HYDRODYNAMIC MESOSCALE FORECAST OF CLOUDINESS AND AEROSOL FORMATIONS EVOLUTION

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We describe in this paper the mesoscale models of cloudiness evolution and impurity dispersal based on the sets of complete hydrothermodynamics equations and turbulent diffusion equations, respectively. The evolution of moisture and temperature fields is described using functions, invariant with respect to water phase transitions namely, the functions of specific moisture content and equipotential temperature. Physical content of the models due to the use of parametrization schemes of the main processes on a subgrid scale as well as their high spatial resolution have made possible their use for simulation and forecasting of cloudiness and fields of aerosol formations.

INTRODUCTION

Solution of many scientific and applied problems is connected with the forecast of dynamics of cloudiness and fields of aerosol formations (AF) of natural and anthropogenic origin. An approach, based on the use of mathematical models built on the base of hydrothermodynamics equations taking account for one or another assumptions, is accepted as most promising and producing better practical results including those for mesometeorological scale. The physical concept of the atmosphere as a compound medium is a basis for constructing the models of such a kind. In this case the parametric consideration of the existing feedbacks enables one to perfect the physical content of the models.

It is well known that a great variety of factors affects the formation of cloudiness and aerosol fields. In connection with the intense anthropogenic pollution of the atmosphere, the above-mentioned factors include both natural and anthropogenic perturbations. They are more pronounced in the case of aerosols that strongly absorb shortwave solar radiation (smoke, dust, soot). The reason is that highly absorbing aerosol causes the horizontally inhomogeneous volume heat release, which makes a great impact on dynamic characteristics of the atmosphere.

Thus, the prediction of AF evolution is not only an independent problem, but it should be used for taking into account nonadiabatic heat fluxes when predicting hydrometeorological fields. The approach proposed to the forecast of cloudiness and aerosol fields is based on the model of moist atmosphere, where the model of the pollution dispersal makes up an individual block. This paper is a continuation of the work done in Ref. 1.

BASIC EQUATIONS OF THE MODEL

Let us consider a limited area $\Omega\times\Omega_t$, corresponding to the typical spatiotemporal scales of simulated mesoscale atmospheric processes, where $\Omega = \{0 \le x \le X;\quad 0 \le y \le Y;\quad z_0 \le z < z_H\}$ and $\Omega_t = \{0 \le t \le t_k\}$. To describe the evolution of cloud fields and thermodynamic regime of the atmosphere, a model has been selected, which is based on the set of complete equations of hydrothermodynamics. In this case, for correct description of water vapor condensation processes and temperature evolution we use the functions, invariant relative to water phase transitions, namely, the equipotential temperature Π and specific moisture content S, see Ref. 2. Taking this fact into account, in the left system of Cartesian coordinates (the x-axis is directed to the east, y-axis – to the north, and z-axis – vertically upward) the model equations are:

a) equations of motion:

$$\frac{\partial \rho u}{\partial t} = -\,\mathrm{div}(\rho u\mathbf{V}) - \frac{\partial P}{\partial x} + \rho f v - \frac{\partial \tau^{xz}}{\partial z} + \rho F_u, \qquad (1)$$

$$\frac{\partial \rho v}{\partial t} = -\,\mathrm{div}(\rho v\mathbf{V}) - \frac{\partial P}{\partial y} - \rho f u - \frac{\partial \tau^{yz}}{\partial z} + \rho F_v; \qquad (2)$$

b) equation of continuity

$$\mathrm{div}(\rho\mathbf{V}) = 0; \qquad (3)$$

c) equations of heat flux and humidity transfer:

$$\frac{\partial \Pi}{\partial t} = -\,\mathbf{V}\mathrm{grad}\Pi + \frac{\varepsilon_r}{\rho C_p} - \frac{1}{\rho C_p}\frac{\partial Q_\pi}{\partial z} + F_\pi\,, \qquad (4)$$

$$\frac{\partial S}{\partial t} = -\,\mathbf{V}\mathrm{grad}S\,\frac{1}{\rho}\left(\frac{\partial Q_K}{\partial z} + \frac{\partial Q_S}{\partial z}\right) + F_S, \qquad (5)$$

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where

$$\Pi = (P_0/P)^{R/C_p}[T + (L/C_p)q]; \quad (6)$$

$$S = \begin{cases} q_m + \delta & \text{in clouds,} \\ q & \text{outside the clouds;} \end{cases} \quad (7)$$

d) equations of static and state:

$$\frac{\partial P}{\partial z} = -\rho g, \quad (8)$$

$$P = \rho R T_v. \quad (9)$$

In Eqs. (1)–(9) the following designations are used: t is time; $\mathbf{V} = (u, v, w)$ is the velocity vector with the components along the axes x, y, and z; ρ is the air density; P is the pressure; f is the Coriolis parameter; Q_π and Q_S are the vertical turbulent fluxes of heat $C_p\Pi$ and moisture content S; Q_K is the sedimentation flux of cloud elements; ε_r is the radiation heat flux, C_p is the specific heat of air at constant pressure; T_v is the virtual temperature; δ is the specific water content of cloudiness; q is the mass fraction of water vapor; q_m is the saturated value of water vapor mass fraction; L is the specific heat of water vapor condensation; g is the acceleration of gravity; $P_0 = 1000$ GPa; R is the universal gas constant; F_u, F_v are the projections of turbulent friction force on the x and y axes, respectively; F_π, F_S are the components taking into account the rate of variation of the corresponding meteorological parameter due to small-scale horizontal turbulent diffusion; τ^{xz} and τ^{yz} are the components of the vector of vertical turbulent momentum flow.

In clouds, where water vapor is saturated, the set of equations (1)–(9) is complemented by the ratio

$$q = q_m = 0.622\ [E(T)/P], \quad (10)$$

where $E(T)$ is the pressure of saturated water vapor at temperature T. An important advantage of the set of equations (1)–(9) is that the phase heat fluxes and transitions of atmospheric water vapor are allowed for directly by Eqs. (4) and (5) due to the invariance of the functions Π and S relative to phase transitions of the atmospheric moisture.

The boundary conditions relative to vertical coordinate are traditional, providing conservation of integral mass of the atmosphere: $\rho w = 0$ at $z = z_H$ and $w = \mathbf{V}\nabla z_0$ at z_0. The procedure of formulation of side boundary conditions is the following. In the boundary zone 120 km wide (six steps of grid) we use "viscous absorption" (the artificial viscosity with a large value of the viscosity coefficient, 10^6 m^2/s, is introduced) and weighted tendencies[3]:

$$\left.\frac{\partial A}{\partial t}\right|_j = \chi_j \left.\frac{\partial A_m}{\partial t}\right|_j + (1 - \chi_j) \left.\frac{\partial A_f}{\partial t}\right|_j, \quad (11)$$

where χ_j is the weighting factor with the following values: $\chi_j = \{0.0,\ 0.4,\ 0.7,\ 0.9,\ 1.0\}$, respectively, for $j = \{0,\ 1,\ 2,\ 3\}$ and $j > 3$, j is the number of a grid point along the normal from the boundary; A_m is the element value in the inner area; A_f is the background value of the element A.

PARAMETRIZATION OF THE SUBGRID-SCALE PROCESSES

For the set of equations (1)–(10) to be closed, we briefly describe the basic peculiarities of parametrization processes of subgrid scale, which are described in detail in Ref. 4.

The horizontal turbulent exchange is considered in the framework of the theory of nonlinear turbulence.[5] The terms F_u and F_v in Eqs. (1) and (2) are:

$$F_u = \frac{\partial \tau^{xx}}{\partial x} + \frac{\partial \tau^{xy}}{\partial y}, \qquad F_v = \frac{\partial \tau^{xy}}{\partial x} + \frac{\partial \tau^{yy}}{\partial y}. \quad (12)$$

The components of tension are the following:

$$\tau^{xx} = -\tau^{yy} = \mu D_1, \qquad \tau^{xy} = \tau^{yx} = \mu D_2, \quad (13)$$

where

$$D_1 = \partial u/\partial x - \partial v/\partial y, \quad D_2 = \partial v/\partial x - \partial u/\partial y.$$

For determining the horizontal turbulence coefficient μ we use the formulation of the closing scheme in the form[6]

$$\mu = k l^2 (D_1 + D_2). \quad (14)$$

Here k is the parameter allowing for variation (in the numerical experiments) of the level of model dissipation; l is the value, characterizing the scale of the simulated process, proportional to the step of the grid region Δn:

$$l = (k_0/\sqrt{2})\Delta n,$$

where $k_0 = 0.4$ is the empirical constant.[4]

The components F_π and F_S in Eqs. (4) and (5) are determined as:

$$F_\pi = \mu \nabla^2 \Pi, \quad F_S = \mu \nabla^2 S. \quad (15)$$

Vertical turbulent fluxes Q_π and Q_S in Eqs. (4) and (5) and the components of the vector of turbulent tension of friction in Eqs. (1) and (2) outside the surface sublayer ($z > z_h$) are presented using the expressions:

$$Q_\pi = -C_p \rho \nu \frac{\partial \Pi}{\partial z}, \quad Q_S = -\rho \nu \frac{\partial S}{\partial z},$$
$$\tau^{xz} = -\rho \nu \frac{\partial u}{\partial z}, \quad \tau^{xy} = -\rho \nu \frac{\partial v}{\partial z}, \quad (16)$$

where ν is the vertical coefficient of turbulence.

Calculations of ν in the boundary layer are made taking into account the nature of atmospheric stability and vertical wind gradient; as to the calculations of ν at altitudes above the boundary layer, we consider only the vertical wind shear.[4]

In the boundary sublayer ($z < z_h$, $z_h \sim 100$ m) the components of vector of tension τ are determined as follows:

$$\tau_h = -C_D \rho |\mathbf{V}_h| \mathbf{V}_h, \tag{17}$$

where C_D is the friction coefficient (in the numerical experiments this coefficient is equal to $C_D = 2.5 \cdot 10^{-3}$); the index h assigns the element value to the level z_h.

The values of vertical heat fluxes Q_π and moisture fluxes Q_S above the sea surface in the layer $z < z_h$ are estimated on the basis of the aerodynamic method,[7] and above the dry land these fluxes are evaluated using the equations of heat balance of the underlying surface. The calculations of Q_π and Q_S are described in Ref. 4.

The account for subgrid convection and stabilization of thermodynamic state of moist atmosphere in the model (1)–(10) is based on the "convective adaptation" method. The iteration (adaptation) procedure of unstable-stratified layer is used. In this case the critical value of the temperature vertical gradient is determined by relative humidity of air.[4]

The flux of cloud elements under the effect of gravity Q_k is described by the relationship[2]:

$$Q_k = -\rho(S - q_m)\tilde{v}, \tag{18}$$

where $\tilde{v}$ is the weighted mean (by mass) rate of cloud element fall, which is defined as the function of vertical cloud extension[3]:

$$\tilde{v} = \tilde{v}_m \exp\{-\beta[(z - z_\mathrm{L})/(z_\mathrm{U} - z_\mathrm{L})]\}. \tag{19}$$

Here $\tilde{v}_m$ is the maximum value, reached at the lower cloud boundary z_L; z_U is the upper cloud boundary; β is the parameter.

Determination of the radiation heat flux ε_r in Eq. (4) is described in Ref. 4.

MODEL OF IMPURITY DISPERSAL

For describing the process of impurity dispersal in the atmosphere, the semiempirical equation of turbulent diffusion is taken as a basis.[9]

Taking into account the above schemes of parametrization of vertical and horizontal turbulent exchange we can write:

$$\frac{\partial \varphi}{\partial t} + \mathbf{V}\mathrm{grad}\varphi - \mathrm{div}_S(\mu\ \mathrm{grad}_S\ \varphi) - \frac{\partial}{\partial z}\nu\frac{\partial \varphi}{\partial t} = \mathbf{d} + \mathbf{f}, \tag{20}$$

where $\varphi = \{\varphi_i,\ i = 1(1)N\}$ is the vector of specific impurity concentrations; φ_i is the concentration of the ith impurity; $\mathbf{j}$ and $\mathbf{d}$ are the vector-functions describing the aerosol sources and sinks, respectively; the subscript S denotes the operators in horizontal directions. Since the dimensions of most of the sources of highly absorbing aerosol are much less than the dimensions of the area Ω, i.e., can be considered as point ones, one source can be expressed in the form:

$$f(\mathbf{x}, t) = \begin{cases} Q\delta(\mathbf{x} - \mathbf{x}_i)(t - t_i) & \text{for a pulsed source,} \\ Q\delta(\mathbf{x} - \mathbf{x}_i), \text{ for a cw source,} & \end{cases} \tag{21}$$

where Q is the intensity of the source at a point with coordinates $\mathbf{x}_i = (x_i, y_i, z_i)$; t_i is the time of the pulse source existence; δ is the delta-function. When solving Eq. (20) we consider parametrically the processes of sedimentation, moist washing out, coagulation, and self-inductive rise, described in Ref. 1.

METHOD OF SOLUTION

Calculation of the impurity concentration is a procedure of a successive solution of Eq. (20) for every component of the vector φ. Numerical calculations by models (1)–(10) enables one to describe the evolution of cloud fields.

The cloudiness is determined in the layers where $S \geq q_m$. Quantitative characteristics, namely, the water content δ, are determined from the relationship $\delta = S - q_m(T)$. In clouds ($S \geq q_m$) in the numerical calculations it is assumed that $\delta^{n+1} = \delta^n + \Delta\delta^{n+1}$ and

$$\Delta\delta^{n+1} = \Delta S^{n+1} - 0.622\frac{L}{RP}\frac{E(T^{n+1})}{(T^{n+1})^2}\Delta T^{n+1}. \tag{22}$$

The temperature is determined from the relationship: $T^{n+1} = T^n + \Delta T^{n+1}$, where

$$\Delta T^{n+1} = \Delta\Pi^{n+1}\left[\left(\frac{P_0}{P}\right)^{0.286} + 0.622\frac{L^2}{RC_pP}\frac{E(T^n)}{(T^n)^2}\right]^{-1}. \tag{23}$$

Outside the clouds ($S < q_m$):

$$\Delta T^{n+1} = \Delta\Pi^{n+1} - (L/C_p)\ q^{n+1},\quad \delta^{n+1} = 0. \tag{24}$$

In Eqs. (22)–(24) the superscripts denote the solution at a proper time layer.

The models formulated provide numerical calculations. For this purpose the grid region Ω^h is introduced into the region Ω. In the vertical direction the grid has 19 levels (counting levels 110, 540, 990, 1460, 1950, 2470, 3010, 3590, 4210, 4870, 5570, 6340, 7190, 8190, 9160, 10360, 11780, 13610, and 16180 m) and separating layers, which, based on the data of the standard atmosphere, are about 50 GPa thick. In a horizontal plane the grid step Δn is taken to be equal to 20 km, and its size is 75×50 nodes.

The numerical solution is based on the splitting of Eqs. (1)–(10), (20) according to physical processes and coordinates. At the first step Eqs. (1)–(10) are solved in two stages. At the first stage the set of equations describing the adaptation of meteorological fields is solved with the use of an evident scheme "forward-backward". At the second stage the problem of advection is solved using the Lux-Vendroff scheme with the initial conditions, obtained as a result of an adaptation step. At the second step the model of impurity transfer (20) is applied. An evident TVD scheme is used at the advection step as well as the Krank-Nikolson scheme at the diffusion step.

CONCLUSION

Models of the atmosphere and aerosol transfer represent an imitation simulating complex. When entering the results of an objective analysis of mesometeorological fields (such an algorithm is cited as an example in Ref. 10) to the complex input and when performing the procedure of initiating meteorological fields, as well as the initial characteristics of aerosol formations and sources, it is possible to develop the prediction of evolution of cloudiness and aerosol formations. Based on the above-mentioned model we have studied the processes of wave cyclogenesis and formation of frontal cloudiness[11] as well as the evolution of aerosol formation of 150×150 km^2 size.

The results of numerical experiments point to a possibility of using the models in theoretical investigations.

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