SENSITIVITY TESTS OF INTERACTION BETWEEN LAND SURFACE PHYSICAL PROCESS AND ATMOSPHERIC BOUNDARY LAYER*

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ABSTRACT

In this paper, an interactive model between land surface physical process and atmosphere boundary layer is established, and is used to simulate the features of soil environmental physics, surface heat fluxes, evaporation from soil and evapotranspiration from vegetation and structures of atmosphere boundary layer over grassland underlying. The sensitivity experiments are engaged in primary physics parameters. The results show that this model can obtain reasonable simulation for diurnal variations of heat balance, soil volumetric water content, resistance of vegetation evaporation, flux of surface moisture, and profiles of turbulent exchange coefficient, turbulent momentum, potential temperature, and specific humidity. The model developed can be used to study the interaction between land surface processes and atmospheric boundary layer in city regions, and can also be used in the simulation of regional climate incorporating a mesoscale model.

Key words: surface process parameterized, soil environmental physics., surface heat fluxes, atmosphere boundary layer structure

I. INTRODUCTION

Physical conditions of continent surface, such as soil moisture, vegetation coverage, vegetation leaf area index and reflectivity, can directly affect mass exchanges and energy exchanges between surface layer and atmosphere, and then affect the construction of atmosphere boundary layer, atmosphere circulation and climate consequently. Therefore either in climate models or in atmosphere boundary layer models, their effects must be considered. In the research of models of micro-meso scale climate and atmosphere boundary layer over different underlying surfaces, they are especially important.

Mass exchanges and energy exchanges between soil, vegetation and atmosphere and interaction process between earth and atmosphere are immensely important for the development of boundary layer. Especially, radiation fluxes, momentum fluxes, sensitive heat fluxes and latent heat fluxes affect the movement of atmosphere and the fields of

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temperature. moisture and precipitation. and then have strong feedback functions on the source and congruence of these physical quantities.

Since Deardorff (1978) brought forward the parameterization of earth-atmosphere interaction, many scientists have done a lot of work on it. Dickinson (1984) put forward a surface model with consideration of effects of vegetations first, and he (Dickinson 1986) consummated the model and developed Biosphere-Atmosphere Transfer Scheme (BATS). At the same time, Sellers et al. (1986) developed a Simple Biosphere model (SiB) based on Dickinson's vegetation model. In addition. Mccumber and Pielke's (1981) multi-layer soil model have been widely used in models of atmosphere circulation and micro-meso scale climate. Different models have developed due to observation data of different territories, so they emphasize particular points. While the common point is that they are all used the method of delamination.

Comparing with the model of NP-89 (Noilhan and Planton 1989), the parameterization method of surface process brought forward in this paper has less parameters. and is more simple and applied. It merely includes some physical parameters of vegetation and soil layer. without consideration of moisture flux between soil and plant root system, and moisture flux on the interface of free-water layer and soil layer. Parameters in this model are defined due to vegetation feature surface of the surface researched and references (Deardorff 1978; Dickinson 1984; 1986; Sellers et al. 1986; Mccumber and Pielke 1981; Noilhan and Planton 1989). This model can be used not only in the interaction of usual continent surface and atmosphere. but the interaction of city surface physical process and atmosphere boundary layer. It can also be coupled with mesoscale atmosphere model in the research of regional climate.

II. MODEL

1. *Basic Equations*

System of equations of two-dimensional numerical model of atmosphere boundary layer is

$$\frac{\partial u}{\partial t} = - u\frac{\partial u}{\partial x} - w\frac{\partial u}{\partial z} - \theta\frac{\partial \pi}{\partial x} + F_u, \tag{1}$$

$$\frac{\partial \theta}{\partial t} = - u\frac{\partial \theta}{\partial x} - w\frac{\partial \theta}{\partial z} + F_\theta, \tag{2}$$

$$\frac{\partial \pi}{\partial z} = - \frac{g}{\theta}, \tag{3}$$

$$\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0, \tag{4}$$

$$\frac{\partial R}{\partial t} = - u\frac{\partial R}{\partial x} - w\frac{\partial R}{\partial z} + F_R, \tag{5}$$

where g is acceleration of gravity; x, z are horizontal and vertical coordinates respectively; u, w are horizontal and vertical wind velocities respectively; θ is potential temperature (K), $\theta=T(\frac{p_0}{p})^{0.286}$; R is specific humidity ($kg \cdot kg^{-1}$); π is Exner function, $\pi=c_p(\frac{p}{p_0})^{0.286}$, with c_p being specific heat at constant pressure. ($1005\ J \cdot K^{-1} \cdot kg^{-1}$); p

being atmospheric pressure, and usually p_0 being to 1000 hPa; F_u, F_R, F_θ are turbulence terms, and φ is used to present u, R and θ respectively. And we get

$$F_\varphi = K_{\mathrm{H}} \frac{\partial^2 \varphi}{\partial x^2} + \frac{\partial}{\partial z}(K_z \frac{\partial \varphi}{\partial z}), \tag{6}$$

where K_{H} is horizontal turbulence exchange coefficient, and it is adopted 1. 0 in this model; K_z presents K_m, K_R and K_θ corresponding to u, R and θ, and usually $K_R = K_\theta$. Quantity of K_m will be discussed in the following text.

2. *Parameterization of Boundary Layer Turbulence*

In this model, K-model is used to parameterize kinetic energy of boundary layer. The hypothesis of mixing length is introduced, so turbulence exchange coefficients and turbulence kinetic energy are related (Yamada 1983):

$$\frac{\partial q^2}{\partial t} = - u\frac{\partial q^2}{\partial x} - w\frac{\partial q^2}{\partial z} + K_{\mathrm{H}}\frac{\partial q^2}{\partial x^2} + \frac{\partial}{\partial z}(K_{q^2}\frac{\partial q^2}{\partial z}) + K_m(\frac{\partial u}{\partial z})^2 - K_\theta \frac{g}{\theta}\frac{\partial \theta}{\partial z} - \frac{B_1 q^3}{L}, \tag{7}$$

where $q^2 = 1/2\,(\overline{u'^2} + \overline{w'^2})$, and q^2 is turbulence kinetic energy; $B_1 q^3/L$ is turbulence dissipation rate, B_1 empirical constant, is adopted 0. 25 in this paper; according to Yamada's empirical equations (Yamada 1983), we have

$$L = \frac{KZ}{(1 + \frac{KZ}{L_\infty})}, \tag{8}$$

$$L_\infty = 0.1 \frac{\int_0^\infty q^2 Z \mathrm{d}z}{\int_0^\infty q^2 \mathrm{d}z}, \tag{9}$$

$K_{q^2} = K_m$, $K_m = 0.5L(q^2)^{1/2}$, $K_{\mathrm{H}} = K_R = K_\theta = 1.35\, K_m$, where K is Karman constant, adopted 0. 4 in this paper.

3. *Parameterization of Surface Energy Balance*

Direct solar radiation flux accepted by the earth's surface is (Kondratyev 1969)

$$Q_K = (t - a)(1 - \alpha)S_0 \cos Z, \tag{10}$$

where $t = 1.03 - 0.08[\frac{(0.000949P + 0.051)}{\cos Z}]^{\frac{1}{2}}$. $P = 980$ hPa ; a is the atmospheric absorption coefficient for the whole solar spectrum, calculated with the following equation:

$$a = \frac{2.9\delta p}{(1 + 141.5\delta p)^{0.635} + 5.925\delta p}, \tag{11}$$

where δp is atmospheric vapor optical thickness, and in this paper $\delta p = \int_0^\infty \rho_w R \mathrm{d}z \approx 10$; S_0 is the solar constant, adopted 1367 W m^{-2};

$$\cos Z = \sin \varphi \sin \delta + \cos \varphi \cos \delta \cos h, \tag{12}$$

where φ geographic latitude, adopted 45. 6°N; δ is solar declination, and

$$\delta = -23.5\frac{\pi}{180}\cos\left(\frac{2\pi(t_j + 16)}{365}\right)\ \text{(radian)}, \tag{13}$$

where t_j represents the t_jth day in a year. and in this paper $\delta = 19°$; h is hour angle. and $h = \pi\frac{12-t_t}{12}$ (radian); t_t is the time in a day. α is surface reflectivity. adopted 0.15; t is an empirical parameter. put forward by Kondratyev with consideration of diffuse reflection of the sky (Kondratyev 1969).

Long wave radiation flux accepted by the earth's surface can be written as (Kondratyev 1969)

$$Q_R = \varepsilon_a \sigma T_a^4 - \varepsilon_s \sigma T_s^4, \tag{14}$$

where σ is Stefan-Boltzmann constant. $\sigma = 5.68\times10^{-8}\ \mathrm{W\cdot m^{-2}\cdot K^{-4}}$; T_a is the atmospheric temperature of referenced layer. which is defined at 10 m height in this paper; T_s is the earth's surface temperature; ε_s is the long wave radiation coefficient of the earth's surface. adopted as 0.98; ε_a is long wave radiation coefficient of atmosphere. calculated by empirical equation below:

$$\varepsilon_a = k e_a^{\frac{1}{7}}, \tag{15}$$

where k is an empirical constant. adopted as 0.4 in this paper; e_a is the vapor pressure of referenced layer. calculated by

$$e_a = \frac{\gamma}{0.622+\gamma}p, \tag{16}$$

where p is atmospheric pressure and γ is mixed specific humidity.

Therefore, net radiation flux accepted by the earth's surface is (Kondratyev 1969)

$$R_n = Q_K + Q_R. \tag{17}$$

In addition. considering heat balance. we have

$$R_n = H + L_E + G. \tag{18}$$

In the equation. H, L_E and G are respectively sensitive heat flux. latent heat flux and soil heat flux between earth and air.

4. *Parameterization of Surface Temperature, Moisture and Heat Fluxes*

(1) *Parameterization of surface temperature* (*Noilhan and Planton* 1989)

Equations of surface temperature parameterization are

$$\frac{\partial T_s}{\partial t} = C_T G - \frac{2\pi}{\tau}(T_s - T_2), \tag{19}$$

$$\frac{\partial T_2}{\partial t} = \frac{1}{\tau}(T_s - T_2), \tag{20}$$

where T_s and T_2 are the surface temperature and the surface daily mean temperature respectively; G is soil heat flux. $G = R_n - H - L_E$; τ is time constant. adopted as 86400 s in this paper. C_T is the thermodynamic coefficient of soil. calculated by

$$C_T = \frac{1}{\dfrac{1-V_{eg}}{C_G} + \dfrac{V_{eg}}{C_V}}, \tag{21}$$

where $C_V = 10^{-3}\ \mathrm{K\cdot m^{-2}\cdot J^{-1}}$; $C_G = 0.6C_{Gsat}\left(\frac{w_{sat}}{w_2}\right)^{\frac{b}{2\ln 10}}$; V_{eg} is vegetation coverage, and when

$V_{eg}=1$ from Eq. (21) we can get $C_T=C_V$, so when $V_{eg}=1$, C_v represents thermodynamic coefficient of the surface; C_{Gsat} is the heat transfer coefficient of saturated soil, adopted as 3.593×10^6 (K • m^2) J^{-1} in this paper; w_{sat} is the volumetric water content of saturated soil; adopted as 0.532 m^3 • m^{-3}; and w_2 is daily mean soil volumetric water content.

(2) *Parameterization of surface moisture* (*Deardorff* 1978; *Noilhan and Planton* 1989)

Equations of surface moisture parameterization are

$$\frac{\partial w_g}{\partial t}=\frac{C_1}{\rho_w d_1}(p_g-E_g)-\frac{C_2}{\tau}(w_g-w_{geq}) \qquad 0\leqslant w_g\leqslant w_{sat}, \tag{22}$$

$$\frac{\partial w_2}{\partial t}=\frac{1}{\rho_w d_2}(p_g-E_g-E_{tr}) \qquad 0\leqslant w_2\leqslant w_{sat}, \tag{23}$$

$$\frac{\partial w_r}{\partial t}=V_{eg}P-(E_v-E_{tr})-R_r \qquad 0\leqslant w_r\leqslant w_{r\max}, \tag{24}$$

where ρ_w is the density of liquid water; p_g is liquid water flux reaching the soil surface, $p_g=0$ in this paper; P is precipitation, $P=0$ in this paper; R_r is the runoff volume of damming off precipitation, $R_r=0$ in this paper; E_g evaporation capacity of the soil; E_{tr} is physiological evapotranspiration flux of vegetation surface; E_v is the total of physiological evapotranspiration flux and leaf surface evaporation flux of the vegetation; w_{geq} is the volumetric water capacity of the soil surface when gravity balances tensile force of capillaries, i.e.

$$w_{geq}=w_2-a\left(\frac{w_2}{w_{sat}}\right)^p\left(1-\frac{w_2}{w_{sat}}\right)^{8p}w_{sat}; \tag{25}$$

$w_{r\max}$ is the maximum of leaf surface damming off precipitation, $w_{r\max}=0.2P_{Veg}L_{AI}$ (mm); d_1 is the thickness of soil layer related with w_g (volumetric water capacity of soil surface), adopted as 0.01 m; d_2 is the thickness of soil layer related with w_2 (daily average volumetric water capacity of soil), adopted as 1 m; a and p are soil coefficients, and their quantities are referenced to NP-89 model (Noilhan and Planton 1989); coefficients C_1 and C_2 are calculated with equations below

$$C_1=C_{1\,sat}\left(\frac{w_{sat}}{w_g}\right)^{\frac{b}{2}+1}, \tag{26}$$

$$C_2=C_{2\,ref}\left(\frac{w_2}{w_{sat}-w_2+w_{ft}}\right), \tag{27}$$

where w_{ft} is a delta value, which makes the above equations valuable when the soil is saturated, and w_{ft} is adopted as 0.05 in this paper.

(3) *Parameterization of surface heat fluxes*

Equation for surface sensible heat flux is (Noilhan and Planton 1989)

$$H=\rho_a c_p\frac{(T_s-T_a)}{R_a}, \tag{28}$$

where T_s is the earth's surface temperature; c_p is the specific heat at constant pressure of air, $c_p=1005$ J • kg^{-1} • K^{-1}; R_a is aerodynamic resistance, $R_a=\frac{1}{C_H|u_a|}$; u_a, ρ_a and T_a are respectively wind velocity, air density and air temperature of referenced layer, and in this paper the altitude of referenced layer is $Z_a=10$ m; C_H is drag coefficient, which is $4.2\times$

10^{-3} in this paper.

Equation for surface latent heat flux is (Noilhan and Planton 1989)

$$L_E = L(E_g - E_v), \tag{29}$$

where L is the latent heat of vaporization of water. $L=2.5\times10^6$ J·kg^{-1}; E_g is the water vaporization flux of the earth's surface, and can be written as :

$$E_g = \frac{(1-V_{eg})\rho_a[h_u q_{sat}(T_s) - q_a]}{R_a}, \tag{30}$$

where h_u is the relative humidity of the earth's surface, and

$$h_u = 0.5[1-\cos(\frac{w_g}{w_{fc}}\pi)] \qquad w_g \geqslant w_{fc}, \tag{31}$$

$$h_u = 1 \qquad w_g \leqslant w_{fc}. \tag{32}$$

In Eq. (30), q_a is the saturation specific humidity of referenced layer; $q_{sat}(T_s)$ is the saturation specific humidity as the earth's surface temperature T_s. By Teten's equation we obtain

$$e_s(T_s) = 6.1\exp(17.269\frac{T_s - 273.16}{T_s - 35.86}), \quad (\text{hPa}) \tag{33}$$

$$q_{sat}(T_s) = 0.622\frac{e_s(T_s)}{p - 0.378e_s(T_s)}. \quad (\text{kg kg}^{-1}) \tag{34}$$

The total of physiological evapotranspiration flux and leaf surface evaporation flux E_v, evaporation of water intercepted by leaf surface E_r and transpiration from leaf surface E_{tr} are calculated respectively by the equations below:

$$E_r = V_{eg}\rho_a\frac{\delta}{R_a}[q_{sat}(T_s) - q_a], \tag{35}$$

$$E_{tr} = V_{eg}\rho_a\frac{1-\delta}{R_a+R_s}[q_{sat}(T_s) - q_a], \tag{36}$$

$$E_V = E_r + E_{tr} = V_{eg}\rho_a\frac{h_v}{R_a}[q_{sat}(T_s) - q_a], \tag{37}$$

where $h_v=\frac{(1-\delta)R_a}{R_a+R_s}+\delta$, with δ being leaf area coverage that dams off precipitation, and $\delta = (\frac{w_r}{w_{r\,max}})^{\frac{2}{3}}$; R_s is the impedance of earth's surface-vegetation system, which can be calculated from

$$R_s = \frac{R_{s\,min}}{L_{AI}}F_1F_2^{-1}F_3^{-1}F_4^{-1}. \tag{38}$$

Here F_1 is the solar shortwave radiation factor which is used for physiological action by leaf surface, and is calculated by

$$F_1 = \frac{1+f}{f+\frac{R_{s\,min}}{R_{s\,max}}}, \tag{39}$$

$$f = 0.55\frac{Q_K}{L_{Q_K}}\frac{2}{L_{AI}}. \tag{40}$$

In Eq. (40), L_{Q_K} is a constant, and for forest it can be adopted as 30 W m^{-2}, and for crop it can be adopted as 100 W m^{-2}, and in this paper it is adopted as 80 W m^{-2}; F_2 is a factor of soil humidity we have

$$F_2 = 1 \qquad w_2 > w_{fc}, \tag{41}$$

$$F_2 = \frac{w_2 - w_{wilt}}{w_{fc} - w_{wilt}} \qquad w_{Wilt} \leqslant w_2 \leqslant w_{fc}, \tag{42}$$

$$F_2 = 0 \qquad w_2 < w_{wilt}, \tag{43}$$

F_3 is a factor of air humidity, portray the effect of vegetation surface impedance by air water vapour, i. e.

$$F_3 = 1 - 0.06(q_{sat}[T_a) - q_a] \qquad q_{sat}(T_a) - q_a \leqslant 12.5 \text{ g/kg}^{-1}, \tag{44}$$

$$F_3 = 0.25 \qquad \text{otherwise}. \tag{45}$$

F_4 is a factor of air temperature, portray the effect of vegetation surface impedance by air temperature, i. e

$$F_4 = 1 - 1.6 \times 10^{-3}(T_o - T_a)^2, \tag{46}$$

where T_o is optimum air temperature of plant stoma open, in this model it is adopted $T_o=$ 298 K

III. INITIALIZATION, BOUNDARY CONDITIONS AND DATA CALCULATING METHODS OF THE MODEL

1. *Initialization of the Model*

This model supposed a grassland area with a horizontal dimension of 30 km, and the altitude of atmospheric boundaric layer is 4 km. Initial wind velocity, potential temperature, specific humidity and turbulence kinetic profile are defined according to observation results in fields. The proper beginning time is 0600 am. In the equations below, I presents horizontal grid point, and J presents vertical grid point.

Initial wind velocity profile (unit: m s^{-1}, negative sign represents direction):

$$u(I,J) = \begin{cases} -4.5\left(\dfrac{Z}{10.0}\right)^{0.14} & Z \leqslant 1500 \text{ m}, \\ u|_{Z=1500 \text{ m}} & 1500 \text{ m} \leqslant Z \leqslant 4000 \text{ m}. \end{cases} \tag{47}$$

Initial potential temperature profile (unit: K):

$$\theta(I,J) = \begin{cases} 298.0 & Z \leqslant 1000 \text{ m}, \\ 298.0 + \dfrac{1.5(Z-1000)}{100} & 1000 \text{ m} \leqslant Z \leqslant 1500 \text{ m}, \\ 305.5 + \dfrac{0.5(Z-1500)}{100} & 1500, \text{ m} \leqslant Z \leqslant 4000 \text{ m}. \end{cases} \tag{48}$$

Initial specific humidity profile (unit: g kg^{-1}):

Z(m)	0	10	20	50	80	100	150	200	250	300
R(g kg^{-1})	10.5	10.5	10.0	9.8	9.5	9.3	9.0	8.8	8.7	8.5

Z(m)	400	500	750	1000	1250	1500	2000	2500	3000	4000
R(g kg^{-1})	8.4	8.1	7.9	7.7	7.5	6.6	6.0	5.0	4.1	4.0

Initial turbulence kinetic energy ($Q = q^2$) (unit: $m^2 s^{-2}$):

$$q^2(I,J) = \begin{cases} 0.1 & Z \leqslant 1000\ \text{m}, \\ 0.05 & 1000\ \text{m} \leqslant Z \leqslant 1500\ \text{m}, \\ 0.01 & 1500\ \text{m} \leqslant Z \leqslant 4000\ \text{m}. \end{cases} \tag{49}$$

Initial daily average temperature of earth's surface (unit: K):

$$T_2(I) = 298.0.$$

Initial moisture of soil

$$w_g(I) = 0.410(\text{m}^3\,\text{m}^{-3}), \quad w_2 = 0.470(\text{m}^3\,\text{m}^{-3}), \quad w_r = 0.0(\text{kg} \cdot \text{m}^{-2}).$$

2. *Boundary Conditions of the Model*

Top border: $Z=4000$ m, $\pi = c_p(\frac{p}{1000})^{0.286} \approx 875.0$;

Bottom border: $U(I,J) = 0$;

Left border: $W(1,J) = W(2,J) = 0$, $T(1,J) = T(2,J)$;

Right border: $W(30,J) = W(29,J) = 0$, $T(30,J) = T(29,J)$.

3. *Data Calculating Methods of the Model*

Considering that near the surface the physical quantities change greatly, we choose the grid altitude following the principle that it is denser near the surface and gets thinner when goes upward. The scheme of difference is a temporal forward difference. Except that special advection items are run by central difference, other items are calculated with forward difference method. In order to keep the stability of difference scheme, the integral time step is set as $\Delta t = 10$ s. Because the initial turbulence kinetic energy in Eq. (49) is supposed, q^2 will gradually match the field of wind and temperature during the process of integration. Therefore, in the first hour of calculus the results are not completely dependable. Further more, because of boundary effect the results on boundary grids are not proper either. Results calculated from other grids and other periods are stable. Given in this paper are the results calculated at a horizontal grid where $I=15$.

IV. RESULTS AND DISCUSSIONS

Parameters of simulation of the interaction between surface physical process and atmosphere boundary layer and parameters of sensitivity experiments in this paper are set according to the characters of the surface vegetation researched by authors and references. All parameters are listed in Table 1. In order to have further research about the effect of vegetation coverage, leaf surface area index, earth's surface reflectivity and soil moisture on the interaction of surface physical process and atmosphere boundary layer, some sensitivity experiments are carried out. Parameters of the sensitivity experiments are listed in Table 2.

Table 1. Evaluation of Essential Variables and Parameters of the Model

Essential variable	Physical meaning	Unit or evaluation
T_s	earth's surface temperature	K
T_2	daily average surface temperature	K
w_g	volumetric water content of surface soil	0.41 $m^3\,m^{-3}$
w_2	daily average soil volumetric water content	0.47$m^3\,m^{-3}$
w_r	precipitation intercepted by leaf surface	kg m^{-3}
d_1	thickness of soil layer related to w_g	0.01 m
d_2	thickness of soil layer related to w_2	1 m
w_{sat}	volumetric water content of saturated soil	0.532 $m^3\,m^3$
w_{wilt}	soil volumetric water content at blasted vegetation point	0.23 $m^3 m^{-3}$
w_{fc}	soil moisture in the crop fields	0.451 $m^3\,m^{-3}$
V_{eg}	vegetation coverage	0.47
L_{AI}	leaf area index	1.9
α	reflectivity of earth's surface	0.17
C_{gsat}	heat exchange coefficient of saturated soil	3.593×10^6
C_{1sat}	parameter of soil type	0.213
C_{2ref}	parameter of soil type	0.8
a	coefficient of w_{geq} equation	1.35
p	coefficient of w_{geq} equation	6.0
b	coefficient of moisture detention curve	5.39
$R_{s\,min}$	minimum leaf surface impedance	40 m s^{-1}
$R_{s\,max}$	maximum leaf surface impedance	5000 m s^{-1}
C_H	drag coefficient	4.2×10^{-3}
T_o	optimal temperature for plants' stoma to open	298 K

Table 2. Parameters of Sensitivity Experiments

	V_{eg}	L_{AI}	α	w_g	w_2
Experiment 1	0.63	2.5	0.12	0.41	0.47
Experiment 2	0.45	1.8	0.18	0.41	0.47
Experiment 3	0.53	2.1	0.15	0.43	0.50

This model concerns about only daytime, so the net radiation (R_n) is positive at almost all the time. Only around the sunset when the long-wave radiation of earth's surface is prominent, is it possible that the net radiation is negative. Of all the heat fluxes, sensitive heat flux (H) can indicate the turbulence heat exchange conditions of atmosphere, and it is the characteristic quantity that shows heat transfer between earth's surface and atmosphere. Latent heat flux (L_E) reflects the status of vapor exchange between earth and atmosphere. This model simulated the conditions of grassland underlying surface of July and August when the soil moisture is very large. Over such a moist surface, evaporation flux of water and evapotranspiration are both very large. The transform from soil moisture to vapor consumes a majority of earth's surface heat, causing that in the heat balance equation the sensible heat flux decreases a lot relatively. And the strength of turbulence exchange is weakened, so the maximal altitude of turbulence mixed layer becomes lower, and the time it takes to reach the maximum is lengthened. Of all the net radiation items accepted by earth's surface, except that some are diffused into the atmosphere by the mode of sensible heat flux and latent heat flux, there still remains a part which is transferred downward into the earth, and this is the soil heat flux (G).

The results in Fig. 1 are calculated with the essential variables and parameters listed in Table 1. From Fig 1, we can see that net radiation is mainly affected by solar altitude, and its maximum appears at 1200 LT(local time). The maximums of latent heat flux and vapor flux at vegetation surface appear prior to that of net radiation, and decrease quickly after 1200 LT. This is because of the quick decrease of moisture in the soil surface layer, which is caused by evaporation and can be proved by the moisture reduction in Fig. 1b. Now plants begin self-protection, and close their stomas. Then the evaporation impedance of vegetation increases quickly, and its maximum appears at about 1300 LT. In addition, with the rapid decrease of latent heat flux, soil heat flux and sensible heat flux go up, and their maximums appear between 1300 LT and 1400 LT. Although soil surface moisture and vapor flux go down rapidly along with the increase of earth's surface temperature and air temperature, the root systems of plants can pump water from deep soil. Therefore, the physiological transpiration and the total amount of leaf surface traspiration of the vegetation reach their maximum at about 1500 LT (Fig. 1d).

The factor that affects sensible heat fluxes most is the temperature difference between referenced layer and earth's surface. From Fig. 1 and Fig. 2 we can find that if we increase vegetation coverage and leaf area index and decrease surface reflectivity with other parameters keeping constant, then the net radiation and latent heat flux increase and sensible heat flux, soil heat flux and evaporation impedance of vegetation decrease. Especially, on the one hand the vapor flux of vegetation surface increases obviously because of increase of vegetation coverage (Fig. 1d, Fig. 2d), and on the other hand, the appearance time of its maximun is delayed. Although the increase of vapor flux of earth's surface is not so obvious, its maximum appears later indeed. Latent heat flux represents the sum of vapor flux of earth's surface and vegetation evapotranspiration that includes evaporation of leaf surface damming off precipitation and physiological transpiration of vegetation. In this paper, precipitation is not considered, so the leaf surface damming off precipitation is zero, and only physiological transpiration of vegetation is taken into account. This physical quantity is closely related with the specific humidity of earth's surface and referenced layer. The maximum of the specific humidity difference between earth's surface and referenced layer appears at about 1300 LT (figure omitted), while the maximum of vegetation evapotranspiration appears at about 1400 LT, so there is a time

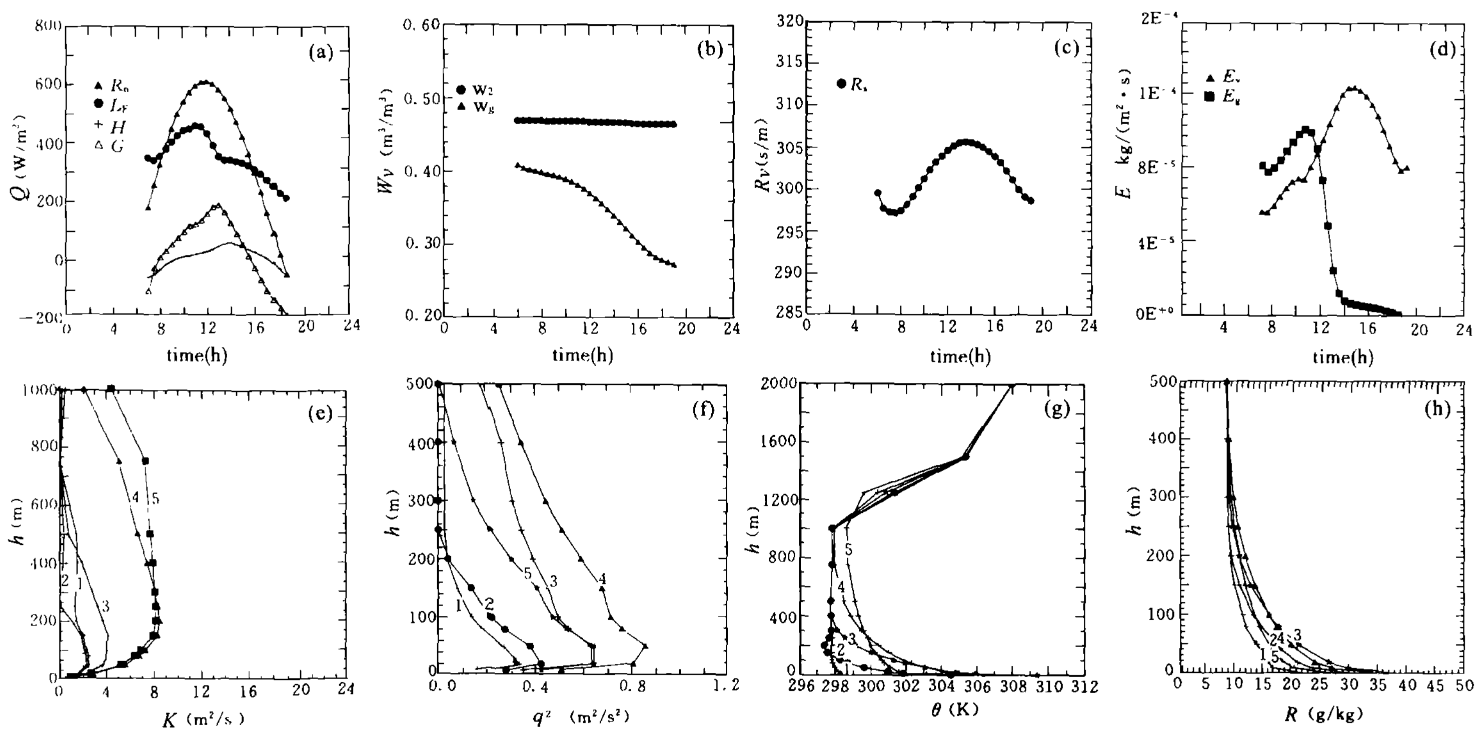

Fig.1. Numerical simulations of surface process and structure of atmosphere boundary layer over grassland underlying surface; (a) heat fluxes (Q), (b) soil volumetric water content(W_V), (c) evaporation impedance of vegetation (R_V), (d) vapor fluxes of surface (E), (e) turbulence exchange coefficient (K), (f) turbulence kinetic energy (q^2), (g) profiles of potential temperature (θ), and (h) profiles of specific humidity (R). InFigs.1e–1h, the numbers 1to 5 indicate time 0800, 1000, 1300, 1500, and 1700LT, respectively.

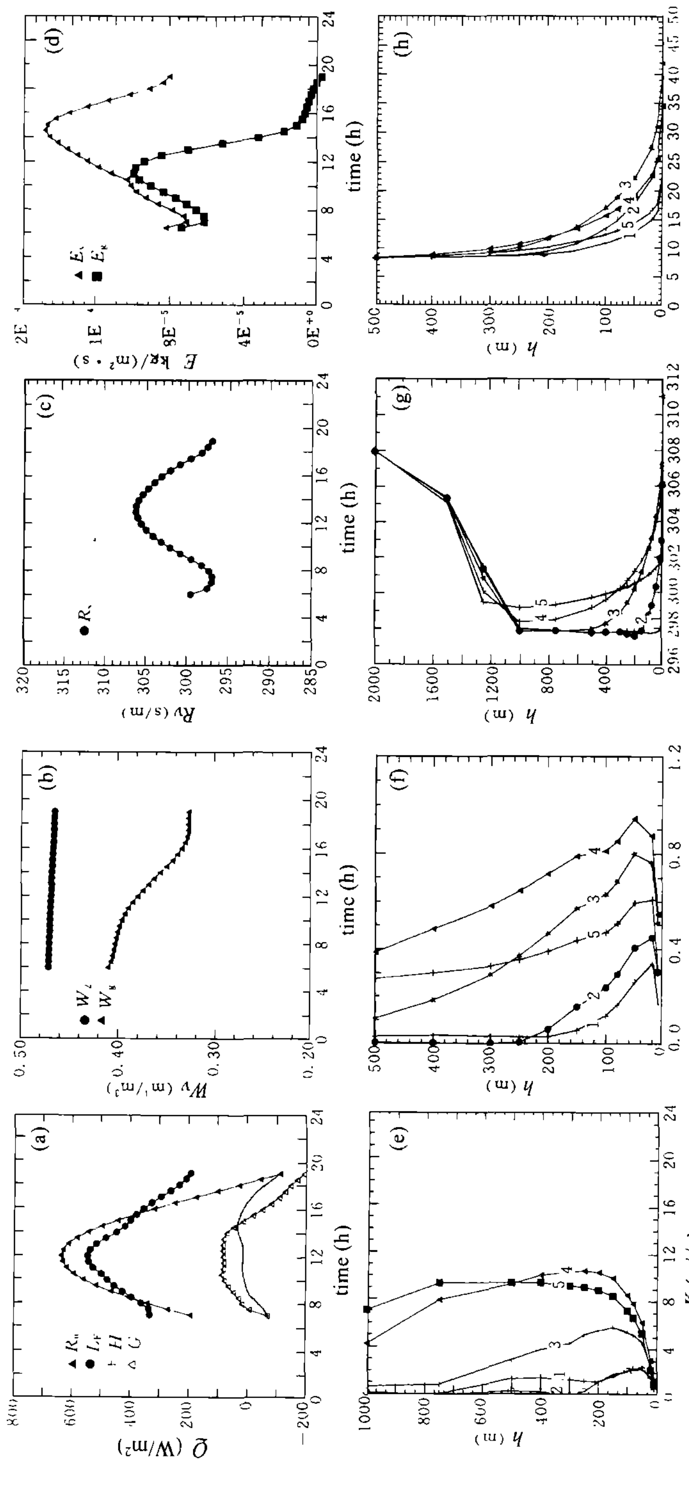

Fig.2. Experiment 1 of simulation of surface process and structure of atmosphere boundary layer, the letters and numbers have the same meaning as those in Fig.1.

lag. The larger the difference is, the stronger the upward vapor transfer is, and the faster the evapotranspiration is. The water source of vegetation evapotranspiration is different from that of evaporation of earth's surface, and it mainly depends on the plant root system pumping water from deep soil. The water is transported inside plants and reaches plant surface in the end. On the other hand, the water of earth's surface evaporation comes mainly from surface soil moisture. Only when the surface soil moisture drops to some degree, and only if the topography is low enough, can under surface water go upward and reach a new balance, which makes the surface layer remain moist. Because the maximums of earth's surface evaporation and vegetation evapotranspiration appear at different time, the maximum of latent heat flux appears between them. From Figs. 1b and 2b, we can see that the decrease of earth's surface moisture accelerates rapidly after 1100 LT, and the less the vegetation coverage is, the faster the decrease rate is. The result of the model above shows that the effect of surface physical quantities on the surface process is obvious.

Then, how about the effect of surface process on atmosphere boundary layer? Comparing panels e to h of Fig. 1 and Fig. 2, we can find that if the soil volumetric water content, the daily average soil volumetric water content and surface reflectivity are reduced, and vegetation coverage and leaf area index are increased with other conditions keeping constant, the profile structure of atmosphere boundary layer is somewhat affected. In sunny days, the main mechanism of the development of mixed layer is the solar heating at earth's surface. Half an hour after sunrise, turbulence mixing starts. Because the hot bulbs of warm air ascend from earth's surface, mixing in mixed layer gets stronger because of static instability. By entraining and mixing the upper weaker turbulence downward, the mixed layer is developed. The final result is that heat, vapor and momentum tend to be vertical homogeneous. Before 1300 LT, roughness of the surface, friction velocity and heat consumption of evapotranspiration increase along with the increase of vegetation coverage, so the development for profiles of turbulence exchange coefficient, turbulence kinetic energy and potential temperature is hampered. After 1300 LT, because of the strong convection turbulence exchange, the development for profiles of turbulence exchange coefficient, turbulence kinetic energy and potential temperature is strengthened.

Vegetation coverage is a main physical parameter of terrain in surface process parameterization. There are two parameters closely related to the vegetation coverage, which are leaf area index and surface reflectivity. Vegetation coverage is related with the physiological period of plants and season alternation. Leaf area index is the ratio of the total leaf surface area to its projected area on the earth's surface. In this model, leaf area index is taken as four times of vegetation coverage. The definition of reflectivity (generally referred to as albedo) is the ratio of the solar radiation, reflected by the earth's surface to the incoming solar radiation. Generally, in visible wave band of short-wave solar radiation, reflectivity gets less with the increase of vegetation coverage. It is not only related with physical features of the surface, but also with the solar altitude, incidence mode of solar radiation and spectrum of solar radiation. And it has the character of diurnal variation.

Following experiments are made for vegetation coverage. Remaining all the other parameters constant, the vegetation coverage is changed, along with which leaf area index and surface reflectivity are changed (Table 2). The purpose of experiments is to check up the effects of the change of parameters on heat fluxes, daily change of vapor flux and other features of atmosphere boundary layer.

With soil volumetric water content remaining unchanged, the diurnal variations of heat fluxes during surface process of Experiments 1 and 2 are shown in Fig. 2a and Fig. 3a respectively. When the vegetation coverage is 0.63, the maximum of latent heat flux is 548 W m^{-2}, and it appears at about 1200 LT. In Experiment 2, when the vegetation coverage is 0.45, the maximum of latent heat flux is 400 W m^{-2}, and it appears at about 1000 LT. Under the surface condition that the vegetation coverage is relatively small, the latent heat flux is less. The main reason is that large vegetation coverage can provide large leaf surface and many plant root systems to pump water from deep soil for evapotranspiration. This point can be seen clearly in Fig. 2d and Fig. 3d. In Fig. 2d that shows the result of Experiment 1, vegetation coverage is 0.63, the vapor flux of soil surface is smaller than the evapotranspiration flux of vegetation all the time, and the maximum of soil surface vapor flux appears at about 1100 LT, while the maximum of vegetation evapotranspiration flux appears at about 1500 LT. In Fig. 3d that shows the result of Experiment 2, the vegetation coverage is 0.45, before 1200 LT, the vapor flux of soil surface is greater than the evapotranspiration flux of vegetation, while after 1200 LT; the evapotranspiration flux of vegetation dominates. This phenomenon indicates that with the increase of vegetation coverage the amount of vegetation evapotranspiration is occupying more proportion of latent heat flux. Comparing Fig. 2d with Fig. 3d, we can also find that the vegetation evapotranspiration of Experiment 1 is about 1.7 times of that of Experiment 2. All these emphasized the importance of vegetation coverage to heat balance and water transfer of surface process. In Fig. 2a and Fig. 3a, comparing sensible heat flux with soil heat flux, we can see that when vegetation coverage is 0.63, the diurnal variation is tardy and peak values are not obvious, while when vegetation coverage is 0.45, the diurnal variation of sensible heat flux is very large and its peak value reaches 250 W m^{-2}(at 1200 LT), and the peak value of soil heat flux appears at about 1300 LT.

Figures 2b and 3b give the diurnal variations of surface soil moisture of Experiments 1 and 2. In Experiment 1 ($V_{eg}=0.63$), the drop range of surface soil volumetric water content is 0.1, while in Experiment 2 ($V_{eg}=0.45$) the drop range is 0.2. This phenomenon is a good complement to the preceding result, i. e. when vegetation coverage is relatively low, the evaporation of soil surface is dominant. That is to say, in a short period of time, more water can evaporate from surface layer of the soil, leading to more water loss of surface soil. In this model the soil is divided into two layers. The surface layer is only 1 cm, thus the surface soil moisture is just related to the liquid water flux of soil surface, evaporation of soil surface and water infiltration due to gravity. This paper has not considered the conditions of precipitation. Concerning the daily average moisture of soil, soil depth is 1 m. And this physical quantity is related to soil surface evaporation and physiological transpiration of vegetation.

Comparing Figs. 2e – 2h with Figs. 3e – 3h, we can see the effects of vegetation coverage on the structure of atmosphere boundary layer. The result shows that when

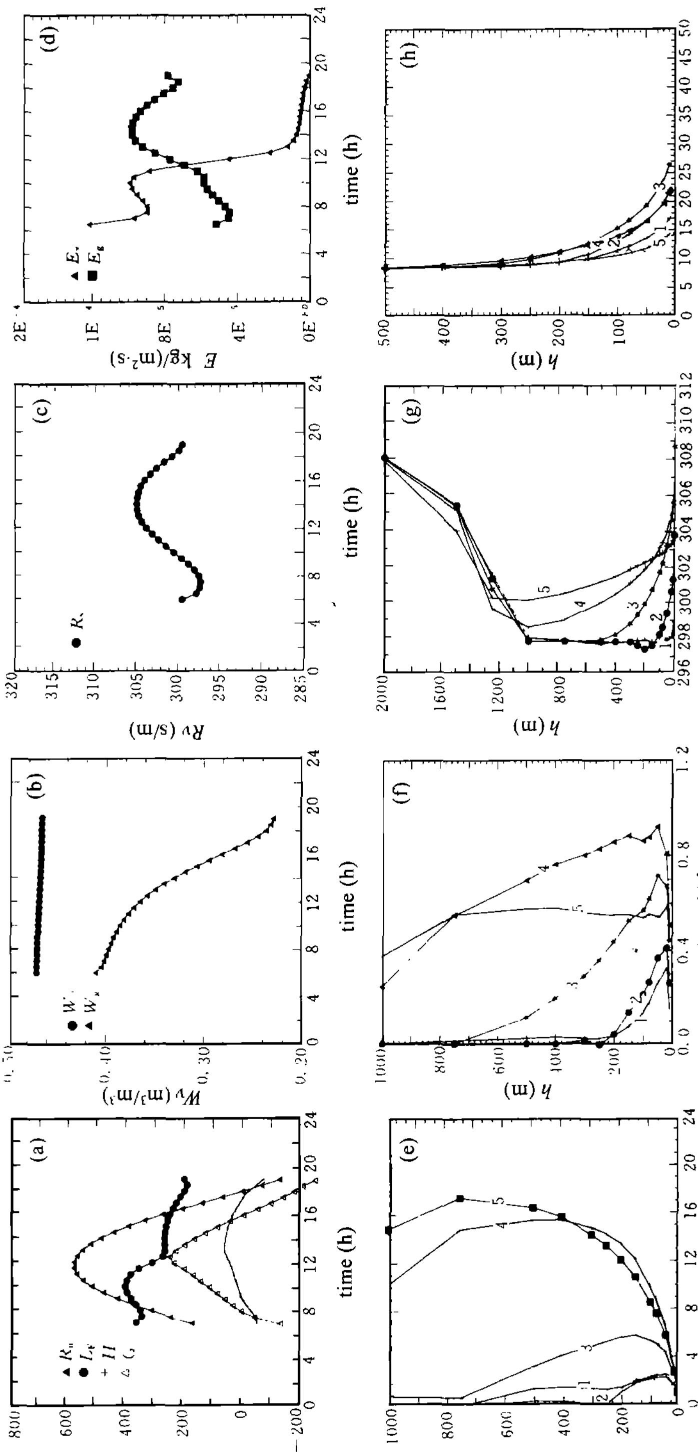

Fig.3. Experiment 2 of simulation of surface process and structure of atmosphere boundary layer, the letters and numbers have the same meaning as those in Fig.1.

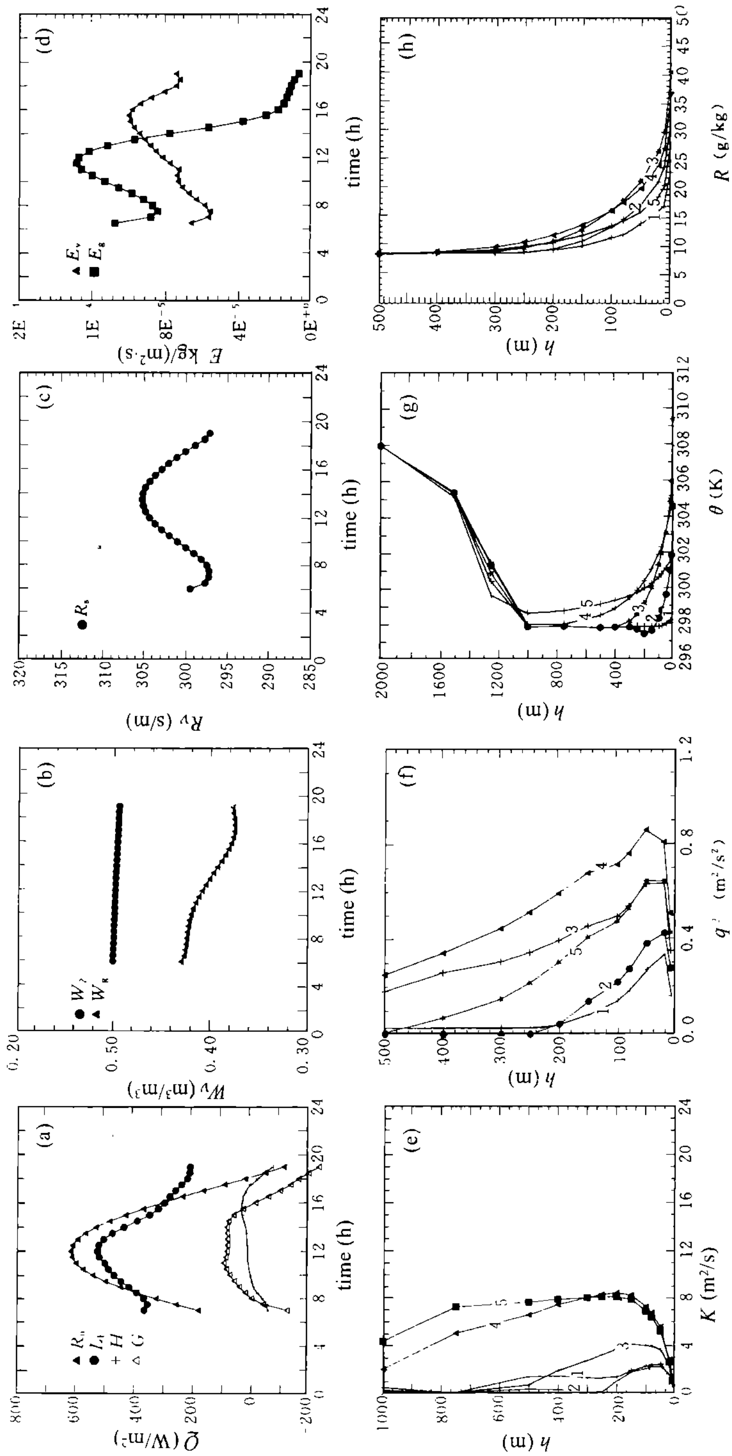

Fig.4. Experiment 3 of simulation of surface process and structure of atmosphere boundary layer, the letters and numbers have the same meaning as those in Fig.1.

vegetation coverage is increased, because of the increase of surface friction resistance the profiles of turbulence exchange coefficient decrease obviously; the profiles of turbulence kinetic energy at 0700 LT and 1500 LT decrease a little, while the profiles of turbulence kinetic energy at 0800, 1000 and 1300 LT increase somewhat. In addition, from the analysis above, we can find out that it is because of the increase of vegetation coverage that augments the latent heat flux consumption of surface solar energy, and weakens profiles of potential temperature at 0700 LT and 1500 LT. But profiles of specific humidity are strengthened obviously.

In Experiment 3, soil moisture is changed, and other physical quantities are the same as in Experiment 1 (Table 2). The purpose of Experiment 3 is to check up the effects of soil moisture on heat fluxes and interactions between daily variations of vapor flux and atmosphere boundary layer in surface processes.

Figures 4a and 4d show the daily variations of heat fluxes and vapor fluxes of Experiment 3. Comparing Fig. 4a with Fig. 1a, we find that latent heat flux of Experiment 3 increases obviously, and the maximum is 520 W m^{-2}, and appearance time is about 1200 LT. While in Fig. 1a, the maximum is 450 W m^{-2}, and its appearance time is about 1100 LT. This indicates that the status of soil moisture may affect both the magnitude and the appearance time of the peak value of latent heat flux. In Fig. 1a the character of daily variation of latent heat flux is because that when soil moisture and daily average soil moisture is relatively small, the solar energy consumed by evaporation is relatively small. Therefore, although latent heat flux consumed by surface evaporation and vegetation transpiration increases quickly with the increase of solar radiation, it is soon confined by the insufficiency of soil water supply. Thus latent heat of evaporation is weakened down rapidly. Then because of the complement of water from deep soil, the decrease rate of latent heat of evaporation gets slower and slower. On the other hand, sensible heat flux increases with the increase of solar radiation, and reaches its maximum at about 1300 LT, and then decreases rapidly. Daily variation of soil heat flux is also very clear, and the maximums of latent heat flux, sensible heat flux and soil heat flux are in turn delayed. In Fig. 4a, because soil moisture and daily average soil moisture are both sufficient, with increases of solar radiation there is no deficiency of water for evaporation and evapotranspiration. Hence, latent heat flux increases steadily with the increase of solar radiation, and when solar radiation reaches its maximum, the latent heat flux of evaporation also reaches its maximum. Because most solar radiation is used for latent heat flux of evaporation, the daily variations of sensible heat flux and soil heat flux are very little, and their peak values are not clear. Comparing Fig. 4d with Fig. 1d, we can find that when soil volumetric water content and daily average soil moisture are increased to some degree, the daily variation rules of vapor flux of soil surface and vapor flux of vegetation surface are completely different. When soil volumetric water content and daily average soil moisture are not sufficient, the vapor flux of soil surface is small, and its maximum appears at about 1100 LT, and then decreases rapidly, while the vapor flux of vegetation surface reaches its peak value at about 1500 LT. When soil volumetric water content and daily average soil moisture are sufficient, the vapor flux of soil surface is larger, and its maximum appears at about 1200 LT, and the value is larger than the counter part value in Fig. 1d; the increase rate of vapor flux of vegetation surface is so tardy that its peak value appears at about 1600 LT, and the maximum is less than the counterpart of Fig. 1d.

The analysis indicates that the effects of vegetation coverage on surface processes and atmosphere boundary layer are immensely sensitive. And the effects are especially sensitive to soil surface evaporation and vegetation evapotranspiration. The effects of soil volumetric water content and daily average soil moisture are sensitive to surface processes, but less sensitive to atmosphere boundary layer.

V. CONCLUSIONS

Both equations of two-dimensional air motions, heating power, continuity and energy and schemes of surface physical process parameterization are used in this model to simulate the interactions of surface physical process and atmosphere boundary layer. Aiming at several physical parameters, sensitivity experiments are made in the model. The results prove that this model can simulate surface heat balance, soil volumetric water content, evaporation resistance of vegetation, daily variations of vapor flux, profiles of turbulence exchange coefficient, turbulence kinetic energy, potential temperature and specific humidity reasonably. The model developed can be used to study the interaction between land surface processes and atmospheric boundary layer in city regions, and it can also be used in the simulation of climate incorporating a mesoscale model. Analysis indicates that this model still has some shortcomings. For example, this model only considers the daytime surface process and boundary layer, but does nothing to nocturnal conditions, and this brings about problems to analyze continuous variations of physical quantities. Moreover, this model does not deal with vegetation and soil surface separately, and does not consider the interactions between vegetation and soil surface and heat storage in vegetation. In addition, because the parameters of this model are selected with considering factors of moist area, vegetation coverage can not be set too little. Otherwise, parameters are not matching with each other, which can lead to insignificant results. Thus conditions with little vegetation coverage are not discussed in the model. This model considers little about the coupling of physical quantities, which makes physical variables are almost independent of each other during the process of simulation. Whether it is necessary to import factors of correlativity to enforce the harmony of the whole is a question worthy of further research in future.

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