A Simple Parameterization of the Surface Fluxes of Sensible and Latent Heat During Daytime Compared with the Penman–Monteith Concept

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ABSTRACT

A comparison is made between two methods for determining the surface fluxes of sensible and latent heat during daytime. The first method, known as the Penman–Monteith approach, incorporates a more complete description of the physics. However, it needs a relatively large number of input parameters, which is inconvenient in many applications. The second method is a modification of the Priestley–Taylor evaporation model, which needs only net radiation, air temperature and an indication of the moisture condition at the surface. Both models are compared on the basis of hourly micro-meteorological data above short grass obtained in the Netherlands during the summer of 1977. The experiments were performed under predominantly unstable conditions [$0 \geq z/L_0 \geq -0.3$; z = (mean) measuring height, L_0 = Obukhov length] with weak or no advection. It appears that, under these environmental conditions, the models have a similar skill. Therefore, the simple parameterization is preferred for practical purposes. It reveals that this result can be partially explained by the fact that the so-called equilibrium latent heat flux density (LE_{EQ}) and vapor pressure deficit are correlated. The method requires further verification for different climatological conditions.

1. Introduction

A simple description of the surface fluxes of heat and water vapor in terms of routine variables is useful for many purposes, such as:

1) The determination of evapo(transpi)ration from the surface, which is required by hydrology and agriculture.
2) The description of the convective atmospheric boundary layer.
3) The estimation of the stability of the air near the ground, e.g., for air pollution problems.
4) The determination of the input of heat and moisture at the ground into the atmosphere for weather-forecast purposes.

For some of these applications, the fluxes must be described in terms of variables which can be forecast, while for others a parameterization is needed in terms of routine weather data observed in the past at standard meteorological stations. In this paper, a parameterization of the surface fluxes, which has the capability to be useful for both categories, will be described. It is a modification of the evaporation model of Priestley and Taylor (1972).

It is the aim of this paper to compare the skill of this simple model with that of the Penman–Monteith approach (Monteith, 1965). This description contains the most complete physics; however, it has the disadvantage that it needs a relatively large number of input parameters.

For the comparisons, a set of micrometeorological data collected at Cabauw, The Netherlands, in the summer of 1977 is used. We will consider hourly values during daytime.

2. Experimental data and environmental conditions

We analyzed a set of micrometeorological data collected at Cabauw, in the central Netherlands, during the period May–August 1977. The measurements were carried out in a 100 × 100 m field covered with short (~8 cm high) grass. Pastures also surrounded the field, so the advection was weak. The surface fluxes of sensible and latent heat were determined with the well-known energy-budget method, using Bowen's ratio (e.g., Sellers, 1965). The latter was evaluated using ventilated psychrometers (Slob, 1978) at 0.45 and 1.10 m, respectively. The vertical differences of dry- and wet-bulb temperatures were measured directly with thermocouples. The net radiation was measured with a net pyrradiometer of the type described by Funk (1959). The soil heat flux was observed with heat flux plates at depths of 5 and 10 cm at three locations. With the aid of the temperature difference between 0 and 2 cm in the ground, the soil heat flux at the surface was obtained using a method developed by W. H. Slob (personal communication, 1982; see Appendix A).

For this study we transformed the 10-min means into hourly averages. Unreliable values were excluded; these mostly refer to situations with rain or fog.

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The environmental conditions under which the experiments were performed are characterized by the following:

- The sensible and latent heat flux density covered a range of about 0–280 and 0–400 W m^{-2}, respectively.
- The net radiation varied between 0 and 550 W m^{-2}.
- The wind speed at 2 m was between 1–7 m s^{-1}.
- The boundary layer height h, which is usually the height of the first inversion, had a typical value of 200 m in the early morning and of 1500 m at noon.
- For the majority of the data the stability parameter z/L_0 (z is the measuring height, which was ~ 1 m in our case, and L_0 is the Obukhov length) was $0 > z/L_0 > -0.3$. Hence the conditions were predominantly unstable.

From this information it usually follows that the boundary layer height is much greater than the measuring height z, i.e., $z/h \ll 1$, while often $h/|L_0| \gg 1$.

Because $z/h \ll 1$, we did our experiments in the constant flux layer, where z/L_0 is the appropriate stability parameter (Tennekes, 1973, 1982).

The Bowen ratio method is based on the assumption that the exchange coefficients for heat and water vapor are the same. Under the conditions mentioned above, i.e., $0 \geqslant z/L_0 \geqslant -0.3$, this appears to be a good assumption (Dyer, 1974).

3. The models

a. Motivation

According to the energy balance equation for the earth's surface, the sum of sensible and latent heat flux densities (H and LE respectively) is given by

$$H + LE = Q^* - G, \tag{1}$$

where Q^* is the net radiant flux density, generally denoted as net radiation, G the soil heat flux density, L the latent heat of vaporization and E the evaporation. For a land surface, G is mostly small with respect to Q^* during daytime. A good estimate for G is (e.g., Burridge and Gadd, 1977):

$$G = 0.1Q^*. \tag{2}$$

It is noted that G can also be determined using techniques which contain more physics. An example is the method of Deardorff (1978). However, for many practical calculations these approaches are too complicated, while often the necessary information, e.g., the soil type, is missing. Therefore, we decided to rely on the empirical expression (2). An experimental verification of (2) will be given in Section 5.

Since net radiation can be evaluated from cloud cover (or duration of bright sunshine), air temperature and global radiation using semi-empirical relations (see Appendix B), our problem is reduced to the determination of the partitioning of the available energy ($Q^* - G$) into sensible and latent heat.

b. The Penman–Monteith model

The most complete expression for the partitioning of ($Q^* - G$) into H and LE is Penman's equation applied to a cropped surface as done, for example, by Monteith (1965) and Rijtema (1965), resulting in

$$LE = \frac{s(Q^* - G) + (\rho c_p \delta e / r_a)}{s + \gamma[1 + (r_c/r_a)]}, \tag{3}$$

where s is the slope of the saturation-vapor-pressure temperature curve, ρ and c_p the density and specific heat of air at constant pressure, γ the psychrometric constant, r_a the aerodynamic resistance for sensible heat (and water vapor) of the air layer between the ground and the height of observation z, r_c the surface resistance, and δe the saturation deficit at z defined by

$$\delta e = e_s(T_a) - e_a, \tag{4}$$

where $e_s(T_a)$ is the saturation vapor pressure at air temperature T_a, and e_a the actual vapor pressure at z.

Eq. (3) applies to extensive areas covered with a uniform vegetation fully shading the ground. Then the surface resistance is mainly determined by physiological factors, except when the foliage is wet. In that case, $r_c = 0$. A detailed survey on the different features of Eq. (3) has been given recently by Monteith (1981). The counterpart of (3) for sensible heat is

$$H = \frac{\gamma[1 + (r_c/r_a)](Q^* - G) - (\rho c_p \delta e / r_a)}{s + \gamma[1 + (r_c/r_a)]}. \tag{5}$$

From (1) and (3) it follows that the surface resistance r_c is given by

$$r_c = \left(\frac{s}{\gamma} B - 1\right) r_a + \frac{\rho c_p}{\gamma} \frac{\delta e}{(Q^* - G)} (1 + B), \tag{6}$$

where B is the Bowen ratio, H/LE. With this relation, r_c can be determined from micrometeorological observations. Then the aerodynamic resistance r_a must be specified. This can be done by applying the Monin–Obukhov similarity theory (e.g., Businger, 1973). However, this requires a rather complicated iteration scheme. Therefore, we decided to use the semi-empirical expression proposed by Thom and Oliver (1977), notably

$$r_a = \frac{4.72[\ln(z/z_0)]^2}{1 + 0.54u}, \tag{7}$$

where u is the wind speed, z the height and z_0 the surface roughness length for momentum. Thom and

Oliver showed that in this expression, which is a modification of a relation given before by Penman (1948), stability effects on r_a are taken into account empirically.[1] In the deduction of (7) it is assumed that the "surface roughness lengths" for heat and water vapor are equal to z_0. This is questionable, but the errors introduced that way are small because the Bowen ratio often is of the order γs^{-1} (see below). Then the first term in the right-hand side of (6) can be ignored and r_c is independent of r_a.

For our purposes Eqs. (3) and (5) are rather inconvenient, since they contain a relatively large number of variables. Therefore, it is worthwhile to search for a simplification of the Penman–Monteith formula. Before doing so, it must be emphasized that although this formula has some empirical elements (notably the determination of r_c is fully empirical), it is a physical reality that the surface fluxes depend on so many variables. The only chance we have to arrive at a simplification of Eqs. (3) or (5) is that some variables are either dominating, are interrelated or have a fairly narrow range of values in practice.

c. *The modified Priestley–Taylor model*[2]

The best-known simplification of Penman's formula is the concept of Priestley and Taylor (1972), who found for saturated surfaces[3] that

$$LE = \alpha \frac{s}{s+\gamma}(Q^* - G), \qquad (8)$$

where α is the so-called Priestley–Taylor parameter. When saturated air passes over a wet surface, α will reach 1 [this was noted by Schmidt (1915) and follows directly from Eq. (3) with $r_c = \delta e = 0$]. However, air seldom is saturated and for saturated surfaces α is found to be ~30% greater than 1, when daily values are concerned (e.g., Priestley and Taylor, 1972; Brutsaert and Stricker, 1979). Eqs. (1) and (8) lead to

$$H = \frac{(1-\alpha)s + \gamma}{s+\gamma}(Q^* - G). \qquad (9)$$

In the moderate climatological regions, $LE > H$, which implies that a relatively small deviation of the Priestley–Taylor concept causes a relatively large error in H. Evidence given by, e.g., de Bruin and Keijman (1979) reveals that a two-parameter model of the type

$$LE = \alpha' \frac{s}{s+\gamma}(Q^* - G) + \beta \qquad (10)$$

is a somewhat better description of LE, where β is a small constant (see below). Then H is given by

$$H = \frac{(1-\alpha')s + \gamma}{s+\gamma}(Q^* - G) - \beta. \qquad (11)$$

As noticed before, the Priestley–Taylor concept originally was restricted to saturated surfaces. In our study it will be applied also to non-saturated cases. Then the parameters α, α' and β will depend on the soil moisture conditions, the soil type, and other factors. Herein we follow authors like Davies and Allen (1973).

We found that the Priestley–Taylor approach becomes clearer by writing

$$\beta = \left\{\frac{\rho c_p}{\gamma}\frac{\delta e}{r_c} - \left[\alpha' - (1-\alpha')\frac{r_a}{r_c}\frac{s+\gamma}{\gamma}\right] \times \frac{s}{s+\gamma}(Q^* - G)\right\}\left[\frac{s}{\gamma}\frac{r_a}{r_c} + \frac{r_a}{r_c} + 1\right]^{-1}, \qquad (12)$$

which follows directly form Eqs. (3) and (10). The (modified) Priestley–Taylor formula can only yield useful results if for each value of r_c one can find an α', so that on the average β is small with respect to the first term on the rhs of Eq. (10). Then a variation in β of, say, a factor of 2, which certainly must be expected when we look to Eq. (12), results in a much smaller change of LE.

According to Eq. (12), β depends on several independent variables such as δe, r_a, r_c and Q^*. Therefore, it is not possible to give *a priori* reasons why β is small or not. Mathematically, it follows from Eq. (12) that β is small if the two terms in the numerator of (12) have approximately the same magnitude, i.e., if

$$\frac{\rho c_p \delta e}{\gamma r_c} \approx \left[\alpha' - (1-\alpha')\frac{r_a}{r_c}\frac{s+\gamma}{\gamma}\right] \times \frac{s}{s+\gamma}(Q^* - G). \qquad (13)$$

Among other things, (13) can hold only when δe and $[s/(s+\gamma)](Q^* - G)$ are positively correlated. (The latter quantity is often denoted as the *equilibrium* latent heat flux density LE_{EQ}. This will be used in the following).

Now, we notice that both δe and LE_{EQ} have a pronounced diurnal variation. That of δe is mainly due to the diurnal cycle of the air temperature. To illustrate this, δe and LE_{EQ} are shown in Fig. 1 as a function of time on a clear day in June 1977. The consequence of this effect is that δe and LE_{EQ} will be positively correlated, which is one of the conditions for the validity of (13). In Section 5 more detailed statistics concerning (13) will be given.

When daily values are considered, δe and LE_{EQ} can

[1] Eq. (7) suggests that z can be varied. However, since Penman's wind function is valid only for $z = 2$ m, Eq. (7) can only be applied to this height.

[2] A water surface or a land surface covered with a thin water layer.

[3] This refers to water surfaces as well as to moist cropped land surfaces.

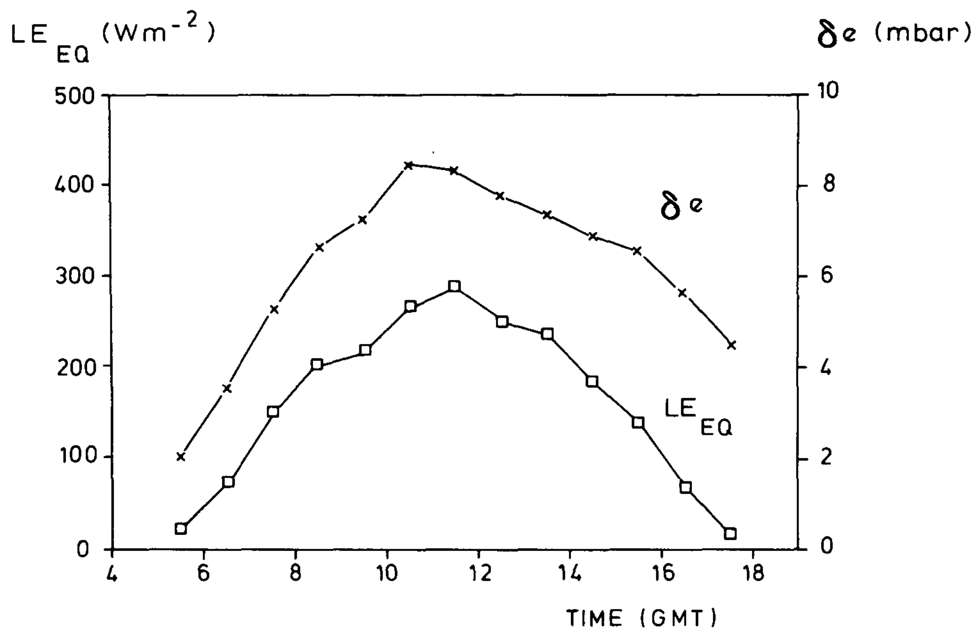

FIG. 1. Diurnal cycle of the water vapor deficit δe and the *equilibrium* latent heat flux density LE_{EQ} for a day in June 1977.

be correlated because both show a pronounced annual cycle.

4. Model comparison

From the above description of the two models under consideration, it follows that both contain at least one parameter which is determined by soil and plant factors. These are r_c in the Penman–Monteith model and α, α' and β in the (modified) Priestley–Taylor approach. Generally, these quantities depend upon many factors, such as availability of soil moisture, soil type, stage of development of the canopy, carbon dioxide concentration, irradiance, etc.[4] Models have been developed to describe, e.g., r_c as a function of these physiological and environmental factors (e.g., Rijtema, 1965), but for the practical problems we are dealing with these procedures are not useful, not least because the necessary information about canopy and soil is often missing. Therefore, in reality the best we can do is to choose appropriate values for r_c, α, etc., under different classes of circumstances, e.g., *wet*, *normal* and *dry* conditions. Our available data set does not cover a range of soil conditions wide enough to define these classes very precisely. But, since our purpose is to compare several models, here this is not a serious problem.

It appears that during the period 3–17 July there was a significant shortage of soil moisture. In the following, this period will be denoted as *dry*, while the remaining days will be characterized as *normal.*

For the model comparison the following strategy is chosen. First, the model parameters will be adjusted for the *dry* and *normal* periods, then the surface fluxes will be evaluated with the different models using these parameters. Finally, a comparison will be made with the measured fluxes.

5. Results

First, we will investigate the skill of Eq. (2) for the determination of the soil heat flux density (G). In Fig. 2, the hourly values of G, determined with the procedure described in Appendix A, are plotted against the corresponding observations of the net radiation Q^*. A visual inspection of this figure reveals that, on the average, Eq. (2) is a rather good approximation, but there is a large random scatter. From a statistical analysis it follows that $\overline{Q^*} = 196$ W m^{-2} and $\bar{G} = 16$ W m^{-2} (an overbar denotes a mean value of the entire data set), the correlation coefficient is 0.85, while for the regression parameter a from $G = aQ^*$, a value of 0.09 is found. In practice, this does not significantly differ from $a = 0.1$, used by Burridge and Gadd (1977). Taking $a = 0.1$, a standard error for G of 9 W m^{-2} is obtained. This is more than 50% of $\bar{G}$; however, it is 5% of $\overline{Q^* - G}$, which is an acceptable scatter for our practical calculations.

In order to get an impression of the variability of the model parameters r_c, α and α' from Eqs. (6), (8)

[4] For a recent review on the dependence of r_c on these factors, see Ziemer (1979).

and (10), respectively, and in order to be able to adjust these quantities for the *normal* and *wet* periods in Fig. 3, the daytime mean hourly values of r_c, α and α' are shown. Only those days were considered from which a complete data set of at least four hours was available.

From Fig. 3a, it is seen that r_c has a considerable variation; it varies between, say, 20 and 280 s m^{-1}. In the normal period it has a typical value of 60 s m^{-1}, while in the dry July period it rises continuously from 100 to 260 s m^{-1} with a mean of $\sim$160 s m^{-1}.

Also, the parameters α and α' (Figs. 3b and c) show a considerable scatter; α varies between 0.6 and 1.5 with a mean of $\sim$1.12 in the *normal* and 0.8 in the *dry* period. The parameter α', from the modified Priestley–Taylor formula (10), shows a very similar behavior. Its mean value in *normal* and *dry* periods is $\sim$0.95 and 0.65 respectively. It is noted that α' is evaluated after β is chosen at 20 W m^{-2} for the entire data set. This mean value of β is visually obtained by plotting the observed values of LE against LE_{EQ}. Having obtained these results, it was decided to choose the following values for r_c, α and α' to be used in the model comparisons:

a) *normal* period: r_c = 60 s m^{-1}, α = 1.12 and α' = 0.95.

b) *dry* period: r_c = 160 s m^{-1}, α = 0.8 and α' = 0.65.

As noted before, parameter β is kept at 20 W m^{-2} for the entire period.

Using these values, the surface fluxes were evaluated with the Penman–Monteith model and the modified and unmodified Priestley–Taylor formula. For LE, some results are shown in Fig. 4. In Table 1, statistical information is given about the skill of the models, referring to the entire data set and including also the sensible heat flux. From the evidence given in Fig. 4 and Table 1, it can be seen that the skill of the modified Priestley–Taylor model is certainly as good as that of the Penman–Monteith equation. This leads to the conclusion that for many practical problems, the modified Priestley–Taylor model must be

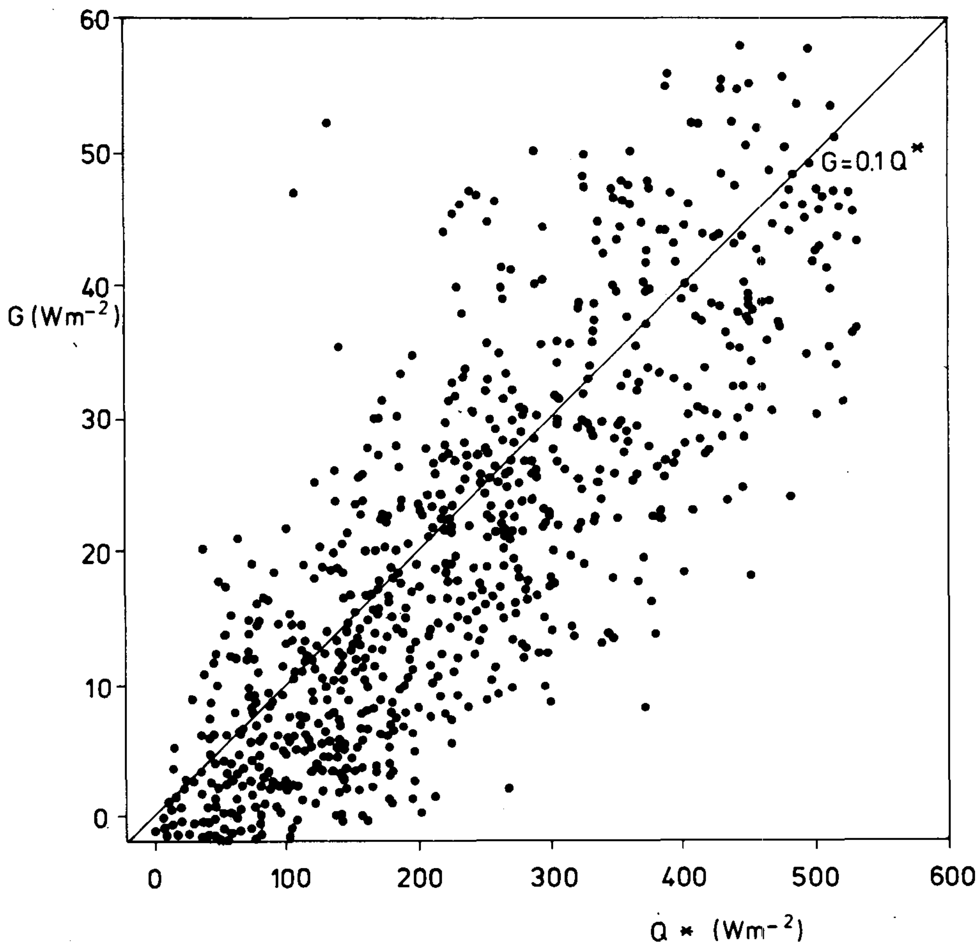

FIG. 2. Soil heat flux density G, plotted against net radiation for 1040 hourly values during daytime.

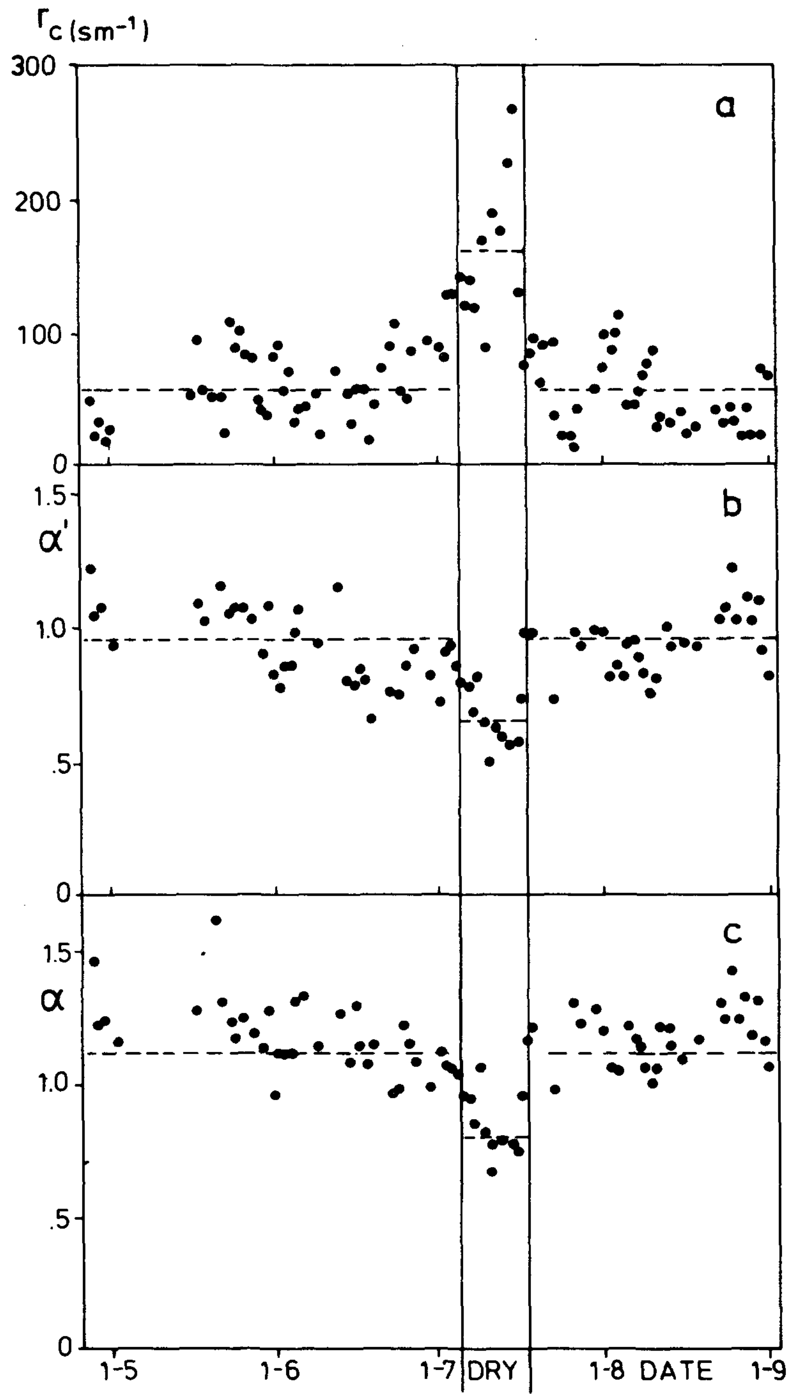

FIG. 3. Daytime averages of: (a) the surface resistance r_c of Eq. (15); (b) the modified Priestley-Taylor parameter α' of Eq. (8); (c) The Priestley-Taylor parameter α of Eq. (6).

preferred. We recall that our measurements refer to predominantly unstable conditions with no advection.

As to be expected, the results of both models for evaporation are better than for the sensible heat flux, mainly due to the fact that mostly LE is greater than H during daytime. From Table 1, it is seen that the modification of the Priestley–Taylor model by adding the constant $\beta = 20$ W m^{-2} improves the skill, especially with respect to H. This can partly be explained by the fact that during the transition hours around sunrise and sunset, H and $(Q^* - G)$ do not change

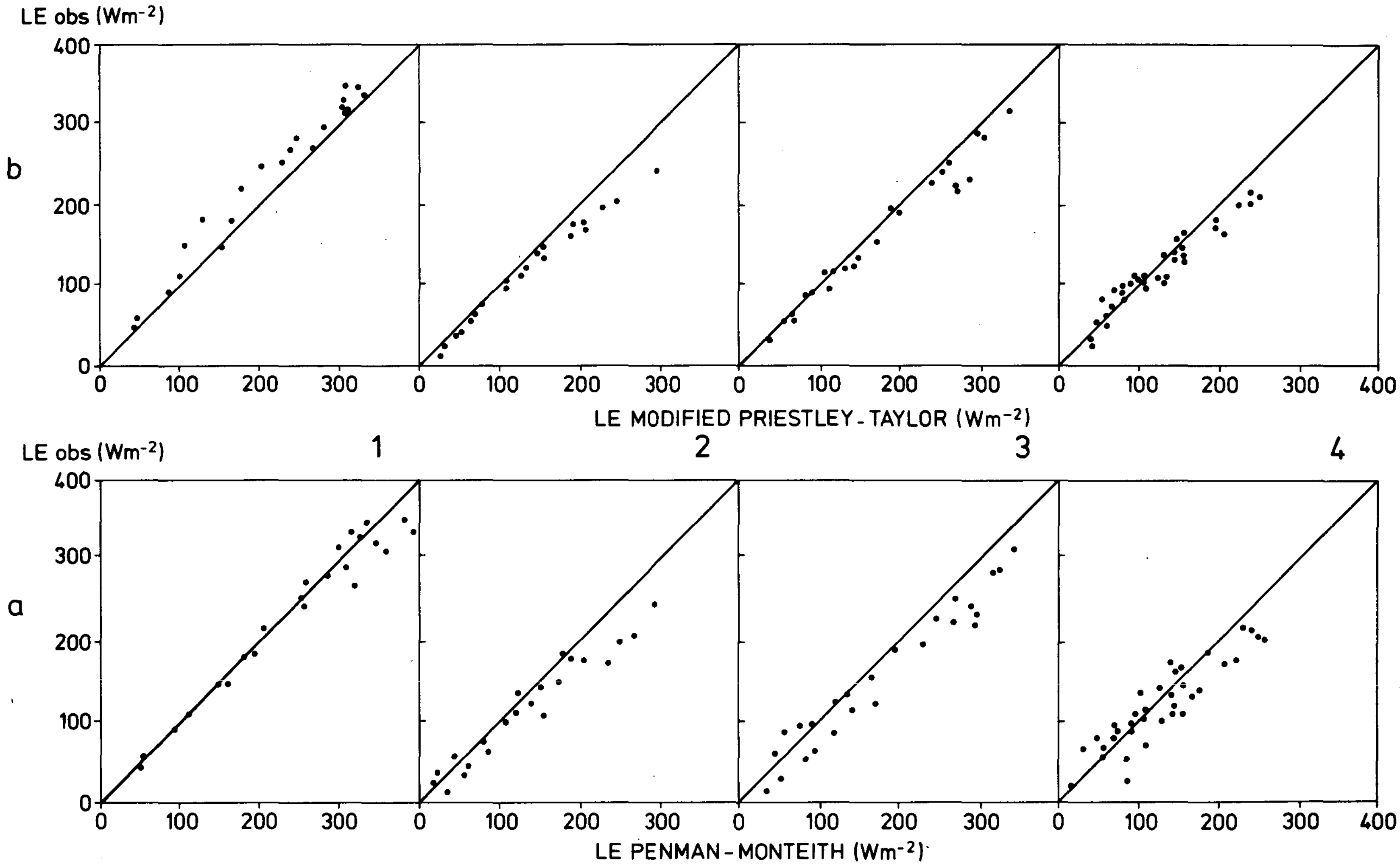

FIG. 4. Comparison of observed hourly averages of the latent heat flux (*LE* obs) with (a) The Penman-Monteith concept of Eq. (3); (b) the modified Priestley-Taylor model of Eq. (8) for the following days: 1) 19 and 25 May (normal period), 2) 4 and 30 June (normal period), 3) 3 and 22 July (normal period), 4) 9, 10 and 16 July (dry period).

TABLE 1. Comparison of observed values (Y) and calculated values (X) of the fluxes of sensible (H) and latent heat (LE) for 1040 hours with H, $LE > 0$. $\bar{Y}$, $\bar{X}$ are the average values of Y, X respectively; r is the correlation coefficient between Y and X, SE is $[\overline{(X-Y)^2}]^{1/2}$; H_1, LE_1 refer to the unmodified Priestley–Taylor model, with $\alpha = 1.12$ for the *normal*, and $\alpha = 0.8$ for the *dry* period. H_2, LE_2 refer to the modified Priestley-Taylor model with $\alpha' = 0.95$ for the *normal* period and $\alpha' = 0.65$ for the *dry* period and $\beta = 20$ W m^{-2} for the entire period. H_3, LE_3 refer to the Penman–Monteith model with $r_c = 60$ s m^{-1} for the *normal* period and $r_c = 160$ s m^{-1} for the *dry* period.

Quantity	$\bar{Y}$	$\bar{X}$	r	SE	$SE/\bar{Y}$
LE_1	125.4	125.1	0.97	23.8	0.19
LE_2	125.4	125.3	0.97	21.2	0.17
LE_3	125.4	126.2	0.95	26.3	0.21
H_1	55.3	55.5	0.92	23.8	0.43
H_2	55.3	55.3	0.92	21.2	0.38
H_3	55.3	54.4	0.90	26.3	0.48

sign at the same time. Especially at the end of the day, it is often observed that H becomes zero earlier than $(Q^* - G)$.

In Section 3c, it is noticed that the good skill of the Priestley–Taylor model can be explained by the two terms of (13) being positively correlated. That this is the case is shown in Fig. 5, where a scatter diagram between these two terms is given. The correlation coefficient is found to be 0.65. There is no direct physical reason for this relatively high value. It is mainly due to the fact that δe and $(s/s + \gamma)(Q^* - G)$ both have a diurnal (and also an annual) cycle. The correlation coefficient between these quantities appears to be 0.7. From Fig. 5 it is further seen that the first term of (13) mostly exceeds the second, resulting in a mean value of β greater than zero. The scatter present in Fig. 5 illustrates the statistical character of the Priestley–Taylor model; in reality β has a relatively large variation. When α' is fixed at 0.95 in the *normal* and at 0.65 in the *dry* period, it appears that β for our data set can vary between -10 and $+100$ W m^{-2}. Excluding the extreme values, we obtain $\bar{\beta} = 23$ W m^{-2} with approximately the same standard deviation.

The final step which must be made, but which falls outside the scope of this paper, is the determination of the net radiation. The way in which this must be done depends upon the available data. As an example we present the results of an estimation scheme de-

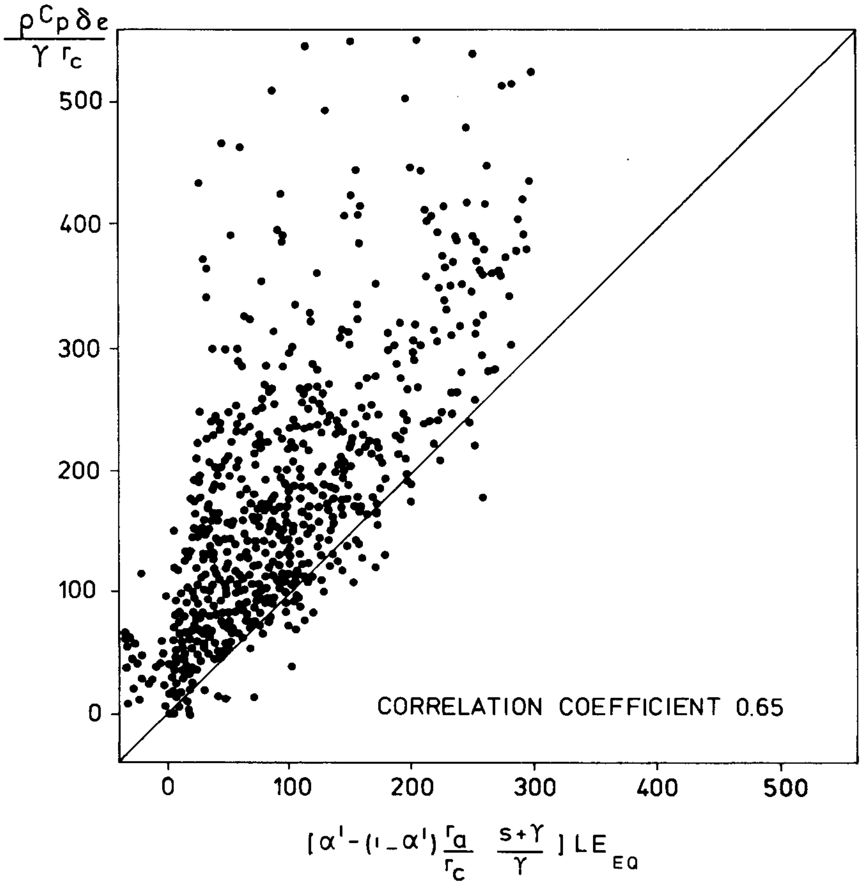

FIG. 5. Scatter diagram of the two terms of (13) for 1040 hourly values during daytime (both in W m^{-2}).

veloped by Holtslag *et al.* (1981), described in Appendix B. Further using Eq. (2), β = 20 W m^{-2} and α' = 0.95 in the *normal* and α' = 0.65 in the *dry* period, H is evaluated with Eq. (11) for 152 randomly selected hours. The results are shown in Fig. 6. It is seen that there is good agreement with the measured values.

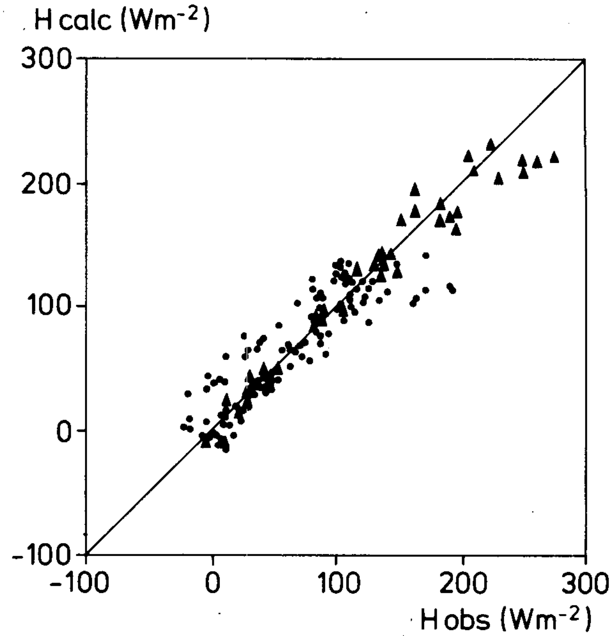

FIG. 6. Comparison of observed hourly averages of the sensible heat flux (H_{obs}) and calculated values (H_{calc}) with Eqs. (9), (2) and a semi-empirical procedure for the net radiation taken from Holtslag *et al.* (1980). Dots represent hours in normal periods and triangles represent hours in the apparently dry fortnight of July.

6. Discussion

From this evidence, it can be concluded that from a practical point of view the modified model of Priestley–Taylor has approximately the same skill as the more complete, but also more complicated, model of Penman–Monteith. Our results refer to predominantly unstable and advection-free conditions and to a short grass cover. We arrived at these results after adjustment of the model parameters which depend on soil and plant factors. The available data set does not allow us to develop complete empirical rules for the determination of these parameters under different soil moisture conditions. In what is denoted in this paper as a *normal* period, the parameter α' from the modified Priestley–Taylor model appears to be ~1 (0.95 to be more precise), while it equals ~0.65 in the so-called *dry* period. The best estimate of β appears to be 20 W m^{-2} during the entire period.

An explanation for the fact that the Penman–Monteith model can be simplified to the modified Priestley–Taylor formula is the fact that the water vapor deficit δe is positively correlated with LE_{EQ}. This correlation is mainly based on the diurnal cycle of both quantities. As a result, the two terms in the numerator of Eq. (12) are counteractive, by which on the average β is small and H and LE become insensitive to variations of δe, r_c and r_a (and thus to variations of wind speed).

We have seen that in the *normal* period α' is found to be 1 and $\beta \approx$ 20 W m^{-2}. On a clear summer day, LE is of the order of several hundreds of W m^{-2}, so that then β can be neglected, by which

$$LE \approx LE_{EQ}, \quad (14)$$

and consequently the Bowen ratio is given by

$$B = \gamma/s. \quad (15)$$

As noted before, this result is also obtained when saturated air passes over a wet surface. On the other hand, in the literature experimental evidence is often reported for (14) for nonsaturated circumstances, but this refers mostly to daily mean values. A review of these observations is given by McNaughton (1976). This author theoretically deduced Eq. (15) for the case of a sudden step in surface wetness, assuming that the reciprocal eddy diffusivity becomes infinitely large with height.

In this study it is shown clearly that $\alpha' \approx 1$ on the average, but there is a large random scatter, while, furthermore, (14) and (15) are not valid in the *dry* July period.

Nevertheless, there remain some interesting questions: Is there a tendency for (short) crops with sufficient water to evaporate at their equilibrium rate LE_{EQ}? Is there a physical or physiological reason for this? If there is a physical cause for (14) and (15), then, in our opinion, this must be searched for in the mechanism of turbulent exchange of heat and mass. It is known that in unstable air, the vertical transfer of heat, mass and momentum is maintained by so-called *convective plumes* (e.g., Kaimal and Businger, 1970). Possibly, within these plumes, the partitioning of the available energy at the surface ($Q^* - G$) into H and LE is described by (15). This can be investigated by studying the behavior of H and E within a plume. This can be done by measuring H and E with eddy-correlation techniques.

It was previously noted that for so-called saturated surfaces, the parameter α from the unmodified Priestley–Taylor model is often found to be 1.26 for daily means of LE. This is not in contradiction with our $\alpha \approx 1.12$ obtained for daytime hourly values. In particular, when it is assumed that during the nighttime, LE is negligibly small, while ($Q^* - G$) is significantly less than zero, it can be shown that the daily mean of α is ~10% greater than its hourly value during daytime. Indeed, it is found that α = 1.26 yields good results for daily values in the summer months.

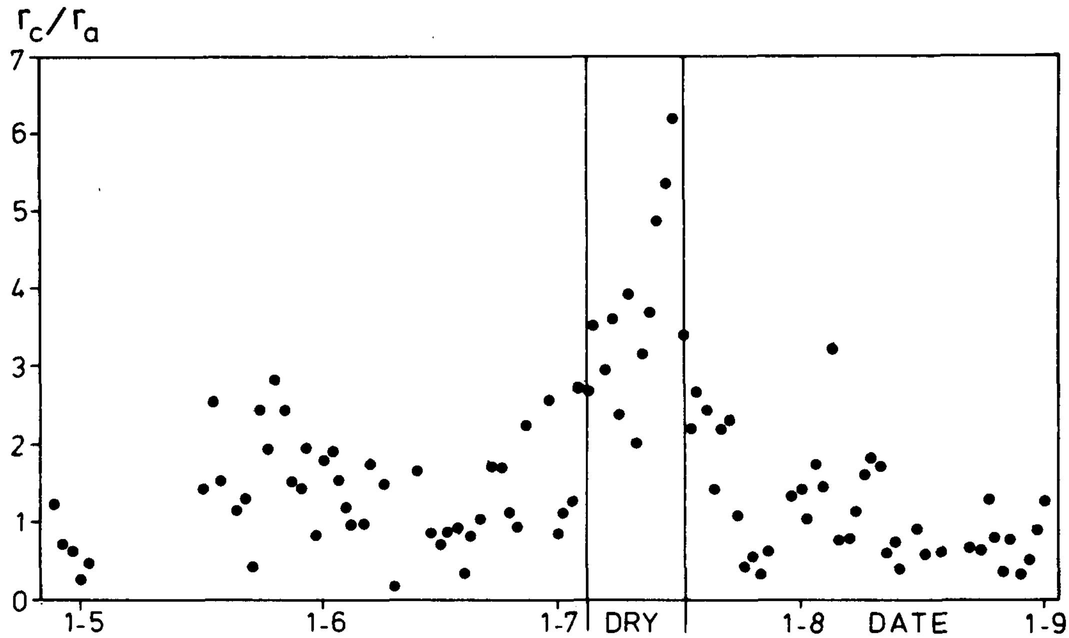

FIG. 7. Daytime averages of the surface aerodynamic resistance ratio r_c/r_a.

Our results apply to a short vegetation for which the aerodynamic and the surface resistance are of the same order. This is illustrated in Fig. 7 where the ratio r_c/r_a is given. Except for the dry July period, r_c/r_a is of the order 1.5, which is in good agreement with the value found by Thom and Oliver (1977) for a similar surface in England. In the case of a tall vegetation with a dry foliage, r_c/r_a is much larger since r_a then is small, due to the great surface roughness. In the limit $r_a \rightarrow 0$, LE reaches (Thom and Oliver, 1977)

$$LE = \frac{\rho c_p}{\gamma} \frac{\delta e}{r_c}, \quad (16)$$

so that LE does not explicitly depend anymore upon $(Q^* - G)$. Thus, for tall vegetation, e.g., forests, it is not to be expected that the Priestley–Taylor model is applicable. Nevertheless, the correlation between δe and $(s/s + \gamma)(Q^* - G)$ also remains in the case of a forest. Therefore, the Priestley–Taylor concept may still be useful. But then one should account for the contribution of evaporation of intercepted water to the total evaporative losses (Shuttleworth and Calder, 1979).

Finally, it must be emphasized that our results refer to the summer months. Then $(Q^* - G)$ is relatively large. In winter, this is no longer true and it must be expected that then the (modified) Priestley–Taylor model will yield unrealistic results.

7. Conclusions

It is shown that, for short vegetation, the hourly fluxes of latent and sensible heat can appropriately be described by a modified Priestley–Taylor model, Eqs. (10) and (11). It has the same skill as the Penman–Monteith equation, which is more complete from a physical point of view, but which requires more input data. An important reason why the Penman–Monteith equation can be simplified to the Priestley–Taylor formula is that the saturation deficit and the equilibrium latent heat flux density (LE_{EQ}) are correlated, due to their both having a diurnal cycle. It should be noted that our results refer to conditions of low advection and an unstable stratification ($0 > z/L_0 > -0.3$). The experiments were performed above a short grass cover. Under different environmental conditions the modified Priestley–Taylor model needs further verification.

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APPENDIX A

The Determination of the Soil Heat Flux

The soil heat flux density G is described by

$$G(z) = -\lambda \frac{\partial T}{\partial z}, \quad (A1)$$

where λ is the thermal conductivity of the soil, T is the soil temperature and z is the depth. In our experimental set-up, G is simultaneously measured at

5 and 10 cm with the soil temperature at 0 and 2 cm in the ground.

The soil heat flux at the surface ($z = 0$) must be evaluated. For this, W. H. Slob (personal communication, 1982) uses the following procedure: It is assumed that $\partial T/\partial z$ at the surface can be approximated by

$$\left(\frac{\partial T}{\partial z}\right)_s = \frac{\Delta T}{\Delta z}, \quad \text{(A2)}$$

where ΔT is the difference between the soil temperature at 2 and 0 cm and $\Delta z = 2$ cm. Furthermore, the assumption is made that λ in the top layer of 2 cm is constant during a period of approximately one day from 0400 GMT till 0400 GMT the next day. The unknown value of λ during the period is obtained as follows: First, the quantity $I(z)$ is introduced, which is defined by

$$I(z) = \int_{t_0}^{t_c+\tau} \times \left| G(z, t) - \frac{G(z, t_0) - G(z, t_0 + \tau)}{\tau}(t - t_0)\right| dt, \quad \text{(A3)}$$

where t is the time, $t_0 = 0400$ GMT and $\tau = 1$ day. It is seen that $I(z)$ is the time integral of the absolute value of $G(z)$ minus the trend over 1 day. It is a measure of the diurnal amplitude of the soil heat flux density.

Now it is assumed that $I(z)$ decreases exponentially with depth. This is the case when $\lambda/(\rho_s c)$, where ρ_s is the density, and c the specific heat capacity of the soil, is constant with depth and when there is no trend. This leads to

$$\frac{I(0)}{I(5)} = \frac{I(5)}{I(10)}, \quad \text{(A4)}$$

where $I(5)$ and $I(10)$ are the values of I at 5 and 10 cm, respectively. The latter can be evaluated from the available data and thus $I(0)$ can be calculated.

Since λ is assumed to be constant over τ, it can be obtained from Eqs. (A1)–(A3) by taking $z = 0$. This yields

$$\lambda = -I(0)\Delta z\left[\int_{t_0}^{t_0+\tau} \times \left| \Delta T(t) - \frac{\Delta T(t_0) - \Delta T(t_0 + \tau)}{\tau}(t - t_0)\right| dt\right]^{-1}. \quad \text{(A5)}$$

Then G follows from (A1).

APPENDIX B

Estimation of Net Radiation from Standard Meteorological Data

Here, a brief description is given of the procedure of Holtslag *et al.* (1981), to estimate the net radiation Q^* from standard meteorological data. Therefore, Q^* is written as

$$Q^* = (1 - A)K^+ + L^+ - L^-, \quad \text{(B1)}$$

where A is the albedo of the surface, K^+ is the incoming shortwave radiation, L^+ is the incoming longwave radiation from the atmosphere and L^- is the outgoing longwave radiation from the surface. Here, A is fixed at $A = 0.25$, which is a typical value for short grass.

The shortwave radiation K^+ is measured on a routine basis at seven locations in The Netherlands. Therefore, in this paper, measurements of K^+ are used. When K^+ is not measured, it can be estimated from solar elevation and total cloud cover (Holtslag *et al.*, 1981; Kasten and Czeplak, 1980).

For the incoming atmospheric radiation L^+, we use (Paltridge and Platt, 1976)

$$L^+ = 5.31 \times 10^{-13}T^6 + 60N, \quad \text{(B2)}$$

where T is the absolute air temperature at screen height and N is total cloud cover.

The outgoing longwave radiation from the surface, L^-, is parameterized by (Holtslag *et al.*, 1981)

$$L^- = \sigma T^4 + c(1 - A)K^+, \quad \text{(B3)}$$

where σ is the Stefan-Boltzmann constant and c is a constant taken at $c = 0.07$. The emissivity of the surface is taken as 1, which is a good approximation for natural surfaces (Sellers, 1965). The second term of the rhs of (B3) accounts approximately for the difference between the surface temperature and the air temperature at screen height T.

REFERENCES

Brutsaert, W., and H. Stricker, 1979: An advection-aridity approval to estimate actual regional evapotranspiration. *Water Resour. Res.*, **15**, 443–450.

Burridge, D. M., and A. J. Gadd, 1977: The Meteorological Office Operational 10-level numerical weather prediction model. Sci. Pap. 34, U.K. Meteorological Office.

Businger, J. A., 1973: Turbulent transfer in the atmospheric surface layer. *Workshop on Micrometeorology*, D. A. Haugen, Ed., Amer. Meteor. Soc., 67–101.

Davies, J. A., and C. D. Allen, 1973: Equilibrium, potential and actual evaporation from cropped surfaces in southern Ontario. *J. Appl. Meteor.*, **12**, 649–657.

Deardorff, J. W., 1978: Efficient prediction of ground surface temperature and moisture, with inclusion of a layer of vegetation. *J. Geophys. Res.*, **83**, 1889–1903.

De Bruin, H. A. R., and J. Q. Keijman, 1979: The Priestley–Taylor evaporation model applied to a large shallow lake in the Netherlands. *J. Appl. Meteor.*, **18**, 898–903.

Dyer, A. J., 1974: A review of flux-profile relationships. *Bound. Layer Meteor.*, **7**, 363–372.

Funk, J. P., 1959: An improved polythene-shielded net radiometer. *J. Sci. Instrum.*, **36**, 267–270.

Holtslag, A. A. M., H. A. R. de Bruin and A. P. van Ulden, 1981: Estimation of the sensible heat flux from standard meteorological data for stability calculations during daytime. *Proc. 11th Int. Tech. Mtg. Air Pollution Modelling and its Applications*, Amsterdam, Plenum.

Kaimal, J. C., and J. A. Businger, 1970: Case study of a convective plume and dust devil. *J. Appl. Meteor.*, **9**, 612–620.

Kasten, F., and G. Czeplak, 1980: Solar and terrestrial radiation dependent on the amount and type of cloud. *Solar Energy*, **24**, 177–189.

McNaughton, K. C., 1976: Evaporation and advection I: Evaporation from extensive homogeneous surfaces. *Quart. J. Roy. Meteor. Soc.*, **102**, 181–191.

Monteith, J. L., 1965: Evaporation and environment. *Symp. Soc. Exp. Biol.*, **XIX**, 205–234.

——, 1981: Evaporation and surface temperature. *Quart. J. Roy. Meteor. Soc.*, **107**, 1–27.

Paltridge, G. W., and C. M. R. Platt, 1976: Radiative processes in meteorology and climatology. *Developments in Atmospheric Science*, Vol. 5, Elsevier, 318 pp.

Penman, H. L., 1948: Natural evaporation from open water, bare soil and grass. *Proc. Roy. Soc. London*, **A193**, 120–145.

Priestley, C. H. B., and R. J. Taylor, 1972: On the assessment of surface heat flux and evaporation using large-scale parameters. *Mon. Wea. Rev.*, **106**, 81–92.

Rijtema, P. E., 1965: An analysis of actual evapotranspiration. *Agric. Res. Rep.*, No. 659, 1–107.

Schmidt, W., 1915: Strahlung und Verdunstung an freien Wasserflächen; ein Beitrag zum Wärmehaushalt des Weltmeers und zum Wasserhaushalt der Erde. *Ann. Hydrog. Marit. Meteor.*, 111–124.

Sellers, W. D., 1965: *Physical Climatology*. University of Chicago Press, 272 pp.

Shuttleworth, W. J., and I. R. Calder, 1979: Has the Priestley–Taylor equation any relevance to forest evaporation? *J. Appl. Meteor.*, **18**, 639–646.

Slob, W. H., 1978: The accuracy of aspiration thermometers. Sci. Rep. 78–1, Roy. Neth. Meteor. Inst., De Bilt, 17 pp.

Tennekes, H., 1973: Similarity laws and scale relations in planetary boundary layers. *Workshop in Micrometeorology*, D. A. Haugen, Ed., Amer. Meteor. Soc., 177–216.

——, 1982: Similarity relations, scaling laws and spectral dynamics. *Atmospheric Turbulence and Air Pollution Modelling*, F. T. M. Nieuwstadt and H. van Dop, Eds., D. Reidel, 37–69.

Thom, A. S., and H. R. Oliver, 1977: On Penman's equation for estimating regional evaporation. *Quart. J. Roy. Meteor. Soc.*, **103**, 345–357.

Ziemer, R. Z., 1979: Evaporation and transpiration. *Rev. Geophys. Space Phys.*, **17**, 1175–1186.