Atmospheric convection and its role in vertical mass transport

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Results of a theoretical study of vertical mass transport under conditions of developed two-layer convection are presented. A spatial eddy-resolving model of an ensemble of thermals, cumulus and cumulonimbus clouds is used to describe motions in the atmosphere. The equations account for the active role of aerosols in processes of cloud and precipitation formation.

Introduction

In the description of mass transport processes in the atmospheric boundary layer (ABL), including convectively driven ones, numerical models are commonly used which are based on semi-empirical diffusion equations and various closure assumptions. In these models, the irregular mesoscale processes in the ABL are parametrized as turbulent (subgrid-scale) processes. However, solutions based on this approach overlook many important structural features of the convective ABL. In eddy-resolving models, processes of atmospheric convection are described using the so-called Large Eddy Simulation (LES) models in which eddies on spatial scales greater than 100 m are simulated by means of non-hydrostatic thermo-hydrodynamic equations, while smaller-scale eddies are parametrized. LES models have proven to be an efficient tool in studies of the internal dynamics and development of clouds.[1,2] The ever-growing interest in this problem stems largely from the ecologic hazard caused by acid rain, the need for cloud parametrization in climate and general circulation models, and the need for a more detailed hydrologic cycle, among others. At the present juncture, domestic simulation studies of cloud dynamics have almost ceased.

In this paper, a spatial eddy-resolving model of ensembles of thermals, cumulus, and cumulonimbus clouds is used to study convective transport of momentum, water vapor, heat, moisture, and aerosols. The active role of aerosols in processes of cloud and precipitation formation is taken into account. The sources of admixtures supplied to the atmosphere are assumed to be the processes of saltation and deflation (windblown transport of aerosol particles from the underlying surface). The admixed particles transported to the cloud layer serve as coagulation nuclei, leading to an intensification of precipitation formation processes. In turn, precipitation washes aerosol out of the atmosphere and stops the supply of aerosol from the underlying surface.

As a baseline model, we use the spatial LES model,[3] which takes account of processes of precipitation formation,[4,5] as well processes of admixture spreading[3]:

$$\frac{\mathrm{d}u}{\mathrm{d}t} = -\frac{\partial \pi}{\partial x} + l(v - v_{\mathrm{g}}) + D_{xy}u + \frac{\partial}{\partial z}K\frac{\partial u}{\partial z}, \quad (1)$$

$$\frac{\mathrm{d}v}{\mathrm{d}t} = -\frac{\partial \pi}{\partial y} - l(u - u_{\mathrm{g}}) + D_{xy}v + \frac{\partial}{\partial z}K\frac{\partial v}{\partial z}, \quad (2)$$

$$\frac{\mathrm{d}w}{\mathrm{d}t} = -\frac{\partial \pi}{\partial z} + g\left(\frac{\theta}{\Theta} + 0.61q - q_{\mathrm{l}}\right) + D_{xy}w + \frac{\partial}{\partial z}K\frac{\partial w}{\partial z}, \quad (3)$$

$$\frac{\mathrm{d}\theta}{\mathrm{d}t} + \frac{\partial \Theta}{\partial z}w = D_{xy}\theta + \frac{L}{c_p}(CR - ER_{\mathrm{c}} - ER_{\mathrm{r}}) + \frac{\partial}{\partial z}K_T\frac{\partial \theta}{\partial z}, \quad (4)$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0, \quad \frac{\mathrm{d}}{\mathrm{d}t} = \frac{\partial}{\partial t} + u\frac{\partial}{\partial x} + v\frac{\partial}{\partial y} + w\frac{\partial}{\partial z},$$

$$D_{xy} = \frac{\partial}{\partial x}K_x\frac{\partial}{dx} + \frac{\partial}{\partial y}K_y\frac{\partial}{\partial y}, \quad (5)$$

$$\frac{\mathrm{d}q}{\mathrm{d}t} = ER_{\mathrm{c}} + ER_{\mathrm{r}} - CR + D_{xy}q + \frac{\partial}{\partial z}K_T\frac{\partial q}{\partial z}, \quad (6)$$

$$\frac{\mathrm{d}q_{\mathrm{c}}}{\mathrm{d}t} = CR - ER_{\mathrm{c}} - k_1(q_{\mathrm{c}} - a) - k_2 q_{\mathrm{c}} q_{\mathrm{r}}^{0.88} + D_{xy}q_{\mathrm{c}} + \frac{\partial}{\partial z}K_T\frac{\partial q_{\mathrm{c}}}{\partial z}, \quad (7)$$

$$\frac{\mathrm{d}q_{\mathrm{r}}}{\mathrm{d}t} = \frac{1}{\rho}\frac{\partial \rho V_{\mathrm{r}} q_{\mathrm{r}}}{\partial z} - ER_{\mathrm{r}} + k_1(q_{\mathrm{c}} - a) + k_3 q_{\mathrm{r}} s^{0.88} + D_{xy}q_{\mathrm{r}} + \frac{\partial}{\partial z}K_T\frac{\partial q_{\mathrm{r}}}{\partial z}, \quad (8)$$

$$\frac{\mathrm{d}s}{\mathrm{d}t} - w_0\frac{\partial s}{\partial z} = -k_3 q_{\mathrm{r}} s^{0.88} + D_{xy}s + \frac{\partial}{\partial z}K_s\frac{\partial s}{\partial z}. \quad (9)$$

Here t is time; u, v, and w are the wind components along the x, y, and z axes; θ is the deviation of the potential temperature from its value $\Theta(z, t)$ for the unperturbed atmosphere; q is the water vapor mixing ratio (humidity); q_{c} is the mixing ratio of cloud

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droplets (cloud liquid water content (LWC)); q_r is the mixing ratio of precipitable water droplets (precipitable LWC); $q_l = q_c + q_r$; s is the mixing ratio of admixed particles; g is the acceleration due to gravity; l is the Coriolis parameter; L is the specific heat of vaporization; c_p is the specific heat of air at constant pressure; K is the coefficient of vertical subgrid-scale turbulent exchange; $K_T = K/\mathrm{Pr}$; $K_s = K/\mathrm{Sc}$; Pr and Sc are the turbulent Prandtl and Schmidt numbers in the ABL; $V_r = 36.34\ (\rho q_r)^{0.1364}$ is the terminal falling velocity of raindrops; T is the mean temperature; $k_1(q_c - a)$ accounts for the autoconversion rate of raindrops and cloud droplets; $k_1 = 0$ for $q_c < a$ and $k_1 = 10^{-3}\ \mathrm{s}^{-1}$ for $q_c \geq a = 10^{-3}$ g/g; ER_c is the evaporation rate of cloud droplets; CR is the condensation rate of water vapor; $CR = 0$ for $q < q_n$, where $q_n = \ = \dfrac{3.8}{p}\exp\left(\dfrac{17.5(T-273)}{T-36}\right)$ is the saturation specific humidity; and p is pressure, T is mean temperature. For $q \geq q_n$, the algorithm for calculating changes in the fields due to condensation has the form

$$q^{i+1} = q_a - (q_n + q_a - q_{na})\left(1 + 0.622\frac{AL}{Pc_p}\right)^{-1};$$

$$\theta^{i+1} = \theta_a + \frac{L}{c_p}(q_a - q^{i+1});\quad q_c^{i+1} = q_{ca} + q_a - q^{i+1},$$

where the subscript a indicates the value of the corresponding variable obtained after calculation of spatial transport; the index i numbers the time steps; $A = \Delta e_{na}/(\Delta T)$, the symbol Δ denotes the difference between the ith and $(i-1)$th time steps; e_n is the saturation water vapor pressure; $k_2\, q_c\, q_r^{0.88}$ is the rate of variation of q_c and q_r due to coagulation of cloud droplets with raindrops; $k_3\, q_r\, s^{0.88}$ is the rate of variation of q_r and s due to coagulation of raindrops with the admixed particles and the washing-out of aerosol from the region of precipitation. In the ER_c calculations, it was assumed that a part of the cloud droplets evaporates instantaneously to give complete saturation of air. If the evaporated cloud water is insufficient to achieve air saturation, rainwater will evaporate at a rate given by the formula

$$ER_r = -\frac{c(Q/Q_n - 1)q_r^{0.88}}{5.48\cdot 10^5 + 4.1\cdot 10^6/e_n},$$

where $c = 1.6 + 5.7 \cdot 10^{-4}\ V_r^{1.5}$.

Initial and boundary conditions

It is assumed that

$$K\frac{\partial u}{\partial z} = c_u|\vec{u}|u,\ K\frac{\partial v}{\partial z} = c_u|\vec{u}|v;\ w = 0;$$

$$\rho c_p K_T\frac{\partial \theta}{\partial z} = H_0,\ q_c = 0,\ \frac{\partial q_r}{\partial z} = 0,$$

$$K_T\frac{\partial q}{\partial z} = c_\theta|\vec{u}|(q - q_0),\ K_s\frac{\partial s}{\partial z} + w_0 s = \beta s - \Gamma \text{ for } z = 0,$$

at the ABL lower boundary, where c_u and c_θ are the resistance and heat exchange coefficients calculated from the model of a quasi-stationary sub-layer; H_0 is a specified heat flux; q_0 is the humidity near the underlying surface (assumed to be equal to the saturation value); Γ and β are parameters of interaction of the aerosol with the underlying surface (see Ref. 3 for more details). It is also assumed that in the upper part of the ABL the atmosphere has a stable stratification and, consequently, that convection decays at the upper boundary of the region $z = H$:

$$u = u_g;\ v = v_g;\ w = 0;\ \frac{\partial \theta}{\partial z} = \gamma;\ q = q_H;$$

$$\frac{\partial q_c}{\partial z} = \frac{\partial q_r}{\partial z} = 0 \text{ for } z = H.$$

The u_g and v_g values are assumed to be known, while γ defines a stable stratification at high altitudes. At the lateral boundaries of the region, periodic conditions are assumed.

The initial conditions are specified in terms of the humidity distribution and the assumption of no clouds or precipitation:

$$\theta = \Theta(z),\ q = Q(z),\ q_c = q_r = 0 \text{ for } t = 0.$$

The field $Q(z)$ is assumed to be uniform in the horizontal direction, and its values are calculated from the relative humidity f_q. The latter is defined as a linear function of z, taking the value $f_q = 90\%$ at the lower boundary and assumed to decrease by 10% per kilometer of altitude. Convection is initialized as a series of weak heat pulses.[5] If a layer with unstable stratification is present near the Earth's surface, heat perturbations trigger the development of convection; otherwise, the perturbations decay.

Let us consider results of model calculations made with different values of H_0 for $u_g = 5\ \mathrm{m \cdot s^{-1}}$ and $v_g = 0$. When $H_0 \leq 10\ \mathrm{W \cdot m^{-2}}$, the upward motions are not strong, and air moisture does not reach the condensation level. In this case, an ensemble of thermals forms, but clouds do not develop. As the supplied heat flux increases to $H_0 = 15\ \mathrm{W \cdot m^{-2}}$, light cloudiness develops with small LWC values. These clouds do not precipitate, and their cloud cover (n) is equal to 0.03–0.05. The cloud layer is located below the inversion level (one-layer cloud convection[5]). Deep penetrative convection takes place for $H_0 = 30\ \mathrm{W \cdot m^{-2}}$. A large instability energy favors the development of strong upwelling motions ($w = 5 - 7\ \mathrm{m \cdot s^{-1}}$), leading to formation of distinct cumulus clouds.

Figure 1*a* shows a contour plot of the relative humidity field f_q (in %) in the horizontal plane at the level $z = 1600$ m at $t = 2$ h into the simulation. The particles, reaching the condensed-phase level, clump together to form closed regions with $f_q = 100\%$; inside these regions, cloud water droplets form.

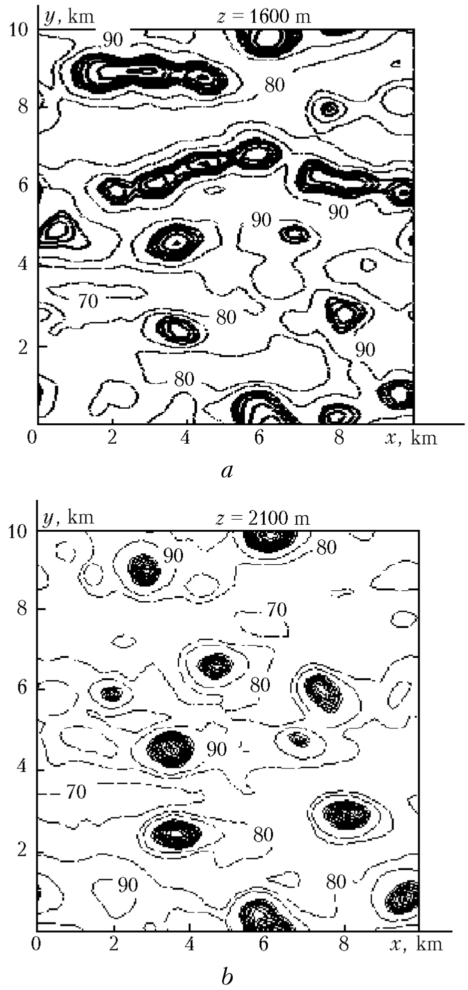

Fig. 1. The horizontal structure of the relative humidity (thin lines) and cloud water (thick lines) fields.

The contours in Fig. 1 are shown by thick lines (with contour interval $\Delta q_c = 10^{-4}$ g/g). The calculated cloud cover is close to $n = 0.2$. Rainwater is absent. Closely located convective cells form cloud streets oriented approximately in the average wind direction, consistent with the observed data.[5,6] The horizontal extent of the cloud streets is 1–4 km, again in agreement with the observations. With increasing altitude, less intense elements of the ensemble decay, and the cloud streets disintegrate (Fig. 1*b*). The vertical size of the clouds is 1–2 km.

The obtained characteristics of the clouds and the entire ensemble are in good agreement with quantitative estimates and data of experimental studies for non-precipitating convective clouds.[5–7] Integrating the equations out to 4 h, we observed the formation of five precipitating clouds 2 km in thickness. Traces of fallen rain ("pools") are oval in shape, slightly elongated along the wind direction with a maximum value of total precipitation equal to 112 g/m^2. The regional average total precipitation was 0.73 g/m^2. Such rainfall is produced by convective clouds with an average thickness of 3 km.[7,8]

The obtained vertical structure of the fields is in good agreement with the observed two-layer cloud convection, when clouds overlie a "dry" convection layer above the inversion level.[5] This is illustrated in Fig. 2*b*, in which curve *1* reflects the vertical structure of the convective heat flux. The dry-convection exchange mechanism dominates in the lower one-kilometer region. The formation of an inversion layer is associated with negative fluxes near the $z = 1300$ m level. At higher altitudes, the convective flux again becomes positive, this time, though, due primarily to latent heat flux (curve *2*), which significantly exceeds the heat transport due to convective motions. The subgrid-scale turbulence (curve *3*) is significant only near the lower boundary.

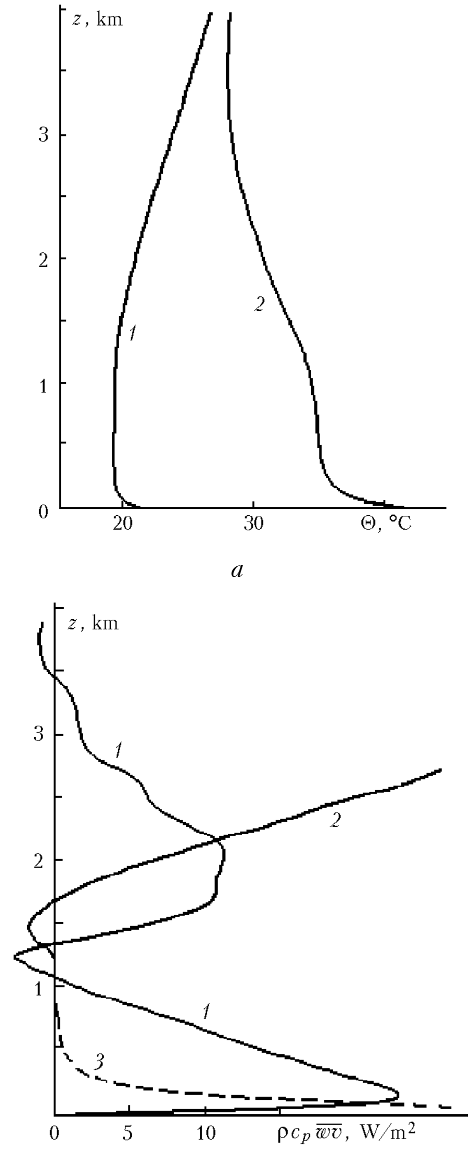

Fig. 2. Mean profiles of (*a*) the potential and pseudo-potential temperature (curves *1* and *2*); and (*b*) the convective, latent, and subgrid-scale heat fluxes (curves *1*, *2*, and *3*).

Figure 2*a* plots the profile of the potential temperature $\Theta(z)$ (curve *1*). As can be seen, for $z > 1300$ m the ABL is stable in the sense of a dry adiabatic process ($\partial\Theta/\partial z > 0$). However, analysis of the pseudopotential temperature $\Theta_p(z)$ (curve *2* in Fig. 2*a*), representing a conservative characteristic of moist-adiabatic processes, shows that its vertical gradient is negative throughout most of the region $z < 3500$ m. This corresponds to unstable temperature stratification in saturated air. Thus, vertical heat, water vapor, and moisture transport leads to a change in the temperature and humidity fields, as well as to cloud and precipitation formation. The influence of convection on the character of the admixture distribution in the vertical direction is also significant: it leads to the formation of a 1-km thick layer with a weakly varying aerosol concentration, and with less than 1% of aerosol penetrating the cloud layer, in agreement with observations.[9,10]

Conclusion

The calculations performed in this study and their comparison with the observed data show that the eddy-resolving model proposed here may serve as an efficient tool for the study of convective transport of heat, water vapor, moisture, and aerosol, for the study of cloud and precipitation formation processes, and for the development of methods of convective parametrization in studies of large-scale processes. The model is easily generalized to the case of "deep" convection, when rain and hail clouds cover the entire troposphere.

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