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ESTIMATION OF REGIONAL EVAPOTRANSPIRATION AND SOIL MOISTURE CONDITIONS USING REMOTELY SENSED CROP SURFACE TEMPERATURES

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PREFACE

This study was effected in the framework of the NIWARS investigations on the application of thermal infrared scanning. It concerns the application of thermal infrared scanning for the measurement of aerial heat and water budgets of cropped surfaces. The study was executed in collaboration with the Department of Hydrology of the Instituut voor Cultuurtechniek en Waterhuishouding on detachment to this institute.

The author is indebted to all co-operators of NIWARS and ICW who made the execution of measurements and processing possible. He expresses his gratitude to Ir. A. Rosema and Dr. R.A. Feddes for their permanent support in physical matters.

It was the intention to have this study published as NIWARS publication 45. As the redaction was not finished before the premature end of all NIWARS activities, it is published provisionally as a note of the ICW.

LIST OF USED SYMBOLS

Symbol	Description	Units
b	root density resistance factor	m
c_p	specific heat per unit mass of air	$J.kg^{-1}.K^{-1}$
Ch6	value of band 6 of Bendix M^2S scanner (0.64 ± 0.02 µm)	-
Ch9	value of band 9 of Bendix M^2S scanner (0.81 ± 0.05 µm)	-
d	zero displacement	m
e_a	actual water vapour pressure in the air at height z_a	Pa
e_c	saturated water vapour pressure in the air at temperature T_c	Pa
E	evapotranspiration flux	$kg.m^{-2}.s^{-1}$
g	acceleration due to gravity	$m.s^{-2}$
G	heat flux into the soil	$W.m^{-2}$
h	crop height	m
H	sensible heat flux	$W.m^{-2}$
k	Von Karman's constant ($k = 0.4$)	-
K	unsaturated hydraulic conductivity	$m.s^{-1}$
K_s	saturated hydraulic conductivity	$m.s^{-1}$
L	latent heat of vaporization	$J.kg^{-1}$
m	pore size distribution factor	-
P_1	stability correction for momentum transport	-
P_2	stability correction for heat transport	-
r_a	turbulent diffusion resistance for heat and water vapour transport	$s.m^{-1}$
r_{plant}	plant resistance for water transport	s
r_s	stomatal diffusion resistance for water vapour transport	$s.m^{-1}$
r_{soil}	soil hydraulic resistance	s
R	reflectance per wavelength λ_i	-
R_1	longwave sky radiation flux	$W.m^{-2}$

Symbol	Description	Units
R_n	net radiation flux	$W.m^{-2}$
R_s	incoming shortwave radiation flux	$W.m^{-2}$
S	standard deviation of the single evapotranspiration calculation	$W.m^{-2}$
T_a	air temperature at height z_a	K
T_c	crop surface temperature at height z_{oh}	K
T_s	sky brightness temperature	K
T_*	apparent surface temperature	K
u	wind velocity at height z_a	$m.s^{-1}$
u_*	friction velocity	$m.s^{-1}$
x	non dimensional wind velocity gradient as a function of Λ	-
z_a	reference level in the atmosphere	m
z_o	crop roughness	m
z_{oh}	crop roughness length for sensible heat	m
z_{om}	crop roughness length for momentum	m
α_l	reflection coefficient for longwave radiation	-
α_s	reflection coefficient for shortwave radiation	-
γ	psychrometric constant	$Pa.K^{-1}$
ε	crop emission coefficient	-
ε_{8-14}	crop emission coefficient for the 8-14 m band	-
$\varepsilon_{*,8-14}$	emission coefficient of the calibration black bodies for the 8-14 m band	-
λ	heat conductivity of soil	$W.m^{-1}.K^{-1}$
λ_i	wavelength spectral reflection	m
Λ	Monin-Obukhov length	m
ρ	density of moist air	$kg.m^{-3}$
$\delta_{\Delta x_1,\Delta x_2}$	correlation between the measurement of x_1 and x_2	-
σ	Stefan-Bolzmann constant ($\sigma = 5.67 \times 10^{-8}$)	$W.m^{-2}.K^{-4}$
$\sigma_{\Delta x_1}$	standard deviation in the measurement of x_1	-
Ψ_a	air entry value	Pa
Ψ_{leaf}	leaf water pressure	Pa
Ψ_{soil}	soil water pressure	Pa

1. INTRODUCTION

In Western Europe potential evapotranspiration of crops generally exceeds effective rainfall in spring and early summer. Under these circumstances, when the for the crop available moisture in the root zone becomes exhausted and capillary rise from the groundwater table does not succeed in supplying sufficient water to achieve the potential evapotranspiration rate, evapotranspiration will be reduced. This results in a lower crop growth and crop production. In such conditions, simultaneously with the decrease in evapotranspiration, the temperature of the crop surface will increase. As distinct physical relations between the surface temperature and the evapotranspiration rate of a crop exist, it is possible to calculate actual evapotranspiration from the crop surface temperature (BARTHOLIC et al., 1972; BROWN, 1974; STONE and HORTON, 1974; HEILMAN et al., 1976). Subsequently the moisture condition of the soil can be estimated from the evapotranspiration rate.

Because of the high cost of measuring actual evapotranspiration over large areas, other evapotranspiration measurements were limited to small areas with spatially uniform hydrological conditions. The use of remotely sensed crop surface temperatures eliminates the need of intensive field measurements over the entire area and offers the possibility to obtain regional evapotranspiration estimates, making water balance analyses for large areas more feasible.

For specific regions, crop production as correlated with the actual crop transpiration rate can be estimated. The extremely dry summer of 1976 affecting large parts of Western Europe emphasized once more the importance of having such regional information.

2. THEORY

Crop surface temperature is determined by heat and moisture flow in the plant - soil - atmosphere continuum. Supply and removal of energy at the crop surface can be expressed in the energy balance equation:

$$R_n + G + H + L.E = 0 \tag{1}$$

where R_n is the net radiation flux ($W.m^{-2}$), G the heat flux into the soil ($W.m^{-2}$), H the sensible heat flux ($W.m^{-2}$) and L.E the latent heat flux ($W.m^{-2}$), the energy equivalent of the evapotranspiration flux E ($kg.m^{-2}.s^{-1}$); L is the latent heat of vaporization of water per unit mass ($J.kg^{-1}$). Fluxes towards the crop surface are given a positive value.

The net radiation flux is the resultant of incoming and outgoing radiation fluxes:

$$R_n = (1 - \alpha_s) R_s + (1 - \alpha_l) R_l - \epsilon \sigma T_c^4 \tag{2}$$

where R_s is the incoming shortwave radiation flux ($W.m^{-2}$), α_s the crop's reflection coefficient for shortwave radiation, R_l the longwave sky radiation flux ($W.m^{-2}$), α_l the crop's reflection coefficient for longwave radiation, ϵ the crop's emission coefficient, σ the Stefan-Boltzmann constant $[(5.67 \times 10^{-8})\ W.m^{-2}.\ K^{-4}]$ and T_c the crop surface temperature (K). For longwave radiation at the crop surface, the sum of ϵ and α_l equals 1, so eq. (2) can be written as:

$$R_n = (1-\alpha_s) R_s + \epsilon(R_l - \sigma T_c^4) \tag{3}$$

The heat flux into the soil is proportional to the temperature gradient and the heat conductivity in the soil:

$$G = -\lambda \frac{\delta T}{\delta z} \tag{4}$$

where z is the depth (m), T the soil temperature (K) and λ the thermal conductivity of the soil ($W.m^{-1}.K^{-1}$). The soil temperature gradient generally increases when the crop temperature increases, but heat conductivity decreases at increasing surface temperatures. The latter because an increase in surface temperature mostly results from a lower water content of the soil. This means that the heat flux into the soil is governed by a kind of buffer system.

The sensible heat flux H can be written as a transport equation, consisting of a driving force, the air - crop temperature gradient ($T_a - T_c$), and a turbulent diffusion resistance for heat transport r_a ($s.m^{-1}$):

$$H = \rho c_p \frac{(T_a - T_c)}{r_a} \tag{5}$$

where ρ is the density ($kg.m^{-3}$) and c_p the specific heat ($J.kg^{-1}.K^{-1}$) of moist air.

The turbulent diffusion resistance r_a is a function of wind velocity u ($m.s^{-1}$), the stability of the atmosphere just above the crop, and of the nature of the surface (crop height, crop structure). Under conditions of neutral stability ($T_c \approx T_a$), r_a can be expressed as a function of only wind velocity and roughness of the surface:

$$r_a = \frac{\ln\left(\frac{z_a - d}{z_{om}}\right) \cdot \left(\frac{z_a - d}{z_{oh}}\right)}{k^2 u} \tag{6}$$

where z_a is an elevation reference level in the atmosphere (m) where wind velocity and air temperature are recorded, d the zero displacement (m), k Von Karman's constant (here taken to be 0.4), z_{om} the roughness length for momentum (m) and z_{oh} the roughness length for sensible heat (m).

When evapotranspiration is reduced crop temperature will rise and unstable conditions will come into being. Due to temperature induced differences in air density, vertical mass as well as heat transport will increase. For such conditions BUSINGER (1966), BUSINGER et al. (1971) and DYER (1967) derived semi-emperical mass and heat transport formulas (hereafter referred to as the Businger-Dyer concept), based on the use of the Monin-Obukhov length Λ (m) as a measure for stability (MONIN and OBUKHOV, 1954):

$$\Lambda = \frac{u_*^3 \rho c_p T_a}{kgH} \tag{7}$$

where u_* is the friction velocity ($m.s^{-1}$) and g is the acceleration due to gravity ($9.813\ m.s^{-2}$). Under unstable conditions ($T_c > T_a$), r_a can be expressed as (c.f. PAULSON, 1971):

$$r_a = \frac{\left[\ln\left(\frac{z_a - d}{z_{om}}\right) - P_1\right] \cdot \left[\ln\left(\frac{z_a - d}{z_{oh}}\right) - P_2\right]}{k^2 u} \tag{8}$$

where P_1 and P_2 are functions of Λ according to:

$$P_1 = 2\ln\left(\frac{1+x}{2}\right) + \ln\left(\frac{1+x^2}{2}\right) - 2\arctan(x) + \frac{\pi}{2} \tag{9}$$

$$P_2 = 2\ln\left(\frac{1+x^2}{2}\right) \tag{10}$$

where:

$$x = \left(1 - 16\,\frac{z_a - d}{\Lambda}\right)^{0.25} \tag{11}$$

Eq. (8) does not hold for extremely unstable conditions when free convection predominates. Practically, for grassland, the Businger-Dyer concept holds for wind velocities of more than about $1\ m.s^{-1}$ at 2 m height.

For stable conditions ($T_c < T_a$), the formulas established by WEBB (1970) can be used. According to BUSINGER et al. (1971) a value of 4.7 was adopted for the constant in his formulas:

$$r_a = \frac{\left[\ln\left(\frac{z_a - d}{z_{om}}\right) + 4.7\,\frac{z_a - d}{\Lambda}\right] \cdot \left[\ln\left(\frac{z_a - d}{z_{oh}}\right) + 4.7\,\frac{z_a - d}{\Lambda}\right]}{k^2 u} \quad \text{for } \Lambda > z_a - d \tag{12}$$

$$r_a = \frac{\left[\ln\left(\frac{z_a - d}{z_{om}}\right) + 4.7\right] \cdot \left[\ln\left(\frac{z_a - d}{z_{oh}}\right) + 4.7\right]}{k^2 u} \quad \text{for } 0 < \Lambda \leq z_a - d \tag{13}$$

Fig. 1 shows for a crop height of 0.20 m the relation between r_a and wind velocity for air - crop temperature gradients varying from -10 to 4 K.

The calculations were performed, assuming:

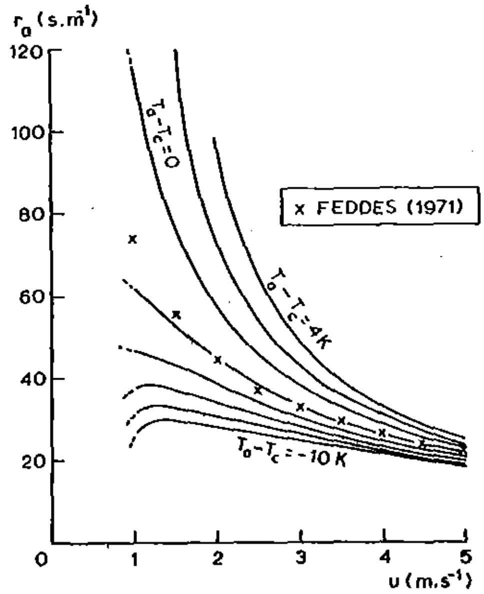

Fig. 1. Theoretically derived relation between the turbulent diffusion resistance r_a and wind velocity u for a crop height of 0.20 m, at different air - crop temperature gradients (T_a-T_c). The values are compared to data of FEDDES (1971)

$$z_{oh} = z_{om} = z_o \tag{14}$$

where z_o is the crop roughness (m) which can be calculated from the crop height h (m) using a simple relation established by MONTEITH (1973):

$$z_o = 0.13\ h \tag{15}$$

The calculated r_a values are compared to values empirically derived by FEDDES (1971). They agree well for high wind velocities. For low wind velocities the values of FEDDES seem to include some instability, which agrees with the climatological conditions during which those values were derived.

Using r_a from eq. (8), (12) and (13) in eq. (5) yields equations in which the sensible heat flux H depends in a rather complicated way on wind velocity, air - crop temperature gradient, and roughness parameters. For unstable conditions these equations can only be solved by iteration techniques (c.f. ROSEMA, 1975). Fig. 2 shows such a relation for three different crop heights and a wind velocity of 2,4 $m.s^{-1}$. A simple relation for a crop height of 0.20 m assuming no influence of stability on sensible heat flux and using eq. (6) to calculate r_a is also given in Fig. 2. It appears that neglecting the influence of stability may cause large errors when calculating H.

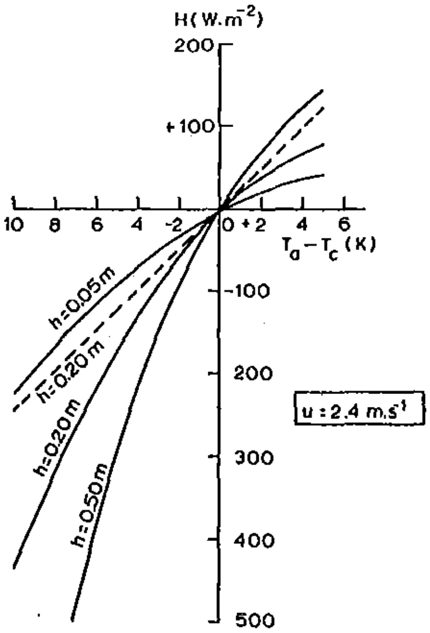

Fig. 2. Theoretically derived relation between sensible heat flux H and the air - crop temperature gradient (T_a-T_c) for three different crop heights h at a wind velocity u = 2.4 $m.s^{-1}$. The dotted line gives the relation (T_a - T_c) without stability correction for h = 0.20 m

Substituting eqs. (3-5) in the energy balance eq. (1) gives:

$$L.E = -\rho c_p \frac{(T_a - T_c)}{r_a} - (1 - \alpha_s) R_s - \varepsilon (R_l - \sigma T_c^4) - G \qquad (16)$$

When the crop surface temperature T_c is measured and the other variables at the right hand side of eq. (16) are also known, actual evapotranspiration can be calculated. For specified meteorological conditions the relation between actual evapotranspiration and crop surface temperature is given in Fig. 3A.

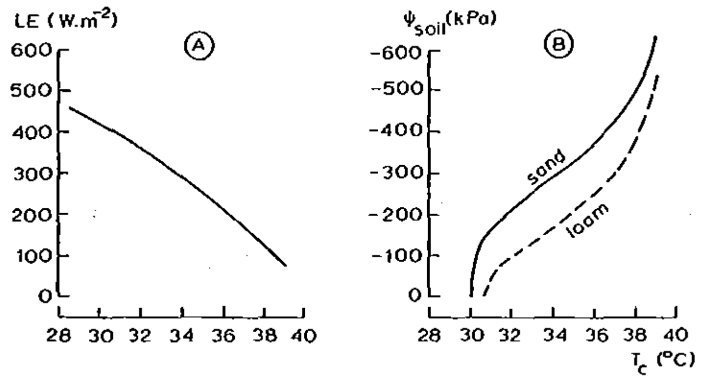

Fig. 3. A, Theoretically derived relation between actual evapotranspiration L.E, and crop surface temperature T_c. B, theoretically derived relation between soil moisture pressure Ψ_{soil} and crop surface temperature T_c for a sandy and a loamy soil. Meteorological conditions: R_s = 749 $W.m^{-2}$, R_l = 330 $W.m^{-2}$, u = 2.4 $m.s^{-1}$, T_a = 30.0 °C, e_a = 1670 Pa, h = 0.20 m, α_s = 0.21, ε = 0.95; $-40 > G > -50$ $W.m^{-2}$
$G = f(T_c)$

The latent heat flux L.E can also be expressed as a transport equation:

$$L.E = \frac{\rho c_p}{\gamma} \cdot \frac{e_a - e_c^*}{r_a + r_s} \tag{17}$$

where γ is the psychrometric constant ($Pa.K^{-1}$), e_a the water vapour pressure in the air (Pa), e_c^* the saturated water vapour pressure of air (Pa) at a temperature T_c, and r_s the stomatal diffusion resistance for water vapour transport ($s.m^{-1}$). For grassland with completely covered soil, the evapotranspiration flux may be supposed to equal the water flux through soil and plant, expressed as (FEDDES and RIJTEMA, 1972):

$$E = \frac{1}{g} \frac{\Psi_{leaf} - \Psi_{soil}}{r_{plant} + r_{soil}} \tag{18}$$

where ψ is the water pressure (Pa) and r the water flow resistance (s) in the appropriate medium. The resistance r_{soil} can be expressed as:

$$r_{soil} = b\left[K\,(\psi_{soil})\right]^{-1} \tag{19}$$

where b is theroot density resistence factor (m), and K (ψ_{soil}) the hydraulic conductivity ($m.s^{-1}$) as a function of ψ_{soil}. Combining Fig. 36 and 37 of RIJTEMA (1965) an empirical relation between ψ_{leaf} and r_s is found:

$$r_s = 4,52 \times 10^{-12}\,\psi_{leaf}^{2.1} \tag{20}$$

It is estimated from Rijtema that this relation is valid for grassland with a mean crop height of 0.14 m. Combining eqs. (16-19), yields a relation between soil water pressure ψ_{soil} and crop surface temperature. This relation is given in Fig. 3B for two soil types and for the specified meteorological conditions. As inaccuracy is inherent to eq. (20) and to the K - ψ_{soil} relation, the calculation of ψ_{soil} will be less accurate than the calculation of L.E.

3. FIELD EXPERIMENTS AND MEASUREMENTS

The study area covers about 200 ha of grassland west of Losser, 7^{o} longitude east and $52^{o}16'$ latitude north. It is part of an area where drought damage occurs, mainly due to groundwater withdrawal for drinking water supply. The deep percolation caused by groundwater withdrawal, manifests itself at ground surface as three almost parallel zones of drought damage (RIJTEMA and BON, 1974). Taking this study into account, three test sites were chosen representing a loamy soil not affected by drought damage, and a loamy soil and a sandy soil both affected by drought damage.

At each test site, several sets of meteorological measurements were performed during the summer of 1975 and 1976. One of these sets is used in this paper to demonstrate the method.

The incoming shortwave radiation was measured with a Kipp CM5 solarimeter. Net radiation of the grass canopy was measured using a S.R.I. 4 net-radiometer, positioned at 1 m height. Incoming longwave radiation was measured using the same type net-radiometer, the lower half covered with a blackbody radiator of known temperature. Air temperature and wet bulb temperature were measured using copper-constantan thermocouples at 2 m height. Wind velocity was measured at 2 m height with a Thiess 0.315 m Ø anemometer. Surface albedo was measured using a S.R.I. 5 solari-albedo-meter. Soil heat flux was measured at 4 and 20 cm depth, using perforated S.R.I. heat flux plates. Soil temperature was measured with copper-constantan thermocouples, respectively at 2, 4, 6, 10, 15, 20, 30 and 50 cm depth.

Crop surface temperature was measured with a Heimann K24 radiation thermometer, with a 37^{o} field of view. The instrument was situated 2.5 m above the canopy. It was field-calibrated, using a tube blackbody radiator of known temperature and emission coefficient.

The output of all these instruments was automatically recorded at 2 minute intervals by a TFDL cassette datalogger.

The physical properties of the soils of the test sites were examined. Soil moisture retention curves were made and saturated hydraulic conductivities were measured. Following LALIBERTE et al. (1968), the K - ψ_{soil} relations were derived. Soil density and soil organic matter in the root zone were measured. In the laboratory, the heat conductivity of undisturbed soil samples was measured for a complete range of soil moisture contents, using the transient needle method (JANSE and BOREL, 1965).

During the meteorological measurements, on each test site the crop height was measured weekly and the root density at 5 cm intervals in the root zone was measured twice. For water balance purposes, groundwater table depth was determined weekly. The soil water content at each test site was measured weekly at 10 cm interval down to 1.30 m with the aid of the γ-transmission method. Rainfall was recorded continuously by an automatic rainfall recorder with an opening of 0.04 m^2 placed at 0.60 m height.

In July the spectral characteristics of grassland were measured at 12 different sites in the study area with the aid of the NIWARS field spectrometer, providing spectral reflectances of about seventy separate wavelength bands in the 0.36 - 2.4 μm wavelength region.

In the summer of 1975 three IRLS recordings of the study area were effected. The first respectively last recording were executed on 29 May and 30 July, both after a relatively wet period. Because of this they are not considered here. The second recording was effected at the end of a relatively dry period at 1345 MET of 8 July 1975.

The crop surface temperature was measured by an airborne multiband scanner (Bendix M^2S), flown at 4000 ft. The instrument scans the earth surface by means of a rotating mirror in scanlines perpendicualr to the flight direction. In one scanline radiation from the earth surface is recorded 825 times. Each separately, integrally scanned, surface element is called a pixel. In the 8 July recording a pixel measures about 2.1 x 3.6 m. From each pixel the thermal radiation in the 8 - 14 μm band was measured. Apart from the measurement of thermal radiation, the reflected short-wave radiation was measured in ten separate wavelength bands. The information received was recorded on magnetic type and converted into eight figure digital numbers, yielding decimal numbers of 0 - 255, directly proportional to the intensity of the radiation received. The scanner measurements of thermal radiation were calibrated to a 'cold' and a 'hot' blackbody radiator within the scanner unit, which temperatures can be adapted to the variety of surface temperatures expected. For the 8 July recording, the blackbody temperatures were set to 5 and 40 ^oC respectively. On 8 and 9 July 1975 an intensive groundtruth of the study area was accomplished. Of each parcel crop height, crop type, soil cover and general state of the crop were recorded. On 8 July between 1300 and 1600 MET, the reflection coefficients α_s of 30 representative parcels were measured.

4. METHODS

4.1. Calculation of momentary evapotranspiration

From the thermal radiation data recorded on magnetic tape, an artificial photograph was produced (fig. 4). This photograph manifests

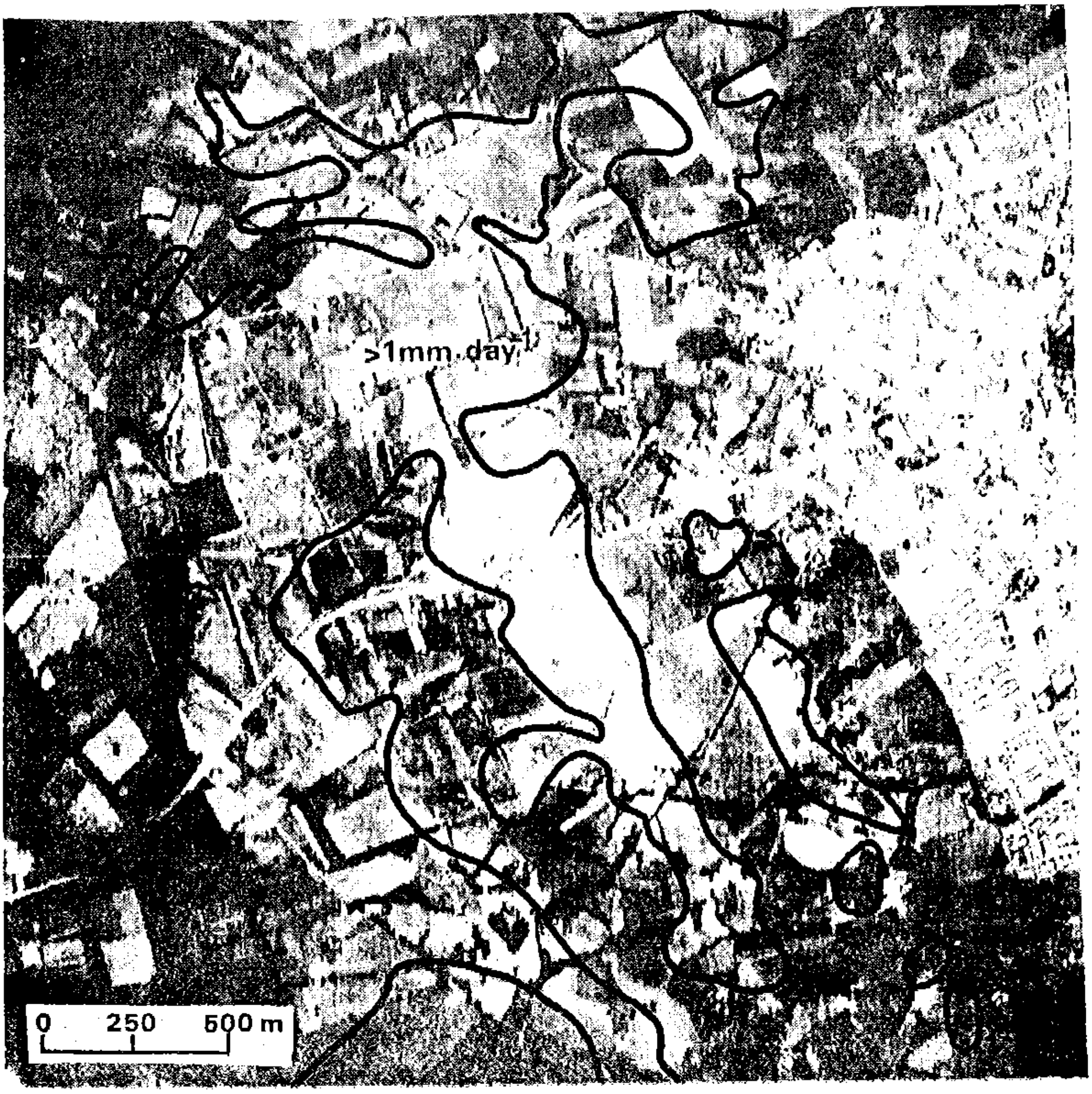

Fig. 4. IRLS image of the Losser area, 1345 MET of 8 July 1975. Light areas have high temperatures, dark areas low temperatures. Areas with a deep percolation in summer of more than 1 $mm.day^{-1}$ are encircled. The image clearly manifests three zones of high crop temperatures and reduced evapotranspiration. These zones correspond well with the encircled zones. (NIWARS IRLS image by Umwelt Data GmbH, Offenbach, W. Germany, through KLM Aerocarto BV, The Netherlands)

clearly some zones of raised crop surface temperature and reduced evapotranspiration. These zones correspond well with the zones of more than 1 mm.day^{-1} deep percolation in the summer (RIJTEMA and BON, 1974).

A quantitative analysis of the scanner measurements was made using computer thermal maps (fig. 5), where each figure represents a pixel over a specified temperature range. Similar maps were made for some reflection bands. From such types of maps, the momentary evapotranspiration at 1345 was computed with eq. 16 for plots with rather uniform surface temperature.

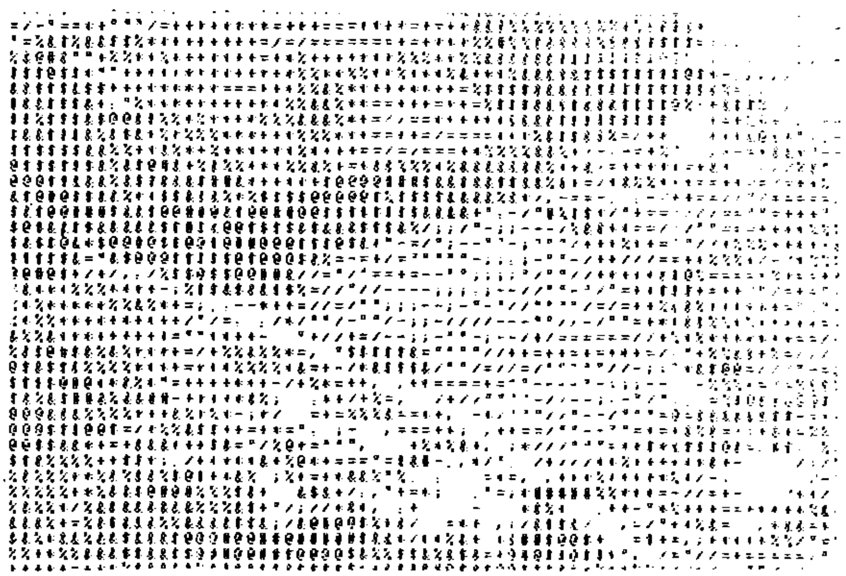

Fig. 5. Computer printout of IRLS image. Each figure represents a pixel over a specified temperature range. Temperatures are ranging from about 25 ^{o}C (#) to 40 ^{o}C (blank)

The variables R_s, R_1, u, T_a, ε and G, at the right hand side of this equation were taken constant for the study area: at the mentioned time R_s was 749 W.m^{-2}, R_1 330 W.m^{-2}, u 2.4 m.s^{-1}, T_a 30.0 ^{o}C, ε was arbitrarily chosen 0.95, and G was established to be -20 W.m^{-2}. The vapour pressure e_a one of the variables needed for the soil moisture pressure calculation (eq. 17-18), was 1670 P_a.

The airborne calibration of the received thermal radiation to the 'cold' and 'hot' blackbody temperatures, furnished the apparent surface temperatures T_*. Neglecting atmospheric influences, the total radiation received by the scanner is approximately proportional to:

$$\varepsilon_{*,8-14}\ \sigma\ T_*^{\ 4} = (\ 1 - \varepsilon_{8-14})\ \sigma\ T_s^{\ 4} + \varepsilon_{8-14}\ \sigma\ T_c^{\ 4} \tag{21}$$

where $\varepsilon_{*,8-14}$ is the emission coefficient of the calibration blackbodies, ε_{8-14} is the emission coefficient of the surface and T_s(K) is the sky brightness temperature, all for the 8 - 14 μm band. For small differences of $(T_c - T_*)$, $(T_c^{\ 4} - T_*^{\ 4})$ equals approximately $4T_c^{\ 3}(T_c - T_*)$. Rearranging eq. 21, we find an expression for the difference between actual and apparent surface temperature:

$$T_c - T_* = \frac{1}{4}\ \left(1 - \frac{\varepsilon_{8-14}}{\varepsilon_{*,8-14}}\right)T_c - \left(\frac{1 - \varepsilon_{8-14}}{\varepsilon_{*,8-14}}\right)\frac{T_s^{\ 4}}{T_c^{\ 3}} \tag{22}$$

For the prevailing conditions, it was chosen arbitrarily that T_s equals 270 k, $\varepsilon_{*,8-14}$ equals 0.99 and ε_{8-14} equals 0.95. For a surface temperature of 303 K, the correction becomes 0.7 K, the value applied in this study.

Radiation reaching the scanner from the earth surface will be influenced by atmospheric attenuation as the scanner also will receive radiation from the atmosphere itself. The combination of those effects may cause errors in the correct measurement of the surface temperature, although it may be possible that they cancel each other. Other investigators (e.g. HEILMAN et al., 1976; SHAW and IRBE, 1972) mentioned rather important atmospheric corrections needed for the thermal band. In this study, scanner-measured crop surface temperatures were compared with crop surface temperatures measured with the radiation thermometer in the field. There was no significant difference.

Also, scanner measurements right below the airplane and measurements at the outmost end of the scanline differ in atmospheric path length. So different atmospheric corrections would be needed. When important, this would cause a systematic difference in the IRLS image. This systematic difference was not found. It appears from the author's experience that atmospheric corrections are not very important. Nevertheless, it is recommended to compare scanner measured surface temperature data with field measurements preferably done with a calibrated radiation thermometer.

The weighted sum of the shortwave radiation measured in all ten reflection bands of the scanner would be a rather exact relative measure for the total reflected shortwave radiation. As the reflection bands of the scanner are not calibrated to absolute radiation values the weights are unknown and it is not possible to compute the crop's reflection coefficient in this way. Analyzing grassland spectral characteristics measured by BUNNIK (pers. comm.), it was found that the reflection coefficient for near-infrared radiation (0.7 - 1.4 µm) is several times higher than that for visible radiation (0.3 - 0.7 µm) (fig. 6). As the incoming shortwave radiation is about equally distributed over near-infrared and visible radiation, the crop reflection coefficient α_s is mainly influenced by the reflection of near-infrared radiation. Comparing field measurements of α_s with band 9 of the scanner (0.81 ± 0.05 µm), being almost proportional to the total near-infrared reflection, gives the relation shown in fig. 7.

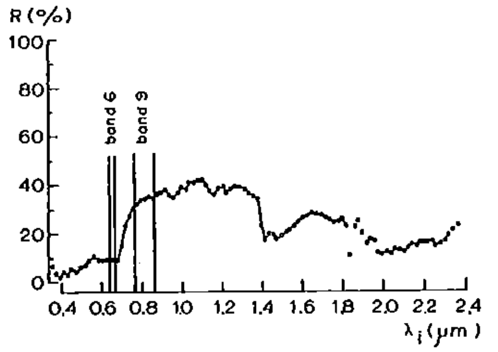

Fig. 6. Spectral reflectance of grassland: reflectance R for wavelength λ_i. Indicated is also the place of band 9 and band 6 of the Bendix M^2S scanner

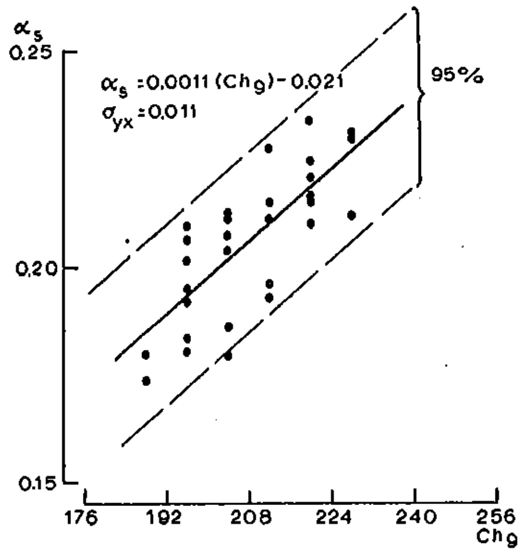

Fig. 7. Relation between the reflection coefficient α_s measured in the field and the value of Bendix M^2S band 9 (Ch 9: 0.81 ± 0.05 µm)

For the study area, the crop height of each parcel was measured and this value was used for the calculation of crop roughness z_o, using eq. 15. For more extensive areas, the measurement of crop height might be difficult. For this reason it was tried to find a correlation between crop height and one or more reflection band values. BUNNIK and VERHOEF (1975) suggested a theoretically based relationship between these parameters. VAN KASTEREN and UENK (1975) found for various grass plots of specified variety and cultural treatment rather high correlations, particularly between crop height and reflection at 0.87 μm, with correlation coefficients varying for separate plots from 0.91 to 0.98. The overall correlation coefficient was 0.84. In the study area, conditions of varieties and cultural treatment are rather variable. Comparing field measurements of crop height to the value of band 9 (0.81 ± 0.05 μm), band 6 (0.64 ± 0.02 μm) and with the ratio of band 9 and band 6, gives a maximum correlation coefficient of 0.67 with a standard deviation for the single observation of grass height of 0.06 m. Accurate grass height estimation from scanner data seems to be impossible. Only some classes in grass height can be distinguished. Fig. 8 shows as an example the correlation between grass height and the ratio of band 9 and band 6.

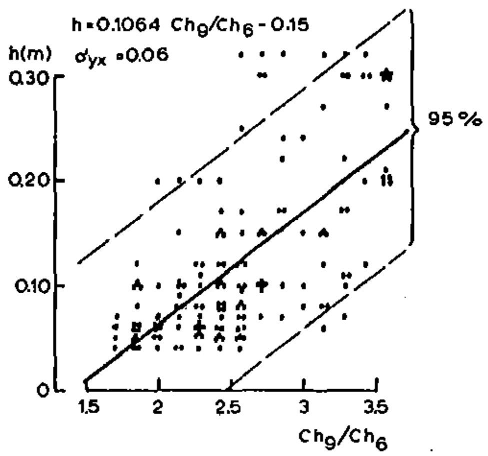

Fig. 8. Relation between crop height h measured in the field and the ratio of the values of band 9 (Ch 9: 0.81 ± 0.05 m) and band 6 (Ch 6: 0.64 ± 0.02 m) of the Bendix M^2S scanner

4.2. Conversion to daily evapotranspiration

To convert the momentary evapotranspiration data into 24-hour estimates of evapotranspiration, being more valuable for hydrological studies, the aid of a simulation model is indispensable. Such a model, the TERGRA model, was developed by the author (SOER, 1977).

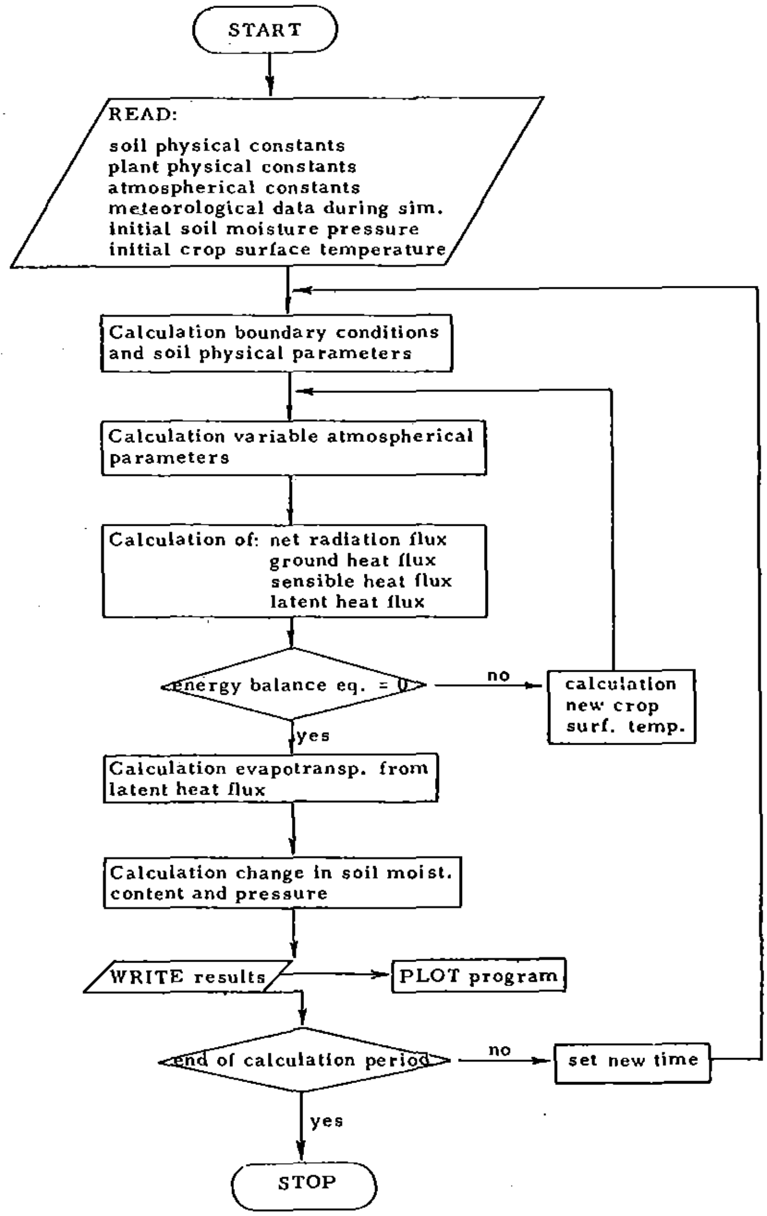

Fig. 9. Flow diagram of the TERGRA model

The model simulates for grassland, under defined meteorological conditions and for different soil moisture regimes, the daily course of crop surface temperature, actual evapotranspiration, net radiation, ground heat flux and dew formation. It is based on the transport equations for heat and moisture flow in soil, plant and atmosphere (eqs. 4-15, 17 and 18). Boundary conditions are the temperature and soil moisture pressure at a reference level in the soil, the energy balance equation at the crop surface (eq. 1) and the temperature and vapour pressure at a reference level in the atmosphere. Also, some relationships between model parameters (e.g. eq. 19) are introduced in the model. A numerical algorithm to solve the transport equations completes the model. Fig. 9 shows a flow diagram of this TERGRA-model.

Fig. 10 shows for 8 July 1975 a model output of the crop surface temperature T_c, and of the input parameters T_a and u. The simulation was performed for a sandy soil at -350 kPa soil moisture pressure. The simulated crop surface temperatures were compared with values measured with the radiation thermometer. The measured crop surface temperatures are systematically higher than the simulated values. This is probably due to a calibration error of the Heimann radiation thermometer. Considering this, measured and simulated values agree well.

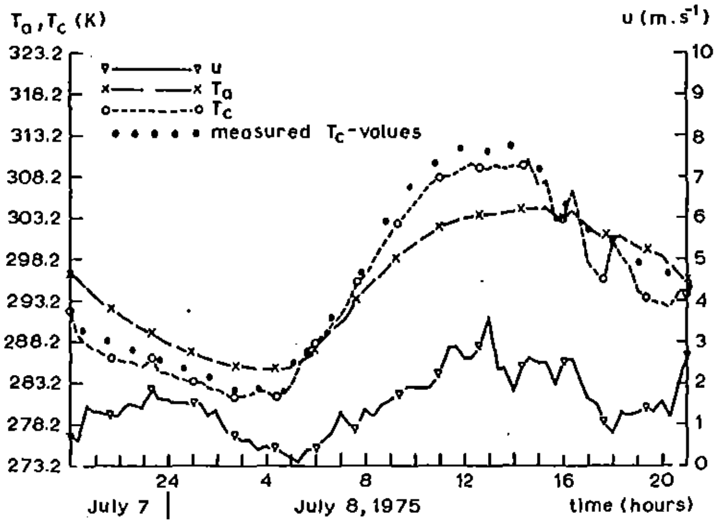

Fig. 10. TERGRA simulation of the crop surface temperature T_c for the input parameters air temperature T_a and wind velocity u. The simulation was performed for a sandy soil with a soil moisture pressure of -350 kPa, for the meteorological conditions of 8 July 1975

Fig. 11 shows for the same day a model output for the energy balance components R_n, G, H and L.E. Fig. 12 shows the model output for the daily course of LE at different soil moisture regimes, also for the meteorological conditions of 8 July 1975. Comparing the momentary evapotranspiration with these curves, a relation between momentary and daily evapotranspiration was established, where the daily evapotranspiration is the integral of the curve which fits the momentary evapotranspiration (fig. 13).

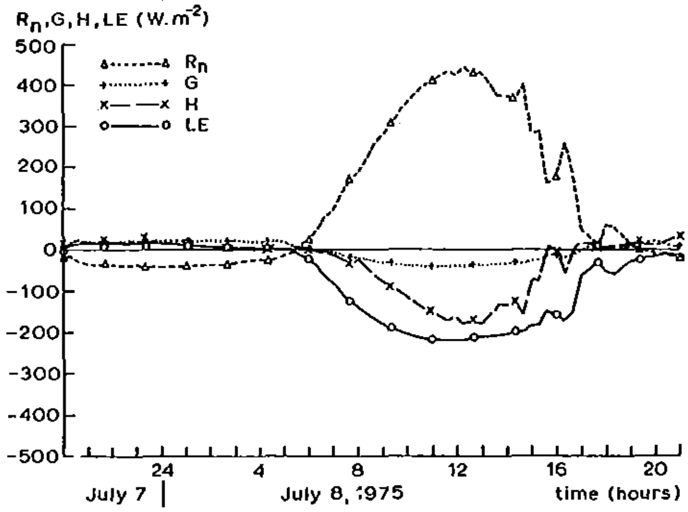

Fig. 11. TERGRA simulation of the energy balance components net radiation flux R_n, heat flux into the soil G, sensible heat flux H and latent heat flux L.E for 8 July 1975

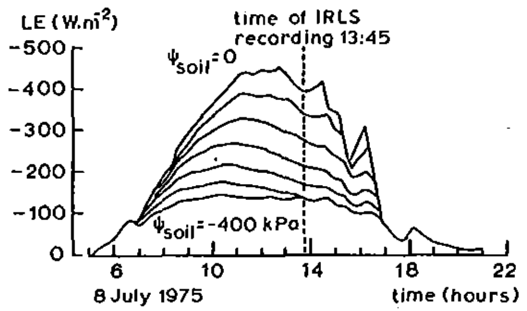

Fig. 12. TERGRA simulation of the daily course of actual evapotranspiration L.E at different soil moisture pressure values Ψ_{soil} for 8 July 1975

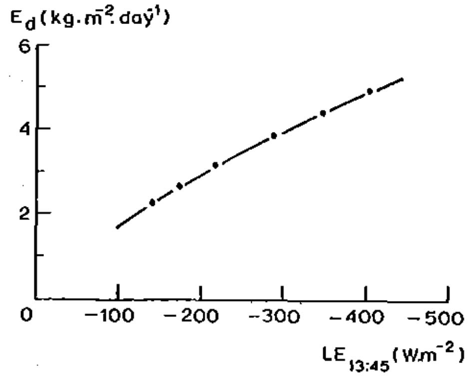

Fig. 13. Relation between the daily evapotranspiration E_d on 8 July 1975 and the momentary evapotranspiration L.E at 1345

4.3. Calculation of soil moisture pressure

Combining equations 16, 17, 18, 19 and 20 the soil moisture pressure can be calculated by an iterative procedure. The soil of the study area was split up in three main types differing in soil physical and rooting caracteristics. For each soil type, r_{plant}, b and $K(\Psi_{soil})$ must be known (eqs. 18 and 19). Comparing rooting depth and intensity with data of FEDDES and RIJTEMA (1972), the r_{plant} and b variables were estimated.

Following LA LIBERTE et al. (1968), the K - Ψ_{soil} relation can be expressed as:

$$K = K_s \cdot \Psi_a^{\ m} \cdot \Psi_{soil}^{-m}$$

where K_s is the saturated hydraulic conductivity (m.s^{-1}), Ψ_a the air entry value (P_a) and m is the pore size distribution factor. The parameters Ψ_a and m were derived from soil moisture retention curves. Table 1 gives r_{plant}, b, K_s, Ψ_a and m for the three main soil types of the area.

Table 1. Soil physical parameters of the three main soil types of the study area

Soil type	b (mm)	r_{plant} (day)	K_s (m.day^{-1})	Ψ_a (kPa)	m
Fine sand	3.0	10 000	2.0	2.5	3.38
Clay loam	3.7	12 300	0.010	2.0	2.39
River deposits	2.4	8 000	0.20	3.0	3.08

5. RESULTS AND DISCUSSION

The calculated daily evapotranspiration rates for the study area were mapped (fig. 14). For grassland, the evapotranspiration rate varies from 0.3 to 5.8 $kg.m^{-2}.day^{-1}$. Net radiation values, calculated simultaneously with the calculation of momentary evapotranspiration, vary from 376 to 485 $W.m^{-2}$. Introducing one constant value of net radiation for the whole area may introduce large errors. Using eq. 17, r_s can be calculated. The r_s-values vary from 0 to $\sim$800 $s.m^{-1}$. The groundtruth indicates that zero values for r_s coincide with wet soil surfaces.

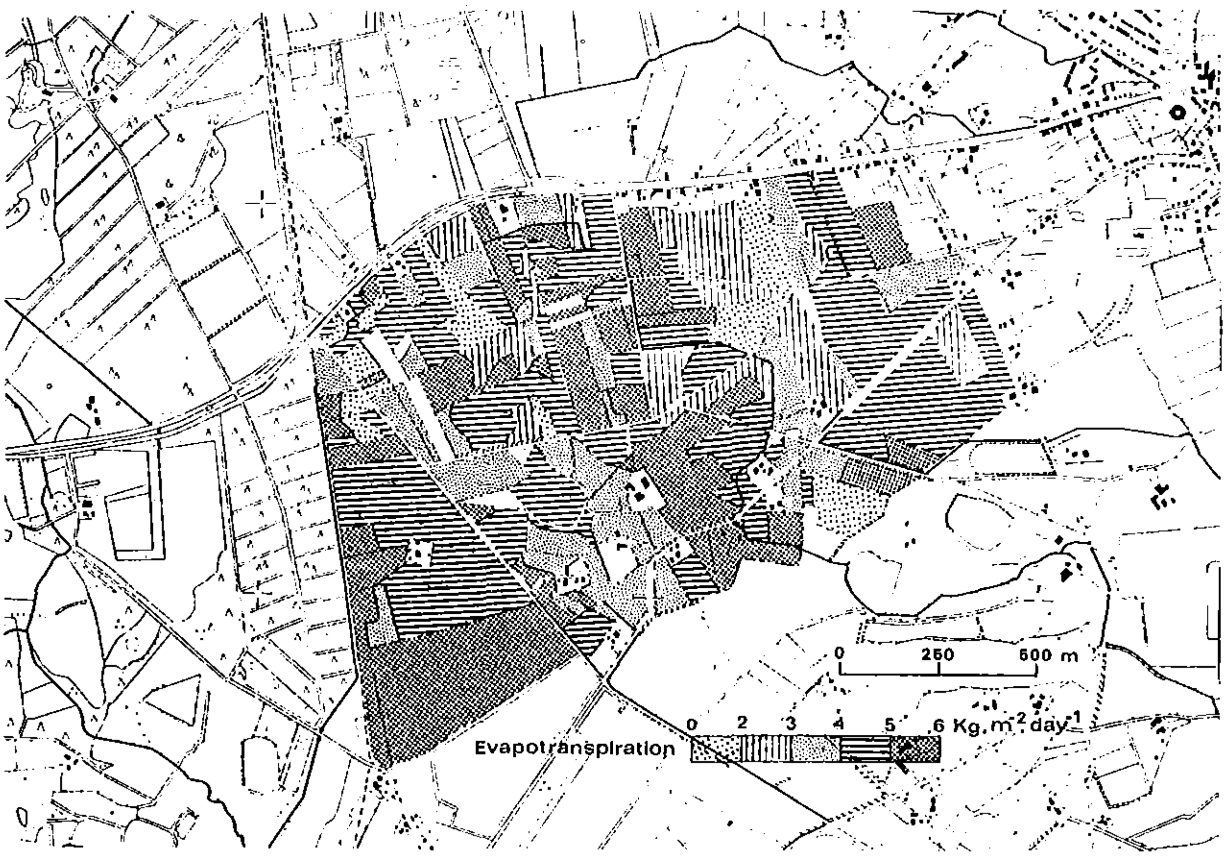

Fig. 14. Daily evapotranspiration of the study area on 8 July 1975

Fig. 15 shows a soil moisture pressure map of the area. Some irregularities probably are caused by inaccuracies in soil moisture characteristics. Difficulties were encountered in computing evapotranspiration of tall grass (over 0.50 m), where the Businger - Dyer concept or the calculation of the roughness length for sensible heat (eqs. 14-15) seem to be less accurate, and in computing soil water pressures of recently cut parcels.

For three parcels, the calculated daily evapotranspiration rates were compared with evapotranspiration rates calculated from water balance estimates, established with the aid of γ-transmission measurements. The differences are within 30%.

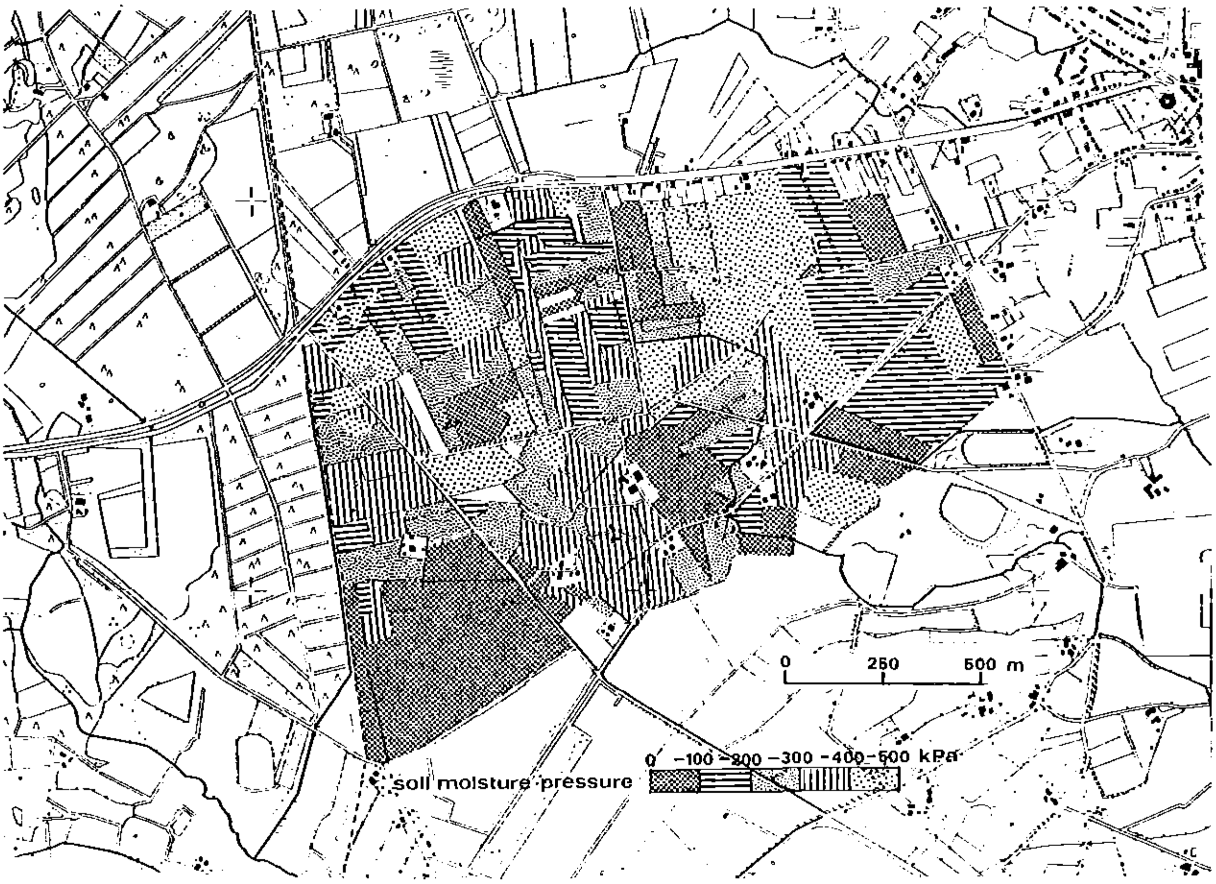

Fig. 15. Soil moisture pressure in the rootzone of the study area on 8 July 1975

An analysis of the accuracy of the calculation of evapotranspiration was performed. It is assumed that the equations used in the algorithm are correct and that the only error introduced in the calculation originates from measurement errors in the values of the necessary variables. This might not be true for all conditions and the actual accuracy may be somewhat lower than the one calculated here. Assuming

$$y = f\,(x_1,\ x_2,\ \ldots,\ x_n,\ a_1,\ \ldots,\ a_n) \tag{24}$$

the standard deviation S of y can be expressed as:

$$S^2 = \left(\frac{\delta f}{\delta x_1}\right)^2 \sigma^2_{\Delta x_1} + \ldots.\ \left(\frac{\delta f}{\delta x_n}\right)^2 \sigma^2_{\Delta x_n} + 2\left(\frac{\delta f}{\delta x_1}\right)\left(\frac{\delta f}{\delta x_2}\right) \varsigma_{\Delta x_1 \Delta x_2}\, \sigma_{\Delta x_1}\, \sigma_{\Delta x_2} + \ldots \text{ etc.} \tag{25}$$

where $\sigma_{\Delta x_1}$ is the standard deviation in the measurement of parameter x_1, etc. and $\varsigma_{\Delta x_1, \Delta x_2}$ is the correlation between the measurement of the parameters x_1 and x_2, etc. Using eq. 16, S for the single evapotranspiration calculation can be computed from eq. 25.

Table 2 gives the result of such a calculation for a crop height of 0.20 m, and for different crop surface temperatures. The standard deviations introduced for R_s, α_s and h were derived from measurements, the standard deviation for R_l was taken proportionally equal to that for R_s, the standard deviation for ε, T_a, u and G were estimated, partly based on meteorological measurements effected in the study area.

At L.E is 430 $W.m^{-2}$, S is 33 $W.m^{-2}$. When L.E is about zero, S is 71 $W.m^{-2}$. Using crop heights determined from remotely sensed reflection values (fig. 8), S will be respectively 41 and 85 $W.m^{-2}$. For the daily evapotranspiration estimate, the proportional inaccuracy will be somewhat higher due to inaccuracies in the TERGRA model. The standard deviation of R_n, mainly caused by the standard deviation of R_s, cannot be reduced significantly. To improve the accuracy of the algorithm, main attention must be given to the determination of the crop surface temperature T_c and the roughness length for sensible heat z_{oh}. Recent work (THOM, 1972; HEILMAN and KANEMASU, 1976)

Table 2. Standard deviations of evapotranspiration calculated with eq. 16 from IRLS images for a crop height of 0.20 m at various T_c-values

Values and standard deviation of input variables			
variable	value	standard deviation	
R_s	749	22	$(W.m^{-2})$
R_l	330	10	$(W.m^{-2})$
α_s	0.21	0.012	
ε	0.95	0.01	
T_a	303.2	0.2	(Kelvin)
u	2.4	0.2	$(m.s^{-1})$
h	.20	0.02	(m)
G	20	7	$(W.m^{-2})$

Values of:				Standard deviation $(W.m^{-2})$ of:			Standard deviation $(W.m^{-2})$ in H caused by:		
T_c (K)	R_n $(W.m^{-2})$	H $(W.m^{-2})$	L.E $(W.m^{-2})$	L.E	H	R_n	T_c	z_o	u
303.2	450	0	430	33	24	22	24	0	0
304.2	444	28	396	37	29	22	29	3	2
305.2	438	60	358	41	34	22	33	7	4
306.2	432	96	315	45	38	22	36	11	5
307.2	425	136	270	49	43	22	39	16	7
308.2	419	178	221	53	47	22	41	21	8
309.2	413	222	171	57	52	22	44	26	10
310.2	406	269	117	61	57	22	46	32	11
311.2	400	318	62	66	62	22	47	37	13
312.2	393	369	5	71	67	22	49	43	15

indicates that z_{oh} must be an order of magnitude lower than the roughness length for momentum z_{om} and that it depends not only on the nature of the surface, but also on the nature of the surrounding air. In the Businger - Dyer concept, z_{oh} is related to the height of T_c in the whole temperature profile, and it might well be possible that this height does not correspond with z_{oh}, as indicated by for example THOM (1972). The determination of z_{oh} rather should be investigated related to this concept than only as a function of the nature of the surface and of the surrounding air.

As no other data were available, ε was estimated at 0.95 and taken constant for the area. In reality, ε may vary slightly from place to place. Intensive studies of emission coefficients will be indispensable to achieve a high accuracy. It may be worthwhile trying to find some correlation between ε and crop physical parameters.

With IRLS, the evapotranspiration is measured in an almost infinite number of pixels within the area. Then, theoretically, the standard error of the measurement of the mean aerial evapotranspiration approaches zero. This implies that the actual mean aerial evapotranspiration can be calculated with about the same accuracy as the potential evapotranspiration with the Penman formula, having the same physical base.

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