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ESTIMATION OF SENSIBLE HEAT FLUX FROM MEASUREMENTS OF SURFACE RADIATIVE TEMPERATURE AND AIR TEMPERATURE AT TWO METERS: APPLICATION TO DETERMINE ACTUAL EVAPORATION RATE

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ABSTRACT

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From measurements of sensible heat flux (H), using the simplified aerodynamic approach, and measurements of surface thermal infrared temperature (T_s), both in arid and semi-arid areas of Tunisia, it is shown that a simple linear relationship can be used to estimate H from the difference $(T_s - T_a)$, T_a being the air temperature at 2 m. It is also pointed out that this relationship can be used in unstable conditions for various types of surface roughness. Application to determine actual evaporation rate is made using the similarity of the H/Rn instantaneous values at noon with the average daily values of H/Rn.

INTRODUCTION

Actual evaporation, E (W m^{-2}), is related to the other components of the energy budget over any surface by

$$E = Rn - H - G$$

Rn, the net radiation, can be measured directly using a net radiometer or can be estimated with quite good accuracy. G, the heat flux into the soil, can be measured directly or neglected in many cases. H, the sensible heat flux, can be measured, but the usual methods employ sophisticated and expensive equipment. However, simple methods for measuring H have been proposed. Itier (1980), Riou and Itier (1982) have proposed a method for measuring H based on a simplified form of the aerodynamic method, using only wind speed and temperature gradients between two levels above the surface. Alternatively,

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Jackson et al. (1977), Seguin and Itier (1983), Hatfield et al. (1983) and Seguin (1984) have used the differences between surface (T_s) and air (T_a) temperature to estimate evaporation. (T_s is the radiative temperature of the surface calculated from satellite thermal infrared (IR) data, T_a is the air temperature at 2 m).

The first method seems to be quite accurate, but it cannot be used for surfaces larger than a few hectares. The second method can be used to estimate evaporation from large areas (hundreds of km^2), but these surfaces must be large and homogeneous. Moreover, the method requires meteorological data from the atmosphere and it is only usable for clear-sky periods. At least, further extensive treatment of satellite data is required to estimate surface temperature.

We have obtained simple and accurate estimates of H by a method based on both the methods described above. From simultaneous measurement of various components of the energy budget and especially the sensible heat flux, H, with the simplified aerodynamic approach, and the measurement of T_a and T_s, we tried to find a relationship between H and $(T_s - T_a)$. Further, we examined the validity of this relationship for different surface roughnesses. T_s has been measured with an IR portable thermometer.

THEORETICAL BACKGROUND

The simplified aerodynamic method

The classical approach

The well-established conventional flux-profile relationship (Busingu et al., 1971) shows that in the surface sub-layer and in the absence of buoyancy effects, the wind profile of the mean wind speed over a uniform and smooth surface is a logarithmic function of z, i.e.

$$\delta u / \delta z = u^* / kz \tag{1}$$

where u is the wind speed, z is the distance from the ground, k is the Von Karman constant and u^* is the friction velocity defined as $\tau = \rho u^{*2}$, τ being the momentum flux.

This logarithmic law is extended to a thermally stratified boundary layer by the Monin–Obukhov similarity theory (Monin and Yaglom, 1971), according to which one can write

$$\delta u / \delta z = u^* / kz \, \Phi M(z/L) \tag{2}$$

Similar relationships can be applied to the flux and gradient of heat

$$\delta \Theta / \delta z = -(\Theta^* / kz) \Phi H(z/L) \tag{3}$$

where Θ is the potential temperature, Θ^* is a scaling parameter for temperature, similar to u^*, and is defined as

$$H = -\rho C_p u^* \Theta^* \quad (4)$$

where H is the sensible heat flux, $\Phi M(z/L)$ and $\Phi H(z/L)$ are the universal functions of the dimensionless stability variable (z/L), where L is the Monin–Obukhov length (Monin and Obukhov, 1954), defined by

$$L = -\rho C p u^3 T / kgH \quad (5)$$

The analytical expressions of ΦM and ΦH, the more frequently used, are the following

$$\Phi M^2 = \Phi H = (1 - 16\, z/L)^{-1/2} \text{ (Dyer and Hicks, 1974)} \quad (6)$$

for unstable conditions, $(z/L < 0)$, and

$$\Phi M = \Phi H = 1 + 5(z/L) \text{ (Webb, 1970)} \quad (7)$$

for stable conditions $(z/L > 0)$.

These relationships are valid only if the measurement height, z, is much larger than the roughness length z_0.

Integration of eq. 2 leads to the classical logarithmic law in stable conditions

$$u(z)/u^* = (\Theta z - \Theta_0)/\Theta^* = (1/k)]\ln(z/z_0) + 5(z/L)] \quad (8)$$

For unstable conditions, Paulson (1970) provided integral functions ψM and ψH of ΦM and ΦH, leading to the following equations

$$u(z) = (u^*/k)[\ln(z/z_0) - \psi M(z/L)]$$

$$\Theta(z) - \Theta_0 = (\Theta^*/k)[\ln(z/z_0) - \psi H(z/L)]$$

with $\psi M = 2\ln[(1+x)/2] + \ln[(1-x^2)/2] - 2\arctan x + \pi/2$, where $x = \Phi M^{-1}$ and $\psi H = 2\ln[(1+x^2)/2]$

u^*, Θ^* and then H from eq. 4 can only be determined using an iterative method from data measured at six levels at least above the surface.

The simplified approach

The Richardson number is an alternative stability parameter, which may be written as

$$Ri = (g/T)(\delta\Theta/\delta z)/(\delta u/\delta z)^2 \quad (9)$$

From eqs. 2–5, we can write

$$Ri = (z/L)\Phi H \Phi M^{-2} \quad (10)$$

and assuming that in unstable conditions (which are our main interest)

$\Phi H = \Phi M^{-2}$, then $Ri = z/L$ (Businger, 1988)

From the set of equations mentioned above, we can also write that

$$H = -\left[\rho C p k^2 z^2 (\delta\Theta/\delta z)(\delta u/\mathrm{d}z)/\Phi M \Phi H\right] \tag{11}$$

or $H = \rho C p k^2 z^2 |\delta\Theta/\delta z|^{3/2} (g/T)^{1/2} |z/L|^{-1/2} \Phi H^{-3/2}$ (12)

Using the dimensionless coefficient, h^*, derived by Priestly (1959) as

$$h^* = k^2 |z/L|^{-1/2} \Phi H^{-3/2} \tag{13}$$

and assuming that in unstable conditions this coefficient may be considered as constant over a small interval (z_1,z_2) and equal to its value for $z = z^* = (z_1 z_2)^{1/2}$, Riou and Itier (1982) proposed the following equation for H

$$H = Ho(1 - 16Ri)^{3/4} \tag{14}$$

where H_0 is the sensible heat flux computed for neutral conditions

$$H_0 = \rho C p k^2 (\Delta T \cdot \Delta u)/[\ln(z_2/z_1)]^2 \tag{15}$$

with $Ri = z^*(\Delta T/\Delta u^2)(g/T)\ln(z_2/z_1)$ (16)

For very high instability ($Ri < -1$), $h^* \cong 1.3$ and

$$H = \alpha |\Delta T|^{3/2} \tag{17}$$

with $\alpha = 1.3 \rho C p (g/T)^{1/2} / [3(z_1^{-1/3} - z_2^{-1/3})]^{3/2}$

The method now only requires measurements of wind speed and temperature at two levels above the ground. It has been compared both with the classical aerodynamic method using five or six levels, the Bowen ratio method and with direct measurements from lysimeters, over Lucerne, by Itier (1980) near Versailles (France), and by the author over an irrigated lawn in the North of Tunisia.

Methods using thermal IR emission of natural surfaces

These methods began to be used at the end of the 1960's and the beginning of the 1970's (Lorenz, 1968; Bartholic et al., 1972; Wiegant and Bartholic, 1970; Brown and Rosenberg, 1973; Stone and Horton, 1974). They have been developed more recently with a view to using satellite thermal IR data to estimate large-scale evaporation. Basically, these methods still use the equation of energy balance

$$E = Rn - H - G$$

in which

$$H = \rho C p (T_s - T_a)/ra \tag{18}$$

Equation 18 is the integral of eq. 11.

Generally, most of the experiments rely on the comparison between E, being calculated from $(T_s - T_a)$, and E being measured or calculated by other methods (lysimeters, aerodynamic method, etc.), Rn being either measured or estimated, and G usually neglected. The main difficulty in using eq. 18 lies in the evaluation of ra in unstable conditions (the more frequent during the day), because it depends on wind speed and surface roughness (Hatfield et al., 1983; Seguin, 1984). In many cases, this has been avoided by establishing a statistical relationship of the form $E-Rn=f(T_s-T_a)$ from experimental data

$E-Rn=0.64\ (T_s-T_a)$ Jackson et al. (1977)

$E-Rn-G=1.07-0.25(T_s-T_a)$ Seguin and Itier (1983)

$E-Rn=0.97-0.25(T_s-T_a)$ Recan (1982)

E and Rn are expressed in millimeters of water per day.

These equations are set up from one time of day measurement of satellite IR thermal data, from which T_s is calculated. The instantaneous values of E are then converted into daily values.

FIELD EXPERIMENTS IN BOTH ARID AND SEMI-ARID AREAS OF TUNISIA

Objectives

Field experiments have been conducted in Tunisia, both in the semi-arid zone of the north and the arid zone of the south. The surfaces were a Chott's surface (Tozeur–Nefta region, a dry salt lake in a desert region), a steppe surface (Gafsa–Gabes region) and a growing wheat crop, 20–80 cm high (Tunis region).

The main objective was to use the differences (T_s-T_a) to develop a statistical relationship which could be used to estimate actual evaporation more simply than by any other methods and using simple equipment. As noted earlier, H is always the most difficult parameter to measure because it needs both sophisticated equipment and a highly skilled technician. Rn is easier to measure, at worst it can be estimated from climatological data and some authors (Seguin and Itier, 1983; Jackson et al., 1983) have shown that Rn may be obtained from one noon-time measurement.

Another objective was to investigate the validity of such a relationship for various surfaces and, if possible, various climatic conditions.

Equipment and method

Sensors of wind speed and temperature measurements

The equipment was composed of a mast on which were fixed the cup anemometers and the dry-bulb thermometers, at two different levels above the surface.

The differences in dry-bulb temperature between the two levels (ΔT), were obtained directly from a differential assembly of two sets of four small thermocouples in series. Each set was housed in a small fully ventilated shield, the device was calibrated before each run. Air temperature was measured separately using a single fine thermocouple with a reference cold junction, put in a standard screen at 2 m above the ground. Its purpose was to allow comparisons with routine data from a meteorological bureau.

A hand-held IR thermometer (type MIKRON 25, differential temperature reading) was used to measure surface temperatures. The spectral response was 7–20μm and the field of view 40°. Each measurement was made every hour or half an hour, according to the weather conditions.

The thermometer was held by hand facing downward with an angle of ~45°, at a height of 1.5 m from the ground. Four measurements were made at each time, the instrument facing successively north, west, south and east. The value taken into account was the average of the four values. As the thermometer was a differential device, all the temperatures were relative to the temperature of a black body, this providing a permanent calibration. The emissivity of the surfaces was assumed to be 1 (Wendler and Eaton, 1983). The accuracy was ±0.2°C.

Sensors of net radiation and soil heat flux

The improved Funk-type net radiometer (Swissteco) was used for measuring net radiation. The polythene dome of the net radiometer was kept inflated and flushed by dry air using a small pump. The net radiometer was fixed on the mast at a height of 1.5 m above the ground.

The heat flux into the ground was measured directly by one heat flux plate manufactured by Middleton and buried 1 cm below the soil surface.

Data logging and calculation of H

Data from the mast were sent every 30 s to a microcomputer through a 12-channel datalogger. Calculations of H were made according to eqs. 15 and 14 in which the approximative value of Ri is given by eq. 16. Average data of Rn H, G and E were available every 3 and 15 min. "Instantaneous" values of H were calculated from average values 3 min before and 3 min after the time of surface temperature measurement.

RESULTS

Region of Tozeur and Nefta

Three field experiments have been carried out in this region of South Tunisia, on two different sites. The first of these was located in the middle of

Chott El Jerid, 30 km east of Tozeur. The soil surface was composed of a sandy-clay crust covered with salt. There was a water table 5 cm below the surface.

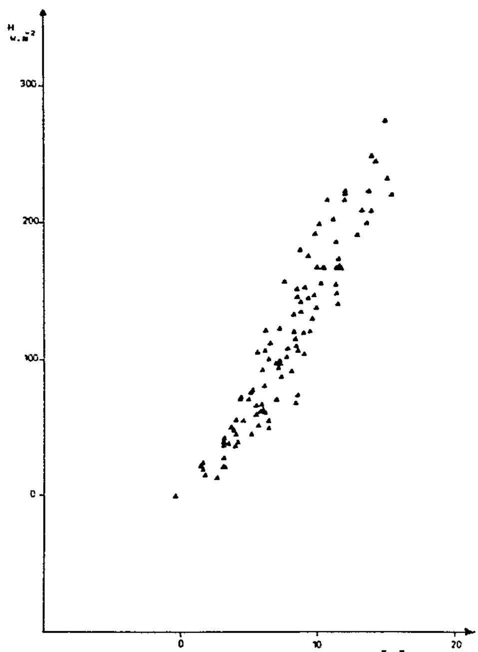
The second site was located several kilometres west of Nefta, between a sandy zone of dunes and the edge of the Chott. There was a water table 1.5 m below the surface. In both cases, there was no vegetation. Average daily wind speed was within the range of 1–4 m s^{-1}, and unstable conditions were prevailing with a Richardson number between -0.25 in the morning and -1.2 in the afternoon.

Statistical relationships found between convective heat flux, H, and soil–air temperature differences were

Tozeur	$H=-28.8+17.3(T_s-T_a)$	with $r=0.94$	(19)
Nefta	$H=-10.4+17.5(T_s-T_a)$	with $r=0.94$	(20)

H is expressed in W m^{-2}. The standard error of the estimate is 30 W m^{-2}. Experimental points are shown in Fig. 1.

These equations summarise the results of 115 hourly values taken between sunrise and sunset, and from both clear days and cloudy days.



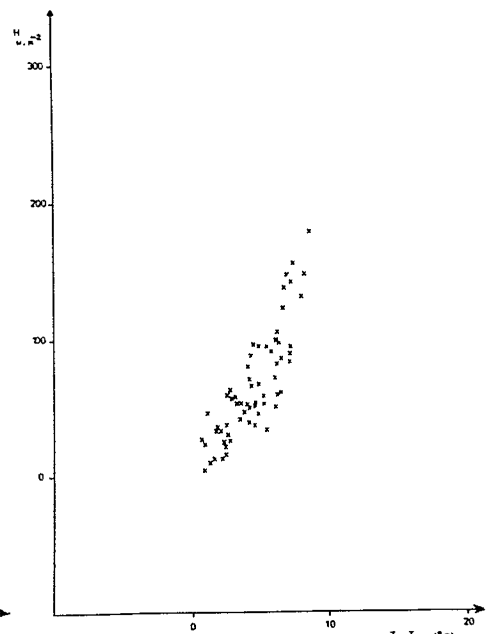


Fig. 1. Relationship between the sensible heat flux, H, and the surface and air temperature differences above dry salt lakes and sand dunes in the Tozeur–Nefta region (South Tunisia).

Fig. 2. Relationship between the sensible heat flux, H, and the surface and air temperature differences above a wheat crop in the Mornag region (North Tunisia).

From the same data, it is also possible to fit a power law equation

$$H=4.95\,|T_s-T_a|^{1.48} \qquad \text{with } r=0.92 \tag{21}$$

This equation, found experimentally, is very similar to the theoretical one derived by Riou and Itier (1982) from the expression of the Priestley coefficient, in the case of highly unstable conditions (eq. 17)

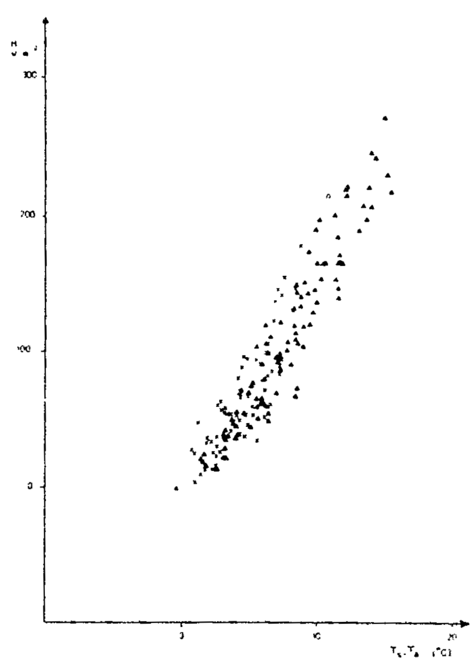
$$H=\alpha\,|\Delta T|^{3/2}$$

where ΔT is the temperature difference between two levels.

Region of Mornag

In contrast, on the Mornag site, in the far north of Tunisia, measurements have been made over wheat at various growing stages. Wind speed conditions were similar to those of the south, but the Richardson number was in the range -0.02 to -0.5. We obtained the following equation

$$H=-6.1+16.1(T_s-T_a) \qquad \text{with } r=0.85 \tag{22}$$



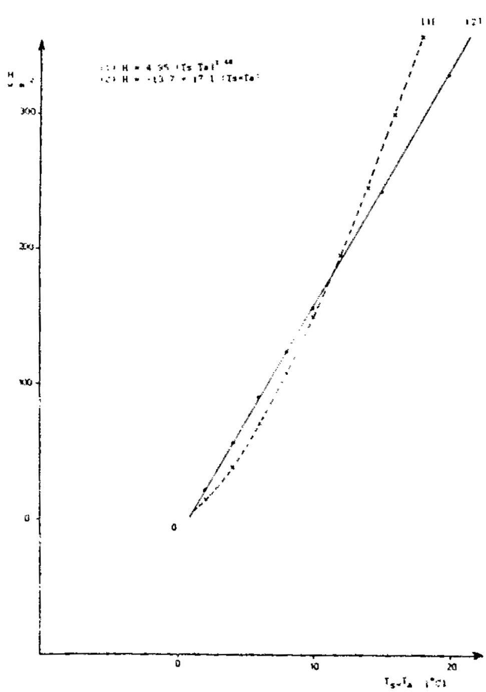


Fig. 3. Whole set of data from the various types of surfaces, showing the relationship between the sensible heat flux, H, and the air and surface temperature differences.

Fig. 4. Experimental curves obtained from the whole set of data.

Experimental points are shown in Fig. 2.

With all the data from Gabes, Tozeur, Nefta and Mornag (a total of 144 pairs, $H, \Delta T$, taken at different hours during the day), the best fit is

$$H = -13.6 + 17.1(T_s - T_a) \qquad \text{with } r = 0.93. \tag{23}$$

Experimental points are shown in Fig. 3.

As can be seen in Fig. 4, for temperature differences $< 15\,^\circ$C, which is mostly the case, this equation leads to H values which are close to those obtained with eq. 21. We can also point out that it seems possible to use the same relationship for different surfaces types, from bare soil surfaces (dry salt lakes from Tozeur and Nefta region), to short vegetation up to 80 cm (wheat crops near Tunis).

APPLICATION OF THIS METHOD TO DETERMINE ACTUAL EVAPORATION RATE

Instantaneous measurements of both radiative surface temperature and air temperature may lead to a satisfactory estimation of H at the same instant, and thus instantaneous actual evaporation, E_i. Finding daily evaporation E_d from E_i can be achieved either by the method proposed by Jackson et al. (1983) or using the H_i/Rn_i ratio at noon.

Using the H/Rn *ratio*

One can observe that on clear days and on uniformly cloudy days, H_i/Rn_i ratios observed around noon (maximum of solar radiation curve) are often very close to H_d/Rn_d ratios (i and d refer, respectively, to instantaneous and total daily values from sunrise to sunset of H, Rn, G). The results for selected days (uniformly clear or uniformly cloudy) and some surface types are shown in Table 1.

Thus, if $H_d/Rn_d \simeq H_i/Rn_i$, we can write $H_d \simeq Rn_d(H_i/Rn_i)$. As $E_d \simeq Rn_d - H_d$, E_d can be expressed by the following equation

$$E_d = Rn_d[1 - (H_i/Rn_i)] - G_d \tag{24}$$

Measurements of G, made at the different sites, showed that in the case of a well developed vegetation, G can be neglected ($G_d < 0.05 Rn_d$), but that with bare soil surfaces, G can fluctuate from 7 up to 30% of the Rn value, according to the soil structure, composition or colour surface, and humidity. We have found average values of G_d/Rn_d for the daylight period close to 0.22.

TABLE 1

Values of H_i/Rn_i and H_d/Rn_d for selected days and various types of surfaces[a]

Date	Site	Hour	H_i/Rn_i	H_d/Rn_d
16/1/86	Gabes	11.00	0.33	
	(d.s.)	12.00	0.41 (0.43)	0.44
		13.00	0.55	
5/3/86	Tozeur	11.30	0.38	
	(ch)	12.00	0.42 (0.43)	0.43
		12.30	0.45	
6/3/86	Tozeur	11.30	0.48	
	(ch)	12.00	0.48 (0.50)	0.52
		12.30	0.54	
13/3/86	Tozeur	11.30	0.56	
	(ch)	12.00	0.59 (0.58)	0.61
		12.30	0.58	
14/3/86	Tozeur	11.30	0.48	
	(ch)	12.00	0.51 (0.49)	0.47
		12.30	0.48	
3/4/86	Mornag	11.00	0.13	
	(wh)	12.00	0.10 (0.13)	0.12
		13.00	0.15	
24/4/86	Nefta	11.00	0.56	
	(sa)	12.00	0.65 (0.67)	0.69
		13.00	0.70	
7/5/86	Mornag	11.45	0.08	
	(wh)	12.00	0.09 (0.08)	0.10
		12.15	0.08	
9/5/86	Mornag	11.20	0.18	
	(wh)	12.00	0.25 (0.21)	0.19
		12.20	0.14	

[a]d.s. = dry steppes; ch = chott (dry salt lake); wh = wheat; sa = sand; H_i/Rn_i = instantaneous values; H_d/Rn_d = average values over the period 9 a.m.–4 p.m., are expressed in solar time.

Calculated examples for 13 March 1986 in Tozeur (Chott surface) and 15 May 1986 in Mornag (wheat crop surface)

13 March 1986, Tozeur

$T_s(12h) = 28.1\,°C$; $T_a(12h) = 16.9\,°C$; $Rn_i(12h) = 345$ W m^{-2}

$Rn_d = 1602$ W m^{-2}; $G_d = 366$ W m^{-2}

From eq. 23 we can write that

$$H_i(12h)=13.6+17.1\ [T_s(12h)-T_a(12h)]$$
$$=178\ \mathrm{W\ m^{-2}}$$

From eq. 24 we can write that

$$E_d=Rn_d[1-(H_i/Rn_i)]$$

$$E_d=1602\ [1-(178/345)]=769\ \mathrm{W\ m^{-2}}$$

$E_d=1.13$ mm and if we take G into account,

$E_d=0.87$ mm

The E_d value measured from all the 15 min data is 0.95 mm.

15 May 1986, Mornag

The same calculation can be made for 15 May 1986 in Mornag, for a wheat crop surface, it leads to the following results

$E_d=3.87$ mm calculated from eqs. 21 and 22

$E_d=3.67$ mm calculated from eqs. 23 and 24 with G being taken into account.

E_d calculated from all the 15 min data is 3.45 mm.

CONCLUSIONS

These results show the following.

(i) Over a fairly large range of wind speed conditions, but provided they are in unstable conditions, which are the most frequent we have encountered during the daylight period, the heat flux H can be estimated quite satisfactorily with measurement of just radiative surface temperature and air temperature.

(ii) In those conditions, the relationship between H and (T_s-T_a) seems to be independent of surface roughness, at least in the interval $0.1>z_0>10$ cm.

(iii) These results confirm that according to Priestley (1959), in the case of free convection, which were the conditions we have mainly encountered in the field, the influence of u^* is very small and any change in u_2-u_1 will not have much effect on H. This is shown by the good agreement found between the theoretical equation derived by Riou and Itier (eq. 17) (using only the temperature difference between two levels above the surface) and the experimental relationship found between (T_s-T_a) and H.

(iv) If Rn can be measured or estimated, an instantaneous value of actual evaporation can be obtained from a simple measurement of the radiative surface temperature. Daily evaporation can then be obtained from the instantaneous measurement using methods proposed by Jackson et al. (1983), or using the noon H/Rn ratio as we have shown.

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LIST OF SYMBOLS

Cp	Heat capacity of air
d	Zero plane displacement
E	Evaporation
g	Gravity
G	Soil heat flux
H	Sensible heat flux
h^*	Priestley coefficient
k	Von Karman constant
L	Monin–Obukov length
Ra	Aerodynamic resistance
Ri	Richardson number
Rn	Net radiation
T, T_1, T_2, T_s, T_a	Temperatures
u, u_1, u_2	Wind speeds
u^*	Friction velocity
z, z_1, z_2	Heights
z^*	$(z_1 z_2)^{1/2}$
z_0	Surface roughness
α	Coefficient
ΔT	Temperature difference between z_2 and z_1
Δu	Wind speed difference between z_2 and z_1
Θ	Potential temperature
Θ^*	Surface layer potential temperature scale
ρ	Density of air
$\Phi M, \Phi H$	Universal flux-gradient functions
$\psi M, \psi H$	Paulson function

REFERENCES

Bartholic, J.F., 1972. Aerial thermal scanner to determine temperatures of soils and crop canopies differing in water stress. Agron. J., 64: 603–608.

Brown, K.W. and Rosenberg, N.J., 1973. A resistance model to predict evapotranspiration to a sugar beet field. Agron. J., 65: 341–347.

Businger, J.A., 1988. A note on the Businger–Dyer profiles. Boundary Layer Meteorol., 42: 145.

Businger, J.A., Wyngaard, J.C., Izumi, Y. and Bradley, E.F., 1971. Flux-profile relationships in the atmospheric surface layer. J. Atmos. Sci., 28: 181–189.

Dyer, A.J. and Hicks, B.B., 1970. Flux-gradient relationship in the constant flux layer. Q. J. R. Meteorol. Soc., 96: 715–721.

Hatfield, J.L., Perrier, A. and Jackson, R.D., 1983. Estimation of evapotranspiration at one time of day using remotely sensed surface temperature. Agric. Water Manag., 7: 341–350.

Itier, B., 1980. Une méthode simplifiée pour la mesure du flux de chaleur sensible. J. Rech. Atmos., 14: 17–34.

Jackson, R.D., Reginato, R.J. and Idso, S.B., 1977. Wheat canopy temperature: a practical tool for evaluating water requirements. Water Resour. Res., 13: 651–656.

Jackson, R.D., Hatfield, R.L., Reginato, R.J., Idso, S.B. and Pinter, P.J., 1983. Estimation of daily evapotranspiration from one time-of-day measurements. Agric. Water Manage., 7: 351–362.

Lorentz, D., 1968. Temperature measurements of natural surfaces using infrared thermometers. Appl. Opt., 7: 1707–1710.

Monin, A.S. and Obukov, A.M., 1954. Basic laws of turbulent mixing in the ground layer of the atmosphere. Tr. Inst. Geofiz. Akad. Nauk SSSR, 24: 163–167.

Monin, A.S. and Yaglom, A.M., 1971. Statistical Fluid Mechanics: Mechanics of Turbulence, Vol. 1. The MIT Press, Cambridge, MA, 769 pp.

Paulson, C.A., 1970. The mathematical representation of wind speed and temperature profiles in the unstable atmospheric surface layer. J. Appl. Meteorol., 9: 857–861.

Priestley, C.H.B., 1959. Turbulent Transfer in the Lower Atmosphere, University Press, Chicago, IL, pp. 39–51.

Recan, M., 1982. Modélisation mathématique du comportement thermique d'un sol nu. Application à la télédétection dans l'infrarouge. These Doct. Ing. Inst. Mec. Fluides, Toulouse, 182 pp.

Riou, Ch. and Itier, B., 1982. Une formulation des flux conduisant à une simplification de la méthode aérodynamique combinée de mesures de l'évapotranspiration réelle. Congres C.I.I.D., Fort Collins, U.S.A.

Seguin, B., 1984. Estimation de l'évapotranspiration à partir de l'infrarouge thermique. 2e Colloque Int. Signatures spectrales d'objects en télédétection, 12–16 September, at Bordeaux.

Seguin, B. and Itier, B., 1983. Using midday surface temperature to estimate evaporation from satellite thermal IR data. Int. J. Remote Sensing, 4: 371–383.

Stone, L.R. and Horton, M.L., 1974. Estimating evapotranspiration using canopy temperatures: field evaluation. Agron. J., 66: 450–454.

Webb, E.K., 1970. Profile relationship: The log-linear range, and extension to strong stability. Q. J. Meteorol. Soc., 96: 67–90.

Wendler, G. and Eaton, F., 1983. On the desertification of the Sahel zone. Climatic Change, 5: 365–380.

Wiegand, L. and Bartholic, J.F., 1970. Remote sensing in evapotranspiration research on The Great Plains. Proceedings of Evapotranspiration in the Great Plains Seminar, Research Committee Great Plains Agricultural Publ. No. 50.