Evapotranspiration of cut over bog covered by *Molinia Caerulea*

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LIST OF SYMBOLS

a	Makkink parameter	(-)
$A_1,...,A_4$	constants for the flux profile function	(-)
b	Makkink parameter	(mmd^{-1})
B	Bowen ratio	(-)
c_p	specific heat at constant pressure	($Jkg^{-1}K^{-1}$)
C_1, C_2, C_3	constants for the σ_θ method	(-)
C_m, C_o, C_w, C_a, C_{soil}	volumetric heat capacity of soil minerals, organic matter, soil moisture, air, and of the weighted average of the soil	($Jm^{-3}K^{-1}$)
d	displacement height	(m)
e	saturation vapour pressure	(mb)
E	evaporation rate	($kgm^{-2}s^{-1}$)
f	crop coefficient	(-)
g	acceleration of gravity	(ms^{-2})
G	soil heat flux density	(Wm^{-2})
H	sensible heat flux density	(Wm^{-2})
H'	factor for the correction for of the soil heat flux plate	
k	Von Kármán constant	(-)
K_d	downward short wave (or global) radiation	(Wm^{-2})
K^*	net short wave radiation	(Wm^{-2})
L	Obukhov length scale	(m)
L_e	heat of vaporization	(Jkg^{-1})
m	ratio of r_a for open water to r_a of the measuring site	(-)
n	ratio of r_a to r_s	(-)
q	specific humidity	(gkg^{-1})
Q^*	net radiation	(Wm^{-2})
r_a	aerodynamic resistance	(sm^{-1})
r_s	bulk stomatal resistance	(sm^{-1})
T	(air) temperature	(K)
T_d	dry bulb temperature	(K)
T_w	wet bulb temperature	(K)
u	wind speed	(ms^{-1})
u_*	friction velocity	(ms^{-1})
z	height above the surface	(m)
z_0	roughness length for momentum	(m)
α	Priestley-Taylor parameter	(-)
γ	psychrometric constant	(mbK^{-1})

Δ	slope of the saturation vapour pressure and temperature	(mbK^{-1})
ε	ratio $\lambda_{plate}/\lambda_{soil}$	(-)
θ	potential temperature	(K)
θ_*	temperature scale	(K)
Θ_m, Θ_o, Θ_w, Θ_a	volume fraction of soil minerals, organic matter, soil moisture and air	(m^3m^{-3})
λ	thermal conductivity	($Wm^{-1}K^{-1}$)
η	ratio of the minor axis to the major axis of the soil heat flux plate	(-)
ρ	density of air	(kgm^{-3})
σ_θ	standard deviation of temperature fluctuations	(K)
τ	shear stress	($kgm^{-1}s^{-2}$)
$\phi_{H,M}$	integrals from level 1 to level 2 of the stability functions for heat and momentum	(-)
$\Psi_{H,M}$	integrated form of the stability functions for heat and momentum	(-)

SUMMARY

During the months May to October in 1988 and in 1989 research is done at Engbertsdijksvenen (The Netherlands) on the evapotranspiration of *Molinia caerulea* growing on cut over bog. In 1988 the flux profile method was used to determine the actual evapotranspiration, and in 1989 three micro-meteorological methods were used to determine the actual evapotranspiration: the Bowen-ratio, the flux profile and the temperature fluctuation method. All three methods use the energy balance to determine the actual evapotranspiration. The data set of the flux profile method had the least amount of data missing. Comparison of the sensible heat flux resulting from different sets of temperature sensors of the flux profile method as well as comparison with the Bowen ratio and the temperature fluctuation method provided evidence of the uncertainty in the flux profile derived sensible heat flux being less than 20%. With the uncertainties in the net radiation estimated at 5% and the soil heat flux density at 20% this results in an uncertainty of daily evapotranspiration of less than 15%. These results support the use of the flux profile data as the actual evapotranspiration data at Engbertsdijksvenen.

Because the groundwater table was relatively high and did not show much fluctuations in 1988 in contrast to 1989 it is taken that in 1988 the actual evapotranspiration is equal to the potential evapotranspiration. To avoid errors due to the evaporation of intercepted precipitation of the vegetation and the layer of litter on the ground, only data with less than 0.2 mm of precipitation were used for the comparison with daily potential evaporation equations. The potential evapotranspiration calculated by the modified Penman equation and the Priestley and Taylor equation was considerably higher than the actual evapotranspiration (flux profile method).

The crop coefficients determined using the Makkink equation to calculate the reference evaporation, resemble on average those of grass taller than 25 cm published by Hooghart and Lablans (1988) (f = 1.1 - 1.2). Although the crop coefficients of *Molinia* tend to be lower (f = 0.8) at the start of the growing season.

The average for dry days during the growing season in the relatively wet year 1988 of the bulk stomatal resistance as defined by Thom and Oliver (1977) is when adjusted for the leaf area comparable (± 65 sm^{-1}) to that of grass (± 65 sm^{-1}) for the Hupselse Beek catchment, the Netherlands. In the dryer year 1989 the average for the growing season of the resistance was somewhat higher, 81 sm^{-1}. When the resistances are not adjusted for the leaf area they are respectively 94 sm^{-1} and 113 sm^{-1}.

1. INTRODUCTION

After the peat mining at the Engbertsdijksvenen (located in the eastern part of the Netherlands) stopped in 1983, the Dutch National Forestry Service started with a program to reestablish the original vegetation dominated by *Sphagnum* species and heather. *Sphagnum* grows at permanently wet sites. However, for the mining of the peat the area was drained, and at present the dominant species is *Molinia caerulea.* One of the policies of the Dutch National Forestry Service is to rise the watertable by damming the drains. This measure should result in the *Sphagnum* species becoming the dominant species of the area again.

To observe the effects of such hydrological measures, it is important to know the different elements of the water balance of the area. For this, a measuring campaign was held from 1987 to 1989. An overview of the data measured and references to the different studies done on these data may be found in Van Amerongen *et al.* (1990). One of the components of the water balance is the evapotranspiration. This was measured in two ways: by the use of micro-lysimeters (Schouwenaars, 1993) and by the use of micro-meteorological methods, which are presented in this report. Some disadvantages of micro-lysimeters in comparison to micro-meteorological methods are: lysimeter measurements only represent a very local condition and the measurements are sensitive for errors with strong fluctuating groundwater tables.

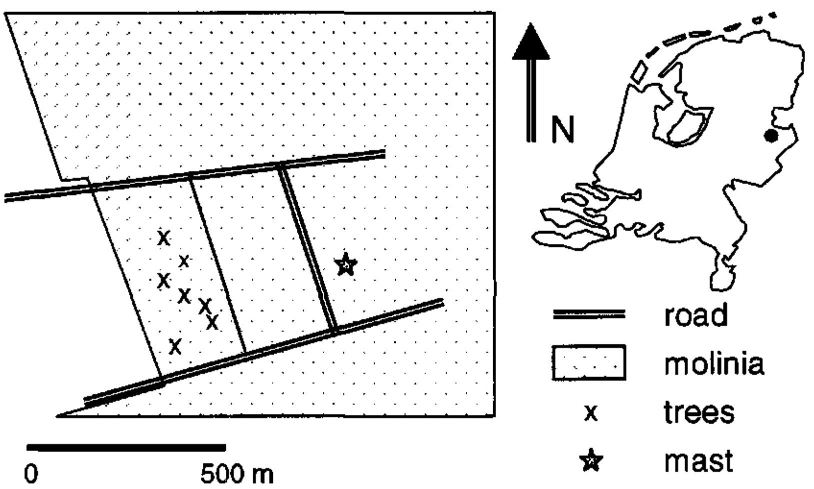

Fig. 1 ***Location of the measuring site.***

The intention of this report is to establish the actual evapotranspiration of *Molinia caerulea* on peat. The results of a measuring campaign in 1988 and 1989 of a site at Engbertsdijksvenen with coordinates 52°28' N and 6°40' E are used (see figure 1). For the actual evapotranspiration rate use is made of the Bowen-ratio, the flux profile, and the temperature fluctuation method. Results of the Bowen ratio method and the temperature method are only available for 1989. A modified form of the Penman equation and the Priestley and Taylor equation are used to see if they give a good estimate of the potential evapotranspiration rate of *Molinia*. The Makkink equation is used to provide reference evapotranspiration values and to produce crop coefficients. Besides crop coefficients to predict the water use of a certain type of vegetation, resistances are also often used. For agricultural crops a lot of research is done on these resistances, but for natural vegetation types such as *Molinia caerulea* there is still a deficiency in knowledge. Some research was done by Duyzer and Bosveld (1988) who found for seven periods divided over different days in the months May and June the surface resistance of *Molinia caerulea* varied between 250 and 400 sm^{-1}. Here we will use a slightly different resistance, the so called bulk stomatal resistance as defined by Thom and Oliver (1977). Resulting values of this resistance at the measuring site will be presented in this report.

To the south east and the north of the measuring site there is a good fetch of ± 400 m, and to the west of ± 150 m. After this the main vegetation remains *Molinia*, but there are a few small trees (birch).

2. PRECIPITATION, WATERTABLE AND LEAF AREA INDEX

In figure 2 the precipitation for the measuring period 23 May to 30 September for the years 1988 and 1989 is depicted. During the measuring period in 1988 87 days with precipitation were registered, accumulating to a total of 394.4 mm with 20 days registering an amount between 0.2 and 0.4 mm d^{-1}. For 1989 a total of 235.3 mm in 58 days was registered, with 20 days having an amount between 0.2 and 0.4 mm d^{-1}. This indicates a dryer growing season in 1989. The daily precipitation data are also shown in the table of appendix B.

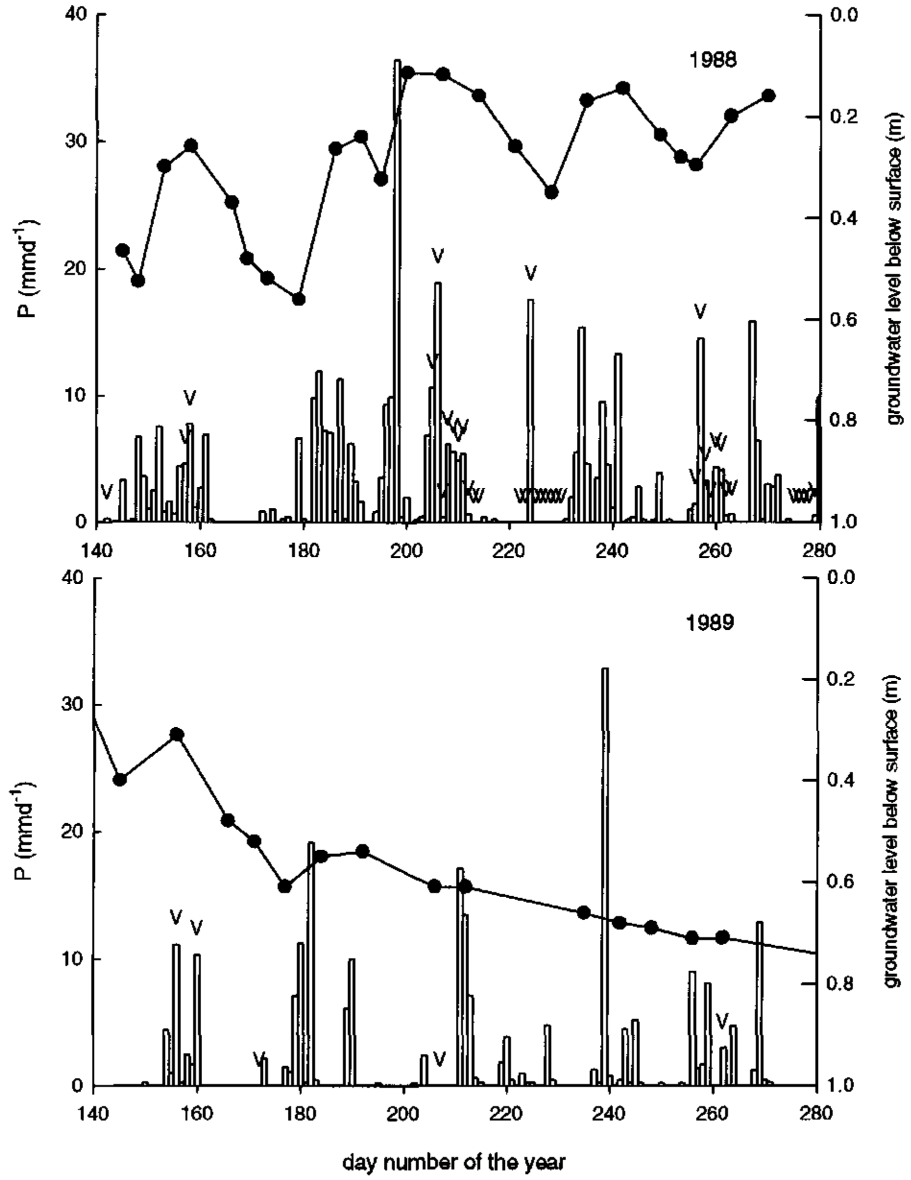

Fig. 2 ***Daily precipitation data 1988 and 1989 represented by the bars (V = missing or incomplete data replaced by data from the KNMI rainfall station Vroomshoop) and groundwater table below the surface of observation well B306 represented by the solid line.***

The relative dryness of 1989 in comparison to 1988 is also demonstrated by the difference in the groundwater table of a monitoring well close to the location of the measuring masts shown in figure 2. Starting at the end of June the groundwater table in 1989 is on average 40 cm below the level of 1988.

In the field the leaf area index *LAI* was only measured in 1988 (Eggink and Vink, 1988). However, the leaf area was measured of the vegetation in the micro-lysimeters for both years (see figure 3). By comparing the *LAI* of 1988 with the *LAI* of 1989 of a lysimeter which represented well the *LAI* of the field in 1988, it was concluded that there was no significant difference in *LAI* for both years (Schouwenaars, pers. comm.).

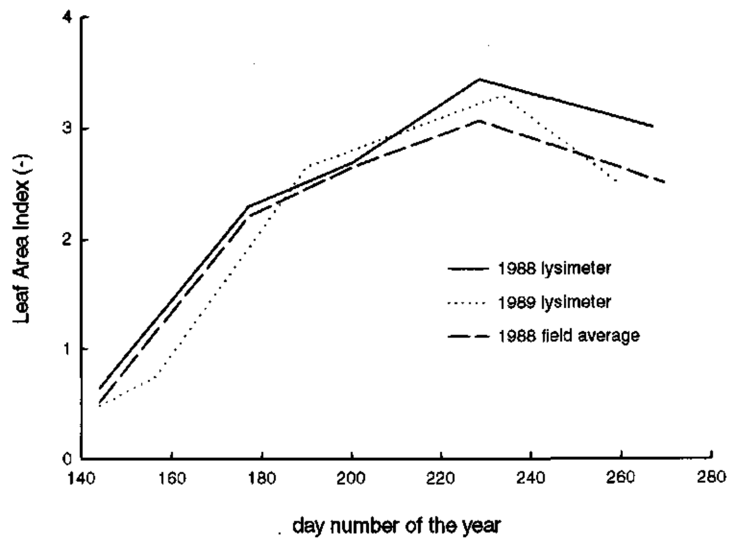

Fig. 3 ***LAI for 1988 as measured in the field and on the lysimeter, and the LAI for 1989 as measured on the same lysimeter.***

3. METHODS

The actual evapotranspiration is calculated with the Bowen ratio, the flux-profile and the temperature-fluctuation method on a 30 minute interval base. The evapotranspiration of a vegetated surface well supplied with water, often called the potential evapotranspiration is calculated on a daily base with a modified Penman and the Priestley and Taylor equation. The reference evapotranspiration is also calculated on a daily basis by the Makkink equation. The Thom and Oliver method is used inversely in combination with the flux profile method to retrieve the daily bulk stomatal resistance.

The energy balance

All three methods used for the derivation of the actual evapotranspiration make use of the energy balance. Both the flux profile and the temperature fluctuation method calculate the sensible heat flux density. The actual evapotranspiration rate is derived as a residue from the energy balance. The Bowen ratio method is a combination equation of the energy balance and the ratio of the sensible and the latent heat flux density.

The energy balance at the surface may be represented by

$$Q^* = L_e E + H + G \qquad (\mathrm{Wm}^{-2}) \tag{1}$$

where Q^* is the net radiation flux density (Wm^{-2}), $L_e E$ the latent heat flux density (Wm^{-2}), H the sensible heat flux density (Wm^{-2}), and G the soil heat flux density (Wm^{-2}). Here L_e is the latent heat of vaporization (Jkg^{-1}) and E the rate of evaporation. ($\mathrm{kgm}^{-2}\mathrm{s}^{-1}$). The energy balance equation is valid for any time interval.

Bowen ratio method

The Bowen ratio B is derived from the energy balance from the underlying surface and is defined as (Brutsaert, 1982):

$$B = \frac{H}{L_e E} = \frac{c_p(\bar{\theta}_2 - \bar{\theta}_1)}{L_e(\bar{q}_2 - \bar{q}_1)} \qquad (-) \tag{2}$$

in which it is assumed that the transfer coefficients of heat and water vapour are equal. Here c_p is the specific heat of air at constant pressure ($\mathrm{Jkg}^{-1}\mathrm{K}^{-1}$), $\bar{\theta}_2$ and $\bar{\theta}_1$ the mean potential temperature (K) at levels 2 and 1, and $\bar{q}_2$ and $\bar{q}_1$ the mean specific humidity (gkg^{-1}) at levels 2 and 1 over time intervals of 30 minutes. The latent heat flux is

$$L_e E = (Q^* - G) \,/\, (1 + B) \qquad (\mathrm{Wm}^{-2}) \tag{3}$$

and the sensible heat flux

$$H = B(Q^* - G) / (1 + B) \qquad (\mathrm{Wm^{-2}}) \tag{4}$$

The method is not applicable if B becomes close to -1, which is often the case during sunset, sunrise and occasionally at night. An other disadvantage of the method is the need of accurate but relatively complicated but accurate measurements of the humidity differences.

Flux profile method

This method may directly provide the latent heat flux, if reliable humidity profiles are measured. However, this is not easy to achieve. To avoid this difficulty, here the profile method is used to calculate the sensible heat flux. The latent heat flux is then determined with the energy budget equation. Thus the difficulties of the measurements of the humidity profiles are avoided, but the uncertainties in the net radiation and the soil heat flux measurements are introduced.

The flux profile method is based on the potential temperature θ (K) and the wind speed u ($\mathrm{ms^{-1}}$) differences

$$\bar{\theta}_2 - \bar{\theta}_1 = \frac{\theta_*}{k}[\Phi_H(\frac{z_2-d}{L}) - \Phi_H(\frac{z_1-d}{L})] \qquad (\mathrm{K}), \tag{5}$$

$$\bar{u}_2 - \bar{u}_1 = \frac{u_*}{k}[\Phi_M(\frac{z_2-d}{L}) - \Phi_M(\frac{z_1-d}{L})] \qquad (\mathrm{ms^{-1}}), \tag{6}$$

where θ_* is the temperature scale (K), u_* is the friction velocity ($\mathrm{ms^{-1}}$), k (= 0.41) the von Karman constant, Φ_H and Φ_M are defined as the integrals from level 1 to level 2 of the stability functions for heat and momentum (-) , z_2 and z_1 the elevation of levels 2 and 1 (m), d is the displacement height (m) and L the stability length (m) as proposed by Monin and Obukhov (Businger and Yaglom, 1971),

$$L = \frac{T}{kg}\frac{u_*^2}{\theta_*} \qquad (\mathrm{m}) \tag{7}$$

with g the acceleration of gravity ($\mathrm{ms^{-2}}$) and T the absolute temperature (K). The friction velocity is defined as,

$$u_* = \sqrt{\frac{\tau}{\rho}} \qquad (\mathrm{ms^{-1}}). \tag{8}$$

where ρ is the density of air ($\mathrm{kgm^{-3}}$) and τ is the shear stress at the surface ($\mathrm{kgm^{-1}s^{-2}}$).

The sensible heat flux H is

$$H = -\rho c_p \theta_* u_* \qquad (\mathrm{Wm^{-2}}). \tag{9}$$

With the integrated form of the stability functions for momentum and heat ψ_M and ψ_H (-) proposed by Dyer (Dyer, 1974 and Paulson, 1970) u_* and θ_* can be calculated

$$u_* = k(\bar{u}_2-\bar{u}_1)\ [\ln(\frac{z_2-d}{z_1-d})-\psi_M(\frac{z_2-d}{L})+\psi_M(\frac{z_1-d}{L})]^{-1} \quad (\mathrm{ms^{-1}}), \tag{10}$$

$$\theta_* = k(\theta_2-\theta_1)\ [\ln(\frac{z_2-d}{z_1-d}) - \psi_H(\frac{z_2-d}{L}) + \psi_H(\frac{z_1-d}{L})]^{-1} \quad (\mathrm{K}). \tag{11}$$

For unstable conditions ($(z\text{-}d)/L < 0$) the experimentally determined expressions are:

$$\psi_M = 2\ln(\frac{1+x}{2}) + \ln(\frac{1+x^2}{2}) - 2\arctan(x) + \frac{\pi}{2} \qquad (-), \tag{12}$$

$$\psi_H = 2\ln(\frac{1+x^2}{2}) \qquad (-), \tag{13}$$

where

$$x = (1-16\frac{z-d}{L})^{1/4} \qquad (-). \tag{14}$$

For stable conditions, $(z\text{-}d)/L > 0$, we used the expression found by Holtslag and de Bruin (1988), which is valid for the entire range of stable conditions

$$\psi_M = \psi_H = -A_1\frac{z-d}{L} - A_2(\frac{z-d}{L}-\frac{A_3}{A_4})\exp(-A_4\frac{z-d}{L}) - \frac{A_2A_3}{A_4} \qquad (-), \tag{15}$$

where $A_1 = 0.7$, $A_2 = 0.75$, $A_3 = 5.0$ and $A_4 = 0.35$. With the above flux profile functions for stable and unstable conditions, H can be found iteratively by solving for L, u_* and θ_*.

For instantaneous wind speeds below 0.7 $\mathrm{ms^{-1}}$ the friction velocity is calculated with the wind speed at the highest sensor and a known roughness length z_0 and displacement height d, instead of the wind profile,

$$u_* = k\ \bar{u}_2\ [\ln(\frac{z_2-d}{z_0}) - \psi_M(\frac{z_2-d}{L})]^{-1} \qquad (\mathrm{ms^{-1}}), \tag{16}$$

This was done because at low wind speeds the possible error in the wind speed differences at two levels may cause unacceptable deviations in u_*-values.

Temperature fluctuation method

The temperature fluctuation method is based on an empirical relation between σ_θ/θ_* and $(z\text{-}d)/L$ (Tillman, 1972), where σ_θ is the standard deviation of the temperature of a fast responding sensor:

$$\frac{\sigma_\theta}{\theta_*} = -\frac{C_1}{(C_2 - \frac{z-d}{L})^{1/3}} \qquad (-), \tag{17}$$

and

$$C_2 = (C_1/C_3)^3 \qquad (-). \tag{18}$$

This relation is only valid for dry unstable conditions. In this study C_1 = 0.95 and C_3 = 2.5 are used (de Bruin, 1982). With measurements of the wind speed at one level and a known roughness length and displacement height a first guess of the friction velocity u_* may be calculated with equation (16) for neutral conditions. The sensible heat flux density from eq. (9) is solved by way of iteration between L, u_* and θ_* from eq. (7), (16) and (17).

However, in a comparison study by de Man (1990) between the results from the flux profile method H_{FP} and the temperature fluctuation method H_{TF}, the following regression equation for clear days above grass in the Netherlands was found:

$$H_{TF} = 0.86 H_{FP} + 1.4 \qquad (\mathrm{Wm^{-2}}) \tag{19}$$

De Man attributed the low H_{TF} to a possible too large time constant of the temperature sensor and/or a filtering method for the removal of the effects of trends and low frequencies which is too rough. The temperature sensor we used to obtain the σ_θ values is a 0.12 mm diameter thermocouple with a sampling frequency of 1 Hz. To remove trends and low frequency effects the 4th order differences of the measured temperature are used to calculate the standard deviation. To avoid the potential inaccuracy by introducing eq. (19) in the iteration procedure we used the friction velocity and the stability length calculated by the flux profile method to calculate directly H_{TF} from eq.'s (16), (17) and (9). On the thus obtained H_{TF} we applied the calibration equation (19) as found by de Man.

Penman (modified)

The best known and probably the most widely used evaporation equation, is the one originally derived by Penman (1948). This equation was intended for the calculation of the evaporation of an open water surface where the heat storage of the water body is negligible (i.e. an evaporation pan). The thus defined evaporation is often called the reference evaporation of open water E_o:

$$E_o = \frac{\Delta}{\Delta+\gamma} Q^*/L_e + \frac{\gamma}{\Delta+\gamma} f(\bar{u})(\bar{e}_s - \bar{e}_a) \qquad (\mathrm{mmd^{-1}}) \tag{20}$$

here Q^* is the net radiation above water (Wm^{-2}), Δ is the slope of the saturation vapour pressure (mbK^{-1}), γ is the psychrometric constant (mbK^{-1}), $\bar{e}_s$ is the mean saturation vapour pressure (mb) for the dry bulb temperature at 2 m, $\bar{e}_a$ the mean actual vapour pressure (mb) at 2 m. The wind function for open water is:

$$f(\bar{u}) = 0.26(1 + 0.54\bar{u}) \qquad (mmd^{-1}mb^{-1}) \tag{23}$$

With $\bar{u}$ the average wind speed (ms^{-1}) at 2 m.

The thus obtained reference evaporation of open water is not the potential evaporation of open water, because the heat flux of the water body is not included. To obtain the potential crop evapotranspiration E_0 has to be multiplied by a factor, called crop coefficient.

For this study we used a slightly modified form of the Penman equation,

$$E_{Pen} = \frac{\Delta}{\Delta+\gamma}(Q^*-G)/L_e + \frac{\gamma}{\Delta+\gamma} f(\bar{u})(\bar{e}_s-\bar{e}_a) \qquad (mmd^{-1}) \tag{25}$$

here the net radiation as measured for water is replaced by the net radiation as measured above the vegetation subtracted by the measured soil heat flux. The wind function for open water was maintained. The thus modified Penman equation estimates directly the potential evapotranspiration of a low vegetation (short grass) well supplied with water (Stricker, 1978).

Priestley and Taylor

If the air becomes saturated the second term of the Penman equation tends to become zero and E_0 becomes the so called equilibrium evaporation. However, large scale weather patterns cause the atmospheric boundary layer to maintain a humidity deficit. Priestley and Taylor (1972) found a multiplier of 1.26 (-), called α, to estimate best this humidity deficit. They defined the evaporation of water surfaces and land surfaces well supplied with water under conditions of minimal advection as:

$$E_{PT} = \alpha\frac{\gamma}{\Delta+\gamma}(Q^*-G)/L_e \qquad (mmd^{-1}) \tag{26}$$

The factor α is not a constant and varies, besides the summer season, throughout the year (De Bruin and Keijman, 1979). There are some indications for a slightly larger value of α (Brutsaert, 1982). A value of 1.28 for α is used for this study.

Makkink

An almost similar equation as used by Priestley and Taylor, was proposed by Makkink (1960):

$$E_{Mak} = a\frac{\Delta}{\Delta+\gamma}\,\frac{K_d}{L_e} + b \qquad (mmd^{-1}), \tag{27}$$

where K_d is the downward short wave radiation (or global radiation) in Wm^{-2} and *a* and

b are constants. To obtain the average potential evapotranspiration in the Netherlands for the summer months of a short grass well supplied with water the values a = 0.65 (-) and b = 0.0 (mm d^{-1}) are used. Presently the method is used to obtain reference evapotranspiration values for the Netherlands. The advantage of this method above the Priestley and Taylor method is that the only needed variables, the global radiation and the temperature, are readily available. Furthermore, the value of 0.65 appears to be valid for all but the winter months.

Thom and Oliver

Thom and Oliver (1977) took the Penman-Monteith equation (Monteith, 1973) as a reference but they defined the aerodynamic resistance term in a different way. Monteith introduced the bulk stomatal resistance r_s and the aerodynamic resistance r_a in the Penman equation. It is assumed that the bulk stomatal resistance characterizes the vapour transfer between stomatal cavities and the leaf surface. For the aerodynamic resistance Thom and Oliver maintained the Penman wind function as it corrects implicitly for the wind profile under unstable atmospheric conditions in the surface layer, but they introduced a correction to Penman's equation for the surface roughness of the crop resulting in the following set of equations:

$$E_{TO} = \frac{\Delta}{\Delta+\gamma(1+n)}(Q^*-G)/L_e + \frac{m\gamma}{\Delta+\gamma(1+n)}f(\bar{u})(\bar{e}_s-\bar{e}_a) \qquad (mmd^{-1}) \qquad (28)$$

with n the ratio of the bulk stomatal resistance r_s (sm^{-1}) to the aerodynamic resistance r_a (sm^{-1}), where r_a is

$$r_a = 4.72[\ln(\frac{z-d}{z_0})]^2 \;/\; (1+0.54\bar{u}) \qquad (sm^{-1}) \qquad (29)$$

The ratio m of the r_a of water as used by Penman with a surface roughness length for open water z_{0P} = 0.00137 m to r_a of the measuring site, defined as

$$m = [\ln(\frac{z-d}{z_{0P}}) \;/\; \ln(\frac{z-d}{z_0})]^2 \qquad (-) \qquad (30)$$

Reference, potential evapotranspiration and the crop coefficient

To obtain the potential evapotranspiration E_{pot} of a certain crop use is made of a crop coefficient f and a reference evapotranspiration E_{ref}:

$$E_{pot} = f\,E_{ref} \qquad (mmd^{-1}) \qquad (31)$$

At present the potential evapotranspiration above grass calculated by the Makkink equation is used by the Dutch Royal Meteorological Institute (KNMI) as the reference evapotranspiration on a daily basis. As may be expected f is close to 1.0 for a short grass (5-15 cm), in the months April through August f = 1.0 and in September f = 0.9. For grass with a height of 25 cm or more f varies between 1.2 in the months April to June and 1.1 in the months July and September (Hooghart en Lablans, 1988). Before, the Penman equation for open water was used on a weekly basis as the reference evaporation. It should be kept in mind that the crop factor for the Makkink equation is different from that for the reference open water evaporation (i.e. Penman). For a more detailed discussion of the two methods the reader is referred to (Hooghart and Lablans, 1988).

4. MEASUREMENTS - CORRECTIONS

Table 1 shows the quantities measured and stored on a 30 minute interval base except for the standard deviation of the temperature which was stored on a 10 minute interval base.

Table 1 Equipment used at Engbertsdijksvenen in 1988 and 1989.

1988

MAST A:
- sensor T_{d1}: dry bulb temperature at 1.34 m, 135°; after 1-08-88 at 1.38 m, 225°,
- sensor T_{d2}: dry bulb temperature at 2.54 m, 135°; after 1-08-88 at 2.59 m, 225°,
- sensor T_{w1}: wet/dry bulb temperature at 1.36 m, 225° (natural ventilation),
- sensor T_{w2}: wet/dry bulb temperature at 2.575 m, 225° (natural ventilation); after 1-08-88 out,
- sensor T_{d3}: dry bulb temperature at 1.38 m, 225°; after 1-08-88 out,
- sensor T_{d4}: dry bulb temperature at 2.595 m, 225°; after 1-08-88 out,
- wind speed at 1.66 m, 225°,
- wind speed at 2.89 m, 225°,
- wind direction,

MAST B:
- incoming shortwave radiation at 2 m,
- reflected shortwave radiation at 2 m,
- net radiation at 2m,
- percentage sun at 2m.

Precipitation at 0.90 m, and
Soil heat flux at 0.03 m depth.

1989

MAST A:
- sensor T_{d1}: dry bulb temperature at 1.62 m, 225°,
- sensor T_{d2}: dry bulb temperature at 2.80 m, 225°,
- sensor T_{w1}: wet/dry bulb temperature at 1.60 m, 225° (natural ventilation),
- sensor T_{w2}: wet/dry bulb temperature at 2.78 m, 225° (natural ventilation),
- sensor T_{d3}: dry bulb temperature at 1.65 m, 135°,
- sensor T_{d4}: dry bulb temperature at 2.81 m, 135°; after 22-06-89 out,
- sensor σ_T: temperature fluctuation sensor at 2.19 m, 135° (10 minute interval),
- wind speed at 1.90 m, 225°,
- wind speed at 3.10 m, 225°,
- wind direction,

MAST B:
- incoming shortwave radiation at 2 m,
- reflected shortwave radiation at 2 m,
- net radiation at 2m,
- percentage sun at 2m.

Precipitation at 0.90 m, and
Soil heat flux at 0.03 m depth.

In 1989 during the period 28-04-89 to 09-06-89 the temperature differences were directly measured using thermocouples. After this period only for the temperature differences between sensors T_{d4} and T_{d3} a thermocouple was used. To determine the other temperature differences the temperatures were separately measured using thermistors as was the case in 1988. The reason for changing back to thermistors was the fact that the internal zero compensation of the thermocouples proved to be unstable.

Correction of the measured net radiation

For the net radiation measurement a CSIRO net radiometer from Middleton was used. The instrument was not ventilated other than by natural ventilation.

In accordance with the results of comparative studies of different net radiometers (e.g. Malhy and Hurk, 1992 and Field e.a., 1992), the measured net radiation data Q^*_{meas} were recalibrated using the measured net short wave radiation K^* and separate calibration values for short x_s = 27.3 $Wm^{-2}mV^{-1}$ and long wave responsivity x_l = 25.15 $Wm^{-2}mV^{-1}$ of the net radiometer.

$$Q^*_{corr} = K^* + \frac{x_l}{\bar{x}}(Q^*_{meas} - K^*\frac{\bar{x}}{x_s}) \qquad (Wm^{-2}) \tag{32}$$

Here $\bar{x}$ = 26.22 $Wm^{-2}mV^{-1}$ is the arithmic mean of the short and long wave calibration values, which were originally used to convert the measured electrical tension into net radiation energy fluxes. Periodically the calibration values for the short wave component and the long wave component of the net radiometer were determined in the laboratory.

Correction of the measured soil heat flux

The soil heat flux was measured by three soil heat flux plates placed in series at a depth of 3 cm. To calculate the soil heat flux at the surface, the following procedure is normally used. The measured values at 3 cm depth are corrected for differences in thermal conductivity between the plates and the soil as proposed by Philip (1961):

$$\frac{G_{plate}}{G_{soil}} = \frac{\varepsilon}{1 + (\varepsilon - 1)H'} \qquad (-), \tag{34}$$

where G is the heat flux of the plate and the soil, ε is the ratio of the thermal conductivity of the plate (λ_{plate} = 0.4-0.5 $Wm^{-1}K^{-1}$) to that of the soil (λ_{soil}), and H' a factor depending on the geometry of the plate. For an oblate spheroid, H' is (Philip, 1961):

$$H' = \frac{1}{1 - \eta^2} - \frac{\eta}{(1 - \eta^2)^3}\arctan\{\frac{(1 - \eta^2)^{1/2}}{\eta}\} \qquad (-) \tag{35}$$

where η denotes the ratio of the minor axis to the major axis of the spheroid. With a thickness of the plates of 0.45 cm and a diameter of 10 cm, H' becomes 0.93.

The thermal conductivity of peat with 80% pore space and a volume fraction of water of 0.4 and 0.8 is respectively 0.29 and 0.50 W m^{-1} K^{-1} (Van Wijk and de Vries, 1963). This results in an average correction factor of the soil heat flux density measured with the plates of 1.007. This can be considered as negligible.

To this corrected soil heat flux density at 3 cm depth the change in heat storage of the layer of soil above the plates is added. The soil heat flux at the surface becomes:

$$G_{surf} = G_3 + \int_{0cm}^{3cm} C_{soil}(z) \frac{\delta T}{\delta t} \, dz \qquad (Wm^{-2}) \tag{36}$$

with T the temperature (K), t the time (s), z the depth (m) and C_{soil} as the soil heat capacity (de Vries, 1963):

$$C_{soil} = C_m\Theta_m + C_o\Theta_o + C_w\Theta_w + C_a\Theta_a \qquad (Jm^{-3}K^{-1}) \tag{37}$$

Here C_m (= 2.26 $Jm^{-3}K^{-1}$), C_o (= 2.50 $Jm^{-3}K^{-1}$), C_w (= 4.18 $Jm^{-3}K^{-1}$), C_a (= 0.001 $Jm^{-3}K^{-1}$), Θ_m, Θ_o, Θ_w and Θ_a denote the volumetric heat capacities and the volume fractions of soil minerals, organic matter, soil moisture and air respectively. The soil heat capacity for peat varies between 3.14 - 4.82 MJ m^{-3} K^{-1} for volume fractions of water varying between 0.4 -0.8 and a porosity of 0.8(van Wijk and de Vries, 1963). Because, the vegetation cover is relatively thick (more than 20 cm) the whole year through, the temporal soil temperature gradient is assumed to be relatively small. Thus, no correction is applied to convert the measured soil heat flux density at 3 cm depth to the soil heat flux density at the surface.

Other corrections on the measurements

Corrections for 1988 data.

The corrections on the temperature differences obtained by calibration of the temperature sensors are presented in table 2. Evident erroneous temperature data were deleted, they occurred especially during the hours of sunrise when errors in the temperature measurements were caused by radiative heating of the sensors. Because small errors in the temperature differences cause big errors in the calculated sensible heat flux using the flux profile method, this is used as a second screening of the temperature differences data.

Table 2 The corrections on the temperature differences (°C) for 1988.

	$\Delta T_{d(2-1)}$		$\Delta T_{w(2-1)}$	$\Delta T_{d(4-3)}$
16h00 20-05-88 to 12h30 27-06-88	$\Delta T_{d(2-1)}$ + 0.01 (°C)		$\Delta T_{w(2-1)}$ + 0.00 (°C)	$\Delta T_{d(4-3)}$ + 0.00 (°C)
13h00 27-06-88 to 17h30 25-07-88	$\Delta T_{d(2-1)}$ + 0.02 $\Delta T_{d(2-1)}$ + 0.01	for $T < 20$ °C for $T > 20$ °C	$\Delta T_{w(2-1)}$ - 0.01 for $T < 20$ °C $\Delta T_{w(2-1)}$ - 0.02 for $T > 20$ °C $\Delta T_{d(4-3)}$ - 0.06 for $24 < T \leq 26$ °C $\Delta T_{d(4-3)}$ - 0.05 for $T > 26$ °C	$\Delta T_{d(4-3)}$ - 0.06 for $T < 10$ °C $\Delta T_{d(4-3)}$ - 0.08 for $10 \leq T \leq 24$ °C
13h30 01-08-88 to 16h00 09-08-88	$\Delta T_{d(2-1)}$ - 0.00762 T + 0.072 $\Delta T_{d(2-1)}$ - 0.08	for $T < 20$ °C for $T \geq 20$ °C		
13h00 15-08-88 to 22h00 30-09-88	$\Delta T_{d(2-1)}$ - 0.0025 T + 0.0875 $\Delta T_{d(2-1)}$ + 0.03	for $T < 23$ °C for $T \geq 23$ °C		

In Figure 4 H is calculated by the flux profile method for the periods in which dry bulb temperature differences for different sensors are available.

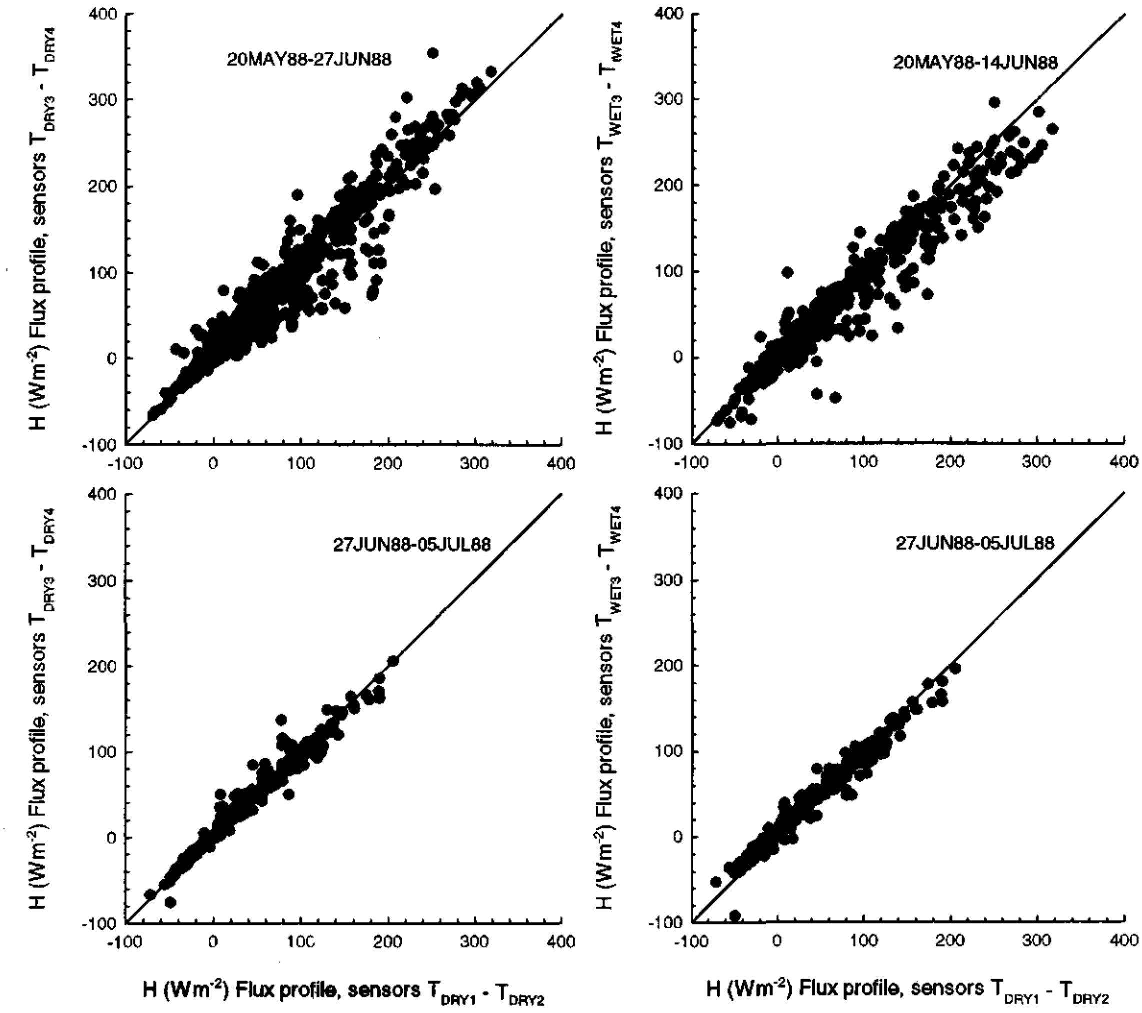

Fig. 4 ***H_{FP} for dry-bulb temperature differences of sensors $T_{d3,4}$ and $T_{d1,2}$ (left hand side graphs) and of sensors $T_{w1,2}$ and $T_{d1,2}$ (right hand side graphs) for the periods 20-05-88 to 27-06-88 and 27-06-88 to 05-07-88.***

It shows that excluding some outliers H of the different sensors corresponds very well. Most of the outliers are caused by radiation errors on the sensors during sunset and sunrise. Additionally these erroneous dry temperature differences of the dry bulb temperature sensors T_{d1} and T_{d2} as well as of the 'wet' bulb (no cotton) temperature sensors T_{w1} and T_{w2} were deleted from the data sets. After this it is concluded that the remaining dry bulb temperature differences of the sensors T_{d1} and T_{d2} are reliable. For the wet bulb temperature differences there is no independent means to check the data.

During night time the measured reflected short wave radiation showed some strange data. It was set to zero if the incoming short wave radiation was zero. Also daytime reflected short wave radiation data were deleted when greater than 350 Wm^{-2}.

The precipitation data stored on the original tapes were calculated using the wrong conversion factor: 0.1733 mm. The correct conversion factor is 0.212 mm. The number of tips were recalculated and multiplied by 0.212 mm to obtain the correct precipitation data.

Corrections for 1989 data.

Table 3 lists the corrections of the temperature differences obtained by calibration of the temperature sensors.

Table 3 Temperature profile corrections (°C) for 1989.

	$\Delta T_{d(2-1)}$		$\Delta T_{w(2-1)}$	
15h30 28-04-89 to 09h00 09-06-89	No corrections needed			
13h30 09-06-89 to 10h00 05-09-89	$\Delta T_{d(2-1)} + 0.00$ $\Delta T_{d(2-1)} - 0.01$	for $T \leq 27$ °C for $T > 27$ °C	$\Delta T_{w(2-1)} - 0.0055\ T + 0.1145$ (°C)	
10h30 05-09-89 to 11h30 03-10-89	$\Delta T_{d(2-1)} + 0.00222\ T + 0.1222$ (°C)		$\Delta T_{w(2-1)} + 0.04$ $\Delta T_{w(2-1)} + 0.03$ $\Delta T_{w(2-1)} - 0.005\ T + 0.125$	for $T \leq 14$ °C for $14 < T \leq 20$ °C for $T > 20$ °C

For the temperature differences directly measured with the thermocouples in the period 28-04-89 to 09-06-89 no correction was needed.

As for 1988 some temperature data were deleted, especially in the morning during sunrise when errors in the temperature measurements were caused by radiation. The dry bulb temperature profiles were checked the same way as for 1988. In figure 5 it can be seen that the dry bulb temperature differences of sensors T_{d1} and T_{d2} are reliable. Less persuasive is the comparison with H of the sensors T_{d3} and T_{d4} (left graph of fig. 5), but this may be due more to sensors T_{d3} and T_{d4} than to T_{d1} and T_{d2}.

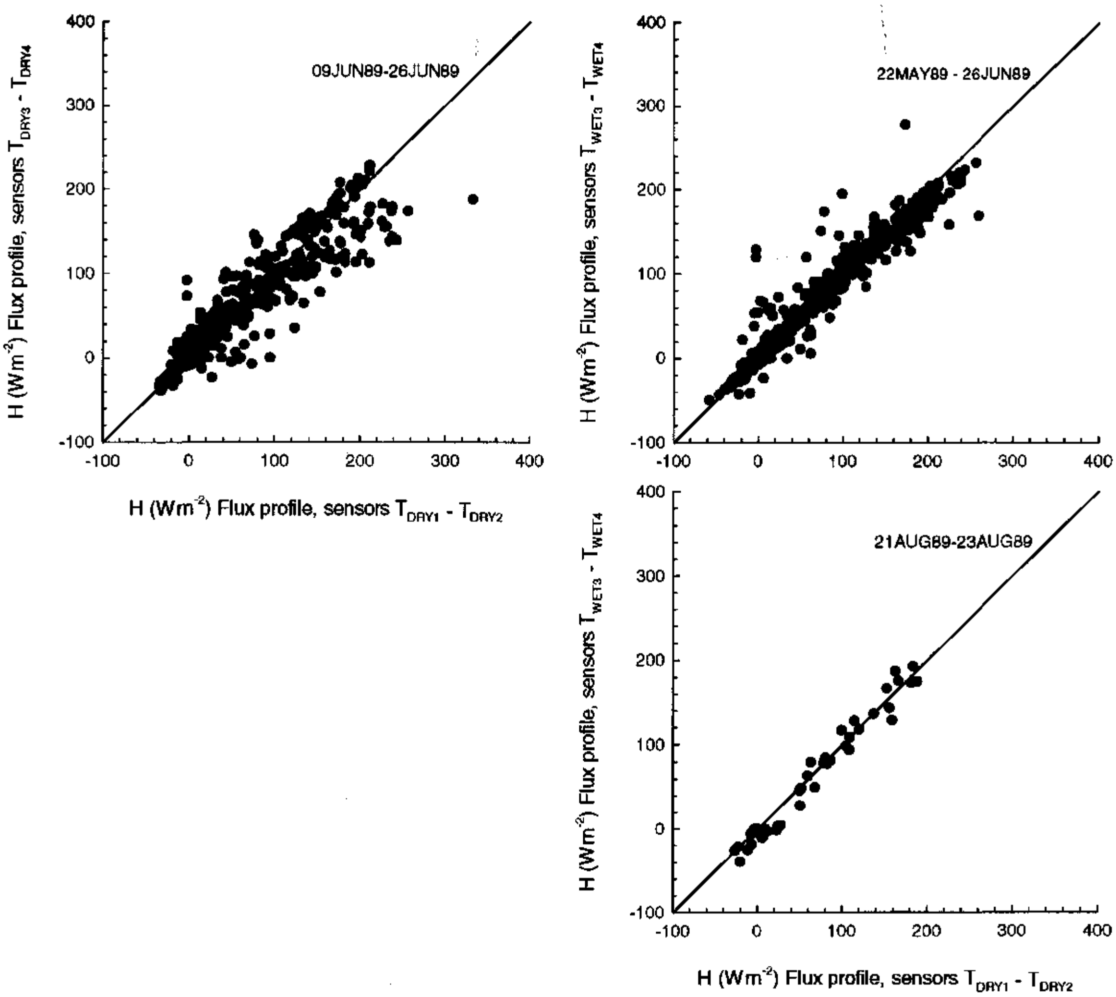

Fig. 5 ***H_{FP} for dry-bulb temperature differences of sensors $T_{d3,4}$ and $T_{d1,2}$ for the period 09-06-89 to 26-06-89 (left hand side graph) and of sensors $T_{w1,2}$ and $T_{d1,2}$ for the periods 22-05-89 to 26-06-89 and 21-08-89 to 23-08-89 (right hand side graphs).***

Additional measurements

To complete the series of the evapotranspiration by Penman and the bulk stomatal resistance by Thom and Oliver, we added for the days when the wet bulb temperature was missing the relative humidity as measured by the KNMI station Twente (located ± 25 km SE of the measuring site). These relative humidities are 24 hour averages from hourly measurements.

The daily precipitation series for the two years were completed using the daily data from the KNMI rainfall station Vroomshoop (see Appendix B).

5. THE SURFACE ROUGHNESS LENGTH AND THE DISPLACEMENT HEIGHT

Molinia shows a strong seasonal dynamics. From November to April only the basal internodes produced during the previous year and the very small new shoots are present. Everything, including the spaces between the tussocks is covered with a layer of litter forming a more or less homogenous layer of 20 - 40 cm thick. A substantial amount of the dead old shoots with an average maximum height of 80 cm remain intact, influencing the wind profile. At the end of May the new shoots start to grow rapidly. The date the maximum biomass is reached varies between August and September. After this the amount of living biomass starts to decline, but depending on wind and rain the dead shoots remain intact for some time (Schouwenaars, 1993 and Berendse *et al.*, 1987).

The surface roughness length, z_0, and the zero plane displacement height, d, are calculated using the logarithmic wind profile under neutral conditions:

$$\bar{u} = \frac{u_*}{k} \ln\left(\frac{z - d}{z_0}\right) \qquad (\mathrm{ms}^{-1}) \qquad (38)$$

The atmosphere was considered neutral for a temperature difference less than plus or minus 0.02 °C between the dry bulb temperatures at level 1 and 2. To reduce the errors in the wind speed measurements, only the wind profiles with a minimum instantaneous wind speed $\geq$ 2.0 ms^{-1} were selected.

Table 4 ***Displacement height d and roughness length z_0.***

	d (m)	z_0 (m)	N
1988	0.362	0.041	235
1989	0.335	0.048	279

A minimization procedure was used to solve z_0 and d in eq. (33). As there is no clear minimum, the results are sensible for the starting values used. However, it was found that different but realistic pairs of z_0 and d values only cause a slight change in the calculated sensible heat flux density. Besides the averages of z_0 and d for each year, we also looked for any consistency with time (different vegetation height) and wind direction, however none was found. In table 4 the average values for the two years are shown for the displacement height and the surface roughness length. N is the number of observations complying with the above mentioned conditions.

Duyzer and Bosveld (1988) found for *Molinia caerulea* growing on peat in June a value for d of 0.4 m and a z_0 of 0.043 ± 0.005 m based on eddy correlation and flux profile measurements. At the time of their measurements the vegetation consisted mainly of 40 cm high tussocks with old dry grass partly filling the spaces between the tussocks. The values found by Duyzer and Bosveld correspond well with the values found at Engbertsdijksvenen. In Canada Ripley and Redmann (1976) studied grassland with *Agropyron dasystachyum* as the major species. They describe this vegetation as having a mean height near 20 cm with a few stems and flowering stalks to 40 cm and occasionally higher. With a dense mat of dead leaves below 10 cm. They found a displacement height d ranging from 0.10 m to 0.15 m and a roughness length z_0 ranging between 0.01 m and 0.04 m with a mean near 0.025 m. Although, the vegetation is somewhat lower than *Molinia caerulea* at Engbertsdijksvenen, the structure of the vegetation with a dense base and a sparse top is the same. At Engbertsdijksvenen the magnitude of d also matches the thickness of the layer of death grass and the height of the tussocks. Also the ratio of z_0 to the height of the stalks minus d ($z_0 \approx 0.1(h_{stalks} - d)$) corresponds well to the findings of Ripley and Redmann (1976).

6. RESULTS AND DISCUSSION

The Flux profile method

As mentioned in the section "Measurements - Corrections" most of the erroneous outliers in figures 4 and 5 are deleted from the dry bulb temperature differences data sets. For several periods this leaves dry bulb temperature differences from 2 or 3 different sets of sensors. But, even with these erroneous data deleted there still remains some scatter between H from the different sets of sensors. However, for periods where the scatter is most pronounced, see e.g. the left hand side graph of figure 5, there is for the same period much less scatter when compared with a different set of sensors, see the top graph at the right hand side of figure 5. Thus, in all cases the sensible heat fluxes resulting from temperature differences of sensors T_{d1} and T_{d2} correspond well to very well with at least the sensible heat fluxes from one other set of sensors. Therefore, we take it that the sensible heat flux density resulting from the dry bulb temperature differences of sensors T_{d1} and T_{d2} is a good to very good approximation of the true sensible heat flux density.

The Bowen ratio

In figure 6 the latent and sensible heat flux density calculated by the flux profile method L_eE_{FP} and H_{FP} are plotted against those calculated by the Bowen ratio method L_eE_{Bow} and H_{Bow}. Because the Bowen ratio method is unreliable when B becomes close to -1 only daytime data (9h00 - 17h00) were plotted. As stated before it is difficult to determine if the wet bulb temperature differences are reliable. Also, a small error in ΔT_{wet} already causes a large error in L_eE_{Bow} (Chatillon, 1988). However, if the dry and especially the wet bulb temperature differences are measured correctly the Bowen ratio method is considered very reliable. Although, for a large part of the measuring period the wet bulb temperature differences are clearly not accurate, it can also be seen that the results of the two methods correspond well if wet and dry bulb temperature differences are correct (as in the period 26-07-89 to 31-07-89), see the second graphs of figure 6.

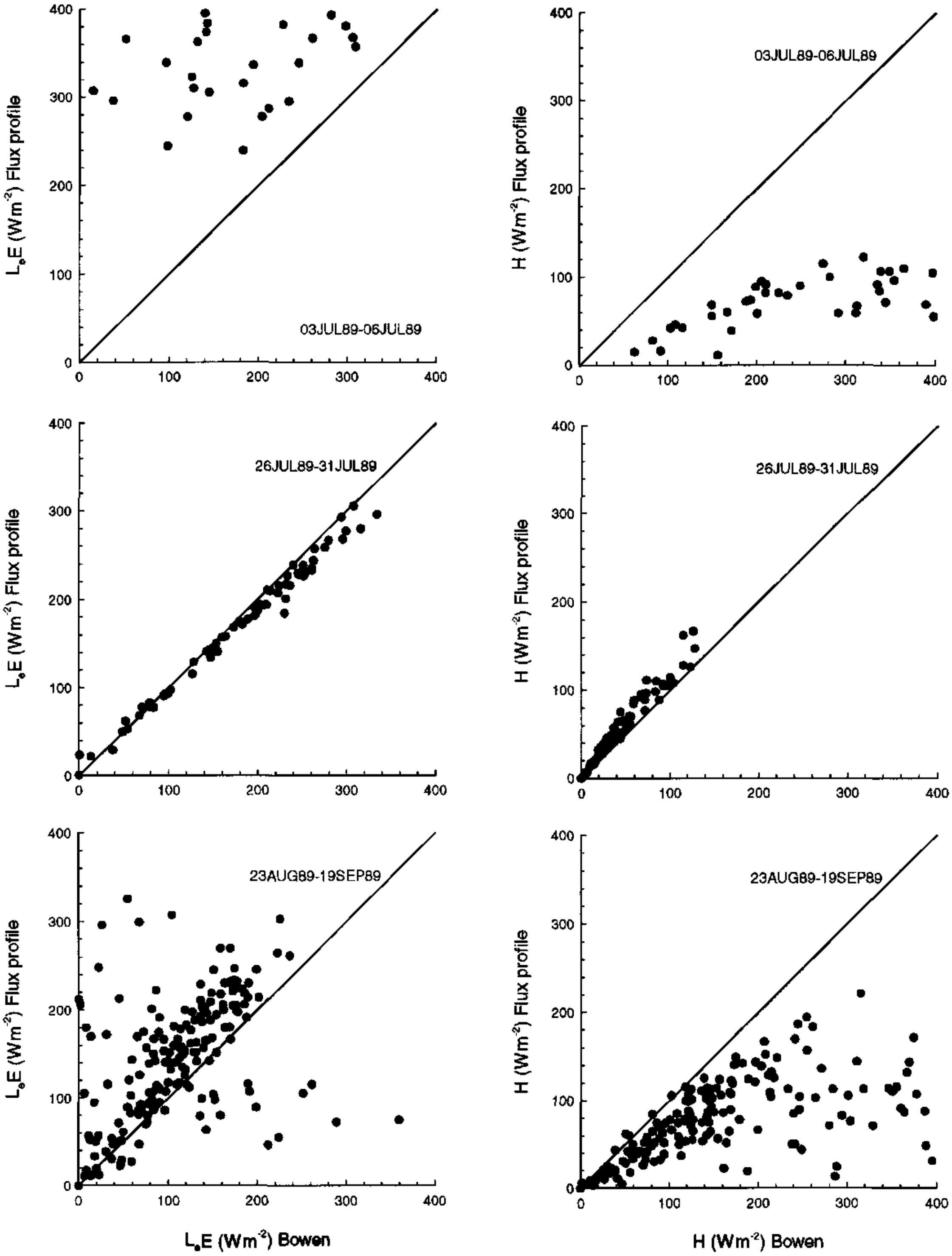

Fig. 6 ***L_eE and H calculated by the flux profile method plotted against L_eE and H calculated by the Bowen ratio method for 3 different periods.***

The Temperature fluctuation method

One of the main advantages of the temperature fluctuation method is the need of relatively simple measurements of temperature and wind speed at only one level. However, as indicated in the section "Methods" there is with the present equipment setup an underestimation of the sensible heat flux density. To overcome this problem, use was made of the friction velocity and the stability length found by the flux profile method in combination with the regression results of the de Man (1990). Research is planned to address this problem. The thus found sensible heat flux density of the temperature fluctuation method H_{TF} compares relatively well with the results from the flux profile method. This is depicted in figure 7, where four different periods in 1989 are plotted. It shows that the temperature fluctuation method has a great potential as a relatively easy and inexpensive but accurate method to measure the daytime sensible heat flux density, especially for remote long-term measuring campaigns. The same good results were found by de Bruin *et al.* (1991) and Moors *et al.* (1994).

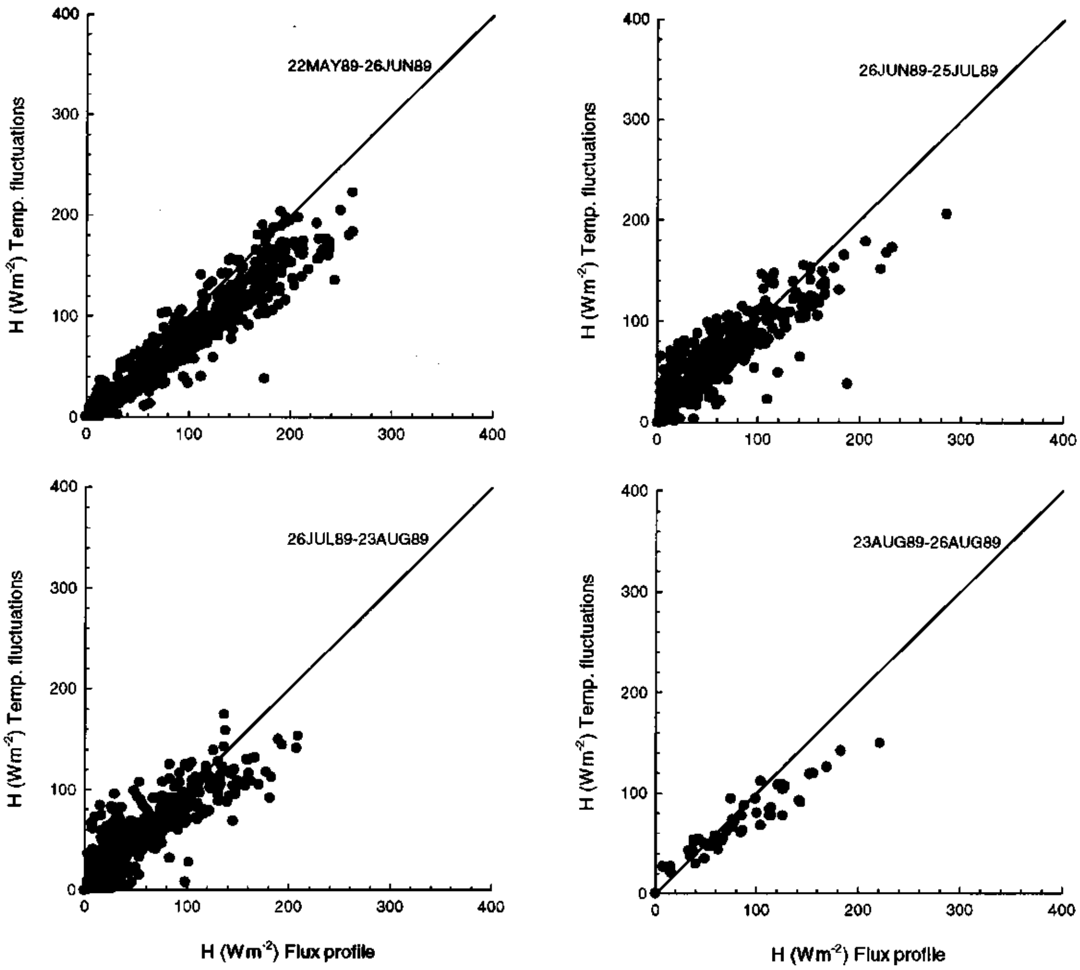

Fig. 7 ***Scatter plots of H calculated by the temperature fluctuation method and by the flux profile method for different periods.***

The latent heat flux density or actual evapotranspiration rate

None of the three methods discussed above provides the latent heat flux density directly. They all need measurements of two other components of the energy balance at the surface i.e. the soil heat flux and the net radiation flux to calculate the latent heat flux density. So, the accuracy of the latent heat flux density depends also on the accuracy of the other two components of the energy balance. The net radiometer used in this experiment proved to give results differing less than 5% in comparison with a four component radiometer (see Moors *et al.*, 1994). The net radiation data are consequently considered accurate. The soil heat flux density data are normally less accurate than the net radiation data, however the ratio of the soil heat flux to the latent heat flux is so small that this inaccuracy will only have minor influences on the latent heat flux density, especially during daytime.

As the Bowen ratio results show large gaps of missing records, mainly attributed to malfunctioning of the wet bulb temperature differences sensors, they are not used for the calculation of the actual evapotranspiration rate. The micro-meteorological method that provided a data set with the least amount of data missing, is the flux profile method. From the comparison of the sensible heat fluxes of this method with those of the Bowen ratio method, together with the comparison of H for different sets of temperature sensors, it is concluded that an uncertainty of less than 20% in H calculated by the flux profile method with the temperature differences of the sensors T_{d1} and T_{d2} is most likely.

The uncertainty in E depends on the uncertainty of each of the other three components of the energy balance as well as their ratio to E. In appendix A some examples are shown of the daily cycle of the different components of the energy balance using the flux profile method on a 30 minute basis. The uncertainty in E will be less than 15% for these relatively humid conditions. The thus obtained 30 minute latent heat flux density data are used to calculate the daily actual evapotranspiration rate to be used later on.

Potential evapotranspiration

The daily data are also listed in appendix B. As far as available the daily data are based on 24 hour sums or averages of the 30 minute data. Missing relative humidity data (i.e. wet-bulb temperature data) are replaced by 24 hour averages based on hourly data from KNMI station Vliegbasis (= air base) Twente.

With the high water table in 1988 it is assumed that there is no restriction in the water supply of *Molinia* in that year, except for may be the second half of June.

The potential evapotranspiration as used here, is the evapotranspiration of a vegetation well supplied with water, but *not* wet. In the case of rainfall, water may be intercepted by the vegetation and directly evaporated into the atmosphere. Thus in such a case the actual evapotranspiration is the total of the transpiration of the vegetation, the evaporation of the intercepted water and the evaporation of the soil. For dry days the evaporation of the intercepted water will be zero and in the case of our site, we will consider the evaporation of the soil as negligible. The latter due to the fact that the ground is covered with a thick layer of litter. This assumption is also supported by the low soil heat flux density measured (see the figures of appendix A). Of course this layer of litter will also intercept water, but this will only contribute to the evaporation of intercepted water when wetted. Thus, for dry days, the actual (evapo-)transpiration will be equal to the potential (evapo-)transpiration, assuming there is no limitation on the water supply. For days with less than 0.4 mm of rain (< 2 tips) the evaporation of the precipitation intercepted by the vegetation and the cover of litter will be considered negligible.

Table 5 ***Results of the regression analysis where $E_{Flux\ Profile}$= slope E_{Penman} (or $E_{P\&T}$ or $E_{Makkink}$).***

		slope	Standard Error of Estimate	R^2	N
1988	Penman	0.864	0.435	0.665	41
	P&T	0.891	0.389	0.827	90
	Makkink	1.034	0.389	0.783	90
$P<0.4 mmd^{-1}$	Penman	0.827	0.390	0.764	23
$P<0.4 mmd^{-1}$	P&T	0.839	0.326	0.887	47
$P<0.4 mmd^{-1}$	Makkink	0.967	0.395	0.834	47
1989	Penman	0.818	0.472	0.825	63
	P&T	0.883	0.463	0.883	93
	Makkink	0.926	0.413	0.817	93
$P<0.4 mmd^{-1}$	Penman	0.807	0.525	0.783	46
$P<0.4 mmd^{-1}$	P&T	0.871	0.430	0.864	71
$P<0.4 mmd^{-1}$	Makkink	0.912	0.543	0.783	71

Comparing figure 8 ($P < 0.4$ mmd^{-1}) with figure 9 (no limitation on P) shows the actual evapotranspiration rate calculated by the Flux profile method is reletavily higher if no limitation on the amount of rain is used (see also the differences in coefficients in table 5.). This implies that the actual evapotranspiration rate is only comparable to the potential evapotranspiration rate when evaporation of intercepted water is negligible, i.e. during dry days. For this we will only consider days with less than 0.4 mm of rain in the following discussion on potential evapotranspiration rates.

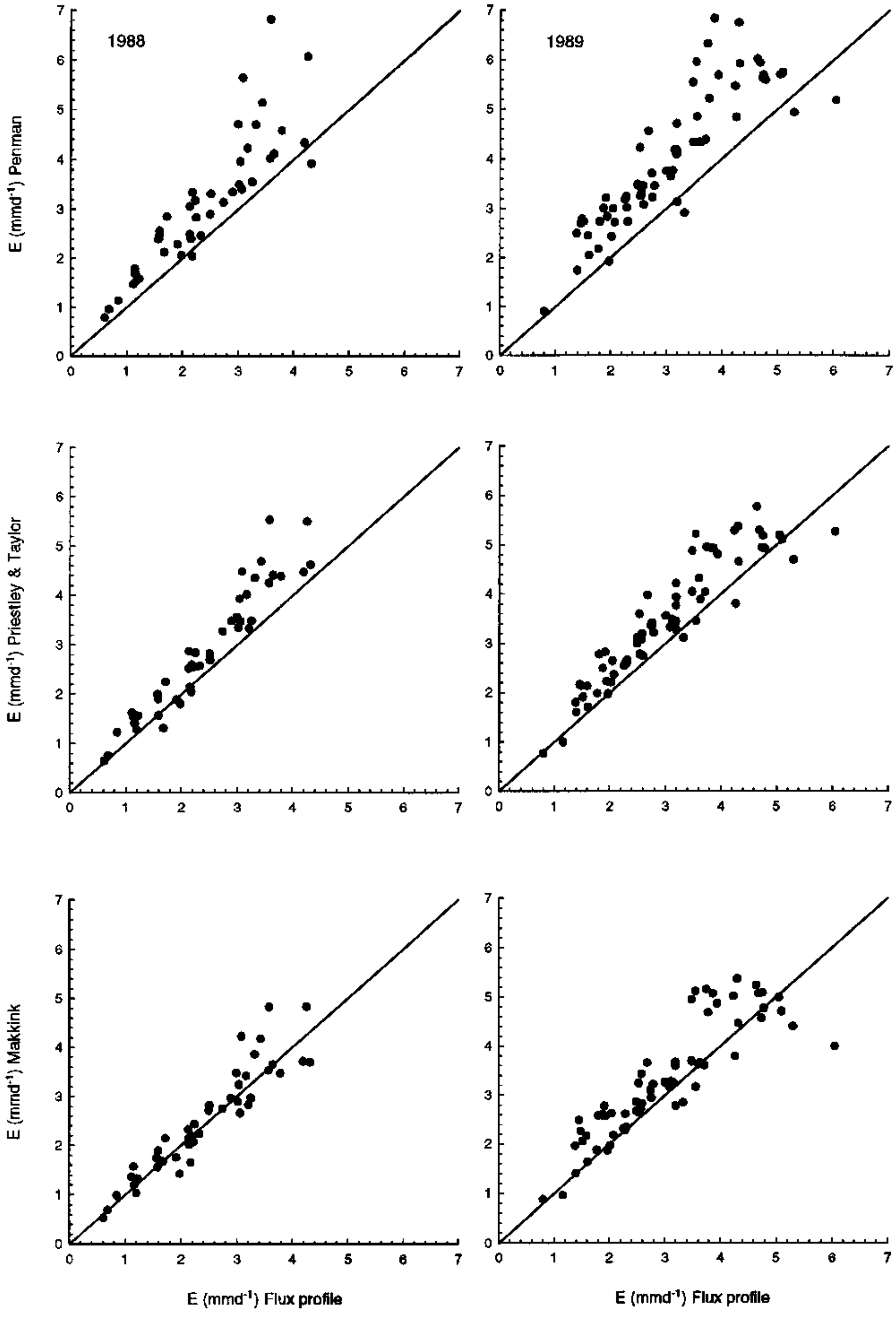

Fig. 8 ***Daily evapotranspiration data of 1988 and 1989 calculated by Penman, Priestley & Taylor and Makkink plotted against the evapotranspiration data calculated by the flux profile method, for days with $P \leq 0.4$ mmd^{-1}.***

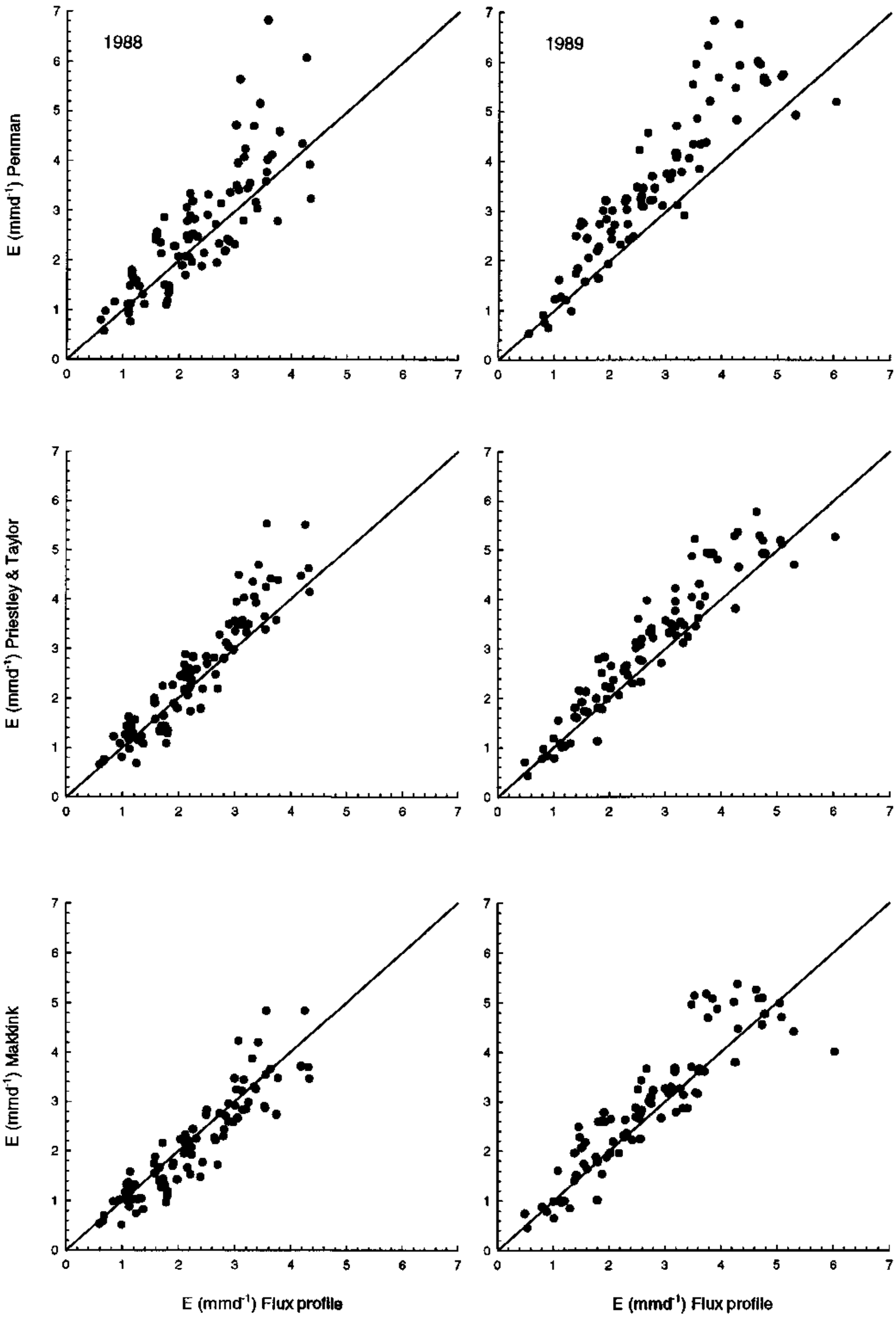

Fig. 9 ***Daily evapotranspiration data of 1988 and 1989 calculated by Penman, Priestley & Taylor and Makkink plotted against the evapotranspiration data calculated by the flux profile method, no restriction for P.***

Modified Penman

Stricker and Brutsaert (1978) noted that the modified Penman equation was a good measure for the potential evapotranspiration of a short grass well supplied with water. The top graphs of figure 8 and 9 show that even without water shortage as is the case in 1988 the modified Penman equation is still overestimating the measured evapotranspiration rate. In the case of *Molinia* this method does not provide a good estimation of the potential evapotranspiration rate. This is even more clear when looking at the daily variation of the ratio of the actual evapotranspiration calculated by the flux profile method and the evapotranspiration rate calculated by the modified Penman equation, as shown in figure 10.

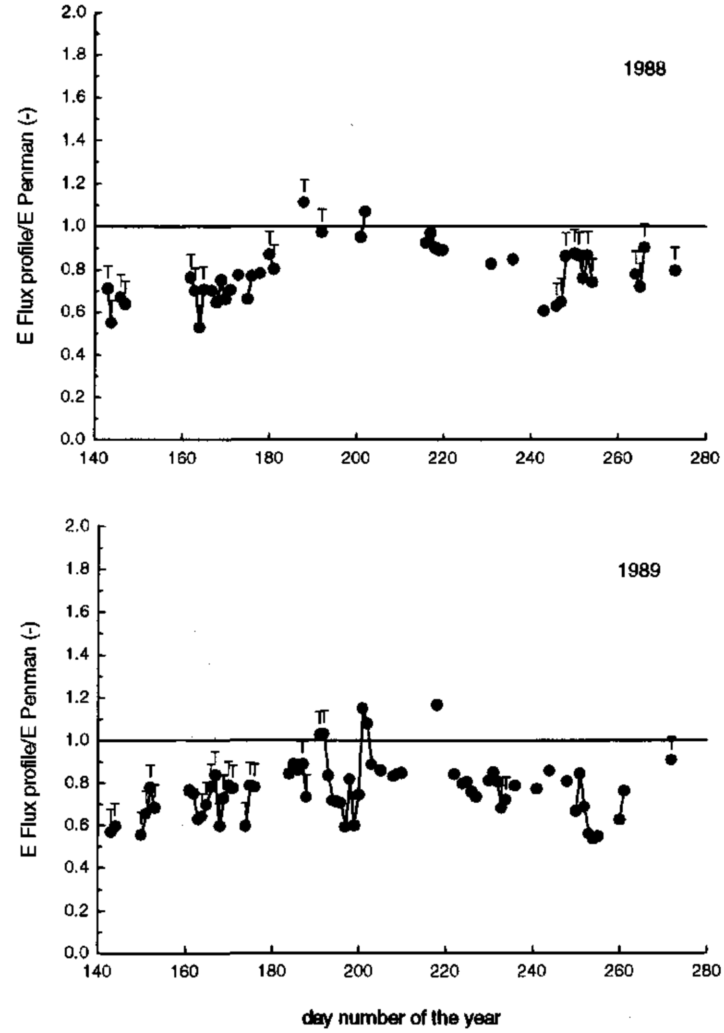

Fig. 10 ***Ratio of E calculated by the flux profile method over E calculated by Penman, $P \leq 0.4$ mmd^{-1} (T=data from air base Twente).***

Only in the month July of 1988 the ratio is close to one, or slightly higher. For the other months the modified Penman equation is overestimating the actual evapotranspiration, and with the assumption that in 1988 the actual equals the potential evapotranspiration, the method is also overestimating the potential evapotranspiration most of the time in 1988.

Priestley and Taylor

The results of the Priestley and Taylor equation are quite similar to those of the modified Penman equation. This is illustrated by the graphs in the middle of figure 8, and the results of the regression analysis in table 5. The more or less similar results of the two methods are caused by the fact that the experiment was done in the growing season when the radiation is the most important driving force of the evapotranspiration. In the winter period the humidity deficit and the windspeed will become more important, causing larger differences between the two methods.

From the present experiment it was found that for *Molinia* growing on peat the potential evapotranspiration is better represented by using a value of 1.01 for the constant α instead of the value 1.28 applied here. This has its background in the relatively low albedo-value causing a Q^*/K_d-value higher than usual.

One remark should be kept in mind when interpreting the above results. In this study for both, the modified Penman and the Priestley and Taylor equation, meteorological data are used measured above *Molinia* on peat. However, in general when applying these two methods for water management purposes, meteorological data will be from a meteorological station above a short grass surface. This will cause differences in the net radiation among others through a different albedo, the saturation vapour pressure deficit, the air temperature and the wind speed. Here no research is done on the magnitude of the differences caused by this.

Makkink (reference evapotranspiration) and crop coefficients

The reference crop evaporation is calculated by the Makkink equation. In the bottom graphs of figures 8 and 9 scatter plots are depicted of the actual evapotranspiration rate calculated by the flux profile method and the reference crop evaporation rate as calculated by Makkink. From table 5 it can be seen that the slope of the regression line is close to one and the results of the two methods are almost similar. In figure 11 the daily variation of the ratio of the actual evapotranspiration calculated by the flux profile method over the reference evapotranspiration calculated by the Makkink equation is plotted.

As explained for the potential evapotranspiration in this figure only days with less than 0.4 mm of rain are selected. The ratio varies in 1988 from a minimum of 0.7 in June and September to a maximum of 1.4 in July. In 1989 the minimum of the ratio varies from 0.6 to 0.7 in June and September. The maximum of 1.5 is found in August and is probably the result of a dry period followed by rain.

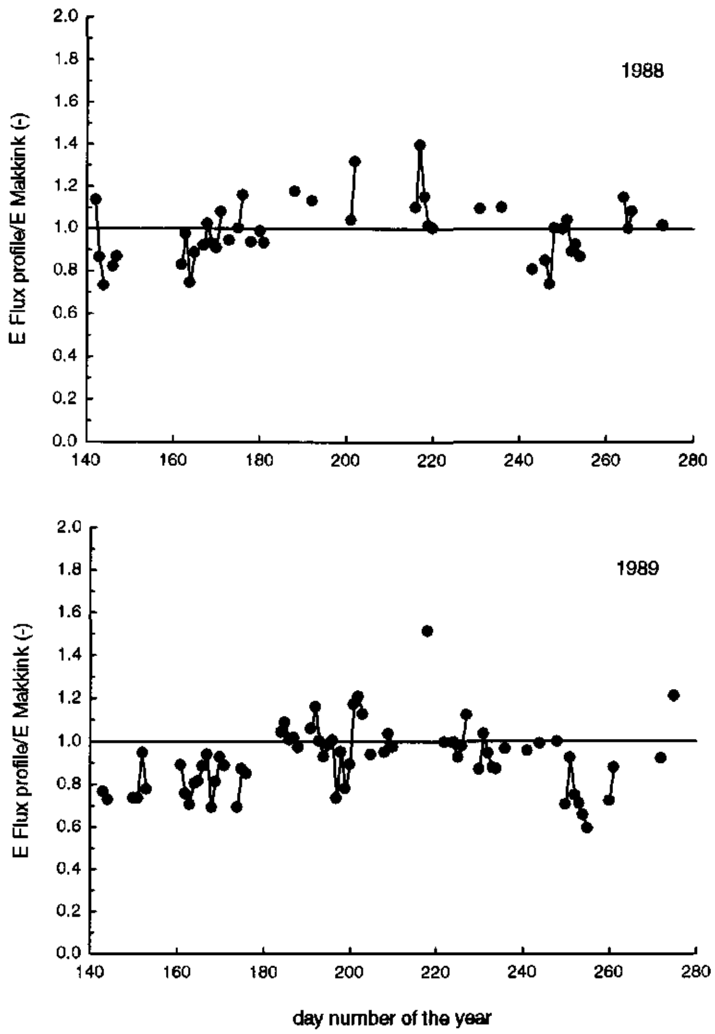

Fig. 11 ***Ratio of E calculated by the flux profile method over E calculated by Makkink, P ≤ 0.4 mmd⁻¹.***

If the actual evapotranspiration is equal to the potential evapotranspiration, the ratio depicted in fig. 11 is the same as the crop coefficient *f*. In table 6 the crop coefficients per decade are listed, with the evapotranspiration by the Makkink equation as the reference crop evapotranspiration. It can be seen that the crop coefficients are lower in 1989 than in 1988, suggesting as expected that the actual evapotranspiration in 1989 is slightly lower than the potential evapotranspiration. Thus, only the crop coefficients of 1988 should be considered to provide the potential evapotranspiration. As there are only data of one year available with potential evapotranspiration, care should be taken in the use of the here calculated crop coefficients. Comparing these results with the values for grass with a height of 25 cm or more (f = 1.1 - 1.2) as supplied by Hooghart and Lablans (1988), then the crop coefficients for *Molinia* found in this study seem slightly lower, especially at the beginning of the growing season. It should be noted that Hooghart and Lablans in contrary to the present study included the evaporation of intercepted water in their derivation of the crop coefficients. This will produce a relatively higher crop coefficient. Comparison of figures 8 and 9 gives an impression of the magnitude of the evaporation of intercepted water. However, as the equipment used for the flux profile method is inaccurate when wet (during rain), the results are no more than an indication of the magnitude of the evaporation of the intercepted water. Possibly, the interception will be better quantified when the results of the present study are compared with those of the lysimeter study of Schouwenaars (1993).

Table 6 ***Decade values for the crop coefficients for the reference evapotranspiration as calculated by the Makkink equation. N is the number of data used.***

		1988		1989	
Month	Dec.	f	N	f	N
May	III	0.81	(3)	0.74	(4)
June	I	0.83	(1)	0.87	(3)
	II	0.93	(8)	0.82	(10)
	III	1.01	(7)	0.80	(3)
July	I	1.15	(2)	1.04	(7)
	II	1.25	(3)	0.96	(10)
	III	1.21	(1)	1.04	(6)
August	I	1.15	(6)	1.14	(3)
	II	1.09	(1)	0.98	(8)
	III	0.96	(3)	0.98	(5)
September	I	0.91	(8)	0.84	(6)
	II	1.15	(1)	0.71	(4)
	III	1.03	(3)	0.92	(1)

Although, the Makkink equation is in general not used to directly estimate the potential evapotranspiration, the bottom graphs of figures 8 and 9 and the regression results in table 5 show that in the case of *Molinia* the results of the Makkink equation are close to the potential evapotranspiration.

The bulk stomatal resistance

The bulk stomatal resistance r_s as formulated by Thom and Oliver (1977) is calculated on a daily basis and assuming the evapotranspiration calculated by the flux profile method as the actual evapotranspiration. In 1988 m = 3.69 and in 1989 m = 4.01 was used. To calculate r_a the measured wind speed data were transformed to the 2 m level assuming a logarithmic profile. In figure 12 the results for both years are shown. Only days with precipitation $\leq$ 0.4 mm are depicted. In 1988 the average r_s is 94 sm^{-1} (N = 46, range 6 - 192 sm^{-1}) and in 1989 113 sm^{-1} (N = 72, range 30 - 205 sm^{-1}). Although there is a large scatter, the overall trend is high r_s values at the beginning and the end of the growing season, and lower values in July. Besides the seasonal trend there is also a marked difference in the r_s values of July for 1988 and 1989. A possible cause is the lower water table starting in July 1989; ± 40 cm lower than in 1988. In August the r_s values of both years become comparable again, probably due to die off. If the assumption is made that the r_s values are directly proportional to the leaf area, then r_s values may be adjusted for this by multiplying by the ratio of *LAI* to the maximum *LAI*. The result of this is also shown in figure 12. Averaged over the whole measuring periods the *LAI* adjusted r_s values are 65 sm^{-1} (N = 46, range 4 - 140 sm^{-1}) and in 1989 81 sm^{-1} (N = 72, range 16 - 174 sm^{-1}).

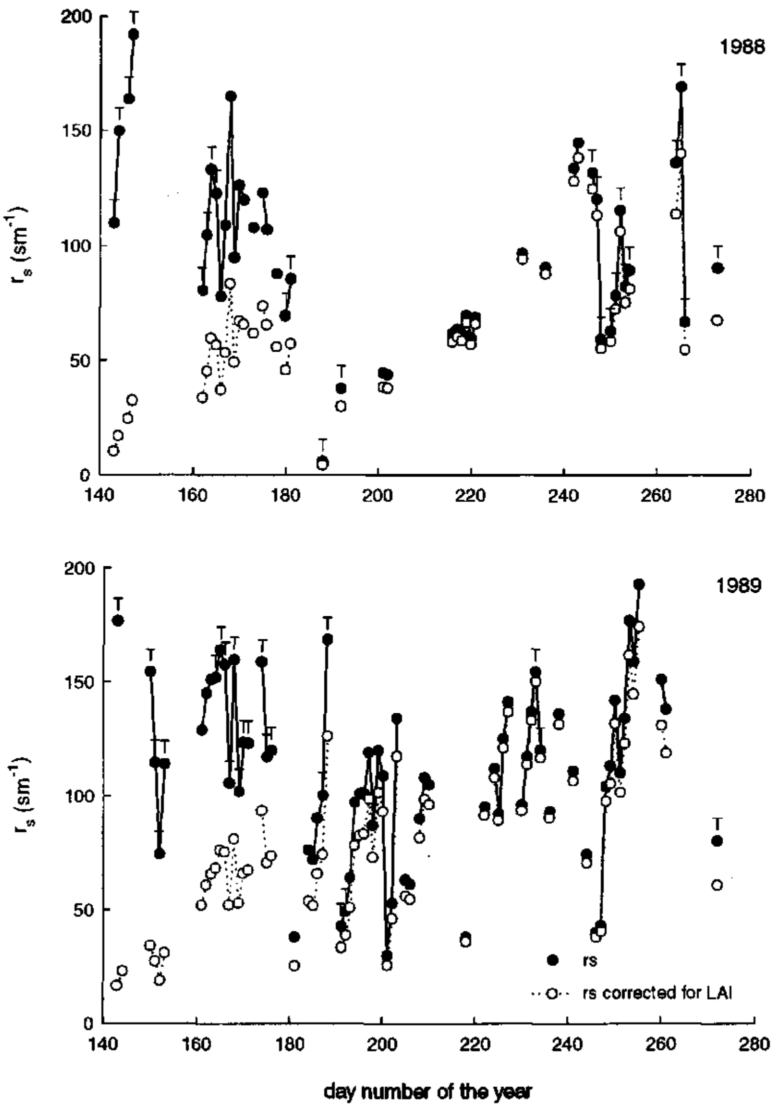

Fig. 12 ***The bulk stomatal resistance as defined by Thom and Oliver, $P \leq 0.4$ mmd^{-1}. The open circles are the r_s values adjusted for the leaf area. (T = missing wet-bulb temperature data replaced by relative humidity data from KNMI station air base Twente)***

For grass an average value in the summer months for r_s of 65 sm^{-1} was found in the Hupselse Beek experimental catchment area. Compared with this the values of r_s for *Molinia* at Engbertsdijksvenen are somewhat higher. However, when the r_s values are adjusted for the leaf area they are comparable to those of grass. The relatively high evaporation rate is in correspondence with the results found by Nichols *et al.* (1980). They noted that the evaporation from moss surfaces plus grasses and sedges was significantly higher than that from the water surface. Also Rutherford *et al.* (1973) reported that the evapotranspiration from a New Hampshire bog was 1.7 times the evaporation from open water.

The relatively high evaporation rate when compared with short grass is also due to the higher amount of available energy for evaporation for *Molinia* than for grass. This is partly caused by the litter covering the ground, which reduces the soil heat flux, and partly by the low albedo for *Molinia* (average albedo ± 0.17), especially at the start and end of the growing season (0.14), when compared to grass (0.24). This would also explain the relatively high potential evapotranspiration rate calculated by the modified Penman equation and the Priestley and Taylor equation when using the meteorological input data measured above *Molinia* at Engbertsdijksvenen.

For a research on dry deposition fluxes of among others *Molinia* at Fochtlooërveen (peat-moorland in Drenthe, The Netherlands) Duyzer and Bosveld (1988) used a different definition for the resistances. The total resistance is split up in three parts: the aerodynamic, the scalar-excess and the surface resistance. Their aerodynamic resistance was calculated using stability functions (see for example the flux profile method). In the formulation of Thom and Oliver the stability corrections are empirically included in the wind function of Penman. The scalar-excess resistance, often called the boundary layer resistance, added to the surface resistance (i.e. stomatal and cuticular resistance) should equal the bulk stomatal resistance as defined by Thom and Oliver.

Using 30 minute averages Duyzer and Bosveld observed for seven days in the period 19 May to 3 June 1985 at Fochtlooërveen a daytime surface resistance for *Molinia* on peat ranging from 250 to 400 sm^{-1} and a mean of 350 sm^{-1}. With the boundary layer resistance ranging from 20 to 40 sm^{-1} and keeping in mind that for the bulk stomatal resistance at Engbertsdijksvenen daily values are used, while for the surface resistance at Fochtlooërveen daytime averages, the magnitude of the resistances at Engbertsdijksvenen seem relatively low. As the measurements at Fochtlooërveen were made at the start of the growing season, the differences may be attributed mainly to differences in moisture conditions of the litter and soil layer.

8. CONCLUSIONS

The flux profile method gives consistent results when the sensible heat fluxes calculated with the 30 minute averages of the temperature differences measured by three different sets of sensors are compared. Also when the latent heat fluxes of the flux profile method are compared with those of the Bowen ratio and the temperature fluctuation method the results are consistent. However, for the Bowen ratio method this good agreement is only found for some periods. This is caused by the high sensitivity of the Bowen ratio method for the wet bulb temperature differences and these are difficult to measure accurately. The results of the temperature fluctuation method are promising as relatively simple and low cost method to measure the sensible heat flux. Improvements of the present measuring method for the temperature fluctuation method may be obtained by changing the filtering method used and/or by using a sensor with a smaller time constant.

As the results of the flux profile method compare well with the other two methods, and there are nearly no data missing for the whole measuring period, these results are used to calculate the daily actual evapotranspiration rate.

In the year 1989 the accumulated precipitation for the measuring period was approximately one third less than for the same period in 1988. This resulted in a lower groundwater table in 1989. With the high groundwater table in 1988 it is assumed that for dry days the actual evapotranspiration rate equals the potential evapotranspiration rate. When only data are used for dry days (P < 0.4 mmd^{-1}) it shows that the modified Penman as well as the Priestley and Taylor equation overestimates the potential evapotranspiration of *Molinia*. When meteorological data measured at the *Molinia* site are used the potential evapotranspiration is better estimated by the Priestley and Taylor equation when the parameter α is changed from 1.28 to 1.01.

The evaporation calculated by the Makkink equation is close to the actual evapotranspiration in 1988, and the crop coefficients, using the results of the Makkink equation as the reference evaporation, are close to those for tall grass (1.1 to 1.2). However, at the beginning of the growing season the crop coefficients for *Molinia* at Engbertsdijksvenen tend to be lower (0.8).

The bulk stomatal resistance found for *Molinia* at Engbertsdijksvenen has an average in 1988 of 94 sm^{-1} and in 1989 of 113 sm^{-1}. This average for the growing season is somewhat high when compared with the resistance for short grass (65 sm^{-1}). When adjusted for the leaf area the average resistance of 1988 is similar to that of short grass. The relatively high evaporation rate is among others due to the low albedo in comparison to short grass.

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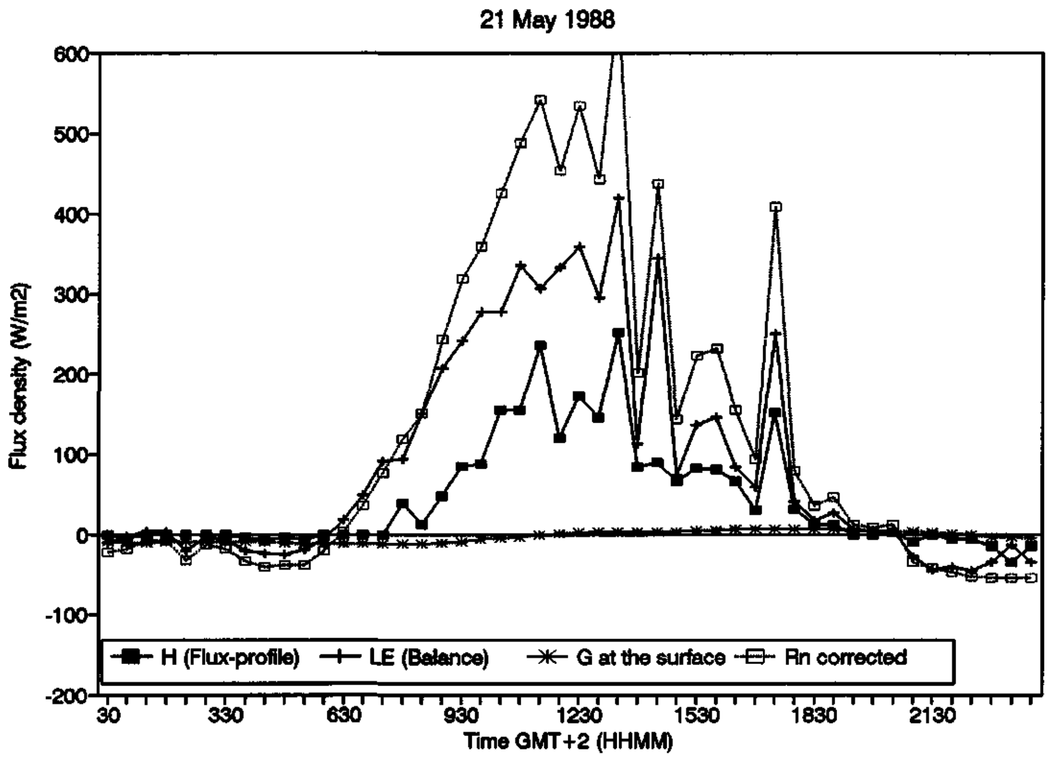
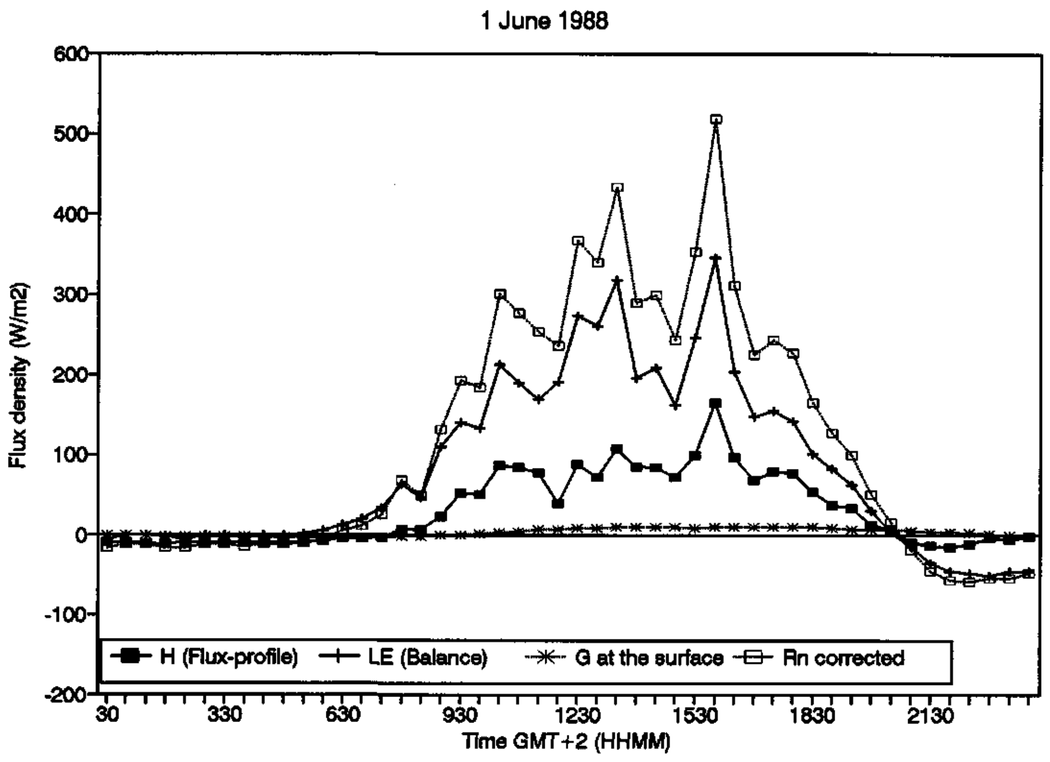
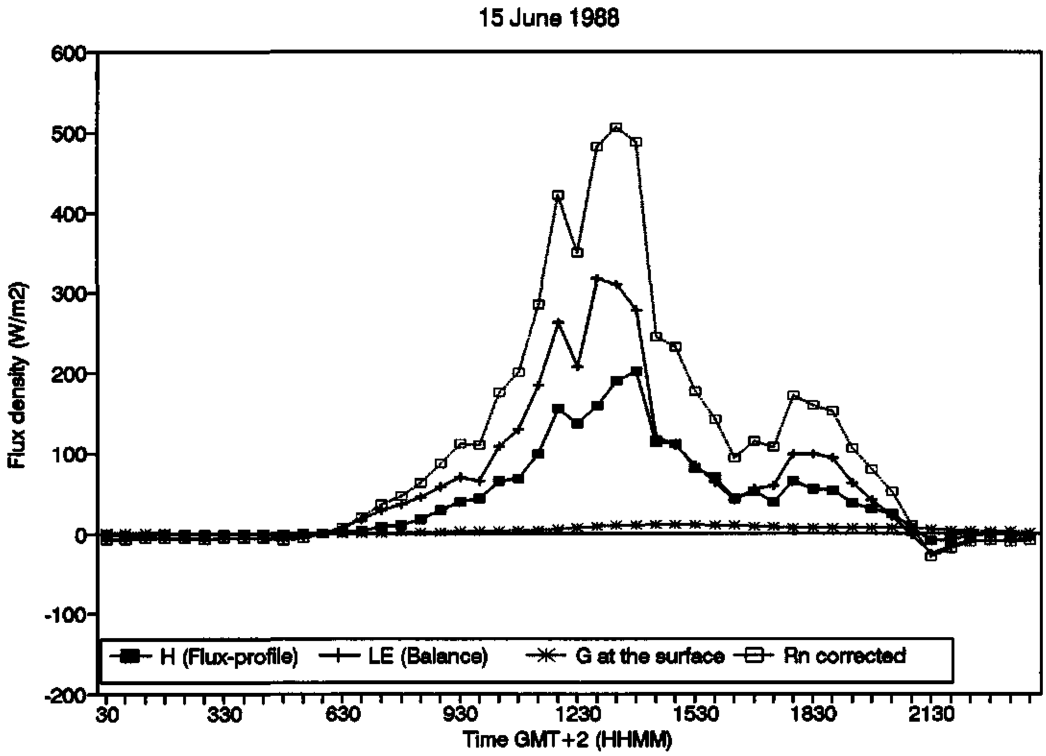
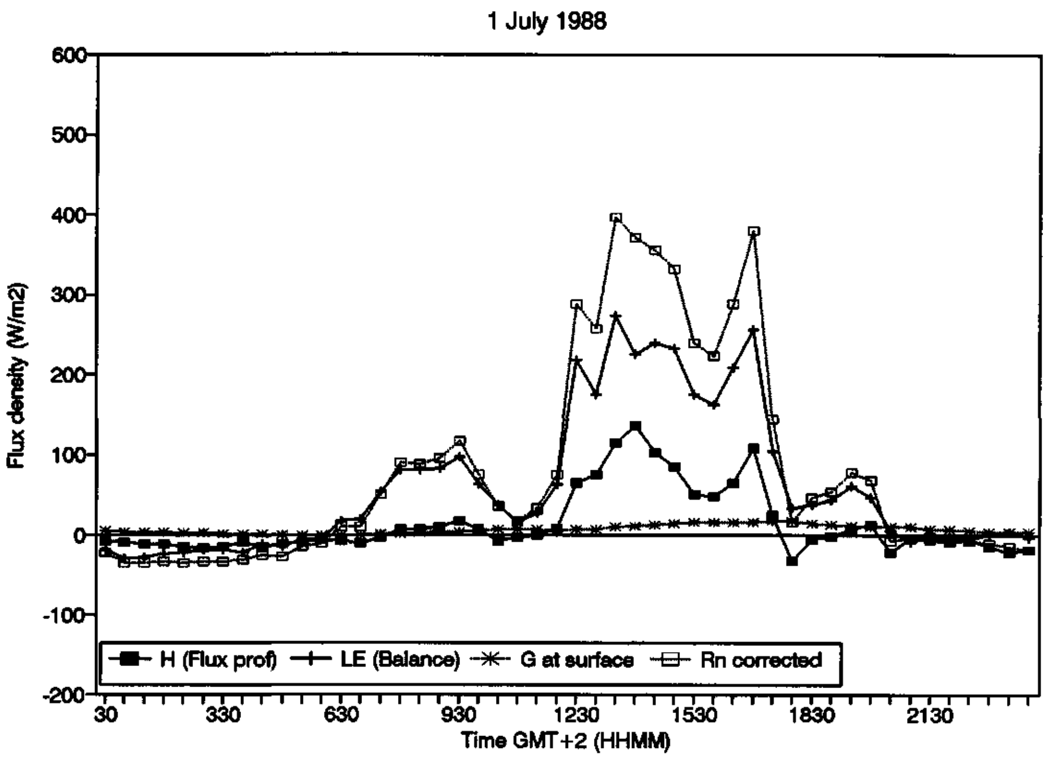
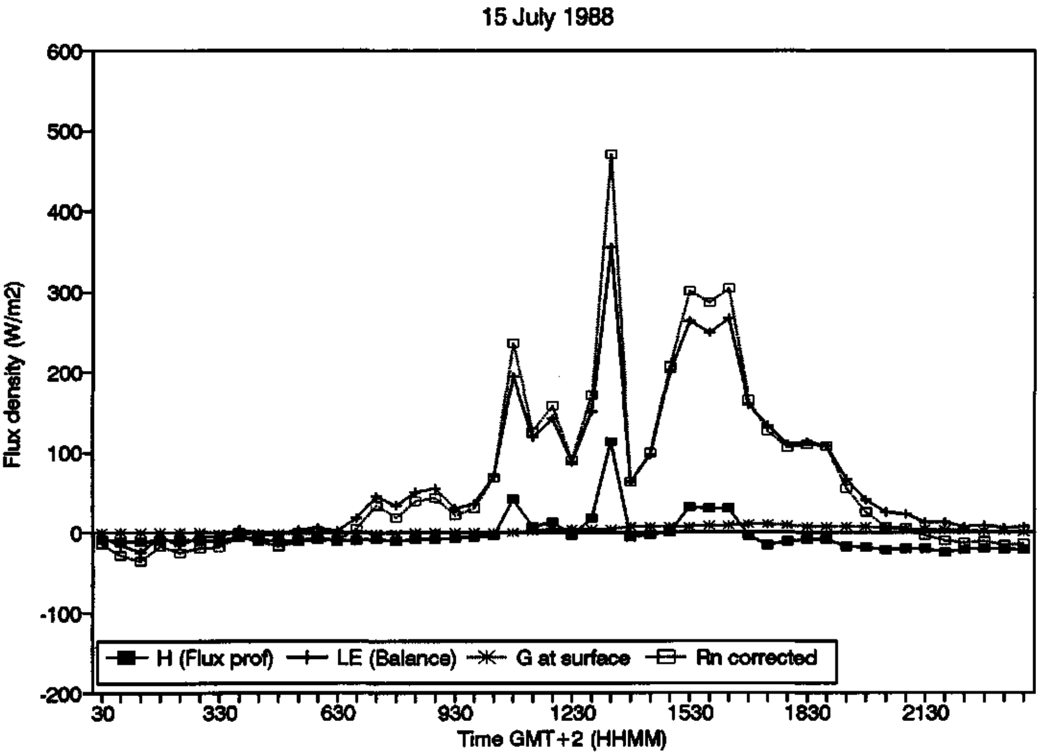
Appendix A Figures of the energy balance at the surface for some dates in 1988 and 1989.


21 May 1988
Flux density (W/m2)
600
500
400
300
200
100
0
-100
-200
H (Flux-profile)
LE (Balance)
G at the surface
Rn corrected
30
330
630
930
1230
1530
1830
2130
Time GMT+2 (HHMM)

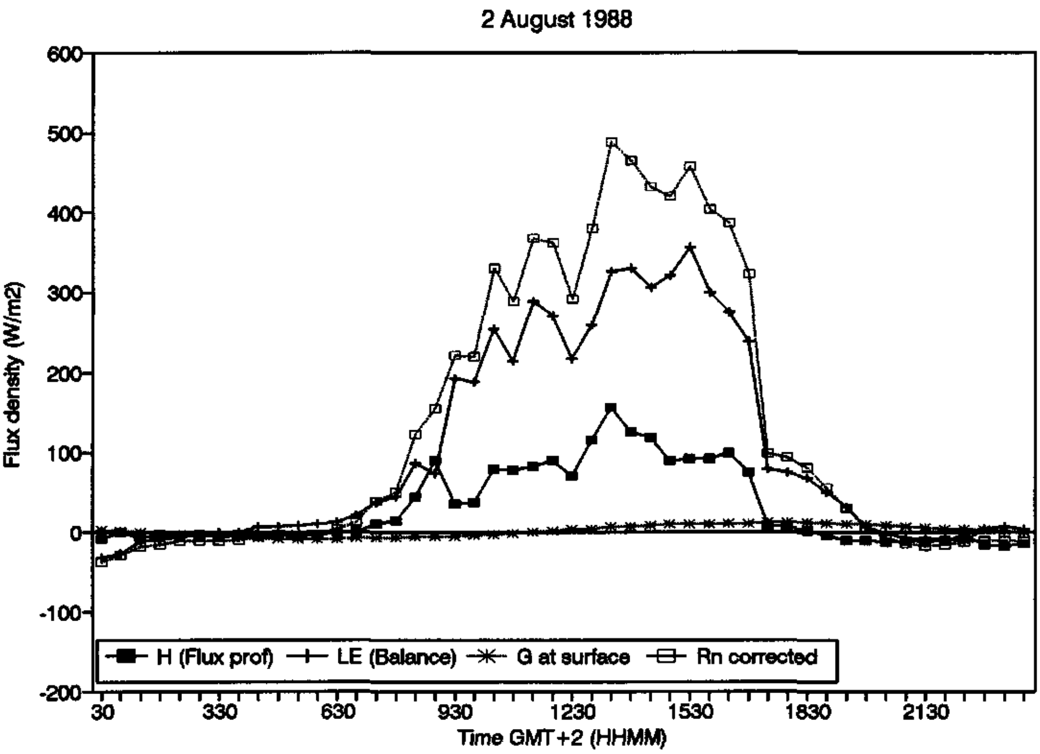

1 June 1988
Flux density (W/m2)
600
500
400
300
200
100
0
-100
-200
H (Flux-profile)
LE (Balance)
G at the surface
Rn corrected
30
330
630
930
1230
1530
1830
2130
Time GMT+2 (HHMM)


15 June 1988
Flux density (W/m2)
600
500
400
300
200
100
0
-100
-200
H (Flux-profile)
LE (Balance)
G at the surface
Rn corrected
30
330
630
930
1230
1530
1830
2130
Time GMT+2 (HHMM)

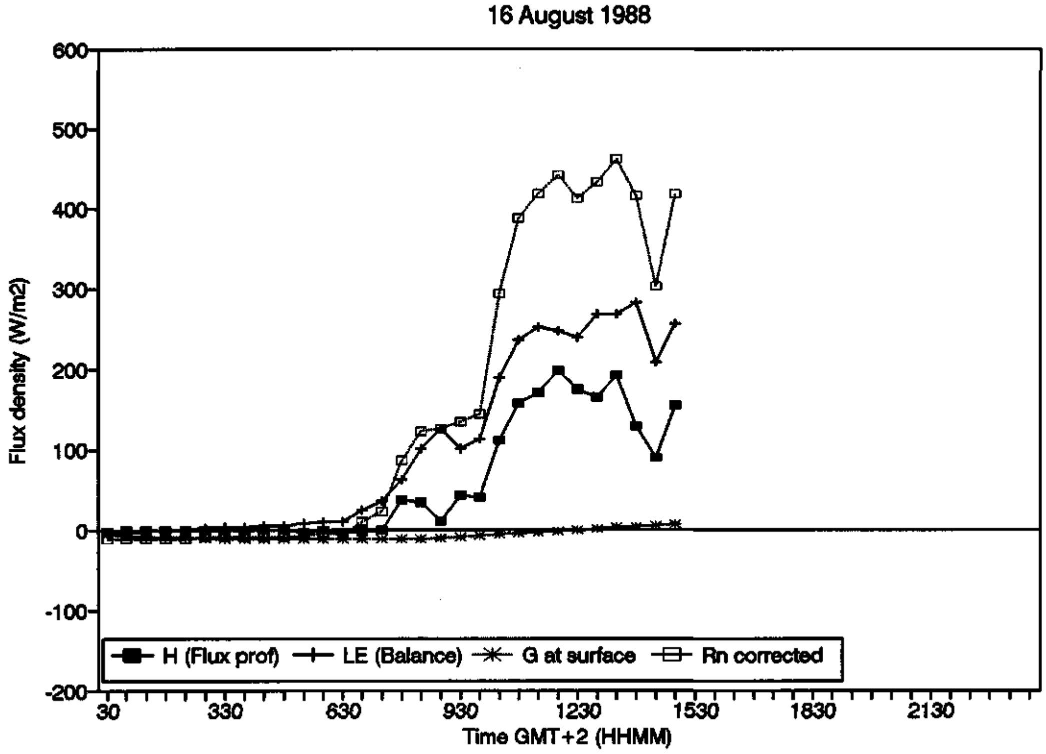

1 July 1988
Flux density (W/m2)
600
500
400
300
200
100
0
-100
-200
H (Flux prof)
LE (Balance)
G at surface
Rn corrected
30
330
630
930
1230
1530
1830
2130
Time GMT+2 (HHMM)

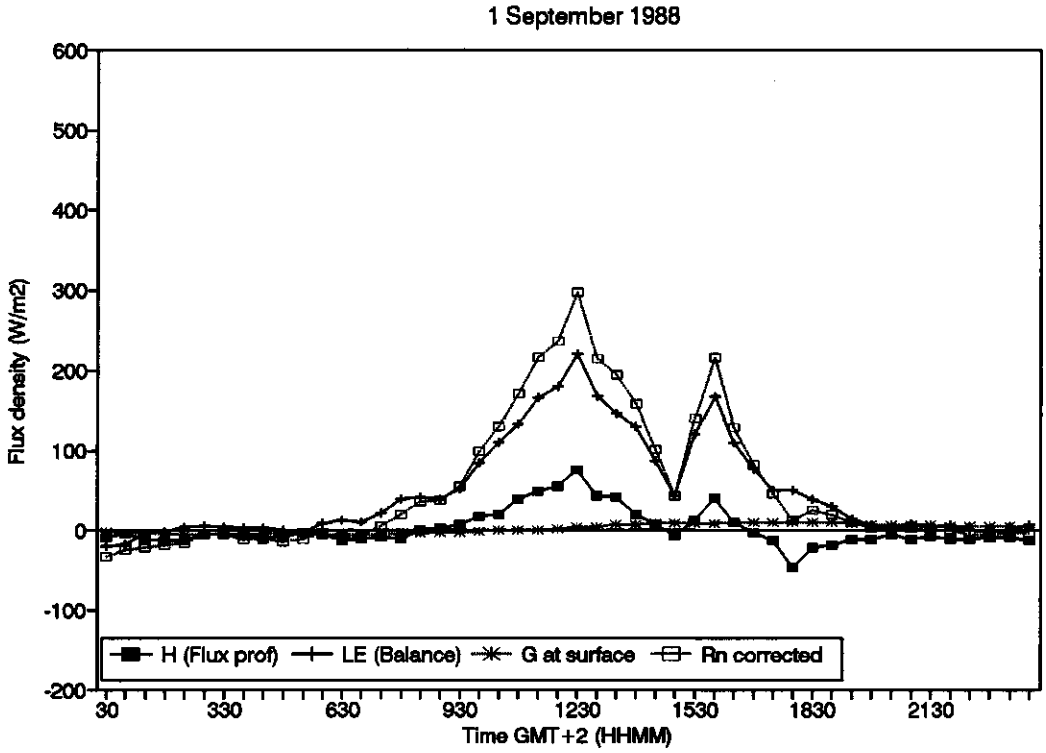


15 July 1988
Flux density (W/m2)
600
500
400
300
200
100
0
-100
-200
H (Flux prof)
LE (Balance)
G at surface
Rn corrected
30
330
630
930
1230
1530
1830
2130
Time GMT+2 (HHMM)

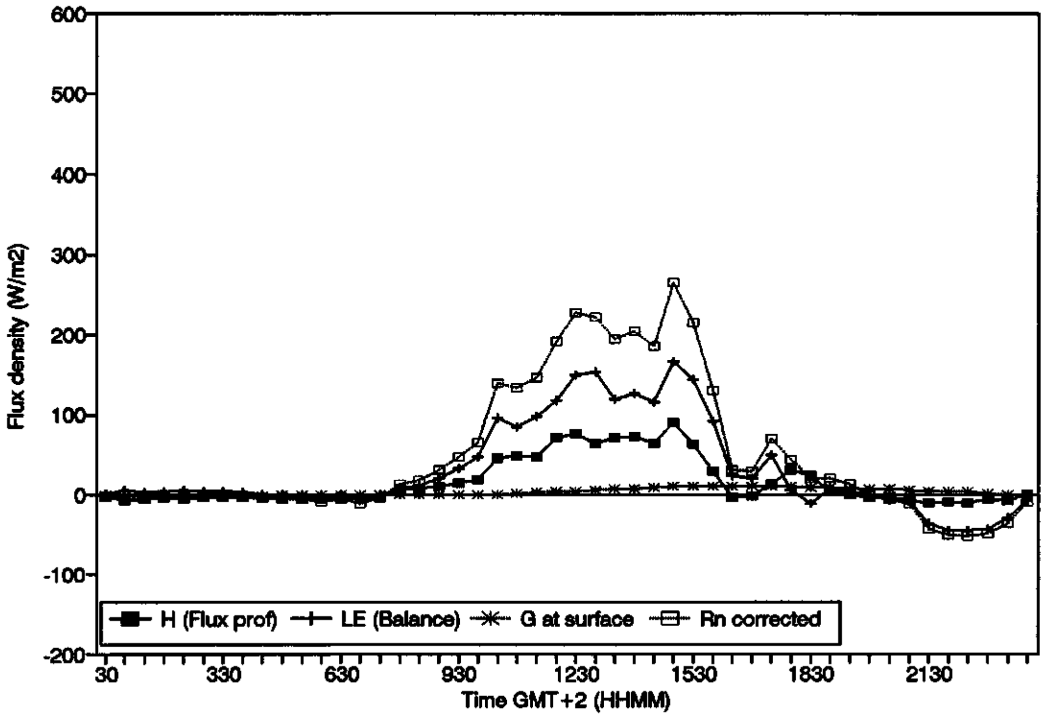


2 August 1988
Flux density (W/m2)
600
500
400
300
200
100
0
-100
-200
H (Flux prof)
LE (Balance)
G at surface
Rn corrected
30
330
630
930
1230
1530
1830
2130
Time GMT+2 (HHMM)

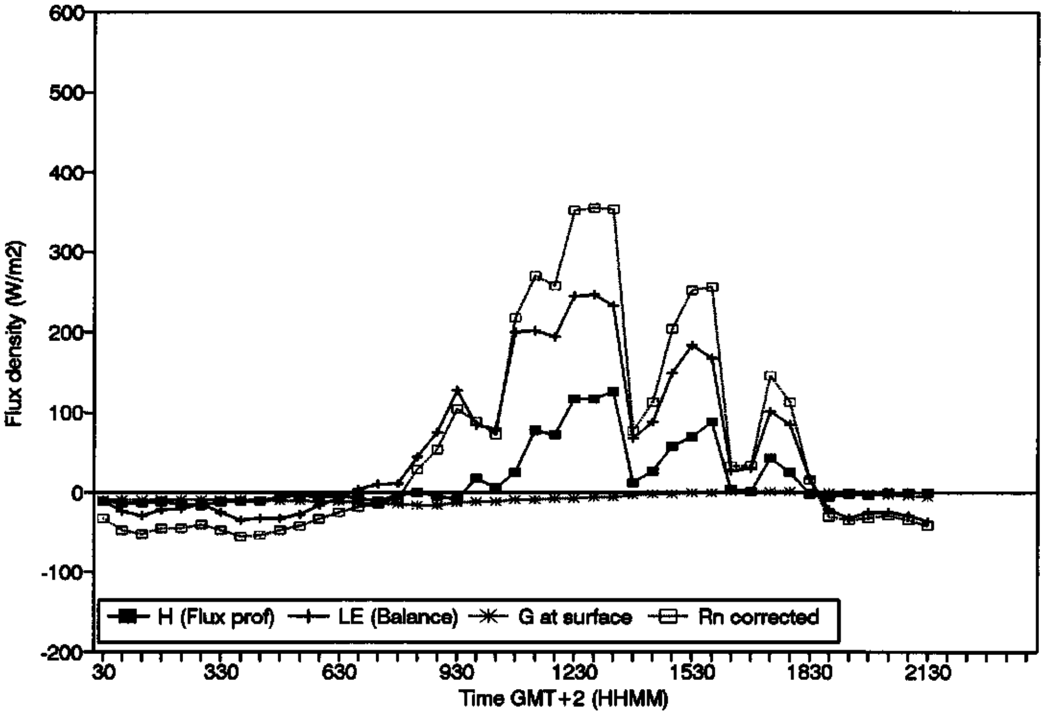

16 August 1988
600
500
400
300
200
100
0
-100
-200
Flux density (W/m2)
H (Flux prof)
LE (Balance)
G at surface
Rn corrected
30
330
630
930
1230
1530
1830
2130
Time GMT+2 (HHMM)


1 September 1988
600
500
400
300
200
100
0
-100
-200
Flux density (W/m2)
H (Flux prof)
LE (Balance)
G at surface
Rn corrected
30
330
630
930
1230
1530
1830
2130
Time GMT+2 (HHMM)

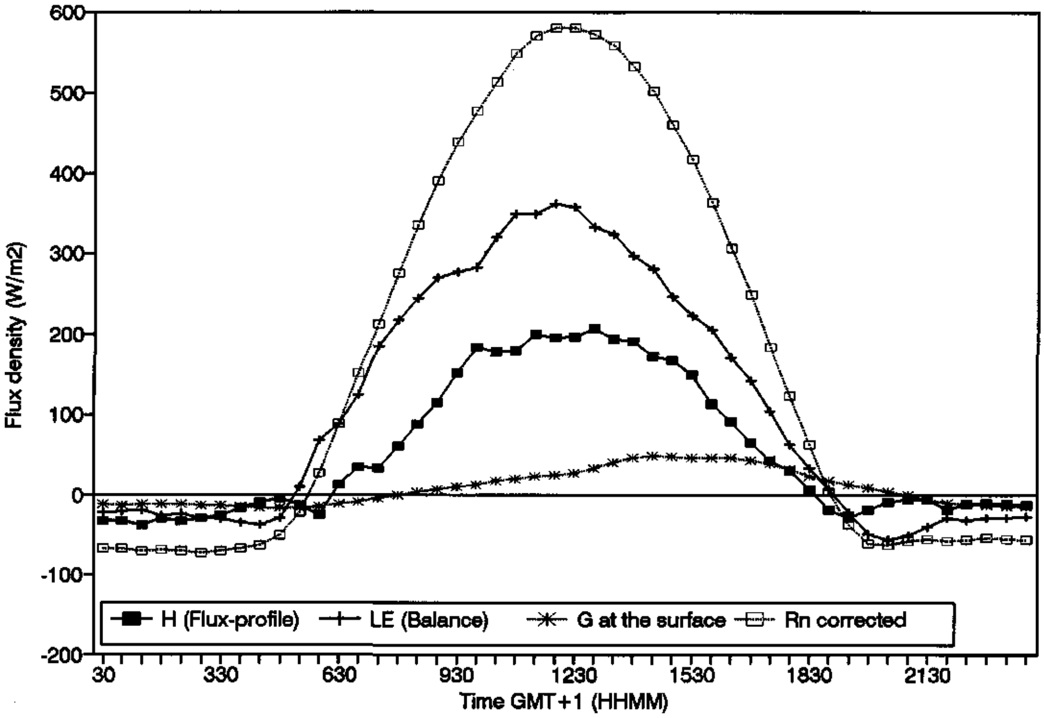
11 September 1988



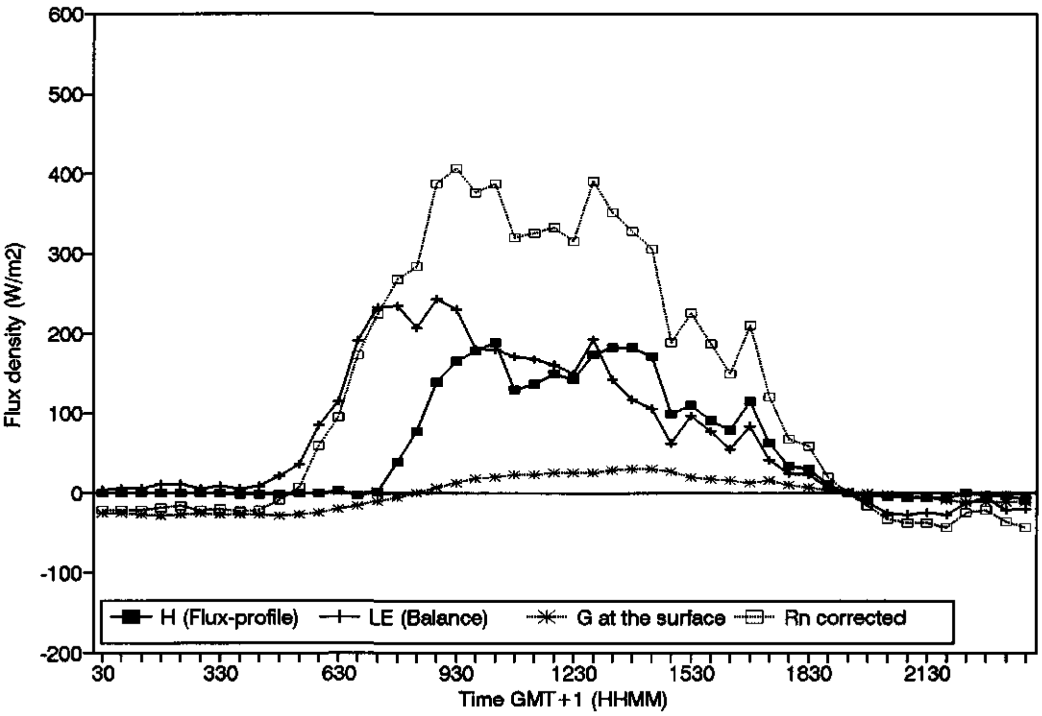
30 September 1988



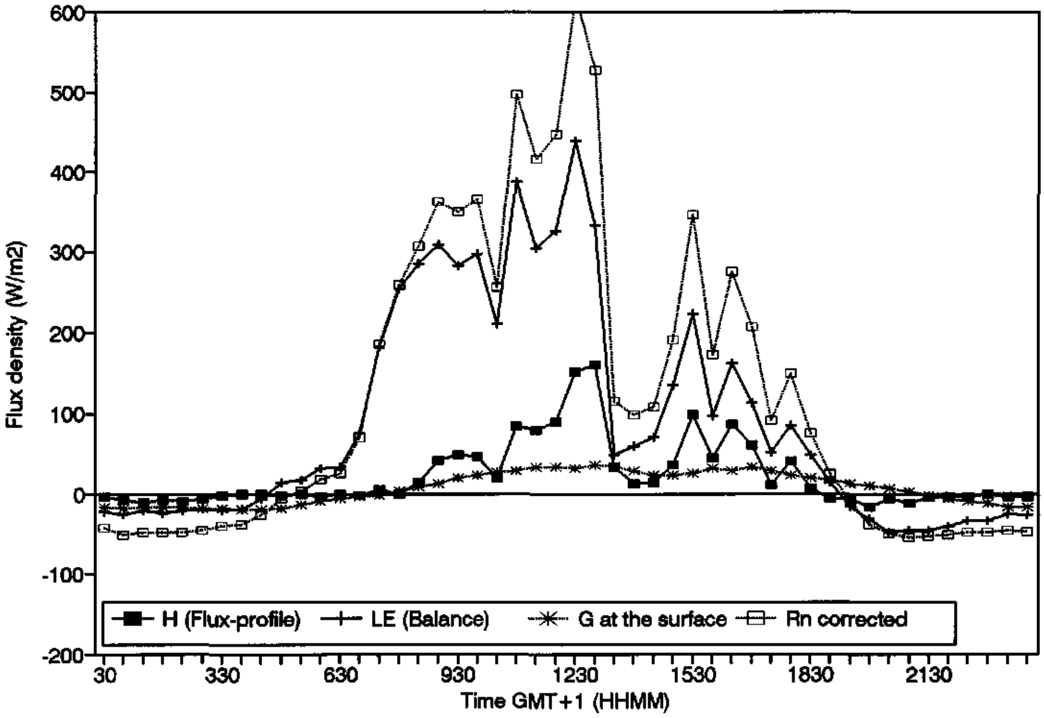
23 May 1989



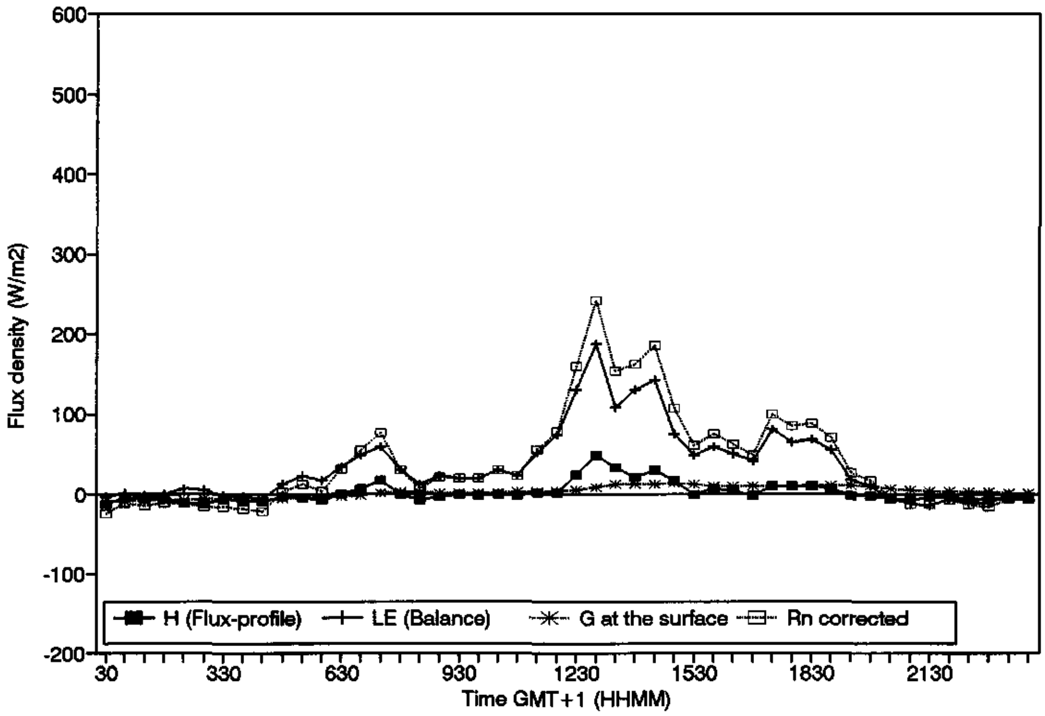
1 June 1989



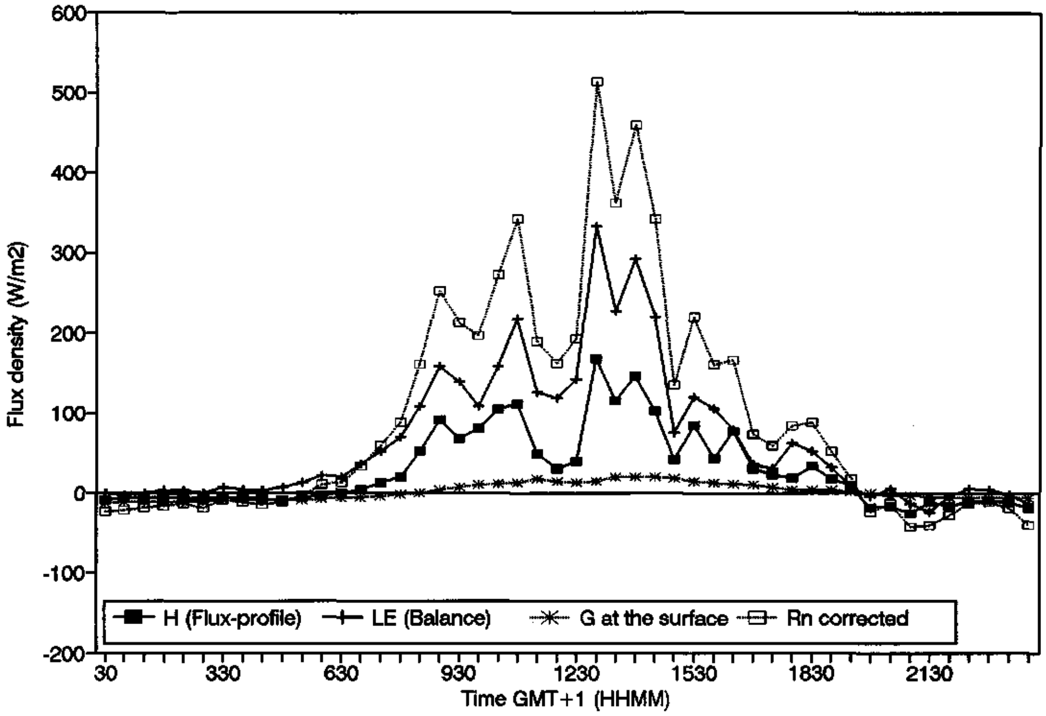
15 June 1989



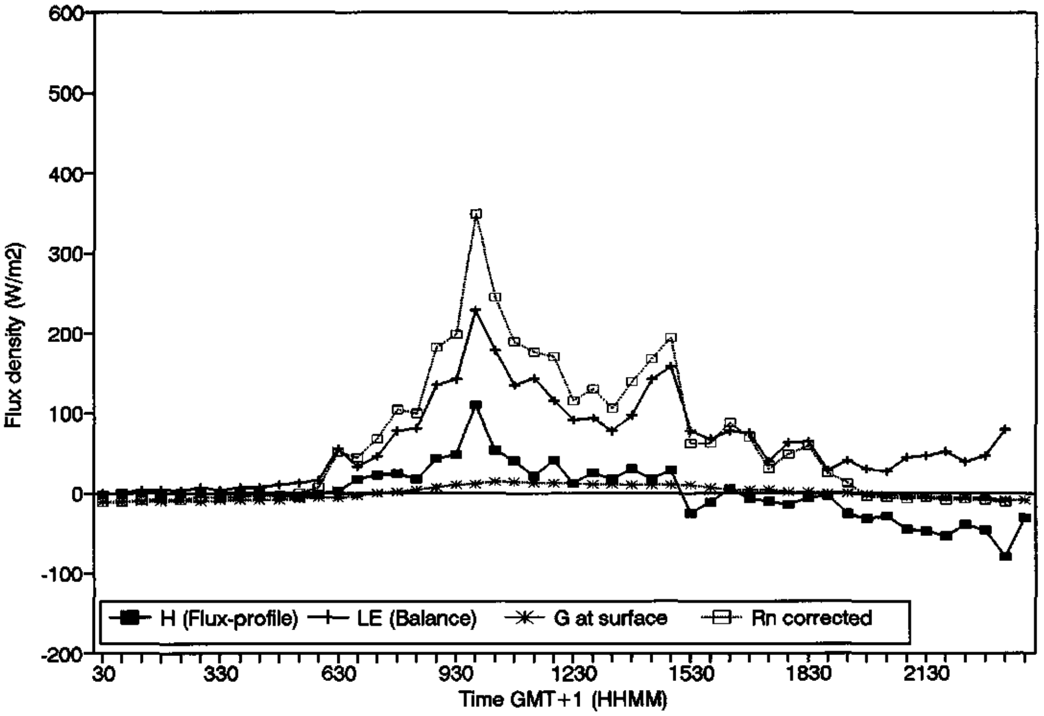
1 July 1989

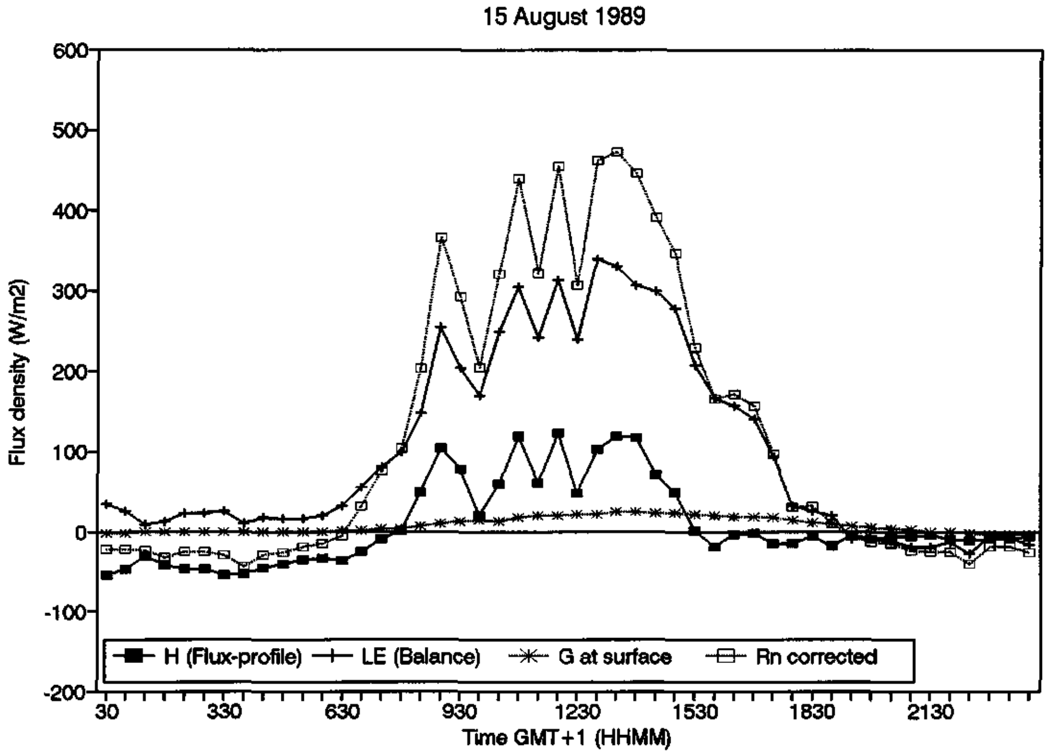


15 July 1989



30 July 1989





15 August 1989
Flux density (W/m2)
600
500
400
300
200
100
0
-100
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30
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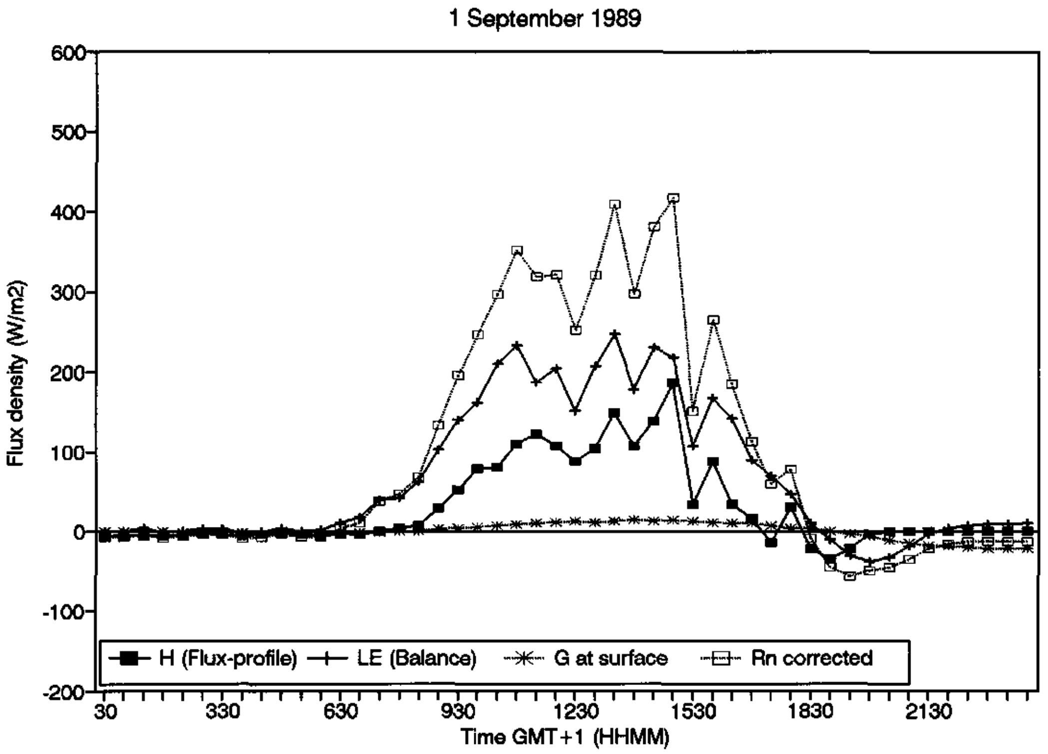
930
1230
1530
1830
2130
Time GMT+1 (HHMM)
H (Flux-profile)
LE (Balance)
G at surface
Rn corrected



1 September 1989
Flux density (W/m2)
600
500
400
300
200
100
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-100
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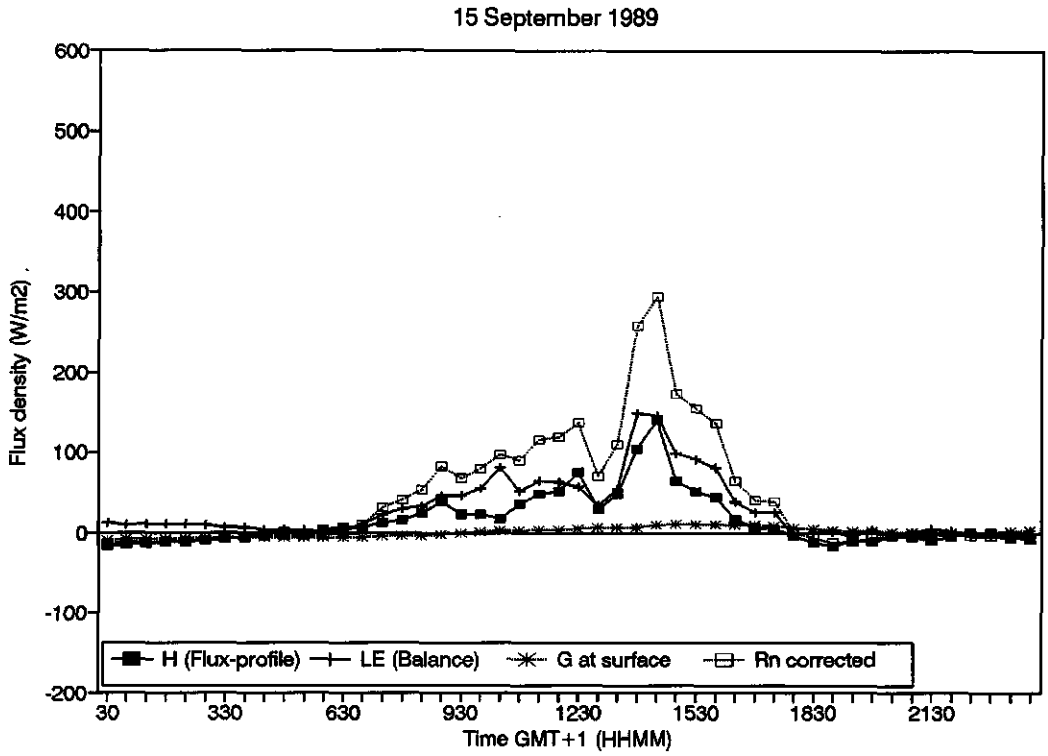
1230
1530
1830
2130
Time GMT+1 (HHMM)
H (Flux-profile)
LE (Balance)
G at surface
Rn corrected



15 September 1989
Flux density (W/m2)
600
500
400
300
200
100
0
-100
-200
H (Flux-profile)
LE (Balance)
G at surface
Rn corrected
30
330
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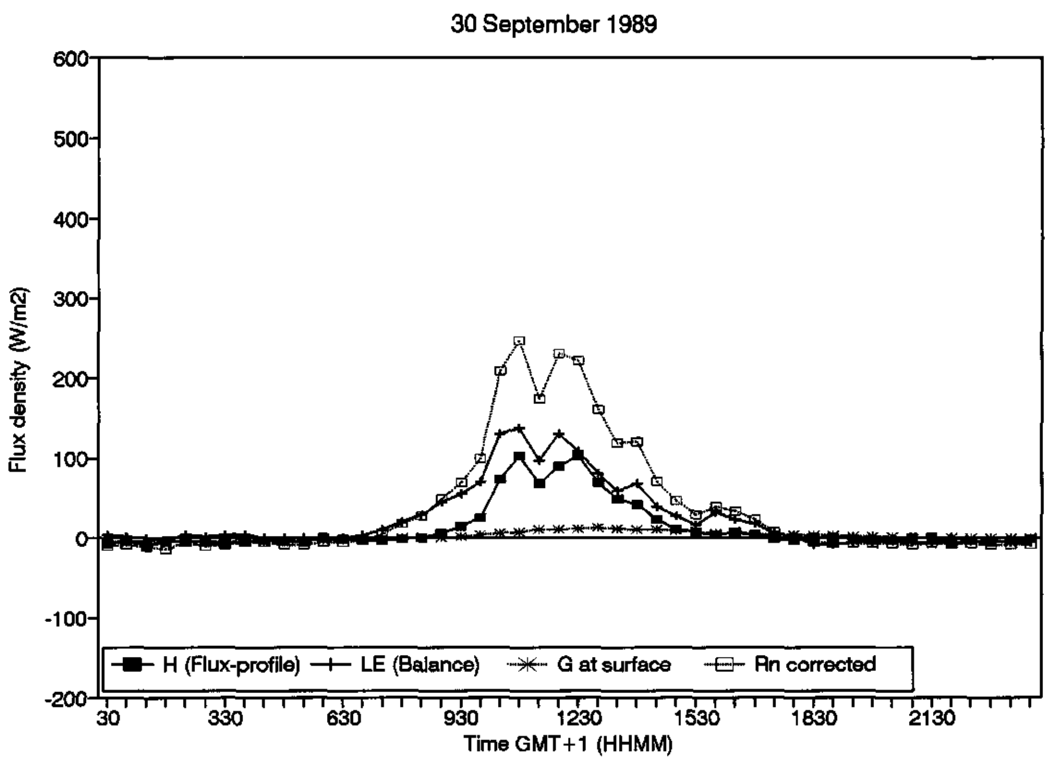
930
1230
1530
1830
2130
Time GMT+1 (HHMM)



30 September 1989
Flux density (W/m2)
600
500
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300
200
100
0
-100
-200
H (Flux-profile)
LE (Balance)
G at surface
Rn corrected
30
330
630
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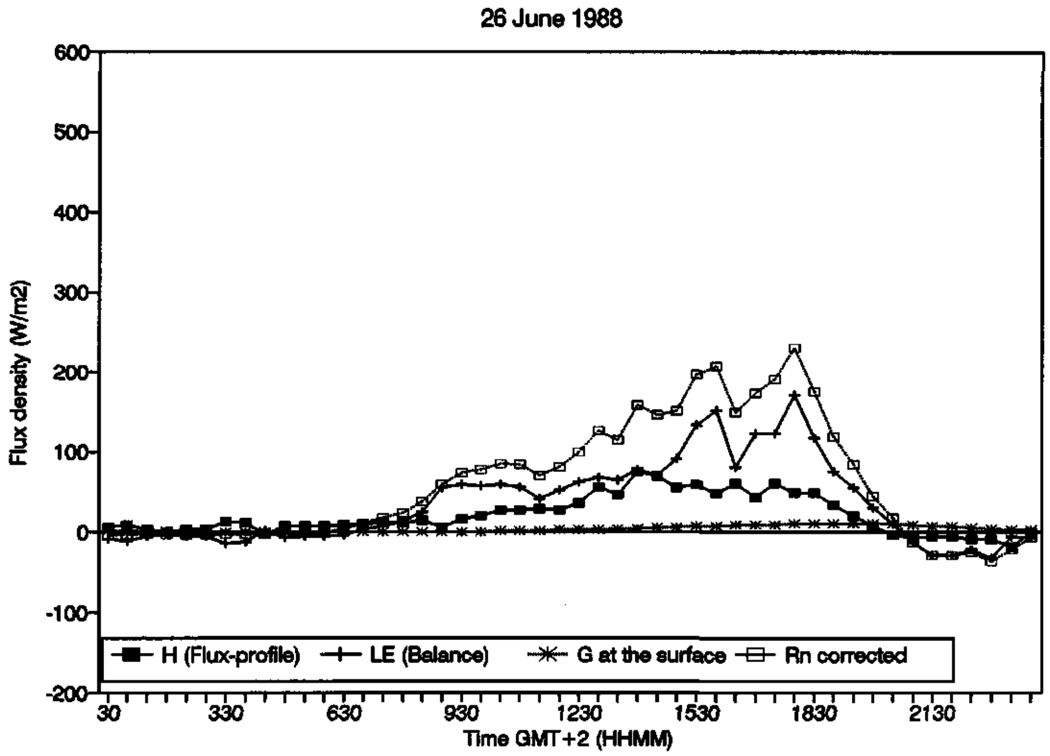
1230
1530
1830
2130
Time GMT+1 (HHMM)


26 June 1988
Flux density (W/m2)
600
500
400
300
200
100
0
-100
-200
H (Flux-profile)
LE (Balance)
G at the surface
Rn corrected
30
330
630
930
1230
1530
1830
2130
Time GMT+2 (HHMM)

Appendix B Tables with the daily precipitation and evapotranspiration data for 1988 and 1989.

Date DDMMYY	PRECIPITATION Start Time	End Time	mm	Daily tot. mmd^{-1}		Flux prof mmd^{-1}	Adj. Pen mmd^{-1}	P&T mmd^{-1}	Mak mmd^{-1}	
210588						3.2				
220588						3.3				
230588						3.1	5.6	4.5	4.2	T
240588	700	800	2.5	3.3		3.1	4.0	3.6	3.2	T
	1030	1130	0.8							
250588				0		3.4	5.1	4.7	4.2	T
260588	1730	1800	0.2	0.2		3.0	4.7	3.6	3.5	T
270588	1930	2230	6.7	6.7		1.1	1.4	1.3	1.3	T
280588	200	330	3.6	3.6		1.1	1.3	1.4	1.1	T
290588	1530	1600	0.2	1		2.0	2.9	2.4	2.2	T
	1800	1900	0.8							
300588	1700	1800	0.8	2.5		2.9	3.6	3.1	2.7	T
	2200	2300	1.7							
310588	1200	1430	7.1	7.5		1.8	2.4	1.4	1.3	T
	1730	1800	0.2							
	2130	2200	0.2							
10688	330	400	0.2	0.8		2.9	3.3	3.0	2.6	T
	530	600	0.2							
	700	730	0.2							
	1100	1130	0.2							
20688	2030	2200	1.6	1.6		2.1	2.7	2.7	2.2	T
30688	100	130	0.2	0.6		3.4	4.0	3.9	3.2	T
	300	330	0.2							
	530	600	0.2							
40688	2400	100	3.6	4.4			2.5	2.5	2.0	
	200	300	0.4							
	1330	1400	0.2							
	1530	1600	0.2							
50688				0	V					
60688				0	V					
70688	800	830	0.4	1.2		0.7	0.6	0.7	0.6	T
	900	930	0.2							
	1130	1200	0.2							
	1230	1300	0.2							
	1630	1700	0.2							
80688	1100	1300	2.7	2.7		1.1	1.2	1.2	1.0	T
90688	1330	1700	5	6.9			0.8	0.9	0.8	
	1730	1900	1.7							
	2200	2230	0.2							
100688	530	600	0.2	0.2		1.1	1.5	1.6	1.4	T
110688				0		1.2	1.7	1.4	1.2	T
120688				0		3.6	6.8	5.5	4.8	T
130688				0		4.3	6.1	5.5	4.8	T
140688				0			5.5	5.5	4.5	

Date DDMMYY	PRECIPITATION Start Time	End Time	mm	Daily tot. mmd^{-1}	Flux prof mmd^{-1}	Adj. Pen mmd^{-1}	P&T mmd^{-1}	Mak mmd^{-1}	
150688				0	2.1	3.1	2.9	2.3	
160688				0	1.6	2.5	1.9	1.6	
170688				0	3.2	4.2	4.0	3.4	
180688				0	1.6	2.4	2.0	1.7	
190688				0	2.2	3.2	2.6	2.1	
200688	500	600	0.4	0.8	2.2	2.5	2.4	1.9	
	1000	1100	0.4						
210688				0	3.0	4.0	3.9	3.2	
220688	930	1100	1	1	3.1	4.1	3.5	2.8	
230688				0	2.2	3.3	2.6	2.2	
240688	300	330	0.2	0.2	1.2	1.6	1.3	1.0	
250688	300	330	0.2	0.4	1.9	2.3	2.3	1.7	
	530	600	0.2						
260688				0	1.2	1.6	1.6	1.3	
270688	2030	2130	8.1						
280688	1130	1200	0.2	0.2	2.9	3.4	3.5	3.0	T
290688				0	2.3	2.8	2.8	2.4	T
300688	2000	2230	9.8	9.8	3.4	4.0	4.0	3.3	T
10788	1000	1130	5.5	11.9	2.1	3.0	2.2	1.9	T
	1730	1800	3						
	2000	2130	3.4						
20788	1230	1300	3	7.2	3.0	2.7	3.0	2.6	T
	1600	1700	0.6						
	1930	2030	3.6						
30788	200	330	1	7	3.7	4.1	3.6	2.7	T
	1500	1600	0.6						
	1900	2230	5.2						
40788	2330	2400	0.2	0.8	4.3	4.8	4.1	3.5	T
	1700	1800	0.8						
50788	800	1230	7.6	11.3	1.1	1.8	1.0	1.0	T
	1400	1430	0.2						
	1830	2200	3.5						
60788	530	600	0.2	0.2	4.3	3.9	4.6	3.7	T
70788	330	400	0.2	6.2	2.7	2.9	2.5	2.2	T
	530	600	0.2						
	1230	1400	0.6						
	1630	1800	5.2						
80788	1730	1800	0.2	3.2	2.8	3.1	2.8	2.3	T
	2200	2330	3						
	2400	30	0.2						
	100	130	0.2						
	400	500	0.8						
	1430	1500	0.4						
90788				1.6	2.8	3.1	2.8	2.4	T

Date DDMMYY	PRECIPITATION Start Time	End Time	mm	Daily tot. mmd^{-1}		Flux prof mmd^{-1}	Adj. Pen mmd^{-1}	P&T mmd^{-1}	Mak mmd^{-1}	
100788				0		4.2	4.3	4.5	3.7	T
110788				0						
120788	530	600	0.2	0.8		2.6	2.7	2.8	2.3	
	1700	1730	0.6							
130788	1230	1300	0.2	3.5		1.8	1.5	1.3	1.1	
	1700	1800	0.8							
	2200	2230	0.6							
	2300	2400	1.9							
140788	1130	1330	3.9	9.3		1.8	1.3	1.3	1.1	
	1400	1700	4.6							
	1730	1800	0.8							
150788	830	1000	2.5	9.9		2.4	1.9	1.8	1.5	
	1330	1500	6.1							
	1630	1700	1.3							
160788	30	100	0.4	36.4		1.4	1.1	1.1	0.8	
	130	1030	9.3							
	1100	1530	20.3							
	1630	1830	5.8							
	2030	2200	0.6							
170788	1830	1930	0.4	0.4		2.4	2.1	2.2	1.8	
180788	100	230	1.8	2		1.8	1.4	1.3	1.2	
	1800	1830	0.2							
190788				0		2.3	2.5	2.6	2.2	
200788	2130	2200	0.2	0.2		2.2	2.0	2.0	1.7	
210788	830	930	0.4	0.4		1.7	1.5	1.6	1.4	
220788	200	230	0.2	6.8		1.8	1.2	1.1	1.0	
	1000	1100	3							
	1130	1200	0.6							
	1300	1330	1.3							
	1600	1700	1.5							
	2000	2030	0.2							
230788	330	500	2.5	4.2						
	600	700	1.5							
	730	800	0.2							
240788				0	V					
250788				0	V					
260788					V					
270788					V					
280788					V					
290788					V					
300788					V					
310788					V					
10888					V					
20888	2300	2400	0.4	0.4		3.5	3.6	3.6	2.9	

Date DDMMYY	PRECIPITATION Start Time	End Time	mm	Daily tot. mmd^{-1}		Flux prof mmd^{-1}	Adj. Pen mmd^{-1}	P&T mmd^{-1}	Mak mmd^{-1}
30888				0		3.3	3.5	3.5	3.0
40888	1930	2000	0.2	0.2		2.0	2.1	1.8	1.4
50888				0		3.1	3.4	3.5	2.7
60888				0		3.6	4.0	4.2	3.5
70888				0		3.6	4.1	4.4	3.7
80888				0			3.6	3.4	3.0
90888					V				
100888					V				
110888					V				
120888					V				
130888					V				
140888					V				
150888					V				
160888				0	V				
170888				0	V				
180888	730	800	0.2	0.2		3.8	4.6	4.4	3.5
190888	430	530	0.4	2		1.7	2.3	1.3	1.4
	700	730	0.2						
	930	1130	1.2						
	1230	1300	0.2						
200888	1500	1600	5.3	5.5		2.2	2.9	2.2	2.1
	2130	2200	0.2						
210888	2400	130	2.1	15.4		1.3	1.3	1.2	1.0
	500	600	0.4						
	630	700	0.2						
	730	1100	3.9						
	1330	1500	1.9						
	1530	1700	4.6						
	1730	1830	1.7						
	2000	2030	0.6						
220888	2400	100	2.3	4.6		2.2	2.5	2.2	2.0
	400	530	1.2						
	1400	1430	1.1						
230888				0		1.9	2.3	1.9	1.8
240888	1600	1630	0.2	3.5		1.3	1.5	1.1	1.0
	1730	1930	3.3						
250888	530	700	4	9.5		2.7	2.3	2.2	1.7
	730	900	1.4						
	1330	1400	0.2						
	1530	1900	3.9						
260888	30	100	0.6	4.5		2.2	2.0	1.7	1.5
	400	430	0.2						
	1030	1200	1.6						
	1500	1530	0.6						

Date DDMMYY	PRECIPITATION Start Time	End Time	mm	Daily tot. mmd^{-1}		Flux prof mmd^{-1}	Adj. Pen mmd^{-1}	P&T mmd^{-1}	Mak mmd^{-1}	
	1600	1630	1.5							
270888	200	230	0.2	1.2		1.1	1.1	1.2	0.9	
	830	1000	0.6							
	1030	1130	0.4							
280888	1900	2030	13.3	13.3		3.6	3.8	3.4	2.9	
290888				0			2.5	1.9	1.9	
300888	530	600	0.2	0.2		1.7	2.9	2.2	2.1	
310888	2000	2100	0.4	0.4		2.1	2.8	2.4	2.2	T
10988	1800	2000	1.8	2.8		1.7	2.1	1.4	1.3	T
	2030	2100	0.2							
	2300	2400	0.8							
20988	1100	1130	0.2	0.2		1.6	2.6	1.6	1.9	T
30988				0		1.2	1.8	1.5	1.6	T
40988	730	800	0.2	0.2		2.1	2.5	2.5	2.1	T
50988	1230	1500	3.9	3.9		1.0	1.0	1.1	1.0	T
60988				0		2.7	3.1	3.3	2.8	T
70988	730	800	0.2	0.2		3.0	3.5	3.3	2.9	T
80988				0		2.5	3.3	2.7	2.8	T
90988				0		2.5	2.9	2.8	2.7	T
100988				0		0.8	1.1	1.2	1.0	T
110988	830	900	0.2	1		1.1	1.5	1.3	1.2	T
	1730	1830	0.8							
120988	630	700	0.2		V					
130988					V					
140988					V					
150988					V					
160988					V					
170988					V					
180988					V					
190988					V					
200988				0		0.6	0.8	0.6	0.5	T
210988				0		0.7	1.0	0.8	0.7	T
220988				0		2.2	2.4	2.1	2.0	T
230988	630	700	0.2	15.9			1.7	0.8	0.9	
	1330	1800	14.9							
	1930	2000	0.2							
	2100	2230	0.6							
240988	1100	1200	0.4	6.4			1.4	0.9	0.6	
	1230	1430	5.6							
	1800	1830	0.2							
	1930	2000	0.2							
250988	2030	2100	0.2	0.2			1.5	1.7	1.3	
260988	1200	1230	0.2	3			0.9	0.5	0.3	
	1330	1400	0.2							

Date DDMMYY	PRECIPITATION Start Time	End Time	mm	Daily tot. mmd^{-1}	Flux prof mmd^{-1}	Adj. Pen mmd^{-1}	P&T mmd^{-1}	Mak mmd^{-1}	
	1530	1630	2.2						
	1930	2000	0.2						
	2100	2130	0.2						
270988	500	600	0.6	2.8	1.0	1.1	0.8	0.5	T
	630	900	1.8						
	1200	1230	0.2						
	2330	2400	0.2						
280988	30	100	0.2	3.7	1.3	1.6	0.7	0.7	T
	400	430	0.2						
	800	1030	3.3						
290988				0	1.7	2.1	1.3	1.7	T
300988	800	830	0.2	0.2		1.6	1.5	1.5	
11088									
21088									
31088									
41088									
51088	1930	2030	1.6						
	2100	2130	1.1						
61088	1130	1800	9.1	9.9					
	1830	2000	0.8						
71088	500	630	1.2	5.1					
	700	900	2.1						
	1100	1200	0.8						
	1730	1900	0.8						
	2300	2330	0.2						
81088	100	400	4.4	9.9					
	500	1030	5.3						
	1100	1130	0.2						
91088	330	400	0.6	0.6					

T = relative humidity data used of KNMI station Twente Vb
V = precipitation measured at KNMI rainfall station Vroomshoop

Date DDMMYY	PRECIPITATION Start Time	End Time	mm	Daily Total mmd^{-1}		Flux prof mmd^{-1}	Adj. Pen mmd^{-1}	P&T mmd^{-1}	Mak mmd^{-1}	
210589				0						
220589				0						
230589				0		3.9	6.8	4.9	5.1	T
240589				0		3.8	6.3	4.9	5.2	T
250589				0						
260589				0						
270589				0						
280589				0						
290589				0						
300589	800	830	0.2	0.2		1.5	2.7	1.9	2.1	T
310589				0		1.6	2.4	2.1	2.2	T
10689				0		2.6	3.3	3.1	2.7	T
20689				0		2.0	3.0	2.6	2.6	T
30689	1430	1500	0.2	4.4			2.1	2.0	2.0	
	1530	1600	0.2							
	1630	2030	2.9							
	2100	2130	0.2							
	2230	2330	0.9							
40689	130	230	0.6	1			1.1	1.3	1.2	
	400	500	0.4							
50689				11.1	V					
60689	2000	2030	0.2	0.2						
70689	2400	30	0.2	2.5		1.1	1.6	1.5	1.6	
	600	630	0.2							
	730	800	0.2							
	830	1100	1.9							
80689	930	1000	0.9	1.7		2.3	2.4	2.5	2.3	
	1100	1130	0.4							
	1900	1930	0.4							
90689	130	300	4.4	10.3	V					
	330	400	0.2							
100689				0		3.2	4.2	3.8	3.6	
110689				0		2.6	3.5	3.2	3.4	
120689				0		3.5	5.5	4.9	5.0	
130689				0		4.3	6.8	5.4	5.4	T
140689				0		3.9	5.7	4.8	4.9	T
150689				0		3.2	4.1	3.4	3.6	T
160689				0		4.8	5.7	5.2	5.1	T
170689				0		3.5	6.0	5.2	5.1	T
180689				0		3.8	5.2	4.9	4.7	T
190689				0		4.7	5.9	5.3	5.1	T
200689				0		4.6	6.0	5.8	5.2	T
210689				0	V					
220689	1630	1700	2.2	2.2						

	PRECIPITATION			Daily	Flux	Adj.			
Date	Start	End		Total	prof	Pen	P&T	Mak	
DDMMYY	Time	Time	mm	mmd^{-1}	mmd^{-1}	mmd^{-1}	mmd^{-1}	mmd^{-1}	
230689				0	1.9	3.2	2.8	2.8	T
240689				0	3.2	4.1	3.9	3.7	T
250689				0	4.3	5.5	5.3	5.0	T
260689	2200	2330	1.5	1.5		6.3	5.3	5.0	
270689	2400	30	0.2	1.1	2.9	3.1	2.7	2.7	
	100	130	0.2						
	1600	1630	0.7						
280689	30	400	6.3	7.1	3.4	4.1	3.2	2.9	
	430	500	0.2						
	900	930	0.2						
	1030	1100	0.4						
290689	930	1000	0.2	11.2	1.0	1.2	0.8	0.6	
	1030	1100	0.2						
	1200	1930	10.8						
300689	130	200	0.2	0.2		4.9	4.8	4.2	
10789	800	1200	11.1	19.1	1.2	1.2	1.0	1.0	
	1300	1400	0.6						
	1500	1600	2.4						
	1830	1900	0.4						
	2130	2200	4.6						
20789	1600	1630	0.4	0.4	2.2	2.3	2.1	1.9	
30789				0	4.7	5.6	4.9	4.6	
40789				0	5.1	5.8	5.1	4.7	
50789				0	4.8	5.6	4.9	4.8	
60789				0	5.1	5.7	5.2	5.0	T
70789				0	4.3	5.9	4.7	4.5	T
80789	1900	2000	6.1	6.1	3.3	4.2	3.5	3.1	
90789	430	600	1.1	10	1.9	2.2	1.8	1.5	
	700	730	0.2						
	1230	1430	8.7						
100789				0	2.0	1.9	2.0	1.9	T
110789				0	3.2	3.1	3.3	2.8	T
120789				0	3.6	4.3	4.3	3.6	
130789				0	2.5	3.5	3.0	2.7	
140789	1330	1400	0.2	0.2	2.3	3.2	2.6	2.3	
150789				0	2.3	3.3	2.6	2.3	
160789				0	2.7	4.6	4.0	3.7	
170789				0	1.8	2.2	2.0	1.9	
180789				0	2.5	4.2	3.6	3.2	
190789				0	2.7	3.7	3.4	3.1	
200789				0	3.3	2.9	3.1	2.9	
210789	530	600	0.2	0.2	5.3	4.9	4.7	4.4	
220789				0	4.3	4.8	3.8	3.8	
230789	730	900	2.4	2.4	2.4	2.5	2.3	2.2	

Date DDMMYY	PRECIPITATION Start Time	End Time	mm	Daily Total mmd^{-1}		Flux prof mmd^{-1}	Adj. Pen mmd^{-1}	P&T mmd^{-1}	Mak mmd^{-1}
240789				0		2.8	3.2	3.4	3.0
250789				0			3.9	4.1	3.6
260789	1500	1530	0.7	0	V				
	1600	1630	0.2						
270789				0		3.1	3.8	3.5	3.3
280789				0		2.0	2.4	2.2	2.0
290789				0		3.1	3.6	3.3	3.2
300789	1500	1630	1.3	17.1			2.2	1.8	1.6
	1700	1800	0.6						
	1830	1900	0.4						
310789	2030	430	23.1	13.4					2.6
	1000	1030	0.2						
	1100	1130	0.7						
	1230	1300	1.3						
	1400	1430	0.2						
	1730	1800	0.7						
10889	2300	30	2.2	7.1					2.8
	300	330	0.2						
	900	930	0.9						
	1230	1330	1.1						
	1400	1430	1.5						
	1500	1530	2.8						
	1600	1700	0.4						
20889	230	330	0.4	0.6					2.3
	1130	1200	0.2						
30889	1800	1830	0.2	0.2					1.6
40889				0					2.7
50889				0					3.3
60889				0		6.0	5.2	5.3	4.0
70889	1200	1400	1.9	1.9			1.6	1.5	1.5
80889	630	700	0.2	3.9		1.6	1.6	1.7	1.7
	1230	1330	3.3						
	1400	1430	0.4						
90889	230	300	0.2	0.4		2.7	3.2	3.3	3.0
	1530	1600	0.2						
100889				0		3.6	4.3	3.9	3.7
110889	1030	1130	0.4	1		1.8	1.6	1.1	1.0
	1200	1300	0.6						
120889	1630	1700	0.2	0.2		1.6	2.0	1.7	1.6
130889	500	530	0.2	0.2		3.0	3.8	3.6	3.3
140889				0		3.2	4.2	3.4	3.2
150889				0		3.6	4.9	3.4	3.2
160889	1800	1830	0.2	4.8		3.6	3.8	3.6	3.1
170889	2230	30	4.8	0.4		3.3	3.8	3.5	3.3

Date DDMMYY	PRECIPITATION Start Time	End Time	mm	Daily Total mmd^{-1}	Flux prof mmd^{-1}	Adj. Pen mmd^{-1}	P&T mmd^{-1}	Mak mmd^{-1}	
	500	530	0.2						
180889				0	2.8	3.5	3.2	3.2	
190889				0	3.7	4.4	4.1	3.6	
200889				0	3.5	4.3	4.0	3.7	
210889				0	3.2	4.7	4.2	3.6	T
220889				0	2.5	3.5	3.1	2.9	T
230889				0					
240889				0	2.5	3.3	2.8	2.6	
250889	730	800	0.2	1.3	2.0	2.6	2.2	2.0	
	830	930	1.1						
260889	1830	1900	0.2	0.2		3.1	2.8	2.2	
270889	2400	800	21.5	32.9		0.9	0.7	0.6	
	830	1230	10.5						
	1500	1530	0.2						
	2230	2300	0.7						
280889	100	230	0.6	0.8	2.6	3.1	2.3	2.2	
	1400	1430	0.2						
290889				0	2.1	2.7	2.4	2.2	
300889	530	600	0.2	0.4	0.6	0.5	0.4	0.4	
	930	1000	0.2						
310889	1800	1830	0.2	4.5	1.1	1.3	1.1	1.0	
	1900	2030	1.5						
	2100	2400	2.8						
10989	1030	1100	0.2	0.2	2.3	2.7	2.7	2.4	
20989	730	900	2.1	5.2	1.3	1.0	1.1	0.8	
	930	1030	1.7						
	1500	1530	0.7						
	1600	1630	0.7						
30989	1300	1330	0.2	0.2		2.5	2.7	2.2	
40989				0		1.8	1.8	1.5	
50989				0	1.4	1.7	1.6	1.4	
60989				0		2.3	2.2	1.9	
70989	400	430	0.2	0.2	1.8	2.7	2.8	2.6	
80989				0	2.6	3.1	2.7	2.8	
90989				0	1.9	2.8	2.2	2.6	
100989				0	1.4	2.5	1.8	2.0	
110989	200	230	0.2	0.2	1.5	2.8	2.1	2.3	
120989				0	1.5	2.7	2.2	2.5	
130989	900	1000	1.1	9	0.5	0.8	0.7	0.7	
	1530	1930	7.3						
	2030	2100	0.2						
	2230	2330	0.4						
140989	300	400	0.6	1.4	1.8	2.3	1.8	1.8	
	800	830	0.2						

Date DDMMYY	PRECIPITATION Start Time	End Time	mm	Daily Total mmd^{-1}		Flux prof mmd^{-1}	Adj. Pen mmd^{-1}	P&T mmd^{-1}	Mak mmd^{-1}	
	900	1000	0.6							
150989	30	100	0.2	1.7		1.0	1.2	1.2	1.0	
	200	230	0.2							
	330	500	1.1							
	1930	2000	0.2							
160989	300	400	1.6	8.1		1.4	1.8	1.6	1.5	
	1100	1200	4.6							
	1930	2030	1.7							
	2130	2200	0.2							
170989				0		1.9	3.0	2.5	2.6	
180989				0		2.3	3.0	2.6	2.6	
190989	830	1030	4.1	3	V					
200989				4.7						
210989				0						
220989				0						
230989				0						
240989				1.2						
250989				12.8						
260989				0.4						
270989				0.2						
280989	1830	1900	0.2	0.4						
290989	2000	2030	0.2	0.2		0.8	0.9	0.8	0.9	T
300989	30	100	0.2	1.8		0.8	0.9	1.0	0.9	
	1730	1930	0.8							
	2030	2200	0.6							
	2230	2300	0.2							
11089	300	400	0.4	0.8		0.9	4.4	0.8	0.8	
	1900	1930	0.2							
	2030	2100	0.2							
21089				0		1.2	5.1	1.0	1.0	T

T = relative humidity data used of KNMI station Twente Vb

V = precipitation measured at KNMI rainfall station Vroomshoop