. In the forties hardly no direct measurements of net radiation existed, so Penman had to estimate Q* using semi-empirical expressions. It is not suprising that later research revealed that Penman's estimation schemes for Q* needed revision. For more details see Holtslag (1987).

Penman ignored the term G in Eq. (1). For his evaporation pan this did not cause serious problems, however, large errors are made if G is neglected considering "real" open water such as lakes and rivers. For a water depth of 10 m G can easily exceed Q*. In section 2.2 a further discussion of open water evaporation is given.

As noted before Eq. (1) applies also to wet land surfaces, i.e. the surface is covered with a thin layer of water. If the surface is dry or partly wetted things become more complicated. In the sixties Monteith (1965) and Rijtema (1965) modified Penman's formula for a dry vegetated surface. This is discussed in section 2.3.

In the late sixties and the seventies a number of micrometeorological measurements of evapotranspiration were collected. It was found that for short well-watered crops this quantity is primarily determined by the available energy (Q*-G). This leads to the formula by Priestley and Taylor (1972), which has been confirmed also for Dutch conditions (Brutsaert, 1982; De Bruin, 1981). This is discussed in section 3.1. Net radiation is well correlated with global radiation (except in winter time). In this way the formula of Makkink (1957) can be obtained from the Priestley-Taylor equation.

As early as 1963 Bouchet realized that the parameters in Penman's equation are not independent. If water vapour or heat are brought into the atmosphere the water vapour deficit $D = e_s(T_a)-e_a$, appearing in the last term of the equation, will be altered. Hence, E and D are interrelated. To describe this effect an additional model for the planetary boundary layer is needed. In the eighties such models for evaporation have been developed e.g. by De Bruin (1983), McNaughton and Spriggs (1986) and Ten Berge (1986). These approaches reveal that formulas by Priestley-Taylor and Makkink have a much stronger physical base than one should expect at first sight. These aspects are discussed by De Bruin and Holtslag (1987).

2.2 Evaporation from open water

Penman's equation (1) describes properly the evaporation from open water. However, its application meets several problems.
First of all, the term G is difficult to determine, whereas it can be of the same order as Q*. Generally G can be written as

$$G = \rho_w c_w h \frac{\partial \tilde{T}_w}{\partial t} \tag{2}$$

where $\tilde{T}_w$ is the water temperature averaged over the depth. For well-mixed water, T_w, is constant with depth. For that case Eq. (2) can be combined with the governing equations leading to Eq. (1). Keijman (1974) showed that then the water temperature T_w is described by a simple differential equation

$$\frac{\partial T_w}{\partial t} + \frac{T_w}{\tau} = \frac{T_e}{\tau} \tag{3}$$

in which the *equilibrium temperature* T_e is given by

$$T_e = T_n + \frac{Q^*_n}{A} \tag{4}$$

and the *time constant* τ by

$$\tau = \frac{\rho_w c_w h}{A} \tag{5}$$

$$\text{where } A = [4\sigma T_n^3 + \frac{s+\gamma}{\gamma r_a} \rho c_p] \tag{5'}$$

Note that T_e and τ are determined solely by meteorological factors.

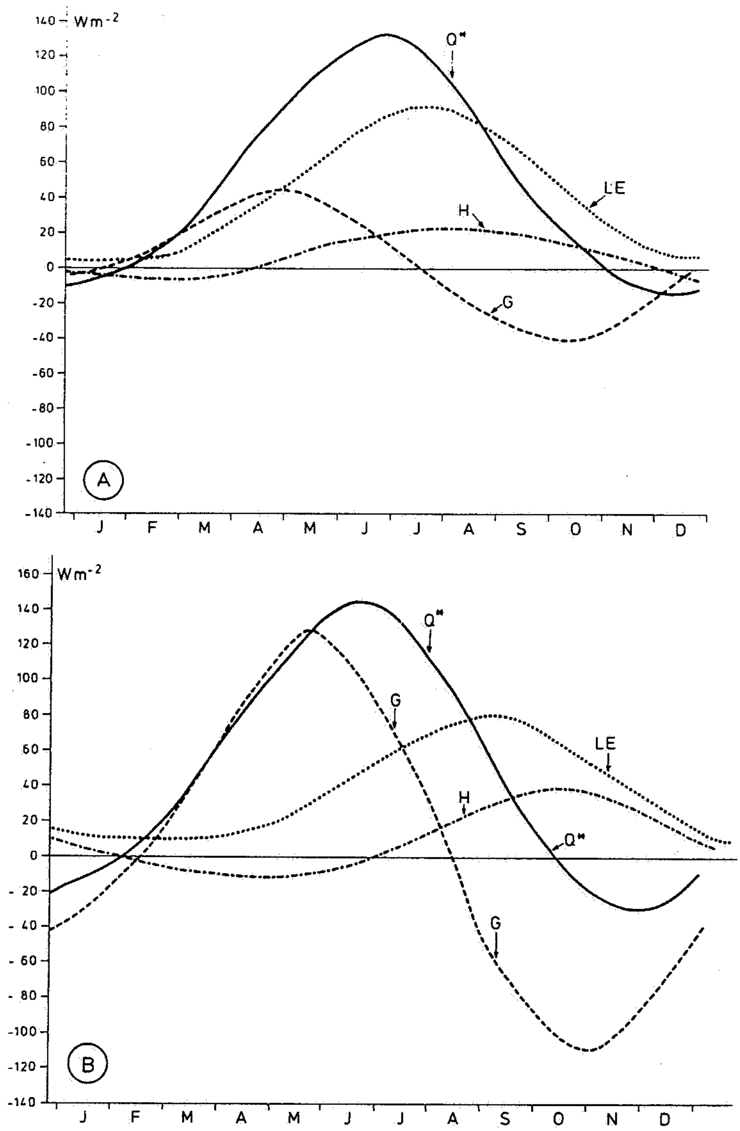

Figure 1 The mean annual cycle of net radiation Q*, sensible and latent heat fluxes H and λE, and the heat storage term G as evaluated by the model: (A) water depth 5 m; and (B) water depth is 15 m. (From De Bruin, 1982.)

In (4) T_n is the wet-bulb temperature at screen height.
Eq. (3) allows the evaluation of T_w and thus also that of G from weather data only. The problem however, is that these weather data have to be observed over the water surface itself. In practice, these data are only available at a nearby landstation. It appears that this problem can be overcome by using an adapted empirical expression for r_a. De Bruin (1982) showed that this approach yields good results for two adjacent lakes with different depths (5 and 15 m) in the Netherlands using the r_a proposed by Sweers (1976).
In Fig. 1 some results of his application of Keijman's model using such an empirical r_a are depicted. Note the significant influence of the water depth on G and through that on E.
In his original paper Penman neglected G. From the above it follows that this is certainly not permitted. Moreover, since Penman fitted his r_a to pan evaporation data, his r_a is not suitable to be applied to "real" open water (see e.g. De Bruin en Kohsiek, 1979).
Moreover, Penman did not take into account the fact that the annual average of the water surface temperature is higher than that of the air temperature. As a result Penman overestimated Q*. Due to these features, the annual E_o-values as published by the KNMI are 10-15% greater than the actual annual evaporation from open water. Thus, the annual values of E_o published by De Bruin (1979) and Buishand and Velds (1980) overestimate evaporation for open water, in spite of the fact that the annual mean of G is about zero.
Wessels (1972) and Schouten and De Bruin (1982) show that Keijman's model can be applied also to rivers (Rhine and Meuse respectively). In these cases the method is used to determine the thermal pollution for these rivers.

2.3 Evapotranspiration from crops

Using the same physics as Penman, Monteith (1965) derived a formula that described the transpiration form a dry (extensive - horizontal uniform) vegetated surface. In international literature this is denoted as the *Penman-Monteith equation*. In the Netherlands the name

of *Rijtema* is added, because this author derived independently a similar formula (Rijtema, 1965).
It reads:

$$\lambda E = \frac{s(Q^*-G) + \rho c_p D/r_a}{s + \gamma\ (1 + r_s/r_a)} \tag{6}$$

where $D = e_s(T_a) - e_a$

(For the symbols see Appendix.)

Experiences show that Eq. (6) successfully describes the transpiration as well as the interceptive loss from different kinds of vegetation such as tall forests, arable crops, heathland and grass.

In Monteith's concept the vegetation layer is described in a very simple way: it is treated as if it were one "big leaf". To this leaf a *canopy resistance* or *surface resistance* is assigned that accounts for the fact that water vapour has to escape from the "stomata" of the "big leaf" to the surrounding air. Within these "stomata" the actual transpiration process takes place (liquid water changes phase here), so that the air within the "stomata" will be saturated at surface temperature T_s.
The Penman-Monteith equation is derived for a dry crop completely shading the ground. If it is covered with a thin water layer r_s becomes zero and the original Penman formula is obtained. So, Eq. (6) describes also the *interception loss* properly as long as the canopy is fully wetted. It is still not clear what the skill of the Penman-Monteith equation is for partly wetted vegetation.

Eq. (6) is not able to describe the evapotranspiration of sparse crops. In that case the evaporation from the soil can be dominant (e.g. De Bruin, 1987).

It appears that the surface resistance, r_s, of a dry crop completely covering the ground has a non-zero minimum value in the case the water supply in the root zone is optimal. For arable crops this minimum is about $r_s = 30\ s.m^{-1}$ (e.g. Russell, 1980). That of a forest is about $150\ s.m^{-1}$.

The canopy resistance is a complex function of incoming solar radiation, water vapour deficit and soil moisture. The relationship between r_s and these environmental quantities varies from species to species and depends also on soil type. It is not possible to measure r_s directly. Usually, it is determined experimentally by using the Penman-Monteith equation, where E is measured independently. The problem is that the aerodynamic resistance r_a has to be known in this approach. Due to the crude description of the vegetation layer this quantity is poorly defined, since it is related to the surface temperature T_s. Because in a real vegetation pronounced temperature gradients occur, it is very difficult to determine T_s precisely. In many studies r_a is determined very simple. This implies that several r_s values published in literature are biased due to errors made in r_a. For more detailed information about the Penman-Monteith equation the reader is referred to recent review papers by McNaughton and Jarvis (1983) and Jarvis and McNaughton (1986).

2.4 Summary of section 2 and recommendations

The above can be summarized as follows:

a. the KNMI E_o-figures are meant to be used for the *crop factor method* to determine the potential crop evapotranspiration;

b. due to several factors a tremendous confusion exists concerning the (physical) meaning of E_o as well as the way it has to be (or is) calculated;

c. the evaporation of "real" open water differs significantly from E_o. The method by Keijman (1974), using the wind function proposed by Sweers (1976), is recommended for the determination of "real" open water evaporation in the Netherlands;

d. the crop factor method is very crude (Feddes, 1987). For cropped surfaces the Penman-Monteith equation (Monteith, 1965) is recommended for more accurate calculations;

e. the methods recommended above under points d. and c. require the same (or similar) meteorological input data as needed by the Penman formula. It is recommended that in the near future these meteorological data are made available at low cost in standard computer compatible form;

f. there will be still a need for practical calculations in the next ten years for an evaporation figure, similar to E_o *which is meant only to be used in the crop factor method*. This figure must meet several requirements:

   i it must have a behaviour similar to E_o;

   ii its calculation has to be simple, the number of meteorological input variables, has to be as small as possible;

   iii it must contain only a few empirical constants;

   iv it has to be obvious that it is an empirical quantity that cannot be "improved" on physical grounds.

We found that the fomula proposed by Makkink fulfills these requirements. This will be discussed in the next section.

3 THE FORMULAS BY PRIESTLEY-TAYLOR AND MAKKINK

3.1 General

Micrometeorological observations over well-watered temperate arable crops reveal that their evapotranspiration depends strongly on net radiation. Furthermore, it appears that the second term in the right-hand of Eq. (6) is typically one-fourth the size of the first term. This leads to the formula proposed by Priestley and Taylor (1972):

$$\lambda E = \alpha \frac{s}{s+\gamma} (Q^{*}-G) \tag{7}$$

where α is a coefficient of value of about 1,2-1,3.

Eq. (7) describes the evaporation loss of both "saturated" land and water surface surprisingly well. For a review of the literature see e.g. Brutsaert (1982).
Usually, G is small for grassland (e.g. De Bruin and Holtslag, 1982). Moreover, it appears that in the Netherlands over grass net radiation is about 0.5 times the incoming short wave radiation in summertime. In this way, one arrives at the formula found by Makkink as early as 1957 for well-watered grassland*):

$$\lambda E = C \frac{s}{s+\gamma} K\downarrow \tag{8}$$

where C is a constant.

At first sight Eqs. (7) and (8) are purely empirical. However, recent research has shown that (on a regional scale) a Priestley-Taylor like formula can be derived by taking into account that evapo-(transpi)ration E and the saturation deficit D are dependent variables. This is due to the fact that if at the surface water vapour and heat are brought into the lower atmosphere the saturation deficit,

*) Note that in his original paper Makkink found $\lambda E = C_1 \frac{s}{s+\gamma} K\downarrow + C_2$. This feature will be discussed later.

D, is changed. In turn this affects E. The relationship between E and D is not a simple one. A coupled model for the atmospheric boundary layer and the surface layer is required. A discussion on this issue is outside the scope of this paper. For this the reader is referred to De Bruin and Holtslag (1987).
Here we adopt the result of recent work, i.e. that on a regional scale the evapotranspiration of a well-watered terrain, covered with a short vegetation, is primarily determined by the net radiation and also by the temperature (through the term $s/(s+\gamma)$). Factors as saturation deficit and wind speed appear to be less important.
This implies that the Priestley-Taylor formula and the related equation by Makkink describe fairly well the evapotranspiration of e.g. grass on a regional scale if there is no short of water. Hence both can serve as an alternative for the KNMI E_o-figure.
Taking into account the requirements for the new evaporation figure, it was decided to choose Makkink's formula in a simplified form. Since it needs as input only *global radiation* and *temperature*, which are observed directly in the Netherlands on a sufficient number of routine stations. The drawback of the Priestley-Taylor formula is that net radiation, Q^*, is needed. This quantity is not measured directly on climatological stations. Moreover, the existing semi-empirical expressions to determine Q^* need a lot of input data and contain several empirical constants. The values of these constants are still uncertain. This was the main reason to choose Makkink's formula.

3.2 Reference crop evapotranspiration according Makkink

Considering the evidence presented in the previous sections it was finally decided to introduce the *reference crop evapotranspiration* according to Makkink *defined* by:

$$E_r = C \frac{s}{s+\gamma} \frac{K\downarrow}{\lambda} \qquad (kg\ m^{-2}\ s^{-1}) \qquad (9)$$

where constant C = 0.65.

This quantity is introduced to replace the KNMI E_o-figure and is meant to be used solely in the crop factor method.
New crop factors belonging to this new evaporation figure E_r are presented by Feddes (1987).

It should be stressed that E_r is not a physical quantity, but approximately, E_r describes the evapotranspiration of well-watered short grass on a regional scale in summertime.

3.2.1 The choise of constant C = 0.65

Originally, Makkink (1957, 1961) proposed a two-constant model: $\lambda E = C_1\ s/(s+\gamma)K{\downarrow} + C_2$. We decided to skip the intercept C_2, since E_r is used only in the growing season. Then E_r is greater than, say, 1 mm/day and a one-constant approach appears to describe Makkink's data also fairly well. Moreover, it is important to note that the choice of C or C_1 and C_2 is arbitrary, since changes in the constant(s) are incorporated directly in the crop factors.
The one-constant approach with C = 0.65 appears to describe reasonably the evapotranspiration of grass (De Bruin, 1981; Keijman, 1982) while it fits fairly well the data presented by Makkink and Van Heemst (1967) for E > 1.5 mm. The crop factors published by Feddes (1987) referring to E_r are based on Eq. (8) with C = 0.65.

3.3 Comparison between E_r (Makkink) and E_o (Penman)

For a comparison between the new evaporation figure, E_r, and the old one, E_o, we analysed data for 1965 through 1985, being the longest period for which the required meteorological input parameters are available. The length of this period is determined primarily by the fact that in 1965 direct routine observations of global radiation started in the Netherlands at more than one KNMI-station*), notably De Bilt, Eelde, Den Helder/De Kooy, Vlissingen and Beek.

*) Note that Wageningen and De Bilt have longer records of K↓.

For these 5 stations E_r and E_o were evaluated per decade*) E_r with Eq. (9) and E_o according the KNMI procedure described by De Bruin (1979) and Buishand and Velds (1980). The decade sums were rounded up to whole mm.
Firstly, lineair regression is applied to all decade totals for each station separately and for the growing season, i.e. April through September. This period consists of 18 decades, so for each calculation 378 pairs of decade totals are analysed.
The results are listed in Table I.

In this Table the mean values of the decade totals of E_r and E_o are listed, their ratios, the regression constants from $E_r = A''E_o$ and $E_r = A'E_o + B$ respectively, the *correlation* coefficient and the standard errors, here defined as $\varepsilon = [\overline{(E_r - A''E_o)^2}]^{\frac{1}{2}}$, where the bar indicates a mean value. It appears that for none of the stations the intercept B differs significantly from zero, so that the regression model $E_r = A''E_o$ is a suitable description of the data set. From the evidence presented in Table I it can be concluded that the correlation between E_r and E_o is high for decade sums and for the entire growing season.
Den Helder/De Kooy and Vlissingen are located nearby or at the sea-shore. Since we are dealing with agricultural problems and the local climate at the coast differs considerably from that inland, it is decided to exclude the data from these two coastal-stations from a further analysis.
We applied the same regression technique described above to the spatial mean decade totals of E_r and E_o using the data for the three inland stations. The results are also listed in Table I.
It can be concluded that regression constant A" shows a spatial variability of less than 2% compared to its mean value of 0.791. For practical calculations this can be ignored, keeping in mind that the crop factor approach in which E_r is meant to be used is a very crude one.

*) A decade is defined here as follows: each month is devided into three decades, being the 1st-10th, the 11th-20th and the 21st-end. So the third decade consists of 8, 9, 10 or 11 days depending on the months.

Table I Growing season

Station	E_r (mm/dec)	E_o (mm/dec)	E_r/E_o -	E_r/E_o -	A" -	A' -	B (mm/dec)	R -	ε (mm/dec)
De Bilt	24.25	30.82	0.787	1.271	0.786	0.776	0.32	0.962	2.09
Eelde	24.42	30.20	0.809	1.237	0.806	0.776	1.0	0.961	2.16
Beek	24.61	31.54	0.780	1.282	0.781	0.791	-0.35	0.955	2.31
Den Helder/De Kooij	26.34	32.73	0.805	1.243	0.805	0.802	0.09	0.951	2.43
Vlissingen	26.0	33.53	0.776	1.290	0.776	0.781	-0.19	0.948	2.42
3 landstations	24.43	30.85	0.792	1.263	0.791	0.782	0.29	0.965	1.99

Mean decade values (1965-1985) of E according to Makkink (E_r) and Penman (E_o); for the five main KNMI-stations as well as the mean of the three inland stations; R is the correlation coefficient and ε is the residual standard deviation. (See further the text.)

Further analyses reveal that constant A" shows a seasonal variation. This was found by applying the linear regression technique to the mean decade totals of the three inland stations for each month separately. Now each calculation concerns 3 x 21= 63 pairs of E_r and E_o. In Table II the results are shown. Herein the mean values and their ratios are listed as well as the correlation coefficients. It is seen that the ratio E_o/E_r, which is needed for the determination of the new crop factors (Feddes, 1987) is month-dependent. It decreases significantly in August and September.

For the new crop factors the ratio E_o/E_r is needed for each decade in April through September. It appears that the direct determined values of E_o/E_r per decade show to much scatter. Apparently, a period of 21 years is too short to obtain stable values. For that reason it was decided to smooth the monthly values "by hand" to obtain decade values. The results are listed in Table II. These smoothed values of E_o/E_r have been used by Feddes (1987) to evaluate the new crop factors related to the evaporation figure according to Makkink. It is realized by the author that the determination of the smoothed E_o/E_r values per decade is rather subjective.

3.4 Dry conditions

As early as 1963 Bouchet pointed out that in the formula by Penman (or related equations) the evapo(transpi)ration is expressed in *dependent* variables. In particular E is interrelated with the water deficit D. This can be illustrated by considering a soil that is drying out. Then the evapotranspiration is *decreasing*, while the air near the ground will become warmer and drier, by which D *increases*. This leads to the conclusion that E and D are (negatively) correlated.

Next we consider the "potential" evapotranspiration E_p. This quantity refers to the imaginary situation that the water supply is plentifull in the root zone. Let D_d and D_p be the water vapour deficit under the actual dry and imaginarily "potential" condition respectively. Obviously, $D_d > D_p$. This implies that *if* E_p *is evaluated with the Penman (or related) formula* using D_d instead of D_p, E_p is *over-estimated*, since under real "potential" conditions D reduces to D_p.

Table II

3 landstations	E_r (mm $decade^{-1}$)	E_o (mm $decade^{-1}$)	E_r/E_o -	Decade	Smoothed value	R
April	17.78	23.16	1.30	I	1.30	0.945
				II	1.30	
				III	1.30	
May	26.44	34.29	1.30	I	1.30	0.971
				II	1.30	
				III	1.30	
June	29.43	38.93	1.32	I	1.31	0.97
				II	1.31	
				III	1.31	
July	29.81	37.79	1.27	I	1.29	0.971
				II	1.27	
				III	1.24	
August	26.47	31.49	1.19	I	1.21	0.948
				II	1.19	
				III	1.18	
September	16.62	19.45	1.17	I	1.17	0.867
				II	1.17	
				III	1.17	

Comparison between "Penman" (E_o) and "Makkink" (E_r) for April through September; spatial (3 landstations) and temporal (1965-1981) mean values per decade. Also the smoothed ratio of Penman/Makkink per decade are given.
R = correlation coefficient.

Note that these considerations apply to extensive areas so that advection is excluded.
The equation of Priestley-Taylor and Makkink do not contain D and, therefore they are not sensitive to the effect described above.
To illustrate this we consider data collected in the very dry summer of 1976 at Cabauw over grass (De Bruin, 1981). A number of days are selected with a mean relative humidity (RH) of 50% or less. For these days E_p we calculated with the Penman-Monteith equation (using r_s = 65 sm^{-1} and the expression for r_a proposed by Thom and Oliver, 1977) and Makkink's formula respectively. The results are listed in Table III with the observed air temperature, wind speed, relative humidity and global and net radiation. It is seen that E_p according to Penman-Montheith is significantly larger than according to Makkink, whereas the first (expressed in energy units) is larger than the observed net radiation. This indicates clearly that the Penman-Monteith equation tends to overestimate E_p, because it is to be expected that the evapotranspiration is less than net radiation.
For the August days also Makkink's formula gives larger values than the observed net radiation. Further investigations reveal that also net radiation depends on the "dryness of the soil". This is probably due to a change in albedo and higher surface temperatures. In the last column of Table III we listed an estimate of net radiation for "potential" conditions, evaluated with an empirical formula developed by Slob (personal communication):

$$Q^* = (1-r)\, K{\downarrow} - 110 \frac{K^{\downarrow}}{K_o^{\downarrow}} \qquad (10)$$

where r is the albedo of the surface taken equal to 0.23 and $K_o^{\downarrow}$ the global radiation at the top of the atmosphere.
Note that Eq. (10) refers to mean daily values and is tested for Dutch conditions only. It is seen that the net radiation for potential conditons is greater than the observed values. Moreover, now the evaporation figure according to Makkink is smaller than the calculated net radiation, whereas that according to Penman-Monteith is still larger.
We conclude that under very dry conditions the Makkink formula shows a more realistic behaviour than the Penman-Monteith equation.

Table III

Date	T	RH	u_2	$K\downarrow$	Q^*	λE_r	λE_p	$\hat{Q}^*_p$
(1976)	(°C)	-	(ms^{-1})	(Wm^{-2})	(Wm^{-2})	(Wm^{-2})	(Wm^{-2})	(Wm^{-2})
03-07	24.1	0.50	1.9	311	151	148	165	167
04-07	23.6	0.49	2.3	307	147	145	169	165
06-07	22.6	0.39	3.0	319	153	148	199	171
22-08	17.3	0.47	3.7	262	94	112	142	124
23-08	17.6	0.46	3.1	256	93	110	139	120
24-08	19.9	0.42	2.4	246	91	110	145	115
25-08	20.4	0.50	1.7	230	81	103	115	107

Comparison between λE_r, according to Makkink and λE_p according to Penman-Monteith, using $r_s = 65\ sm^{-1}$ and r_a proposed by Thom and Oliver (1977) for a number of (very) dry days in 1976. Also the observed air temperature (T), relative humidity (RH), global radiation ($K\downarrow$) and net radiation (Q^*) are listed. Data from Cabauw.

Q^*_p is an estimate of net radiation for "potential" conditions.

3.5 Wintertime

The arguments leading to Makkink's formula apply only to the "summer season" April through September, since then radiation is the main driving-force for evaporation. In the winter season this is no longer true and the physical ground for Makkink's formula then is lacking. However, since its use is confined to the crop factor method, in principle, this does not matter.

Most reliable data-sets of evapo(transpi)ration concern summertime conditions. This is primarily due to instrumental problems; direct evaporation measurements under wintertime conditions are extra-ordinarily difficult to carry out. For that reason not much is known about winter evaporation. Water balance studies have revealed that de evaporation loss of catchments covered with aerodynamically rough vegetation, such as (pine)trees and heather, exceeds significantly E_o. This applies even to grass (Thom and Oliver, 1977). The main reason for this feature is the fact that in wintertime the aerodynamic term in the Penman-Monteith equation often is dominant, whereas the aero-dynamic resistance, r_a, strongly depends on the surface roughness. Moreover, Penman's r_a refers to a very smooth surface (Thom and Oliver, 1977; Keijman, 1981). Stricker (1981) reports good results in wintertime using the Thom-Oliver version of the Penman-Monteith equation for the *Hupselse Beek catchment* in the Netherlands. He uses a time-step of one day.
From the above it must be concluded that neither Makkink's formula nor the Penman equation is applicable in wintertime.
Several catchment areas in the Netherlands are pastures. In wintertime precipitation is on the average one order greater than evaporation, so for water balance calculations over a month or so, E needs not to be known very accurately. Often, E_o or 0.8 E_o is taken as first estimate. Note that from the discussion above it appears that this leads to an underestimation of E.

The question arises whether the new figure E_r can be used for these rough water balance calculations. For that reason we compared E_r and

E_o also for the "winter months" October-March. The results are shown in Table IV, which is simular to Table II. It is seen that:

1) $E_r > E_o$ in October-February;
2) E_r and E_o are virtually non-correlated in November-January, illustrating the fact that radiation is no longer the drying-force in winter;
3) March behaves as a "summer month", so that Makkink's formula can be used from about 1st of March.

Since E_o appears to underestimate E of grassland in wintertime and E_r then is some mm/decade larger than E_o, it is concluded that *on the average* E_r can be used in wintertime for *rough* water balance calculations for catchments covered with pastures. For short periods (less than 1 month or so) this is certainly not true. We recall that Makkink's formula has no physical base in wintertime.

Table IV

3 landstations	Makkink (mm decade^{-1})	Penman (mm decade^{-1})	Penman/Makkink -	R
October	9.19	9.01	0.98	0.709
November	3.77	3.06	0.81	0.471
December	2.11	0.71	0.34	-0.260
January	2.57	1.28	0.50	0.166
February	4.96	4.46	0.90	0.615
March	10.31	13.90	1.35	0.881

As Table II, but now for October-March

4 SUMMARY

In the first part of this paper the background and applicability of Penman's equation is briefly discussed. It is pointed out that this equation is used in practice primarily empirically. Commonly, the so-called open water evaporation, E_o, which is evaluated with the Penman formula, is multiplied with a suitable crop factor to obtain an estimate of the "potential" evapotranspiration. This *crop factor method* appears to be rather crude. For more accurate calculations the Penman-Monteith equation, using a canopy or surface resistance, is recommended. It is shown that E_o - in spite of its name - cannot be used for "real" open water. For that case the model of Keijman is recommended.
An important drawback of the use of E_o is the fact that there excists a lot of confusion about the way it is or has to be calculated. This confusion is the main reason that the KNMI has decided to stop the routine publication of E_o.
Since there is still a need for an evaporation figure, similar to E_o, which can be used in the crop factor approach, one has searched for an alternative for E_o.

The formula proposed by Makkink appears to be very suitable for this purpose.
A new evaporation figure based on Makkink's expression is introduced. It is called the *reference crop evapotranspiration* and is denoted by E_r and defined by Eq. (9).

The second part in this paper is devoted to E_r. Its background is discussed. In addition a comparison is presented between E_r and the "old" figure E_o. This comparison reveals that in the growing season (April through September) the two quantities are correlated very well. The behaviour of E_r in wintertime and under very dry conditions is discussed also.
Finally, it is noted that Feddes (1987) derived crop factors, which allow the use on the crop factor method, using E_r as reference.

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APPENDIX

Symbol	Definition	Units
c_p	specific heat of air at constant pressure	$J\ kg^{-1}\ K^{-1}$
c_w	specific heat of water	$J\ kg^{-1}\ K^{-1}$
e_a	water vapour pressure at screen height	mbar
$e_s(T_a)$	saturation water vapour pressure at T_a	mbar
h	water depth	m
r	albedo	-
r_a	aerodynamic resistance	$s\ m^{-1}$
r_s	canopy or surface resistance	$s\ m^{-1}$
s	slope of saturation water vapour temperature curve at T_a	$mbar\ K^{-1}$
t	time	s
A	exchange coefficient (Eq. 5')	$W\ m^{-2}\ k^{-1}$
A', A"	regression constants	-
C, C_1	constants	
C_2	constant	$W\ m^{-2}$
D	water vapour saturation deficit	mbar
E	evapo(transpi)ration	$kg\ m^{-2}\ s^{-1}$ (or mm/decade)
E_o	"open water evaporation" according to Penman	idem
E_r	reference crop evapotranspiration	idem
G	soil heat flux density or change per second of heat stored per m^2 in water body	$W\ m^{-2}$
Q*	net radiation	$W\ m^{-2}$
Q^*_n	net radiation if surface temperature is T_n (Eq. 4)	$W\ m^{-2}$
$K^{\downarrow}$	global radiation	$W\ m^{-2}$
$K^{\downarrow}_o$	global radiation at the top of the atmosphere	$W.m^{-2}$
RH	relative humidity	-
T_a	air temperature	K
T_e	effective temperature (Eq. 4)	K

Symbol	Definition	Units
T_n	wet-bulb temperature at screen height	K
T_w	water temperature	K
α	Priestley-Taylor parameter	-
γ	psychrometric constant	mbar K^{-1}
ρ	density of air	kg m^{-3}
ρ_w	density of water	kg m^{-3}
σ	Stefan-Boltzmann constant = 5.6710^{-8}	W m^{-2} K^{-4}
τ	time constant (Eq. 5)	s

Note: In literature Penman's equation sometimes is written as

$$\lambda E = \frac{sQ^* + \gamma\lambda E_a}{s+\gamma} \quad \text{with}$$

$$\lambda E_a = f(u)\ [e_s(T_a)-e_a]$$

where f(u) is a wind function.

Apparently $r_a = \frac{\rho c_p}{\gamma f(u)}$

(In his original paper Penman used mm/day as unit for E and therefore his windfunction contained λ.)

CROP FACTORS IN RELATION TO MAKKINK REFERENCE-CROP EVAPOTRANSPIRATION

R.A. Feddes

1 GENERAL RELATIONSHIPS OF EVAPOTRANSPIRATION

The actual evapotranspiration of a cropped surface, E, can be considered as the sum of evaporation of intercepted water, E_i, evaporation from the soil surface, E_s, and the transpiration of the (dry) crop leaf surface, E_t:

$$E = E_i + E_s + E_t \tag{1}$$

If under the governing meteorological conditions enough water is available for evapotranspiration of the soil and the crop (and if the meteorological conditions are unaffected by the evapotranspiration process itself) one considers evapotranspiration to be maximal. For the condition that both the crop surface and the soil surface are wet, eq. (1) reads as:

$$E_{max} = E_i + E_{sp} + E_{tp} \tag{2}$$

where E_{max} is the maximum possible evapotranspiration of a cropped surface, E_{sp} is potential soil evaporation and E_{tp} the potential transpiration. For large uniform fields advection is negligibly small such that the magnitude of maximum possible crop evapotranspiration depends on the meteorological conditions (such as radiation, air temperature, windspeed and air vapour pressure) and on the type and structure of the crop.

If the crop surface is dry, i.e. $E_i = 0$, but water supply to both roots and soil surface is still optimal, maximum possible crop evapotranspiration reduces to potential crop evapotranspiration, E_p, according to:

$$E_{max} = E_p = E_{sp} + E_{tp} \tag{3}$$

During periods with and without precipitation, the maximum possible evapotranspiration of a cropped surface can be theoretically approximated by the equation (RIJTEMA, 1965; FEDDES, 1971):

$$E_{max} = \underbrace{\frac{s + \gamma}{s + \gamma(1+r_s/r_a)} (E_w - E_i)}_{E_p} + E_i \qquad (mm.d^{-1}) \tag{4}$$

where: s = slope of the saturation water vapour pressure temperature curve at air temperature ($m.bar.K^{-1}$)

γ = psychrometer constant

r_s = crop or surface resistance ($s.m^{-1}$)

r_a = diffusion resistance for water vapour transfer of the air layer between the ground surface and screen height ($s.m^{-1}$)

E_w = wet crop evapotranspiration, i.e. the theoretical evaporation flux of a fictitious water surface with the albedo and aerodynamic resistance of the crop. For a wet crop E_w is synonymus with E_i

E_w can be calculated from a modified Penman equation (MONTEITH, 1965; RIJTEMA, 1965) as:

$$\lambda E_w = \frac{s(Q^*-G) + c_p \rho_a (e_s - e)/r_a}{(s+\gamma)} \qquad (W.m^{-2}) \tag{5}$$

(To convert λE from $W.m^{-2}$ to $mm.d^{-1}$ one has to multiply λE with $86{,}400/\lambda \approx 0.0352$ at 20°C, with 86,400 being the number of seconds in 24 h).

where: λ = latent heat of vaporization of water ($J.kg^{-1}$)

Q^* = net radiation flux density ($W.m^{-2}$)

G = soil heat flux density ($W.m^{-2}$)

c_p = specific heat of air at constant pressure ($J.kg^{-1}.K^{-1}$)

ρ_a = density of the air ($kg.m^{-3}$)

e = water vapour pressure at screen height (mbar)

e_s = saturated vapour pressure at air temperature at screen height (mbar)

In case not enough soil water is available to meet the demand set by the atmosphere to the crop-soil surface, evapotranspiration will be reduced. Then photosynthesis and growth is reduced, hence final crop yield will be reduced.

Remarks at eq. (4):

- E_i can be derived daily from measured interception-precipitation curves (FEDDES, 1971; HOYNINGEN HUENE, 1981);
- under conditions that the crop is partly wet and/or the soil is not completely covered by the crop, values of r_s may change considerably;
- under conditions of a dry crop that covers completely the soil $E_i = 0$ and $E_{sp} = 0$, hence $E_{max} = E_{tp}$, and eq. (4) reduces to:

$$\lambda E_{tp} = \frac{s(Q^*-G) + \rho_a c_p (e_s - e)/r_a}{s + \gamma(1 + r_s/r_a)} \qquad (W.m^{-2}) \qquad (6)$$

2 CROP FACTORS IN RELATION TO PENMAN-OPEN WATER EVAPORATION

In order to simply compute maximum possible evapotranspiration of a certain crop, E_{max}, one often relates this quantity empirically to the evaporation of a hypothetical shallow water surface, often called 'open water evaporation', E_o. This quantity E_o has thus no strict physical meaning because it describes for the prevailing weather conditions the evaporation of a water surface that does not exist! E_{max} is related to E_o simply through a crop factor, g, according to:

$$E_{max} = g\,E_o \qquad (mm.d^{-1}) \tag{7}$$

with E_o being calculated according to PENMAN (1948) as:

$$\lambda E_o = \frac{sQ_w^* + \gamma\lambda E_a}{s + \gamma} \qquad (W.m^{-2}) \tag{8}$$

where: Q_w^* = net radiation flux of a hypothetical water surface $(W.m^{-2})$

and

$$\lambda E_a = f(u)(e_s - e) \qquad (kg.m^{-2}.s^{-1}) \tag{9}$$

with: $f(u)$ = function of the wind speed, being defined as $f(u)$ = $3.7 + 4.0\,\bar{u}_2$ $(W.m^{-2}.mbar^{-1})$

$\bar{u}_2$ = average wind speed at 2 m height $(m.s^{-1})$

The wind function f(u) holds for the evaporation pan of Penman, i.e. for advective conditions. For an actual water surface this function is too large. For more information about the theoretical background of eq. (9), see DE BRUIN (1987). Note again that in case of $E_i = 0$, E_{max} in eq. (7) reduces to E_p (see eq. 3)!

In Table 1 crop factors g are listed as being presently used in agricultural applications.

On applying these g-values one has to keep in mind the way E_o has been computed. In practice one takes E_o often from the monthly reports of the Royal Meteorological Institute (KNMI). Before 1971 computation of monthly E_o-values were based upon inputs of daytime averages of the air temperature, humidity and on values of global radiation that were computed from sunshine duration observations.

Table 1 Decade values for the crop factor g related to open water evaporation E_0 (after WERKGROEP LANDBOUWKUNDIGE ASPECTEN, 1984)

	April			May			June			July			August			September		
	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3
Grass	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8
Cereals	0.5	0.6	0.7	0.8	0.8	0.8	0.9	0.9	0.9	0.8	0.7	0.6	0.5	-	-	-	-	-
Maize	-	-	-	0.4	0.5	0.6	0.7	0.8	0.9	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0
Potatoes	-	-	-	-	0.5	0.7	0.8	0.9	0.9	0.9	0.9	0.9	0.9	0.9	0.9	0.6	-	-
Sugar beets	-	-	-	0.4	0.4	0.4	0.6	0.8	0.8	0.9	0.9	0.9	0.9	1.0	1.0	1.0	0.9	0.9
Leguminous plants	-	0.4	0.5	0.6	0.7	0.8	0.9	0.9	0.9	0.8	0.6	-	-	-	-	-	-	-
Plant-onions	0.4	0.5	0.5	0.6	0.6	0.7	0.8	0.8	0.8	0.8	0.8	0.8	0.8	-	-	-	-	-
Sow-onions	-	0.3	0.4	0.4	0.5	0.5	0.6	0.6	0.7	0.8	0.8	0.8	0.8	0.8	0.8	0.6	-	-
Chicory	-	-	-	-	-	-	0.4	0.4	0.4	0.6	0.8	0.9	0.9	0.9	0.9	0.9	0.9	0.9
Winter carrots	-	-	-	-	-	-	0.4	0.4	0.4	0.6	0.8	0.9	0.9	0.9	0.9	0.9	0.9	0.9
Celery	-	-	-	-	-	0.4	0.5	0.5	0.5	0.6	0.6	0.7	0.8	0.9	0.9	0.9	0.9	-
Leek	-	-	-	-	0.4	0.4	0.4	0.4	0.5	0.5	0.6	0.6	0.7	0.8	0.8	0.8	0.8	0.8
Bulb/tube crops	-	-	-	-	0.4	0.5	0.5	0.7	0.9	0.9	0.9	1.0	1.0	1.0	1.0	1.0	1.0	1.0
Pome/stone-fruit	0.8	0.8	0.8	1.1	1.1	1.1	1.2	1.2	1.2	1.3	1.3	1.3	1.1	1.1	1.0	1.0	1.0	1.0

Since 1971 E_0 is calculated for periods of 10 days. Then, however, 24-hour averages of temperature and humidity were applied. (The computation procedure is such that the meteorological input data are first averaged over the 10-day period and then inserted into the Penman-equation (eq. 8)). The E_0-values calculated according to this new procedure turned out to be approximately 10% lower than these calculated before 1971. Therefore KNMI decided to add a fixed amount, depending on the station and period considered, to the E_0-values calculated according to the new procedure. For full details, the reader is referred to VAN BOHEEMEN, 1977; DE BRUIN en KOHSIEK, 1977; BUISHAND en VELDS, 1980; DE BRUIN en LABLANS, 1980; DE GRAAF, 1983; VAN BOHEEMEN et al., 1986; LABLANS, 1987.

Crop factor data as shown in Table 1 are usually derived from soil water balance experiments, especially from sprinkling experiments where water is applied in quantities such that potential evapotranspiration is reached.

The water balance of the soil accounts for the incoming and outgoing fluxes of a soil compartment. This compartment can for example be the root zone, the profile over a large depth of 150 cm, or even a homogeneous layer as small as 10 cm.

In sprinkling studies one often considers the soil water balance of the root zone only. The change in water storage ΔV_r yields a given infiltration (including irrigation) F, plus net upward flow through the bottom Q_r, minus outflow, i.e. evapotranspiration E:

$$\Delta V_r = F + Q_r - E$$

or

$$E = F + Q_r - \Delta V_r \qquad \text{(mm)} \tag{10}$$

The problem with eq. (10) is that it is very difficult to evaluate Q_r properly. This flow is the resultant of capillary rise and percolation. Often one does not consider capillary rise: what has been percolated through the root zone is simply lost. In the presence of a groundwater table which influences the moisure conditions in the root zone, eq. (10) cannot be applied. Then one should take into account in detail the water transport in the subsoil below the root zone. Hence all the errors in determining F, Q_r and ΔV_r will be reflected in the quantity E. Therefore the crop factors of Table 1 have to be considered as factors that have been determined over average periods of 7 to 14 days, with considerable possible errors.

Another aspect is the degree of variation of crop cover over time. An example of it is presented for potatoes and sugar beets in Fig. 1 for optimum sprinkled fields. The crop cover development will vary with the species and may be different from year to year. Hence the variation of crop factors over time is not fixed, as suggested in Table 1, but may be different from year to year.

An aspect also to be taken into account with sprinkled experimental fields is that during most of the time the soil surface is dry, while the crop is still well supplied with water. Then eq. (2) changes into:

$$E_{max} = E_i + E_s + E_{tp} \tag{11}$$

where E_s is thus the <u>actual</u> soil evaporation. The drier the period/

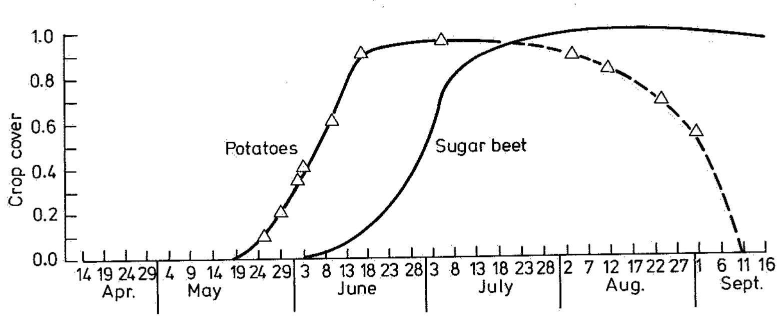

Figure 1 The degree of variation of crop cover for potatoes during the growing season of 1981 and of sugar beets during the growing season 1983 on optimum sprinkled experimental fields at Sinderhoeve, Renkum.

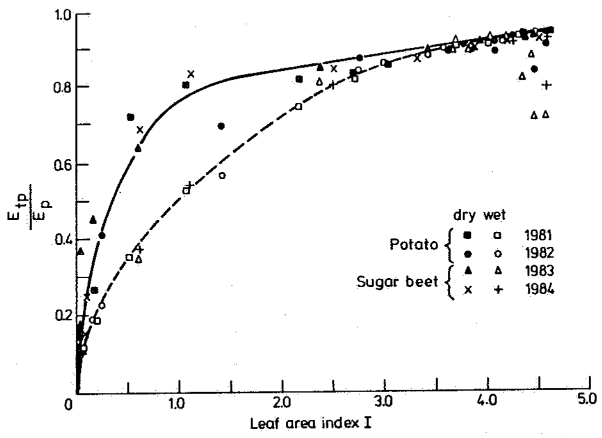

Figure 2 Potential transpiration E_{tp} over potential evapotranspiration E_p (= $E_{max}-E_i$) as a function of leaf area index I when either the soil surface is wetted every day or the soil surface is dry most of the time

season, the smaller the quantity (E_i+E_s). Eq. (11) thus describes the practical situation one often encounters in the field.

On soils with partial soil cover such as row crops in the beginning of their growth stage the condition of the soil, dry or wet will considerably influence the partitioning of E over E_s and E_{tp}. Figure 2 gives an idea about the computed variation of E_{tp}/E_p (with $E_p = E_{max}-E_i$) with the leaf area index I for a potato crop with optimal water supply to the roots for a dry and a wet (applying eqs. 2 and 4) soil. Assuming that E_p is the same for both dry and wet soil conditions, it appears that for $I < 1$ at increasing drying of the soil and thus decreasing E_s, E_{tp} will increase with about a factor 1.5 to 2. For $I > 2$-2.5, E is almost independent of the condition of the soil surface. This result agrees with findings of FEDDES (1971) on red cabbage that the soil must be covered for about 70 to 80% ($I = 2$) before E becomes constant.

The g-values of Table 1 originate mainly from field water balance experiments. Fig. 2 shows that it is rather difficult to estimate evapotranspiration in relation to crop development. Hence for leaf area indexes $I < 2$ the g-factors of Table 1 may only be considered as orders of magnitude.

For grass with a height of 5-15 cm a g-factor of 0.8 will do. This value is based on the WERKCOMMISSIE VOOR VERDAMPINGSONDERZOEK (1984). They report on the basis of 11 years of lysimeter experiments for periods with a low evapotranspiration demand (80% probability of exceedance) $\bar{g} = 0.73$; for periods with a high demand (10% probability of exceedance) $\bar{g} = 0.77$; as overall average they report $\bar{g} = 0.75$. One has to realize that in water balance studies precipitation may often be underestimated because of wind influence on the rain gauge. This error has the tendency of underestimating g. Also errors may arise due to inconsistencies of eq. (7) by which g is dependent on the influence of meteorological parameters (see, for instance, ROMIJN, 1985). Van BOHEEMEN et al. (1986) performed computations on grass of 5-15 cm high using equations such as eq. (4), and found also an overall g-value of 0.8. Based on similar type of computations one will find that for grass of 15-25 cm high $g = 0.85$ and for heights ≥ 25 cm $g = 0.9$.

The g-factors of Table 1 for potatoes and sugar beets have been derived from careful soil water balance measurements with sprinkling experiments at Sinderhoeve during 1981-1984 (see HELLINGS et al., 1982).

The g-factors for maize are now being investigated and will be derived more precisely from soil water balance and micrometeorological (Bowen-ratio) experiments held at the same field during 1985 and 1986 (to be published).

Note that the g-values of Table 1 were derived from fields with different local conditions and agricultural practices. These local effects may thus include size of fields, advection, irrigation and cultivation practices, climatological variations in time, distance and altitude, and soil water availability.

3 REFERENCE-CROP EVAPOTRANSPIRATION ACCORDING TO MAKKINK

Instead of taking the evaporation of a hypothetical water surface as a reference to calculate maximum possible crop evapotranspiration, one can also take the evapotranspiration of a reference crop, i.e. of 'standard' grass 8 to 13 cm high, well supplied with water. Analogous to eq. (7) one can formulate:

$$E_{max} = f \cdot E_r \qquad (mm.d^{-1}) \qquad (12)$$

where f is a new crop factor and E_r is the maximum possible evapotranspiration of grass according to MAKKINK (1957).

For conditions in the Netherlands (KEIJMAN, 1982; De BRUIN, 1987) the Makkink relationship can be expressed as:

$$\lambda E_r = 0.65 \frac{s}{s + \gamma} K\downarrow \qquad (W.m^{-2}) \qquad (13)$$

where $K\downarrow$ is global radiation ($W.m^{-2}$). Eq. (13) has the advantage that easily measurable quantities as global radiation and air temperature (to determine s) will sufficiently accurate describe evapotranspiration. [To describe reference-crop evapotranspiration for different climatological conditions in the world DOORENBOS and PRUITT (1977) have used a modified Penman equation. VOS et al. (1987) have developed the computer program CRIWAR to predict crop evapotranspiration and crop irrigation water requirements based on this approach.]

The new crop factors f can be derived from the old factors g by equating the right hand sides of eq. (12) and eq. (7):

$$f \cdot E_r = g \cdot E_o$$

or

$$f = \frac{E_o}{E_r} \cdot g \tag{14}$$

The multiplication factor E_o/E_r has been derived by DE BRUIN (1987) from 10-day period averages during the growing season from the meteorological stations De Bilt, Eelde and Beek for the period 1965-1985. By multiplying the ratios E_o/E_r with the g-factors of Table 1, the new f-factors can be obtained. However, irrigularities in f-values occurred: sudden jumps/falls which could not be physically based but originated from the computation procedure. Therefore the ratios of E_o/E_r were smoothed (De BRUIN, 1987). These smoothed values are listed in Table 2.

Table 2 The ratio E_o/E_r over the various 10-day periods of the growing season as averaged over the period 1965-1985 for De Bilt, Eelde and Beek (after De BRUIN, 1987)

April			May			June			July		
1	2	3	1	2	3	1	2	3	1	2	3
1.30	1.30	1.30	1.30	1.30	1.30	1.31	1.31	1.31	1.29	1.27	1.24

August			September		
1	2	3	1	2	3
1.21	1.19	1.18	1.17	1.17	1.17

Table 3 Crop factors f as related to Makkink reference-crop evapotranspiration

	April			May			June			July			August			September		
	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3
Grass	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	0.9	0.9	0.9	0.9
Cereals	0.7	0.8	0.9	1.0	1.0	1.0	1.2	1.2	1.2	1.0	0.9	0.8	0.6	-	-	-	-	-
Maize	-	-	-	0.5	0.7	0.8	0.9	1.0	1.2	1.3	1.3	1.2	1.2	1.2	1.2	1.2	1.2	1.2
Potatoes	-	-	-	-	0.7	0.9	1.0	1.2	1.2	1.2	1.1	1.1	1.1	1.1	1.1	0.7	-	-
Sugar beets	-	-	-	0.5	0.5	0.5	0.8	1.0	1.0	1.2	1.1	1.1	1.1	1.2	1.2	1.2	1.1	1.1
Leguminous plants	-	0.5	0.7	0.8	0.9	1.0	1.2	1.2	1.2	1.0	0.8	-	-	-	-	-	-	-
Plant-onions	0.5	0.7	0.7	0.8	0.8	0.9	1.0	1.0	1.0	1.0	1.0	1.0	1.0	-	-	-	-	-
Sow-onions	-	0.4	0.5	0.5	0.7	0.7	0.8	0.8	0.9	1.0	1.0	1.0	1.0	1.0	0.9	0.7	-	-
Chicory	-	-	-	-	-	-	0.5	0.5	0.5	0.8	1.0	1.1	1.1	1.1	1.1	1.1	1.1	1.1
Winter carrots	-	-	-	-	-	-	0.5	0.5	0.5	0.8	1.0	1.1	1.1	1.1	1.1	1.1	1.1	1.1
Celery	-	-	-	-	-	0.5	0.7	0.7	0.7	0.8	0.9	1.0	1.1	1.1	1.1	1.1	1.1	-
Leek	-	-	-	-	0.5	0.5	0.5	0.5	0.7	0.7	0.8	0.8	0.8	1.0	0.9	0.9	0.9	0.9
Bulb/tube crops	-	-	-	-	0.5	0.7	0.7	0.9	1.2	1.2	1.2	1.2	1.2	1.2	1.2	1.2	1.2	1.2
Pome/stone-fruit	1.0	1.0	1.0	1.4	1.4	1.4	1.6	1.6	1.6	1.7	1.7	1.7	1.3	1.3	1.2	1.2	1.2	1.2

One has to realize that the values listed in Table 2 are averages taken over a population of 'average', 'dry' and 'wet' years, that will certainly not be homogeneously distributed. A statistical analysis would be necessary to make more precise statements about it.

Multiplication of the g-values of Table 1 with the smoothed E_0/E_r ratios of Table 2 result in the final f-values being presented in Table 3.

The f-values for grass in Table 3 apply to a grass height of 5-15 cm. For heights 15-25 cm: f = 1.1 for the months April-July and f = 1.0 for August-September. For heights >25 cm: f = 1.2 for April-June and f = 1.1 for July-September.

All the remarks that were made concerning the g-factors of Table 1 are of course also valid for the f-factors of Table 3. So one has always to be careful in applying crop factor data. They should not be considered as being absolutely true. Moreover, they may be liable to change in the future when more experimental data become available.

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CLIMATOLOGICAL DATA ON EVAPORATION IN THE NETHERLANDS; PAST, PRESENT AND FUTURE

W.N. Lablans

ABSTRACT

Data on evaporation have been published by the Royal Netherlands Meteorological Institute for a number of stations since 1911.
Over the period 1911 - March 1987 the evaporation-data were calculated according to Penman's method.
As various modifications of the original method have been introduced, both inside and outside KNMI, confusion has arisen about the precise significance of the various data-sets published in the KNMI climatological bulletins and in monographs. Therefore a Working Group, with representatives of KNMI and the TNO Committee on Hydrological Research as members, advised that KNMI should stop using the Penman method and in the future should adopt a method proposed by G.F. Makkink.
In this paper the various time series of evaporation published by KNMI are discussed and the future practice of the calculation and the dissemination of evaporation-data by KNMI is described.

1 INTRODUCTION; SOME ASPECTS OF PENMAN'S METHOD

In 1948 H.L. Penman published his well known algorithm for the estimation of evaporation and evapotranspiration. Shortly thereafter this method was adopted by C. Kramer for use in the Netherlands.

Unfortunately it has been appeared that Penman's method can easily lead to confusing results.
One of the reasons is that since 1948 Penman and others have published modified and extended versions of the method, therefore to indicate data as "calculated according to Penman" can be ambiguous.
To identify the sources of confusion we have to look into the Penman formula (version 1948) which reads in our notation:

$$\lambda E = \frac{s(Q^* - G) + \gamma\lambda E_a}{s + \gamma} \qquad (1)$$

with

$$\lambda E_a = f(u)\ (e_s(T_a) - e_a) \qquad (2)$$

and

$$f(u) = \alpha u_2 + \beta \qquad (3)$$

The parameters are given in Table 2.
It is essential for the Penman method that - for applications where G can be neglected - E can be calculated from data obtained by standard climatological observations.
We see however in Table 2 that none of the parameters in the Penman formula are observed climatological data. Therefore Table 2 also shows from which climatological data the values for the parameters are derived before E can be calculated.
In the following sections there is a discussion on how, in the course of time, data sets on open-water evaporation have been calculated on the basis of Penman's formula, at KNMI.
For background information on the way Penman derived the formula from physical principles the reader is referred to the relevant literature, e.g. Penman (1948, 1956), De Bruin (1979), Buishands and Velds (1980).

From the above it follows that the algorithm for the calculation of evaporation according to Penman is only complete when in addition to the basic equation (1), the ways in which the values for the various parameters are derived from climatological records are also specified.

Differences in such procedures can easily lead to slightly different numerical results of the calculations and they form one of the sources of the confusion.
Moreover it must be mentioned that in eq. (1) evaporation is expressed as a momentary water vapour flux density, while the calculation practice always pertains to periods of 24 hours or a multiple. The averaging procedures used to adapt the observational data to the required input for the calculation also have some influence on the numerical results, as Penman's formula is not linear in all parameters.

2 THE CALCULATIONS EXECUTED BY KRAMER

In his famous paper of 1948 Penman put forward his algorithm for evaporation (E), and discussed its applications to evaporation from open water (E_o), bare soil (E_B) and turf (E_T).
Kramer chose the algorithm for E_o as the basis for an investigation of differences in the mean evaporation for various parts of the Netherlands.
He discussed in great detail all the procedures he used to obtain values for the parameters in Penman's formula from the climatological records.
Here we will reproduce some brief examples to show in what way such technical details can give rise to differences in the final results.

One of the problems Kramer was faced with was that, unlike Penman, he did not have at his disposal observations of wind speeds at 2 m height, so that in applying Penman's windfunction (3) *) he had to estimate u_2 from observations made at other heights.

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*) To indicate that eq. (3) is valid only for wind speeds at 2 m height, according to mathematical convention (3) should strictly read $f(u_2) = \alpha u_2 + \beta$, but the notation $f(u) = \alpha u_2 + \beta$ as used by Penman is currently in use.

To do this Kramer assumed the wind profile to be logarithmic, using an average surface roughness for all stations.
Another interesting detail is the way Kramer derived the daily input data from the climatological records.
Different from Penman, who used averages over a 24 hours period, Kramer used the averages of three daytime readings. He was well aware of the bias this would bring about compared to Penman's calculations. He reasoned however, that daytime values of temperature would suit his purpose better, as the inclusion of nighttime conditions would have a smoothing effect on the regional differences he was investigating. In other respects Kramer closely followed Penman's algorithm.
The bias thus introduced is a reason that the notation E_o can be ambiguous, but it should also be noted that Penman was not rigorous in this respect. This follows for instance from table 1 which is taken from Penman (1956).

Table 1 E_o values (inches) for Lake Hefner as published by Penman (1956)

Month	Observed	Calculated	
		Uncorrected	Corrected
Aug. 1950	6.8	7.4	7.8
Nov. 1950	0.0	2.4	5.7
Feb. 1951	0.4	1.9	1.0
May 1951	4.4	6.1	4.0
Aug. - July *)	54.9	57.5	50.0

The uncorrected values for E_o are calculated using a modified version of Penman's formula of 1948.
The corrected values for E_o make allowance for the changes in heat storage.

*) Presumably August 1950 - July 1951.

From Penman (1956) it follows that the notation E_o has been used for data calculated with different wind functions, and even for results of a version of the Penman formula where the heat storage term is not neglected. So the seeds of confusion about the precise meaning of the notation E_o were sown in early times.
Kramer's calculations are of paramount importance for all evaporation data subsequently calculated at KNMI, as basic components such as the wind function and the reflection coefficient of the surface have never been changed. Also, when changes in the calculation procedures were introduced which would cause a systematic difference in the numerical results, corrections have been introduced to keep the data statistically consistent with Kramer's time series.
Kramer has calculated time series for 12 stations over the period 1933-1953 which are published in his monograph in the form of monthly values. These results were obtained by applying the Penman formula to the monthly averages of the daily values for the input parameters.

3 EVAPORATION DATA PUBLISHED IN THE CLIMATOLOGICAL BULLETINS OF KNMI

By January 1st 1956 it was decided to publish data on evaporation in the climatological bulletins of KNMI, for five stations, Den Helder, Eelde, De Bilt, Vlissingen and Beek (figure 1).
The method chosen for calculating the data was the method as developed by Kramer (1957).
This decision is open to criticism, as the bias introduced by Kramer, was retained, relative to the original algorithm of Penman (1948).
A good point however, is that in this way the data in the climatological bulletins were made consistent with the time series calculated by Kramer.
In 1961 it was decided to extend the number of stations to 15.
Since then E_o-data have always been given in the climatological bulletins for 15 stations, but due to changes in the observation network the total number of stations for which during some period of time E_o-data have been published amounts up to 21 stations as can be seen in figure 1.

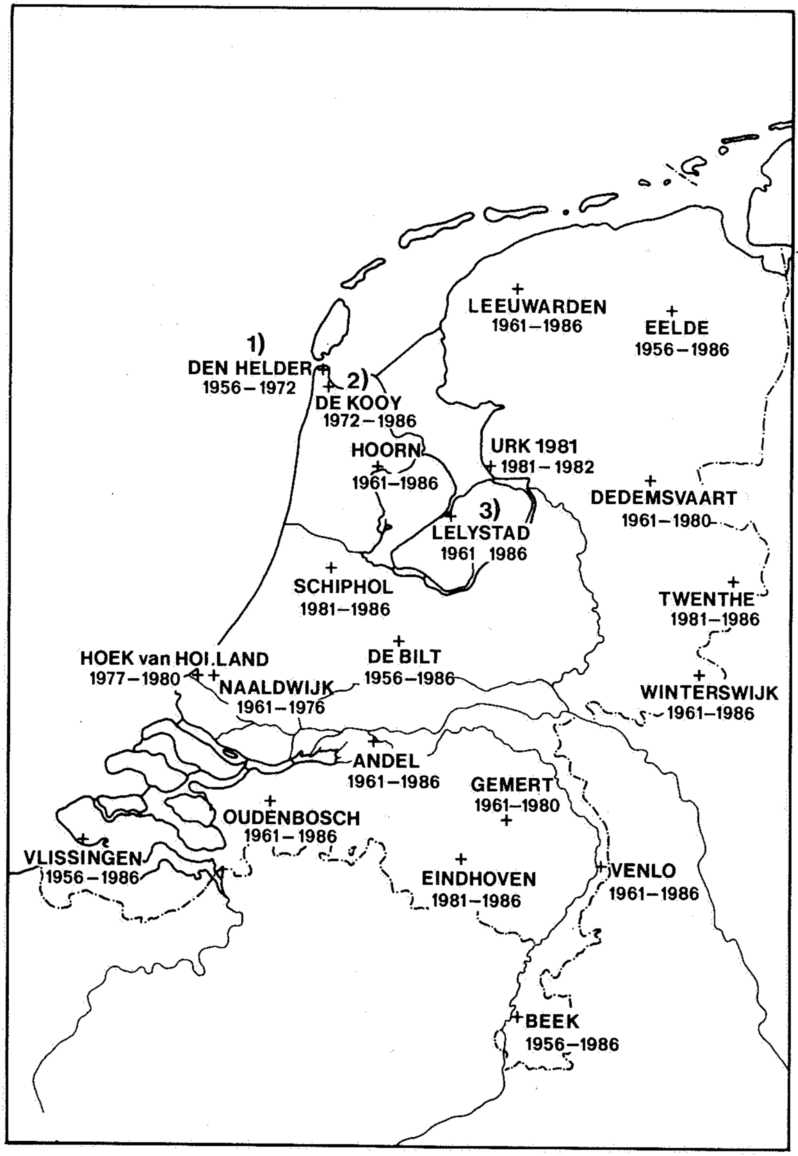

Figure 1 Stations and periods for which E_o-data have been published in the climatological bulletins of KNMI
1) Till July 1972; 2) From August 1972; 3) Not in 1981 and 1982.

In 1971 an important change in the calculation method was made. By that time the climatological stations were equipped with recording instruments, so that 24 hour averages for the input data became available. It was decided to execute the calculations from then onwards with these data, according to Penman (1948).
However, this introduced a decrease in E_o-values by about 10%. In order to retain statistical homogeneity in the time series, correction terms were calculated.
From a statistical analysis it appeared that the required corrections were indeed of the order of 10%, but that they differed from station to station and also depended on the season (De Bruin, 1979).
Unfortunately, since 1971 the network of stations has not remained unaltered. For new localities introduced in the climatological bulletins E_o-values had to be generated by interpolation from data calculated for stations where both the required observations and correction factors were available. Ultimo 1986 the ratio of the calculated and estimated E_o-data in the climatological bulletins had decreased to 10 against 5 stations. As this ratio only could deteriorate in future this is one of the reasons that an alternative for the existing practice was required.
The climatological bulletins over the period 1956-1971 contained only monthly values for E_o, calculated from input-data averaged over the month. From 1971 so-called decade values were provided, which means that for each month E_o-data are given for two ten-day periods and for the rest of the month separately.
Later on it appeared that there was a need for daily estimates of evaporation. In 1981 it was therefore decided to calculate daily values of a first estimate of E_o for five stations. This was done with a faster calculation procedure, which involved the use of measured solar radiation instead of sunshine duration. (De Bruin and Lablans, 1980).
The daily values have been disseminated by radio in the growing seasons of the years 1981-1986 under the name "reference-evaporation".
To avoid confusion with the corresponding E_o-data, the daily values have not been archived, as their summation differs somewhat from the E_o-values and - being a first estimate - they lost their significance as soon as the E_o-data became available.

The data published in the climatological bulletins and how they were derived will be discussed in more detail in the final report of the Working Group (KNMI/TNO, 1988).

4 DATA ON "OPEN WATER EVAPORATION", E_o PUBLISHED IN MONOGRAPHS

Various time series of E_o, calculated at KNMI, now exist:

- Kramers time series for 12 stations over the period 1933-1953;
- time series of various length and for various stations published in the monthly climatological bulletins;
- time series published by De Bruin (1979) over the period 1911-1975;
- time series published by Buishand and Velds (1980) over the period 1911-1979.

The data in these publication overlap considerably.
It should however be noted that De Bruin and Buishand and Velds have put much effort into screening the available material with respect to the quality of the data and, in particular their statistical homogeneity. When E_o-data are used which are not included in these monographs a recommendation is made to check the consistency of the data with these publications.
In figure 2 a survey is given of the locations for which time series have been published in the monographs. It should be noted that some of the time series of Buishand and Velds have a regional character as they are composed of data from 2 to 4 climatological stations.

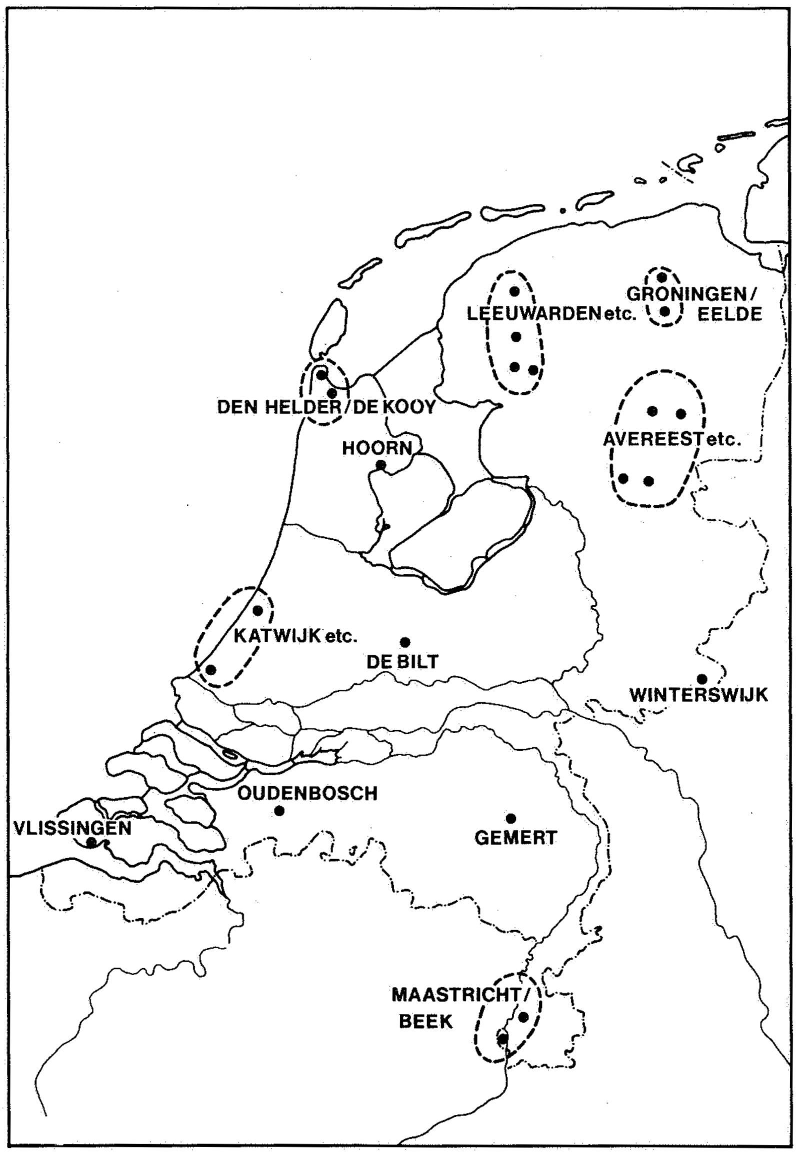

Figure 2 Geographical survey of the time series published in monographs over the period 1911 - 1979, after Buishand and Velds (1980). Data from stations within dotted lines have been published as a single regional time series.

5 THE PRESENT SITUATION (ULTIMO 1986)

The situation by the end of 1986 was unsatisfactory. The publication of the evaporation-data in the climatological bulletins evokes the suggestion that data are provided on a physical climatological phenomenon with an accuracy comparable to that of standard climatological data. In fact the user of the data is supplied only with a rather rough estimate of evaporation. This leaves the user the difficulty of assessing from what applications of the data satisfactory results may be expected.

E.g. for open water, differences between the actual evaporation and E_o-values may amount to 20% in spring and autumn (De Bruin and Kohsiek 1979) and in winter far larger differences have been reported. (Penman, 1956; see Table 1).

To improve this situation the KNMI/TNO Working Group considered several possibilities:

- To select or design a recommendable version of Penman's method.
- To terminate the dissemination of evaporation-data by KNMI and to advise hydrologists and agronomists to calculate evapo(transpi) ration from climatological data with algorithms designed for their special purposes and their requirements for accuracy.
- To select or design an alternative to Penman's method for the calculation of estimated evaporation by KNMI.

The first possibility was rejected, as this would suggest that Penman's method can be improved in such way that a high degree of accuracy can be achieved. Moreover the existing confusion would be exacerbated, as again time series of Penman-data would be introduced, slightly different from the existent data sets.

Arguments in favour of terminating the calculations at KNMI were outweighed by the apparent wish that KNMI should continue to provide an estimate of evaporation comparable in quality to the Penman-data.

It was therefore decided to investigate the possibility of selecting an alternative to the Penman method.

It appeared that a version of the formula of Makkink (1957, 1961) for the potential evapotranspiration of grass was a good basis to define a "reference crop evapotranspiration, E_r" as an alternative for E_o-data in climatological practice, (De Bruin, 1987).

In particular the fact that only two climatological quantities are needed for the calculation of E_r (solar radiation and air temperature) opens up the possibility of performing the required calculations for a network of stations of a sufficient density over a long period of time to come.

6 THE FUTURE

De Bruin (1987) has defined the requirements for the future practice for the calculation and dissemination of evaporation-data by KNMI.
From a practical point of view we may add that:

- calculations should be executed using daily climatological data as input; evaporation data over longer periods should only be obtained by the summation of daily values;
- the data should be provided throughout the year;
- when an evaporation figure is given for a geographical position it should be made clear whether the figure is calculated from climatological data observed at that location or whether it has been derived from data obtained elsewhere, e.g. by interpolation.

For the potential evapotranspiration of short grass Makkink proposed the expression:

$$\lambda E_M = C_1 \frac{s}{s + \gamma} K^{\downarrow} + C_2 \qquad (4)$$

The best values for C_1 and C_2 are obtained by agronomical research on the potential evapotranspiration of grass.

In climatological practice it is better to abstract from this agronomical problem by defining a hypothetical reference crop for which the potential evapotranspiration is defined with postulated, fixed, values for C_1 and C_2. Accordingly it has been decided to calculate, from April 1st 1987, a reference crop evapotranspiration with $C_1 = 0{,}65$ and $C_2 = 0$.
The potential evapotranspiration of real crops (including grass) can then be estimated with a system of crop factors, as explained in this volume by Feddes and by De Bruin.

The values for the crop factors may reflect, in the course of time, new results of agronomical research on potential evapotranspiration.
It should be emphasized that the above implies that E_r is not defined as a physical quantity, but by a so-called operational definition:

$$\lambda E_r = 0,65 \frac{s}{s+\gamma} K^{\downarrow}$$

Values for E_r will be calculated from daily values of global radiation and temperature for a network of stations as shown in figure 3.
Some of the drawbacks related to the dissemination of E_o-data will still be attached to the future practice.
For example, as well as in the case of E_o-data, the selection of applications of E_r-data so, that satisfactory results may be expected, can only be made by skillful hands.
In particular in winter-time neither E_o nor E_r can be considered as usable estimates for actual evaporation. The decision to disseminate E_r-data also in winter therefore requires some justification.
Firstly, from year to year the periods of winter weather are irregularly distributed over the winter half year.
It is therefore desirable to calculate the E_r-data throughout the year and to decide on their significance afterwards.
Furthermore it should be noted that E_o-data have often been used for rough estimates of evaporation the year around.
It holds both for E_o and E_r that, in view of the low absolute values in winter, the accuracy of the estimates of quantities such as a yearly summation will not be affected appreciably by the inclusion of the winter period in the statistics.
Data on E_r, both for decades and monthly, will be published in the Monthly Climatological Bulletins of KNMI. Besides this, data (also daily values) can be obtained on request on shorter notice.

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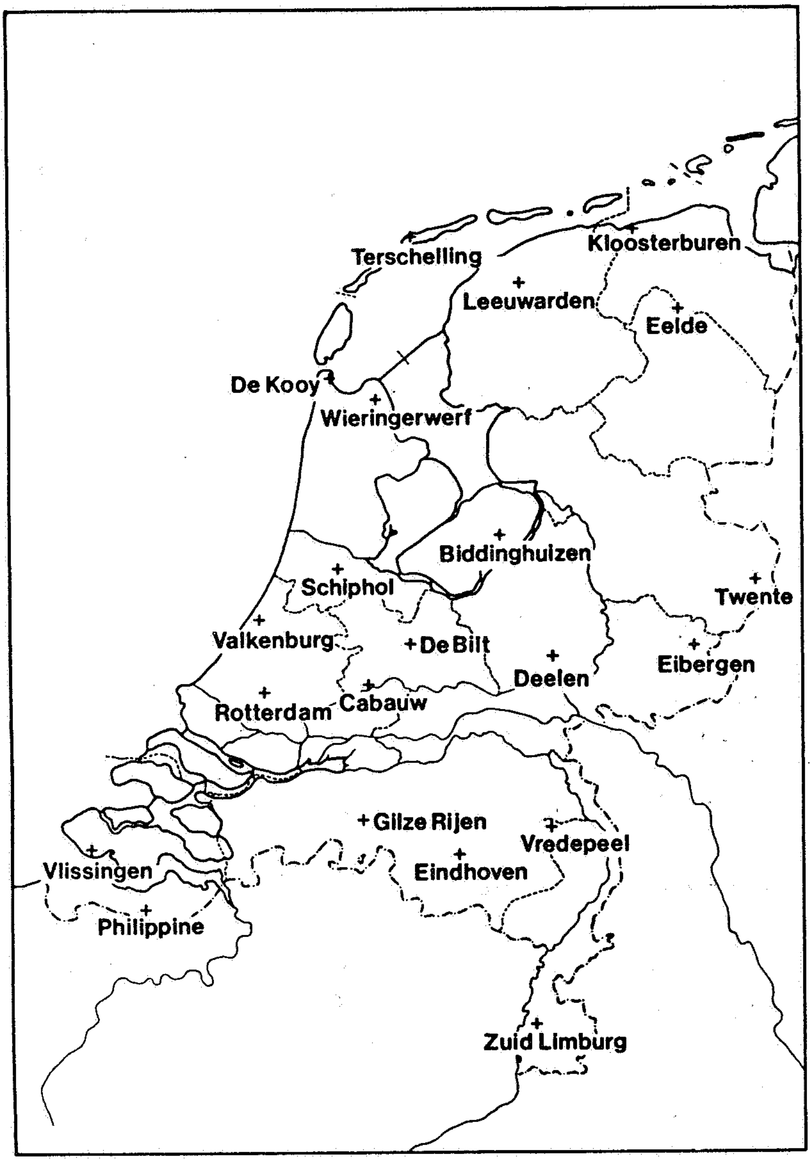

Figure 3 First draft for the network of stations for the observation of global radiation and temperature. It is planned that three more stations will be added in agricultural districts.

Table 2 List of symbols

For the parameters in Penman's formula it is indicated in brackets from what climatological data the numerical values have been derived in the Netherlands climatological practice.

Symbol	Definition	Unit
E	Evaporation	kg m^{-2} s^{-1}
E_o	"Open water evaporation" according to Penman	kg m^{-2} s^{-1}
E_M	Potential evapotranspiration of grass according to Makkink	kg m^{-2} s^{-1}
E_a	Isothermal evaporation	kg m^{-2} s^{-1}
Q^*	Net radiation [T_a, sunshine duration, e_a]	W m^{-2}
G	Soil heat flux density or change per second of heat stored per m^2 in water body	W m^{-2}
$K^{\downarrow}$	Global radiation	W m-2
T_a	Air temperature at screen height	K
e_a	Water vapour pressure at screen height [T_a, r.h.]	mbar
$e_s(T_a)$	Saturation water vapour pressure at temperature T_a. [T_a]	mbar

Symbol	Definition	Unit
r.h.	Relative humidity	%
γ	Psychometric constant [T_a]	mbar K^{-1}
s	Slope of the curve of saturation water vapour pressure versus temperature at T_a. [T_a]	mbar K^{-1}
λ	Specific heat of evaporation of water [T_a]	J kg^{-1}
u, u_2	Wind velocity at 2 m height [wind observations at various heights]	m s^{-1}

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EVAPORATION AND WEATHER: INTERACTIONS WITH THE PLANETARY BOUNDARY LAYER

H.A.R. de Bruin and A.A.M. Holtslag

1 INTRODUCTION

Due to their peculiar thermodynamic, optical and other properties water substances play an important role in phenomena which are related to "weather" and "climate". The behaviour of the atmosphere would be much simpler if water vapour was absent in the earth's atmosphere.

Water vapour plays a part in the formation of fog, clouds and precipitation. Its strong absorption bands in the infrared region are crucial in the so-called "greenhouse-effect", by which the mean temperature at the earth's surface is about 288 K instead of 254 K (being the mean temperature in the absence of the greenhouse-effect). Moreover, water vapour affects the vertical stability of the atmosphere by which the pressure of water vapour tends to increase vertical atmospheric motion and thus precipitation.

The main source for atmospheric water vapour is evaporation at the earth's surface. In spite of the fact that meteorologists recognize the importance of water vapour for atmospheric processes, until now evaporation is hardly ever described properly in models developed for e.g. weather forecasts or climate studies.
In these models the dynamics and thermodynamics of the atmosphere are described very detailed, however surface processes such as evaporation are generally treated as independent boundary conditions.

The other part of our story is the development in the last decades of evaporation research carried out by hydro- and agrometeorologists, primarily to solve practical hydrological and agricultural problems. Evaporation models developed for these purposes usually describe in detail the plant-soil system, but take the properties of the overlying air as independent boundary conditions.

Recently, it has been recognized both by meteorologists and hydro- and agrometeorologists that evaporation and properties of the lower atmosphere are no independent variables.
As a result, there is an increasing interest of meteorologists and climatologists in land surface processes, including evaporation, on the other hand the hydro- and agrometeorologists have made a start with including planetary boundary layer theory in their evaporation models.

It is the objective of the first part of this paper to illustrate, using simple examples, the interrelation between evaporation on one hand and the temperature and humidity of the planetary boundary layer on the other. From these examples it will be made clear why the evaporation formula by Priestley and Taylor (1972) or that by Makkink (1957) works so well.

In the second part of this paper (section 5) a brief description is given of a meteorological model developed for short term weather forecasts in which evaporation plays an important role. We start with a brief description of the planetary boundary layer.

2 THE PLANETARY BOUNDARY LAYER

Processes that take place at the earth's surface affect directly the lowest layer of the atmosphere. This layer is denoted as the *atmospheric* or *planetary boundary layer* (PBL). Generally, the flow within the PBL is turbulent. The turbulent state of the PBL appears to be primarily determined by the wind speed, surface roughness and the surface fluxes of sensible heat and water vapour (H and E, respectively).

If the PBL is heated from below, i.e. the vertical surface flux density of sensible heat, H, is positive, the PBL is *unstably* stratisfied. Then relatively warm (and less dense) air is near the surface, whereas at greater height the air is cooler and thus more dense. This state occurs during daytime. On the other hand the PBL is *stable* if $H < 0$, i.e. the surface is cooling. Finally, if H is small and wind speed is large the PBL is *neutrally* stratisfied.

Evaporation, E, plays an important role in this story. First of all, through the energy balance at the earth's surface:

$$H + \lambda E = Q^* - G \qquad (1)$$

where Q^* is net radiation and G is soil heat flux. For given $Q^* - G$, λE determines directly H and thus indirectly the turbulent state of the PBL. Moreover, water vapour affects the air density. In this way, E influences directly the stability of the lowest atmosphere.

Generally, the terms of the surface energy balance above land show a diurnal cycle. This is illustrated in Figure 1, which shows a typical example for the diurnal cycle in summertime on a cloudless day at Cabauw, the Netherlands. The data of Figure 1 are discussed in De Bruin and Holtslag (1982).

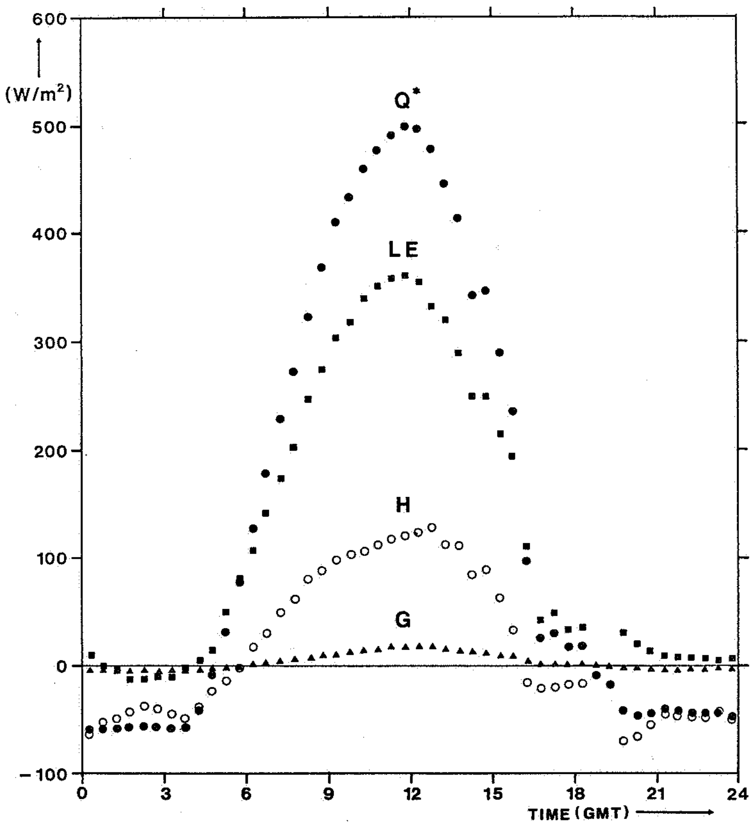

Figure 1 The observed diurnal variations of the components in the surface energy balance at Cabauw on a cloudless day in summertime (May 31, 1978)

It is outside the scope of the present paper to describe in detail the state-of-the-art of the present PBL-research. For this the readers are referred to textbooks such as that edited by Nieuwstadt and Van Dop (1983). Here we will confine ourselves to the unstable PBL, since, usually, most evaporation occurs during daytime.

Under clear sky conditions the unstable PBL is most simple to describe. Then, the PBL is often well-mixed. For simplicity we will restrict ourselves to this case.

A schematic picture of the well-mixed PBL is given in Figure 2. Up to $z = h$ (= the PBL-height), q and θ are constant with height at q_m and θ_m respectively, due to turbulent mixing. At $z = h$ the PBL is capped by an inversion; for $z > h$ the air is stable and is characterized by $d\theta/dz = \gamma_\theta$ and $dq/dz = \gamma_q$. Usually, the transition layer between the well-mixed layer and the stable air aloft is small, so that the profiles can be approximated as shown in Figure 2, i.e. at $z = h$ the θ and q-profiles show a jump of respectively $\Delta\theta$ and Δq.

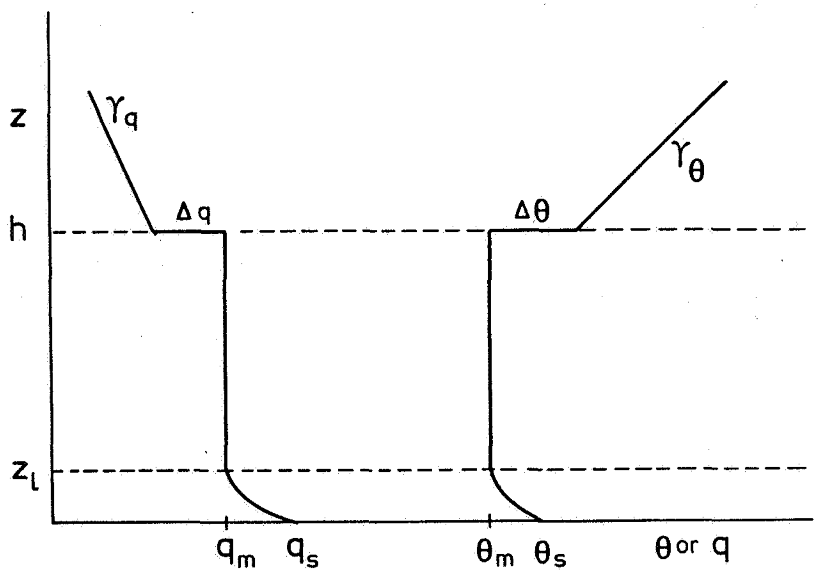

Figure 2 Profiles of potential temperature (θ) and specific humidity (q) in a well-mixed atmospheric boundary layer

The lowest part of the PBL, between $z = 0$ and $z = z_1$, is called the surface layer (or constant flux layer). Herein the gradients of θ and q are sharp: going down from the top of the surface layer ($z = z_1$) to the ground, θ and q increases rapidly from θ_m and q_m to the surface values θ_s and q_s respectively. Usually, $z_1 = 0.1$ h, so that the heat capacity and the capacity to store water vapour in the surface layer are small compared to those of the entire PBL.

The processes taking place in the well-mixed PBL can be described briefly as follows. In first approximation the PBL is transparent for shortwave (solar) radiation, implying that there is no direct heating of the PBL by the sun. The surface is heated by solar radiation and, in its turn, the surface heats the PBL, which leads to convective production of turbulence in the PBL. Moreover, wind produces mechanically turbulence due to wind shear induced by surface roughness. In clear days with sufficient solar radiation, the turbulence is vigorous enough to mix the PBL above the surface layer.

Due to turbulent eddies that intrude into the stable air aloft the well-mixed layer, air from above the inversion is entrained into the PBL. This entrainment process is primarily determined by surface heating (thus by the sensible heat flux density H). As a result of the surface heating h growth from about 100-200 m in the early morning up to 1-2 km in the late afternoon in summertime clear sky conditions. This is illustrated in Figure 3 for the period of Figure 1. For cloudy skies and also in wintertime, the diurnal variation of h is much less (as it is for H). A further discussion is given by Holtslag (1987).

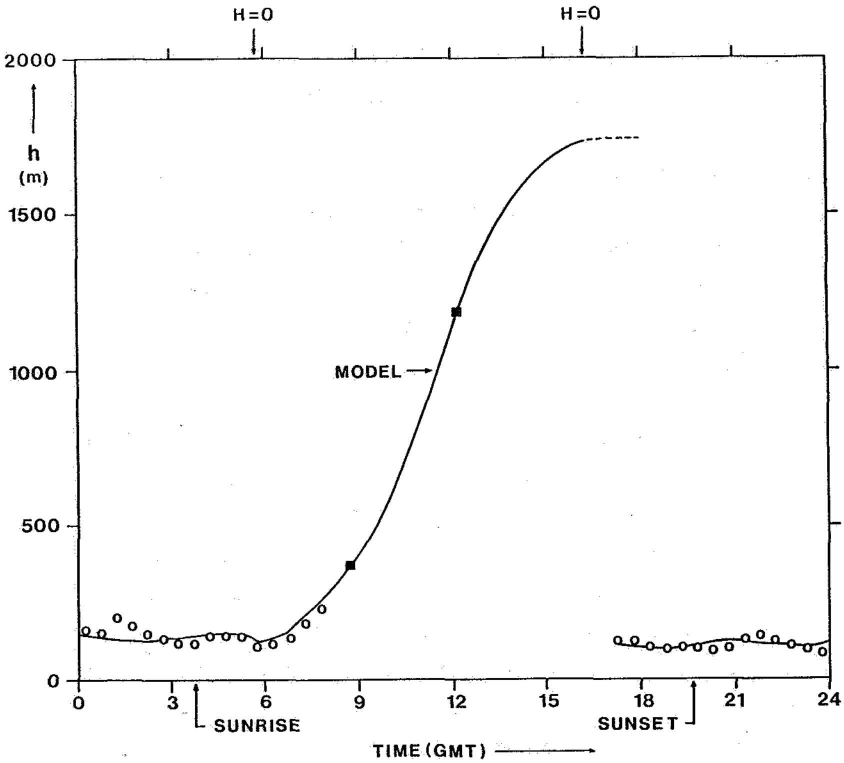

Figure 3 The diurnal variation of the turbulent PBL height h for the period of Figure 1. Indicated are the moments of sunrise, sunset and H = 0. Dots indicate observations of h with an acoustic sounder, squares are estimates of h obtained from temperature profiles. The indicated line is based on model calculations (see Van Ulden and Holtslag, 1985; Holtslag, 1987).

In the picture of the well-mixed PBL of Figure 2 it is tacitly assumed that the surface heating (H) is an independent variable, i.e. H and θ_m are supposed to be independent. In reality this is not true. If θ_m is increased due to surface heating, there is a tendency to decrease H, because the difference between θ_m and θ_s decreases. Similar things apply to the surface evaporation, E, and q_m.

In the next section we will discuss this interrelation between H and θ_m or E and q_m. Firstly, we will consider the very simple case of the *closed box model*. Herein the PBL-height, h, is taken constant. Next the more realistic case is considered, where h is allowed to grow. In that case *entrainment* is taken into account.

3 THE CLOSED BOX MODEL

To illustrate the relationship that exists between the surface fluxes of sensible heat and water vapour on one hand and the temperature and humidity of the air near the ground we firstly consider the simple case where the planetary boundary layer (PBL) is assumed to be a closed box. This closed box model has been used by Perrier (1980) and in a somewhat different form by McNaughton (1976). Moreover, it is described by McNaughton and Jarvis (1983).

It is assumed that the PBL is well-mixed above the surface layer, implying that the potential temperature, θ, and specific humidity, q, are constant with height. Within the surface layer, gradients of θ and q are allowed. Here the Penman-Monteith equation (De Bruin, 1987) applies. Under these conditions the depth of the surface layer is typically one tenth of the PBL-height, h. The PBL is capped by an inversion, which is assumed to act as an impermeable lid for heat and water vapour.

If at the surface sensible heat and water vapour are supplied the (potential) temperature, θ_m, and specific humidity, q_m, of the well-mixed PBL will increase according to

$$\frac{\partial \theta_m}{\partial t} = \frac{H}{\rho c_p h} \quad \text{and} \quad \frac{\partial q_m}{\partial t} = \frac{E}{\rho h} \tag{2}$$

In Eq. (2) advection is ignored (see section 5).

We define now:

$$D_m = q_s(\theta_m) - q_m \tag{3}$$

where $q_s(\theta_m)$ is the saturated specific humidity at θ_m. D_m is a measure for the specific humidity deficit of the PBL. Differentiating Eq. (3) and combining the results with (2) yields

$$\rho c_p h \frac{\partial D_m}{\partial t} = s\ H - \gamma(\lambda E) = s\ A-(s+\gamma)\lambda E \tag{4}$$

where $s = \frac{dq_s}{d\theta}$ at θ_m, $\gamma = {}^{c}p/\lambda$ and $A = Q^*-G = H + \lambda E$.

From Eq. (4) it is seen that D_m and E are interrelated in the case of the closed box model. This result can be combined with the Penman-Monteith equation, which can be written as

$$\lambda E = \frac{sA}{s+\gamma^*} + \frac{\rho c_p D_m/r_a}{s+\gamma^*} \tag{5}$$

where $\gamma^* = \gamma(1 + r_s/r_a)$, r_s is the canopy resistance and r_a is the atmospheric resistance of the surface layer.

From Eqs. (4) and (5) λE can be eliminated. This leads to a simple first order differential equation for D_m that can be written as

$$\frac{\partial D_m}{\partial t} = \frac{D_{eq}-D_m}{\tau} \tag{6}$$

where the *equilibrium saturation deficit* D_{eq} is defined by

$$D_{eq} = \frac{Ar_a}{\rho c_p}\ \frac{s\ \gamma^*-s\gamma}{s + \gamma} \tag{7a}$$

and the *time constant* is given by

$$\tau = \frac{hr_a\ (s + \gamma^*)}{(s + \gamma)} \tag{7b}$$

D_{eq} can be regarded as a forcing function. If D_{eq} and τ are constant in time D_m will approach D_{eq} and $\frac{\partial D_m}{\partial t}$ tends to zero. Then, it follows from Eq. (4) that λE reaches its equilibrium rate, λE_{eq}, defined by

$$\lambda E_{eq} = \frac{s}{s+\gamma} A \tag{8}$$

Hence, for a closed PBL the Priestley-Taylor (1972) equation is obtained with $\alpha = 1$ (see Eq. 15). Although this example is not realistic, since h is not constant, it illustrates clearly that E and D_m are interrelated, and that finally in first order E is independent of D_m and determined primarily by net radiation (usually $Q^* >> G$ during daytime).

In the next section a more complete PBL model will be described.

4 A MIXED LAYER MODEL, INCLUSIVE ENTRAINMENT

In reality the PBL height is not constant as assumed in the previous section. Due to turbulent eddies, created within the well-mixed layer (primarily by surface heating), the PBL will grow, since these eddies intrude into the stable air aloft. As a consequence this air, which is relatively warm and dry, is entrained into the well-mixed layer affecting its (potential) temperature, θ_m, and its specific humidity, q_m.

According to Tennekes (1973) and others Eq. (2) has to be replaced by

$$\frac{\partial \theta_m}{\partial t} = \frac{H}{\rho c_p h} + \frac{\Delta\theta}{h}\,\frac{\partial h}{\partial t} \tag{9a}$$

$$\frac{\partial q_m}{\partial t} = \frac{E}{\rho h} + \frac{\Delta q}{h}\,\frac{\partial h}{\partial t} \tag{9b}$$

where $\Delta\theta$ and Δq are shown in Figure 2 as the jumps at $z = h$. Note that again advection terms are ignored, so only local effects are considered.

The last terms of (9) describe the effect of the growth of the PBL height and the resulting entrainment of warmer and drier air. Because the entrainment process is primarily steered by surface heating the last term in the r.h.s. of (9a) often appears to be proportional to the first, so

$$\frac{\partial \theta_m}{\partial t} = \frac{(1+c)\,H}{\rho c_p h} \tag{10}$$

where c = 0.2 (Driedonks, 1982).

Let θ_t and q_t be the values of θ and q of the stable air at the top of the PBL ($z = h$). Then $\theta_t = \theta_m + \Delta\theta$ and $q_t = q_m + \Delta q$ (usually, $\Delta q < o$). Moreover, we define $D_t = q_s(\theta_t) - q_t$.

It can be shown from Eq. (9) that approximately

$$\frac{\partial D_m}{\partial t} = \frac{s\,H - \gamma\lambda E}{\rho c_p h} + \frac{D_t - D_m}{h}\,\frac{\partial h}{\partial t} \tag{11}$$

This equation is the equivalent of Eq. (4) concerning the closed box model with h constant. Again the last term describes the effect of entrainment. Note that now h neither D_t are constant. According to Driedonks (1981) $\partial h/\partial t$ is approximately

$$\frac{\partial h}{\partial t} \cong \frac{(1+2c)}{(1+c)\,\gamma_\theta}\,\frac{\partial \theta_m}{\partial t} \tag{12}$$

The latter equation applies under convective conditions, provided $h \gtrsim 3h_o$ (h_o = initial value of h just after sunrise).

After some algebra it can be shown from Eqs. (10)-(12) that

$$\rho c_p h \frac{\partial D_m}{\partial t} = (s + s^*)\ H - \gamma \lambda E \tag{13}$$

where $s^* = \dfrac{(D_t - D_m)\ (1+2c)}{\gamma_\theta\ h}$

It appears that also for this more general case an equation similar to Eq. (4) can be derived. This implies that also in this case under stationary conditions D_m strives to an "equilibrium" value D_{eq}; if $\partial D_m/\partial t$ vanishes, λE approaches.

$$\lambda E = \frac{A}{1 + \frac{\gamma}{s+s^*}} \tag{14}$$

Defining parameter α by

$$\lambda E = \alpha \frac{s}{s+\gamma} A \tag{15}$$

it is seen that now

$$\alpha = \frac{s+s^*}{\gamma+s+s^*} \left(\frac{s+\gamma}{s}\right) \tag{16}$$

Consequently, $\alpha > 1$ if $s^* > 0$. Then $D_t > D_m$, i.e. the stable air has a larger specific humidity deficit than the well-mixed layer. It is also possible that $\alpha < 1$, then s^* has to be negative or $D_t < D_m$.

Whether D_m is less or greater than D_t depends on several factors, notably: a) the surface fluxes H and λE and b) D_t itself, i.e. the "dryness" of the stable air aloft.
To solve the set of equations listed above an additional equation for the surface fluxes is needed, e.g. the Penman-Monteith equation (5). In this way, one arrives at a set of coupled differential equations, which can be solved only if the initial and boundary conditions are

known. It is outside the scope of this paper to discuss this matter in detail. For special cases De Bruin (1983) and McNaughton and Spriggs (1986) gave a solution. In Figure 4 some results of the paper by De Bruin (1983) are presented. Herein the calculated day-time variation of parameter α is shown. It is seen that around noon $\alpha = 1.3$ if $r_s = 0$; $\alpha = 1$ if $r_s = 60\text{-}90\ sm^{-1}$ and $\alpha < 1$ if $r_s > 100\ sm^{-1}$. This is in good agreement with observations. Note that the more complete approach by McNaughton and Spriggs (1986) yields similar results (Jarvis and McNaughton, 1986).

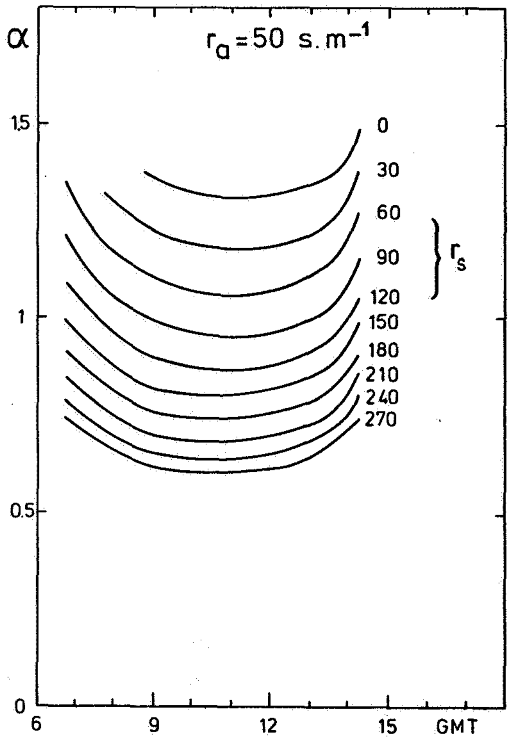

Figure 4 Daytime variations of the computed α for different values of r_s with $r_a = 50\ sm^{-1}$ (De Bruin, 1983).

For practical calculations the above result is important. Normally, it appears that a Priestley-Taylor type of estimate can be used for the surface fluxes E and H. This result can be used for rough estimates of evaporation. An example is Makkink's formula discussed by De Bruin

(1987). In the next section an example of another application is given, notably for weather forecast purposes.

Finally, it is noted that recently authors such as Ten Berge (1986) and Pan and Mahrt (1987) coupled models for the PBL and the surface fluxes. These authors considered the case of bare soil. Note that then the Penman-Monteith equation can not be used.

5 A SIMPLE PBL MODEL FOR SHORT RANGE WEATHER FORECASTING

It is the objective of this section to present an example of a simple PBL-model that includes a sub-model for the surface fluxes. The governing equations are extensions of Eq. (9) and are given by

$$\frac{d\theta_m}{dt} = \frac{H}{\rho c_p h} + \frac{\Delta\theta}{h}\left(\frac{dh}{dt} - w_h\right) \tag{17a}$$

$$\frac{dq_m}{dt} = \frac{E}{\rho h} + \frac{\Delta q}{h}\left(\frac{dh}{dt} - w_h\right) \tag{17b}$$

where the d/dt terms at the left denote the total change of mean temperature θ_m and humidity q_m. In general we can write for θ_m (and similar for q_m)

$$\frac{d\theta_m}{dt} = \frac{\partial\theta_m}{\partial t} + U\frac{\partial\theta_m}{\partial x} + V\frac{\partial\theta_m}{\partial y} \tag{18}$$

where the first term at the r.h.s. of (18) is the local change of θ_m as in Eq. (9a), and the second and third term of the r.h.s. of (18) are known as advection terms. These terms take account for the change of θ_m (and q_m) by horizontal transport. The vertical movement is taken into account in w_h of Eq. (17).

It can be shown that Eqs. (17) are also valid during stable conditions e.g. in cases for which the temperature and humidity profiles are not uniform in the PBL (see Driedonks et al, 1985). In such cases the profiles in the PBL need to be described as a function of relative height e.g. z/h. In that case the mean development of θ and q, denoted by θ_m and q_m are calculated with Eq. (17).

In a weather prediction model for the PBL, Eq. (17) need to be solved. This means that initial conditions are needed for temperature and humidity together with the surface fluxes H and E. Moreover the influence of advection needs to be calculated. In Reiff et al (1984) it is shown that for forecasting the development of the PBL more than 12 hours ahead, advection has to be included.

A manner to take the influence of advection into account is discussed by Reiff et al (1984) and Driedonks et al (1985). They consider an "air mass transformation model", in which the development of the PBL is calculated along predicted trajectories. Figure 5 gives an example of such trajectories starting at different locations on different pressure levels, but ending at the same time at a given location (here the arrival time is March 25, 1987, 13.00 local time for De Bilt, the Netherlands). At each trajectory the value of the pressure at the arrival time is given. The lowest trajectory (1000 mb) is thought to be representative for the transport of the boundary layer. The starting time of the trajectories is March 24, 1987, 01.00 local time, e.g. 36 hrs. before the arrival time. These trajectories can be calculated with a weather prediction model, like the one of the European Centre (ECMWF).

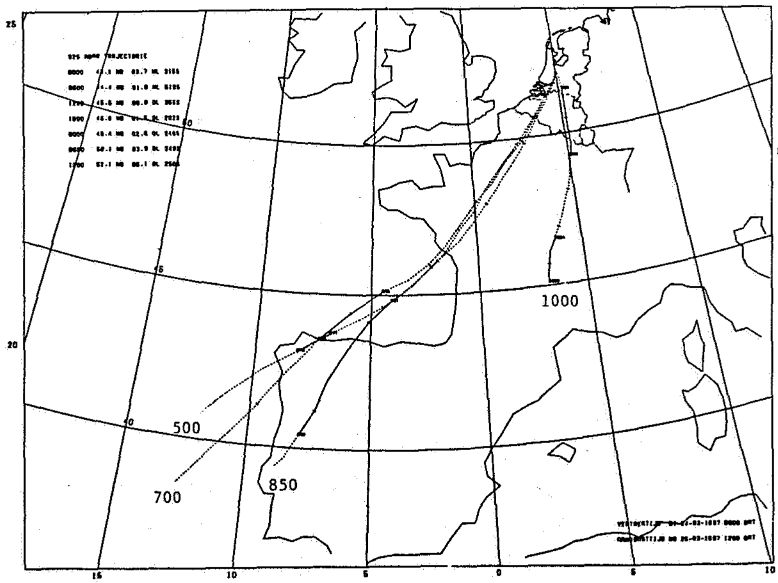

Figure 5 Predicted trajectories ending at De Bilt, the Netherlands on March 25, 1987, 13.00 local time for indicated pressure levels in De Bilt (see further explanation in text).

In the source area of the trajectories, observations of radiosounds are used to construct an initial temperature and humidity profile for the boundary layer and for the atmosphere aloft. With this information Eq. (17) can be used to calculate the total rate of change of θ_m and q_m, provided H and λE are known. This cycle is repeated every 10 minutes until the place of arrival has been reached.

It is characteristic that the surface fluxes need to be described in terms of other predictable quantities to solve the PBL equations. During daytime the surface fluxes H and E are parameterized with the Priestley-Taylor approach (see the preceeding sections), G is related to Q* and Q* is parameterized in terms of predicted total cloud cover and solar elevation. Here the findings of Holtslag and Van Ulden (1983) are used. These authors show the type of uncertainty, which has to be expected for this kind of applications.

During nighttime the latent heat flux is generally small, and the sensible heat flux is strongly influenced by wind speed. In the present model the results of Holtslag and Van Ulden (1982) are used. These results were recently generalized by Van Ulden and Holtslag (1985) and Holtslag and De Bruin (1987). As an example of the typical behaviour of the surface fluxes with wind speed during nighttime we present Figure 6 (adopted from Holtslag and De Bruin, 1987). In Figure 6 u_{*N} is related to windspeed U_z, by

$$u_{*N} = \frac{kU_z}{\ln\left(\frac{z}{z_o}\right)} \tag{19}$$

where k is the Von Kármán constant ($k \cong 0.4$), z is the height above the surface and z_o is the so-called effective roughness length for momentum. Here $z = 10$ m and $z_o = 0.15$ m are used. So u_{*N} can be interpreted as a scaled wind speed with respect to the surface roughness conditions.

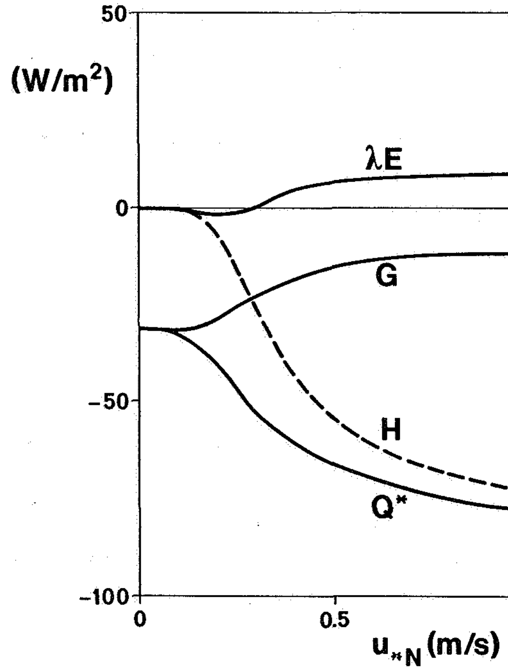

Figure 6 The variation of the terms in the surface energy balance of Eq. (1) during nighttime stable conditions over land, according to model calculations of Holtslag and De Bruin (1987). Here u_{*N} is defined by Eq. (19).

From Figure 6 it is seen that for small wind speeds $\lambda E < 0$, so condensation occurs at the surface. For larger wind speeds also during nighttime evaporation will occur. Generally the absolute value of λE is small compared with the other terms. On the other hand H is strongly influenced by wind speed and its magnitude is the same order as Q* for large wind speeds.

So far we have discussed the surface fluxes above land surfaces. When the air passes over the sea other types of parameterizations are needed. In these circumstances the fluxes are often taken proportional to the temperature and humidity differences between the sea and the air. Details are given in Reiff et al (1984).

Application of Eqs. (17) in the above described manner ultimately leads to forecasts of the temperature and humidity profiles in the PBL up to 36 hours ahead. Also the boundary layer height is obtained in this way. From this section it might be clear that the surface fluxes have a strong impact on the predictions. Results of such forecasts are discussed by Reiff et al (1984) and Driedonks et al (1985). In Reiff (1987) a review is given on the forecasting of clouds and fog in the PBL.

6 SUMMARY

In this paper we have discussed the interaction of the surface fluxes with the planetary boundary layer (PBL). After a description of the main PBL characteristics we have illustrated the physical background of the Priestley-Taylor approach. Subsequently the findings are applied into a PBL-model for short range weather forecasting of temperature and humidity profiles.

ACKNOWLEDGMENT

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INTERNATIONAL INVESTIGATIONS OF LARGE-SCALE EVAPORATION

W.J. Shuttleworth

ABSTRACT

Computer models of the earth's atmosphere used for weather and climate forecasting currently contain very simple descriptions of land-surface hydrology, with little or no recognition of variations in surface vegetation. This paper describes three recent or on-going international experiments designed to improve our knowledge of surface processes, and their description at a scale and complexity consistent with that required in such atmospheric models.

The first is a two year, single-site micrometeorological/hydrological study in central Amazonia, the results of which now provide calibration of tropical forest vegetation in new land-surface descriptions specifically designed for use in climate models. The second, the Hydrologic Atmospheric Pilot Experiment (HAPEX), was carried out during 1986 in South-West France. It provided detailed measurements of weather and surface-flux variables simultaneously over several agricultural and forest crops, and will investigate their integration to larger scale using aircraft, satellite and catchment data. The last study, the First ISLSCP Field Experiment (FIFE), is part of the International Satellite Land Surface Climatology Programme (ISLSCP) and is taking place during 1987/8 in Kansas, USA. It will investigate and evaluate the potential use of satellite data for routine climate monitoring, and climate model calibration.

1 INTRODUCTION: MACROHYDROLOGY

The simple observation that the earth's climate within continents differs from that over the oceans demonstrates that surface processes influence weather. Computer models of the earth's atmosphere have recognized this fact in elementary form, but in general have very simple formulations of land-surface hydrology. Despite the coarseness of this description such models have been successful in demonstrating two effects. Firstly, the earth's climate, as simulated in such models, is indeed sensitive to large changes in simple properties such as albedo, surface roughness and soil moisture (see for example Charney et. al., 1977; Sud et. al., 1985; Shukla and Mintz, 1982). Secondly, they have shown that water vapour and energy entering the atmosphere from the ground at one place can travel large distances before returning to the surface elsewhere (see for example, Eagleston, 1986). The implication is that changes in surface processes, perhaps generated by human activity, may well have significant and possibly detrimental consequences on the climate locally, and possibly at considerable distance.

To the hydrologist the reliability, or otherwise, of the detailed quantitative predictions made with such General Circulation Models (GCM's) should not influence the seriousness with which we regard their general predictions. Even if the probability that their forecasts are correct is as low as fifty per cent, the consequences on hydrology, and through this on human well-being, are, in general, so severe that they must be taken seriously. Hydrologists must respond positively to this new challenge.

Research built around this response represents an important, growing and internationally recognized area of hydrological interest, which has come to be called 'Macrohydrology'. The most important and novel aspect of Macrohydrology is that limited resources, both in the computers used to model climate and in the experimental and observational data used to calibrate them, necessitate the creation of average process descriptions relevant to large areas. The area of interest is very significantly greater than that at which hydrologists are accustomed to working, and is typically in the order 400 x 400 km.

Macrohydrology encompasses two broad areas:

(a) incorporating hydrological expertise into improving the description of land-surface properties in the GCM's themselves, and thereby improving the reliability of their climate and weather prediction, and

(b) interpreting the hydrological consequences of any predicted climate change in terms which affect human well-being through the hydrological cycle, such as changes in flood/drought frequency, available water resource and agricultural environment.

This paper outlines international, collaborative experiments designed to provide information which relates to the first of these.

At this stage such experiments are, in some measure, speculative and currently they tend to demand more expertise from process specialists in the hydrological community, to input modelling ideas and physical insight. They will, however, also require increasing support from catchment specialists to provide the integrated, long-term calibration of the ensuing large area average description.

2 THE AMAZON REGION MICROMETEOROLOGY EXPERIMENT (ARME)

This experiment took place as an Anglo-Brazilian collaboration over undisturbed tropical rain forest at a site 25 km North-East of the city of Manaus in the central Amazon basin. The data collection extended over two years, from September 1983 to September 1985, with routine collection of hourly meteorological data above the 35 m high canopy, measurements of integrated rainfall interception loss, and measurements of soil moisture and tension all maintained over this period. In addition three intensive campaigns were carried out with considerably enhanced data collection involving the measurement of radiation components, eddy-correlation measurements of surface-energy and momentum transfer, temperature, humidity and windspeed profiles, and plant physiology studies. Campaigns occurred in September 1983, from

July to September 1984, and from March to September 1985.

These data have since been analyzed in micrometeorological, hydrological and plant physiological terms, and also to provide a description of the water balance at this central Amazonian site (see for example, Shuttleworth et. al., 1984 A, Shuttleworth et. al., 1984 B, Shuttleworth et. al., 1985, Lloyd and Marques, 1987). More importantly in the context of this paper, the data are currently being used to provide calibration of the tropical forest biome in new models of land-surface-energy partition specifically designed for inclusion in computer weather and climate models, such as the SiB model (Sellers et. al., 1986; Sellers and Dorman, 1987) and the BATS model (Dickinson, 1984). The calibration of the SiB model, illustrated in Figure 1, is already completed.

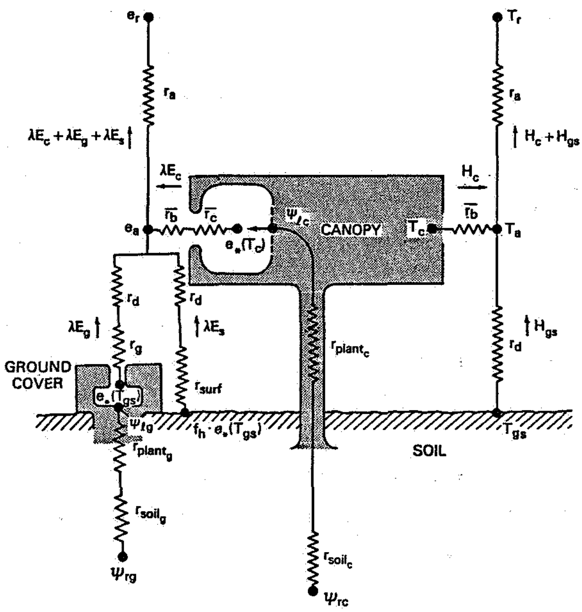

Figure 1 Schematic Diagram of the SiB Model

Clearly the fact that the data used in this calibration is collected at a single site in central Amazonia may limit the credibility of the calibration in continental, let alone global, terms. With this in mind, attempts are now in hand to exploit recent developments in remote sensing theory (e.g. Sellers, 1985) to provide extrapolation of the more important vegetation related parameters. The still extensive existence of a full forest cover over large regions of Amazonia may mean this is a uniquely relevant area in which to apply such ideas. The bulk canopy (stomatal) conductance of the forest is an important control on energy partition, and preliminary investigation suggests further research into a relationship between canopy conductance and satellite measurements of the ratio of surface reflectance in the near infrared and visible regions of the radiation spectrum. Although speculative at this stage, such research may at least provide a basis for estimating the possible error involved in assuming the spatial constancy of the single point calibration, and the sensitivity of climate predictions to this error can then be tested.

3 THE HYDROLOGIC ATMOSPHERIC PILOT EXPERIMENT (HAPEX)

HAPEX is the first attempt to design and implement a complex, multi-disciplinary, multi-site experiment, with diverse techniques simultaneously deployed towards the central objective of providing measurement and modelling of land-surface-energy partition at a scale approaching that used in GCM's. As such it represents an experiment in carrying out experiments of this type. It was stimulated by several international organizations, notably the World Meteorological Organization under the World Climate Research Programme, and funded by both national and international agencies. It took place during 1986 under the management of the Centre National de Recherches Meteorologiques in Toulouse (see Andre et. al., 1986).

The experimental site was an area 100 km x 100 km in South-West France, which was selected as already having a considerable network of automatic weather stations, and with past and non-going collection of hydrological catchment data over significant portions of the experimental area. About 60 per cent of the selected site is covered with agricultural crops of diverse species, while the remainder is an established forest of Maritime Pine.

Routine meteorological data, surface energy-flux data, rainfall and runoff data and soil moisture soundings were collected over an extended period approaching one year. A single intensive study period, lasting about 10 weeks, took place starting in May 1986.
During this the long-term data were supplemented with intensive air-crafts measurements, frequent radio soundings, additional surface energy-flux measurements (notably over the forest), and with the detailed collection of plant physiological and botanical data.

The primary thrust of the experiment is towards investigating the techniques and procedures involved in integrating the different surface-energy partition measured at many sites and over diverse crops to a much larger scale. Such investigation involves the use of experimental techniques, such as hydrological (catchment) integration, aircraft measurement, boundary layer sounding, and remote sensing from both air-craft and satellite. It also involves the use of analytic or numerical techniques, such as the application of meso-scale meteorological models as an integrating mechanism. Figure 2 shows a preliminary but encouraging comparison between surface measurements of sensible heat flux made at a single site in the forested portion of the study area, and measurements over a wider area of forest deducted from changes in atmospheric temperature between radio-sonde ascents on June 16th, 1986.

Currently analysis is concentrating on the quality control, initial interpretation and intercomparison of the several data sources. Emphasis in the attempt to formulate and integrated description is presently orientated towards the calibration and use of meso-scale meteorological models in this role.

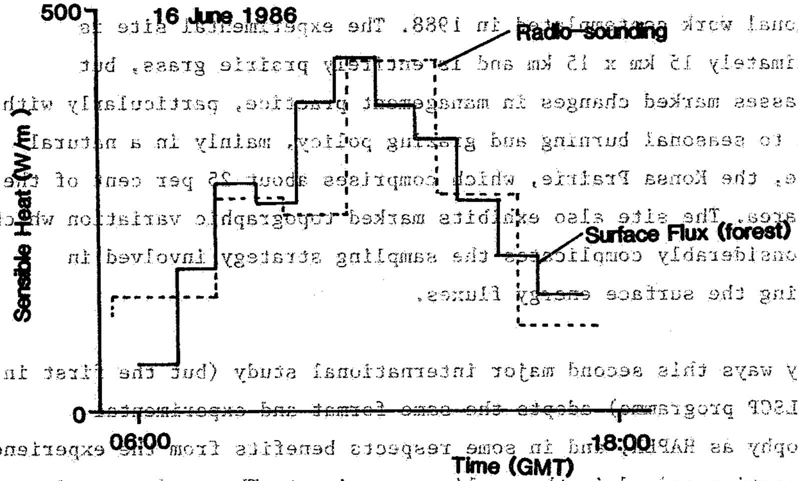

Figure 2 Preliminary comparison between surface measurements of sensible heat flux and estimates from boundary layer radio soundings.

4 THE FIRST ISLSCP FIELD EXPERIMENT (FIFE)

The dual aims of the International Satellite Land Surface Climatology Project (ISLSCP) are to conduct research oriented towards evaluating and implementing the use of satellite based remote sensing as a tool to provide:

(a) routine global monitoring of climate over land surfaces, and

(b) techniques for calibrating the land surface descriptions in models of global climate.

These ambitious, long-term goals are apparent in FIFE, whose emphasis is towards an initial, concerted attempt to provide sufficient simultaneous satellite and ground truth data to test remote sensing as the required integration and extrapolation mechanism required by Macrohydrology.

The experiment will take place in central Kansas during 1987, with additional work contemplated in 1988. The experimental site is approximately 15 km x 15 km and is entirely prairie grass, but encompasses marked changes in management practice, particularly with regard to seasonal burning and grazing policy, mainly in a natural reserve, the Konsa Prairie, which comprises about 25 per cent of the study area. The site also exhibits marked topographic variation which very considerably complicates the sampling strategy involved in measuring the surface energy fluxes.

In many ways this second major international study (but the first in the ISLSCP programme) adopts the same format and experimental philosophy as HAPEX, and in some respects benefits from the experience and expertise gained in the earlier experiment. The study area is much less, and the surface flux sampling frequency commensurately greater; and there are four intensive study periods spread through the growing season rather than one long intensive session near the beginning.

The most notable feature of this, largely NASA sponsored, experiment is the level of financial, instrumental and manpower commitment which will be deployed. It is likely that the FIFE data set will cost perhaps 20 million dollars to produce. The anticipated data output is summarized in the Appendix. Clearly with this level of commitment to data production, and with a commensurate commitment to its analysis and interpretation, FIFE must represent an important initial test of the potential relevance of remote sensing to the integration and extrapolation of land-surface properties.

5 CONCLUDING REMARKS

This paper attempts to draw the attention of the Dutch hydrological community to the growing existence of an important new area of hydrology, Macrohydrology, whose main characteristic is the scale at which simple, but adequate, descriptions of surface hydrology are required. Already a major international effort has developed, and three of the initial experiments are outlined. The ARME experiment was a single point, micrometeorological study, one of whose major objectives was to provide calibration of new land-surface models specifically designed for inclusion in GCM's. HAPEX and FIFE are both important multisite experiments. The HAPEX site comprised measurements over different crops and investigates experimental and numerical integration techniques over large area scales. The FIFE site comprises differently managed samples of one crop, with difficult topography, and directs primary attention towards remote sensing as the required integration tool.

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APPENDIX
OVERVIEW OF DATA TO BE GENERATED BY THE FIFE EXPERIMENT

1 Satellite Data

GOES Data	Data type:	Visible counts (6 Bit) and IR brightness temperatures.
	Resolution:	4 and 8 km pixels
	Frequency:	Hourly as available
	Coverage:	30 x 30 pixel array centred on the FIFE site.
AVHRR Data	Data Type:	5 channel AVHRR HRPT and GAC data
	Resolution:	HRPT data consists of 1 x 1 km contiguous pixels. GAC data consists of 4 x 1 km averages of HRPT 1 km data along every third LAC scan line.
	Frequency:	Twice per day (0230 and 1420 local time)
	Coverage:	All scan lines centred within ± 2° latitude of the FIFE site.
TOVS Data	Data type:	Calibrated HIRS-2 and MSU radiances
	Resolution:	20 km FOV for HITS-2. MSU analyzed to HIRS resolution.
	Frequency:	Twice per day (0230 and 1430 local time)
	Coverage:	7 x 9 pixel array centred on FIFE site.

NASA will independently acquire all the available day-time Landsat Multispectral Scanner (MSS) and Thematic Mapper (TM) data over the site. Additionally, data from the French satellite SPOT may be available.

2 Conventional Meteorological data

Hourly surface reports within 250 km x 250 km area.
Temperature and moisture profile from nearest radio-sonde stations.
Selected NMC upper air data at nearest analyzed grid points.

3 Surface and airborne observations
3A Long-term measurement network

32 Automatic Weather Stations - measuring temperature, humidity, wind speed and direction, soil temperature, reflected solar radiation, net radiation, surface temperature, precipitation, soil moisture, global radiation, direct and diffuse solar radiation, photosynthetically active radiation and longwave radiation.

Terrestrial Water Budget - measurements of surface and subsurface runoff, precipitation, and sample hillslope and local soil hydrology studies.

3B Measurements during four intensive field campaigns

Aircraft Remote Sensing Measurements
Microwave: 21 cm Multi-beam radiometer
Multispectral: 0.45-12.5 μm

ER-2 High Altitude Overflights

Aircraft Measurements of heat and moisture flux

Gust probe equipped aircraft, to measure fluxes of sensible and latent heat, and momentum.

Atmospheric Boundary Layer Soundings

Doppler Radar

Radio sondes

SODAR profiles

Supplementary Radiation Measurements

Spectra-radiometers

Pyrheliometer and Pyranometers

Thermal Infrared Radiometers

Surface Mass and Energy Exchanges

12 stations measuring Bowen ratio and momentum flux.

5 additional eddy correlation measurements at selected sites.

Biophysical Measurements

Hydrological and Physical Parameters of the soil

Hydraulic conductivity

Soil water tension

Soil moisture content

Soil heat capacity and thermal conductivity

Notes

Notes

TNO COMMITTEE ON HYDROLOGICAL RESEARCH

PROCEEDINGS AND INFORMATION

No. 1. Proceedings of Technical Meetings 1-6 (with summaries in English), 1952.

1. Investigations into the water balance of the Rottegatspolder.
2. The water supply for crops I.
3. Observations of groundwater levels.
4. Investigations by drain gauges in the Netherlands.
5. The water supply for crops II.
6. The problem of the increasing salinity of ground and surface water in the Netherlands.

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8. Model research on groundwater flows.
9. Measurements and improvement works in basin of brooks.
10. Geo-electrical research.

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12. Quality requirements for surface waters.

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Groundwater pollution.

No. 22. Proceedings of Technical Meeting 32 (with summaries in English), 1976.
Systems approach to the management of water resources.

No. 23. Proceedings of Technical Meeting 33 (in English), 1977.
Precipitation and measurements of precipitation.

No. 24. Proceedings of Technical Meeting 34 (in English), 1978.
Urbanization and water management.

No. 25. Proceedings of Technical Meeting 35 (in English), 1979.
The relation between water quantity and water quality in studies of surface waters.

No. 26. Proceedings of Technical Meeting 36 (in English), 1980.
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No. 27. Proceedings of Technical Meeting 37 (in English), 1981.
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