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ONE DIMENSIONAL MODEL OF ATMOSPHERIC BOUNDARY LAYER

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Abstract

A one-dimensional boundary layer model as a combination of a boundary layer model and a two layer soil model is presented. The model forecasts the tendencies of the potential temperature (θ), specific humidity (q) and horizontal component of wind due to turbulent mixing in the boundary layer. Soil moisture incorporated through the two layer soil model is to improve the accuracy of the surface fluxes. It is expected that the output of the model shall consist realistic values of surface fluxes for applications in simulation of boundary layer /local weather forecast/air pollution dispersion.

Keywords: Atmospheric Boundary Layer, Modelling of Land Surface Processes, Land-Atmosphere Interactions

1. INTRODUCTION

The interaction between land surface and atmosphere is an important problem in climate system. Land surface provides energy, water vapour and other substance to the air, meanwhile, the atmosphere provides an external environment to the processes of meteorology, hydrology and ecology near land surface. The region where meteorological variables, such as wind velocity, temperature and humidity adjust from their values in the free-atmosphere to the boundary conditions at the Earth's surface is called the atmospheric boundary layer (ABL). ABL plays a vital role in the exchange processes of heat, moisture and momentum from surface aloft. The land surface processes have a significant impact on ABL processes.

The numerical modelling of land surface processes is an effective way to understand the physical mechanism of exchange processes in the ABL. The growing interest in the interactions between the atmosphere and the underlying earth's surface is driven by the realisation that the surface fluxes determine, the steady state of the atmosphere to a great extent. Development of adequate parameterisation schemes for different physical processes have, therefore become an important area of research. The surface fluxes are of crucial importance not only because they determine the steady state of the atmosphere but also because they determine the mean profiles of the surface layer and the atmospheric boundary layer. The fluxes further play an important role in evolution of the weather and climate system.

A relatively simple one dimensional model of Atmospheric Boundary Layer(ABL) is described here, in which a soil atmosphere system that incorporates the interactions between soil and atmosphere in the parameterisation of surface fluxes is also included.

2. ATMOSPHERIC BOUNDARY LAYER MODEL

2.1 Prognostic Equations

The model forecasts the tendencies due to turbulent mixing of the potential temperature (θ), specific humidity (q) and horizontal component of the wind (u and v). To simplify presentation, only the vertical diffusion terms due to boundary-layer turbulent mixing and the advection terms due to a prescribed vertical motion field are considered in the prognostic equations (Troen and Mahrt, 1986).

The prognostic equations for atmospheric boundary layer are

$$\frac{\partial Vh}{\partial t} = \frac{\partial}{\partial z}\left(K_m \frac{\partial Vh}{\partial z}\right) - W\left(\frac{\partial Vh}{\partial z}\right) \tag{1}$$

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left(K_h\left[\frac{\partial \theta}{\partial z} - \gamma\theta\right]\right) - W\left(\frac{\partial \theta}{\partial z}\right) \tag{2}$$

$$\frac{\partial q}{\partial t} = \frac{\partial}{\partial z}\left(K_h \frac{\partial q}{\partial z}\right) - W\left(\frac{\partial q}{\partial z}\right) \tag{3}$$

where v, θ, q and h are the mean wind, potential temperature, specific humidity and boundary layer height respectively and K_m and K_h are the diffusivity coefficients for momentum and heat. γ_θ (Km^{-1}) is the counter gradient correction for the potential temperature required for weakly stable conditions of surface layer.

2.2 Parameterisation of γ_θ , K_m

This γ_θ is parameterised for the boundary layer as

$$\gamma_\theta = \begin{cases} 0 & for \quad stable \\ C\dfrac{(w'\theta')_s}{w_s h} & for \quad unstable \end{cases} \tag{4}$$

C a non dimensional constant and Ws is a velocity scale, given by

$$w_s = u_* \phi_m^{-1}\left(\frac{Z_s}{L}\right) \tag{5}$$

Here u_* surface friction velocity, Z_s is top of the surface layer , L is Monin-Obukhov length, and ϕ_m is a non dimensional profile function.

In free convection conditions as $V \to 0$, $u_* \to 0$, W_s takes the form

$$W_s \to \left[\frac{7Z_s gk(w'\theta'_v)_s}{\theta_{sv}}\right]^{1/3} \tag{6}$$

The coefficient of diffusivity for momentum (K_m, m^2s^{-1}) is parameterised as

$$K_m = W_s hk\frac{z}{h}\left(1-\frac{z}{h}\right)^p \qquad \text{(stable)} \tag{7}$$

$$K_m = u_* \phi^{-1}\left(\frac{z}{L}\right) hk\frac{z}{h}\left(1-\frac{z}{h}\right)^2 \qquad \text{(unstable)} \tag{8}$$

The coefficient of eddy diffusivity for heat is related to K_m in terms of turbulent Prandtl number (Pr) as:

$$K_h = K_m . P_r^{-1} \tag{9}$$

with

$$P_r = \left[\frac{\phi_h\left(\frac{z}{L}\right)}{\phi_m\left(\frac{z}{L}\right)} + C \quad k\left(\frac{z}{h}\right) \right]_{z=z_s} \quad \text{unstable} \qquad (10)$$

which is determined as the value at the top of the surface layer (z_s= 0.1 h) using surface layer similarity theory. P_r is assumed to be a constant for stable and neutral cases.

The non-dimensional profile functions for shear and temperature gradients (Equation 10) are

$$\phi_m = \begin{cases} 6.0 & \text{(very stable)} \\ 1.0 + 5.0\, z/L & \text{(stable)} \\ (1.0 - 15\, z/L)^{-1/3} & \text{(unstable)} \end{cases} \qquad (11)$$

and

$$\phi_h = \begin{cases} 6.0 & \text{(very stable)} \\ 1.0 + 5.0\, z/L & \text{(stable)} \\ (1.0 - 15\, z/L)^{-1/2} & \text{(unstable)} \end{cases} \qquad (12)$$

These are the functions of height coordinate (Z) and the Monin Obukhov length scale (L). The Monin Obukhov length scale (L) is defined in terms of surface virtual potential temperature (θ_{sv}) and virtual heat flux at surface $(w'\theta'_v)_s$ and frictional velocity u_* as

$$L = -\frac{\theta_{SV} u_*^3}{gk\left(w'\theta'_v\right)_S} \qquad (13)$$

2.3 Determination of boundary layer height

The boundary layer height (Z_i) as a base of low level inversion is determined as

$$Z_i = \frac{Ri_{cr}\theta_{ov}\left|V(h)\right|^2}{g(\theta_v(h) - \theta_{ov}^*)} \tag{14}$$

where Ri_{cr}, is a non dimensional critical Richardson number, θ_{ov}, the reference Virtual potential temperature at Ist model level, the $\theta_v(h)$, virtual potential temperature at model level h, V (h), horizontal wind velocity at level h, g(m s^{-2}), acceleration due to gravity. This approach to determine boundary layer height requires the specification of a low-level potential temperature (θ^*_{ov}) which is given by

$$\theta_{ov} \qquad \text{(stable)}$$

$$\theta^*_{ov} = \theta_{ov} + C\frac{(\overline{w'\theta'_v})_s}{w_s} \qquad \text{(unstable)} \tag{15}$$

When the boundary layer is unstable, the virtual potential temperature at the top of the surface layer is enhanced by thermal effects in an amount that is proportional to the surface sensible heat flux. In the neutral limit as w_s approaches u_*. The correction to the surface temperature vanishes so that $\theta^*_{ov} \rightarrow \theta_{ov}$ with the result that the modified bulk Ri number reduces to the usual one.

3. SURFACE LAYER MODEL

The counter gradient correction (γ_θ) is evaluated in terms of the surface flux of potential temperature. The surface fluxes are parameterised for stable case following Mahrt (1987) and for unstable case following Louis et.al. (1982) modified by Holtslag and Beljaars, (1989) as

$$u_*^2 = C_m\left|V_0\right| \tag{16}$$

$$(\overline{w'\theta'})_s = C_h(\theta_s - \theta_o) \tag{17}$$

$$(\overline{w'q'})_s = C_h(q_s - q_o) \tag{18}$$

where C_m and C_h are the exchange coefficients for momentum and heat respectively (ms^{-1}). $|V_0|$ is the wind speed evaluated at the first model level above the surface. The potential temperature (θ_o) and specific humidity (q_o) are taken as their representations at the first model level while the potential temperature (θ_s) and specific humidity (q_s) at the surface are obtained from the surface energy balance.

The surface exchange coefficients are defined as

$$C_m = k^2 |V_o| \frac{F1(z, z_{oM}, Ri_B)}{\left(\ln\left(\frac{z}{z_o}\right)\right)^2} \tag{19}$$

$$C_h = \frac{k^2}{R} |V_o| \frac{F2(z, z_{oM}, z_{oH}, Ri_B)}{\ln\left(\frac{z}{z_o}\right) \ln\left(\frac{z}{z_{oH}}\right)} \tag{20}$$

where k is von Karman's constant (0.4), R is the ratio of the drag coefficients for momentum and heat in neutral limit estimated as 1.0 (Holtslag and Beljaars, 1989). C_m and C_h are the functions of wind speed $|V_0|$, evaluated at first model level above the surface, height of the first model level (z), roughness length (z_{oM}) which depends on surface characteristics, and Ri_B the bulk Richardson number for surface layer. C_h is also a function of the roughness length for heat (z_{oH})

The function F1 in equation (19) is defined as

$$- e^{-aRi_B} \qquad \text{for stable}$$

$$F1 = 1 - \frac{10\ Ri_B}{1 + 75\left[\frac{k^2}{\ln(z/z_{oM})^2} 10\right] \left[-Ri_B \frac{z}{z_{oM}}\right]^{1/2}}, \text{ unstable}$$

(21)

similarly, the function F2 in equation (20) as:

$$F2 = \begin{cases} e^{-aRi_B} & \text{for stable} \\ 1 - \dfrac{15\ Ri_B}{1 + 7.5\left[\dfrac{k^2}{\ln(z/z_{oM})\ln(z/z_{oH})}10\right]\left[-Ri_B \dfrac{z}{z_{oM}}\right]^{1/2}}, & \text{unstable} \end{cases} \quad (22)$$

where a is a constant. In the surface layer the bulk Richardson number is defined as

$$Ri_B = \frac{gz(\theta_{ov} - \theta_{sv})}{\theta_{ov}|V_o|^2} \quad (23)$$

The tendency equations of the surface layer are the same as those of boundary layer except for eddy diffusivities which are given by:

$$K_m = u_* kz\Phi_m^{-1}\left(\frac{z}{L}\right)\left(1 - \frac{z}{h}\right)^p \quad (24)$$

$$K_h = K_m P_r^{-1} \quad (25)$$

q_s and θ_s the only variables to close the surface layer model, can be obtained through soil model and surface energy balance.

4. SOIL MODEL

4.1 Total evaporation:

Total evaporation (E) is obtained by summing up the direct soil evaporation (E_{dir}), transpiration (E_t) and the canopy transpiration(E_c).

$$E = E_{dir} + E_t + E_c \tag{26}$$

The direct evaporation (E_{dir}, ms^{-1}) at the air-soil interface (Z=0) is given by

$$E_{dir} = \left[-D(\Theta)\left(\frac{\partial \Theta}{\partial z}\right)_o - K(\Theta_o) \right](1-\sigma_f) + I(1-\sigma_f) \tag{27}$$

The coefficients of diffusivity (D, m^2s^{-1}) and hydraulic conductivity (K, ms^{-1}) are functions of the volumetric water content (Θ) (Mahrt and Pan, 1984). Through extremes of wet and dry conditions, the coefficients D and K can vary by several orders of magnitude and therefore cannot be treated as constants. I is the infiltration rate (ms^{-1}), σ_f is the plant shading factor (between 0 and 1). Θ is the non dimensional volumetric water content.

The prognostic equation equation for the nondimensional volumetric water content (Θ)

$$\frac{\partial \Theta}{\partial t} = \frac{\partial}{\partial z}\left(D(\Theta)\frac{\partial \Theta}{\partial z}\right) + \frac{\partial K(\Theta)}{\partial z} \tag{28}$$

The canopy evaporation of free water (E_c) is formulated as

$$E_C = E_p \sigma_f \left(\frac{C^*}{S'}\right)^n \tag{29}$$

where S'(m) is the saturation water content for a canopy surface, a connstant, C^* is the canopy water content which is increased by precipitation and reduced by evaporation. The potential evaporation is used to compute canopy transpiration as well as evaporation through transpiration which in turn is used to calculate total evaporation. n is (nondimensional) taken to be 0.5 (Pan and Mahrt, 1987),

The potential evaporation can be evaluated by surface energy balance.

4.2 Surface Energy Balance and potential evaporation

At the surface, the energy balance is obtained from the surface energy partitions as:

$$(1-\alpha)S\downarrow + L\downarrow - \sigma\theta_S^4 = G + H + L.E \tag{30}$$

where each term is expressed in Wm^{-2}. α is the non dimensional coefficient is the surface albedo and is a function of surface characteristics, s is the downward solar radiation (defined as positive downward). L is the downward atmospheric radiation (positive downward) and $\sigma\ \theta_s^4$ is the terrestrial radiation (positive upward), the coefficient σ is the Steafan-Boltzmann constant (5.6696×10^8 W m^{-1} K). G is the soil for ground heat flux (positive downwards), H is the sensible heat flux and LE is the latent heat flux. The last term LE is the latent heat flux where L is the latent heat of phase change (W m^{-2}) , The soil or ground heat flux G is defined as

$$G = K_T(\Theta)\left(\frac{\partial T}{\partial z}\right)_{z=0} \tag{31}$$

where Θ is the volumetric water content as defined above, K_T is thermal conductivity ($Wm^{-1}K^{-1}$) and T is soil temperature.

Soil thermodynamics is treated with a prognostic equation for soil temperature (T) such that

$$C(\Theta)\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left[K_T(\Theta)\frac{\partial T}{\partial z}\right] \tag{32}$$

where C is volumetric heat capacity (C, J m^{-3} k^{-1}) of the soil, both C and K_T are functions of the soil water content (Θ), where as the coefficient of thermal conductivity, (K_T) is a highly non-linear function of Θ and increases by several orders of magnitude from dry to wet soil conditions.

The sensible heat flux is calculated as

$$H = \rho_O C_p C_h (\theta_s - \theta_o) \quad (33)$$

which is a function of the air density (ρ_o), the specific heat for air, the heat exchange coefficient (C_h, as given in equation 20) the difference between the surface temperature (θ_s) and the air potential temperature at the first model level (θ_o).

The potential evaporation (E_p) is used to calculate the total evaporation.

This can be evaluated using surface energy balance (equation 30) as

$$(1-\alpha)S\downarrow + L\downarrow - \sigma\theta_s^4 = G + H + L_V E_p \quad (34)$$

$\sigma\theta_s^4$ the upward longwave radiation is linearised as

$$\theta_s^4 = \sigma T_O^4 \left(1 + 4 \left[\frac{(\theta_s - T_o)}{T_O} \right] \right) \quad (35)$$

G,H, given in equation 31 and 33 respectively.

For a saturated surface, replacing temperature appropriate for the potential evaporation, the sensible heat flux can be

$$H' = \rho C_p C_h (\theta_s - \theta_o) = \rho C_p C_h [(\theta_s - T_o) - (\theta_o - T_o)] \quad (36)$$

Then the surface energy balance equation can be put as

$$(1-\alpha)S\downarrow + L\downarrow - \sigma T_O^4 - 4\sigma T_O^4 \left(\frac{\theta_s - T_o}{T_o} \right)$$

$$= G + \rho C_p C_h [(\theta_s - T_o) - (\theta_o - T_o)] + L_v E_p \quad (37)$$

solving for $L_V E_p$

$$L_v E_p = (1-\alpha)S\downarrow + L\downarrow - \sigma T_O^4 - G + \rho C_p C_h(\theta_O - T_O)$$

$$-(\theta_s - T_O)\left[\frac{4\sigma T_O^4}{T_O} + \rho C_p C_h\right] \tag{38}$$

The latent heat flux is approximated as

$$L_v E_p = \rho \ \ L_v C_h(q_s^* - q_O) \tag{39}$$

$$= \rho L_v C_h((q^*_s - q^*_0) + (q^*_o - q_o)) \tag{40}$$

$$= \rho L_v C_h\left[\left(\frac{q_s^* - q_O^*}{\theta_s - T_O}\right)(\theta_s - T_O) + (q_O^* - q_O)\right] \tag{41}$$

$$\approx \rho L_v C_h\left[\left(\frac{dq^*_s}{dT}\right)(\theta_s - T_O) + (q_O^* - q_O)\right] \tag{42}$$

where q_s^* is the surface saturation specific humidity. q_0 and q_0^* are the actual and saturation specific humidities at the first model level respectively. Solving for $\theta_s - T_o$ from above expression

$$\theta_s - T_O = \frac{\left[\dfrac{L_v E_p}{\rho L_v C_h} - (q_O^* - q_O)\right]}{\dfrac{dq^*}{dT}} \tag{43}$$

substituting the above expression for $\theta_s - T_o$ in equation (38), where $\rho\, T_o \approx \rho_{sfc} / R_d$, and solving for E_p

$$E_p = \left[\frac{\left(\frac{R_n}{\rho C_p C_h} + (\theta_o - T_o) \right) \Delta + (r+1) A}{\Delta + r + 1} \right] \frac{\rho C_p C_h}{L_v} \qquad (44)$$

$$R_n = (1-\alpha) S\downarrow + L\downarrow - \sigma T_o^4 - G$$

$$\Delta = \frac{dq_s}{dT}^{*} \frac{L_v}{C_p} \qquad\qquad r = \frac{4\sigma T_o^4 R_d}{P_{sfc} C_p C_h}$$

$$A = (q_o^* - q_o) \frac{L_v}{C_p}$$

Incorporation of evaporation through transpiration (E_t) is in the following manner

$$E_t = E_p \sigma_f k_v \frac{\sum_{t=1}^{2} \left[\Delta z_i g(\Theta_i) \right] \left[1 - \left(\frac{C^*}{S'} \right)^n \right]}{\sum_{t=1}^{2} \left[\Delta z_i \right]} \qquad (45)$$

where k_v is a nondimensional plant resistance factor or plant coefficient (PC) with a value between 0 and 1. The canopy resistance (RC) may be used instead of the plant coefficient. The non dimensional transpiration rate function $g(\Theta)$ is given by

$$g(\Theta) = \begin{cases} 1, & \Theta \ge \Theta_{ref} \\ \frac{\Theta - \Theta_{wilt}}{\Theta_{ref} - \Theta_{wilt}}, & \Theta_{ref} \ge \Theta > \Theta_{wilt} \\ 0, & \Theta_{wilt} \ge \Theta \end{cases}$$

(46)

Θ_{ref}, Θ_{wilt} are the transpiration limits to an upper reference value which is Θ value where transpiration begins to decrease due to a deficit of water and the plant wilting factor which is the θ value where transpiration stops (Mahrt and Pan, 1984).

4.3 Determination of surface specific humidity

The surface specific humidity (q_s) is calculated from

$$q_S = q_O + \frac{E}{\rho_O C_h} \tag{47}$$

where q_0 is the specific humidity at first model level, ρ_0 is the air density at surface (kg m^{-3}), C_h is the exchange coefficient for moisture and E is the total evaporation.

4.4 Computation of Surface Temperature

The surface potential temperature (θ_s) also is obtained by the surface energy balance equation (30) in which the potential evaporation E_p is substituted for the actual evaporation E as a fraction of it such that

$$(1-\alpha)S\downarrow + L\downarrow - \sigma\theta_S^4 = G + H + \beta \quad L_V E_p \tag{48}$$

or

$$(1-\alpha)S\downarrow + L\downarrow - \sigma T_O^4 - 4\sigma T_O^4\left(\frac{\theta_S - T_O}{T_O}\right)$$

$$= G + \rho\, C_p C_h\left[\left(\theta_S - \theta_O\right) - \left(\theta_O - T_O\right)\right] + \beta \quad L_V E_p \tag{49}$$

writing $F = (1 - \alpha)S\downarrow + L\downarrow$

and $$G = K_T(\Theta)\frac{\left(\theta_S - T_{1soil}\right)}{\Delta z}$$

The surface energy balance equation can take the form

$$\frac{F - \sigma T_O^4}{\rho C_p C_h} - \frac{K_T(\Theta)(\theta_s - T_{soil})}{\rho C_p C_h \Delta z} = (r+1)(\theta_s - T_O) - (\theta_O - T_O) + \frac{\beta L_v E_p}{\rho C_p C_h}$$

(50)

combining terms and solving for θ_s

$$\theta_s = \frac{T_0 + \dfrac{\dfrac{F - \sigma T_O^4}{\rho C_p C_h} + (\theta_O - T_O) - \dfrac{\beta L_v E_p}{\rho C_p C_h}}{(r+1)} + \dfrac{K_T(\Theta) T_{1soil}}{\Delta z \rho C_p C_h (r+1)}}{1 + \dfrac{K_T(\Theta)}{\Delta z \rho C_p C_h (r+1)}}$$

(51)

4.5 External Radiational Forcing

The total downward radiation is a combination of incoming solar (short wave) and downward atmospheric (long wave) radiation and is obtained as follows:

$$S\downarrow \;=\; [1-(1-t)\,CLC^n]S_{cs}\downarrow \qquad (52)$$

where $S\downarrow$ is the net incoming solar radiation (below clouds but above the ground), t is a fraction dependent on the solar radiation transmitted through the clouds which depends on sun angle following Liou (1976), CLC is the fractional cloud cover, n is an empirically derived coefficient , and $S_{cs}\downarrow$ is the clear sky solar radiation adjusted for solar elevation. When n = 1, t is the actual fraction of solar radiation transmitted through the clouds.

$$L\downarrow \;=\; \varepsilon\sigma\, T^4_{ref} + c_2\, CLC \qquad (53)$$

where $L\downarrow$ is the downward atmospheric radiation ($W\,m^{-2}$); ε is the emissivity of the atmosphere, a function of the temperature and moisture at the reference level, T^4_{ref} is the temperature at the reference height, CLC is the fractional cloud cover, and c_2 is an empirically derived constant equal to 60 $W\,m^{-2}$. The expression is a parameterised form following Scatterlund (1979).

5. Summary

The one-dimensional Atmospheric Boundary Layer model, briefly described above is a combination of a boundary layer model and a two layer soil model. The equations used in this composite model are comprehensive to approximate the important physical processes under a variety of diverse atmospheric conditions. The equations of the model describe the evolution of the thermodynamic properties (e.g. the potential temperature θ and the specific humidity q) due to the exchange of heat, water vapor and momentum at the earth's surface and due to entrainment of air from above the boundary layer.

The soil model gives direct evaporation (E_{dir}), considering the soil moisture at two levels. The potential evaporation (E_p) could be computed using radiative fluxes, wind speed, moisture deficit and atmospheric stability. Different finite differencing techniques could be employed for the numerical simulation of each of the different physical processes depending on the stability and the characteristics of the terms.

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List of symbols

ρ	Air density at surface (kg m^{-3})
C_h	Exchange coefficient for heat
C_m	Exchange coefficient for momentum
E	Total evaporation
E_{dir}	Direct evaporation
E_p	Potential evaporation
E_t	Evaporation due to transpiration
E_c	E Evaporation due to canopy
Θ	Volumetric water content
K	Coefficient of diffusivity (m^2 s^{-1})
D	Hydraulic conductivity (m s^{-1})
G	Ground heat flux
s	Solar radiation
α	Albedo
$g(\Theta)$	Non-dimensional transpiration rate
v	Mean wind (m s^{-1})
θ	Potential temperature (^{0}K)
q	Specific humidity (gm kg^{-1})
γ	Counter gradient correction
k_m	Coefficient of diffusion for momentum
k_h	Coefficient of diffusion for heat
$w'\theta'$	Surface heat flux
w_s	Velocity scale
φ_m	Non -dimensional profile function for shear
φ_h	Non -dimensional profile function for heat
u^*	Frictional velocity (ms^{-1})
L	Monin Obhukhov length
k	Von Karman" constant
R_i	Richardson number
h	Boundary layer height
P_r	Prandtl number