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A PBL-Radiation Model for Application to Regional Numerical Weather Prediction

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ABSTRACT

Often in the short-range limited-area numerical weather prediction (NWP) of extratropical weather systems the effects of planetary boundary layer (PBL) processes are considered secondarily important. However, it may not be the case for the regional NWP of mesoscale convective systems over the arid and semi-arid highlands of the southwestern and south-central United States in late spring and summer. Over these dry regions, the PBL can grow quite high up into the lower middle troposphere (600 mb) due to very effective solar heating and hence a vigorous air-land thermal interaction can occur. The interaction representing a major heat source for regional dynamical systems can not be ignored.

The present study focuses on the development of an one-dimensional PBL-radiation model. The model PBL consists of a constant-flux surface layer superposed with a well-mixed (Ekman) layer. The vertical eddy mixing coefficients for heat and momentum in the surface layer are determined according to the surface similarity theory, while their vertical profiles in the Ekman layer are specified with a cubic polynomial. Prognostic equations are used for predicting the height of the nonneutral PBL. The atmospheric radiation is parameterized to define the surface heat source/sink for the growth and decay of the PBL. A series of real-data numerical experiments has been carried out to obtain a physical understanding how the model performs under various atmospheric and surface conditions.

This one-dimensional model will eventually be incorporated into a mesoscale prediction system. The ultimate goal of this research is to improve the NWP of mesoscale convective storms over land.

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The PBL program used in this air-land interaction research is a modified version of the marine PBL code written by Dr. S. Yoh of Drexel University. I appreciate valuable discussion and comments on this work by my colleagues Dr. F. Robertson of NASA/MSFC and Dr. M. Kalb of USRA and their interest in this research. Also, I wish to thank Ms. J. Srikishen for technical help in carrying out numerical experiments, and friends at USRA for a delightful summer.

1. INTRODUCTION

Often in the short-range limited-area numerical weather prediction (NWP) of extratropical weather systems the effects of planetary boundary layer (PBL) processes are considered secondarily important. This is probably true over land in winter or over the oceans in summer when the PBL is generally very shallow. However, it may not be the case for the regional NWP of mesoscale convective systems over the arid and semi-arid highlands of the southwestern and south-central United States. Over these arid regions, in late spring and summer the PBL can grow quite high up into the lower middle troposphere (600 mb) due to very effective solar heating. This can cause vigorous air-land thermal interaction. The interaction representing a very significant heat source/sink for regional dynamical systems can not be ignored.

The present study will focus on the development of an one-dimensional model coupling the PBL processes and atmospheric radiative transfer. The model PBL consists of a constant-flux surface layer superposed with a well-mixed (Ekman) layer. The vertical eddy (turbulence) mixing coefficients for heat and momentum in the surface layer will be determined according to the surface similarity theory, while their vertical profiles in the Ekman layer are specified with a cubic polynomial. Prognostic equations are formulated for predicting the height of nonneutral PBL. The atmospheric radiation including solar and infrared components is parameterized to define the surface heat source/sink for the growth and decay of the PBL. This one-dimensional model will eventually be incorporated into the limited-area mesoscale prediction system (LAMPS) (Perkey,1976; Chang *et al.*, 1981) for regional NWP over land.

The model levels range from 0 to 16 km with relatively higher resolution in the PBL. A discussion of PBL physics and techniques for solving the atmospheric radiative transfer in this study is included in Sections 3 and 4, respectively. The results of three numerical experiments conducted on the PBL-radiation model are described in Section 5.

2. OBJECTIVES

The ultimate goal of this research is to improve the NWP of mesoscale convective storms over arid land. Before incorporating the PBL-radiation model into LAMPS, we like to carry out a series of real-data numerical experiments to have a physical understanding how the one-dimensional model performs under various atmospheric conditions. The specific objectives include:

1) To assess the impact of clouds on the surface heat sources and on the PBL development.

2) To determine the sensitivity of the PBL height to its initial value.

3) To understand the impact of surface conditions such as roughness length, albedo, the Bowen ratio on the structure of the PBL.

4) To understand the role of the static stability in the growth and decay of the PBL.

3. PLANETARY BOUNDARY LAYER

The PBL is a layer of atmosphere on the order of 1 km in depth above the earth's surface. This thin layer of air is characterized by small-scale turbulence of spatial dimension no greater than 1 km. In the free atmosphere above the PBL, the turbulent motions are considerably weaker. The PBL processes represent a consequence of interaction between the lowest layer of air and the underlying surface. The interaction can significant impact on the dynamics of the upper air flows.

The influences of the small-scale eddies on large-scale (model resolable scale) atmospheric circulations may be included in the model equations as described below. These equations are formulated on the (x,y,z,t) coordinate system, in which x is along the latitudinal (west to east), y is along the longitudinal (south to south) circles on the earth's surface, while z is perpendicular to the x-y plane.

3.1 The basic equations

Equations of horizontal motion:

$$du/dt = fv - \rho^{-1}\partial p/\partial x - \rho^{-1}[\partial \rho\langle u'u'\rangle/\partial x + \partial \rho\langle u'v'\rangle/\partial y + \partial \rho\langle u'w'\rangle/\partial z] \quad (1)$$

$$dv/dt = -fu - \rho^{-1}\partial p/\partial y - \rho^{-1}[\partial \rho\langle u'v'\rangle/\partial x + \partial \rho\langle v'v'\rangle/\partial y + \partial \rho\langle v''w'\rangle/\partial z] \quad (2)$$

Hydrostatic equation:

$$\partial p/\partial z = -\rho g \tag{3}$$

Equation of state for air:

$$p = \rho R\theta(p/1000)^{0.28571} \tag{4}$$

Mass conservation:

$$d\rho/dt = -\rho(\partial u/\partial x + \partial v/\partial y + \partial w/\partial z) \tag{5}$$

Energy conservation:

$$d\theta/dt = H + \rho^{-1}[\partial\langle-\rho u'\theta'\rangle/\partial x + \partial\rho\langle-v'\theta'\rangle/\partial y + \partial\rho\langle-w'\theta'\rangle/\partial z] \tag{6}$$

Moisture conservation:

$$dq/dt = Q + \rho^{-1}[\partial\langle-\rho u'q'\rangle/\partial x + \partial\rho\langle-v'q'\rangle/\partial y + \partial\rho\langle-w'q'\rangle/\partial z] \tag{7}$$

where $d/dt = \partial/\partial t + u\partial/\partial x + v\partial/\partial y + w\partial/\partial z$. There are seven basic large-scale variables in the above system: three velocity components in the x, y, and z directions, respectively, u ,v, and w; pressure p; air density ρ; potential temperature θ; and specific humidity q. Other symbols include the Coriolis parameter f, heat source H, moisture source Q, and the gas constant for dry air R.

The prime ' denotes the small-scale turbulent fluctuations, while the bracket ‹ › denotes a time average over a period much longer than the time scale of the turbulent fluctuations. In the u- and v-equations, the first term represents the Coriolis force due to the earth's rotation, and the second term represents the pressure gradient force. The six covariances (e.g., $-\rho\langle u'u'\rangle$) are referred to as eddy stresses representing eddy or turbulent fluxes of momentum. Similarly, there are eddy flux of heat in the θ equation and eddy flux of moisture in the q equation.

3.2 Closure problem

Considering the basic variables only, we note that the above set of equations forms a closed system. Given initial conditions, the future states of the seven basic unknowns can be determined in principle by numerical integration. However, with the eddy flux terms representing additional dependent variables additional equations are required to close the system. This is so-called closure problem in dealing with atmospheric turbulence.

In the free atmosphere, for the atmospheric circulations of horizontal scale larger than 100 km and time scale on the order of one day or longer the contribution to the local change ($\partial / \partial t$) due to the eddy stresses as well as the eddy fluxes of heat and moisture is insiginificant in comparison with those due to large-scale processes (e.g., pressure gradient force, advection, etc.). In NWP models, we may neglect the eddy flux terms in the above set of equations. This is known as the zero-order closure.

In the PBL, however, the eddy flux terms may have the same order of magnitude as the other terms and can not be neglected under most situations. To close the system without neglecting PBL processes in a NWP model, we may consider the following two closure approaches.

a) First-order closure

A well-known and commonly used technique in PBL meteorology for the first-order closure is the so-called K-theory. The theory simply states that the turbulent flux of any physical quanties such as momentum and heat may be treated analogously to molecular diffusion. For example, for the vertical eddy flux of heat, we may write

$$\langle \theta' w' \rangle = K \partial \theta / \partial z$$

which indicates the vertical mixing of heat as a result of turbulence is parameterized in terms of the large-scale variables. Similar expressions can be formulated for the eddy mixing of momentum. A more detailed discussion of the K-theory and the procedures to implement the theory to PBL modeling in this research is presented in Section 3.3.

b) Higher-order closure

The concept consists of deriving prognostic equations for the eddy flux

terms from the basic equations. The derived equations contain higher-order correlations. For example, in the case of second-order closure **triple correlations such as** $\langle u'u'v' \rangle$ will appear in the prediction equation for $\langle u'u' \rangle$. Thus, the closure problem remains and closure assumptions have to be made at a higher level to complete the system. Moreover, the introduction of the equations for the eddy components in NWP models will drastically increase the demand for computer time. Also, the current observations are insufficient to provide an accurate initial condition for the eddy fluxes. To date, higher-order closure approaches are not yet desirable for NWP models.

3.3 First-order closure PBL model

As shown in the basic equations, the eddy mixing of momentum, heat, and moisture includes both horizontal and vertical components. Due to the horizontally stratified nature of the atmosphere, the gradient of u, v, θ, and q is much greater in the vertical than in the horizontal direction. The eddy mixing in the horizontal is negligible comparing to its counterpart in the vertical. Hence, with the use of the K-theory the central issues in PBL research involve the determination of the vertical profiles of K for momentum, heat, and moisture as well as the depth of the PBL within which the profiles are applied.

In formulating equations for the K profiles, a two-layer model is developed. The PBL is devided into a constant-flux surface layer superposed with a well-mixed Ekman layer.

a) Surface layer

This is a very thin turbulent layer within a depth less than 50 m above the surface. In this layer, the turbulent fluxes are nearly constant with height and the vertical wind profile depends on the static stability and surface roughness. There are mechanical turbulence driven by the vertical wind shear as well as thermal turbulence driven by buoyancy (dry convection). The former decreases rapidly with height, while the latter varies slowly with height.

For a surface layer of neutral or near-neutral stability ($\langle \theta' w' \rangle \approx 0$ or mechanical turbulence only, the vertical wind profile is very close to the well-known logarithmic form

$$u = (u_*/k) \ln (z/z_0) \tag{8}$$

and the corresponding K, eddy viscosity for both u and v, is

$$K = kzu_* \tag{9}$$

where $k = 0.35$ the von Karmon constant, u_* the friction velocity, and z_0 the roughness length. Over a given surface, z_0 is a known constant, while u_* can be evaluated from the observed wind profile. Thus, K is determined. Note that K increases upward suggesting the eddy size is proportional to the height above the surface.

For a nonneutral surface layer, there will be vertical heat exchanges between air and the underlying surface. We will have predominant mechanical turbulence if the layer is stable ($\partial\theta/\partial z > 0$), or both mechanical and thermal turbulence if the layer is unstable ($\partial\theta/\partial z < 0$). According to the similarity theory summarized in the book by Haltiner and Williams (1980), the vertical gradients of the wind and potential temperature can be expressed as follows, respectively.

$$\partial u/\partial z = [(u_*/k)/z]\, \phi_m(z/L) \tag{10}$$

$$\partial\theta/\partial z = [(\theta_*/k)/z]\, \phi_h(z/L) \tag{11}$$

$$L = \theta_0 u_*^2/(kg\theta_*) \tag{12}$$

where θ_* analogous to u_* representing a temperature scale related to the vertical eddy heat flux, and θ_0 the potential temperature at the surface.

The length scale L, called the Monin-Obukhov length derived based upon dimension analysis, measures the relative role between mechanical and thermal forcing in generating the turbulence. L is negative in unstable air with upward heat flux, while positive in stable air with downward heat flux. Also, physically the absolute value of L may be interpreted as the height above which the mechanical turbulence becomes insignificant compared to the thermal turbulence. Thus, as air becomes more unstable, the absolute value of L becomes smaller. ϕ_m and ϕ_h are two nondimensional universal functions determined empirically from observations. The corresponding eddy viscosity K_m and eddy conductivity K_h are

$$K_m = ku_* z/\phi_m \tag{13}$$

$$K_h = ku_* z/\phi_h \tag{14}$$

For unstable air

$$\phi_m = [1 - 15z/L]^{-1/4} \quad \text{and} \quad \phi_h = 0.74[1 - 9z/L]^{-1/2} \tag{15}$$

For stable air

$$\phi_m = [1 + 4.7z/L] \qquad \text{and} \qquad \phi_h = 0.74 + 4.7z/L \tag{16}$$

For moisture flux, K_q known as eddy diffusivity is assumed to be the same as K_h.

b) The Ekman layer

The vertical extent of the PBL is variable. Its depth may range from 100 m (e.g., the nocturnal PBL) in stable air to over 1 km (e.g., the daytime PBL over arid land) in unstable air. Out of the total depth the surface layer occupies less than 10% of the lowest part, and the rest is the well-mixed Ekman layer. Because of its elevation, it is much more difficult to conduct observational study in the Ekman layer than in the surface layer. As a result, there is lack of a simple and elegant theory, such as the surface similar theory, which may be used to define the K profiles in this upper part of the PBL. Here, a cubic polynomial similar to the O'Brien's formulation (1970) except for the constraint of zero slope at the top of the PBL is adopted for the the K profiles in the Ekaman layer.

The cubic polynomial has the following form:

$$K(z) = K(z_s)\,[1-9.75z+8.75z^3] + \partial K(z_s)/\partial z\ \ [z_e-z_s]\,[z-3.5z^2+2.5z^3] + K(z_m)\,[125z^2\,(1-z)/12] + K(z_e)\,[z^2(5z-2)/3] \tag{17}$$

where $K(z_s)$, $K(z_e)$, and $K(z_m)$ are K evaluated at the top of the surface layer (z_s), at the top of the Ekman layer (z_e), and at the middle of the PBL (z_m), separately. $K(z_s)$ and $\partial K(z_s)/\partial z$ are obtained from the surface layer physics discussed earlier, and $K(z_e)$ has to be specified. Also, K has the maximum value at z_m. In the current PBL model z_s is set at 25 m and $K(z_e)$ is set equal to 1 m^2s^{-1}.

c) Height of the PBL

To apply the K profiles, we must know the depth of the PBL. Based on theoretical and observational studies, for the neutral PBL, Panofsky and Dutton (1984) suggested

$$z_e = 0.2u_*/f \tag{18}$$

For the unstable PBL, the Deardorff's (1974) prognostic equation is used.

$$dz_e/dt = 1.8(w_*^3 + 1.1u_*^3 - 3.3u_*^2fz_e)/(z_e^2\sigma + 9w_*^2 + 7.2u_*^2) \tag{19}$$

where $w_* = [g\theta_*z_e/\theta_0)]^{1/2}$, the vertical velocity scale; $\sigma = (g/\theta_0)\partial\theta/\partial z$, the static stability at the surface. For the stable PBL, the following equation suggested by Nieuwstadt and Tennekes (1981) is used.

$$\partial z_e/\partial t = -u\partial z_e/\partial x - v\partial z_e/\partial y + (z'-z_e)/t_s \tag{20}$$

where $z' = 0.4(u_*L/f)^{1/2}$, the equilibrium height; $t_s=0.75[\theta(z_e)-\theta_0](d\theta_0/dt)^{-1}$, the time scale.

d) Extremely unstable PBL

Recently, Yoh (1989) suggested that under extremely unstable conditions some modifications of the surface similarity theory were necessary. In very unstable air, $-L$ approaches the order of meters and free convection responsible for generating large local eddies plays a key role in the vertical eddy mixing. It is inappropriate to use u_*, which is strongly influenced by the mean wind, as a velocity scale to determine the eddy fluxes. Also, because $\phi_m(z/L)$ and $\phi_h(z/L)$ are accurate for a relatively small $|z/L|$ (< 5), both functions may only be applied at the level of a few meters in height.

Yoh subdivided the surface layer into two layers: the lowest layer of 1.22 m upward from the suface, in which the modified similarity theory will be applied, and 1.22-25 m layer, in which the mixed layer scaling will be applied. The major difference between the modified and the original similarity theory discussed earlier is the incorporation of w_*, a free convection scaling velocity, in computing the modified friction velocity, u_+ and Monin-Obuhkov length L_+. The mixing length scaling technique is designed to relate u, v, θ, and q at 1.22 m to those at 25 m. This information provides an input for the modified surface similarity. The relationship between u_+ and L_+ is identical to that between u_* and L (see Equ. 12). After L_+ is determined, Equations (13)-(16), with L replaced by L_+, are used to obtain Ks at 25 m. Finally, the K profiles in the Ekman layer under very unstable conditions can be determined from (17).

e) Surface heat source

From the above discussion, it is clear that under the nonneutral conditions surface heat fluxes exert great influences on the K profiles in the PBL. The intensity of these heat fluxes depends on to a large extent the temperature difference between air and the surface. To solve numerically the PBL equations, we often need the surface temperature as one of lower boundary conditions.

Over the oceans, water responds slowly to atmospheric radiation and the temporal variation of water temperature is small. A constant sea surface temperature (SST) field at its initial observed values is adequate for model PBL in a short-range (< 48 h) NWP. Over land, due to its quick response to solar heating the temporal variation of surface temperature is rather large.

A diurnal character is expected in the PBL height as well as the K profiles. Thus, a mechanism for determining the surface temperature and the associated heat flux is needed for properly resolving the essential features of the PBL over land. This is particularly true over arid land.

In this study, the land surface temperature will be calculated based on a heat budget consideration. The surface heat budget involves infrared and solar irradiance, and sensible and latent heat flux. Also, the surface heat budget is well known to be a strong function of cloud cover and moisture content in air. In the next section, a brief description of the treatment of radiation processes as well as the procedures to obtain the surface temperature is presented.

4. RADIATIVE ENERGY AND HEAT BUDGET AT THE SURFACE

In modeling radiative transfers the usual constraints of a numerical model, i.e., not to be too time-consuming, must be considered. In this study only water vapor, which is one of the prognostic variables in the model equations, is regarded as a radiatively active gas in the model atmosphere. Ozone and carbon dioxide are not considered in radiation computation because the model top is set at the lower stratosphere. In the troposphere, for a short-range NWP carbon dioxide contribution to the atmospheric radiation processes is negligible comparing with that due to water vapor.

To solve the transfer equations for infrared radiation a simplified method based on the broad-band emissivity technique (Staley and Jurica, 1970) is used. The technique has been successfully applied in many NWP studies (Danard, 1969; Chang, 1980). With respect to infrared radiation, clouds and the earth's surface are treated as a blackbody radiator. For modeling solar radiation, an empirical technique designed for the UCLA general circulation model (Haltiner and Williams, 1980) is used. Some modifications have been made to accommodate multi-layer clouds in the current model atmosphere. The reflection of insolation at the cloud top and the earth's surface, and the absorption of insolation by clouds are considered. The intensity of solar radiation varies with the zenith angle which is computed as a function of location and time.

The sum of the insolation and the net infrared (downward minus upward) radiation at the surface is taken as the total available surface heat source in the day or sink at night. The heat source/sink is partitioned into sensible heat flux, H_s, and latent heat flux, H_l, according to the Bowen ratio (Sellers, 1967). The surface temperature is then related to sensible heat flux by a bulk aerodynamic formula of the form

$$H_s = \rho c_p C_d \, [\sqrt{(u^2+v^2)}] \, (\theta_s - \theta_a)$$

where c_p specific heat capacity of air at constant pressure, θ_s the surface potential temperatures, and θ_a the potential temperature at 25 m. The drag coefficience, C_d, is defined as

$$C_d = k^2[\ln(z/z_t) - \phi_h(z/L)]^{-1}[\ln(z/z_0) - \phi_m(z/L)]^{-1}$$

where z_t the roughness length for temperature.

5. RESULTS

Disregarding the large-scale processes in the basic equations shown earlier and with the use of K-theory for the small-scale eddies, we reduce the u, v, θ, and q equations to the following forms.

$$\partial u/\partial t = -\rho^{-1}\partial(\rho K_m \partial u/\partial z)/\partial z$$

$$\partial v/\partial t = -\rho^{-1}\partial(\rho K_m \partial v/\partial z)/\partial z$$

$$\partial \theta/\partial t = -\rho^{-1}\partial(\rho K_h \partial \theta/\partial z)/\partial z$$

$$\partial q/\partial t = -\rho^{-1}\partial(\rho K_h \partial q/\partial z)/\partial z$$

These equations in conjunction with the hydrostatic equation and the equation of state for air constitute a closed system for the one-dimensional PBL-radiation model

The model has 15 levels in the vertical (z in Table 1). The PBL structure is resolved in terms of the lowest 6 levels. The time step used for the model integration was 1 minute and the radiation computation was updated every 30 minutes. The input sounding (Table 1) for the experiments was taken in Kansas at 1200 GMT on 11 June 1985 during the PRE-STORM. Other PBL parameters required for integrating the equations include the roughness length = 20 cm, the Bowen ratio = 0.85, and the surface albedo = 0.9. Major findings of three model experiments are presented below.

Figure 1 summarizes the results of experiment under a clear sky. The model simulated PBL height, θ at the surface and 25-m level, heat and momentum fluxes in the surface layer, and the eddy conductivity at the 25-m level are shown as a function of local time for a 24-h period starting at 6 am (0 h on the figure). The initial PBL height was set at 375 m. During the first 12-h, solar heating resulted in a large amount of upward heat flux. The unstable PBL grew to about 800 m at the noon and reached a maximum of about 1050 m near 6 pm. The maximum air temperature lagged behind the largest temperature difference between the surface and air as well as the maximum upward heat flux by about 6 h. The diurnal range for the surface temperature was close to 25°C and for the air temperature was about 4°C.

In the period after sunset and before sunrise the stable environment prevailed. Air was warmer than the surface and slightly downward heat flux was observed. The stable PBL was much shallower than the unstable one. There was a sharp drop in the height and the eddy conductivity during the transition period from the unstable to stable PBL but a relatively minor decrease in the momentum flux took place. This simulated the collapse of unstable PBL after the cut-off of the surface heat source and the slow re-development of stable PBL at night.

Figure 2 is similar to Fig. 1 except for the experiment under a cloudy sky. A layer of low cloud was placed between the 1.25 km and 2 km model levels. Comparisons between Figs. 1 and 2 reveal that

1) The model cloud decreased the unstable PBL height in the daytime, while increased the stable PBL height slightly at night. The vertical mixing in the unstable PBL was suppressed as indicated by the lower K value.

2) The model cloud had modulated the surface temperature considerably

resulting in much smaller diurnal range. Also, it caused few degree cooling of air in late afternoon and slightly warming of air in the early morning.

3) The model cloud reduced the temperature differences between the surface and air and consequently the heat fluxes, but had relatively minor impact on the momentum fluxes.

Figure 3 shows the results from the third experiment. In this experiment, the model integration started at the local noon instead of 6 am and the initial PBL height was set at 750 m. The main purpose of this experiment is to see how the model behaves during the transition periods from unstable to stable then back to unstable air. Also, it will reveal how sensitive the model is to the initial PBL height. Comparing Figs. 3 with 1, we find that except for the first few hours the model behaved very similarly between the two experiments. For examples, in both cases the PBL grew to slightly over 1 km at 6 pm, and the diurnal ranges in the surface and air temperatures in the two experiments were very close.

6. CONCLUSIONS AND RECOMMENDATIONS

An one-dimensional model coupling the PBL processes and atmospheric radiation has been developed. The simulated results indicate that the model is capable of producing physically realistic solutions. A few more sensitivity tests with different soundings and surface parameters will be desizable for further understanding. For example, we may alter the roughness length or the Bowen ratio and see how the one-dimensional system responds to the changes. We hope to obtain some PBL observations for detailed model verifications in the near future. Comparisons with other approaches in modeling the PBL processes will also be very interesting. After carefully assessment of its performance under various conditions, the PBL-radiation scheme will then be incorporated into the LAMPS.

Table 1 Observed data used in the numerical experiments

z (km)	u (m/s)	v (m/s)	θ (K)	q (g/kg)
16.0	10.	-5.8	403	0.01
14.0	32.2	3.0	359	0.05
12.0	15.2	20.9	341	0.1
10.5	15.9	16.3	337	0.2
9.0	21.7	8.0	332	0.3
7.5	18.3	12.0	327	0.3
6.0	18.9	14.4	322	2.0
4.5	7.8	-10.0	316	2.8
3.0	-3.4	-5.9	310	3.3
2.0	-5.6	-7.3	304	6
1.25	-6.0	-8.5	299	9
0.75	-4.7	-9.4	297	10
0.375	-1.1	-6.0	296	11
0.25	2.5	-3.0	295	12
0.0	2.5	-3.0	295	12

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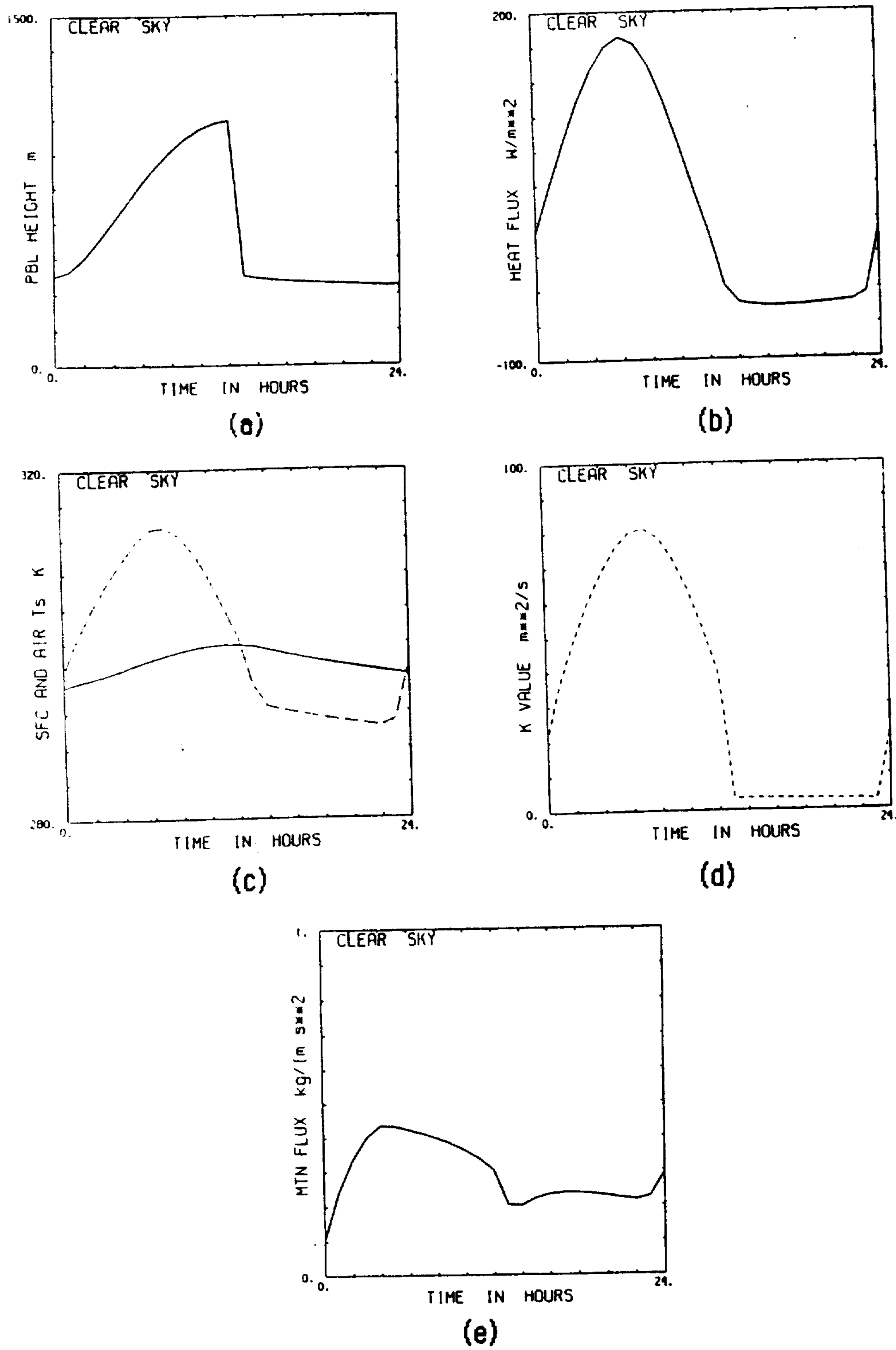

Fig. 1 Model simulated (a) the PBL height, (b) heat flux, (c) ground (dashed line) and 25-m air temperatures, (d) eddy conductivity, and (e) momentum flux as a function of local time starting at 6 am for the case of clear sky.

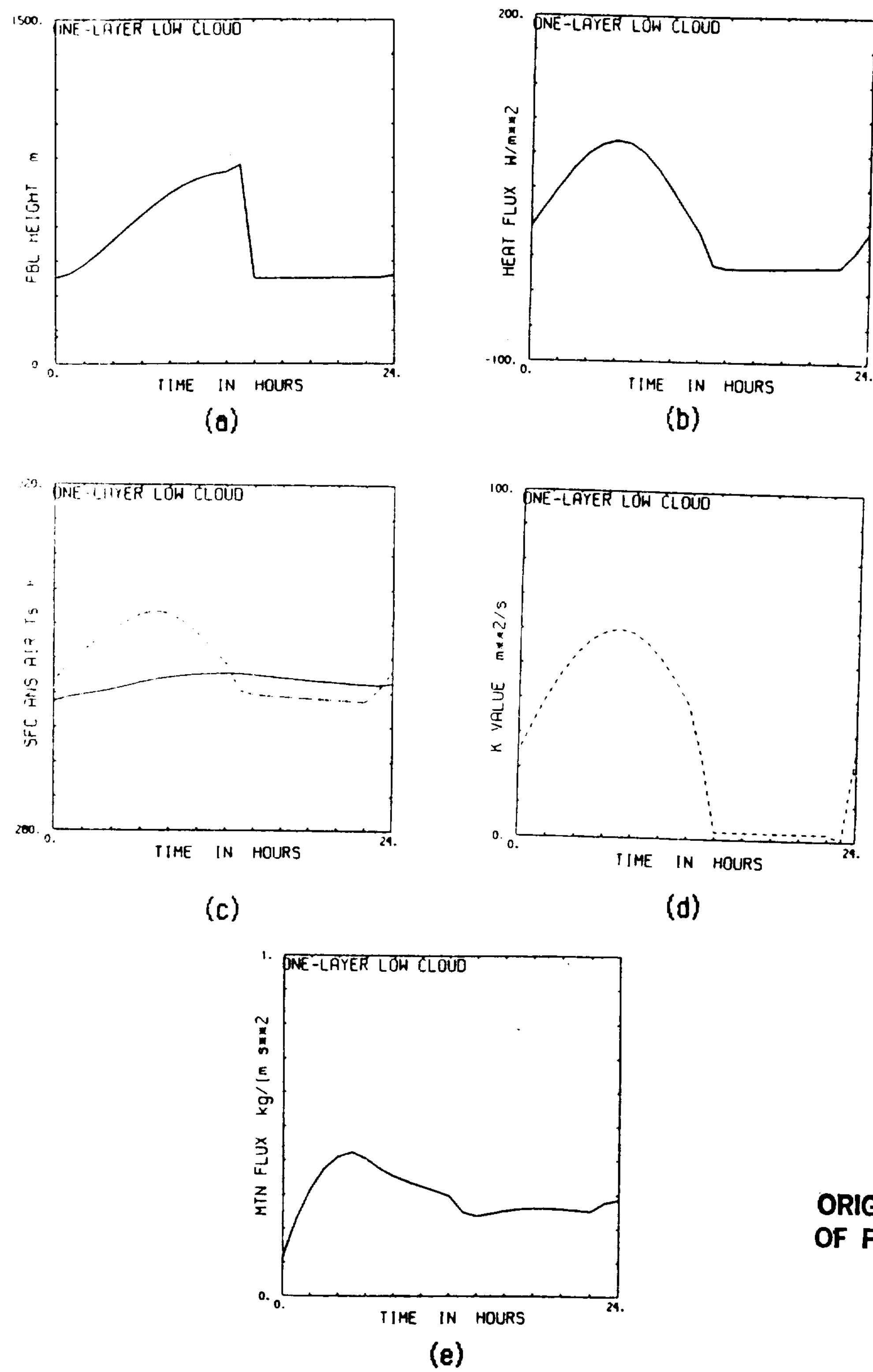

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Fig. 2 Similar to Fig. 1 except for the case of cloudy sky.

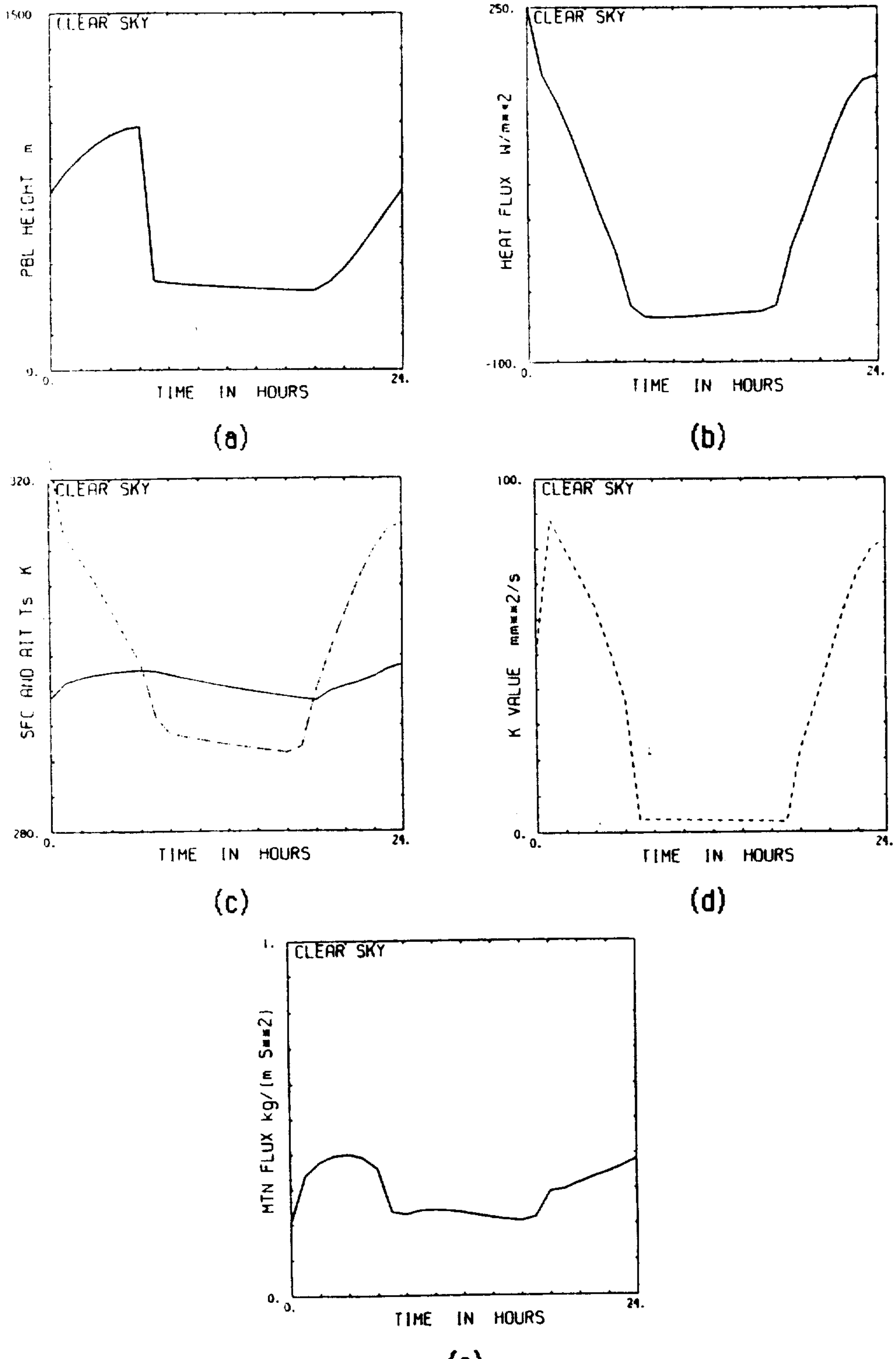

Fig. 3 Similar to Fig. 1 except for starting at noon. The initial PBL height is set at 750 m.