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Using midday surface temperature to estimate daily evaporation from satellite thermal IR data

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Abstract. The practical experience obtained by our participation in the European project TELLUS (a part of the HCMM programme) led us to state that the thermal inertia concept and sophisticated models are useful for understanding basic processes and for performing informative simulations, but cannot be used for a real estimation of evaporation.

On the other hand, a simplified procedure like that proposed by Jackson *et al.* (1977) appears feasible, but its theoretical basis and its field of application are subject to discussion. Two main questions arise, namely, the conditions underlying the use of one instantaneous measurement to estimate daily integrated values, and the significance of the adjustment parameter B in the simplified relationship $ET_{\mathrm{d}} = Rn_{\mathrm{d}} - B(Ts - Ta)_{\mathrm{i}}$ which relates the daily evaporation to daily net radiation by means of one measurement of the surface and air temperature, at a given time of day.

These questions are discussed with reference to results of recent experiments. A theoretical analysis of the B parameter was undertaken to explain the difference between the value (0·25) derived from an experiment at Crau (south of France) and the value (0·64) proposed by Jackson *et al.* (1977) for conditions in Phoenix. The influences of wind velocity, thermal stratification and surface roughness are discussed. Consequences of the practical use of the simplified procedure are presented.

1. Introduction

The possibility of using thermal IR data for estimating evaporation from satellite measurements has been extensively studied during the past few years. The basic processes, such as combining the equation of surface energy budget and the laws of convective exchanges, have been tested by small-scale ground experiments (Bartholic *et al.* 1972, Brown 1974, Stone and Horton 1974, Idso *et al.* 1975, Heilman and Kanemasu 1976). Recently available satellite data (e.g. NOAA-6, TIROS-N, HCMM and METEOSAT) allow large scale estimations of evaporation to be attempted.

2. Some conclusions from the TELLUS programme

The TELLUS programme, a group of European research laboratories co-ordinated by the Joint Research Center of European Communities (Ispra, Italy), was a contribution to the global HCMM experiment of NASA. From the large scientific exchange generated by that programme during its 3 years, we have derived a number of practical conclusions concerning the estimation of evaporation using satellite thermal data (which represent our personal point of view).

2.1. *Limitations of thermal inertia for estimating evaporation*

The concept of thermal inertia, upon which the HCMM project was based, appears to adequately express qualitative variations in water availability (Price 1980, Reiniger *et al.* 1981). However, its practical use for a routine quantitative estimation of evaporation is limited by the relatively few combined day–night data sets. Existing data sets are useful for testing models such as TELLUS (Rosema *et al.* 1979), but do not allow realistic estimations of evaporation on a long-term basis.

Moreover, as far as surface-moisture problems are concerned, differences due to moisture variations are considerably greater during the day than at night. Typical temperature variability due to dryness differences amounts to 10–15° during the day compared to 1–3° during the night. Night-time measurements are informative concerning the structure of vegetative canopies (Boissard *et al.* 1981), but their contribution to thermal inertia is low as far as water availability is concerned. We have explored the use of day-time data only, thus avoiding the technical limitations of the combined day–night passes.

2.2 *Models*

The limitation of one satellite acquisition per day (when available) leads to the use of models like TELLUS (Rosema *et al.* 1979), or TERGRA (Soer 1980) in order to extrapolate to the entire day. However, a large number of micrometeorological parameters (thermal properties of soil, surface roughness, wind velocity, etc.) must be specified. Obtaining these parameters is possible for extensively equipped test areas, but cannot be obtained for use in remote sensing over large regions that include a wide range of natural surfaces. The models are useful tools for simulating the course of basic processes and for understanding the influence of the various parameters involved, but are too complicated to be used as long-term estimation procedures. The problem is to determine to what extent the loss of precision due to rough assumptions may affect the significance of the final results.

2.3. *Results from the Crau experiment*

The experiment performed in Crau (Seguin *et al.* 1982) showed that using a ground measurement of surface temperature (thus obviating the problem of accuracy of satellite data), the long-term performance of a continuous estimation of evaporation using the classical approach

$$ET = Rn - S - H \tag{1}$$

with

$$H = \rho Cph(Ts - Ta) \tag{2}$$

was not really better than crude approximations such as $ET = Rn$ (for humid conditions), or $ET = ET_{eq} = Rn\Delta/(\Delta + \gamma)$ (for dry conditions). In equations (1) and (2), ET is the evaporation, Rn is the net radiation, S is the ground heat flux, H is the sensible heat flux, ρ is the air density, Cp is the specific heat of air, h is the exchange coefficient at the height z of measurement of the air temperature (Ta) and Ts is the surface temperature. The term ET_{eq} is the equilibrium evaporation corresponding to the first term of the Penman equation. The range of precision of these methods was 10–20 per cent for a 100 day period.

If the estimate of the absolute value of ET over large regions is not improved by use of surface temperatures, then their usefulness lies in delineating spatial differences in ET between dry and wet zones, a measure that the crude estimations ($ET = Rn$ and $ET = ET_{eq}$) cannot give.

We believe that the main use for remote sensing, in its present stage, lies in the description of spatial variations of ET more than in estimating an overall regional value for ET. Since sophisticated models also do not specify spatial variability, simplified procedures are of interest if they can express the spatial variations of ET and estimate the absolute values within a tolerable range of error (10–20 per cent). That seems to be the case for the method of estimation for daily evaporation proposed by Jackson *et al.* (1977), which will be extensively discussed in the following sections.

3. Discussion of Jackson *et al.*'s (1977) simplified procedure

The basic idea consisted of relating daily evaporatin (ET) to daily net radiation (Rn) and the instantaneous temperature difference ($Ts - Ta$) near midday, i.e.

$$ET_d = Rn_d - B(Ts - Ta)_i \tag{3}$$

First, this relation obviously depends on the assumption that an instantaneous measurement of $Ts - Ta$, which is related to the instantaneous value of the sensible heat flux H_i by the means of the exchange coefficient h_i, adequately expresses the influence of the daily value of H in the energy balance equation applied to the whole day. Moreover, the nature of the relation between $Ts - Ta$ and H needs to be analysed. A question arises as to whether it is feasible to express a statistically-derived mean value for the exchange coefficient h_i as a constant (B) when complex models clearly show a marked influence of several micrometeorological parameters.

Before examining theoretical aspects of the above assumption, it is informative to establish that the procedure is well founded from an experimental point of view. Two experiments were carried out in the Avignon region (south-east France). The results have been published (Seguin 1979, Seguin *et al.* 1982) and will not be repeated here except for the main conclusions.

The first experiment (Seguin 1979) was performed during the summer of 1977 on a grass field (70 × 120 m). The data, given in figure 1 (dots), show a quasilinear relationship between $ET_d - Rn_d$ and $(Ts - Ta)_i$, but somewhat different from the line

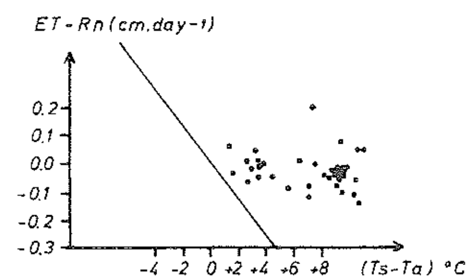


Figure 1. Relationship between $ET - Rn$ and $Ts - Ta$ from the Avignon experiment (dots) and from the Phoenix experiment (line) (from Seguin 1979).

representing Jackson's results. The second experiment (Seguin *et al.* 1982) was on a regional scale and was based upon measurements over large homogeneous units of the Crau plain. Simultaneous measurements over irrigated (50 km^2) and dry (170 km^2) zones of the Crau are given in figure 2.

The general trend of the data is similar to the results of the first experiment. The number of measured points allowed the calculation of the relationship

$$ET_d = Rn_d + 1{\cdot}0 - 0{\cdot}25(Ts - Ta) \tag{4}$$

With ET and Rn expressed in millimetres per day. The corresponding relation given by Jackson *et al.* (1977) is

$$ET_d = Rn_d - 0{\cdot}64(Ts - Ta) \tag{5}$$

Although other experiments are needed to establish the validity of this procedure under different environmental conditions, these experiments confirm that the relationship holds, at least for a given location and for a given surface. The results indicate that a unique value of $Ts - Ta$ at one time of the day may be useful for predicting daily evaporation (at least for clear days). Concerning the question of the constancy of B, the results are less positive, since we found a value fairly different from the original parameter reported by Jackson *et al.* (1977).

Although the above statements are supported by experimental data, they must be examined on a more general basis if the simplified procedure is to have global applicability.

4. Midday temperature data as representative for the whole day

A detailed analysis is not necessary to establish that only data for clear days can be used with this method. For the case of intermittent cloudiness or changing weather, the representativeness of midday data has to be considered in terms of probability. This limitation would not apply to continuous cloudy weather, but

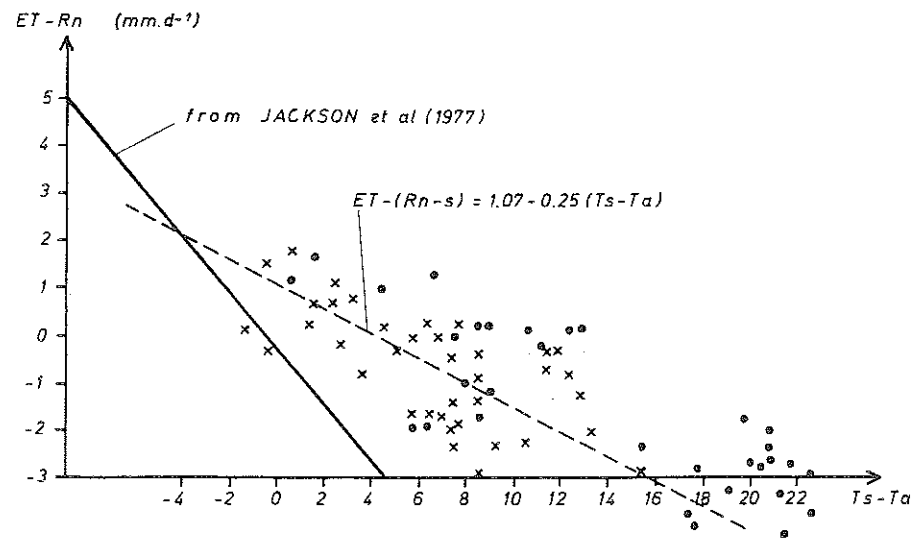


Figure 2. Relationship between $ET - Rn$ and $Ts - Ta$ from the Crau experiment (×, irrigated; ●, dry) during the summers of 1978 and 1979, compared to the original relationship of Jackson *et al.* (1977) (solid line) (from Seguin *et al.* 1982).

thermal IR data cannot be obtained from satellites or aircraft under these conditions.

Limiting the discussion to clear days, the well known regular course of climatic parameters during the day suggests that midday values can be representative of the entire day. That general statement is supported by recent results (Itier and Riou 1982) which were based upon an experiment in Voves (centre of France). Midday values of the partition H_i/Rn_i were found to be in good agreement with the same partition of the daily integrated fluxes H_d/RN_d. Measurements were performed over two surfaces, wheat stubble and sugar-beets (figure 3). These results indicate that the use of midday instantaneous values to represent daily values has no serious limitation other than the restriction of using only clear-day data.

5. Analytical expression for the adjustment parameter B

The above cited values for B were empirically derived by statistical regression of experimental data. The parameter B needs to be discussed from an analytical point of view to evaluate its dependence upon micrometeorological parameters.

On a daily scale, the soil heat flux S_d can be neglected and we can write equation (1) as

$$ET_d = Rn_d - H_d \tag{6}$$

We assume, as did Itier and Riou (1982), that

$$H_d/Rn_d = H_i/Rn_i \tag{7}$$

as a first-order approximation. Equation (7) implies that the Bowen ratio $\beta = H/ET$ is more or less constant throughout the day. This is not strictly true because β is known to vary, especially at night. However, the magnitude of the midday periods of daily evaporation is such that the resulting error is not large.

With the above assumption we can write

$$ET_d = Rn_d - H_i Rn_d/Rn_i = Rn_d - (Rn_d/Rn_i)\rho Cph_i(Ts - Ta)_i \tag{8}$$

Equation (8) shows that B can be statistically derived by regression of $ET_d - Rn_d$ values versus the corresponding $(Ts - Ta)_i$ measurements. It follows that

$$B \simeq \overline{(Rn_d/Rn_i)\rho Cph_i} \tag{9}$$

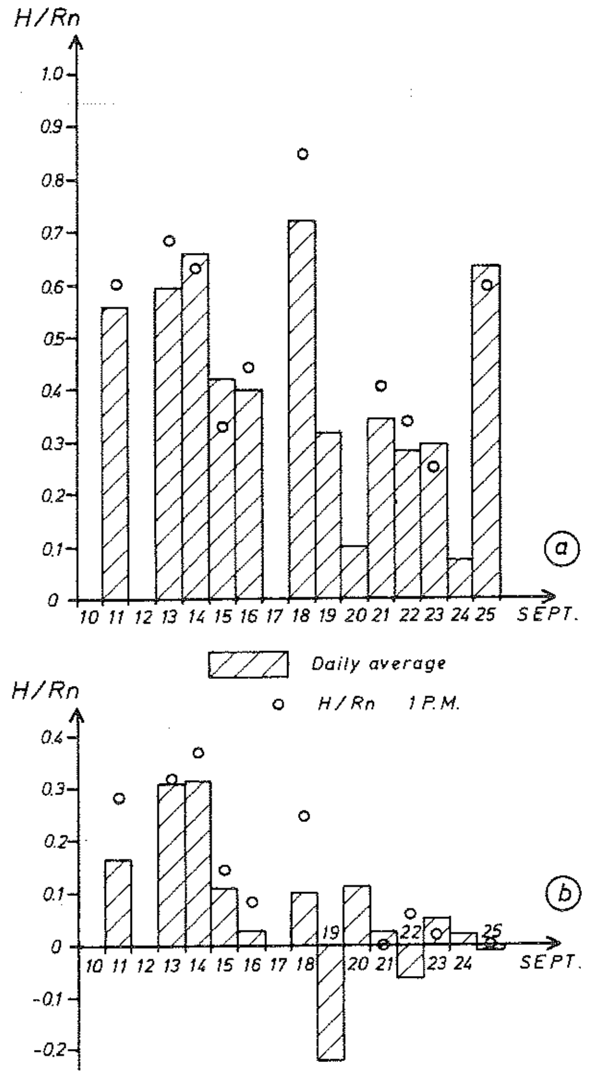
If we assume that the ratio Rn_d/Rn_i for clear days is reasonably constant (this will be justified later), we obtain

$$B \simeq (Rn_d/Rn_i)\overline{\rho Cph_i} \tag{10}$$

The term B may be defined as a 'mean exchange coefficient' which is weighted by the ratio Rn_d/Rn_i. The ratio expresses the relative contribution of midday net radiation to the integrated whole day radiative exchange.

A detailed analysis would be needed to establish the range of variation of this ratio. We have restricted our discussion to clear summer days, the hypothesis of a constant mean value may be assumed as a first approximation. The analysis of 3 years of data (for summer months in Avignon) gave a value of 0.3 ± 0.03 for Rn_d/Rn_i.

Within the limits of a first-order approximation, the exchange coefficient h_i can



10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	Dates
0.2	0	0.3	0.1	0	0	0.3	0	0	0.7	11.8	0.1	1.5	0.2	2.2	0	Rain fall (mm.)

Figure 3. Comparison of the daily H_d/Rn_d values to the instantaneous H_i/Rn_i values at 13.00 hours on sunny days for a period during September 1980 in Voves (France): (*a*) above wheat stubble, (*b*) above sugar-beet (from Itier and Riou 1982).

be written for the neutral case as

$$h_i = \frac{k^2}{[\log\ (z/z_0)]^2}\, u_i(z) \tag{11}$$

where k is the von Karman constant ($\simeq 0{\cdot}4$), $u_i(z)$ is the wind velocity at height z and z_0 is the roughness parameter.

Taking $z = 2$ m (a standard height), $z_0 = 1$ cm (typical for a grass surface) and

$u(2\,\mathrm{m}) = 3\,\mathrm{m/s}$ as a middle-range value, we calculate for B a typical value of 0·22, which is in good agreement with experimental values (0·25 for our case). The corresponding value for a wheat canopy (z_0 being taken as 10 cm) would be about 0·66, which compares favourably to the value of 0·64 obtained for the Phoenix data.

From an analytical point of view, these crude calculations can only be expected to indicate the order of magnitude of B. Since the calculated values are reasonably close to the statistically derived values, it appears worth while to use the analytical expression (10) to establish the range of significance of the parameter B.

Even if the ratio Rn_d/Rn_i may vary somewhat, the evident source of variation of B stems from the variation of h_i, which is known to be related to wind velocity, thermal stratification and surface roughness.

Using the first-order expression for neutral conditions (11), it follows that B is proportional to u, and may vary by a factor of 6 when z_0 increases from 1 mm to 10 cm (which roughly corresponds to a transition from a short-grass cover to tall crops). The hypothesis of one single value for B to be applied to various climates and surfaces, appears unacceptable, even as an approximation for long time periods that damp day-to-day variations. It is obvious that a detailed analysis concerning the influence of the various parameters (wind velocity, thermal stratification and surface roughness) is necessary.

6. The influence of wind velocity and thermal stratification

The approximation for h_i (equation (11)) must be corrected for stability effects. This could be achieved using the commonly accepted universal functions of Ri (the Richardson number) or z/L (L is the Monin–Obukhov length), but the complete treatment would require iterative procedures, which are rather complicated. However, for the case of unstable stratification ($Ts - Ta > 0$), which was the conditions encountered in the Crau experiment, a direct procedure is possible (Itier 1980). We have tried to adapt this procedure to obtain analytical expressions for B, for the unstable case.

6.1. *The unstable case ($Ts - Ta > 0$)*

The unstable case is the normal day-time situation, especially near noon when the simplified procedure is usually aplied. Under sufficiently high instability conditions ($|\mathrm{Ri}| > 0{\cdot}015$), H may be written as (Itier 1980, Itier and Riou 1982):

$$H = \alpha (\Delta T)^{3/2} \tag{12}$$

with ΔT being the difference between the air temperatures at two levels z_1 and z_2, and α is expressed by

$$\alpha = a \frac{Cp(g/T)^{1/2}}{5{\cdot}2(a_2^{-1/3} - z_1^{-1/3})^{3/2}} \tag{13}$$

with $a = 1{\cdot}3 \pm 0{\cdot}3$, if $|\mathrm{Ri}| > 0{\cdot}015$.

Assuming $Rn_d/Rn_i = 0{\cdot}3$ (keeping in mind the limitations due to Richardson-number values), and extrapolating the relations (12) and (13) down to the surface by considering $T_{(z_1)} = Ts$ when $z = z_0$, and $T_{(z_2)} = Ta$ when $z = 2\,\mathrm{m}$, we obtain the relationships

$$z_0 = 1\,\mathrm{mm}, \qquad ET_d - Rn_d = 0{\cdot}021(Ts - Ta)^{3/2}$$

$$z_0 = 2\,\mathrm{mm}, \qquad ET_d - Rn_d = 0{\cdot}033(Ts - Ta)^{3/2}$$

$$z_0 = 1\,\text{cm}, \qquad ET_d - Rn_d = 0{\cdot}080(Ts - Ta)^{3/2}$$

$$z_0 = 10\,\text{cm}, \qquad ET_d - Rn_d = 0{\cdot}360(Ts - Ta)^{3/2}$$

These relationships are reported in figure 4, together with the experimental equation (4) obtained from our experiment in Crau.

The results given in figure 4 indicate that the analytical calculations are in agreement with the experimental results for the Crau experiment, which included two different surface conditions (irrigated pastures, $z_0 = 1$ cm; and dry-land short grass, $z_0 = 2$ mm).

An independent confirmation of these results was reported by Recan (1982), who used a coupled heat and mass transfer model for porous media applied to the Crau meteorological parameters. This author obtained the following relationships:

for sandy soil,

$$ET_d - Rn_d = 0{\cdot}97 - 0{\cdot}25(Ts - Ta) \tag{14}$$

and for clay

$$ET_d - Rn_d = 0{\cdot}94 - 0{\cdot}25(Ts - Ta) \tag{15}$$

which agree quite well with the empirical results obtained from the Crau experiment (equation (4)).

To be complete for the unstable case, we must also consider the conditions corresponding to $|\text{Ri}| < 0{\cdot}015$, which were not directly calculated by Itier and Riou (1982). Within this range, the effects of thermal stratification are almost negligible, so we can use the expressions for neutral stability already treated in §5. Taking $u(2\,\text{m}) = 3$ m/s, we obtain from equation (11) values of B ranging from 0·17 for $z_0 = 2$ mm to 0·22 for $z_0 = 1$ cm. Since values of $|\text{Ri}| < 0{\cdot}015$ are generally associated with wind velocities (at 2 m height) larger than 3 m/s, a global mean value of 0·25 for B for the unstable case $(Ts - Ta > 0)$ appears to be completely consistent with analytical calculations.

6.2. *The stable case* $(Ts - Ta < 0)$

The stable case is not the usual situation for midday conditions. It can be encountered as the result of advection. The case for particular conditions such as

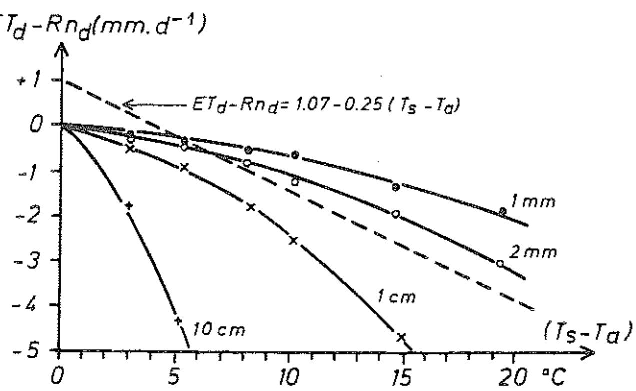


Figure 4. Comparison of the theoretically derived relationship between $ET_d - Rn_d$ and $(Ts - Ta)$, for various values of z_0, and the experimentally derived equation (4).

irrigated fields surrounded by dry lands will not be considered because, in general, remote-sensing applications are usually concerned with large surfaces and the results obtained from small fields cannot be extrapolated. However, large-scale advection may operate for some situations, for example, during dry summers in the Great Plains of the U.S.A. A detailed re-examination of the Phoenix data reveals that the original relationship proposed by Jackson *et al.* (1977) corresponds to advective situations, since the experimental data show that $Ts-Ta<0$.

A direct analytical treatment similar to that developed for the unstable case can be made using the universal stability functions (Webb 1970). The exchange coefficient for stable conditions h_s is

$$h_s=\frac{k^2}{[\log(z/z_0)+5z/L]^2}u(z) \tag{16}$$

which, after some manipulations and using equation (11) for h_n for neutral conditions, gives

$$h_s=h_n(1-5R_i)^2$$

The expression for B under stable conditions is

$$B_s=B_n(1-5R_i)^2$$

where B_n is the expression for B under neutral conditions. Taking R_i at $z=(2z_0)^{1/2}$, we obtain a reduction factor

$$B_s/B_n=[1-0{\cdot}2(Ts-Ta)/u^2]$$

which is equal to 0·8 for $u=3$ m/s and 0·56 for $u=2$ m/s, with $Ta-Ts=5°$ for both cases.

This result indicates that typical values of B would be lower under stable conditions by a factor of about 0·8. For a surface having a roughness of 1 cm (B about 0·18), the range of observed $Ts-Ta$ values cannot explain the high value of B (0·64) obtained at Phoenix. In addition to a high surface roughness the discrepancy could be attributed to the small dimensions of the experimental plots (12 × 90 m) which may create local advective effects not representative of large-scale regional advection.

6.3. *Preliminary conclusions*

From the above analytical treatments we can state the following:

(1) The experimental values of B derived from the Phoenix and the Crau studies are consistent with a simplified analysis of energy exchanges.

(2) As a first approximation, the effects of wind velocity and thermal stratification can be reduced to two cases, unstable ($Ts-Ta>0$) and stable ($Ts-Ta<0$), with the corresponding values of B being about 0·25 and 0·18, respectively. We should remember that, for any given day, large deviations from the mean value may be noted because of variations in wind velocity. The simplified procedure must be applied to long-term periods (at least several weeks), so that daily extremes are damped and the mean values can be used with confidence.

The calculations on which the above tentative conclusions were derived included only the influence of climatic parameters, although the surface-roughness factor was noted several times. A detailed discussion appears warranted to establish the effect of the surface roughness.

7. The influence of surface roughness

Equation (11) and the data in figure 4 demonstrate the strong influence of surface roughness on exchange processes. For operational purposes, the value of the parameter B to be used in the simplified procedure must be closely related to the type of surface encountered. Since a precise value of z_0 cannot be specified for remote-sensing purposes, an approximate classification containing four classes is suggested,

1 mm	smooth bare soil
1 mm to 1 cm	bare soil, grass, pastures, etc.
1 cm to 10 cm	wheat, tall crops, shrubs, etc.
10 cm to 1 m	orchards, forests

Using equation (11) or figure 4, associated ranges of B could be developed.

While the above procedure is acceptable, it should be considered with caution because the true effect of surface roughness is not readily apparent, especially for tall canopies. During the past few years a number of authors have reported results about the non-validity of the Reynolds analogy near rough surfaces and the consequent excess resistance for heat transfer as compared to momentum transfer (Owen and Thomson 1963, Chamberlain 1968, Brutsaert 1975, Thom 1976, Garratt 1978). We will not present a detailed analysis of that effect here, but we will discuss its consequences, namely a very damped influence of surface roughness on turbulent exchanges. An illustration of this effect is presented in figure 5, where the variation of the global heat-exchange coefficient C_H is shown to be a highly damped function of z_0 compared to the momentum coefficient C_d (Brutsaert 1975). Not taking into account the excess resistance, as discussed above, corresponds to assuming $C_H = C_d$, and we can easily see the marked difference between the two coefficients, especially for high surface roughnesses.

Recan (1982) found that the heat-exchange coefficient was increased by 1·6–1·8 when z_0 increased from 1 mm to 1 cm. This can be compared to multiplying the

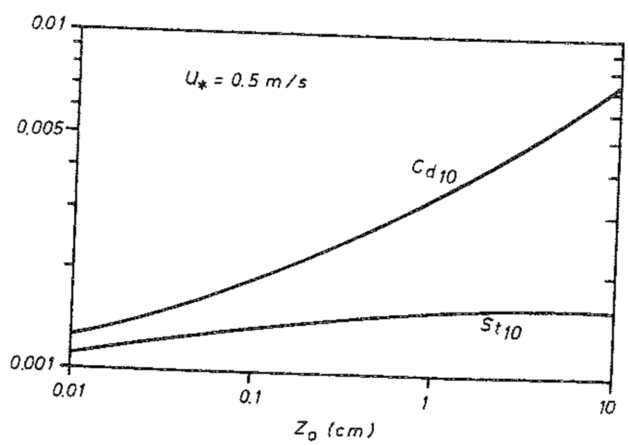


Figure 5. The effect of surface roughness z_0 on the heat-transfer coefficient $C_H = St$ and the momentum transfer coefficient C_d taken at a reference height of 10 m (adapted from Brutsaert 1975).

coefficient by 4 when the excess resistance is neglected (with the same change in z_0). We have calculated the variation of the heat-exchange coefficient for the range of z_0 from 1 mm to 10 cm by several procedures: two that do not incorporate the excess resistance (equations (11) and (13)) and two that do (Seguin 1973, Thom 1976). Results of the calculations are given in the table.

Relative variation of the heat-exchange coefficient as a function of z_0 using several modes of calculation (1 mm was used as a reference).

Mode of calculation	z_0 1 mm	2 mm	1 cm	10 cm
Equation (11)	1·0	1·19	2·06	6·4
Equation (13)	1·0	1·45	3·82	17·3
Seguin (1973)	1·0	1·10	1·25	1·24
Thom (1976)	1·0	1·17	1·78	4·0

Although these results are only indicative of the problem, we can draw the following conclusions:

(1) Taking into account the excess resistance dampens the influence of the surface roughness.

(2) For high values of z_0, the differences between the several modes of calculation are sufficiently great that two ways of estimating evapotranspiration are evident, and we must choose one or the other.

If we accept the calculations of Brutsaert (1975), Recan (1982) and Seguin (1973) we could assume a limited variation of h_i with surface roughness and then develop a 'universal' value for B. It could then be of the order of 0·25 for the unstable case ($Ts-Ta>0$) and about 0·18 for advective conditions ($Ts-Ta<0$). On the other hand, if the excess resistance is ignored, or its effect is less marked as in the Thom formulation, the previous suggestion of classifying surfaces into several main groups with adjusted B parameters for each group must be adopted. As previously noted, the results of the Phoenix experiment may support that point of view if the observed value of 0·64 can be explained by the roughness length of about 10 cm for a wheat canopy. That point is uncertain, however, because of possible effects of small-scale advective perturbations.

At present, it is not possible to go further than these tentative conclusions. This is especially true for tall canopies, where the general question of expressing the surface temperature as a function of canopy structure is not resolved, and the purely aerodynamical effect of surface roughness cannot be separated from the radiative effect. We believe that judgements upon the significance of the thermal IR emission from a tall canopy as related to its structure can only be made after additional research has been completed.

8. Conclusion

The analytical treatments we have presented provide a means of evaluating the simplified procedure proposed by Jackson *et al.* (1977) on a theoretical rather than an empirical basis.

Considering only clear days, the main idea of relating daily evaporation to midday surface-temperature data appears well founded. The question of universality of the parameter B is, however, open to discussion. If we can accept considerable variability in daily values and modest precision (say 20 per cent) for results summed over a 2–4 week period, a simple partition into the unstable case ($Ts - Ta > 0$) and the advective case ($Ts - Ta < 0$) would be sufficient. For these two cases we propose the following relationships,

$$Ts - Ta > 0, \qquad ET_d - Rn_d = 1 \cdot 1 - 0 \cdot 25(Ts - Ta)$$

$$Ts - Ta < 0, \qquad ET_d - Rn_d = -0 \cdot 18(Ts - Ta)$$

These relationships apply only to 'medium rough' surfaces (typically with z_0 ranging from 1 mm to 1 cm). For other surfaces (especially for tall crops like orchards and forests), the hypothesis that these relationships are valid cannot be rejected. However, for these surfaces the excess resistance and the determination of Ts as a function of radiative exchanges inside the canopy structure, must be evaluated. For this case, four or five values that correspond to various degrees of surface roughness could be developed, instead of a 'universal' value.

Overall, the simplified procedure is not in contradiction with theoretical analyses, and can be considered as a valuable first step for using thermal IR satellite data to estimate evaporation. Its precision is not high, but it compares favourably to current ground estimates that are subject to spatial variations. The difficulty of obtaining meaningful surface-temperature data from satellites (problems include atmospheric effects and calibration) needs to be considered when making operational computations. The correct use of the simplified procedure requires the availability of at least one correct ground reference, for calibration purposes, if one is to derive spatial variations of evaporation from satellite data.

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