ONE-DIMENSIONAL SIMULATION OF TEMPERATURE AND MOISTURE IN ATMOSPHERIC AND SOIL BOUNDARY LAYERS

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1. INTRODUCTION

Meteorologists are interested in modeling the vertical flow of heat and moisture through the soil in order to better simulate the vertical and temporal variations of the atmospheric boundary layer. In the present study the one-dimensional PBL model of Diertele (1979) is modified by the addition of transport equations to be solved by a finite difference technique to predict soil moisture. The model of Diertele is a refinement of the two-dimensional model of Bornstein (1975) and Bornstein and Robock (1976) in which a one-dimensional time-dependent model of the planetary boundary layer simulates the vertical distribution of wind, temperature, and water vapor.

2. FORMULATION

Time dependent solutions for the vertical distribution of wind, temperature, and water vapor in the planetary boundary layer (PBL) can be obtained from equations used by Bornstein (1975) in his URBMET (for urban meteorology) model. This system of equations describes a two-dimensional (x-z plane) boundary layer model, with a lower, analytical "constant flux" layer of 25 meters, and an upper, numerical transition layer of 1900 meters.

A one-dimensional version of the model valid over homogeneous terrain was developed by Diertele (1979) to include the effects of: 1) heating due to divergence of the long and short wave radiative fluxes, and 2) an energy conservation equations for the air-ground interface, including radiative, sensible, and latent heat fluxes. Use of this equation requires inclusion in the model of a subsurface layer for prediction of soil sensible heat flux. In the current formulation, the differential equations of soil heat and moisture fluxes are solved using an explicit finite difference technique to obtain soil temperature and moisture respectively. Instead of using a parametrization for solving surface latent heat flux, it is obtained more accurately from the calculations of surface moisture and temperature.

a. Transition Layer

The equations governing atmospheric variables in a one-dimensional, hydrostatic, Boussinesq transition layer were given by Bornstein (1975). With the additional assumptions discussed above, the equations become

$$\frac{\delta \zeta}{\delta t} = f\frac{\delta v}{\delta z} + \frac{\delta^2}{\delta z^2}(K_M \zeta) \qquad (1)$$

$$\frac{\delta v}{\delta t} = f(u_g - u) + \frac{\delta}{\delta z}\left(K_M \frac{\delta v}{\delta z}\right) \qquad (2)$$

$$\frac{\delta \theta}{\delta t} = \frac{\delta}{\delta z}\left(K_H \frac{\delta \theta}{\delta z}\right) - \frac{1}{\rho_m c_p}\frac{\delta Q_N}{\delta z} \qquad (3)$$

$$\frac{\delta q}{\delta t} = \frac{\delta}{\delta z}\left(K_q \frac{\delta q}{\delta z}\right) \qquad (4)$$

$$u = \frac{\delta \psi}{\delta z} \qquad (5)$$

$$\zeta = \frac{\delta u}{\delta z} = \frac{\delta^2 \psi}{\delta z^2} \qquad (6)$$

where all symbols are defined in Appendix.

The verticity and stream function approach, used in the original model, is retained in the present one-dimensional study for convenience. The eddy diffusivities K_M, K_H, and K_V (set equal to K_H) are specified using the interpolation formula of O'Brien (1970). Details of the formulation of radiative flux divergence which enters into the thermal energy equation (2-3) for PBL, can be found in Santhanam (1980) as can the details of the equations for the constant flux surface boundary layer. Only a cloudless and pollutant free atmosphere is considered in the present study, which therefore restricts the generality of the model.

b. Soil Layer

The flow of moisture through soil can be written as

$$\frac{\delta \eta_s}{\delta t} = \frac{\delta}{\delta z}\left(D_\eta \frac{\delta \eta_s}{\delta z}\right) + \frac{\delta}{\delta z}\left(D_T \frac{\delta T_s}{\delta z}\right) + \frac{\delta}{\delta z}(K_W), \quad (7)$$

where the capillary action term (first term on the right hand side of the equation) is generally larger than the convection term (middle term) and the gravity term (last term). The flow of heat through the soil can be written as

$$\frac{\delta T_s}{\delta t} = \frac{\delta}{\delta z}\left(\frac{\lambda'}{c} \frac{\delta T_s}{\delta z}\right) + L_E \rho_w \frac{\delta}{\delta z}\left(\frac{D_{\eta v}}{c} \frac{\delta \eta_s}{\delta z}\right), \quad (8)$$

where the first term on the right hand side of the equation represents sensible heat flux and the second term represents latent heat flux. All values of the moisture dependent diffusivities (D_N, D_{NV}, and D_T), as well as the hydraulic conductivity (K_W) and the modified thermal conductivity (λ') were obtained from Sasamori (1970). One difference between current formulation and that of Sasamori (1970) is the present use of moisture dependent values of soil heat capacity and density.

c. Grid and Boundary Conditions

The one-dimensional version of the interlaced grid of Fromm (1964) computes vorticity and stream function values at grid locations vertically displaced by one-half of a grid interval from locations of the velocity components. Studies by Clark (1970) and Taylor and Delage (1971) discuss the superiority of variable grid spacing in achieving high resolution near the surface, and the grid network used in the present model possesses such resolution. A complete list of boundary conditions used in the present simulation can be found in Santhanam (1980).

In order to obtain the following equation for the time-varying surface temperature, a balance is assumed at the surface between the computed net radiant flux and convection in the atmosphere, conduction into the soil, and evaporation:

$$Q_R + Q_{L\downarrow} - Q_{L\uparrow} - \epsilon_s \sigma T_g^4 + \rho_m c_p u_* \theta_*$$

$$+ \rho_m L_E u_* q_* + \lambda' \frac{\delta T_s}{\delta z} + L_E \rho_w D_{\eta v} \frac{\delta \eta_s}{\delta z} = 0. \quad (9)$$

In order to obtain the following similar equation for surface moisture, a balance is assumed between the atmospheric surface water vapor flux and soil surface moisture flux:

$$\rho_m u_* q_* + \rho_w D_\eta \frac{\delta \eta_s}{\delta z} + \rho_w D_T \frac{\delta T_s}{\delta z} + \rho_w K_W = 0. \quad (10)$$

Equations (9) and (10) are partial differential equations with two unknowns, i.e., moisture and temperature. Solutions of these two equations are obtained by using a Newton-Ralphson double iterative technique (Carnahan et al., 1969).

The specific humidity at ground can be obtained from

$$q_g = q_s \times h_g, \quad (11)$$

where q_s is obtained from an expression given by Murray (1967) and h_g is obtained from equations from Philip (1957) and Nappo (1975).

3. SOLUTION

At each time step, the following procedure is performed: determination of soil layer profiles; construction of constant flux profiles; determination of the new values of ϕ, q, η, and U at the surface and at the lowest finite determination of the new ϕ, q, ζ, Ψ, u, v, and U profiles using newest available values of all parameters. Details can be found in Santhanam (1980).

4. RESULTS

This simulation attempts to reproduce the temperature and moisture fields in the atmospheric and soil boundary layers. Initial conditions for these simulations were taken as those existing at the beginning of the 5th general observational period at O'Neill. Nebraska, 1230 LST on August 24, 1953 (Lettau and Davidson, 1957). Numerical integration was carried out for a 48 hour period.

Computed soil moisture values near the surface (approximately 0.25 cm below surface) are compared with those of Sasamori (1970) in Fig. 1. Soil moisture displays a minimum value in the afternoon as the surface dries because evaporation is greater than the moisture flux from below. A maximum value is found after midnight. Wetness values decrease from 5.4 % at 1200 LST on the first day to 3.1% at 1200 LST on the next day, indicating a continuous depletion of soil moisture. The depletion arises as the moisture flux from below the surface is less than surface evaporation and as nighttime surface recondensation is also less than daytime evaporation. In general, the present results are similar to those of Sasamori results are similar to those of Sasamori (1970).

Computed surface relative humidity values (Fig. 2) compare well with those of Nappo (1975), who obtained his values by dividing observed mixing ratio values at 10 cm into the atmosphere (which he assumed equal to that at the surface) by the saturation mixing ratios. However, his specific humidity values which were obtained by multiplying relative humidity by saturation specific humidity (which was not explicitly presented in his original paper) do not match the present predictions (Fig. 3) in amplitude and phase. However, the present calculations do agree somewhat better with those of Sasamori (1970).

Estimated Bowen ratios (of sensible flux to latent heat flux) from various models are shown in Fig. 4. Those of Suomi (computed from observations of net radiation and soil heat flux) show an unrealistic jump at 1500 LST due to a too small latent heat flux. In addition, his values beyond 18 hours of simulated time, as well as those of Lettau (Lettau and Davidson, 1957), which are obtained in a similar manner to those of Suomi, do not agree with the present results and those of Sasamori (1970). These later results as well as the present ones, are from complete PBL formulations, while those of Suomi and Lettau are associated with simple parametrizations.

5. CONCLUSION

The two dimensional URBMET urban planetary boundary layer model has been simplified to one dimension and modified to perform a detailed analysis of the flow of soil moisture into the atmosphere. Model calculations of temperature and moisture in the soil and atmospheric boundary layers were validated against existing observations at O'Neill, Nebraska.

Results have indicated that simple models with parametrizations of the surface latent heat flux do well in predicting surface relative humidity but do not do well for specific humidity, because they do well in predicting surface temperature. Calculations show that models using empirical techniques to evaluate the surface energy balance do not predict Bowen ratios as well as those incorporating a soil layer.

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APPENDIX

List of Symbols

Variables and Constants - Roman Alphabet

c	soil heat capacity
c_p	specific heat of dry air
D_η	total soil moisture diffusivity
$D_{\eta v}$	vapor moisture diffusivity
D_T	total soil thermal diffusivity
f	Coriolis parameter
h_g	surface relative humidity
K_H, K_M, K_q	eddy exchange coefficients for heat, momentum, and water vapor
K_W	hydraulic condictivity
L_E	latent heat of vaporization
q, q_g, q_s, q_*	specific humidity, surface q, saturated q, and friction q
Q_E, Q_H	vertical flux of latent and sensible heats
$Q_{L\uparrow}$, $Q_{L\downarrow}$	upward and downward directed long-wave radiation
Q_N, Q_R	net all-wave and solar radiation at the surface
t	time
T_g, T_s	surface and soil temperature
u, u_g	horizontal component of the wind and geostropic wind in x-direction
u_*	friction velocity
v	horizontal component of the wind in the y-direction
x, y, z	horizontal coordinate in direction of geostropic wind, perpendicular to geostropic wind and vertical coordinate

Variables and Constants - Greek Alphabet

ϵ_s	wave length integrated surface emissivity
ζ	modified y-component of vorticity
η_g	soil wetness
θ, θ_*	potential temperature and friction θ
λ'	modified soil thermal conductivity
ρ_m	constant space averaged atmospheric density
ρ_w	density of water
σ	Stefan-Boltzman constant
Ψ	stream function

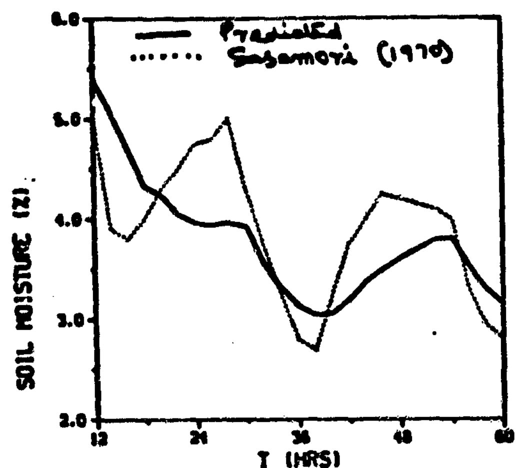

Figure 1.
Predicted soil moisture (by volume) at 0.25 cm below surface, where 12, the initial time, corresponds to 1200 LST.

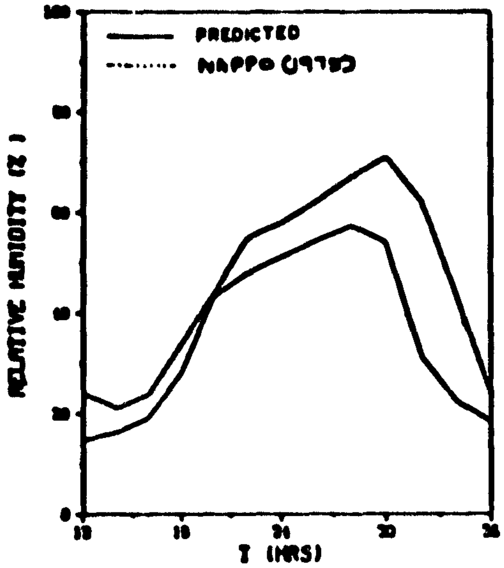

Figure 2.
Predicted surface relative humitidy (%)

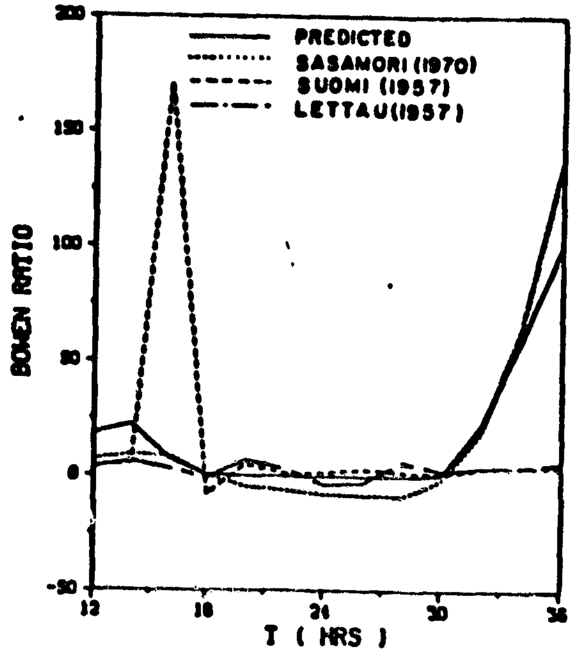

Figure 4.
Predicted Bowen ratios.

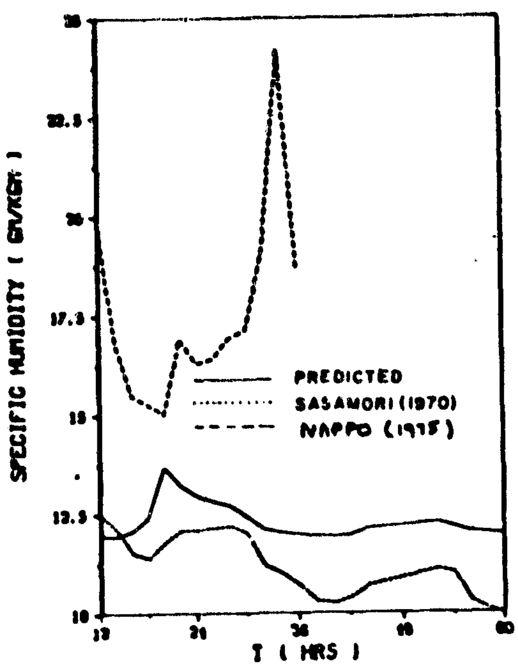

Figure 3.
Predicted surface specific humidity.