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Effects of air-land thermal interaction on the energetics of planetary boundary layer over NW India

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ABSTRACT. A one dimensional boundary layer model has been used to investigate the influence of solar radiation and the large diurnal variation of surface temperature on the low-level turbulence in the atmosphere over NW India. The model is based on Eulerian conservation equations for momentum, heat, moisture and pollutants and is capable of examining the transport and diffusion processes in the atmospheric boundary layer. The simulation experiment has been conducted for a period of 48 hours based on initial set of meteorological observations collected during 24-26 April 1966 at New Delhi in NW India and the model results of vertical structure of the atmospheric boundary layer and the turbulent heat fluxes at the interface are discussed in this paper.

1. Introduction

In recent years, considerable effort is being devoted to study the turbulent entrainment in the atmospheric boundary layer. The atmospheric boundary layer is characterized by the large vertical gradients of wind velocity, temperature and humidity. Starting from a low value at sunrise, the depth of atmospheric boundary layer generally increases by turbulent entrainment druing daylight hours and is observed more predominently over the semi-arid or arid regions of NW-India during hot and dry summer season. This turbulence results primarily due to diurnal variations of surface temperature caused by the exchange of radiation at the air-soil interface and is maintained by the buoyancy associated with a positive surface heat flux. Thus, in order to understand the low-level turbulence in the atmospheric boundary layer, a detailed knowledge of thermal interaction between the atmosphere and the earth's surface is desired.

The objective of this paper is to describe some preliminary results of a numerical study which was aimed at evaluating the effect of thermal exchange at the air-land interface on the turbulent structure of the boundary layer over NW India.

2. The model description

In the model formulation of this study, the principal processes parameterized are boundary layer turbulence and cloud-dependent radiative heating. The model is unique in that it provides solutions to the primitive equations including complex sources and sinks of momentum, heat and moisture using a fine vertical grid in the coupled atmopshere-soil boundary layer.

The general form of the model equation is the Eulerian conservation equation given by :

$$\frac{\partial X_i}{\partial t} + \mathbf{V}\cdot \nabla X_i = \frac{\partial}{\partial z}\left[K_i\,(R_i, z)\ \frac{\partial X_i}{\partial z}\right] + A_i \quad (1)$$

where the dependent variables X_i and the associated source terms A_i are as listed in Table 1. In this equation t is time, $\mathbf{V}$ is three dimensional velocity vector, z is height, K_i is the eddy exchange coefficient for the dependent variables and R_i is the Richardson number.

The Eqn. (1) is non-linear through the dependence of the eddy exchange coefficients on the Richardson number. This implies a dependence on the vertical gradients of the horizontal velocity components, temperature and humidity. The

TABLE 1

Index	Dependent variable	Source (sink) term
1	U (eastward wind component)	Geostrophic wind
2	V (northward wind component)	Geostrophic wind
3	T (temperature)	Solar and infrared radiation, latent heating/cooling
4	q (specific humidity)	Evaporation
5	P (pollutant)	Pollutant source term

internal source terms also take on non-linear dependence on the dependent variables of the problem in the temperature equation. The values of dependent variables are prescribed as a function of time at the upper and lower boundaries. Geostrophic winds at any height in the atmospheric layer are computed from the upper boundary value and from the horizontal temperature gradients by the use of thermal wind equation. Vertical velocities at any height relative to terrain are computed from the horizontal velocity gradients by use of the continuity equation and is assumed to be zero at the interface.

In the soil layer, temperature is the only dependent variable and is governed by the equation

$$\frac{\partial I}{\partial t} = \frac{\partial}{\partial z}\left\{ K_e(z) \ \frac{\partial T}{\partial z}\right\} \tag{2}$$

where $K_e(z)$, the soil thermal diffusivity, is specified as a function of height. For the soil subsurface, albedo, density, specific heat, thermal diffusivity and moisture parameter are specified.

At the air-soil interface, the dependent variables are computed at each time step. The specific humidity q_I is computed from

$$q_I = M.q_s + (1-M).\ q_1 \tag{3}$$

where q_1 is the humidity at the first atmospheric grid level and q_s is the specific humidity at saturation. The moisture parameter M takes intermediate fractional values between 0 (dry soil) and 1 (water) for moist soil surfaces.

The evaporation E is computed from

$$E = \rho_a K_{ea} \ \frac{\partial q}{\partial z} \tag{4}$$

where ρ_a is the air density and K_{ea} is the vertical diffusion coefficient for the atmosphere.

The interface temperature is computed by an iterative solution to the non-linear heat balance equation

$$T = \left[\frac{1}{\varepsilon\sigma}(R_H + R_a + R_X + P_T - A - S - LE)\right]^{1/4} \tag{5}$$

where ε is emissivity, σ is Stefan's constant, R_H is net solar energy absorbed at the surface, R_a is infrared radiation incident on the interface, R_X is artificial heat source, P_T is heat energy due to precipitation, A and S are sensible heat fluxes from the interface to atmosphere and soil respectively, L is latent heat and E is surface evaporation. The details of the calculation of each of these terms have been given by Jacobs (1973). The solar and infrared radiative fluxes are calculated by the procedure described elsewhere (Lal 1976). The temperature at the interface is assumed to be continuous for the calculation of individual terms in the heat balance equation.

The pollutant concentration at the interface is calculated from a flux value using the formula

$$P(z_I, t) = P(z_{I+1}, t) - \frac{F.\triangle z}{K_e(z_{I+1}, t)} \tag{6}$$

where F is the surface flux of the pollutant. At higher elevation the pollutant is assumed to be instantaneously mixed in the layer that is bounded by the mid-points of the layers adjacent to the grid level. This results in a source term $S_p = F_e/\rho\triangle z$ where, F_e is the flux of pollutant at elevated height.

The Richardson number in the atmosphere is computed as a function of height. In the atmospheric layer, Richardson number is assumed to be zero at the first grid level adjacent to the interface for computational convenience.

The vertical diffusion coefficients are internally computed in the model as a function of the vertical gradients of wind, humidity and temperature. The exchange coefficient formulas used here for the atmospheric layer are consistent with the profile formulas discussed by Pandolfo (1966). The following limiting values for the exchange coefficients are imposed in the atmospheric layer :

$$10^4 \leqslant K_{e,m} \leq 10^7,\ z > 100 \text{ m}$$

$$\wedge \leqslant K_{e,m} \leq 10^7,\ z < 100 \text{ m}$$

where $\wedge = z_0^2$, z_0 being the roughness height.

The radiative source terms in the temperature equation are computed numerically as a function of time of day, day of the year and prescribed cloud state as well as the vertical distributions of temperature and humidity.

The numerical solutions for equation (1) in the one dimensional model are obtained on an arbitrarily spaced vertical grid using a specified time step. The finite difference analogues to the diffusion equations are adapted from an implicit, three time-level scheme. Weighted centred differences approximate the vertical derivatives for the vertical grid. Upwind differences approximate the horizontal spatial derivatives for the advection terms.

3. The input data

The specific data requirements for the model simulation are listed in Table 2.

Our simulation experiment was based on input meteorological data for the period 24-26 April 1966 collected at New Delhi. A non-uniform grid system was used and the vertical layer extended from 10 cm below the air soil interface to 1500 metres above the surface. The grid levels used for the simulation are listed in Table 3.

The initial time for the simulation was chosen to be 1200 hr (Indian Standard Time) of 24 April 1966 and a 48 hour simulation was carried out. The length of time step was 2·4 minutes which made 50 time steps equal to 2 simulation hours. The temperature, humidity and wind components were provided as initial conditions at all atmospheric grid levels except at the upper boundary where the dependent variables were specified at each time step. The horizontal gradients of air and soil temperature, wind and humidity were input to the model and remained constant throughout the simulation. A time-varying cloud cover representative of the observed clouds during the simulation period was input to the model. A vertical profile of the aerosol concentration was input to the model. The thermal properties of the soil used in the simulation and the physical parameters assigned at the interface are listed in Table 4.

4. The results

A comparison of model simulated meteorological conditions and the actual observations is made. The major components of thermal exchange at the air-soil interface as simulated by the model are also discussed here.

TABLE 2

Condition	Input
Initial	Vertical profiles of dependent variables
Upper and lower boundary	Specification of dependent variables for the duration of simulation
Gradients	Horizontal gradients of the dependent variables
Clouds	Cloud levels and types as a function of time for the simulation period
Interface	Geostrophic winds

TABLE 3

Hight (depth) of vertical grid levels (m)
1500·0
1250·0
1000·0
850·0
700·0
600·0
500·0
400·0
250·0
150·0
100·0
75·0
50·0
25·0
4·0
1·0
0·5
0·0
0·0
0·05
0·10

TABLE 4

Parameter	Value
Albedo	0·2
Soil heat conductivity	0·0022
Soil specific heat	0·33
Soil density	1·78
Thermal diffusivity	0·0037
Roughness height (cm)	100
Moisture parameters	0·1

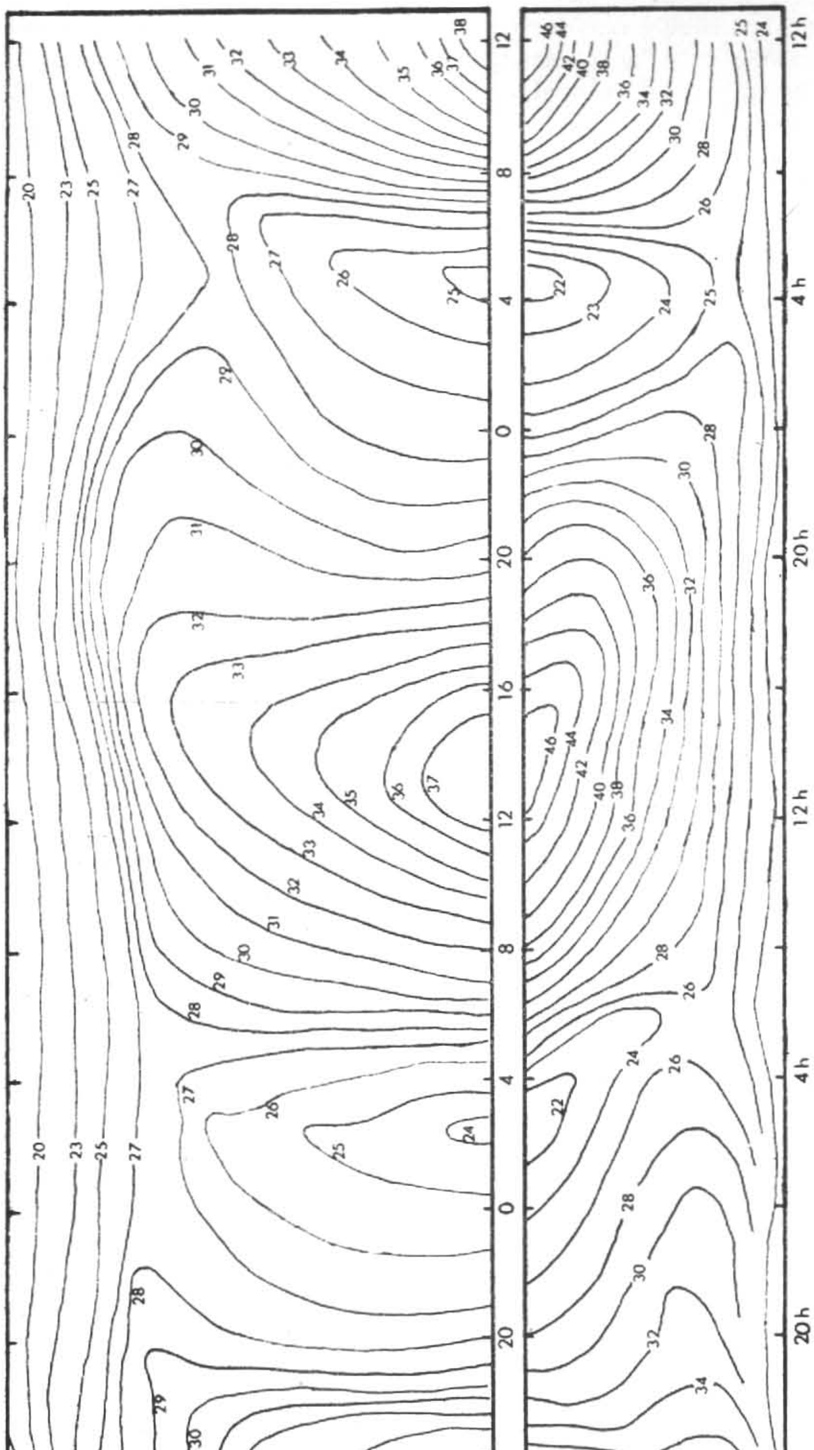

Fig. 1. The model simulated temperature variations (°C) in the atmospheric (upper) and soil (lower) boundary layers

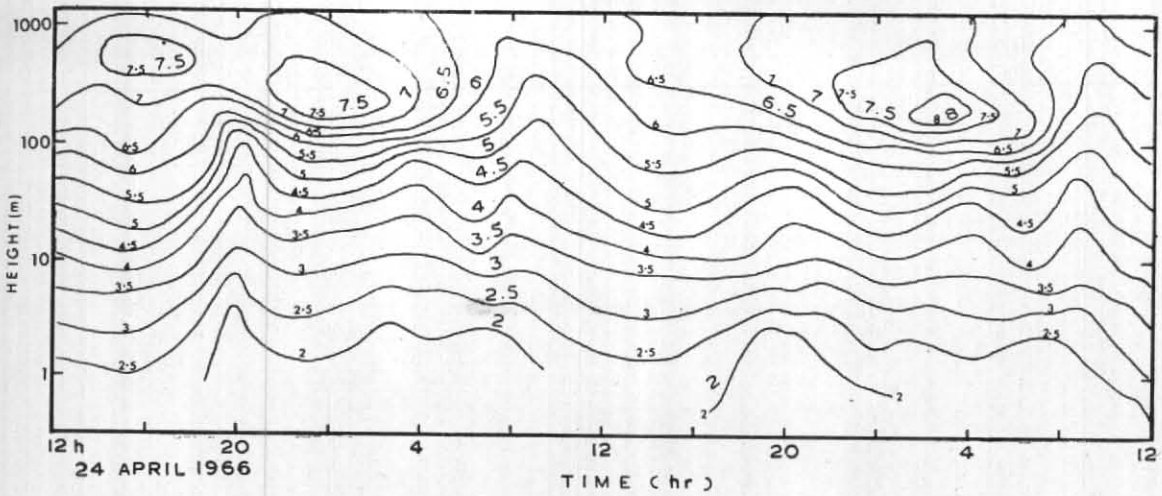

Fig. 2. The model simulated wind velocity field (m/sec) in the atmospheric boundary layer

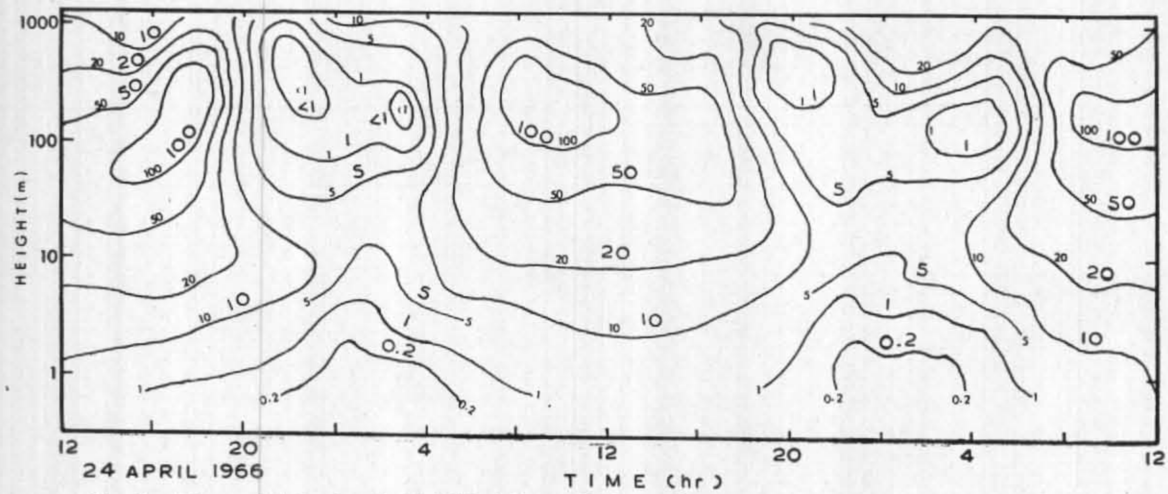

Fig. 3. The model simulated eddy diffusivity coefficient (m^2/sec) in the atmospheric boundary layer

The simulated temperature variations in the atmospheric and soil boundary layers are shown in Fig. 1. The agreement with the observations is fairly good except that the predicted temperature close to the surface is about 2° to 3°C lower than the observed temperature throughout the period of simulation. The predicted interface temperature reaches to about 50°C in the daytime and falls to 20°C at night; its diurnal range attains to 30°C. This is in contrast to the variations in the air layer close to the surface where diurnal range is of the order of 8°C. Under such conditions, superadiabatic lapse rate of air temperature appears especially stronger near the interface. The atmosphere tends to be isothermal near the surface in the evening and night comparing favourably with observation.

The time variation of simulated eddy diffusivity coefficient is given in Fig. 3. The eddy diffusivity profiles show an increase in turbulent activity during the day when it increases in the lowest portion of the mixed layer, reaches a maximum at heights about half the boundary layer thickness and decreases close to unity at the top of the boundary layer. The largest values of eddy diffusivity occur during

TABLE 5

Diurnal variation of heat balance components at the air-soil interface (cal cm^{-2} min^{-1})

Time (hr)	S	B	L	V
12 (24 April)	0·72	0·18	0·38	0·16
14	0·66	0·10	0·39	0·17
16	0·21	—0·03	0·24	0·00
18	—0·12	—0·09	0·02	—0·05
20	—0·10	—0·09	0·00	—0·01
22	—0·09	—0·08	0·00	—0·01
0 (25 April)	—0·09	—0·07	0·00	—0·02
2	—0·09	—0·06	0·00	—0·03
4	—0·09	—0·04	—0·01	—0·04
6	0·07	—0·01	0·08	0·00
8	0·38	0·16	0·13	0·09
10	0·63	0·19	0·31	0·13

TABLE 6

Daily totals of the components of thermal exchange from 12 hr on 24 to 10 hr on 25 April 1966 (cal cm^{-2} day^{-1})

	S	B	L	V	Total exchange
Positive	320·4 (80·7%)	75·6 (19·1%)	186·0 (46·8%)	66·0 (16·6%)	
Negative	69·6 (17·5%)	56·4 (14·2%)	1·2 (0·3%)	19·2 (4·8%)	397·2 (100%)
Net exchange	250·8 (100%)	19·2 (7·7%)	184·8 (73·7%)	46·8 (18·6%)	

S : Net radiative flux (short and longwave) received (+) or given off (—) at the interface

B : Heat flux through conduction into (+) or out of (—) the soil

L : Sensible heat flux to (+) or from (—) the air

V : Latent heat due to evaporation (+) or condensation (—) of moisture at the interface

$$S = B + L + V$$

the day and are of the order of 100 m^2 sec^{-1} which is in agreement with other investigations (Orlanski *et al.* 1974; Deardorff 1967). The evening and night time temperature and eddy diffusivity profiles exhibit some interesting features. In about three to four hours after the surface starts cooling at around 1400 hours, turbulence is virtually extinguished in the bulk of the boundary layer. This unusual behaviour of the nocturnal boundary layer has been observed elsewhere by Kaimal *et al.* (1976). The discontinuities in the boundary layer depth near sunrise and sunset are indicative of sharp changes in turbulent activity associated with high surface cooling rates.

The components of thermal exchange at the air-soil interface as simulated by the model are listed in Table 5. It is apparent that S, the heat gain at the interface by radiation, begins somewhat later than the sunrise and attains its maximum at about noon. During the night negative values of S, controlled only by the terrestrial radiation, are nearly constant. A retardation of about half an hour is observed for the beginning time of positive B to that of positive S in the early morning hours. L and V both are positive (upward heat flux and evaporation) during the day and then maxima occurs about one hour after the noon.

The main difference of the mode of thermal exchange at the interface between day and night time can be observed in the daily totals of positive and negative thermal exchange components (Table 6). The model simulated heat flux to the air layer through the ground surface is very intense in the daytime. In fact, 45% of the daily total exchange of heat is conveyed into the air by L as against 17% of the downward flux conveyed into the soil by conductive process B. In daily net exchange, 74% of the accumulated heat by radiative process is expended by L as against only 8% by B. The turbulence in the lowest 50 metre is too small to transport all of this heat across the air-land interface with the result that a large super-adiabatic lapse rate is established in the lowest layer and the resulting instability allows warm buoyant parcels of varying sizes to rise through the air, thereby transporting the heat to higher layers where it is finally mixed into the environmental air.

5. Conclusions

The results demonstrate that the model, given appropriate input data, reproduces the near-surface atmospheric conditions over a semi-arid region. The maximum departure between simulated and the observed interface temperature is of the order of 2° to 3°C. The simulated near-surface wind field is in good agreement with the observation.

While we are yet to fully examine the major thermal exchange components and their contribution to the large scale turbulent eddies in the lowest atmospheric layer, it is hoped that the incorporation of a suitable convective parameterization scheme into the model would account better for the bulk effects of sub-grid scale processes.

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DISCUSSION

P.S. Srivastava : Have you considered extending this model to three dimensions?

Author : No.

Mukut B. Mathur : I presume that the lower boundary is in the soil. Do you have a separate surface layer or do you have one layer from interface to 1500 m?

How is your K (exchange coefficient) calculated? What is its formulation for stable and unstable conditions?

Author : The lowest grid level is in the soil and the atmospheric layer extends from the interface to 1500 m.

The exchange coefficient formulas for the atmospheric layer for different stability conditions are consistent with the profile formulas as discussed in detail by Pandolfo (JAS 1966). Computed values for eddy exchange coefficients for extreme stability are, however, meaningless whenever they fall below estimates for minimum values in the atmospheric boundary layer. The values over a reasonable range are therefore allowed for the atmosphere.