Modeling of local transport of natural aerohydrosol in "atmosphere – water body" system

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A two-dimensional hydrodynamic model is proposed for description of atmospheric aerosol transport from land to water and impurity migration in the water medium. Turbulence characteristics in the both natural media are determined by solving K–ε equations. Transport of soil-sand aerosol to water area is considered.

Introduction

In studies of the interaction between the atmosphere and a water body by mathematical simulation, these two natural objects are usually considered separately with attention focused at one of them and another taken into account by means of simple parameterizations. In Refs. 1 and 2, which describe these two objects as a single system, the fine near-surface structure, in particular, the mechanism of turbulent energy transfer through the atmosphere/water interface, has defied direct description because of insufficient vertical resolution. The problem of the interaction between the atmosphere and ocean has been formulated in Ref. 1, but the issue of matching the turbulence fields in the both media has not been touched on.

In this paper, we consider the problem of local dynamic interaction of the atmospheric boundary layer (ABL) and a water body based on the 1D approximation of physical processes. Such an approach allows detailed consideration of processes of mass and energy exchange near the atmosphere/water interface because it allows using a finite-difference grid with as small as needed separation between nodes. The process of turbulent kinetic energy transfer from the ABL to the water body is explicitly described in the model by directly sewing together the turbulent characteristics on the surface.

The problem stated below can be considered as a particular case of the problem of modeling diffusion of soil-sand aerosol into the atmosphere and its transport from shore to the water area.

Models of turbulence in the atmosphere and water body

Let us write the system of equations for the non-stationary horizontally homogeneous ABL in the form:

$$\frac{\partial U}{\partial t} + W\frac{\partial U}{\partial z} = l\,(V - V_G) + \frac{\partial}{\partial z}\nu\frac{\partial U}{\partial z},$$

$$\frac{\partial V}{\partial t} + W\frac{\partial V}{\partial z} = -l\,(U - U_G) + \frac{\partial}{\partial z}\nu\frac{\partial V}{\partial z}, \qquad (1)$$

where U and V are the sought components of the horizontal velocity vector; W is the prescribed field of the large-scale vertical velocity; U_G and V_G are the geostrophic wind components; l is the Coriolis parameter; ν is the coefficient of the vertical turbulent exchange.

Turbulent exchange in the ABL is simulated based on the b–ε equations of the semi-empiric theory of turbulence[3]:

$$\frac{\partial b}{\partial t} + W\frac{\partial b}{\partial z} = \frac{\partial}{\partial z}\nu\frac{\partial b}{\partial z} + P - \varepsilon,$$

$$\frac{\partial \varepsilon}{\partial t} + W\frac{\partial \varepsilon}{\partial z} = \frac{1}{\sigma}\frac{\partial}{\partial z}\nu\frac{\partial \varepsilon}{\partial z} - c_1\frac{\varepsilon}{b}P - c_2\frac{\varepsilon^2}{b}, \qquad (2)$$

$$\nu = c_\nu\frac{b^2}{\varepsilon},$$

where b is the kinetic energy of turbulence (KET), ε is the rate of dissipation, $P = \nu\,(U_z^2 + V_z^2) - \nu\lambda\Theta_z/\mathrm{Pr}$ is the source of KET generation, λ is the buoyancy parameter, Pr is the atmospheric Prandtl number, c_ν, c_1, c_2, and σ are empiric constants; the potential temperature $\Theta(z)$ is assumed to be prescribed here.

Let us use the Ekman equations in the form

$$\frac{\partial \tilde U}{\partial t} = l\tilde V + \frac{\partial}{\partial z}\tilde\nu\frac{\partial \tilde U}{\partial z}, \quad \frac{\partial \tilde V}{\partial t} = -l\tilde U + \frac{\partial}{\partial z}\tilde\nu\frac{\partial \tilde V}{\partial z}, \qquad (3)$$

for description of flow in the water body. Here $\tilde U$ and $\tilde V$ are components of the drift velocity (from here on parameters with tilde correspond to those in the ABL, but in the water medium).

The equations of the b–ε model for the water body have the form

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$$\frac{\partial \tilde{b}}{\partial t} = \frac{\partial}{\partial z} \tilde{\nu} \frac{\partial \tilde{b}}{\partial z} + \tilde{P} - \tilde{\varepsilon},$$

$$\frac{\partial \varepsilon}{\partial t} = \frac{1}{\tilde{\sigma}} \frac{\partial}{\partial z} \tilde{\nu} \frac{\partial \tilde{\varepsilon}}{\partial z} - \tilde{c}_1 \frac{\tilde{\varepsilon}}{\tilde{b}} \tilde{P} - \tilde{c}_2 \frac{\tilde{\varepsilon}^2}{\tilde{b}}, \quad (4)$$

$$\tilde{\nu} = \tilde{c}_\nu \frac{\tilde{b}^2}{\tilde{\varepsilon}},$$

where $\tilde{P} = \tilde{\nu}\,(\tilde{U}_z^2 + \tilde{V}_z^2) - \tilde{\nu} g \beta_T\, T_z/\tilde{\mathrm{P}}\mathrm{r}$, g is the gravity acceleration, β_T is the coefficient of thermal expansion of water, $T(z)$ is temperature, $\tilde{\mathrm{P}}\mathrm{r}$ is the Prandtl number in water; the values of the constants $\tilde{c}_\nu$, $\tilde{c}_1$, $\tilde{c}_2$, and $\tilde{\sigma}$ are borrowed from Ref. 4.

Let us consider now boundary and initial conditions. The vertical structure of the area is determined as follows: H is the upper boundary of the ABL, h is the thickness of the near-water air layer for which the assumption of constant turbulent fluxes is valid, $z = 0$ is the air/water interface, and $z = \tilde{H}$ is the bottom boundary of the area in the water body.

The following conditions are stated for Eqs. (1)–(4):

$$U = U_\mathrm{G}, \;\; V = V_\mathrm{G}, \;\; b = 0, \;\; \frac{\partial \varepsilon}{\partial z} = 0 \;\; \text{at } z = H; \quad (5)$$

$$\nu \frac{\partial U}{\partial z} = c_u |\mathbf{U}| (U - \tilde{U}_0), \; \nu \frac{\partial V}{\partial z} = c_u |\mathbf{U}| (V - \tilde{V}_0),$$

$$\frac{\partial b}{\partial z} = 0, \quad \varepsilon = \frac{c_\nu\, b^2}{\nu_h} \quad \text{at} \quad z = h; \quad (6)$$

$$\tilde{\rho}_0\, \tilde{\nu} \frac{\partial \tilde{U}}{\partial z} = \rho_0\, \nu \frac{\partial U}{\partial z}, \quad \tilde{\rho}_0\, \tilde{\nu} \frac{\partial \tilde{V}}{\partial z} = \rho_0\, \nu \frac{\partial V}{\partial z},$$

$$\tilde{\rho}_0\, \tilde{b} = \rho_0\, b, \;\; \tilde{\rho}_0\, \tilde{\varepsilon} = \rho_0\, \varepsilon \;\; \text{at } z = 0; \quad (7)$$

$$\tilde{U} = \tilde{V} = 0, \;\; \tilde{b} = 0, \;\; \frac{\partial \tilde{\varepsilon}}{\partial z} = 0 \;\; \text{at } z = \tilde{H}, \quad (8)$$

where c_u is the resistance coefficient, c_Θ is the coefficient of heat exchange of air mass with water, ν_h is the value of ν generated in the near-water layer at $z = h$, $\tilde{U}_0$ and $\tilde{V}_0$ are the components of the drift velocity on the water surface, and ρ_0 and $\tilde{\rho}_0$ are the air and water densities, respectively.

The stationary fields U, V, b, ε, $\tilde{U}$, $\tilde{V}$, $\tilde{b}$, and $\tilde{\varepsilon}$, obtained by integrating Eqs. (1)–(8) were taken as the initial conditions.

Models of transport of aerohydrosol

Describing transport of an admixture, we take into account advection in the land-to-water direction, since due to it the aerosol is transported to the water area. Assuming homogeneity of the processes along the shoreline, we direct the x-axis horizontally along the normal to the shore. The equation of turbulent diffusion in the ABL is written in the form:

$$\frac{\partial S}{\partial t} + U \frac{\partial S}{\partial x} + (W - W_\mathrm{g}) \frac{\partial S}{\partial z} = \frac{\partial}{\partial z} \nu_s \frac{\partial S}{\partial z}, \quad (9)$$

where S is the admixture concentration, W_g is the gravitational rate of sedimentation, $\nu_s = \nu/\mathrm{Pr}_s$, Pr_s is the diffusion Prandtl number.

The hydrosol concentration is denoted C, and thus for the water medium we have:

$$\frac{\partial C}{\partial t} + \tilde{U} \frac{\partial C}{\partial x} - \tilde{W}_\mathrm{g} \frac{\partial C}{\partial z} = \frac{\partial}{\partial z} \tilde{\nu}_c \frac{\partial C}{\partial z}, \quad (10)$$

where $\tilde{\nu}_c = \tilde{\nu}/\mathrm{Pr}_c$.

Let us solve equations (9) and (10) in the area $x > 0$ (above the water area) under the following boundary conditions:

$$S = 0 \quad \text{at} \quad z = H; \quad (11)$$

$$\nu_s \frac{\partial S}{\partial z} = 0, \; \tilde{\nu}_c \frac{\partial C}{\partial z} + \tilde{W}_\mathrm{g}\, C = W_\mathrm{g}\, S \text{ at } z = 0; \quad (12)$$

$$\frac{\partial C}{\partial t} = \tilde{W}_\mathrm{g} \frac{\partial C}{\partial z} \;\; \text{at } z = \tilde{H}. \quad (13)$$

It should be noted that Eqs. (12) provide continuity of the admixture flow at transition through the interface.

The boundary conditions along the horizontal are stated at the water edge:

$$S = S_\mathrm{b}(z, t), \;\; C = 0 \;\; \text{at } x = 0, \quad (14)$$

where S_b is the aerosol concentration in the area $x \le 0$ (above the land). Let us use the 1D (along z) analog of Eq. (9) for calculating S_b, assuming $\partial S/\partial x = 0$. Formulating the lower boundary condition, we assume that the underlying surface is horizontally homogeneous and covered by dry monodisperse sand. The concentration on the surface is determined by the equation

$$\nu_s \frac{\partial S_\mathrm{b}}{\partial z} + W_\mathrm{g}\, S_\mathrm{b} = \beta S_\mathrm{b} - \Gamma \;\; \text{at } z = 0, \quad (15)$$

where Γ is the mass of saltatory particles blown out of the surface, β characterizes the rate of their entrapment into turbulent diffusion. The source strength Γ was calculated in accordance with the technique proposed in Ref. 5, which studies the saltation and diffusion mechanisms in the wind-sand flow near the surface. According to Ref. 5, deflation arises as the speed of airflow exceeds some threshold value, at which the Froude number $\mathrm{Fr} = \rho_0\, u_*^2/\rho_d\, gd$ for aerosol particles is equal to 10^{-2}. Here d and ρ_d are the diameter and density of particles, respectively; u_* is the dynamic velocity.

At the initial moment the fields S and C are assumed zero.

Analysis and conclusions

The problem (1)–(15) was solved numerically by quantization of the equations and boundary conditions in space and time using the conservative procedure for description of the fields of velocity and turbulence and the monotonic procedure with conservation of the first moments for the admixture transport equations. As a result of quantization in x, the 2D equations (9) and (10) are reduced to the set of 1D problems in terms of z, which can be integrated consecutively downstream the flow from the point corresponding to the shoreline. The 1D non-stationary problem presented by Eqs. (9), (11), (14), and (15) is solved at this point. For convenience of numerical realization, Eq. (15) was replaced with its exact analog which allows the boundary condition to be displaced from the surface $z = 0$ to the first computational level $z = h$.

The values of W_g and $\tilde{W}_g$ were determined by the Stokes equation; it was taken $\beta = W_g$ in Eq. (15).

Formation of the fields of velocity and turbulence in the ABL and water body depends on the value of the geostrophic velocity as well as stratification of the both natural media. Let us set the stable thermal conditions with the standard value Θ_z in the atmosphere and $T_z = 0.1°C/m$ in the water body. Assume that $U_G = 10$ m/s, and the size of soil particles $d = 3$ μm. This value of the geostrophic velocity gives rise to development of turbulence with $\nu \approx 6$ m/s^2 in the ABL, and the drift speed is about 5 m/s. Analysis of the Froude number shows that its value for the obtained dynamic parameters is close to the critical one, but not exceeding it, i.e., admixture flow from the surface is absent.

Further integration was performed with smooth increase of the flow speed up to $U_G = 15$ m/s. The increase of the speed leads to intensification of turbulent pulsations and development of the processes of saltation and turbulent diffusion of particles into the atmosphere. Finally, the admixture transport to the water area due to advection is formed. Coming onto the surface, the admixture is absorbed, and then it migrates in the water medium under the effect of turbulent mixing, drift stream, and its own gravitational sedimentation.

Figure 1 shows isolines of the concentration fields S and C (in g/m^3) in the area $0 < x < 22$ km. The fields are obtained at the moment $t = 18$ h from the beginning of saltation. It is seen that the thickness of the horizontal dust stream does not exceed 200 m near the source and decreases with the distance from the shore. The admixture concentration in water is two orders of magnitude higher than that in air because of the marked difference between the rates of gravitational sedimentation.

Formation of the maximum in the field of C (lower part of the figure) is caused by the following circumstance. As the flow speed in the ABL increases in the initial period, dynamic characteristics in the near-ground layer are transformed. As this takes place, the source strength Γ increases from 3.7 to 5.5 g/m^2 s. Hence, the admixture flow from the surface increases with time until the wind increase stops ($t \approx 8$ hours). At this time the aerosol mass in the ABL is maximum. Consequently, sediment is maximum too. As the turbulent conditions adapt to the new values of U_G, the admixture flow from the surface gradually decreases, approaching the new stationary value.

Displacement of the maximum along the horizontal from the shore (by about 3 km) can be explained by the joint effect of atmospheric advection and drift in water.

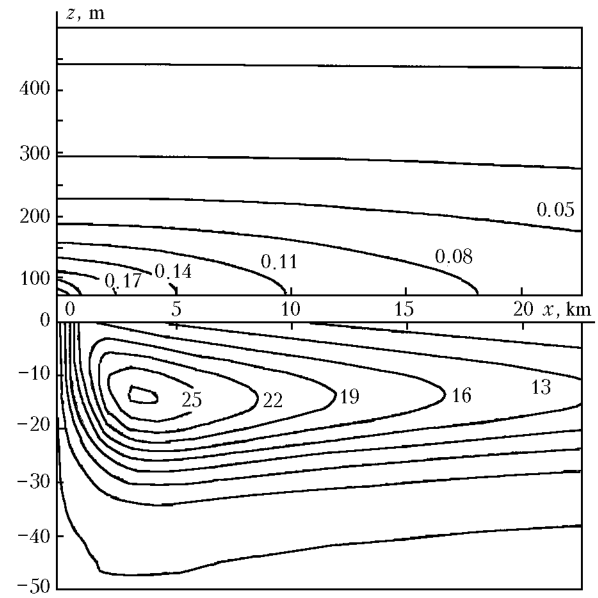

Fig. 1. Field of admixture concentration (in g/m^3) in the atmosphere ($z > 0$) and water body ($z < 0$) at $t = 18$ hours.

Computation by the set of scenarios on the effect of temperature profiles on the peculiarities of the admixture spread has shown that stratification of the ABL is a significant parameter. Thus, turbulent exchange intensifies at the neutral stratification of the atmosphere, and the admixture flow rises relatively high. This leads to the decrease of the concentration in the near-water layer. As a result, the admixture inflow into water is almost halved, and the aforementioned extreme of the field of C is not formed. The effect of density stratification of the water body proves to be weaker due to its passive role of a recipient and dominating mechanism of gravitational sedimentation.

The large-scale vertical motion can significantly change the spatial distribution of the sediment. For example, if the linearly increasing profile of W is prescribed up to the value of 3 cm/s at $z = 1000$ m (let us note that $W_g = 7$ cm/s for this version of computation) characteristic of cyclonic vortices, then the peak of C decreases four times due to transport of a portion of the admixture to the upper layers and due to the decrease of the concentration near the water surface. Under anticyclone conditions, the horizontal path of particles decreases ($W < 0$), and the range of the source contracts to the boundary $x = 0$.

Thus, the proposed model of the turbulent interaction between the atmosphere and the water body gives the qualitatively correct pattern of admixture redistribution in each of these natural media. This gives reason to assert that the applied approach is realistic and it is necessary to optimize the model in the aspect of comparison with experimental data. The further steps are, first of all, generalizing the statement of the problem for description of the thermal conditions of the atmosphere and water body and considering thoroughly the turbulent interaction taking into account wind roughness.

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