A Method for Coupling a Parameterization of the Planetary Boundary Layer with a Hydrologic Model

J. D. LIN

Civil Engineering Department, University of Connecticut, Storrs, CT 06268

SHU FEN SUN

Institute of Mechanics, Chinese Academy of Sciences, Beijing, China

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ABSTRACT

Deardorff's parameterization of the planetary boundary layer is adapted to drive a hydrologic model. The method converts the atmospheric conditions measured at the anemometer height at one site to the mean values in the planetary boundary layer; it then uses the planetary boundary layer parameterization and the hydrologic variables to calculate the fluxes of momentum, heat and moisture at the atmosphere–land interface for a different site. A simplified hydrologic model is used for a simulation study of soil moisture and ground temperature on three different land surface covers. The results indicate that this method can be used to drive a spatially distributed hydrologic model by using observed data available at a meteorological station located on or nearby the site.

1. Introduction

Spatially distributed hydrologic models (Beven, 1985) designed to simulate or predict the hydrologic cycle on the basis of the principles of water mass and thermal energy balances require meteorological data over each grid in a watershed. Observed data are, at best, only available at a few meteorological stations, located in the watershed or nearby. To utilize the data from these stations to drive the hydrologic model, modification of the observed data is necessary because local atmospheric conditions above various land surface covers are never uniform over the entire watershed.

Rutter et al. (1975) made use of data taken some distance from a forested area with a modification of wind velocity to study the evapotranspiration from the forest. They used the formula given in the Meteorological Office Handbook (1969) to extrapolate the data measured at the anemometer height to a higher level in the planetary boundary layer (PBL). The extrapolated data were then used to calculate the values at a site having a different surface roughness under neutral or near-neutral conditions. Sellers (1981) modified the wind profile from one site for use at another and included the effect of nonneutrality and different surface roughness. His approach was a combination of Ayra's (1975) description of shear stress induced by geostrophic winds and Paulson's (1970) wind and temperature profiles. The basic assumption in these spatial extrapolations is that the geostrophic wind is the same over the gaging station and the site.

Deardorff's (1972) parameterization of the PBL, originally designed for use in an atmospheric general circulation model, is adapted to link the atmospheric conditions at the midlevel of the PBL to those at the anemometer height and at the lower boundary of the PBL just above the land surface cover. Data at the anemometer height can be converted to the mean values in the PBL, and sensible and latent heat fluxes can be directly estimated on a nearby site by using these mean values and specified lower boundary conditions. As long as the distance between these sites is less than the horizontal characteristic length of the PBL (of the order of PBL height), the atmospheric condition at the midlevel of the PBL should remain nearly the same, despite the difference in the land surface cover and the hydrologic conditions. However, in order to use this parameterization with a hydrologic model, it is necessary to further link the lower boundary conditions to the hydrologic conditions at the atmosphere–land interface. This paper proposes a method for the coupling, which can use meteorological data measured at one site to drive a hydrologic model applied at another site with a quite different land surface cover. This method is applied to a simulation study of soil moisture and ground temperature for three different land surface covers for 7 days, using a simplified version of the ground hydrologic model developed by Lin et al. (1979).

2. Conversion of the data observed at a meteorological station to the mean values in the PBL

The parameterization of Deardorff (1972) provides the following relation of atmospheric conditions between the midlevel of the PBL and the anemometer height:

$$\frac{u_m - u_a}{u_*} = 8.4\left(1 - \frac{50h}{L}\right)^{-0.16} \quad (1a)$$

$$\frac{u_*(\theta_{vm} - \theta_{va})}{-F} = 7.3\left(1 - \frac{5.8h}{L}\right)^{-0.47} \quad (1b)$$

for the unstable case ($F > 0$), and

$$\frac{u_m - u_a}{u_*} = 8.4 + \frac{0.6h}{L} \quad (2a)$$

$$\frac{u_*(\theta_{vm} - \theta_{va})}{-F} = 7.3 + \frac{0.6h}{L} \quad (2b)$$

for the stable or neutral case ($F < 0$ or $F = 0$), where the Monin-Obukhov length is

$$L = -\frac{u_*^3}{kgF}\theta_{vm} \quad (3)$$

and

$$F = F_\theta + 0.61\theta_m F_q \quad (4)$$

Given the friction velocity, u_*, and the kinematic fluxes of sensible heat and moisture, F_θ and F_q, Eqs. (1) to (4) can be used to convert the velocity and virtual potential temperature at the anemometer height to the mean values in the PBL. An explicit solution for the stable case is

$$\theta_{vm} = \frac{-b + (b^2 - 4ac)^{0.5}}{2a} \quad (5a)$$

and

$$u_m = u_a + u_*(8.4 + 0.6h/L) \quad (5b)$$

where

$$a = u_*^4$$

$$b = 7.3u_*^3 F - \theta_{va} u_*^4$$

$$c = -0.6hkgF^2$$

For the unstable case, a numerical solution must be sought. After θ_{vm} is obtained, θ_m and q_m can be partitioned by

$$\theta_m = \theta_a + (\theta_{vm} - \theta_{va})F_\theta/F \quad (6a)$$

and

$$q_m = q_a + (q_{vm} - q_{va})F_q/F \quad (6b)$$

In order to find u_*, F_θ and F_q for the unstable case [$(\theta_{va} - \theta_{vs}) < 0$], the formulas of Paulson (1970) may be used. These are

$$\frac{u_a}{u_*} = \frac{1}{k}\left(\ln\frac{z_a}{z_0} - \ln\frac{1+x^2}{2} - 2\ln\frac{1+x}{2} + 2\tan^{-1}x - 0.5\right) \quad (7a)$$

and

$$\frac{(\theta_{va} - \theta_{vs})u_*}{-F} = \frac{R}{k}\left(\ln\frac{z_a}{z_0} - 2\ln\frac{1+y^2}{2}\right) \quad (7b)$$

where

$$x = (1 - 15z_a/L)^{0.25} \quad (7c)$$

$$y = (1 - 9z_a/L)^{0.25} \quad (7d)$$

For the stable case [$(\theta_{va} - \theta_{vs}) > 0$],

$$\frac{u_a}{u_*} = \frac{1}{k}\left(\ln\frac{z_a}{z_0} + \frac{4.7z_a}{L}\right) \quad (8a)$$

and

$$\frac{(\theta_{va} - \theta_{vs})u_*}{-F} = \frac{R}{k}\left(\ln\frac{z_a}{z_0} + \frac{4.7z_a}{RL}\right) \quad (8b)$$

Similar to Eqs. (1) to (3), the stable case can be solved explicitly as

$$F = u_*(au_* - u_a)/b \quad (9a)$$

$$u_* = \frac{(2-R)u_a - ((u_a R)^2 + 4b(1-R)(\theta_{va} - \theta_{vs}))^{0.5}}{2a(1-R)} \quad (9b)$$

where $a = 1/k \ln(z_a/z_0)$, $b = 4.7z_a g/\theta_{vm}$ and θ_{vm} is approximated by θ_{va}. Again, the solution for the unstable case must be solved numerically. It should be noted that the above equations were derived from conditions over a fairly flat, uniform terrain. Hogstrom (1974) has shown the effect of deviations from "ideal" conditions on the flux-profile relations, especially for the stable case.

In Eqs. (7) and (8), the virtual potential temperature θ_{vs} is calculated by assuming that the air temperature at the lower boundary of the PBL equals the temperature of land surface cover. However, a further link to the hydrologic conditions is necessary in order to obtain the specific humidity q_s, based on the definition of evaporation from bare soil or transpiration from a vegetation canopy.

The evaporation from bare soil is given by

$$E_s = \beta_s E_p = \rho_a(q_s - q_a)/r_{av} \quad (10)$$

where the potential evaporation

$$E_p = \rho_a(q^*(T_g) - q_a)/r_{av} \quad (11)$$

and β_s is a scaling factor, depending on the level of available soil moisture. The transpiration from vegetation is given by

$$E_c = \beta_c E_u = \rho_a(q_s - q_a)/r_{av} \quad (12)$$

where the unstressed transpiration (Federer, 1977) is

$$E_u = \rho_a(q^*(T_c) - q_a)/(r_{av} + r_c) \quad (13)$$

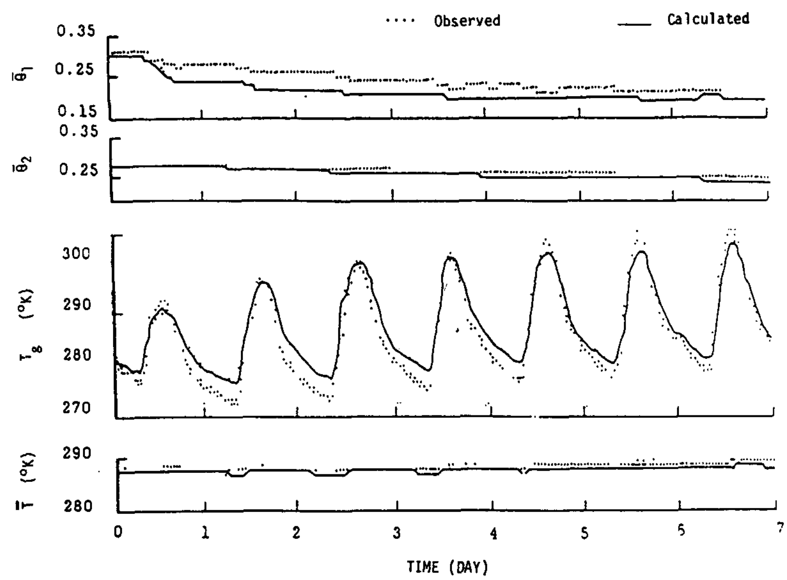

FIG. 1. Calculated and measured soil moisture content and temperature of bare soil in Arizona.

and β_c is the scaling factor, depending on available soil moisture, evaporative demand of the atmosphere and canopy resistance (Sun, 1985). Equations (10) and (11) can be solved for bare soil:

$$q_s = \beta_s q^*(T_g) + (1 - \beta_s) q_a \tag{14a}$$

and Eqs. (12) and (13) for vegetation:

$$q_s = \beta_c q^*(T_c) r_{av}/(r_{av} + r_c) + \beta_c q_a (1 - r_{av}/(r_{av} + r_c)) \tag{14b}$$

3. Calculation of the fluxes of sensible heat and moisture over the site of the hydrologic model

By assuming the same mean values in the PBL over the two sites, the fluxes of sensible heat and moisture can be calculated by Deardorff's parameterization as follows. The stability condition is expressed in terms of a bulk Richardson number,

$$R_{iB} = \frac{gh(\theta_{vm} - \theta_{vs})}{\theta_{vm} u_m^2} \tag{15}$$

For the stable case ($R_{iB} < 0$), the coefficients of friction and heat transfer are

$$C_u = C_{uN}(1 - R_{iB}/R_{ic}) \tag{16a}$$

and

$$C_\theta = C_{\theta N}(1 - R_{iB}/R_{ic}) \tag{16b}$$

where R_{iB} should be less than $0.9R_{ic}$. The C_{uN} and $C_{\theta N}$ are the coefficients of friction and heat transfer for the neutral case depending on h and z_0. They are

$$C_{uN} = \left(\frac{1}{k} \ln \frac{0.025h}{z_0} + 8.4\right)^{-1} \tag{16c}$$

and

$$C_{\theta N} = \left(\frac{R}{k} \ln \frac{0.025h}{z_0} + 7.3\right)^{-1} \tag{16d}$$

For the unstable case ($R_{iB} < 0$),

$$C_u = [C_{uN}^{-1} - 25 \exp(0.26x - 0.03x^2)]^{-1} \tag{17a}$$

and

$$C_\theta = (C_{\theta N}^{-1} + C_u^{-1} - C_{uN}^{-1})^{-1} \tag{17b}$$

where $x = \log_{10}(-R_{iB}) - 3.5$. All coefficients for the unstable case must be subject to adjustment, due to occurrence of the free convection [see Deardorff (1972)].

After C_θ and C_u are obtained, the kinematic vertical fluxes F can be partitioned into kinematic fluxes of sensible heat and moisture by Eq. (6), using the values at the lower boundary of the PBL instead of those at the anemometer height.

4. The hydrologic model

The hydrologic model used in this study is a simplified version of the large-scale hydrologic model developed by Lin et al. (1979) and later refined by Sun (1985). The two-layer model consists of a surface layer and a lower layer for soil moisture and ground temperature.

The average volumetric soil moisture contents, θ_1 in the surface layer and θ_2 in the lower layer, are governed by

$$d_1 \frac{d\theta_1}{dt} = -U_1 - q_{12} \tag{18a}$$

$$d_2 \frac{d\theta_2}{dt} = q_{12} - q_{23} - U_2 \tag{18b}$$

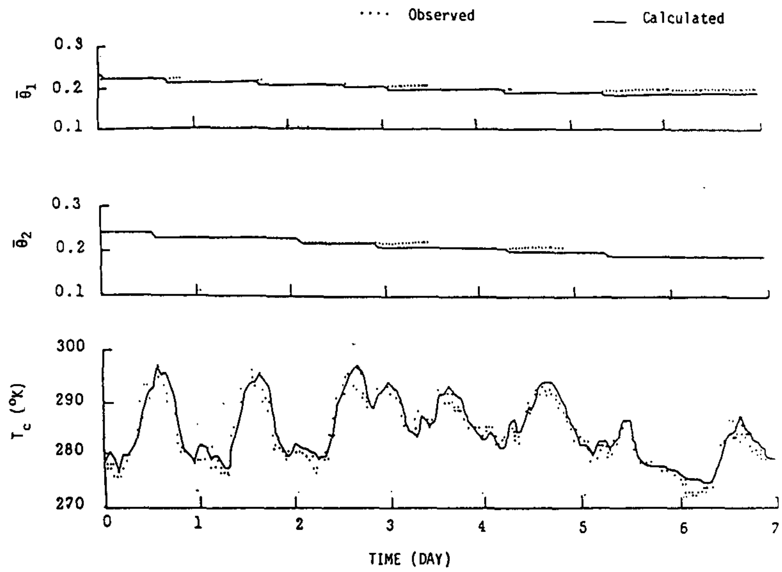

FIG. 2. As in Fig. 1, but for winter wheat in Arizona.

where U_i is the water uptake rate from layer i. For bare soil, U_1 is the evaporation and $U_2 = 0$. For vegetation, the transpiration is partitioned into U_1 and U_2, i.e.,

$$E_c = \rho_w(U_1 + U_2) \tag{18c}$$

The ground surface temperature T_g is predicted by the force-restore method (Lin, 1980). The rate equation is

$$\frac{dT_g}{dt} = \frac{2G}{C_s d} - \frac{2\pi(T_g - \bar{T})}{\tau} \tag{19}$$

where G is the heat flux into the bare soil surface or vegetated soil surface. The G can be derived from the energy balance on the ground surface. The rate equation for the depth-averaged temperature is

$$\frac{d\bar{T}}{dt} = \frac{G}{C_s d(365\pi)^{1/2}} \tag{20}$$

For a bulk layer of vegetation over the ground surface, the canopy temperature T_c is given by (Sun, 1985)

$$C_c h_c \frac{dT_c}{dt} = M \tag{21}$$

where M is the rate of heat storage in the canopy; it can be derived from the energy balance in the bulk canopy layer.

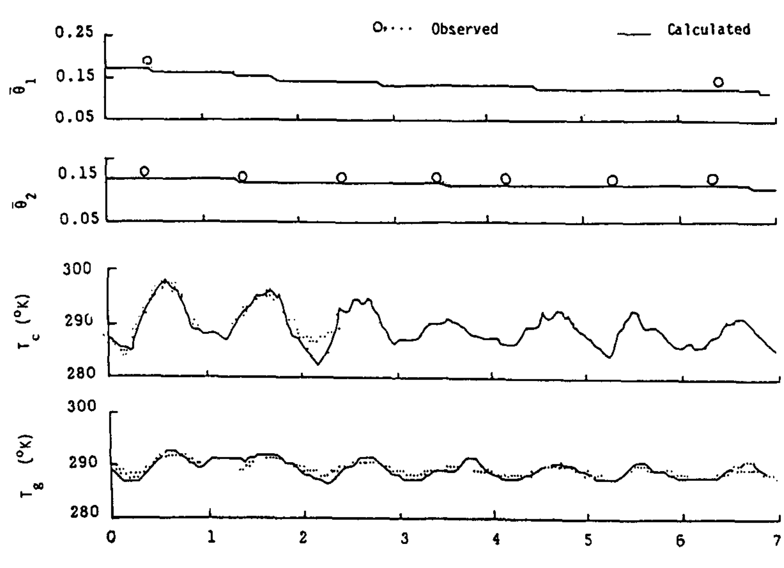

FIG. 3. As in Fig. 1, but for Douglas fir in British Columbia.

5. The simulation study

To evaluate the method, three sets of field data were used for the simulation study. Two are from a half acre site in Phoenix, Arizona (Jackson, 1971, 1973, 1976). The first set is for bare soil surface while the second is for winter wheat. The third one is from a study of evapotranspiration of Douglas fir in British Columbia (Black, 1979). The mean values in the PBL, which were used to drive the hydrologic model, were calculated from the measured atmospheric data at the instrument (anemometer) height. The simulation was conducted for a period of 7 days.

Although the meteorological data in these experiments were taken at the same site, the simulation results indicate that this method can provide a satisfactory prediction of soil moisture and temperature, respectively, for three distinctively different land surface covers. Comparisons of moisture and heat fluxes are not shown here because not all of these fluxes were measured by the experiments. The simulated and measured values of soil moisture and ground temperature are summarized in Figs. 1 to 3. In the bare soil simulation, the calculated values of θ_1 in the 10-cm surface layer decreases more rapidly as compared with the measured values. However, a better agreement was found in the averaged soil moisture contents in the 50-cm root zone, which is defined in this study by $(\theta_1 d_1 + \theta_2 d_2)/(d_1 + d_2)$. The measured surface temperature appears to have a larger diurnal amplitude than the predicted. These discrepancies might be caused by the large gradients of soil moisture and temperature that exist very close to the surface. The results for the vegetation canopies show good agreement in soil moisture and canopy and ground temperatures. Thus, it appears that the proposed method can be used to drive a hydrologic model by using observed data at a meteorological station located on or nearby the site.

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APPENDIX

List of Symbols

C_c	volumetric heat capacity of canopy
C_s	volumetric heat capacity of soil
C_u	friction coefficient u_*/u_m
C_{uN}	friction coefficient in neutral case
C_θ	heat transfer coefficient $-F/(u_*(\theta_{vm} - \theta_{vs}))$
$C_{\theta N}$	heat transfer coefficient in neutral case
c_p	specific heat of air
d_i	depth of layer i in root zone
d	penetrating depth of diurnal temperature wave
E_c	evapotranspiration from canopy
E_p	potential evapotranspiration
E_s	evaporation from bare soil
E_u	unstressed transpiration from canopy
F	kinematic vertical flux as defined in Eq. (4)
F_q	kinematic moisture flux
F_θ	kinematic flux of sensible heat
G	heat flux into ground surface
g	gravitational acceleration
H	sensible heat flux
H_c	sensible heat flux from canopy
H_s	sensible heat flux from bare soil
h	thickness of planetary boundary layer
h_c	height of canopy
κ	von Karman constant (=0.35)
L	Moning-Obukhov length
M	rate of heat storage in canopy
p	air pressure
q	specific humidity
q_{ij}	moisture flux between soil layer i and j
$q^*(T)$	saturated specific humidity at temperature T
R	0.74
R_{iB}	bulk Richardson number
R_{ic}	critical Richardson number (=3.05)
r_{av}	aerodynamic resistance to moisture transfer
r_c	canopy resistance to transpiration
$\bar{T}$	depth-averaged temperature of soil layer
T_c	canopy temperature
T_g	ground surface temperature
U_i	water uptake rate in layer i
u	air velocity
u_*	friction velocity
z_a	anemometer height
z_0	roughness height
β_c	transpiration scaling factor
β_s	evaporation scaling factor
θ	potential temperature $[=T(1000/p)^{0.288}]$
θ_v	virtual potential temperature $[=\theta(1 + 0.61q)]$
θ_i	volumetric moisture content in layer i heat conductivity of soil
ρ_a	air density
ρ_w	water density
τ	diurnal period (=86 400 sec.)

Subscript

a	value at anemometer height
m	mean value in PBL
s	surface value in the air above land cover

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