Efficient Prediction of Ground Surface Temperature and Moisture, With Inclusion of a Layer of Vegetation

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An efficient time-dependent equation for predicting ground surface temperature devised by Bhumralkar (1975) and Blackadar (1976) is tested against a 12-layer soil model and compared with five other approximate methods in current use. It is found to be generally superior if diurnal forcing is present and very much superior to the use of the insulated surface assumption. An analogous method of predicting ground surface moisture content is presented which allows the surface to become moist quickly during rainfall or to become drier than the bulk soil while evaporation occurs. These improved methods are not of much relevance unless the main influences of a vegetation layer are included. An efficient one-layer foliage parameterization is therefore developed that extends continuously from the case of no shielding of the ground by vegetation to complete shielding. It includes influences of both ground and foliage albedos and emissivities, net leaf area index, stomatal resistance, retained water on the foliage, and several other considerations. When it is tested against data for wheat measured by Penman and Long (1960), it appears quite adequate despite the many simplifying assumptions. The parameterization predicts that errors of up to a factor of 2 in evapotranspiration can be incurred by ignoring the presence of a vegetation layer.

1. INTRODUCTION

The determination of ground surface temperature within a numerical weather prediction model is usually accomplished by solution of a surface energy balance equation. A troublesome component, however, is the soil heat flux, which apparently requires the time-dependent solution for soil temperature within six or more layers of soil for reasonably good accuracy [*Benoit,* 1976]. Since the number of layers in the soil may then be comparable to the number of atmospheric layers comprising the model, abbreviations are usually sought for dealing with the soil heat flux. However, the abbreviations presently in use are very crude.

The prediction models of the Geophysical Fluid Dynamics Laboratory (GFDL) of Princeton have ignored the soil heat flux entirely [see *Manabe et al.,* 1974], as has also a version of the UCLA two-level model [*Gates,* 1975]. The National Center for Atmospheric Research (NCAR) model attempts to improve upon this omission by assuming that the soil heat flux, $G = -\lambda(\partial T/\partial z)_0$, is one third the sensible heat flux H_s to the atmosphere. Here λ is the soil thermal conductivity, T is temperature, and subscript zero refers to evaluation at the ground surface. (As is customary, G is defined as positive when directed downward; atmospheric fluxes will be defined as positive when directed upward.) A list of symbols used in this paper is found in the notation list at the end of this paper. The proportionality constant, $\frac{1}{3}$, was chosen by *Kasahara and Washington* [1971] on the basis of the study of *Sasamori* [1970], who simulated a measurement period of the O'Neill experiment [*Lettau and Davidson,* 1957]. Another possibility is to assume that G is proportional to the net radiative flux R_{net}; a proportionality constant of about -0.4 is suggested by the study of *Idso et al.* [1975*a*], while a value of -0.19 when the net radiative flux is downward and -0.32 when it is upward has been recommended by *Nickerson and Smiley* [1975] for O'Neill type conditions. However, since the negative soil heat flux equals the sum of all the atmospheric fluxes (as stated by the surface energy balance equation), any assumption that it is proportional to any particular component, or partial set of such components, seems dangerously nongeneral.

An entirely different approach to the soil heat flux has been proposed by *Shaffer and Long* [1975]. It is based upon the analytic solution to the diffusion equation [*Carlslaw and Jaeger,* 1959], which makes use of a time-weighted summation of all past values of G and theoretically requires that on each time step a fresh summation be constructed from all past values. In practice, a regrouping and truncation of terms within the summation is performed periodically. However, it is not clear if this method can be made as efficient as the use of, say, six soil layer temperatures with layer thickness increasing with depth while retaining high accuracy.

A method developed independently by *Arakawa* [1972] and by the British Meteorological Service [*Corby et al.,* 1972; *Rowntree,* 1975] utilizes a rate equation for the ground surface temperature T_g, dependent upon forcing by the sum of the atmospheric energy fluxes. It seems promising because it allows the temperature of a slab of soil to depend explicitly upon soil properties in the proper way; however, as was pointed out by *Bhumralkar* [1975], it omits the influence of soil heat flux on the underside of the slab.

Recently, *Bhumralkar* [1975] and *Blackadar* [1976] have independently proposed a method similar to the above T_g rate equation, which, however, contains the mechanism by which a deeper soil layer can influence the surface temperature. This method appears even more promising and is still much more efficient than the use of multiple soil layers.

With the multiplicity of methods presently in use it seems appropriate to point out their similarities and differences and to compare their accuracies for different types of soil.

The prediction of specific humidity of the air at the ground surface is in one respect usually treated slightly better than the prediction of T_g because an additional time-dependent equation for soil wetness is usually carried [*Manabe,* 1969; *Washington and Williamson,* 1977]. However, for purposes of short-range prediction this method also needs improvement. The evaporation rate at the ground, E_g, is assumed to be given by

$$E_g = \rho_a c_{H0} u_a \alpha[q_{\mathrm{sat}}(T_g) - q_a] \quad (1a)$$

$$E_g = \rho_a c_{H0} u_a (q_g - q_a) \quad (1b)$$

$$\alpha = \min(1, W_2/W_k) \quad (1c)$$

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where W_2 is the net soil moisture (depth of extracted liquid water) within a thick upper layer of soil, W_k is the critical depth this layer is capable of holding before the surface is considered to act as if it were saturated, q_{sat} is the saturation specific humidity at the ground surface temperature T_g and pressure p_s, q_a is the specific humidity of the air at height z_a within the surface layer, and q_g is the surface value of q. Also, ρ_a is the air density, and u_a is the wind speed at $z = z_a$, to which height the bare ground moisture or heat transfer coefficient c_{H0} applies.

In conjunction with (1) the time-dependent equation for W_2 is

$$\partial W_2/\partial t = -(E_g - P)/\rho_w \qquad 0 \le W_2 \le W_{max} \tag{2}$$

where P is the precipitation rate (mass of water exchanged per unit area and unit time), ρ_w is the density of water, and W_{max} is the maximum value of water depth, or field capacity, which exceeds W_k. (Runoff is considered to prevent W_2 from exceeding W_{max}.)

The definition of q_g implied by (1) is

$$q_g = \alpha q_{sat}(T_g) + (1 - \alpha)q_a \tag{3a}$$

but an obvious restriction that should be recognized [*Benoit*, 1976; *Rowntree*, 1975] is

$$q_g \le q_{sat}(T_g) \tag{3b}$$

The main shortcoming of (1) is that E_g does not respond to short-period occurrences of precipitation and evaporation which only gradually change W_2 according to (2) and therefore only gradually change q_g and E_g according to (1). For example, a rainfall of 1 cm in 3 hours would increase W_2 from 4 to 5 cm, say, relative to a saturated value of 10 cm, while in this period E_g as predicted from (1) would only increase 10% of the way toward its wet surface potential value. Instead, one would prefer a simple method which treats the actual surface and allows it to become saturated after only a short period of rainfall and to dry out substantially following evaporation. An accurate estimation of E_g is very important for the prediction of T_g.

Finally, there is little knowledge of how accurate the prediction of ground temperature T_g need be, when in actuality a complicated vegetative ground cover is usually present in the situation toward which the calculation of T_g is applied. It might be thought that the GFDL method, which assumes an insulated lower boundary, effectively treats the very case of a foliage-covered ground surface and actually applies to the estimation of mean foliage temperature. However, this interpretation neglects the facts that (1) significant amounts of heat and moisture may diffuse from the ground up through the vegetative canopy, (2) the foliage transpires at a rate not closely dependent upon W_2, and (3) on a large scale a significant fraction of the surface, either ground or rocks, is usually exposed to solar and atmospheric radiation and is not totally shielded by vegetation.

The purpose of this paper is to overcome the shortcomings described above. *Bhumralkar*'s [1975] and *Blackadar*'s [1976] approximate method of calculating T_g will be tested against the other methods which have been mentioned. An analogous approximate method for estimating soil surface moisture and q_g will be presented and tested. Finally, a highly simplified parameterization of the influence of a vegetative layer will be presented, along with a comparison of how its inclusion affects T_g and the net fluxes of heat and moisture to the atmosphere.

2. BRIEF DESCRIPTION OF T_g PREDICTION METHODS

The various methods to be compared are designated and described in Table 1. In this table, $H_A = -G$ is the sum of the fluxes in the atmosphere at the ground, ρ_s is the soil density, c_s is the soil specific heat, $d_1 = (\kappa_s \tau_1)^{1/2}$ is proportional to the depth reached by the diurnal temperature wave, κ_s is the soil thermal diffusivity, τ_1 is a period of 1 day, and T_2 is a deep soil temperature to be discussed later.

a. Multiple soil layer model. In the comparative tests, T_{gm} from a 12-layer soil model will be considered the 'true' value of T_g with which the value from the more approximate methods will be compared. The vertical coordinate transformation $\zeta = \ln(1 + z/\delta)$, $\delta = 1$ cm, is used, along with equal intervals in ζ of $\Delta\zeta = 0.385$. Soil grid points are located at depths of 0, 0.47, 1.11, 2.17, 3.66, 5.84, 9.05, 13.76, 20.69, 30.86, 45.80, 67.75, and 100.0 cm. Below the 100.0-cm depth the soil heat flux is assumed to be zero. The second-derivative soil diffusion terms are finite differenced by the Dufort-Frankel method by using a time step of $\Delta t = 15$ s. Only for the sake of accuracy is the T_{gm} time step made this small, and results were essentially the same as if second-order spatial differencing with a forward time step had been used. The soil flux, $G = -\lambda(\partial T/\partial z)_0$, is obtained from $-\lambda(T_1 - T_{gm})/z_1$, where $z_1 = 0.47$ cm. No consideration is given to any heating effects of water phase change within the soil below the surface or heat transport by water movement.

b. Solution of surface energy balance equation. This equation is

$$H_A = \epsilon_g \sigma T_g^4 + H_{sg} + L \cdot E_g - (1 - \alpha_g)S^{\downarrow} - \epsilon_g R_L^{\downarrow} = -G \tag{4}$$

TABLE 1. Methods of Calculating T_g

Method	Designation of T_g	Description	Reference
Multiple (12) soil layers	T_{gm}	Finite difference solution of diffusion equation for $T(z)$; $G = -\lambda(\partial T/\partial z)_0$	*Carlslaw and Jaeger* [1959]; *Benoit* [1976]
Insulated surface	T_{gi}	$G = 0$	*Gates et al.* [1971]; *Manabe et al.* [1974]
H_{sg} dependence	T_{gs}	$G = 1/3 H_{sg}$	*Kasahara and Washington* [1971]
R_{net} dependence	T_{gr}	$G = -0.19R_{net}$ $R_{net} < 0$ (down) $G = -0.32R_{net}$ $R_{net} > 0$ (up)	*Nickerson and Smiley* [1975]
H_A forcing	T_{gf}	$\partial T_g/\partial t = -\pi^{1/2}H_A/(\rho_s c_s d_1)$	*Arakawa* [1972]; *Corby et al.* [1972]; *Rowntree* [1975]
Force restore rate equation	T_{gfr}	$\partial T_g/\partial t = -2\pi^{1/2}H_A/(\rho_s c_s d_1) - (2\pi/\tau_1)(T_g - T_2)$	*Bhumralkar* [1975], *Blackadar* [1976]

where ϵ_g is the emissivity of the ground surface in the infrared, σ is the Stefan-Boltzman constant, H_{sg} is the sensible heat flux at the ground to the atmosphere, L is the latent heat of condensation, α_g is the ground albedo, $S\downarrow$ is the magnitude of the shortwave radiative flux, and R_L is the downcoming longwave radiative flux. The second, third, and fourth methods listed in Table 1 replace $-G$ on the right of (4) with the indicated assumption and then solve for T_g; the last two methods also make use of H_A.

Within H_A the three terms on the left depend upon T_g, and the two nonlinear terms, $\epsilon_g\sigma T_g^4$ and $L\cdot E$, are linearized for convenience. That is, with superscript (n) referring to the nth time step index, $[T_g^{(n+1)}]^4$ is approximated by

$$[T_g^{(n+1)}]^4 = [T_g^{(n)}]^4 + 4[T_g^{(n)}]^3[T_g^{(n+1)} - T_g^{(n)}]$$

and

$$q_{sat}(T_g^{(n+1)}) = q_{sat}(T_g^{(n)}) + (\partial q_{sat}/\partial T)_{T_g^{(n)}} \cdot [T_g^{(n+1)} - T_g^{(n)}]$$

where $\partial q_{sat}/\partial T$ is obtained from the Clausius-Clapeyron equation at $T = T_g^{(n)}$ and $q_{sat}(T_g^{(n)})$ is obtained from *Tetens'* [1930] equation. The linear solution of (4) for $T_g^{(n+1)}$, making use of the known value of $T_g^{(n)}$, is then straightforward.

In (4), H_{sg} is specified analogously to E_g in (1*b*):

$$H_{sg} = \rho_a c_p c_{H0} u_a (T_g - T_a) \quad (5)$$

where c_p is the specific heat at constant pressure and T_a is the air temperature at height z_a within the surface layer.

In the solar radiation term of (4), $S\downarrow$ is assigned the time-dependent value appropriate to 45° latitude on March 21, attenuated by 15% (or more when clouds are specified to be present).

In the last term of (4), $R_L\downarrow$ is prescribed by

$$R_L\downarrow = [\sigma_c + (1 - \sigma_c)0.67(1670q_a)^{0.08}]\sigma T_a^4 \quad (6)$$

where the factor in brackets is the parameterization for the effective emissivity of the air proposed by *Staley and Jurica* [1972] for clear skies ($\sigma_c = 0$) and σ_c is the cloud fraction.

c. H_A forcing method. The method of *Arakawa* [1972], *Corby* [1972], and *Rowntree* [1975] is here called the H_A forcing method because $T_g = T_{gf}$ is in that method driven by H_A:

$$\partial T_{gf}/\partial t = -(\pi)^{1/2} H_A/(\rho_s c_s d_1) \quad (7)$$

d. Force restore method. *Bhumralkar*'s [1975] and *Blackadar*'s [1976] approach is here called the force restore method because the forcing by $-H_A$ is modified by a restoring term which contains the deep soil temperature T_2:

$$\partial T_{gfr}/\partial t = -c_1 H_A/(\rho_s c_s d_1) - c_2(T_{gfr} - T_2)/\tau_1 \quad (8a)$$

with

$$c_1 = 2\pi^{1/2} \qquad c_2 = 2\pi \quad (8b)$$

For the short-range studies that Blackadar had in mind, T_2 can be treated as a constant and estimated as he suggested from the mean air temperature over the previous 24 hours. For longer-range purposes (longer than about 3 days), T_2 may be calculated from

$$\partial T_2/\partial t = -H_A/(\rho_s c_s d_2) \quad (9)$$

where

$$d_2 = (365\kappa_s\tau_1)^{1/2} \quad (10)$$

is $\pi^{1/2}$ times the e folding depth of the annual temperature wave. If one wishes d_2 to represent the latter depth, then the factor $\pi^{1/2}$ should multiply H_A in (9), in analogy to (7).

The last term in (8*a*) is seen to restore T_{gfr} exponentially toward the mean (or deep) soil temperature if H_A is removed. The constants c_1 and c_2 are chosen to yield the exact solution for a sinusoidally varying soil surface heat flux G after any transients have died away. For $G = A\cdot\sin(2\pi t/\tau_1 + \pi/4)$ this exact solution is [*Sellers*, 1965]:

$$T_g = T_2 + \{A\tau_1/[(2\pi)^{1/2}\rho_s c_s d_1]\} \sin(2\pi t/\tau_1) \quad (11)$$

where t is time and A is the amplitude of G.

The constants chosen by Blackadar for c_1 and c_2 in (8*a*) are slightly different: 3.72 and 7.4, respectively, to account for higher harmonics of the diurnal cycle.

The method of *Bhumralkar* [1975] is exactly the same as that of (8*a*) and (8*b*) except that he treated the temperature within a thin 1-cm slab of soil just below the surface. Instead, if his 1-cm slab thickness is denoted by δ, say, and carried through to his (15), then that equation reduces to (8) as $\delta \to 0$.

In using the T_{gfr} rate equation (8) the terms within H_A which contain T_{gfr} are also linearized and treated by the Crank-Nicolson method: i.e., T_{gfr} values are expressed as the average of values at the (n) and $(n + 1)$ time levels for $\partial T_g/\partial t$ being expressed by $(T_g^{(n+1)} - T_g^{(n)})/\Delta t$. The H_A forcing rate equation for T_{gf} in (7) is treated exactly the same as it is in (8), except that c_1 is taken as $\pi^{1/2}$ and c_2 as 0.

3. RESULTS OF TESTING (8) AND OTHER APPROXIMATE METHODS FOR T_g

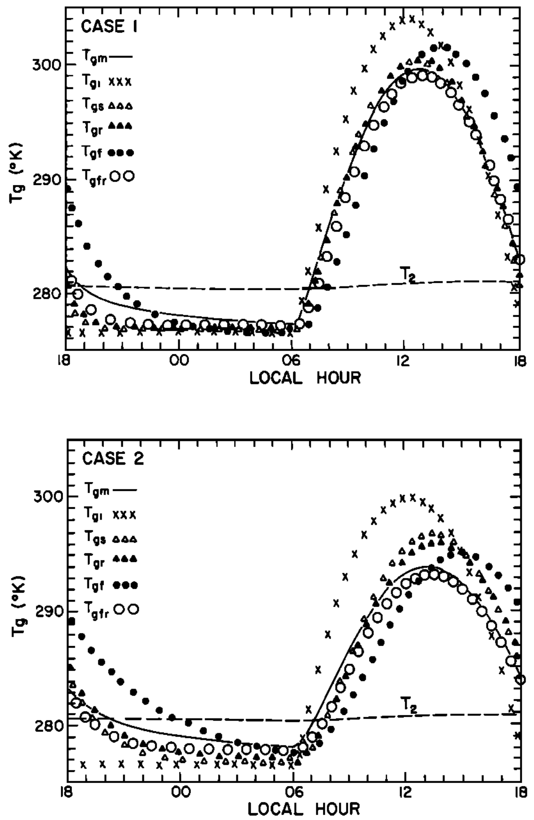
In the tests to be described in this section, (1), (2), and (3) were utilized to obtain q_g and E_g; a modified method for obtaining q_g is described in section 4. The value of c_{H0} utilized in (1) and (5) was 0.0025.

Five different sets of soil parameters and other relevant values were stipulated, as indicated in Table 2. They are chosen to be roughly consistent with data presented by *Sellers* [1965], *Lettau* [1951], and *Lettau and Davidson* [1957].

a. Constant atmospheric forcing except for variable solar radiation. In this set of tests the atmospheric temperature T_a and specific humidity q_a at $z = z_a$ were held constant at 280 K and 5×10^{-3}, respectively, except in case 5 of a simulated deep snow layer for which they were changed to 270 K and 3.0 ×

TABLE 2. Soil Parameters

	Case 1	Case 2	Case 3	Case 4	Case 5
Property					
κ_s, cm^2 s^{-1}	0.0040	0.0120	0.0020	0.0015	0.0027
$\rho_s c_s$, cal cm^{-3} K^{-1}	0.37	0.56	0.30	1.00	0.10
ϵ_g	0.90	0.95	0.80	0.94	0.90
α_g	0.25	0.15	0.40	0.10	0.65
$\bar{W}_2$, cm	2	8	0	(sat)	(sat)
W_k, cm	11	15	(12)	(sat)	(sat)
Description	O'Neill average	clay pasture	dry quartz sand	still muddy water	snow



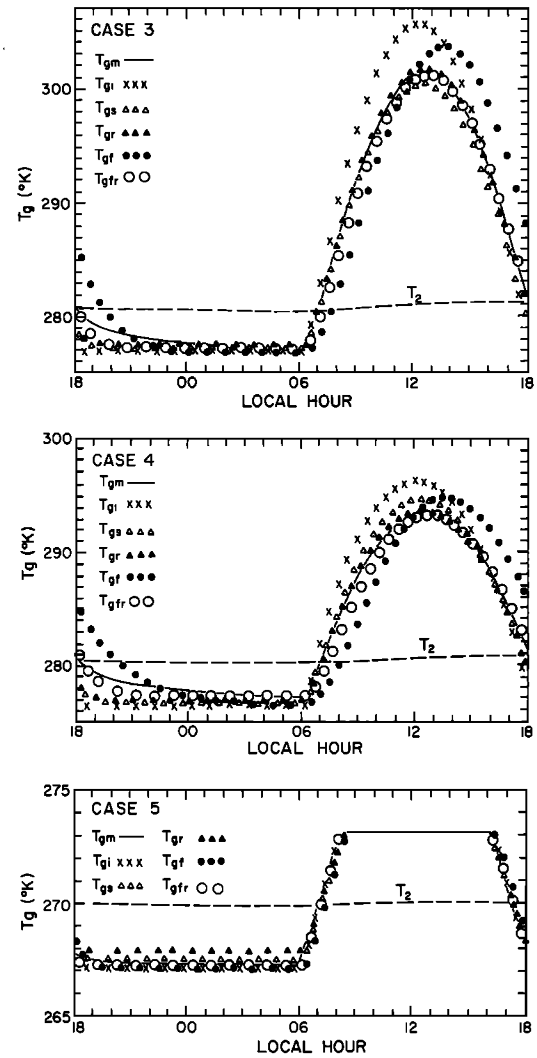


Fig. 1. Ground surface temperatures calculated for the second 24 hours of a 2-day period as a function of the local time of day. The different methods listed in the upper left corner of Figure 1*a* are explained in Table 1, and the key for the different symbols is also in the upper left corner. Figures 1*a*–1*e* refer to cases 1–5, respectively, of Table 2 for constant atmospheric forcing except for variable solar radiation. Values are only plotted about every hour rather than every 10 min.

10^{-3}. The wind speed u_a was held constant at 4 m s^{-1}. No clouds or precipitation were introduced. Initial soil temperatures, specified at about local sunset time, were assigned the value of T_a.

Predicted bare ground values of T_{gm}, T_{gi}, T_{gs}, T_{gr}, T_{gf}, and T_{gfr} are shown in Figures 1*a*–1*e* for a 1-day period with clear skies. A time step of 10 min was used (except for 15 s for T_{gm}). Emphasis is given to the 10-min time step because that value is typical of the value used within a global circulation model.

In the figures we notice that T_{gi} precedes the true solution in phase by about an hour, while T_{gf} lags it by 1–2 hours. In cases 1–4, T_{gi} overshoots the peak amplitude by $2\frac{1}{2}$–6 K and T_{gf} by 1.5–2.0 K. The approximate methods have the most difficulty with case 2, for which the medium has the greatest values of κ_s and λ. In this case, T_{gs} overshoots the peak value by 3 K, T_{gr} overshoots it by 2 K, and T_{gfr} undershoots it by 0.8 K. The approximate methods have difficulty in simulating the evening hours when G is relatively most important and, except for T_{gf}, predict T_g values 1–2 K too low.

In case 4 (still muddy water) the occurrence of evaporation at the full potential rate somewhat moderates the diurnal amplitude and causes the various solutions to be more similar to each other than they would be otherwise. In case 5 (snow layer) all the approximate methods appear more satisfactory because of the melting temperature constraint. However, because of the insulating property of snow, T_{gi} is in this one case (with Δt = 10 min) superior to the other approximate methods; T_{gr} is then the least accurate.

The relative error

$$e_r = [\langle(T_g - T_{gm})^2\rangle]^{1/2}/\Delta T_{gm}$$

during the second 24-hour period of the comparisons is presented in Table 3 for each of the approximate methods. The quantity T_g in e_r here stands for the calculated ground temperature by the approximate method tested, T_{gm} in e_r always refers to the T_{gm} calculation with the 15-s time step, the angle brackets are the average over the second 24-hour period, and ΔT_{gm} is the diurnal range in T_{gm}. Overall, for a 10-min time step the error in T_{gfr} is seen to be only about 0.62 as large as for T_{gs} or T_{gr} and only one third as large as for T_{gi}. The results confirm the assertion of *Bhumralkar* [1975] that T_{gf} does not realistically reproduce the diurnal variation of T_g.

A test was made on the fidelity of T_{gm} with its 15-s time step by comparing it with the exact solution (11) for G varying sinusoidally. The e_r value for T_{gm} relative to T_g (exact), after 9 days, averaged only 0.008 for the five different sets of soil properties, and extreme values of T_{gm} on the second day differed from those on the ninth day by no more than 0.68% of the diurnal range. Hence it is concluded that second-day values of T_{gm} were a viable standard against which to compare the more approximately calculated temperatures.

Tests were also conducted on the influence of the time step upon the approximate methods, and results are included in Table 3. As was pointed out by a reviewer, for constant atmospheric forcing but variable solar radiation the values of T_{gi}, T_{gs}, and T_{gr} would reach maxima at the time of maximum solar radiation (noon) were it not for the linearization procedure already discussed. This procedure causes the calculated times of maxima to lag more and more as Δt is increased. In addition, the soil heat flux assumptions involved in the T_{gs} and T_{gr} solutions also involve quantities evaluated on the previous time step, and this explains why the T_{gs} and T_{gr} maxima lag the maximum insolation more than the T_{gi} maximum does. Thus it turned out that the T_{gi} solution has least error for a time step between 10 and 30 min, while the T_{gs} and T_{gr} solutions have least error for $\Delta t \approx 10$ min. The T_{gfr} values are seen to retain superior accuracy even for large time steps.

Included in Table 3 are e_r values for $\Delta t = 10$ min and for equatorial latitude. In this test the initial soil temperature and T_a were increased to 287 K. The e_r values are little changed because of the normalization by the range of T_{gm}, which, as is indicated in the table, increased by up to 38%.

In other tests with a 10-min time step the assumption $G = -0.4R_{\text{net}}$ was tried, but this led to e_r values about twice as large in cases 1, 3, and 4. Also, the assumption $G = -0.10R_{\text{net}}$ (R_{net} incoming) and $-0.50\, R_{\text{net}}$ (R_{net} outgoing) utilized by *Gadd and Keers* [1970] was tested, but e_r values were larger in all cases by an average factor of 1.37.

The steady state form of (8) was also tested in the five cases, but ground temperatures then were substantially less accurate than the T_{gfr} values.

In one test the force restore method was tried with $c_1 = 3.72$ and $c_2 = 7.4$, as suggested by *Blackadar* [1976]. The e_r values for a 10-min time step then were slightly smaller for cases 4 and 5, the same for case 3, and slightly larger for cases 1 and 2 than with the use of $c_1 = 3.55$ and $c_2 = 6.28$. Results do not appear to be overly sensitive to the choice of c_1 and c_2.

Other conclusions for the case of diurnal forcing which may be drawn from Table 3 are (1) for time steps of ½ hour or greater, only the force restore method retains significant skill relative to the assumption $G = 0$; (2) the assumptions relating G either to H_{sg} or to R_{net} are nearly equally successful; and (3) the methods of H_{sg} or R_{net} dependence are about as successful as the force restore method for simulation of O'Neill type conditions when a 10-min time step is used but generally not as successful otherwise.

b. Random plus diurnal forcing. It is of interest to test (8) and the other approximate methods when the atmospheric forcing is partially random and contains fluctuations of period much shorter than the diurnal cycle. For this purpose the wind speed, air temperature, and specific humidity were allowed to vary as first-order Markovian processes (values continuous in time but time differences random) with standard deviations of 1.2 m s^{-1}, 1.0 K, and 0.5×10^{-3}, respectively, relative to the same mean values mentioned before. In addition, the solar radiation was attenuated by the factor $(1-0.6\sigma_c)$, where σ_c is a similarly varying cloud fraction with a mean value of about 0.3 and a standard deviation of about 0.1. The cloud fraction also influenced $R_L{}^{\downarrow}$ according to (6). Again, (1) and (3) were utilized to obtain q_g and the evaporation rate; no precipitation was allowed, and W_2 was held constant.

The random forcing was reapplied on each time step of the approximate method and therefore not on every time step of the 15-s T_{gm} calculations.

Results for cases 1 and 2 identified in Table 2 are shown in Figures 2*a* and 2*b*. For the most part the comparative behavior

TABLE 3. Relative 24-Hour Averaged Error e_r in Ground Temperature Calculated by the Five Approximate Methods for the Five Cases and Four Different Time Steps

Δt, min	T_g Designation	Case 1	Case 2	Case 3	Case 4	Case 5	Case Average Error
				45° Latitude			
5	T_{gi}	0.140(0.163)	0.276(0.294)	0.103(0.101)	0.138(0.161)	0.026(0.037)	0.137(0.151)
	T_{gs}	0.075(0.079)	0.159(0.137)	0.057(0.054)	0.092(0.086)	0.038(0.024)	0.084(0.076)
	T_{gr}	0.067(0.072)	0.128(0.112)	0.047(0.048)	0.077(0.070)	0.080(0.051)	0.080(0.071)
	T_{gf}	0.143(0.133)	0.199(0.188)	0.118(0.087)	0.144(0.111)	0.060(0.070)	0.133(0.118)
	T_{gfr}	0.041(0.044)	0.062(0.061)	0.033(0.036)	0.040(0.041)	0.022(0.037)	0.040(0.044)
10	T_{gi}	0.134(0.149)	0.270(0.340)	0.097(0.121)	0.133(0.141)	0.026(0.049)	0.132(0.160)
	T_{gs}	0.053(0.047)	0.121(0.138)	0.044(0.052)	0.076(0.074)	0.040(0.029)	0.077(0.068)
	T_{gr}	0.043(0.041)	0.097(0.117)	0.029(0.045)	0.059(0.063)	0.081(0.054)	0.062(0.064)
	T_{gf}	0.147(0.123)	0.203(0.194)	0.122(0.104)	0.149(0.121)	0.067(0.069)	0.138(0.122)
	T_{gfr}	0.043(0.039)	0.064(0.066)	0.035(0.039)	0.043(0.049)	0.029(0.030)	0.043(0.045)
30	T_{gi}	0.119(0.122)	0.251(0.347)	0.087(0.108)	0.119(0.129)	0.077(0.063)	0.131(0.178)
	T_{gs}	0.101(0.097)	0.213(0.264)	0.067(0.075)	0.083(0.079)	0.095(0.077)	0.112(0.118)
	T_{gr}	0.116(0.117)	0.226(0.264)	0.074(0.078)	0.076(0.073)	0.113(0.074)	0.121(0.121)
	T_{gf}	0.163(0.162)	0.218(0.282)	0.139(0.142)	0.166(0.144)	0.097(0.090)	0.157(0.164)
	T_{gfr}	0.056(0.059)	0.072(0.094)	0.050(0.056)	0.057(0.048)	0.063(0.059)	0.060(0.063)
60	T_{gi}	0.129(0.155)	0.236(0.386)	0.120(0.118)	0.125(0.163)	0.172(0.133)	0.156(0.191)
	T_{gs}	0.215(0.211)	0.288(0.250)	0.156(0.137)	0.178(0.196)	0.188(0.137)	0.205(0.186)
	T_{gr}	0.229(0.220)	0.287(0.256)	0.173(0.150)	0.174(0.190)	0.184(0.139)	0.209(0.191)
	T_{gf}	0.192(0.176)	0.243(0.198)	0.169(0.150)	0.195(0.207)	0.160(0.118)	0.192(0.170)
	T_{gfr}	0.082(0.084)	0.093(0.089)	0.078(0.073)	0.086(0.100)	0.133(0.110)	0.094(0.091)
	ΔT_{gm}, K	22.5(26.1)	15.9(14.0)	24.4(25.2)	16.3(16.1)	5.9(6.8)	
				0° Latitude			
10	T_{gi}	0.119(0.130)	0.231(0.295)	0.095(0.116)	0.114(0.125)		0.111(0.167)
	T_{gs}	0.049(0.042)	0.107(0.121)	0.047(0.052)	0.064(0.063)		0.067(0.070)
	T_{gr}	0.044(0.036)	0.092(0.106)	0.031(0.044)	0.068(0.073)		0.059(0.065)
	T_{gf}	0.134(0.113)	0.186(0.180)	0.120(0.103)	0.133(0.112)		0.143(0.127)
	T_{gfr}	0.041(0.036)	0.049(0.051)	0.037(0.040)	0.033(0.039)		0.040(0.042)
	ΔT_{gm}, K	28.4(33.9)	20.0(18.0)	33.6(34.2)	19.6(19.9)		

Values listed first are for constant air properties; those in parentheses are for varying air properties.

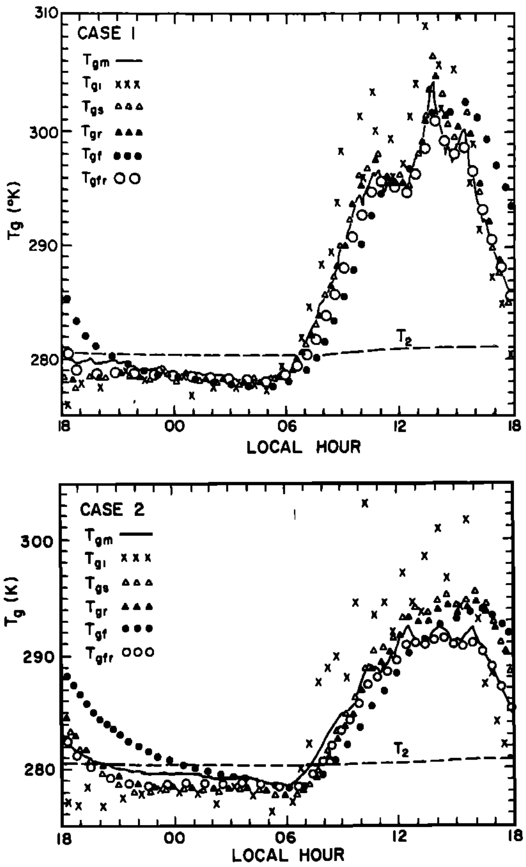


Fig. 2. Ground surface temperatures calculated over a 2-day period with the atmospheric forcing being partially random. Symbols are as in Figure 1*a* for the calculation methods listed in Table 1. Figure 2*a* is for case 1 of Table 2, and Figure 2*b* for case 2.

of the approximate methods is very similar to that portrayed in Figures 1*a*–1*e* so that cases 3–5 are not presented. T_{gfr} values have degraded only very slightly and, on the average, remain superior to the other T_g approximations tested for all values of the time step, as shown in Table 3 by the e_r values given in parentheses. A new feature which these e_r values disclose is that the relative error in T_{gi} in the presence of some random forcing is increased by 20–30% for time steps of 10 min or more, owing to the lack of any simulation of thermal inertia of the soil.

c. Random forcing only. One may question if, in the presence of random forcing but in the absence of diurnal forcing, T_{gfr} retains any skill over T_{gi}. The force restore method makes direct use of the assumption that a predominant forcing of period τ_1 exists. A random forcing comparison test was therefore conducted, exactly as it was in section 3*b* above, except that all solar forcing was removed. Much smaller ranges in T_{gm} then resulted, of only about 1–3 K in cases 1–4 and 5 K in case 5. The case average e_r values which were then obtained for a time step of 10 min are 0.28 for T_{gi}, 0.12 for T_{gs}, 0.14 for T_{gr}, 0.17 for T_{gf}, and 0.18 for T_{gfr}. Thus T_{gfr} and the other approximate methods do retain very significant skill in relation to the assumption $G = 0$. Although T_{gs} is found to be the superior method here, its advantage over T_{gfr} is not great, since the absolute errors were only of the order of 0.5 K.

4. FORCE RESTORE TREATMENT OF GROUND SOIL MOISTURE

a. Method. Let us denote the volume fraction of soil moisture by w (volume H_2O/volume soil, or depth of liquid H_2O/depth soil), and let w_k denote the value of w above which the surface is considered to act as if it were saturated. Upon assuming that most of the vertical movement of w within the soil can be described by a diffusion process, we may postulate the existence of an equation for $\partial w_g/\partial t$, where w_g is the ground surface value of w. The result is

$$\frac{\partial w_g}{\partial t} = -C_1 \frac{(E_g - P)}{\rho_w d_1'} - C_2 \frac{(w_g - w_2)}{\tau_1} \qquad (12)$$

$$0 \le w_g \le w_{\max}$$

where ρ_w is the density of liquid water, C_1 and C_2 are constants analogous to c_1 and c_2, d_1' is a depth to which the diurnal soil moisture cycle extends, w_2 is the vertically averaged value of w over a thicker layer d_2' below which the moisture flux is negligible, and $w_{\max}$ is the maximum value of w_g which exceeds w_k. Runoff of precipitation reaching the ground is considered to occur when w_g exceeds $w_{\max}$.

Although the coefficients C_1 and C_2 are somewhat uncertain and depend upon soil type, the data of *Jackson* [1973] for Adelanto loam suggest that

$$C_1 = 0.5 \qquad w_g/w_{\max} \ge 0.75 \qquad (13a)$$

$$C_1 = 14 - 22.5\,[(w_g/w_{\max}) - 0.15] \qquad (13b)$$

$$0.15 < w_g/w_{\max} < 0.75$$

$$C_1 = 14 \qquad w_g/w_{\max} \le 0.15 \qquad (13c)$$

and

$$C_2 = 0.9 \qquad (14)$$

with

$$d_1' = 10 \text{ cm} \qquad d_2' = 50 \text{ cm} \qquad (15)$$

The same data show how strong the influence of the diurnal cycle is upon the soil moisture fraction in the uppermost 5 mm of bare soil and provide justification for using the diurnal period in the denominator of the last term in (12). Earlier data of *Schiff and Dreibelbis* [1949] lack the very high resolution necessary for evaluation of the two constants but also suggest that C_1 is usually larger than C_2.

The time-dependent equation for w_2 is

$$\partial w_2/\partial t = -(E_g - P)/\rho_w d_2' \qquad 0 \le w_2 \le w_{\max} \qquad (16)$$

For comparison with (1)–(3) it may be noted that

$$w_2 d_2' = W_2 \qquad (17)$$

if we identify d_2' as the thickness of the layer containing the water depth W_2. Multiplication of (16) by d_2' then yields (2).

Having now a prediction equation for w_g, we may redefine (3*a*) to read

$$q_g = \alpha' q_{\text{sat}}(T_g) + (1 - \alpha')q_a \qquad q_g \le q_{\text{sat}}(T_g) \qquad (18a)$$

The form similar to (1*a*) which E_g takes upon using (18*a*) is

$$E_g = \rho_a c_{H0} u_a \alpha' [q_{\text{sat}}(T_g) - q_a] \qquad (18b)$$

where

$$\alpha' = \min(1, w_g/w_k) \qquad (18c)$$

For convenience, d_1' in (15) has been taken to be constant and its dependence upon w_g absorbed into C_1. Thus C_1 decreases with increasing w_g in (13*b*), which corresponds to stage 2 of a drying soil [*Idso et al.*, 1974] and to the soil moisture diffusivity's being an increasing function of soil moisture content. Equation (13*a*) corresponds to stage 1 (potential evaporation); (13*c*) corresponds to stage 3 (evaporation rate governed by vapor transfer and adsorption), except that (18*b*) is not really applicable [*Philip*, 1957]. It is nevertheless used even in stage 3, since the evaporation rate is then very small and its absolute error of estimate is also small.

Equation (18*a*) does not give the unrealistic behavior that (3*a*) does of q_g reaching its saturation value only when the mean soil moisture value W_2 reaches its saturated value. Instead, in (18*a*), local surface saturation can occur during moderate precipitation while w_2 is still far below its critical or saturated value. Following cessation of precipitation, w_g in (12) starts dropping below $w_{\max}$ owing to both evaporation and redistribution to the deeper soil and may even drop far below w_2 during a sunny afternoon because of evaporation.

Another advantage of using (12) is that the soil albedo may be formulated to be dependent upon w_g; the dependency is much more accurate and unique if w_g is used rather than w_2 [see *Idso et al.*, 1975*b*].

It should be recognized that (18), although an improvement over (3*a*), still suffers the inconsistency that the height of evaluation of q_a within the surface layer does not appear explicitly.

The behavior of (12)-(16) and (18) will be examined by utilizing this set of equations in place of (1)-(3). However, a demonstration of the behavior will be postponed until the latter part of the next section.

5. INCLUSION OF A LAYER OF VEGETATION

a. The assumptions and parameterizaton. A single layer of vegetation which has negligible heat capacity is assumed to be present. Its density will be characterized by the single quantity σ_f, which is an area average shielding factor associated with the degree to which the foliage prevents shortwave radiation from reaching the ground. The limits of σ_f are $0 \leq \sigma_f \leq 1$, $\sigma_f = 0$ signifying no foliage and $\sigma_f = 1$ signifying complete radiative blocking. Estimates of σ_f for some different plant types obtained from *Geiger* [1965] are 0.82 for meadow grass, 0.95 for clover 30 cm high, 0.83 for winter rye 80 cm high, 0.30 for summer barley 12-15 cm high, 0.98 for a 3-m-high thicket of young elms with dense undergrowth in summer, 0.50 for the same thicket after the leaves have fallen, and from 0.4 to 0.98 for various stands of either deciduous or evergreen trees.

The canopy will be treated as a bulk layer, and it will be necessary to distinguish between the heat or moisture transfer coefficient applicable to the ground surface underneath a canopy c_{Hg}, that applicable to bare ground c_{H0}, and that applicable to the top of a dense canopy c_{Hh}. The assumption will be made that for a dense canopy, $c_{Hg} = c_{Hh}$; the simplest interpolation which will be used here for c_{Hg} in general is

$$c_{Hg} = (1 - \sigma_f)c_{H0} + \sigma_f c_{Hh} \tag{19}$$

Within the canopy the mean wind which both ventilates the foliage and promotes weak heat and moisture fluxes from the ground surface will be denoted by u_{af} and prescribed by

$$u_{af} = 0.83\sigma_f c_{Hh}^{1/2}u_a + (1 - \sigma_f)u_a \tag{20}$$

In the absence of foliage this in-canopy wind is seen to revert to u_a and in the presence of dense foliage to become $0.83 \cdot c_{Hh}^{1/2}u_a$. The factor 0.83 comes from the studies reported by *Legg and Long* [1975], *Thom* [1972], *Webb* [1975], and *Geiger* [1965] with the considerations that within a relatively dense canopy, u_{af} is approximately 0.3 of the wind speed at foliage top height, the zero displacement height in the wind profile is about three fourths of the foliage top height, the roughness length is about one third of the difference between foliage top height and displacement height, and $c_{Hh}^{1/2}u_a$ is roughly the friction velocity.

The air in close proximity to the foliage is assumed to take on properties intermediate between above-canopy air properties at $z = z_a$, foliage surface properties, and ground surface properties:

$$T_{af} = (1 - \sigma_f)T_a + \sigma_f(0.3T_a + 0.6T_f + 0.1T_g) \tag{21a}$$

$$q_{af} = (1 - \sigma_f)q_a + \sigma_f(0.3q_a + 0.6q_f + 0.1q_g) \tag{21b}$$

where T_{af} refers to the mean temperature of the air within the foliage, T_f is a representative temperature at the foliage surface itself, and q_{af} and q_f are analogous specific humidities. Formula (21) prescribes T_{af} and q_{af} to be the same as T_a and q_a, respectively, in the absence of foliage; for maximum foliage cover they prescribe T_{af} and q_{af} to be influenced still by above-canopy air as well as by the foliage and ground. The sum of the three coefficients in the last terms of (21) must be unity, and the foliage influence is made several times stronger than the ground influence because of the greater foliage surface area within a relatively dense canopy. There is little other information for guidance except from sensitivity studies which show little change in results if the present coefficients in the last terms of (21) are changed to (0.45, 0.45, 0.1).

The sensible heat transfer from a representative leaf will be assumed to be given by

$$H_{\text{leaf}} = \rho_a c_p c_f u_{af}(T_f - T_{af}) \tag{22a}$$

where c_f is a dimensionless heat transfer coefficient which takes into account both sides of the leaf and T_f is the representative leaf temperature. If the net leaf area index N over some region is defined as the total one-sided leaf area of the foliage relative to the ground area of the same region, the expression for the net sensible heat flux from the foliage to the surrounding air, H_{sf}, per unit horizontal ground area, is approximately

$$H_{sf} = 1.1NH_{\text{leaf}} = 1.1N\rho_a c_p c_f u_{af}(T_f - T_{af}) \tag{22b}$$

The factor 1.1 roughly accounts for the effects of those stalks, stems, twigs, and limbs which exchange heat but do not transpire.

The dimensionless heat transfer coefficient is assumed to be given by

$$c_f = 0.01[1 + 0.3\ (\text{m s}^{-1})/u_{af}] \tag{23}$$

The value 0.01 is derived from the study of *Kumar and Barthakur* [1971] for forced convection over several different types of plants. The other factor in (23) is a free convection enhancement; the general importance of free convection within the canopy has been emphasized by *Allen and Lemon* [1972].

There is obviously a rough relation between N and σ_f, here assumed to be given by

$$N = 7\sigma_f \tag{24}$$

which is consistent with results reported in *Allen and Lemon* [1972] for a corn crop and by *Monteith et al.* [1965] for barley. N does not much exceed a number like 7 because there is then insufficient light to support additional growth within the canopy. Although isolated trees may appear to have larger values of N, they usually do not when surrounding open space is taken into account.

Generalizing from *Monteith and Szeicz* [1962], we find that the evaporation rate per unit area from a representative leaf, E_{leaf}, is

$$E_{\text{leaf}} = \rho_a c_f u_{af}[q_{\text{sat}}(T_f) - q_{af}]r'' \tag{25a}$$

where

$$r'' = 1 - \delta_c[r_s/(r_s + r_a)]\,[1 - (W_{\text{dew}}/W_{\text{dmax}})^{2/3}] \tag{25b}$$

represents the fraction of potential evaporation, δ_c is a step function which is zero if condensation is occurring onto the leaf (if $q_{af} > q_{sat}(T_f)$) and is unity otherwise, r_s is a generalized stomatal resistance and r_a is the 'atmospheric resistance,' W_{dew} is the mass of any liquid water retained on the foliage per unit ground area, and W_{dmax} is the maximum value of W_{dew} beyond which runoff to the ground occurs.

The net foliage evaporation rate per unit horizontal ground area, E_f, is then

$$E_f = NE_{leaf} = N\rho_a c_f u_{af}[q_{sat}(T_f) - q_{af}]r'' = r''(E_f)_{pot} \qquad (25c)$$

where $(E_f)_{pot}$ is the potential (maximum possible) foliage evaporation rate. By definition of the leaf area index N the evaporative resistance of the canopy is only $1/N$ as large as that of the representative leaf.

With the inclusion of r'' in (25c), E_f may be seen to be the transpiration rate when $\delta_c = 1$ and no dew or retained water is present, to be the rate of condensation onto the foliage when $\delta_c = 0$, and to be the rate of transpiration plus evaporation of dew when $\delta_c = 1$ and $W_{dew} > 0$. When evaporating dew is present, the fraction of the foliage surfaces coated with moisture is assumed to be given by $(W_{dew}/W_{dmax})^{2/3}$; the reason for the fractional exponent is explained later. For $\delta_c = 1$ and $W_{dew} = 0$ (the usual daytime case), (25c) and (25b) together represent the transpiration rate as formulated by Monteith and Szeicz.

The transpiration rate itself, per unit ground area, is seen to be given by

$$E_{tr} = \delta_c(E_f)_{pot}[r_a/(r_s + r_a)]\ [1 - (W_{dew}/W_{dmax})^{2/3}] \qquad (26)$$

since the quantity in the second set of brackets represents the fractional foliage surface not covered by dew.

The facts that many types of leaves transpire from only the underside and that older leaves transpire less than newer ones are supposed to be accounted for by a representative choice of generalized stomatal resistance r_s:

$$r_s = 2.0\ (\text{s cm}^{-1})[S_{max}{\downarrow}/(S{\downarrow} + 0.03S_{max}{\downarrow}) + \mathcal{S} + (w_{wilt}/w_s)^2] \qquad (27)$$

$S_{max}{\downarrow}$ is the maximum noon incoming solar radiation which can be achieved, $\mathcal{S}$ is a seasonal dependence, w_{wilt} is a wilting point value of soil moisture relative to its saturated value, and $w_s = (0.9w_2 + 0.1w_g)$ is a soil moisture value in the root zone assumed to lie closer to the bulk value w_2 than to the surface value w_g. The specified dependence of r_s upon solar radiation is suggested from studies of *Cline and Campbell* [1976] for forests, *Waggoner and Reifsnyder* [1968] for certain crops, and *Monteith et al.* [1965] for barley. The minimum value of 2 s cm^{-1} is near the lower side of the range of values suggested by these investigators, by *Fetcher* [1976] for lodgepole pine, and by *Monteith and Szeicz* [1962] for various types of foliage when the difference between leaf resistance and canopy resistance is considered. The last term within the brackets of (27) is designed to give a strong enhancement to the transpirative resistance if w_s drops close to or below w_{wilt} in magnitude. Besides the seasonal dependence, the two dependencies, daylight and soil moisture, appear to be the most important [*Cline and Campbell,* 1976] of many factors involved in the transpirative resistance. In temperate latitudes, $\mathcal{S}$ is set to zero during the growing season and to a value much larger than unity during the rest of the year.

With this approach of Monteith and Szeicz the atmospheric resistance is simply

$$r_a = (c_f u_{af})^{-1} \qquad (28)$$

since the transfer coefficient c_f is defined with respect to the mean flow inside the foliage layer rather than to the mean flow above the foliage.

The conservation equation for W_{dew} is taken to be

$$\partial W_{dew}/\partial t = \sigma_f P - (E_f - E_{tr}) \qquad 0 \le W_{dew} \le W_{dmax} \qquad (29)$$

since the difference, $E_f - E_{tr}$, just represents evaporation or condensation of liquid water from or onto the foliage surfaces. When dew is evaporating, $E_f - E_{tr}$ becomes $(W_{dew}/W_{dmax})^{2/3}(E_f)_{pot}$. The reason for specifying a fractional power dependence upon W_{dew}/W_{dmax} is to allow dew to evaporate more rapidly than it does at an exponentially decreasing rate; for unit exponent the dew would never quite disappear, while for zero exponent it would evaporate much too fast with the implication that the dew is present in a continuous thin film over the entire leaf as evaporation proceeds. Formulations (25b) and (29) are a compromise between those two extreme positions and simulate the dew's occupying only a fraction of the leaf area during its evaporation and the entire area during its formation.

Condensation which may occur on the soil surface is not treated as dew but is simply added to the bulk soil moisture budget.

A gross energy budget for the foliage layer must be established in order to estimate T_f. The values at the top of the canopy being denoted by subscript h, those at the ground by subscript g, and the direction of radiative fluxes by arrows, the assumption of no canopy heat storage leads to

$$S_h{\downarrow} + R_{Lh}{\downarrow} - S_h{\uparrow} - R_{Lh}{\uparrow} - (S_g{\downarrow} + R_{Lg}{\downarrow} - S_g{\uparrow} - R_{Lg}{\uparrow}) = H_{sh} - H_{sg} + L(E_h - E_g) \qquad (30)$$

where S is the shortwave and R_L the longwave flux, H_s the sensible heat, and E_h the evapotranspiration as seen at level $z = h$. On the left side of (30), $S_h{\downarrow}$ and $R_{Lh}{\downarrow}$ are assumed known, while by definition of σ_f, $S_g{\downarrow}$ is given by

$$S_g{\downarrow} = (1 - \sigma_f)S_h{\downarrow} \qquad (31a)$$

By definition of α_g the reflected flux $S_g{\uparrow}$ is given by

$$S_g{\uparrow} = \alpha_g(1 - \sigma_f)S_h{\downarrow} \qquad (31b)$$

The upward longwave flux just above the ground, $R_{Lg}{\uparrow}$, is obtained by interpolating with σ_f between the expression applicable above bare soil and that applicable just above soil overlain with a dense canopy:

$$R_{Lg}{\uparrow} = (1 - \sigma_f)[\epsilon_g\sigma T_g^4 + (1 - \epsilon_g)R_{Lh}{\downarrow}] + \sigma_f[\epsilon_g\sigma T_g^4 + (1 - \epsilon_g)\epsilon_f\sigma T_f^4]/(\epsilon_f + \epsilon_g - \epsilon_f\epsilon_g) \qquad (31c)$$

For $\sigma_f = 1$ this expression for $R_{Lg}{\uparrow}$ reverts to that for the radiative flux between two parallel surfaces [*Fleagle and Businger,* 1963] of emissivities ϵ_g and ϵ_f. For $\sigma_f = 0$ the expression similarly accounts for the upward reflection of $R_{Lh}{\downarrow}$ from the ground when $\epsilon_g < 1$. The three remaining radiative fluxes are similarly obtained:

$$S_h{\uparrow} = (1 - \sigma_f)\alpha_g S_h{\downarrow} + \sigma_f\alpha_f S_h{\downarrow} \qquad (31d)$$

$$R_{Lh}{\uparrow} = (1 - \sigma_f)[\epsilon_g\sigma T_g^4 + (1 - \epsilon_g)R_{Lh}{\downarrow}] + \sigma_f[\epsilon_f\sigma T_f^4 + (1 - \epsilon_f)R_{Lh}{\downarrow}] \qquad (31e)$$

$$R_{Lg}{\downarrow} = (1 - \sigma_f)R_{Lh}{\downarrow} + \sigma_f[\epsilon_f\sigma T_f^4 + (1 - \epsilon_f)\epsilon_g\sigma T_g^4]/(\epsilon_f + \epsilon_g - \epsilon_f\epsilon_g) \qquad (31f)$$

With these substitutions of (31) into (30) and with the definitions

$$H_{sf} \equiv H_{sh} - H_{sg}$$

and

$$E_f \equiv E_h - E_g$$

(30) becomes

$$\sigma_f\left[(1-\alpha_f)S_h{}^{\downarrow} + \epsilon_f R_{Lh}{}^{\downarrow} + \frac{\epsilon_f\epsilon_g}{(\epsilon_f+\epsilon_g-\epsilon_f\epsilon_g)}\sigma T_g{}^4 - \frac{(\epsilon_f+2\epsilon_g-\epsilon_f\epsilon_g)}{(\epsilon_f+\epsilon_g-\epsilon_f\epsilon_g)}\epsilon_f\sigma T_f{}^4\right] = H_{sf} + LE_f \quad (32)$$

which is to be solved for T_f. It may be noted that σ_f in (32) can be cancelled from all terms (since the right-hand terms contain N, which contains σ_f), allowing T_f to be obtained even in the limit of just no foliage. It may also be noticed from (31d) that α_f in (32) is the foliage albedo as would be measured with a radiometer above a dense canopy, so that variable leaf angle relative to the sun and to the zenith is incorporated within its definition. For ϵ_f and ϵ_g close to unity the coefficient of the $\epsilon_f\sigma T_f{}^4$ term in (32) is about 2, which accords with the expectation that the foliage layer emits longwave radiation both upward and downward.

Within a sunlit canopy there will be a wide range of individual leaf temperatures differing by as much as 12 K [*Miller*, 1971], and the intent is for the leaf energy budget equation (32) to yield a representative value for T_f, which taken along with the values for T_{af}, q_{af}, u_{af}, c_f, N, and r_s will yield the correct values for sensible heat flux and transpiration rate from foliage to air.

The foliage surface specific humidity q_f needed in (21b) is obtained from

$$q_f = r'' q_{\text{sat}}(T_f) + (1 - r'')q_{af} \quad (33a)$$

with the usual restriction

$$q_f \le q_{\text{sat}}(T_f) \quad (33b)$$

Expression (33a) is derived by equating E_f from (25c) with the alternate expression involving q_f:

$$E_f = N\rho_a c_f u_{af}(q_f - q_{af}) \quad (34)$$

Because of the ground cover the soil energy balance of (4) must be modified in order to obtain T_g needed in (32). With H_{sg} denoting the sensible heat flux at the ground surface and E_g the evaporation rate there, we make the simplest possible generalization of (5) and (1b):

$$H_{sg} = \rho_a c_p c_{Hg} u_{af}(T_g - T_{af}) \quad (35a)$$

$$E_g = \rho_a c_{Hg} u_{af}(q_g - q_{af}) \quad (35b)$$

Equation (35) makes use of the property that $(c_{Hg}, u_{af}, T_{af}, q_{af}) \to (c_{H0}, u_a, T_a, q_a)$ as $\sigma_f \to 0$; (35a) and (5) further neglect the small distinction between temperature and potential temperature differences within the foliage and atmospheric surface layers.

The ground surface energy balance then becomes

$$-G = H_{sg} + LE_g - (1-\alpha_g)S_g{}^{\downarrow} + R_{Lg}{}^{\uparrow} - R_{Lg}{}^{\downarrow} \quad (36)$$

with $S_g{}^{\downarrow}$ given by (31a), $R_{Lg}{}^{\uparrow}$ by (31c), and $R_{Lg}{}^{\downarrow}$ by (31f). When the force restore method is used to get T_g, the soil surface heat flux $-G$ in (36) replaces H_A in (8) and in (9).

The soil properties used in (8) and (9) appear to depend more upon soil moisture than soil type, and a parameterized dependence for λ and $\rho_s c_s$ upon w_g and w_2 is therefore attempted, following *Benoit* [1976] and *Sellers* [1965]. At the ground surface and in the bulk mean, respectively, we parameterize $\rho_s c_s$ by

$$(\rho c)_g = 0.27 + w_g \quad \text{cal cm}^{-3}\text{ K}^{-1}$$
$$(\rho c)_2 = 0.27 + w_2 \quad \text{cal cm}^{-3}\text{ K}^{-1} \quad (37)$$

and parameterize $\lambda = \rho_s c_s \kappa_s$ and $d_1 = (\kappa_s \tau_1)^{1/2}$ by

$$\lambda_g = 0.001 + 0.004(w_g)^{1/3} \quad \text{cal cm}^{-1}\text{ s}^{-1}\text{ K}^{-1}$$
$$\lambda_2 = 0.001 + 0.004(w_2)^{1/3} \quad \text{cal cm}^{-1}\text{ s}^{-1}\text{ K}^{-1} \quad (38a)$$

$$d_{1g} = [\tau_1\lambda_g/(\rho c)_g]^{1/2}$$
$$d_{12} = [\tau_1\lambda_2/(\rho c)_2]^{1/2} \quad (38b)$$

Now, in (8) it is not clear how best to define an optimal value of $\rho_s c_s d_1$ when the ground surface soil properties differ significantly from the bulk soil properties. However, tests were made using the 12-layer soil model with w_g/w_2 being either a factor of 5 greater than or smaller than unity and with a logarithmic distribution of w with depth (linear distribution in the coordinate ζ). Improved results were obtained by defining

$$r' = 0.30 + 0.05 w_g/w_2 \qquad 0.3 < r' \le 1 \quad (39a)$$

and utilizing

$$\rho_s c_s d_1 = r'(\rho c)_g d_{1g} + (1 - r')(\rho c)_2 d_{12} \quad (39b)$$

Without this subparameterization and with only the use of the w_2 moisture value, the root-mean-square error in T_g averaged over a day was 2.6 K relative to a diurnal range of 22.7 K when the surface was moist and 1.3 K relative to a diurnal range of 21.4 K when the surface was dry. With the parameterization these errors were reduced to 0.6 and 0.9 K, respectively. Much more work is needed on this subject, however.

The ground albedo α_g is also made a function of ground surface moisture as follows:

$$\alpha_g = 0.31 - 0.17 w_g/w_k \qquad w_g \le w_k$$
$$\alpha_g = 0.14 \qquad w_g > w_k \quad (40)$$

as suggested by the study of *Idso et al.* [1975b].

The quantity q_g needed in (35b) and in (21b) is obtained not from (18a) but from the slightly generalized equation

$$q_g = \alpha' q_{\text{sat}}(T_g) + (1 - \alpha')q_{af} \quad (41a)$$

again provided that

$$q_g \le q_{\text{sat}}(T_g) \quad (41b)$$

The soil moisture budget equations are modified slightly from those in sections 1 and 4 to allow for the effect of transpiration:

$$\partial w_g/\partial t = -C_1(E_g + 0.1E_{tr} - P_g)/(\rho_w d_1') - C_2(w_g - w_2)/\tau_1 \quad (42)$$

$$\partial w_2/\partial t = -(E_g + E_{tr} - P_g)/(\rho_w d_2') \quad (43)$$

where P_g is the precipitation rate felt at the ground surface:

$$P_g = P(1 - \sigma_f) \qquad W_{\text{dew}} < W_{\text{dmax}}$$
$$P_g = P \qquad W_{\text{dew}} \ge W_{\text{dmax}} \quad (44)$$

and C_1 and C_2 are given by (13) and (14). Since most of the moisture supply for transpiration is considered to lie beneath the uppermost layer of relative moisture content w_g, only 0.1 of E_{tr} is allowed to influence w_g in (42). In (43), moisture transpiring from the foliage is considered to be piped directly from the bulk layer of mean content w_2, while that condensing onto the foliage at night is assumed to remain in situ (unless $W_{\text{dew}} > W_{\text{dmax}}$) until eventual evaporation.

The main purpose for the development of the equations in this section is to obtain modified expressions for the vertical fluxes from the ground foliage system to the atmosphere. These are the sensible heat flux $H_{sh} = H_{sg} + H_{sf}$, given from (35*a*) and (22*b*) as

$$H_{sh} = \rho_a c_p c_{Hg} u_{af}(T_g - T_{af}) + 1.1 N \rho_a c_p c_f u_{af}(T_f - T_{af}) \quad (45)$$

the evapotranspiration rate $E_h = E_g + E_f$, given from (35*b*) and (25*c*) as

$$E_h = \rho_a c_{Hg} u_{af}(q_g - q_{af}) + N \rho_a c_f u_{af}[q_{\text{sat}}(T_f) - q_{af}]r'' \quad (46)$$

and the upward directed shortwave radiative flux $S_h \uparrow$ from (31*d*) and upward directed longwave flux $R_{Lh} \uparrow$ from (31*e*).

An outline of the method of solution which will be used for obtaining these vertical fluxes is as follows:

1. Specify σ_f, along with soil and foliage roughness, α_f, initial conditions, above-canopy air properties, and incoming radiative fluxes or precipitation; calculate N from (24) and c_{Hg} from (19).
2. Calculate α_g from (40), u_{af} from (20), c_f from (23), r_a from (28), r_s from (27), and r'' from (25*b*).
3. Using $q_{\text{sat}}(T_f)$ and q_{af} from the previous time step, diagnose q_f from (33). Here a more economical method of calculating q_{sat} may be employed than that in section 2, where testing of a large time step precluded stepping along the Clausius-Clapeyron equation.
4. Diagnose q_{af} from (21*b*) and q_g from (41), the two being solved simultaneously, and T_{af} from (21*a*). Previous values of T_f, T_g, q_f, and q_g are utilized.
5. Diagnose T_f from the foliage energy budget equation (32) by using a time step linearization of the outgoing longwave radiation term and foliage saturation humidity term as described in section 2.
6. Obtain $\rho_s c_s d_1$ from (37)–(39) and the components of G from (31*a*), (31*f*), and (35), and then obtain G from (36).
7. Predict updated values for T_g by the force restore method (8), and T_2 from (9), by using G from step 6.

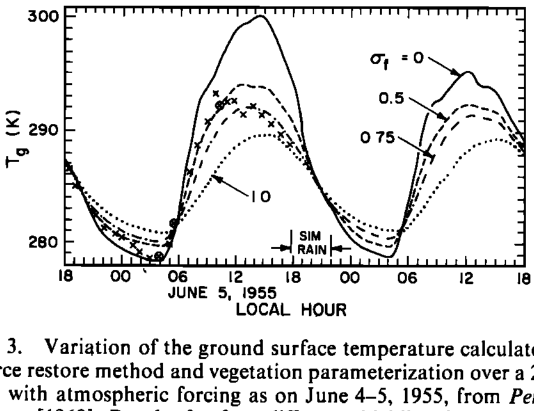


Fig. 3. Variation of the ground surface temperature calculated by the force restore method and vegetation parameterization over a 2-day period with atmospheric forcing as on June 4–5, 1955, from *Penman and Long* [1960]. Results for four different shielding factors, σ_f, are shown, along with extrapolated observed values denoted by crosses.

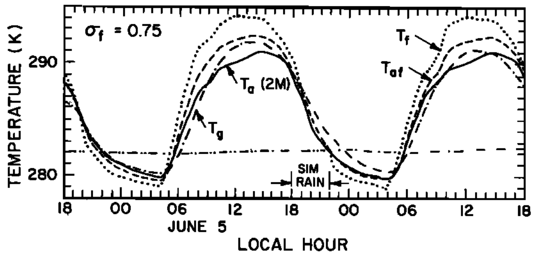


Fig. 4. Variation of the predicted temperatures T_f, T_{af}, T_g, and of T_a over the 2-day period for a shielding factor of 0.75.

8. Obtain E_{tr} from (26) and P_g from (44), and then predict w_g from (42), w_2 from (43), and W_{dew} from (29).
9. Obtain the vertical fluxes to the atmosphere from (45), (46), (31*d*), and (31*e*).
10. Repeat steps 2–9 on subsequent time steps.

b. Test of the vegetation layer parameterization. A partial test is made by simulating the conditions reported by *Penman and Long* [1960] within a dense wheat crop in England on June 4–5, 1955. It is incomplete because measurements of σ_f, soil moisture, soil temperatures, mean foliage radiometric temperature, evapotranspiration, heat flux, and radiation balance during this period were not made or reported and because sampling errors of temperature and humidity within the crop can be appreciable.

There was apparently little or no cloudiness during this period, and solar radiation appropriate for latitude 51.8°N, attenuated by 15% to allow for absorption by ozone, water vapor, and dust, is prescribed. Parameters w_g and w_2 are assumed to be 0.20 and 0.25 initially, w_k to be 0.30, $w_{\max}$ to be $1.33 w_k$, and w_{wilt} to be 0.10. Soil properties are taken from (37) and (38) and, for $w_g \approx 0.25$, agree well with those reported by the authors for their site in 1956. Initially (hour 18), T_g was taken as 286.6 K and T_2 as 282 K. W_{dmax} was set to $0.1\sigma_f$ g cm^{-2}.

The wheat crop was extra thick and about 31 cm high. Although Penman and Long mention that it was typical of a canopy which completely shades the ground, it is assumed here, on the basis of earlier discussion of σ_f, that σ_f was 0.75. The values of c_{H0} and c_{Hh}, referred to a height of 2 m, are taken to be 0.0057 and 0.0096; the corresponding roughness lengths are 1 and 3 cm, respectively, with a zero displacement height within foliage of 23 cm. Emissivities are assigned the values $\epsilon_f = \epsilon_g = 0.95$ and for albedo, $\alpha_f = 0.20$ and α_g as taken from (40).

Air temperatures, specific humidity, and wind speed at the z_a (2 m) height for the 24-hour period beginning at 1800, June 4, are specified from Penman and Long's Figure 2. During this period, u_a varied between 0.25 m s^{-1} at night and 4.8 m s^{-1} in the afternoon. To account for winds too weak to measure, u_a is

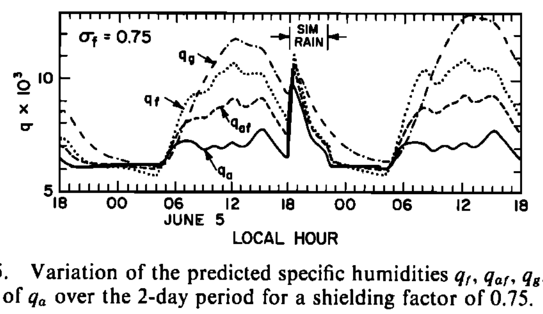


Fig. 5. Variation of the predicted specific humidities q_f, q_{af}, q_g, and of q_a over the 2-day period for a shielding factor of 0.75.

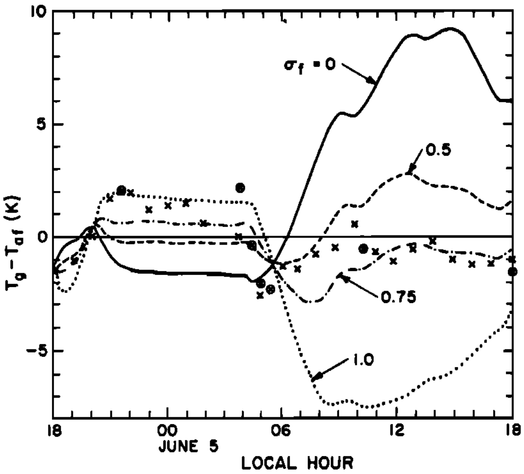


Fig. 6. Variation of $T_g - T_{af}$ during the 24-hour period from June 4 to 5, 1955, as calculated from the model for three values of σ_f, curves, and observational estimates from *Penman and Long* [1960], crosses. Circled crosses are obtained from observations with greater vertical resolution within the canopy.

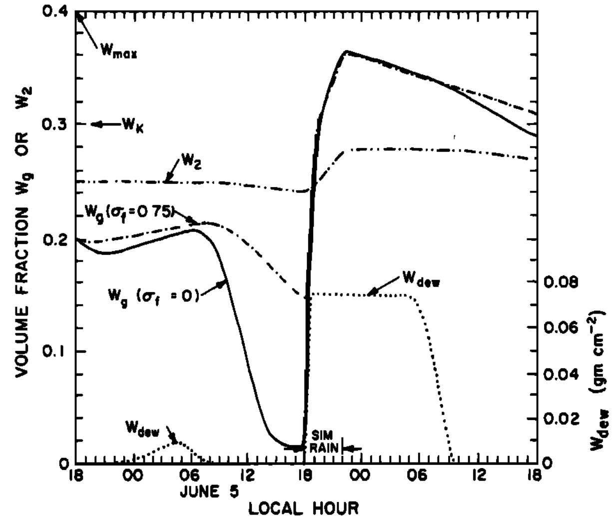


Fig. 8. Predicted variation of surface (w_g) and bulk (W_2) soil moisture, and of W_{dew} (right-hand scale) over the 2-day period which includes a 4-hour period of simulated rain, for $\sigma_f = 0.75$.

constrained not to fall below 0.3 m s^{-1} and u_{af} not below 0.15 m s^{-1}. A second 24-hour period is also treated with conditions identical to those of the first, except for simulation of 2 cm of rainfall between hours 18 and 22, accompanied then by overcast skies and a relative humidity of 95%. The numerical integration utilized a 10-min time step.

The diurnal variation of T_g calculated for the 2-day period for $\sigma_f = 0.75$ is shown in Figure 3; included are calculated T_g values obtained for $\sigma_f = 0, 0.5$, and 1.0 for comparison, all else being held the same, and observational estimates from *Penman and Long* [1960]. The observed values (crosses) are in-canopy downward extrapolations from nickel resistance thermometer measurements at 5.0 and 30 cm except in three instances (circled crosses) which make use also of measurements at 2.5 and 7.5 cm. Although neither the model nor the measurements are so good that the crosses fall along any particular σ_f curve, the value of 0.75 selected for σ_f yields satisfactory T_g values overall. The observed maximum at hour 10 seems peculiar but could be associated with the microdensity of the wheat canopy in the immediate vicinity of the instruments in relation to the solar azimuth angle. The parameterization predicts a 2.5-fold reduction in the diurnal range of T_g from $\sigma_f = 0$ to $\sigma_f = 1$ and a corresponding lag in the time of the occurrence of peak

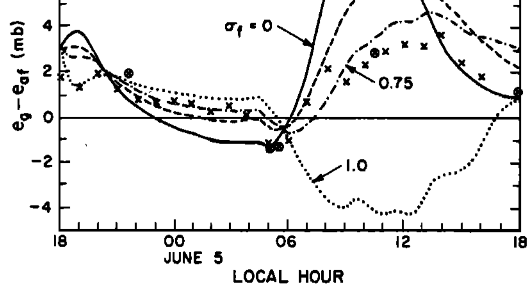


Fig. 7. Variation of the vapor pressure difference $e_g - e_{af}$ during the 24-hour period from June 4 to 5, 1955, as calculated from the model for $\sigma_f = 0.75$, dash-dot curve, and observational estimates from *Penman and Long* [1960], crosses.

ground temperature of about 1 hour. The maximum bare ground temperature is 5 K less on the second day than on the first because of the effect of simulated rainfall during the intervening evening.

The diurnal variations calculated for T_f, T_{af}, T_g, and the prescribed variation of T_a are shown in Figure 4, and those for q_f, q_{af}, q_g, and q_a are shown in Figure 5, all for $\sigma_f = 0.75$.

The representative foliage temperature is seen to exceed substantially the maximum value which T_g reaches during midday but to be up to 5° cooler than T_g for bare soil. T_f exceeds the surrounding air temperature T_{af} by up to 2 K. This result is consistent with data compiled by *Geiger* [1965, Figures 136 and 137] and observations [*Waterhouse*, 1955] which show the air near the top of a crop during the day to be warmer than either the air above or the ground below. The air must therefore have been heated by foliage considerably warmer than itself. However, this conclusion has been disputed by *Lomas et al.* [1971], and the difficulty in obtaining representative measurements is probably responsible for the uncertainty. If the model is rerun with the stomatal resistance increased by a factor of 100, but with all else the same, the maximum foliage temperature reaches a value 5.5 K warmer (12.0 K warmer for $\sigma_f = 1$), which is an example of 'foliage fever.'

A downward jump in T_f is discernible in Figure 4 just following the simulated rainfall; this is associated with the decrease in downcoming longwave radiation after the skies clear.

Figure 5 shows that q_f becomes less than q_g and q_{af} at night, thus promoting condensation onto the foliage, while q_f exceeds q_g during the early morning hours. The peak between hours 18 and 21 of June 5 is caused by the simulated rainfall which wets both the foliage and ground. The peak in q_f between 0700 and 0930 on June 6 is associated with warming of the extra-wet foliage before the retained water evaporates. Other wiggles are associated with variations in the observed values of q_a, T_a, and u_a utilized.

A sensitive measure of the performance of the parameterization is the temperature difference $T_g - T_{af}$, since T_g and T_{af} each comprises rather independent elements of the model. This difference is shown in Figure 6 for $\sigma_f = 0, 0.5$, and 1.0 as well as for 0.75. Estimated values from the observations (crosses

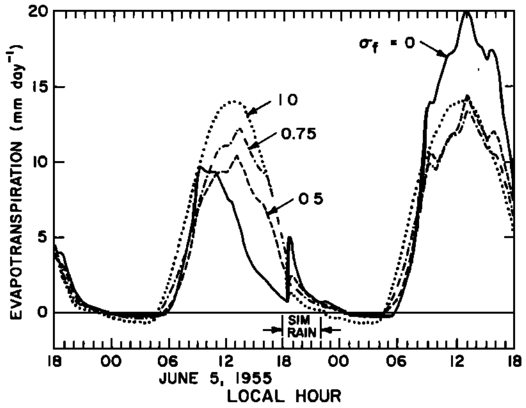


Fig. 9. Predicted variation of the evapotranspiration rate over the 2-day period for four values of the shielding factor.

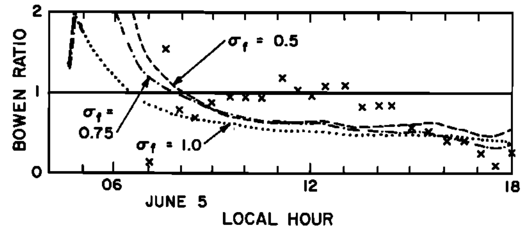


Fig. 11. The Bowen ratio above the foliage during the daytime of the June 5 simulation of *Penman and Long*'s [1960] data.

and circled crosses) are also presented, with in-canopy measurements at a height of 30 cm utilized as a measure of T_{af}. Although predicted differences are seen to be critically dependent upon σ_f, the constant value of $\sigma_f = 0.75$ captures the observed behavior fairly well in both phase and magnitude. We see that the sensible heat transfer from ground to canopy is positive at night and negative in the daytime, just opposite from the behavior experienced over bare ground.

A similar plot of the vapor pressure difference, $e_g - e_{af}$, is shown in Figure 7. Here the $\sigma_f = 0.75$ curve agrees even closer with the observations than could be expected. The positive values of the difference at night mean that part of the dew formation then comes from moisture supplied from the ground, a process called 'distillation' by *Monteith* [1957].

Predicted values of w_g, w_2, and W_{dew} during the 48-hour period for $\sigma_f = 0.75$ are presented in Figure 8. Predicted dewfall commences at 2240 (sunset being calculated for 2016), and W_{dew} reaches a maximum value of 0.08 mm per unit ground area by sunrise (predicted for 0408 h). Although the dewfall was not measured, Penman and Long stated that the night of June 4–5 at their site 'was one of exceptionally intense dew formation,' heavy dew being characterized by 0.15 mm or more [*Long*, 1958]. (For $\sigma_f = 1$ the predicted dewfall reached 0.16 mm.) Evaporation of the dew is predicted to have occurred by 0740 for $\sigma_f = 0.75$.

Although the ground is emitting moisture at night while the foliage is receiving it, the rate of dew accumulation is predicted to exceed E_g more and more as the night proceeds. Shortly after onset of dew, E_g exceeded the dew accumulation rate, but by 0400 the latter exceeded E_g by a factor of 16. Distillation was therefore unimportant after the first hour of dewfall. If the model is rerun with all wind speeds doubled but all else the same, the maximum dew accumulation is 7% less. This result can be traced mainly to a potential for increased dewfall through increased turbulent mixing to be more than compensated by warmer foliage at night and a smaller difference in $q_{af} - q_{sat}(T_f)$.

The ground surface moisture fraction, w_g in Figure 8, is predicted to decrease considerably between 0800 and 1800 on June 5 owing to evapotranspiration, but the amount of decrease for bare soil under similar conditions is very marked. During the simulated rainy period, w_g quickly increases beyond the saturated value. Following cessation of the rain the restoring term in (42) causes the soil surface moisture content to decrease slowly, as does evaporation the following day. The retained rainwater on the foliage starts disappearing after sunrise, but because of the increased amount in comparison with dewfall it is not predicted to vanish until 0940 on June 6.

The evapotranspiration rate E_h of (46) is shown in Figure 9 for this case of $\sigma_f = 0.75$ and also for $\sigma_f = 0.0$, 0.5, and 1.0. Unfortunately, there is no observational check from Penman and Long, except that the net value predicted over the first 24-hour interval, which is 0.43 cm, lies within the range of values 0.33–0.61 cm that they found using a large field balance during June 12–19, 1957, with generally clear skies for a wheat crop at the same site. It may be mentioned that the net 24-hour evapotranspiration is predicted to be 1.9 times larger for $\sigma_f = 1$ than for $\sigma_f = 0$, all else being held constant. This result follows from the enhanced dryness of the soil surface which develops during the day in the absence of foliage. To capture this effect, a model needs to include both a foliage parameterization and a prediction of soil surface moisture content as in (12). The skewness of the daytime June 5 E_g curve for $\sigma_f = 0$ is also caused by the progressive drying of the soil surface and is typical of stage 2 evaporation [*Idso et al.*, 1974].

If u_a is doubled in this test of the model, while all else is being held the same, the net 24-hour evapotranspiration for $\sigma_f = 0.75$ is found to be increased from 0.43 to only 0.49 cm. This somewhat surprising result stems from the foliage surfaces' being about 2 K cooler, owing to increased ventilation (and transpiration), and this leads to a correspondingly greater reduction in leaf surface saturation humidity due to its nonlinear temperature dependence.

If the stomatal resistance is halved and all else is held constant, the net 24-hour evapotranspiration for $\sigma_f = 0.75$ is found to increase from 0.43 to 0.51 cm. If r_s is increased by 50%, the net 24-hour evapotranspiration is found to decrease from 0.43 to 0.38 cm.

For $\sigma_f = 0$ a pronounced peak in E_h occurs during the simulated rainy period (Figure 9) when the ground becomes wet while it is still relatively warm. The bare soil evaporation rate is seen to be much increased on the day following the simulated rain, whereas the transpiration rate is little affected.

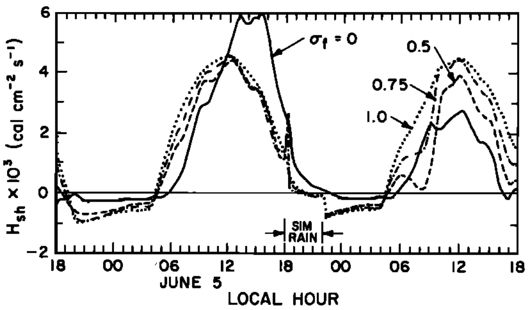


Fig. 10. Predicted variation of the net sensible heat flux to the atmosphere over the 2-day period for the same four values of the shielding factor.

The sensible heat flux to the atmosphere, H_{sh} in (45), is shown in Figure 10 for the same four values of σ_f. An upward spike occurs briefly at the beginning of the rainfall period, associated with increased downward longwave radiation to which the foliage quickly responds. If this rainfall is not simulated, the maximum sensible heat flux over the bare soil increases by 12% on the second day in comparison with the first and then exceeds H_{sh} for $\sigma_f = 0.75$ by 78%.

The investigation of Penman and Long was not designed for the purpose of obtaining the sensible heat flux, and as a gross check of the H_{sh} values obtained here the Bowen ratio (H_{sh}/LE_h) is compared with Bowen ratios measured by *Black and McNaughton* [1971] above a young Douglas fir forest in British Columbia in summer. This comparison can be made because there is little in this parameterization to distinguish a field of grass from a forest. The present model is seen to predict Bowen ratios in rough agreement with those measured above the forest, as indicated by Figure 11. Daytime values of Bowen ratio near ½ over meadowland on clear days in spring and summer are reported by Geiger. One estimate of Bowen ratio from the data of Penman and Long can be made for 1000–1100 h, June 5, and is 0.48 ± 0.15; at that time the model gives a value of 0.62 for $\sigma_f = 0.75$. Hence the model prediction of H_{sh} appears reasonable.

c. Discussion of vegetation parameterization. Although there are some existing foliage models [*Paltridge,* 1970; *Waggoner and Reifsnyder,* 1968], these are multilayer (e.g., eight layers within the foliage) and do not treat the fluxes from the soil surface underneath the canopy nor allow the canopy density to approach small values.

The present parameterization has purposely been kept extremely simple, from the viewpoint of an agricultural meteorologist, so that it might be useful to atmospheric modelers who cannot afford to become too entwined in a host of vegetative canopy details. Another reason for the simplicity is that the parameterization should be capable of working reasonably well for a given large-scale shielding factor, whether the foliage cover is homogeneous in space with $\sigma_f = 0.5$ or is very dense over half the area and nonexistent over the other half. Because of the linear interpolations upon σ_f built into this parameterization, it is subject to either interpretation of the foliage distribution.

The additional computer running time required of this parameterization is not great. The numerical program which involves prediction of the vegetation layer properties as well as of T_{gfr}, T_2, W_g, and W_2 was found to require only 1.5 times the computer time of a similar program stripped of all vegetative canopy and related statements. (Both programs were stripped of all superfluous or output statements.) The parameterization could therefore be made considerably more comprehensive without acquiring excessive computer requirements. Additional foliage variables requiring computer storage are T_f, W_{dew}, and q_{af}. Presumably σ_f will also be a variable which could be made a function of season and location (or latitude, elevation, seasonal soil moisture, and land use).

It is perhaps worth mentioning some of the considerations omitted from this parameterization. These include effects of stratification or free convection upon both c_{H0} and c_{Hh}; influence of different types of canopy, for given σ_f, upon mean wind, temperature, humidity, thermal radiation, and leaf area index within the canopy; different rates at which air within the canopy may be modified in temperature and humidity; influence of solar elevation angle upon σ_f, α_f, and α_g; dependence of liquid water retention upon canopy height and type; influence of soil water potential and of atmospheric temperature and humidity upon stomatal resistance; influence of soil properties upon the prediction of w_g; interaction of snowfall with a foliage layer, and effects of a thin layer of organic litter atop the soil.

As the parameterization now stands, its chief components subject to tuning are two of the three factors which sum to unity in (21*a*) and (21*b*), the stomatal factor 2 s cm^{-1} appearing in (27), the factor 0.83 in (20), the factor 7 for the maximum leaf area index in (24), and the values assigned to c_f, α_f, W_{dmax}, and w_{wilt}. To a certain extent, different types or combinations of canopies could be represented by somewhat different values of these parameters.

6. Summary and Conclusions

Bhumralkar's [1975] and *Blackadar*'s [1976] 'force restore' method of predicting ground surface temperature has been tested against five other approximate methods that are about equally efficient numerically and do not require calculation of multiple soil layer temperatures. In cases with substantial diurnal solar forcing this force restore method is found in general to be superior to the other methods. When the atmospheric forcing has a substantial random component to it, the insulated surface assumption is found to be especially poor. If no substantial diurnal forcing is present, the force restore method is still found to be superior to the insulated surface method. The procedure of relating the soil heat flux to either the sensible heat flux or the net radiative flux yields surface temperatures of intermediate but surprisingly acceptable accuracy if the time step does not exceed 10 or 20 min.

A method is presented for predicting ground surface moisture which is analogous to the force restore method of predicting surface temperature. The specific humidity at the surface is then related to the ground surface moisture content rather than to the bulk soil moisture content as is present practice. This permits evaporation to dry out the ground surface and so reduce the evaporation rate from bare soil in comparison with evapotranspiration. Also the method permits the new variable, ground surface moisture, to be treated with comparable care and time scale resolution as is afforded ground surface temperature.

A simple parameterization for a vegetation layer is developed and tested against observations of *Penman and Long* [1960] and others. It involves solution of an abbreviated energy budget equation to obtain the temperature of a representative foliage element and diagnosis of mean air temperature and humidity within the vegetation layer. The force restore method is utilized to calculate ground surface temperature and moisture. With no shielding of the ground from solar radiation ($\sigma_f = 0$), one recovers the bare soil values of ground temperature and surface heat flux. With complete shielding ($\sigma_f = 1$) the diurnal ground surface temperature wave is strongly damped but is still present because of heat and moisture exchange between the ground and the vegetation. With intermediate shielding of $\sigma_f = 0.75$ in simulation of the data of *Penman and Long* [1960], the difference between in-canopy and ground surface temperature and the corresponding difference in vapor pressure are predicted surprisingly well during the course of a day. The evapotranspiration rate above a dense foliage layer is predicted to exceed the bare soil evaporation rate by a factor of about 2, while the sensible heat flux in the afternoon over bare soil can typically exceed that over the foliage by a similar factor. There is therefore strong need to make use of a simple parameterization of the vegetation layer if one wishes to improve upon the method of calculating ground surface temperature and at the same time predict more accurate fluxes to the atmosphere.

The introduction of this vegetation parameterization does add some 50% to the computer time otherwise necessary for calculating the bulk plus ground surface values of temperature and moisture. However, this modest expansion of a minor part of an atmospheric prediction model seems fully warranted in view of the gross errors which can occur when the foliage layer is ignored. Introduction of a foliage layer which may have variable density and an albedo different from that of the ground seems mandatory for testing a climate theory like that of *Charney et al.* [1975], for example, which involves interactions between surface albedo, soil moisture, large-scale weather and precipitation, and crop type and amount. In their paper, Charney et al. do point out the need for including a model of the biosphere within the atmospheric model.

NOTATION

Arabic

c specific heat in general.

c_f dimensionless heat or moisture transfer coefficient for the foliage element.

c_H dimensionless heat or moisture transfer coefficient applicable to bare soil, c_{H0}; to soil under a canopy, c_{Hg}; or to the top of a dense canopy, c_{Hh}.

c_p specific heat of air at constant pressure.

c_s specific heat of soil.

c_1, c_2 constants in force restore rate equation (8) for ground surface temperature.

C_1, C_2 coefficients in the rate equation (12) for ground surface moisture.

d_1 a soil depth influenced by the diurnal temperature cycle, equal to $(\kappa_s \tau_1)^{1/2}$.

d_1' a soil depth (10 cm) influenced by the diurnal soil moisture cycle.

d_2 a soil depth influenced by the annual temperature cycle, equal to $19.1 d_1$.

d_2' a soil depth (50 cm) influenced by seasonal moisture variations.

e vapor pressure.

E evaporation rate in general.

$(E_f)_{\text{pot}}$ potential evaporation rate from foliage.

e_r root-mean-square error of approximately calculated ground temperature, relative to the diurnal range.

E_{tr} foliage transpiration rate.

G soil heat flux at the surface (positive when directed into the soil).

H_A sum of fluxes to atmosphere (positive when directed upward).

H_s sensible heat flux (positive when directed upward).

K degrees Kelvin.

L latent heat of vaporization.

n time step index.

N net leaf area index.

P precipitation rate (mass per unit time and area).

q specific humidity in general.

q_{af} a mean specific humidity of air within a canopy.

$q_{\text{sat}}(T)$ saturation specific humidity at temperature T.

r_a atmospheric resistance, equal to $(c_f u_{af})^{-1}$.

r_s generalized stomatal resistance (dimensions of inverse velocity).

r' soil moisture interpolation factor.

r'' fraction of potential evaporation rate from foliage.

$R_L{\downarrow}$ downward directed longwave radiative flux.

$R_L{\uparrow}$ upward directed longwave radiative flux.

R_{net} net shortwave and longwave radiative flux.

$S{\downarrow}$ downward directed shortwave radiative flux.

$S{\uparrow}$ upward reflected shortwave radiative flux.

$\mathcal{S}$ seasonal dependence of stomatal resistance.

t time.

T absolute temperature.

T_{af} a mean air temperature within a canopy.

T_f foliage or leaf surface temperature.

T_g ground surface temperature.

T_{gf} T_g calculated by the H_A forcing method.

T_{gfr} T_g calculated by the force restore method.

T_{gi} T_g calculated from the insulated surface assumption.

T_{gm} T_g calculated from a multilayer soil model.

T_{gr} T_g calculated from the R_{net} dependence assumption.

T_{gs} T_g calculated from the H_{sg} dependence assumption.

T_1 soil temperature at depth z_1 of multilayer model.

T_2 mean soil temperature over layer of depth d_2.

u_{af} a mean wind speed within a canopy.

w volumetric concentration of soil moisture (dimensionless).

w_k critical or saturated value of w.

$w_{\max}$ maximum value of w.

W_2 soil moisture content (depth of liquid water) within the depth d_2.

W_{dew} mass of liquid water retained by foliage per unit horizontal ground area.

W_{dmax} maximum value of W_{dew} beyond which runoff to soil occurs.

W_k field capacity, or saturated value of W_2.

$W_{\max}$ maximum or runoff value of W.

$w_s = 0.9 w_2 + 0.1 w_g$.

w_{wilt} a wilting point value of w.

z height above surface or depth below surface.

z_1 thickness of first soil layer in multilayer model.

Subscripts

a reference 'anemometer level' height.

f foliage surface.

g value at the ground surface.

h height just above the top of the canopy.

0 evaluation at the surface of bare soil.

2 mean soil values averaged over the layer d_2.

Greek

α degree of bulk soil water saturation.

α' degree of soil surface water saturation.

α_f foliage albedo.

α_g ground surface albedo.

δ soil depth increment, equal to 1 cm.

δ_c step function, equal to 1, except 0 during condensation.

Δ a difference operator.

ϵ_f foliage emissivity.

ϵ_g ground surface emissivity.

ζ transformed depth coordinate in soil, equal to $\ln(1 + z/\delta)$.

κ_s soil thermal diffusivity.

λ thermal conductivity of soil.

σ Stefan-Boltzman constant.

σ_c cloud fraction.

σ_f foliage shielding factor of ground from shortwave radiation, area average.
τ_1 diurnal period.
ρ_a density of air.
ρ_s density of soil.
ρ_w density of water.

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