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Documentation for Program SOILSIM
A Computer Program for the Simulation
of Heat and Moisture Flow in Soils
and Between Soils, Canopy and Atmosphere

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SOILSIM Program Description
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INTRODUCTION

SOILSIM is a digital model of energy and moisture fluxes in the soil and above the soil surface. It simulates the time evolution of soil temperature and moisture, temperature of the soil surface and plant canopy the above surface, and the fluxes of sensible and latent heat into the atmosphere in response to surface weather conditions. The model is driven by simple weather observations including wind speed, air temperature, air humidity, and incident radiation. The model is intended to be useful in conjunction with remotely sensed information of the land surface state, such as surface brightness temperature and soil moisture, for computing wide area evapotranspiration. The model was developed by Peter Camillo and reported on in a number of papers (Camillo and Schmugge, 1983; Camillo, Gurney and Schmugge, 1983; Camillo, 1986;). Prior to his death in 1988, Camillo had substantially modified the program to include an option for a "two layer" model structure which allowed separate values to be solved for the soil surface and canopy temperatures. A working version had been developed of the revised model which was used with data from the HAPEX exercise (Camillo, 1988). However, this FORTRAN code was not well described and was not usable by the uninitiated. An early version of the model was carefully documented by Camillo and Schmugge (1981). Some of the substantial changes made to the FORTRAN code after this time were partially described informally by Camillo in several draft documents. This description has been prepared from Camillo's original and revised documentation to the extent possible. Description of the scientific and mathematical basis of the modifications and additions to the FORTRAN code had to be worked out by reverse engineering the code in conjunction with the references with which he was working. In the present documentation, Camillo's original text is retained where it applies. The section describing the canopy modelling was written by Field. Procedures have been added so that simulator now computes local mean time in addition to simulator time. This was added to rectify the times of simulator output and observations. Also the declination of the sun is now computed from data on latitude and day of the year. The list of references has been updated to include additions to the model and results from model simulations.

In brief, the approach integrates a pair of differential equations for soil heat and moisture flux in each soil layer to satisfy certain deep soil and soil surface boundary conditions. The chief boundary condition at the soil surface is the total flux of heat between the soil and air above the soil. The fluxes of radiation, sensible and latent heat above the soil surface, whether bare or covered with a plant canopy, are computed iteratively as responses to the forcing induced by the observed weather observations. The solved above-surface "boundary fluxes" are such that their sum at the soil surface equals the surface value of the solved soil heat flux. The soil surface temperature must satisfy both the below-surface and above-surface fluxes. When a canopy is present, temperatures for plant leaves and canopy air are calculated. These may be either the same as the surface temperature (a one layer model) or different from it (a two layer model). Comparing surface temperature or canopy temperature with observations provides a test of the quality of energy budget solution. Alternatively, one may compare observations of soil moisture with calculated values.

The model was originally written to run on an IBM 3060. The model has been ported to the DEC VMS environment to run on VAX computers by Karen Humes, University of Arizona. It was ported from this version to the UNIX environment on a SUN 386i workstation by John Kuhn and Richard Field at the University of Delaware Center for Remote Sensing. Rather extensive modifications were necessary were to provide file interfacing to UNIX and the windowing environment of the SUN operating system (Sun OS 4.0). The program has been satisfactorily tested on both the Sun 3 and Sun 386i computers. In what follows only those changes to variables which appear in the FORTRAN code and affect running the model are discussed. Discussion of the code interfacing the FORTRAN code to UNIX and the X Windowing System, written in both FORTRAN and C, will be discussed at a later time.

MATHEMATICAL DESCRIPTION

1. MODEL EQUATIONS

A. Soil Heat and Moisture Flux Equations

The slow movement of heat and moisture in a porous medium such as soil can be described by diffusion-type equations (Nielson et. al., 1972). In the classical diffusion theory, the flux (the amount of a substance crossing a unit area per unit time) is proportional to the negative of the gradient of the concentration. The proportionality factor is the diffusion coefficient. The best known example of this kind of flow is embodied in Darcy's Law (Hillel, 1971). The wetness flux is

$$q_\theta = -K(\theta)\vec{\nabla}(\psi(\theta)-z) \tag{1}$$

where q_θ is the flux (cubic centimeters of water per square centimeter per second (cm sec^{-1}), K(θ) is the hydraulic conductivity (cm sec^{-1}), $\psi(\theta)$ is the matric potential (cm), and z is the distance from some reference point. The term ψ - z is the hydraulic head and is the potential energy of the soil water (matric plus gravitational energy) per unit weight of water. The function ψ is called the matric potential and is the energy per unit weight required to overcome the capillary and adhesion forces that bind the water to the soil. Because work must be done to remove water from an unsaturated soil, ψ is negative. The distance z is the gravitational potential energy per unit weight. K and ψ are functions of volumetric moisture θ (cm^3 water per cm^3 medium). In this application, it is assumed that soil properties change only with depth; thus the gradient is a derivative with respect to z. Therefore, Equation 1 may be written as

$$\begin{aligned} q_\theta &= -K(\theta)\left(\frac{d\psi}{dz}-1\right) \\ &= -K(\theta)\frac{d\psi}{d\theta}\frac{d\theta}{dz}+K(\theta) \end{aligned} \tag{2}$$

The second line follows from the chain rule of differentiation.
Defining a diffusion coefficient

$$D(\theta) = K(\theta)\frac{d\psi}{d\theta} \tag{3}$$

yields, when inserted into Equation (2)

$$q_\theta = -D(\theta)\frac{d\theta}{dz}+K(\theta) \tag{4}$$

The first term in Equation 4 is the diffusion contribution to the moisture flux due to a wetness gradient.

There is a large body of experimental evidence indicating that thermal gradients induce moisture flow (Nielsen et al., 1972). For example, if a uniformly moist soil sample is enclosed in a horizontal cylinder and is subjected to a thermal gradient, moisture flows from the warm toward the cool end. As field soil temperatures are always changing, an isothermal model such as Equation (4) is not complete; a theory that treats both heat and moisture flow in soils is necessary. In the following description, diffusion-type expressions for both heat and moisture fluxes are presented. The derivation closely follows the work of Philip and De Vries (1957). Contributions to heat and wetness fluxes that are proportional to wetness and temperature gradients are described. The conservation of mass and energy is the invoked to derive the partial differential equations that describe the variation with time of temperature and moisture profiles.

The diffusive flux of water vapor in a porous medium is modeled as

$$\vec{q}_v = -D_{atm} f(\varepsilon,\theta)\vec{\nabla}\rho \tag{5}$$

where q_v = vapor flux density (gm cm^{-2} sec^{-1})
D_{atm} = molecular diffusivity of water vapor in air (cm^2 sec^{-1})
f = tortuosity and porosity function
ρ = density of water vapor (gm cm^{-3})

Equation (5), with f = 1, describes the diffusion of water vapor in air (Eagleson, 1970). The factor f represents the reduced volume available for vapor diffusion in the soil matrix due to obstacles such as soil particles and liquid water which adheres to them. An experimentally determined graph of f as a function of θ can be parameterized by a linear function (Jackson et al., 1974)

$$f(\theta) = \alpha(\varepsilon - \theta) \tag{6a}$$

where ε = soil porosity
α = constant less than 1

The diffusivity D_{atm} is a function of temperature and can be adequately modeled by the equation (Kimball et al., 1976)

$$D_{atm} = D_o\left(\frac{T}{273.16}\right)^{1.75} \tag{6b}$$

where D_o = 0.229 (cm^2 sec^{-1}) and T is the absolute temperature.

The gradient in Equation 5 is to be evaluated in terms of moisture and temperature gradients as these are the dependent variables of the model. This can be accomplished by using the relationship between vapor density and relative humidity

$$\rho = \rho_o h = \rho_o \exp[(\psi g)/(RT)] \tag{7a}$$

where ρ_o = density of saturated water vapor

h = relative humidity
g = gravitational acceleration constant
R = gas constant for water vapor = 4.615 X 10^6 (ergs gm^{-1} K^{-1})

The vapor density ρ_o depends on temperature and can be approximated by (Kimball et al. 1976)

$$\rho_o(T) \cong \exp[R_o - (R_1/T)] \tag{7b}$$

where R_o = 6.0035
R_1 = 4975.9 (K)
T = temperature (K)

Equation 7 can be derived from the laws of thermodynamics. Assuming water vapor behaves as an ideal gas, an expression can be readily obtained relating the vapor pressure, the temperature, and the chemical potential of the gas. The chemical potential and the matric potential of liquid water are related because they both represent the free energy of the respective phases and the two phases are in equilibrium. The gas density is proportional to the partial pressure..

The gradient in Equation 5 can be expressed in terms of temperature and moisture gradients as follows:

$$\begin{aligned} \vec{\nabla}\rho &= \vec{\nabla}(\rho_o h) = \rho_o \vec{\nabla} h + h \vec{\nabla}\rho_o \\ &= \rho_o\left(\frac{\partial h}{\partial T}\vec{\nabla} T + \frac{\partial h}{\partial \theta}\vec{\nabla}\theta\right) + h\left(\frac{\partial \rho_o}{\partial T}\vec{\nabla} T + \frac{\partial \rho_o}{\partial \theta}\vec{\nabla}\theta\right) \end{aligned} \tag{8}$$

The derivative of h with respect to T can be computed from Equation 7 according to

$$\begin{aligned} \left(\frac{1}{h}\right)\frac{\partial h}{\partial T} &= \frac{-\psi g}{RT^2} + \left(\frac{g}{RT}\right)\left(\frac{\partial \psi}{\partial T}\right) \\ &= \frac{-\ln h}{T} + \ln h\left(\frac{1}{\psi}\right)\left(\frac{\partial \psi}{\partial T}\right) \end{aligned} \tag{9}$$

The matric potential is dependent on temperature through the dependence of the surface tension of water on temperature, which is responsible for the capillary force that binds the water to the soil matrix. Therefore, ψ is proportional to surface tension σ (Philip and De Vries, 1957) and we may write

$$\left(\frac{1}{\psi}\right)\frac{\partial \psi}{\partial T} = \left(\frac{1}{\sigma}\right)\frac{\partial \sigma}{\partial T} \tag{10}$$

A table giving surface tension at a pressure of one atmosphere as a function of temperature (Eagleson 1970) can be fit with the exponential

$$\sigma(T) = \sigma_o \exp[-\gamma(T-273.16)] \tag{11}$$

where σ_o = 75.9 (dyne cm^{-1})
θ = 2.09 x 10^{-3} (K^{-1})
T = temperature (K)

The derivative of ψ with respect to T can be computed using Equations 10 and 11. Equation 9 therefore is

$$\frac{dh}{dT} = -h \ln h \left(\gamma + \frac{1}{T}\right) \tag{12}$$

The θ dependence of h is, from Equation 7

$$\frac{dh}{d\theta} = \left(\frac{g}{RT}\right)\frac{d\psi}{d\theta} = h \ln h \frac{d}{d\theta} \ln\psi \tag{13}$$

Matric potential ψ typically changes by four to six orders of magnitude over the range of wetness values normally found in unsaturated soils. A comparison of Equations (12) and (13) shows that the variation of h with θ is much larger than the variation of h with T, at least over the range of temperatures found in soils (273 to 310 K). Therefore, relative humidity h in Equation (8) is considered to be only a function of θ.

Since water vapor behaves approximately like an ideal gas, its density depends primarily on pressure and temperature. Therefore, ρ_o can be assumed to be a function of temperature only, with no dependence on θ. With h depending only on θ, and ρ_o depending only on T, Equation (8) becomes

$$\vec{\nabla}\rho = \rho_o\left(\frac{\partial h}{\partial \theta}\right)\vec{\nabla}\theta + h\left(\frac{\partial \rho_o}{\partial T}\right)\vec{\nabla}T \tag{14}$$

Inserting Equation (14) into Equation (5) and using Equation (13) for $dh/d\theta$ yields

$$\vec{q}_v = -D_{T,vap}\vec{\nabla}T - D_{\theta,vap}\vec{\nabla}\theta \tag{15a}$$

where

$$D_{T,vap} = D_{atm}\alpha(\varepsilon - \theta)\, h\left(\frac{d\rho_o}{dT}\right) \tag{15b}$$

$$D_{\theta,vap} = \frac{D_{atm}\alpha(\varepsilon - \theta)\rho_o g h}{RT}\left(\frac{d\psi}{d\theta}\right) \tag{15c}$$

This is the sought for diffusion expression for the vapor flux. Diffusion coefficients $D_{T,vap}$ and $D_{\theta,vap}$ respectively called the "thermal vapor diffusivity" and the "wetness vapor diffusivity" depend on both θ and on T.

The liquid flux can be computed from "Darcy's law" (Equation (1)). The gradient of ψ in terms of moisture and temperature gradients is

$$\vec{\nabla}\psi = \frac{\partial\psi}{\partial\theta}\vec{\nabla}\theta + \frac{\partial\psi}{\partial T}\vec{\nabla}T \tag{16}$$

Equations (10) and (11) give the derivative of ψ with respect to temperature:

$$\frac{\partial\psi}{\partial T} = -\gamma\psi \tag{17}$$

Thus, the liquid flux is

$$q_1 = K\vec{\nabla}z - D_{\theta,liq}\vec{\nabla} - D_{T,liq}\vec{\nabla}T \tag{18a}$$

where

$$D_{\theta,liq} = K\left(\frac{\partial\psi}{\partial\theta}\right) \tag{18b}$$

$$D_{T,liq} = -K\gamma\psi \tag{18c}$$

The total moisture flux, q_θ (gm cm^{-2} sec^{-1}), is the sum of the vapor and liquid fluxes

$$\vec{q}_\theta = \vec{q}_v + \vec{q}_1 \tag{19}$$

This can be written in a diffusion form by adding Equations (15) and (18) to obtain

$$\vec{q}_\theta = -D_\theta\vec{\nabla}\theta - D_T\vec{\nabla}T + K\vec{\nabla}z \tag{20a}$$

where

$$D_\theta \equiv D_{\theta,liq} + D_{\theta,vap} \tag{20b}$$

$$D_T \equiv D_{T,liq} + D_{T,vap} \tag{20c}$$

The volumetric water content, θ, is the volume of moisture per unit volume of soil. Because the density of water is 1 gm cm^{-3}, θ also represents the mass of water per unit volume of soil, assuming that the water includes the liquid phase plus the gas phase. As θ represents the mass and q_θ is the mass flux, they are related by the continuity equation

$$\frac{d\theta}{dt} + \vec{\nabla}\cdot\vec{q}_\theta = 0 \tag{21}$$

This is a partial differential equation involving θ as a function of depth and time. An analogous diffusion equation can be derived to describe the time dependence of the temperature profile as a function of the soil heat flux.

Fourier's heat flow equation gives the heat flow due to a temperature gradient as

$$\vec{q}_{h,T} = -\lambda \vec{\nabla} T \tag{22}$$

where $q_{h,T}$ is the temperature-driven heat flux (calories cm^{-2} sec^{-1}) and λ is the thermal conductivity of the medium (cal cm^{-1} sec^{-1} K^{-1})

To apply this equation to heat transfer in the soil, the effective thermal conductivity of the soil-water-air system must be known. A generally accepted model (De Vries, 1975) gives λ as a weighted average over the thermal conductivity of each soil constituent:

$$\lambda = \frac{f_w \lambda_w + k_s f_s \lambda_s + k_a f_a (\lambda_a + \lambda_{vap})}{f_w + k_s f_s + k_a f_a} \tag{23}$$

where f_w, f_s and f_a are the volumetric fractions of the liquid, solids, and air constituents respectively. (It should be noted that f_w and θ are the same, and the porosity, ε, is equal to $f_w + f_a$.) The thermal conductivities of each component are λ_w, λ_s, and λ_a. Factor k_s represents the ratio of the average thermal gradient in the solid constituents of the soil to the average thermal gradient in water. It also depends on the shape and orientation of the soil grains. For spheroid-shaped particles, factor k_s is given by

$$k_s = \frac{2}{3}\left[1 + \left(\frac{\lambda_s}{\lambda_w} - 1\right) g_s\right]^{-1} + \frac{1}{3}\left[1 + \left(\frac{\lambda_s}{\lambda_w} - 1\right)(1 - 2g_s)\right] \tag{24}$$

where g_s is the shape factor and is equal to 1/2 for cylinders of infinite length, 1/3 for spheres, and 0 for disks of infinite radius. The complete model uses a sum over various soil solids, but only one representative solid is allowed in this program. The weight factor k_a for air can be determined from Equation (24) with λ equal to the thermal conductivity of dry air. The air shape factor g_s in this case would have no physical meaning and is usually treated as a variable function of moisture that must be determined for each soil type. Therefore the air shape factor k_a (and not g_a) is input for the air phase.

The latent heat absorbed or emitted by the soil as the wetness changes state between the liquid and vapor phases can be an important cause of temperature fluctuations. This heat can be included in the heat flux by assuming that the vapor flux carries with it a heat flux due to the latent heat of vaporization that it absorbed from the soil when it evaporated. This heat flux carried by the vapor phase is

$$\vec{q}_{h,v} = L\vec{q}_v$$

where L is the latent heat of vaporization (cal gm^{-1}) and q_v is the vapor flux (Equation (15)). Both thermal and moisture gradients contribute to q_v and therefore contribute to $q_{h,v}$. The moisture contribution is computed by inserting the appropriate term from Equation (15) into the above equation to obtain

$$\vec{q}_{h,v(\theta)} = L D_{\theta,vap} \vec{\nabla} \theta \tag{25}$$

The temperature gradient contribution from Equation (15) is included by increasing the apparent thermal conductivity of the air-filled pores, where the vapor phase primarily exists. This vapor has thermal conductivity λ_{vap} and carries heat flux $-\lambda_{vap}\nabla T$ according to Fourier's law, where ∇T is the temperature gradient in the pore. However, this heat can also be represented by the thermal term in Equation (15) with porosity factor f set equal to 1 as

$$-D_{atm} h \left(\frac{d\rho_o}{dT} \right) \vec{\nabla} T$$

by equating these two expressions for the same heat flux, the vapor conductivity is found to be

$$\lambda_{vap} = L D_{atm} h \left(\frac{d\rho_o}{dT} \right) \tag{26}$$

Therefore, the total heat flux in the soil is

$$\vec{q}_h = \vec{q}_{h,T} + \vec{q}_{h,v(\theta)}$$

where λ is given by Equation (23) and includes the vapor thermal conductivity.

The total thermal energy per unit volume of medium at temperature T is CT, where C is the volumetric heat capacity (cal cm^{-3} K^{-1}) and T is the temperature in kelvins. The conservation of heat energy leads to an equation similar to the conservation of mass for water (Equation (21)

$$\frac{d(CT)}{dt} + \vec{\nabla} \cdot \vec{q}_h = 0 \tag{28}$$

The volumetric heat capacity of the soil is computed as a sum over the capacities of the constituents (De Vries, 1975) by

$$C = f_s C_s + f_w C_w + f_a C_a \tag{29}$$

Fractions f_s f_w f_a are the volumetric constituents of solid, water, and air; and C_s, C_w, and C_a are the heat capacities of the constituent solids, water, and air respectively.

Equations (21) and (28) describe the time dependence of soil moisture and temperature profiles. In this application, only vertical fluxes are considered; this constitutes a stratified model of the soil. Therefore, the gradient operator can be replaced by a derivative with respect to soil depth. Thus, the moisture and heat fluxes are, from Equations (20) and (27), given by

$$q_\theta = -D_\theta \left(\frac{d\theta}{dz} \right) - D_T \left(\frac{dT}{dz} \right) - K \tag{30a}$$

$$q_h = -\lambda \left(\frac{dT}{dz} \right) - L D_{\theta,vap} \left(\frac{d\theta}{dz} \right) \tag{30b}$$

The time derivatives of the moisture and temperature profiles, from Equations (21) and (28), are

$$\frac{d\theta}{dt} = -\frac{dq_\theta}{dz} \tag{31a}$$

$$\frac{dT}{dt} = \frac{1}{C}\left(\frac{dq_\theta}{dz}\right) \tag{31b}$$

B. Boundary Conditions for the Soil Flux Model Equations

To solve these equations, boundary conditions must be supplied for both wetness and temperature at the air/soil interface and in the bottom layer of the soil profile. In principle, either the fluxes q_θ and q_h or the variables θ and T could be specified. In the simulator both heat and moisture fluxes q_θ and q_h are computed at the soil surface. In this way the effects of the environment (ie, rainfall, evapotranspiration, radiation, etc.) on the evolution of the soil temperature and moisture profiles can be modeled. At the bottom of the profile a mixture of flux and variable boundary conditions are used. One can specify constant wetness, a downward wetness flux equal to the hydraulic conductivity of the bottom layer, a flux equal to zero, or a constant flux. The temperature in the bottom layer is held constant. When both temperature and moisture profiles are modeled, the surface fluxes can be found by the solution of the heat balance equation

$$S = R + LE + H \tag{32}$$

All fluxes are given as flux densities and are positive downward. S is the heat absorbed by the soil, R is the net radiation flux density, L·E is the evapotranspiration heat flux density, and H is the sensible heat flux density. This equation can be written with the temperature at the soil surface as the only unknown variable. After finding the solution, the soil heat flux density at the surface, q_h, is set equal to S, and the soil surface moisture flux density q_θ is set equal to E.

The heat flux absorbed by the soil can be evaluated by using the discrete form of equation (22)

$$S = -\lambda_1\left(\frac{T_1 - T_s}{Z_1}\right) \tag{33}$$

where λ_1 = thermal conductivity of the surface soil layer
T_1 = temperature at the center of the surface soil layer
T_s = temperature at the soil surface
Z_1 = depth to the center of the first layer

1. Incident radiation

The net radiation R is usually divided into average net short and net long wavelength components (Eagleson, 1970)

$$R = R_{short} + R_{long} \tag{34a}$$

The net short wave component of R at the surface is

$$R_{short} = (1 - A)(1 - B) I_c \tag{34b}$$

where A = surface short wave albedo (constant)
B = fraction of short wave radiation absorbed by the cloud cover
I_c = insolation at the Earth's surface for a cloudless sky

The insolation at the surface, I_c,is usually one the the observed meteorological inputs to the simulator. However, when MODSOL = 1 the simulator employs an empirical model for I_c

$$I_c = I_o (1 - d) \left(\frac{\sin\alpha}{\exp\left(n \left(c_0 + c_1 \frac{\log_{10}(\sin\alpha)}{\sin\alpha} \right) \right)} \right) \tag{34c}$$

in which

$$\sin\alpha = \sin\delta \sin\phi + \cos\delta \cos\phi \cos\tau \tag{34d}$$

where I_o = short wave solar energy flux density incident at the top of the Earth's atmosphere (0.033 cal cm^{-2} sec^{-1})
n = air turbidity factor (n ~ 2-5)
α = angle between the sun and the local tangent plane (solar altitude)
ϕ = local latitude
δ = angle between the sun and the plane of the celestial equator (in radians) (-23 deg $\leq \delta \geq$ 23 deg)
τ = hour angle of the sun (radians) = $W_d(t - 12)$
t = hour of the day
$W_d = \pi/12$ (rad/hour = 7.2722 x 10^{-5} rad/sec)
c_o = scattering model intercept (0.128) (SCAT0)
c_1 = scattering model slope (0.054) (SCAT1)
d = short wave absorption by clouds. During rain storms d = (0.5); else d = CLDABS (ATTEN, else CLDABS=CLDIN(I))

The solar declanation, δ, is approximated from the day of the year according to (Bras, 1990).

$$\delta = \frac{23.45\,\pi}{180} \cos\left[\frac{2\,\pi}{365}(172 - D)\right] \tag{34e}$$

In order to rectify the times of simulator output and observations made at a site, simulator noon is adjusted to clock noon using an approximation to the equation of time (EQT) (Nautical Almanac Office, 1987). Compute first the number of fractional days (FD) into the year in as of noon on day D, universal time.

$$FD = D + CNOON + [\lambda_g(\pi/180)(43200/\pi)]\,(24/3600) \tag{34f}$$

The equation of time, in seconds, for noon at the location of the simulation is approximated by

$$EQT = 60\,[-7.66 \sin((0.9856\,FD + TC1)\frac{\pi}{180}) - 9.78 \sin((1.9712\,FD + TC2)\frac{\pi}{180})] \tag{34g}$$

and the simulator noon is calculated

$$\mathrm{TNOON} = \mathrm{CNOON} - \mathrm{EQT}$$

where D = the day of the year ($1 \le D \le 365$ or 366) in universal time. This may be the first day of the simulation or the day mid way through a multi day simulation; it does not change during the simulation.

FD = fractional days into the year (universal time, in days)

λ_g = longitude of the local meridian (west longitude is negative) (in degrees and fractional degrees

TNOON = seconds from integration start to simulator noon

CNOON = 43200 (number of seconds in 12 hours)

TC1 = first adjustable constant (adjustable for current year)

TC2 = second adjustable constant (adjustable for current year)

If the observations are recorded in day light saving time the appropriate adjustment will be made by the program.

When the net incident short wave radiation is to be modeled (when IRNET equals any of 0, 2, or 3, or MODSOL=1) the surface albedo, A_{sur}, must be modeled. This is done, when the sun is above the horizon, according to a polynomial

$$A_{sur} = c_1 + c_2\sin\alpha + c_3(\sin\alpha)^2 + c_4(\sin\alpha)^3 \tag{34g}$$

where the c's are fit to the surface conditions. Default values are $c_1 = 0.22$, $c_2 = -0.08$, $c_3 = c_4 = 0$, which produces an albedo of 0.15 when the sun is 58 degrees above the horizon. When the sun is below the horizon, A_{sur} is set to 0.

The net short wave is $I_{net} = I_o\,(1 - A_{sur})$.

The contribution to the net radiation from the long wavelength part of the spectrum is modeled by

$$R_{long} = \varepsilon_s(\sigma\,\varepsilon_a T_a^4 - \sigma T_s^4) \tag{34h}$$

where T_a = air temperature (K)
ε_a = the emissivity of the air
T_s = surface temperature (K)
σ = Stefan-Boltzman constant
ε_s = emissivity of the surface

The emissivity of the air is modeled as (Eagleson, 1970)

$$\varepsilon_a = 0.74 + 0.005\, e_a \tag{34i}$$

where e_a is the vapor pressure of the atmosphere in millibars. The vapor pressure and air temperature are supplied by the user, and should be taken from the same height above the surface. Equation (34h) is a mathematical statement of the assumption that the air and land surface radiate with emissivity of ε_a and ε_s respectively.

It was Camillo's intention that if a canopy were present, the net short wave radiation (short wave absorbed by the canopy) would be computed from an analytical solution to the equation for multiple scattering which models short wave radiation in a plant canopy (Camillo, 1987). Although this is not yet implemented, the result of Camillo's development is briefly outlined here.

The single scattering phase function p is the probability of scattering from the incident direction (u',ϕ') to (u,ϕ) and is modeled by

$$p(u,\phi;u',\phi') = \omega\,[1 + x\cos(\Omega)]$$

where ω = the single scattering albedo
x = a measure of ansiotropy in p
Ω = the scattering angle.

Direct solar and diffuse sky radiation are integrated seperately over the hemisphere. The bidirectional canopy flux is expressed divided by the hemispherically integrated solar radiation incoming at the top of the canopy. I is therefore the ratio of the scattered flux to the flux incident normal to the surface.

$$I(L,u,\phi) = f\, I_{h0}(L,u,\phi) + (1 - f)\, I_d(L,u,\phi) \tag{35}$$

where $I(L,u,\phi)$= flux of scattered radiation in the canopy at optical depth L in the direction u,ϕ
f = the fraction of diffuse (sky) radiation in the incident short wave flux
I_{h0}= the short wave radiant flux in the canopy from diffuse radiation
I_d = the short wave radiant flux in the canopy from direct solar radiation
L = the leaf area index (optical depth within the canopy)
u = the cosine of the zenith angle of the radiation
ϕ = azimuth angle of the radiation

The directional reflectance due to the diffuse sky radiation flux in (35) is

$$I_{ho}(L,u) = A_o \exp\left(\frac{k_o L}{2}\right)\left(\frac{1 + a_o u}{1 - k_o u}\right) + B_o \exp\left(\frac{-k_o L}{2}\right)\left(\frac{1 - a_o u}{1 + k_o u}\right) \tag{36a}$$

and the directional reflectance due to the direct solar flux in (35) is

$$\begin{aligned} I_d(L,u,\phi) = {} & A_1 \exp(k_o L)\left(\frac{1 + a_o u}{1 - k_o u}\right) \\ & + B_1 \exp(-k_o L)\left(\frac{1 - a_o u}{1 + k_o u}\right) \\ & + h_o \exp\left(\frac{-G L}{u_o}\right)\left(\frac{1 + b_o u}{1 + \frac{G u}{u_o}}\right) \\ & + h_1 \exp\left(\frac{-G L}{u_o}\right)\left(\frac{\sqrt{1 - u^2}}{1 + \frac{G u}{u_o}}\right)\cos(\phi_o - \phi) \end{aligned} \tag{36b}$$

where G = the average projection of the leaves in the direction u_o

u_o, ϕ_o = the direction of the sun

the constants A_o, B_o, A_1, B_1, k_o, a_o, b_o, h_o, and h_1 are all explicit functions of the total canopy leaf area index, single scattering albedo, phase function ansiotropy parameter, soil albedo, and $G(u_o)$

The canopy albedo (the hemispherically integrated reflected radiation over all wavelengths) is

$$I^-(o) = \frac{1}{2\pi}\int_0^{2\pi} d\phi \int_0^1 u \, du \;\; I(0, -u, \phi) \tag{37a}$$

and can be evaluated analytically with Equations (35) and (36). Similarly, the flux at the bottom of the canopy incident on the soil surface is

$$I\ (L_T) = \frac{1}{2\pi}\int_0^{2\pi} d\phi \int_0^1 u \, du \;\; I(L_T, u, \phi) \tag{37b}$$

where L_T is the total leaf area index (integrated from the top of the canopy to the soil surface). The signs indicate radiation direction, positive toward the soil surface. Then, with the soil albedo, A_s, (assumed Lambertian), the short wave radiation absorbed by the soil (RS_s) and canopy (RS_c) are

$$RS_s = (1 - A_s)(I^{+}(L_T) + \exp(-G L_T / u_o)) S_o \tag{38a}$$

$$RS_c = (1 - I^{-}(o)) S_o - R S_s \tag{38b}$$

where S_o is the hemispherical integral of the incident flux. The exponential function in Equation (38a) accounts for the direct solar radiation which passes through the canopy without being absorbed.

Equations (38a) and (38b) are evaluated for both visible (ω~0.2) and near IR (ω~0.8) wave bands. If a two layer model (ICAN=1) is specified, the short wave radiation absorbed by the canopy (SLCAN) and soil (SLSOIL) are modelled by the simulator (in SUBROUTINE SUN) through the following expressions which must partition I_{NET}, the solar radiation absorbed by the surface.

$$I_{soil} = I_{NET} (-0.039 + 0.13 \sin\alpha) \tag{39a}$$

$$I_{can} = I_{NET} (0.818 - 0.048 \sin\alpha) \tag{39b}$$

I_{NET} is either observed or computed by the program from the albedo and radiation models in SUBROUTINE SUN. These coefficients are current default in the program. It should be noted that they depend on the leaf area index of the canopy and must be modified as LAI changes. One approach to accomplishing this is through Camillo's canopy reflectance model (Camilo, 1987).

The photosynthetically active radiation (PAR) absorbed by the canopy, F, at a particular canopy optical depth L, may be estimated by the product of the average projection of leaves in the direction u multiplied by the radiation intensity in the downwards direction (negative u), integrated over all angles. That is

$$F(L) = \frac{1}{2\pi} \int_0^{2\pi} d\phi \int_0^1 du\, G(u) (I(L, -u, \phi) + \exp(-GL/u_o)) \tag{40}$$

where G(u)= the average projection of the leaves in the zenith direction θ, where u = $\cos\theta$
I = Equation (35) evaluated for the value of ω appropriate to photosynthetically active radiation (PAR); this is ω~0.2.

With the following approximation for G (Goudriaan, 1977)

$$G(u) = \phi_o + \phi_1 u$$

the integral may be evaluated analytically. In the current program the absorbed PAR is modeled (in SUBROUTINE SUN) according to

$$F()_{par} = I_{NET}(0.193 + 0.44 \sin\alpha) \qquad (41)$$

Again the coefficients must be evaluated for each site according to its leaf area index, perhaps with the aid of Camillo's canopy radiation model.

The net long wave radiation absorbed by the canopy, RL_c, and soil, RL_s, are (Ross, 1981)

$$RL_c = (1 - T_r)(\sigma\varepsilon_a T_a^4 + \sigma\varepsilon_s T_s^4 - 2\sigma\varepsilon_c T_c^4) \qquad (42a)$$

$$RL_s = T_r \sigma \varepsilon_a T_a^4 - \sigma \varepsilon_s T_s^4 + (1 - T_r)\sigma\varepsilon_c T_c^4 \qquad (42b)$$

where T_a = the observed air temperature (K)
T_c = the canopy leaf temperature (K)
T_s = the soil surface temperature (K)
T_r = the canopy transmission coefficient for long wave radiation.

T_r is calculated according to

$$T_r = 2\int_0^1 u \, du \exp(-G(u) \, L_T/u \qquad (43)$$

where L_T = the total leaf area index
$G(u)$= the average projection of the leaves in the zenith direction θ, where $u = \cos\theta$

2. Latent and Sensible Heat Fluxes above the Soil Surface.

The environment above the soil surface is modelled in two alternative schemes according to whether the soil surface is considered to be directly coupled to the atmosphere or coupled in series through a plant canopy layer. In the one layer model, the soil and plant canopy surfaces are modelled as parallel pathways, both directly coupled to the atmosphere in which the meteorological observations are taken. The surface is partitioned into bare soil and canopy covered fractions. The sensible and latent heat fluxes from the two fractions are summed proportionately. In the two layer model, the soil surface interacts with the plant material of the overlying canopy and the canopy air. Canopy surfaces also interact with the canopy air. The canopy air, in turn, interacts with the atmosphere above the canopy. Modeling of the radiation flux and the resistances to latent and sensible heat flux differs somewhat between the one layer and two layer model. In both cases, the surface temperature, T_s, which is solved is that temperature common to the latent, sensible, and long wave radiation flux models which simultaneously satisfies the below ground heat and moisture fluxes and the energy fluxes to the atmosphere.

a. General models for heat fluxes above the soil surface

General models for the latent (L·E) and sensible (H) heat fluxes are

$$L{\cdot}E = -\frac{\rho c_p}{\gamma}\frac{e_s - e_a}{r_s + r_a} \tag{44a}$$

$$H = -\rho c_p \frac{T_s - T_a}{r_a} \tag{44b}$$

where the following definitions apply:

L = the latent heat of vaporization (586 cal gm^{-1} of water at 273 K)
E = the mass of water evaporated (cm/cm^2)
ρ = air density (1.15 X 10^{-3} gm cm^{-3})
c_p = specific heat of air (0.24 cal gm^{-1})
γ = psychrometric constant (0.66 mb K^{-1})
e_s = vapor pressure at the surface (mb)
e_a = vapor pressure of the air above the surface (mb)
T_s = temperature of the surface (K)
T_a = temperature of the air above the surface (K)
r_a = aerodynamic resistance (sec cm^{-1})
r_s = surface resistance (sec cm^{-1})

The surface vapor pressure is computed from

$$e_s = h\, e_{sat} \tag{45a}$$

where h is the relative humidity of the surface according to Equation (7a) for a bare soil surface or h = 1 for plant surfaces. The saturated vapor pressure at the surface, e_{sat}, is modelled as a function of the surface temperature

$$e_{sat} = \frac{\rho_{sat} T_s (T_s) R_{gas} T_s}{1000} \tag{45b}$$

where R_{gas} is the gas constant for water vapor and ρ_{sat} is the density of water vapor at saturation, Equation (7b).

b. Expressions for flux resistances

The variables r_s and r_a represent the resistance to the diffusion of water vapor from inside the evaporating surface to just outside the surface and then into the atmosphere, respectively.

The aerodynamic resistance to exchange between the surface and the atmospheric boundary layer above the surface, r_a, may be modelled from the similarity theory of the atmosphere diabatic boundary layer wind profile, (see for instance Brutsaert, 1984) as

$$r_a = \frac{\left[\ln\frac{z-d}{z_o} - \Psi_2\right]}{u_* k}$$

$$= \frac{\left[\ln\frac{z-d}{z_o} - \Psi_1\right]\left[\ln\frac{z-d}{z_o} - \Psi_2\right]}{k^2 u_z} \tag{46}$$

where z = the height of the meteorological data measurements
z_o = the roughness length of the surface
k = von Karman's constant (0.4)
d = the zero plane displacement
u_z = observed wind speed at height z (cm sec^{-1})
Ψ_1 = the stability correction for the wind speed profile
Ψ_2 = the stability correction for the temperature profile

The model may be run without employing a stability correction, implying nearly neutral atmospheric conditions (IAERO=0) in which case Ψ_1 and Ψ_2 are set to 0. When IAERO=1 the program calculates the stability corrections iteratively using Paulson's (1970) model. The three stability conditions recognized depend on the difference between the air and surface temperatures:

1. Neutral case, $|T_s - T_a| < 0.1$ K, $\Psi_1 = \Psi_2 = 0$.

For the other conditions the Monin-Obukhov length, $\mathcal{L}$, is calculated according to a profile formulation

$$\mathcal{L} = \frac{T_a u^2}{g(T_a - T_s)\ln\left(z - \frac{d}{z_o}\right)}\left(\frac{1 - \Psi_2}{1 - \Psi_1}\right)^2$$

2. Unstable case, $T_a - T_s < -0.1$ K

$$\Psi_1 = 2\ln\left(\frac{1+x}{2}\right) + \ln\left(\frac{1+x^2}{2}\right) - 2\tan^{-1}(x) + \frac{\pi}{2}$$

$$\Psi_2 = 2\ln\left(\frac{1+x^2}{2}\right)$$

where $x = (1 - 16\, z/\mathcal{L})^{0.25}$

3. Stable case, $T_a - T_s > -0.1$ K

a) if $z/\mathcal{L} \leq 1$, then $\Psi_1 = \Psi_2 = -5(z - z_o)/\mathcal{L}$
b) if $z/\mathcal{L} > 1$, then $\Psi_1 = \Psi_2 = -5(z - z_o)$

The surface to which the surface resistance r_s applies consists of two fractions: bare soil and plant canopy surfaces. Separate resistances are modeled for each.

No generally accepted model of bare soil resistance is known at this time, so the simulator employs a function of soil moisture (Camillo and Gurney, 1986).

$$r_{soil} = R_s(0) + R_s(1)\,(\theta_{sat}(1) - \theta(1)) + R_s(2)\,(\frac{\theta_{sat}(1)}{\theta(1)} - 1) \tag{47}$$

where $R_s(0)$, $R_s(1)$, and $R_s(2)$ = input constants (sec cm^{-1})
$\theta_{sat}(1)$ = saturation soil moisture content of top soil layer input by the user
$\theta(1)$ = the modelled soil moisture in the top soil layer at the time the resistance is calculated

The current default for the soil surface resistance in the program is 0.

The resistance coefficient for plant canopy surfaces, r_{can}, is modelled as a bulk stomatal resistance depending on the photosynthetically active radiation (PAR) absorbed by the canopy and three environmental stress factors: vapor pressure deficit, soil moisture deficit, and canopy temperature (Sellers 1986, Camillo, unpublished). (See for instance the discussion by Gates 1982).

Experimental data from cereals and grains support a leaf stomatal conductance (the reciprocal of resistance) which is a linear function of the photosynthetically active radiation (PAR) incident normal to the leaf (Choudhury and Monteith, 1986)

$$\frac{1}{r_{leaf}} = \frac{1}{a_1}(b_1 + F) \tag{48}$$

where F is the PAR and a_1 and b_1 are empirical constants. The total canopy stomatal conductance is the integral of the leaf conductances over the canopy:

$$\frac{1}{r_{st}} = \frac{1}{a_1}\int_o^{L_T} dL\,(b_1 + F(L))$$

$$r_{st} = \frac{a_1}{b_1 L_T + \int_0^{L_T} F(L)\,dL} \tag{49}$$

where the integral may be analytically evaluated (for individual cases).

The functional forms for the stress factors are not well known. They are included in the following way (Charles-Edwards and Ludwig, 1974; Sellers, 1989).

The stress factor due to vapor pressure deficit is given by

$$f(VPD) = 1 - VAPSTR * VPD \tag{50a}$$

where VAPSTR = a stress factor input by the user (Default = 0)

$VPD = e_{sat}(T_a) - e_{meas}$

$e_{sat}(T_a) = (R_{gas}\ T_a\ \rho_o) / 1000$

e_{sat} = saturation vapor pressure of canopy air

e_{meas} = vapor pressure according to meteorological observation

ρ_o = saturation vapor density from equation 7b using T_a

T_a = the air temperature

The stress factor due to dry soil conditions (high crown potential) is

$$f(\Psi_c) = 1 - \exp(\psi_{str}\ (\Psi_{cmin} - \Psi_c)) \tag{50b}$$

where Ψ_c = the crown potential computed during the simulation

Ψ_{cmin}= a user selected minimum crown potential

ψ_{str} = a user selected stress coefficient

The stress factor due to canopy temperature $f(T_c)$ is

$$FTMP = \left[\frac{T_{CENT}-(-1.0E4)}{30.0-(-1.0E4)} \times \frac{1.0E4-T_{CENT}}{1.0E4-(-1.0E4)}\right]^{TSTEXP} \tag{50c}$$

where

$$TSTEXP = \frac{TSTRH-TSTRO}{TSTRO-TSTRL} = \frac{1.0E4-30.0}{30.0-(-1.0E4)} = 0.994$$

and T_{CENT} is the canopy temperature in celsius.

These three stress factors are combined

$$f(T_c,VPD,\Psi_c) = f(VPD)\ f(\Psi_c)\ f(T_c); \qquad f(T_c,VPD,\Psi_c) \geq 0.001 \tag{50d}$$

The bulk canopy resistance, r_{can}, is then, from Equations (49) and (50c), calculated according to

$$r_{can} = \frac{a_1}{b_1 \times L_T + PAR1} \frac{1}{f(T_c, VPD, \psi_c)}$$

where $a_1 = 1.2 \times 10^{-2}$ (RCAN0)
$b_1 = 6 \times 10^{-4}$ (RCAN1)
PAR1 = the photosynthetically active radiation absorbed by the canopy (Equation (40)).

Thus, any increase in one of the three contributors to environmental stress will increase the canopy resistance. In the absence of accepted functions for the stress factors, they may all be set to zero, f = 1, and rely on fitting the canopy resistance coefficient a_1. If a_1 is not constant from day to day at a site, this may reflect need to provide estimated functions for the stress factors. (One might take advantage of the expectation that these stress factors vary with soil moisture, air temperature, and light conditions to estimate them from available field data.) In the current program, f() = 1.

c. Surface water contribution to the evaporative flux

The model allows for water loss by transpiration and by evaporation from the soil surface; evaporation from water accumulated on both plant and soil surfaces; and on dew formation. The accumulation of surface water from interception and dew formation is expressed as a fraction of a maximum possible depth of average accumulation for canopy, x_c, and soil, x_s, respectively, as

$$X_c = h_c/h_{cmax} \qquad \textbf{(51a)}$$
$$X_s = h_s/h_{smax} \qquad \textbf{(51b)}$$

where h_c and h_s are the respective depths of accumulation.

These are obtained as solutions to linear, first order differential equations (Equations (52a) and (52b)) (Massman, 1983).

$$\frac{dh_c}{dt} = (1-p)R_o - [D_o + d_o R_o] X_c + E_c' \qquad \textbf{(52a)}$$

$$\frac{dh_s}{dt} = pR_o - [D_o + d_o R_o] X_c + E_s' - KS \qquad \textbf{(52b)}$$

In Equation (52a) and (52b) the first term is the rate at which the rain is intercepted by the corresponding surface, the second term is the rate at which intercepted water drips off the canopy, and the third term is the evaporation or dew formation rate. R_o is the rain rate, pis the fraction of rain which falls directly through the canopy to the soil surface, and D_o and d_o are empirical constants taken to be 0.12 mm/hr and 0.3 respectively. The fraction p is modelled as $p = \exp(-G(1)/L_T)$ (Sellers et al., 1986), where G(1) is the average projection of the leaves in the direction of the incoming rain. KS is the saturated conductivity of the soil.

Loss of standing water from the canopy and soil surfaces is then modelled by

$$E_c' = -\frac{\rho c_p}{\gamma} \frac{e_l - e_{ca}}{r_{ca}} \cdot \begin{matrix} 1, & e_l, e_{ca} \ (dew) \\ X_c, & e_l \leq e_{ca} \ (transpiration) \end{matrix} \tag{53a}$$

$$E_s' = -KS - \frac{\rho c_p}{\gamma} \frac{e_s - e_{ca}}{r_{sa}} \cdot \begin{matrix} 1, & e_s, e_{ca} \ (dew) \\ X_s, & e_s \leq e_{ca} \ (transpiration) \end{matrix} \tag{53b}$$

where r_{ca} = the aerodynamic resistance over the canopy
r_{sa} = the aerodynamic resistance over bare soil

The surface water heights are updated by

$$h_s, c(t + \Delta t) = h_{s,c}(t) + \Delta t \frac{dh_{s,c}(t)}{dt} \tag{54}$$

where Δt is the integrator step size. The heights are constrained to be between zero and their maximum allowed values.

d. The one layer model

In the one layer model the latent and sensible heat fluxes are computed seperately for the bare soil and canopy. The computation for soil and for canopy is each weighted according to the fraction of the surface which is bare soil and plant covered. The solved surface temperature, T_s, common to canopy and soil surfaces, satisfies all the energy budget terms at the soil-air interface. The fluxes are then computed using T_s. The heat fluxes are given by

$$H_s = -\rho c_p f_s \frac{T_s - T_a}{r_{sa}} \tag{55a}$$

$$H_c = -\rho c_p (1 - f_s) \frac{T_s - T_a}{r_{ca}} \tag{55b}$$

When there is no standing water on soil or plant surfaces the latent heat fluxes are given by

$$L \cdot E_s = -\frac{\rho c_p}{\gamma} f_s (1 - f_{sw}) \frac{e_s - e_a}{r_{sa} + r_{soil}} \tag{55c}$$

$$L \cdot E_c = -\frac{\rho c_p}{\gamma} (1 - f_s)(1 - f_{pw}) \frac{e_c^* - e_a}{r_{ca} + r_{can}} \tag{55d}$$

where f_s = the fraction of the surface that is bare soil
$1-f_s$ = the fraction of the surface that is covered by vegetation
f_{sw} = the fraction of maximum surface water on the soil

f_{cw} = the fraction of maximum surface water on the canopy
$1-f_{sw}$ = the fraction of dry soil surface
$1-f_{cw}$ = the fraction of dry plant surface.
e_c^* = canopy vapor pressure, saturated at canopy temperature

When there is standing water on soil or plant surfaces the latent heat fluxes are given by

$$L{\cdot}E_s = -\frac{\rho\, c_p}{\gamma} f_s (1-f_{sw}) \frac{e_s{*}-e_a}{r_{sa}} \tag{55e}$$

$$L{\cdot}E_c = -\frac{\rho\, c_p}{\gamma} (1-f_s)(1-f_{pw}) \frac{e_c^* - e_a}{r_{ca}} \tag{55f}$$

where e_s^* = soil surface vapor pressure, saturated at T_s
e_c^* = canopy surface vapor pressure, saturated at T_s

e. The two layer model

For the two layer model the situation is more complex. The flux network consists of transfers between soil surface and canopy air; leaf surfaces and canopy air; and canopy air and the above canopy air. The resistance network is comprised, as in the one layer model, of the soil surface resistance, r_{soil}, and soil aerodynamic resistance, r_{sa}; bulk canopy resistance, r_{can}, and canopy aerodynamic resistance, r_{ca}. However, there is the additional aerodynamic resistance between canopy air and the outside air, r_a. The two aerodynamic resistances r_{sa} and r_{ca} depend on the wind speed profile within the canopy. The wind speed and turbulent diffusion coefficient profiles are approximated from their values at the top of the canopy, (Choudhry and Monteith, 1988) which are estimated according to

$$u_h = \frac{u_{obs}\ \ln\frac{h-d}{z_o}}{\ln\frac{z-d}{z_o}} \tag{56a}$$

$$K_h = u_{obs} \frac{k^2(h-d)}{\ln\frac{z-d}{z_o}} \tag{56b}$$

where u_h = the wind speed at the top of the canopy
u_{obs} =the observed wind speed
h = the height of the canopy
d = the zero plane displacement of the canopy
z = the height at which the wind speed is observed
z_o = the roughness length of the canopy
k = von Karman's constant (0.4)

The aerodynamic resistance between canopy leaf surfaces and canopy air, r_{ca}, is

$$r_{ca} = \frac{\frac{\alpha}{2}\frac{a}{L_T\sqrt{u_h}}}{1-\exp(-\frac{\alpha}{2})} \tag{57}$$

where α = an attenuation coefficient (default = 3.0)
a = a constant input by user $(sec/cm)^{½}$ (default = 3.0)
L_T= the (total) canopy leaf area index

The aerodynamic resistance between soil surface and canopy air, r_{sa}, is

$$r_{sa} = \frac{h\exp(\alpha')}{\alpha' K_h}\left[\exp\left(\frac{-\alpha' z_o'}{h}\right) - \exp\left(\frac{-\alpha'(z_o+d)}{h}\right)\right] \tag{58}$$

where α' = a diffusion damping coefficient (= 3.0)
z_o' = the roughness length of the soil.

The canopy air temperature, T_{ca}, is computed iteratively from equations (60a), (60b), and (60c) letting

$$H_c + H_s - H = 0$$

during the integration interval. Solving for T_{ca} gives

$$T_{ca} = \frac{r_{ca}r_{sa}T_a + r_a r_{sa}T_c + r_a r_{ca}T_s}{r_{ca}r_{sa} + r_a r_{sa} + r_a r_{sa}} \tag{59}$$

A similar procedure using equations (60d), (60e), and (60f) letting $L{\cdot}E_c + L{\cdot}E_s - L{\cdot}E = 0$ permits computing the vapor pressure in the canopy air space. The procedure weights the resistances in r_1 and r_2 according to the fraction of standing water on soil and canopy surfaces. The soil resistance r_{soil} and bulk canopy resistance r_{can} are obtained according to Equations (47) and (50e) respectively.

The fluxes into the canopy air space and from the canopy air into the air above are then computed according to equations (60a through 60f)

$$H = \frac{\rho\, c_p (T_{ca} - T_a)}{r_a} \tag{60a}$$

$$H_c = \frac{\rho c_p (T_c - T_{ca})}{r_{ca}} \tag{60b}$$

$$H_s = \frac{\rho c_p (T_s - T_{ca})}{r_{sa}} \tag{60c}$$

$$L{\cdot}E = -\left(\frac{\rho c_p}{\gamma}\right)\frac{(e_{ca} - e_a)}{r_a} \tag{60d}$$

$$L{\cdot}E_c = -\left(\frac{\rho c_p}{\gamma}\right)\frac{(e_c^* - e_{ca})}{r_2} \tag{60e}$$

$$L{\cdot}E_s = -\left(\frac{\rho c_p}{\gamma}\right)\frac{(e_s - e_{ca})}{r_1} \tag{60f}$$

where the subscripts c, ca, a, and s refer respectively to the canopy, the canopy air space, the air above the canopy, and the soil surface. The subscript r_1 and r_2 are defined

$r_1 = r_{sa}$	for wet soil surfaces
$r_1 = r_{sa} + r_{soil}$	for dry soil surfaces
$r_2 = r_{ca}$	for wet canopy surfaces
$r_2 = r_{ca} + r_{can}$	for dry canopy surfaces.

The latent heat fluxes from wet and dry surfaces are added in proportion to their respective fractions in a manner analagous to that used in the one layer model.

C. Solution of the surface energy budget

To solve the heat balance equation for T_s, one starts by rewriting Equation (33) as (Hillel, 1977)

$$T_s = \frac{Z_1}{\lambda_1} S + T_1 \tag{61}$$

Inserting the right hand side of the heat balance equation (32) for S gives

$$T_s = T_1 + \frac{Z_1}{\lambda_1}\{R(T_s) + L{\cdot}E(T_s) + H(T_s)\} \tag{62}$$

The dependence of all three modeled flux terms on the same surface temperature, T_s, is thus established. This equation, having units of temperature on both sides, is of the form

$$T_s = F(T_s)$$

and can be solved by the method of successive approximations. A trial value for T_s is chosen, $F(T_s)$ is evaluated, and a new value for T_s results. This procedure can be repeated until satisfactory convergence is obtained. In the simulator the air temperature T_a is used as a start value, and a maximum of four (default) iterations are allowed. Tests have shown that the process converges after one or two iterations. Convergence is defined as the absolute value of the change in T_s between iterations being less than 0.1 K. The process will always converge if the magnitude of the derivative of F is less than 1,

$$\left|\frac{dF}{dT_s}\right| < 1$$

This derivative is a complicated function of changing meteorological variables, so an analytical study of the conditions required for convergence is not feasible. However, from Equation (62) it is clear that this derivative is proportional to Z_1, the depth to the center of the first soil layer.

Periods of rainfall can be modelled The user supplies the number of rain storms, the start and stop times of the rain (t_o and t_1), and the total accumulation (r_{tot}) for each one. A constant rate throughout each time interval is assumed and calculated as

$$r = \frac{r_{tot}}{r_1 - r_o} \tag{63}$$

The short wave radiation attenuation factor during each rain storm must also be supplied. This number is equivalent to the cloud attenuation factor B in equation (34b).

During periods of rain, the evapotranspiration and sensible heat fluxes are set equal to zero (LE and H, Equation (32)), and the soil wetness flux at the surface q_θ, is set equal to the rain rate, Equation (63).

It is possible to remove all temperature dependence from the simulation. (See the description of the NAMELIST parameter ITEMPS). In this case the temperature profile, soil heat fluxes, and atmospheric heat fluxes are not modelled. However, the evapotranspiration flux must still be estimated to provide the surface moisture boundary condition. To do this a Gaussian function of time is supplied,

$$E(t) = -E_{max} \exp[-k(t - t_{max})^2] \tag{64}$$

where t_{max} is the time of maximum demand and E_{max} is the rate at this time. The variable k which determines the width of the Gaussian can be related to the integrated daily rate E_{day} as follows:

$$\begin{aligned} E_{day} &= \int dt \, |E(t)| \sim E_{max} \int_{-\infty}^{\infty} dt \exp[-k(t - t_{max})^2] \\ &= E_{max} \sqrt{\frac{\pi}{k}} \end{aligned} \tag{65}$$

This gives

$$k = \pi \left(\frac{E_{max}}{E_{day}} \right)^2 \tag{66}$$

The user supplies t_{max}, E_{max}, and E_{day}. The simulator computes k from Equation (66), and then Equation (65) is used to model the evapotranspiration flux.

For some simulations it may be simpler to specify the integrated daily total and the width of the Gaussian. The maximum rate E_{max} can be computed from these two. The exponential slope k equals $1/t_e^2$, where t_e is the time interval between the maximum rate and the time when the rate falls to 1/e of this value. Setting Equation (66) equal to $1/t_e^2$ and solving for E_{max} gives

$$E_{max} = \frac{1}{\sqrt{\pi}} \frac{E_{day}}{t_e}$$

Therefore the user can compute the required input E_{max} from E_{day} and t_e.

It is also possible to model the surface temperature and the heat balance equation (and thereby the effect of the meteorological variables on evapotranspiration) without modeling the soil temperature profile. The surface temperature T_s and average subsurface temperature T are modelled by the force restore method (Lin, 1980). The mathematical model is

$$\frac{dT_s}{dt} = \frac{2S}{a} - \frac{2\pi}{\tau}(T_s - T) \tag{67a}$$

$$\frac{dT}{dt} = \frac{S}{a\sqrt{365\pi}} \tag{67b}$$

where S is the heat flux absorbed by the soil and

$$a \equiv \sqrt{\frac{\lambda c \tau}{\pi}}$$

In this expression λ is the thermal conductivity and c is the heat capacity of the soil surface layer and τ is the number of seconds in a day. The thermal conductivity and heat capacity are computed using Equations (23) and (29).

Since Equation (67) gives the time dependence of the surface temperature explicitly, T_s is made one element of the state vector and is therefore known. Therefore, no iteration is required to solve the heat balance equation. The terms R, LE, and H are evaluated using the state vector value for T_s, and S is computed from Equation (32).

D. Root Model

A model of soil water depletion by plant roots has been included as an extra term in the equation of continuity, Equation (31a)

$$\frac{d\theta}{dt} = -\frac{dq}{dz} - Q(z,t) \tag{68}$$

The sink term Q (1/sec) in Equation (68) is positive when water flows from the soil to the plant. The mathematical model is (Hillel, 1977; Gardner, 1964)

$$Q(z,t) = \frac{\Phi_s(\theta,z) - \Phi_p(t)}{\Omega_s + \Omega_p} \tag{69}$$

where $\Phi_p(t)$ = the crown potential (cm)
$\Phi_s(\theta,z)$ = the total potential energy of the soil water (cm)

$$\Phi_s(\theta,z) = \psi(\theta) - z \tag{70}$$

where ψ = the matric potential
$-z$ = gravitational potential (z is depth below the surface and is positive).

The plant potential Φ_p (cm) in Equation (69) varies with time but is assigned the same value throughout the root system. The soil resistance Φ_s (cm-sec) is inversely proportional to the soil conductivity and the quantity of active roots

$$\Omega_s = \frac{1}{B\,K(\theta)\,P(z)} \tag{71}$$

where B = constant
$K(\theta)$ = hydraulic conductivity (cm/sec)
$P(z)$ = relative root density at depth z (cm/cm^3)

The resistance to flow in the roots

$$\Omega_p(z) = r/P(z) \tag{72}$$

where r = specific resistance to flow in the roots (sec/cm).

Using Equations (71) and (72) for the resistances and Equation (70) for the soil water potential energy in Equation (69), after rearranging, gives

$$Q = \frac{BKP[\psi - z - \Phi_p]}{1 + BKr}$$

$$Q = \frac{BKP[\psi - z - \Phi_p]}{1 + \frac{\Omega_p}{\Omega_s}} \tag{73}$$

The important model parameters are the relative root density P(z) and the ratio of the resistances Ω_p/Ω_s. No loss of generality results from setting B=1, since its value can be absorbed into the definitions of P and r. Since Q is proportional to P, multiplying P at all depths by a constant would only change the rate at which the moisture profiles evolve. Since P has the dimensions of $1/cm^2$, it is commonly thought of as the length of active roots per volume of soil. As yet there is no experimental evidence that this is true; the model only requires that P(z) represent the relative ability of the roots to absorb water at each depth. The plant potential Φ_p, commonly referred to as the crown potential, is modelled as a response to an atmospheric evapotranspiration demand function.

The discrete model of the sink term as used in this simulator is

$$Q_j = \frac{K_j P_j[\psi_j - z_j - \Phi_p]}{1 + rK_j} \tag{74}$$

Q_j is the value of the sink in the j^{th} soil layer, and z_j is the depth to the center of this layer. K_j and ψ_j are the hydraulic conductivity and matric potential of the soil water in the layer. The relative root density in each layer P_j and the specific resistance of the plant roots r are input parameters. The crown potential $T_p(t)$ is modeled as a response to a known transpiration demand function $E_{pl}(t)$. The crown potential is computed by requiring that the integral of the sink terms over the soil profile be equal to E_{pl}. In its discrete form, this integral is

$$E_{pl} = \sum_{j=1}^{N} Q_j dz_j \tag{75}$$

where dz_j is the thickness of the j^{th} layer and N is the number of layers in the profile model. Using Equation (60) for Q_j and solving for the crown potential gives

$$\Phi_p(t) = \frac{E_{pl}(t) + \sum_{j=1}^{N} K_j P_j (\psi_j - z_j) dz_j}{\sum_{j=1}^{N} K_j P_j dz_j} \tag{76}$$

Both E_{pl} and ψ_j are negative, so T_p is also negative. Its magnitude can be large if either the demand is large, the soil is dry (so $|\psi_j|$ is large), or both. The magnitude of Φ_p must be less than the magnitude of the wilting point Φ_w, which is the largest potential for water intake the plant can create before wilting. Thus the crown potential must satisfy the inequality

$$\Phi_w \leq \Phi_p \leq 0$$

If $\Phi_p < \Phi_w$ the simulator will set $\Phi_p = \Phi_w$. Once Φ_p is calculated the sink term can be evaluated for each layer using Equation (74). It must be positive for all layers, to correspond to flow from soil to roots. Any of the Q_j which are negative are set equal to zero. This procedure is used to accommodate experimental evidence that water flow from plant roots to the soil is negligible (reference 8).

The transpiration demand E_{pl} is computed from equation (60e).

D. Soil Hydraulic Properties

Both matric potential and hydraulic conductivity as functions of volumetric wetness may be modelled for a wide range of soil types and textures following Clapp and Hornberger, 1978

$$K(\theta) = K_s \left(\frac{\theta}{\theta_s} \right)^{2b+3} \tag{77a}$$

$$\psi(\theta) = \psi_s \left(\frac{\theta}{\theta_s} \right)^{-b} \tag{77b}$$

where θ_s is the volumetric wetness at saturation, K_s and ψ_s are the conductivity and matric potential respectively at saturation, and b is a parameter determined primarily by the soil texture. Representative values are 4 for sand to 11 for clay.

Instead of using this model, the user may choose to provide tables of hydraulic conductivity and matric potential as functions of soil moisture. The simulator will perform a linear interpolation within the table (or linear extrapolation for soil moisture values outside the range of those supplied). An example of such a table is presented in the Appendix. The user specifies the choice of model or table look-up via the NAMELIST variable MODHYD.

2. METHOD OF SOLUTION (To be added later, unchanged from the original)

3. PROGRAM SOILSIM: NAMELIST INPUT AND SELECTED OTHER VARIABLES

(University of Delaware version)
Variables Definition
2 May 1990

Equation numbers refer to the attached SOILSIM Program Description.

This is the main system input. The NAMELIST name is INPUT, and it is read on unit-5 Subscripts run from 1 to NL (number of soil layers) unless otherwise indicated. Variables indicated with (*) have been removed from the University of Delaware version and those indicated by (**) have been added.

I. INTEGRATOR CONTROL VARIABLES

NAME	TYPE	DEFAULT	DESCRIPTION
TSTOP	R*8	8.64D4	Duration of simulation (seconds).
TNOON	R*4	4.32D4	(*)Seconds from integration start to noon - times before noon positive, afternoon negative.
HMAX	R*8	1.8D3	Maximum integration step size if IFORCE=1
IFORCE	I*4	1	Force integration step size to remain less than HMAX (0=no, 1=yes)
H	R*8	1.0D0	Initial step size (seconds) if IFORCE=1; otherwise H is set to HMAX/512).
WATERR	R*4	1.0E-3	Moisture error tolerance (E_i, Eq. 103)
TEMERR	R*4	1.0	Temperature error tolerance (E_{i+NL}, Eq. 103)
ED	R*4	5.0	Error window parameter (Eq. 103)
ITEMPS	I*4	0	Select soil temperature model: 0 = no temperature 1 = model soil temperature profile 2 = use force restore method
JBOT	I*4	1	Bottom boundary condition for moisture 0 = 0 flux 1 = constant soil moisture in bottom layer 2 = downward flux equal to hydraulic conductivity. 3 = constant flux set by user
QBOT	R*4	0.0	Moisture flux at bottom of profile (cm/sec) if JBOT=3

The following variables are for establishing the simulation noon (TNOON) in terms of local mean clock time and for calculating the declanation of the sun.

XLONG	R*4	0.0	(**)Longitude of site in degrees (west longitude is negative)
XLAT	R*4	0.0	(**)Latitude of the site in degrees (north latitude is positive)
DAYNUM	R*4	0.0	(**)Day of the year ($1 \leq$ DAYNUM ≤ 364 or 365) (Eq. 34e)
IDST	I*4	0	(**)Clock time is Standard time or Daylight saving time (Eq. 34f) (0 = standard time, 1 = daylight saving time)
TC1 1987)	R*4	4.02	(**)First constant in equation of time (degrees) (default values for year (Eq. 34f)
TC2 See	R*4	17.41	(**)Second constant in equation of time (degrees) (default for year 1987; Almanac for Computers, U. S. Naval Observatory for current values) (Eq. 34f)

II. OUTPUT CONTROL VARIABLES

NAME	TYPE	DEFAULT	DESCRIPTION
DTOUT	R*4	1800.0	Output period (seconds)
ITABLE	I*4	1	Indicator for amount of print output: 0 = none 1 = NAMELIST only 2 = NAMELIST and boundary conditions 3 = NAMELIST, boundary, soil conditions.
NFUNCT	I*4	0	(*)Number of moisture and temperature profiles per plot page.
NWATRS	I*4	0	(*)Number of soil layers for which moisture is to be plotted as a function of time.
INDXW	I*4	10*0 (I,I=1,10)	(*)Indices of soil layers for moisture versus time plots
ITRANS	I*4	0	(*)Plot transpiration flux (0=no, 1=yes)
NWFLUX versus	I*4	0	(*)Number of soil boundaries for which moisture flux is to be plotted time.
INDXWF	I*4	10*0 (I,I=1,10)	(*)Indices of soil boundaries for moisture flux versus time plots.
NTEMPS	I*4	0	(*)Number of soil layers for which temperature is to be plotted versus time.
INDXT	I*4	10*0	(*)Indices of soil layers for temperature versus time plots.
ISRFT	I*4	0	(*)Plot surface temperature versus time (0 = no, 1 = yes).
NTFLUX	I*4	0	(*)Index giving number of graphs of heat flux versus time
INDXTF	I*4	10*0 (I,I=1,10)	(*)Indices of soil layers for which soil heat flux is to be plotted versus time.
IHBAL	I*4	0	(*)Plot components of surface energy balance? (0=no, 1=yes)
IPR	I*4	10	(*)Output unit for printer plots. (Tables are printed on unit 6).
IUPRT	I*4	7	(**)FORTRAN unit number for formatted output
BASENAME	CHARACTER*128		(**)Pathname and basefile name for input and output files. Extensions added to distinguish the files. Input files are ".in", ".met", ".hyd". Output files are ".out" for meteorological output; ".prt" for formatted output. Additional extensions may be used to identify the file type, ie ".ps" for postscript.

The following 10 parameters are lower and upper limits for the axes of the printer plots. If a set of lower and upper limits are equal, the actual limits will be determined from the data plotted.

WL, WH	I*4	0.0,0.0	(*)Limits for moisture plots.
WFL,WFH	I*4	0.0,0.0	(*)Limits for moisture flux plots.
TL, TH	I*4	0.0,0.0	(*)Limits for temperature plots.
TFL, TFH	I*4	0.0,0.0	(*)Limits for heat flux plots.
HBL, HBH	I*4	0.0,0,0	(*)Limits for heat balance Eq. plots.

Indices to control accumulation of moisture and heat in soil at boundaries

NWCUMS	I*4	0	Number of soil boundaries for which cumulative soil moisture fluxes are to be computed. (<=10).
IXWCUM	I*4	10*0 (I,I=1,10)	Indices of boundaries for cumulative moisture fluxes
NHCUMS	I*4	0	Number of soil boundaries for which cumulative heat fluxes are to be computed (≤10)
IXHCUM	I*4	10*0 (I,I=1,10)	Indices of boundaries for cumulative heat fluxes
NDISK	I*4	0	Indicator if moisture and temperature profiles are to be output to disk (0=no,1=yes).
IUDISK	I*4	12	Unit number for profile output

(*) Indicates variable added to the unix version.

III. Variables for Defining the Soil Profile

NL	I*4	200	Number of soil layers ($2 \leq NL \leq 200$)
DZ(I)	R*4	200*1.0	Thickness of soil layers (cm)
WATER(I)	R*4	200*0.25	Initial volumetric moisture of soil layers.
TFORCE	R*4	2*293	Initial force restore soil temperatures (Eq. 67)
TEMPS(I)	R*4	200*293.0	Initial temperature of soil layers (degrees K).
MODHYD	I*4	1	Source for soil hydraulic functions: 0 = table look-up 1 = Clapp and Hornberger model
IHDUNT	I*4	9	Unit number for look-up tables.
SATW(I)	R*4	200*0.3	Saturation moisture (θ_s, Eq. 77)
SATP(I)	R*4	200*-10.0	Saturation matric potential (cm) (ψ_s, Eq. 77b)
SATK(I)	R*4	200*1E-4	Saturation hydraulic conductivity (cm/sec) (K_s, Eq. 77a)
EB(I)	R*4	200*5.0	Soil texture parameter (b, Eq. 77)
PORSTY(I)	R*4	200*0.45	Soil porosity (ε, Eq. 5, 15)
TCONDS(I)	R*4	200*2.5E-3	Thermal conductivity of soil solids (cal cm^{-1} sec^{-1} K^{-1}) (λ_s, Eq. 23)
VHCAPS	R*4	200*0.5	Volumetric heat capacity of soil solids (cal cm^{-1} sec^{-1} K^{-1}) (C_s, Eq. 29)
SHAPE(I)	R*4	200*0.33	Shape factor (g_s, Eq. 24)
FACTKA	R*4	1.4	Weight factor for air (k_a, Eq. 23)
TCONDW	R*4	1.3E-3	Thermal conductivity of water (cal cm^{-1} sec^{-1} K^{-1}) (λ_w, Eq. 24)
TCONDA	R*4	5.967E-5	Thermal conductivity of air (cal cm^{-1} sec^{-1} K^{-1}) (λ_a, Eq. 24)
VHCAPW	R*4	1.0	Volumetric heat capacity of water (cal cm^{-3} K^{-1}) (C_w, Eq. 29)
VHCAPA	R*4	3.0E-4	Volumetric heat capacity of air (cal cm^{-3} K^{-1}) (C_a, Eq. 29)
ALPHA	R*4	0.667	Tortuosity factor (α, Eq. 6a, 15)

GNU	R*4	1.0E0	Diffusivity constant (Eq. 15b, 15c) (Looks like a fudge factor)
GAMMA0	R*8	2.09D-3	Surface tension parameter (1/C) (for γ in Eq. 11)
GAMMA1	R*8	0.0D0	Surface tension parameter for γ in Eq. 11)
IROOTS	I*4	0	Use root model (0=no, 1=yes)
ROOTS(I)	R*8	200*0.0D0	Root density profile (1/cm**2) (P(z), Eq. 71 ,72)
SPRES	R*8	1.0D6	Root specific resistance (sec/cm) (r, Eq. 72)

IV. Variables at the Air/Soil Interface

Name	Type	Default	Description
MUNIT	I*4	8	Unit number for meteorological data set
ICAN	I*4		Index for one layer or two layer model (0=1 layer, 1=2 layer model) (In the two layer model the soil surface is coupled to atmosphere serially through the canopy air space. Otherwise, both canopy and soil surface are coupled in parallel to the atmosphere).
XLAT	R*4	45.0	Latitude (degrees) (ϕ, Eq. 34d)
ZHGHT	R*4		Height of meteorological data measurements (cm) (z, Eq. 46)
IAERO	I*4	0	Include stability corrections in aerodynamic resistance model. (0=no, 1=yes)
MAXAER	I*4	1	Max iterations in AERO calculation
RAERR	R*4		Error criterion for iteration in AERO
DTNEUT	R*2	0.2	Maximum Ts-Ta for neutral conditions in AERO
NITERS	I*4	4	Maximum number of iterations for HBE solution of Ts
TMPITR	R*4	0.1	Temperature error criterion in HBE solution
ITRANS	I*4	0	Index to plot transpiration rate (0=no, 1=yes)
ML	I*4	80	This variable has not been used to compute ... looks like page size
MC	I*4	132	This variable has not been used to compute ... looks like page size
THMIN	R*4	0.05	Minimum surface soil moisture which will support evaporation.
SFRAC	R*8	0.1D0	Fraction of ET demand allocated to bare soil evaporation (f, Eq. 55)
EMAX	R*8	3.0D-5	Maximum rate for evaporation model (cm/sec) (E_{max}, Eq. 64, 65, 66).
EMAXT	R*8	4.68D4	Time of maximum rate (seconds since start of

			simulation day (t_{max}, Eq. 65).
EDAY	R*4	1.0	Daily evaporation (cm) (E_{day}, Eq. 65).
NRAINS	I*4	0	Number of rain storms ($\leq$ 10).
RNSTRT(I)	R*4	10*0.0	Start times of rain storms (seconds since simulation start) (I=1,10) (t_o, Eq. 63)
RNSTOP(I)	R*4	10*0.0	Stop times of rain storms (seconds since simulation start) (t_1, Eq. 63)(I=1,10)
RNTOT(I)	R*4	10*0.0 (I=1,10)	Total accumulation for each storm (cm) (r_{tot}, Eq. 63)
HSL0	R*4	0.0	Initial soil dew interception (cm)
HPL0	R*4	0.0	Initial plant dew interception (cm)
HSLMAX	R*4	0.0	Maximum soil dew interception (cm) (Eq. 51)
HPLMAX	R*4	0.0	Maximum plant dew interception (cm) (Eq. 51)
ATTEN(I)	R*4	10*0.5 (I=1,10)	Short wave attenuation factor for each storm (Eq. 34c)
TURB	R*4	2.0	Turbidity factor (n, Eq. 34c).
SUNDEC	R*4	0.0	Sun declination (degrees) (δ, Eq. 34d)
EMLONG	R*4	0.96	Long wave emissivity (ε_s, Eq. 34e).
RHOVP0	R*4	6.0035	Soil water vapor density coefficient (R_o, Eq. 7b)
RHOVPT	R*4	4975.9	Soil water vapor density coefficient (R_1, Eq. 7b)
DATM0	R*4	0.229	Diffusion coefficient of water vapor at 0 degrees C (cm_2 sec_{-1}) (D_o, Eq. 6b)
DATMT	R*4	0.055E0	Vapor diffusion model constant (appears unused at present)
LHEAT	R*8	586.0D0	Latent heat of vaporization of water (cal/gm)
EMISS0	R*4	0.74	Offset constant for emissivity of atmosphere. (Eq. 34g)
EMISS1	R*4	0.005	Slope constant for emissivity of atmosphere. (Eq. 34g)
PSYCHR	R*4	0.66	Psychrometric constant (mb/K) (γ, Eq. 53, 55)
DAIR0	R*4	1.29E-3	Density of air at 0 C used to compute σ c_p
DAIR1	R*4	0.0045E-3	Air density model constant (not used in current program)

CP	R*4	0.24	Specific heat of air (c_p, Eq. 53, 55)
RS0	R*4	0.0	Soil surface resistance offset (sec/cm) (r_0, Eq 47)
RS1	R*4	0.0	Soil surface resistance slope (r_1, Eq. 47)
RS2	R*4	0.0	Coefficient of 1/theta in soil surface resistance (sec/cm) (Eq. 47)
VON	R*4	0.4	Von Karman's constant (k, Eq. 46)
Z0SOIL	R*4	0.2	Soil roughness length (cm) (z_o', Eq. 58)
Z0CAN	R*4	25.0	Canopy roughness length (cm) (z_o, Eq. 46, 56, 58)
DISP	R*4	75.0	Zero plane displacement (cm) (d, Eq. 46, 56, 58)

V. Parameters to describe the canopy

RCAN0	R*4	1.2E-2	Stomatal resistance constant (a_1, Eq. 49, 50e)
RCAN1	R*4	6.0E-4	Stomatal resistance constant (b_1, Eq. 49, 50e)
RCA0	R*4	3.0	RCA resistance constant ($(\text{sec cm}^{-1})^{1/2}$) (a, Eq. 57)
WNDDMP	R*4	3.0	Wind damping coefficient in canopy model (α, Eq. 57)
DIFDMP	R*4	3.0	Diffusivity damping coefficient (α', Eq. 58)
CPMIN	R*8	-1.5D4	Limiting value of crown potential (cm) (Ψ_{cmin}, Eq. 50b)
PSISTR	R*4	1.0E-4	Crown potential stress coefficient (Ψ_{str}, Eq. 50b)
VAPSTR	R*4	0.0	Vapor pressure deficit stress coefficient (1/MB) (V_{str}, Eq. 50a)
RSTOM0	R*4	0.0	(not currently used)
RSTOM1	R*4	0.0	(not currently used)
TSTR0	R*4	30	Temperature stress coefficient (Eq. 50c)
TSTRL	R*4	-1.0E4	Temperature stress coefficient (Eq. 50c)
TSTRH	R*4	1.0E4	Temperature stress coefficient (Eq. 50c)
CD	R*4	0.2	(not currently used)
HCAN	R*4	100.0	Canopy height (cm) (h, Eq. 56, 58)
RLAI	R*4	2.0	Leaf Area Index (L, Eq. 49, 57)

EMCAN	R*4	1.00	Canopy emissivity (ε_c, Eq. 42a)

VI. Parameters for radiation modeling

IRNET	I*4	3	Index for computing radiation balance. (see METEOR and SUN, HBAL, PET) (This needs more work) 0= no value supplied; program will model incident 1= net short wave supplied, model computes net long wave. 2= net long wave radiation supplied 3= net all-wave radiation supplied, 4= ???incoming solar supplied, atmospheric long wave modelled 5= ??? model computes atmospheric long wave.
MODSOL	I*4	0	Index whether measured or modeled incident solar will be used for radiation absorbed by canopy. (0=measured, 1=modeled)
SOIABS	R*8	-0.039D0, .13D0,2*0.0D0	Coefficients for solar radiation absorbed by soil (Eq. 39a)
CANABS	R*8	0.818D0, -0.048D0, 2*0.0D0	Coefficients for solar radiation absorbed by canopy (Eq. 39b)
PARINC	R*8	0.193D0, 0.44D0,2*0.0	Coefficients for PAR absorbed by canopy (Eq. 41)
RLTRA	R*4	0.0	Long wave transmissivity in canopy (T_t, Eq.42)
CALB	R*8	.220D0, -0.08D0,2*0.0D0	Canopy albedo coefficients (Eq. 34f)
ALB	R*4	0.3	Surface short wave albedo (Eq.34f)
CTRANS	R*4	0.2	Cloud transmissivity to short wave (not used in current program)
CLOUDS	R*4	0.0	Fractional cloud cover (? B, Eq. 34b) (not used in current program)
SHORT0	R*4	0.033333	Solar constant in incoming short wave model coefficient (cal cm^{-2} sec^{-1})
SCAT0	R*4	0.128	Short wave model constant (Eq. 34c)
SCAT1	R*4	0.054	Short wave model constant (Eq. 34c)
PLFRAC	R*4	1.0	Canopy interception fraction ((1-p), Eq. 52)

METEOROLOGICAL DATA INPUT

The meteorological data needed to drive the simulations are input in NAMELIST format. The NAMELIST name is MET, and the data set is on the unit specified by the parameter MUNIT in the main NAMELIST input data set INPUT.

Each data value (air temperature, vapor pressure, wind speed, and optionally radiation) is used throughout a user-specified time interval. Using the parameter IREUSE, one may also specify that the first 24 hours of data be reused for succeeding 24 hour periods.

Following is a description of the data elements in MET:

NAME	TYPE	DEFAULT	DESCRIPTION
NMETS	I*4	0	Number of values in each data array (≤ 1000)
DELTA	R*4	3600.0	Time interval (seconds) for which each data value is used.
IREUSE	I*4	0	Reuse first 24 hours of data for succeeding days (0=no, 1=yes)
IROUT	I*4	0	Indicator showing type of radiation data on this data set to be used by simulator. (Called IRCODE in METEOR and IRNET in CANOPY; PET; and SUN). Components not supplied with meteorological data will be supplied by models (Eq. 34b for net short wave, Eq. 34e for net long wave) (0 = none 1 = Net short wave 2 = Net long wave 3 = Net all wave) 4 = Incoming solar radiation 5 =)

The following are the data arrays, maximum length 1000.

TMPAIR	R*4	1000*293.0	Air temperature (degrees K)
VAPORS	R*4	1000*15.0	Vapor pressure (mb)
WINDS	R*4	1000*100.0	Wind speed (cm/sec)
SHORT	R*4	1000*0.0	Net short wave radiation (cal cm^{-2} sec^{-1})
SLONG	R*4	1000*0.0	Net long wave radiation (cal cm^{-2} sec^{-1})
RTOT	R*4	1000*0.0	Net all wave (total) radiation (cal cm^{-2} sec^{-1})
CLDIN	R*4	1000*0.0	Fraction of the solar radiation absorbed by the cloud cover (B, Eq. 34b). This is used only when solar radiation is modelled (IROUT = 0 OR 2)
SOLAR	R*4	1000*0.0	Downwelling solar radiation (cal cm^{-2} sec^{-1})

HYDRAULIC DATA INPUT

The look-up tables for the hydraulic conductivity and matric potential as functions of the volumetric soil moisture are supplied in the NAMELIST data set named HYD. The data set is on the unit number specified by the parameter IHDUNT in the main NAMELIST data set, INPUT.

The interpolation is linear between the nearest moisture values in the table. If the soil moisture is either smaller than the smallest table entry or larger then the largest table entry a linear extrapolation is performed. To simplify the interpolation logic, the soil moisture values in the look-up table must be evenly spaced. Therefore this option may be used only when modelling a homogeneous soil profile.

The following describes the elements in HYD:

NAME	TYPE	DEFAULT	DESCRIPTION
NTHETA	I*4	0	Number of table entries (< 500)
THETA(I)	R*4	none	Volumetric soil moisture.
COND(I)	R*4	none	Hydraulic conductivity (cm/sec)
HEAD(I)	R*4	none	Matric potential (cm)

DESCRIPTION OF SIMULATOR OUTPUT

The first two pages of output show values of the variables in NAMELIST INPUT.
Page 1 has the physical constants which vary with soil depth, and page 2 has values for all other variables.

OUTPUT AT A PARTICULAR TIME

TIME — elapsed (simulator) time since 0 hours of the first simulation day.
the format is DDDHHMMSS.MM where
DDD = days
HH = hours
MM = minutes
SS = seconds
MM = milliseconds

CLOCK TIME — (**) Local clock time (Standard time or Day Light Saving time) of current output. Hours and decimal fractions of an hour.

CUMULATIVE WETNESS VARIABLES

TOTAL IN PROFILE is the amout of water (cm) in the profile, calculated from

$$TOTAL = \sum_{i=1}^{NL} DZ(i)\,\theta(i)$$

The next lines give integrated moisture fluxes at the boundaries specified by the NAMELIST parameters NWCUMS and IXWCUM. The next line shows values of parameters associated with the surface energy balance. These are the total, soil, and plant evaporation, and, if IAERO = 1, the Monin-Obukhov length, stability corrections Ψ_1 and Ψ_2, (Equation (46)) bare soil resistance (Equation 47), and aerodynamic resistance (Equation 46). The next lines show values of the surface energy fluxes.

SURFACE TEMPERATURE

If the temperature profile is modelled (ITEMPS=1), then this is the solution of Equation 62. If the force-restore method is used (ITEMPS=2) then this is the solution of Equation 67a.

Note that the equations numbers that follow must be updated after Method of Solution is added.

VALUES OF SOIL VARIABLES

DEPTH - Depth to the center of the layer
MOISTURE - Volumetric moisture
M FLUX - Moisture flux (cm/sec) at the top of the layer (Equation 78A). The last entry in this column is the moisture flux at the bottom of the profile.
SINK - Q(z,t) (1/sec), (Equation 68)
HYD COND - Hydraulic conductivity (cm/sec) of moisture in the layer.
PHEAD - Matric potential (cm)
DWDT - $d\theta/dt$, (Equation 68, 78)
TEMP - Layer temperature (K)

H FLUX - Heat flux (cal cm^{-2} sec^{-1}) at the top of the layer. The last entry is the heat flux at the bottom of the profile.
TCOND - Layer thermal conductivity (cal cm^{-1} sec^{-1} K^{-1}).
VHCAP - Layer volumetric heat capacity (cal cm^{-3}) (Equation 29)
DTEMPDT - dT/dt (Equation 78c).

Information at each output time can also be output to disk or tape. The NAMELIST variable NDISK controls whether this is done, and IUDISK is the FORTRAN unit number of the DD card which points to the output data set. This can be a sequential data set on disk or tape.

The output records are unformatted. The record length should be at least 8*NL+80 bytes, where NL is the number of soil layers. This is 2*NL + 20 words per data record. The records are written by subroutine DSKOUT. They can be read with C program convert_out.c.

One header record is output at the start of the simulations, and one data record is output at each simulator output time. The header record contains the number of soil layers and thickness of each layer. Each data record contains the following: output time, soil moisture in each layer (NL values), temperature in each layer (NL values), surface temperature (either the force-restore or that from the heat balance equation solution, depending on whether ITEMPS is 2 or 1), soil heat flux, net radiation, latent heat flux, sensible heat flux, bare soil evaporation rate, and the plant transpiration rate. The record formats are depicted below.

Structure of Unformatted Output Data Record Created by SOILSIM

Header Record

Words	Contents
1	Binary zero
2	Number of soil layers, Interger*4
3 ↓ NL+ 2	Layer thickness, Real*4 ↓
NL + 3 ↓ 2NL + 2	Zero, Real*4 ↓

Data Record

Words Contents

1-2	Output time (DDDHHMMSS.SS), Real*8
3 ↓ NL + 2	Soil Moisture, Real*4 ↓
NL + 3 ↓ 2NL + 2	Soil Temperature, Real*4 ↓
2NL + 3	Surface Temperature (K), TSURF, Real*4
2NL + 4	Net heat + radiation flux at the soil-air interface (cal cm^{-2} sec^{-1}), SABS, Real*4
2NL + 5	Net short and long radiation summed for soil and canopy (cal cm^{-2} sec^{-1}), RNET, Real*4
2NL + 6	Evapotranspiration (cal cm^{-2} sec^{-1}), ETHEAT, Real*4
2NL + 7	Sensible heat flux (cal cm^{-2} sec^{-1}), SENTOT, Real*4
2NL + 8	Evapotranspiration from soil (cm^3 cm^{-2} sec^{-1}), ESOIL, Real*4
2NL + 9	Evapotranspiration from canopy (cm^3 cm^{-2} sec^{-1}), EPLANT, Real*4
2NL + 10	Net solar absorbed by soil (cal cm^{-2} sec^{-1}), SLSOIL, Real*4
2NL + 11	Net solar absorbed by canopy (cal cm^{-2} sec^{-1}), SLCAN, Real*4
2NL + 12	Net long wave radiation absorbed by soil (cal cm^{-2} sec^{-1}), RNSOIL, Real*4
2NL + 13	Net long wave radiation absorbed by canopy (cal cm^{-2} sec^{-1}), RNCAN, Real*4
2NL + 14	Temperature of canopy leaves (K), TCANL, Real*4
2NL + 15	Temperature of canopy air (K), TCANA, Real*4
2NL + 16	Total evapotranspiration from soil and canopy (cm^3 cm^{-2} sec^{-1}), ETOT, Real*4
2NL + 17	Sensible heaf flux from soil (cal cm^{-2} sec^{-1}), SENSOI, Real*4
2NL + 18	Sensible heat flux from canopy (cal cm^{-2} sec^{-1}), SENCAN, Real*4
2NL + 19	(**)Modeled net solar radiation absorbed by canopy + soil (cal cm^{-2} sec^{-1}), SLTOT, Real*4
2NL + 20	(**)Modeled PAR absorbed by canopy. PARCAN, Real*4

Other Significant Variables

RCAN00	Canopy stomatal resistance (r_{st}, Eq. 49)
ECANA	Canopy air vapor pressure
ECANL	Canopy air space vapor pressure
TCANA	Canopy air temperature
TCANL	Canopy leaf temperature
VPDCAN	Canopy air vapor pressure deficit
SCALE	Ratio of plant potential transpiration to actual transpiration
VSCALE	Scale factor of 1/1000 used in Eq. 45b.
CON1	$\rho c_p/(L\gamma)$
CON2	ρc_p
HSL	Interception on soil surface (Eq. 52b)
HPL	Interception on plant canopy surfaces (Eq. 52a)
SLRAT	Ratio of depth of surface water to its maximum depth on soil surface (Eq. 51b).
SLRAT1	Complement of LSRAT (1-SLRAT)
PLRAT	Ratio of depth (HPL) of water on plant surfaces to its maximum depth (HPLMAX) (Eq. 51a.
PLRAT1	Complement of PLART (1-PLRAT)
CP1	CON1 * plant cover fraction * SLRAT / RCA (Eq. 55)
CP2	CON1 * plant cover fraction * SLRAT1 / (RCA + RCAN) (Ea. 55)
VAP0S	Saturation vapor pressure of soil surface layer
VAPSOI	Actual calculated vapor pressure of the soil
VAPZ0	Observed vapor pressure of atmosphere
DAVP	The difference VAPSOI - VAPZ0
ESOIL	Evaporation from soil
SENSOI	Sensible heat flux from soil
SENCAN	Sensible heat flux from canopy

CPOT	Crown Potential, (Eq. 69)
TCENT	Temperature Celsius
T0	273.16 K, 0° Celsius
TSTEXP	Exponent on canopy stress temperature function
SINA	Sine of the solar altitude angle
CANALB	Canopy albedo (CALB(1) + SINA)
CANM1	Canopy absorption coefficient (1 - CANALB)
RSOLAR	Downwelling solar radiation
RSHORT	Net short wave radiation
RLONG	Net long wave radiation above canopy
RNET	Net allwave radiation above canopy
RNCAN	Net long wave radiation from canopy
RNSOIL	Net long wave radiation from soil surface

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