LIDAR STUDIES OF FLUCTUATIONS OF AEROSOL CONCENTRATION IN THE GROUND ATMOSPHERIC LAYER

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The analysis of lidar and acoustic anemometers data on atmospheric fluctuations allowed us to derive the initial approximation of a versatile function for variance of fluctuations of the backscattering coefficient in the ground atmospheric layer on Monin–Obukhov theory of similarity.

The Monin–Obukhov theory of similarity for turbulent regime in a stratified medium is the base for physical description of the state of the ground atmospheric layer. The basis of this theory is hypothesis on independence of the turbulent regime of medium motion on the molecular air constants and the absence of the effect of underlying surface. The theory allows one to construct scales for measuring wind velocity u^*, temperature T^* and length L (the Monin–Obukhov scale) and then, passing to the non–dimensional variables, to obtain the versatile profiles of principal meteorological parameters and the versatile functions of the statistical characteristics in the ground air layer.[1]

Analogously the versatile functions can be defined for aerosol particles concentration, if the particles are considered as a passive conservative admixture. The truth of the latter is theoretically proved in Refs. 2 and 3 and experimentally substantiated by contact methods[4,5] and optical measurements.[3,6]

In this paper we take an attempt on semiquantitative definition of the versatile function that connects the variance of aerosol concentration fluctuations with the parameter characterizing the hydrodynamic situation of the atmosphere. Usually the Richardson number Ri is used as such a parameter, because it uniquely describes the thermal stratification[1]

$$\mathrm{Ri} = \frac{g}{T}\left(\frac{\partial T}{\partial z} + \gamma_a\right) \Big/ \left(\frac{\partial u}{\partial z}\right)^2,$$

where $\gamma_a = 0.0098$°C/m is the dry adiabatic lapse, g is the acceleration due to gravity, and z is the height. The number Ri determines the relative contribution of convective and dynamic factors to the turbulence energy and can be easily determined in practice by Eq. (5) from the readings of two pairs of sensors of temperature and wind velocity placed at different levels within the ground atmospheric layer. The number Ri is equal to 0 for indifferent (neutral) stratification when temperature lapse taken absolutely is equal to γ_a.

Let us consider the backscattering coefficient as an optical parameter that characterizes the fluctuations of concentration. According to the theory of similarity, one may write for the lapse of the backscattering coefficient $\beta(z)$

$$\frac{\partial\beta}{\partial z} = \frac{\beta^*}{L^*} f_\beta(\zeta), \tag{1}$$

where β^* is the measurement scale of the backscattering coefficient, $f_\beta(\zeta)$ is some versatile function, and $\zeta = z/L$ is the dimensionless height. According to the theory, the scale β^* should be determined by the turbulent flux of particles F_β through the unit horizontal plate. Then on the basis of the relation of dimensions one may write

$$\beta^* = -\frac{F_\beta}{\mathrm{k}\rho u}, \tag{2}$$

where κ is the versatile Karman constant and ρ is the atmospheric density. Let us note that β^* serves as a scale only in the case when the flux F_β is constant with height. It generally is the definition of the ground layer for the field of concentration of atmospheric particles. According to Ref. 7, the ground layer height h is 20–200 m for the momentum flux and, therefore, for the wind velocity field, depending on meteorological conditions. The analogous magnitude estimates are obtained for the fluxes of temperature and moisture.[7] It allows us to suppose that the height of the ground layer for concentration of particles should not be essentially different from the noted values.

As is seen from Eq. (2), to find β^* it is essential to know the turbulent flux F. However, direct experimental measuring the turbulent fluxes of heat, moisture or particles is quite complicated technical problem.[8] So in practice the proper scales are estimated on the base of lapse measurements. Particularly, one can estimate the flux of aerosol particles using the following relationship:

$$F_\beta = K_\beta \frac{\partial\beta}{\partial z}, \tag{3}$$

where K_β is the coefficient of turbulent exchange for the particles concentration. Combining Eqs. (2) and (3) we obtain the expression for the scale in general form

$$\beta^* = -\frac{K_\beta}{\mathrm{k}\rho u^*}\frac{\partial\beta}{\partial z}. \tag{4}$$

It follows from this that to estimate β^* it is essential to have the information on behavior of the turbulent exchange coefficients and wind velocity first of all. Since the coefficient K_β depends on the concrete meteorological conditions, its value is not constant.[1] This circumstance does not allow one to estimate the scale β^* directly, when it is not possible to determine K_β. Nevertheless, if some assumptions realize, it is possible to obtain the versatile functions for the aerosol particles concentration.

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Integrating Eq. (1) for the stratification being close to indifferent, one can obtain the formulas called in publications as "linear+logarithmic" model of the ground atmospheric layer[1,7]

$$\beta(z) = \beta(z_0) + \beta^*[\ln z/z_0 + \alpha_\beta z/L], \quad (5)$$

where the coefficient α is essential for correction of the profile of passive admixture for the stratification different from neutral one, Ri = 0, since the dimensionless height $\zeta = z/L$ is the classical parameter of hydrostatic stability that is uniquely connected with the number Ri. For neutral stratification when Ri = 0, the Monin–Obukhov scale is $L = \pm\infty$, and, therefore, the parameter ζ is also equal to zero. The logarithmic law for the profiles of meteorological parameters in the ground layer follows from the theory of similarity. The relationships analogous to Eq. (5) but different in coefficient α are true for the profiles of wind velocity, temperature, humidity, and other atmospheric admixtures.

However, in contrast to the noted meteorological elements, we have not succeeded in finding the data on the coefficient α for aerosol profiles in the literature.

In this connection, to obtain the end formulas let us consider only the neutral stratification (or close to that), i.e., when Ri = 0. In this case Eq. (5) can be simplified

$$\beta(z) - \beta(z_0) = \beta^* \ln z/z_0 ,$$

and one can write for the dimensionless variance of fluctuations

$$f_\beta = \sqrt{\sigma_\beta^2}/\beta^* = (\sqrt{\sigma_\beta^2}/<\beta>) <\beta> \ln z/z_0/[\beta(z) - \beta(z_0)].$$

Fixing the point within the ground atmospheric layer ($z = z_1 > z_0$) and assuming $<\beta> = [\beta(z_1) + \beta(z_0)]/2$ we obtain

$$f_\beta = 0.5(\sqrt{\sigma_\beta^2}/<\beta>)\,[(B + 1)/(B - 1)]\ln z_1/z_0 , \quad (6)$$

where $\beta = \beta(z_1)/\beta(z_0)$.

As is seen from Eq. (6), the expression obtained for the versatile function makes it possible to work with optical parameters presented in the relative dimensionless form. In its turn, this makes it possible to use in data processing the values of lidar signals directly without absolute calibration of the lidar. These signals are proportional to the values of backscattering coefficient, if the optical depth is small.

To study the features of fluctuations of aerosol particles concentration we performed the experimental investigations by means of a LOZA–3 laser locator[9] and two acoustic anemometers[10] during summer and fall seasons.

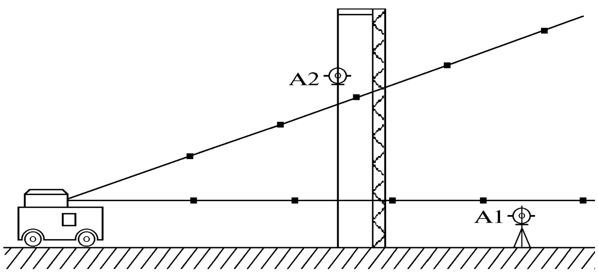

FIG. 1. Diagram of the performance of the experiment using a lidar and acoustic anemometers.

Measurements were carried out near Tomsk above the plane underlying surface covered with grass according to the experiment diagram given in Fig. 1. The first anemometer A1 was placed at a height of 2.5 m, and the second one A2 was placed at a meteorological mast 20 m in height. Both anemometers measured two horizontal components of wind velocity with frequency of 4 Hz. The mean value of wind speed and the mean wind direction were determined from these data as well as the variance of wind velocity fluctuations and the Richardson number Ri within the atmospheric layer of 20 m.

Lidar measurements were carried out along horizontal direction and at an angle of 3.2°. The sounding paths and anemometers were spaced at interval of no more than 200 m. The frequency of recording lidar returns was 2 Hz, and the spatial resolution along the path was 7.5 m. Measurements were carried out simultaneously, the cycle involved two sets.

In the first set the profiles of lidar returns in sounding along both directions were being written to computer memory over 2 minutes. The ratio between these signals was allowing us to estimate vertical distribution of $\beta(z)/\beta(z_0)$, where $z_0 = 2.5$ m. In the second set the sounding was being performed at an angle of 3.2° over 10 minutes at 1–s intervals, and the information on the fluctuations of concentration of aerosol particles was being accumulated. As the sounding was being stopped, we were calculating the normalized rms deviation $\delta\beta = \sqrt{\sigma_\beta^2}/<\beta>$ of the backscattering coefficient in the height range from 2.5 up to 36 m.

To span the greatest amount of the values Ri and profiles $\beta(z)$ occuring in the atmosphere, we carried out the observations at different times of the day. The most typical behavior of vertical distribution of the ratio $\beta(z)/\beta(z_0)$ depending on the value Ri is shown in Fig. 2.

The situation when aerosol particles concentration increases with height (curve *1*) occurs, as a rule, in the morning, when there is a ground temperature inversion (Ri > 0) blocking up the aerosol under the inversion. Then during daytime the Earth's surface is well warmed up (Ri < 0), inversion fails, and aerosol is brought upward by the convective fluxes. As a result, the aerosol particles concentration decreases with height (curve *3*) and the aerosol generation occurs mainly near the surface. In the case of indifferent stratification (Ri ≃ 0) the layer near the ground is, as a rule, well mixed, and the profile of the backscattering coefficient is practically constant within all the layer (curve *2*). Thus, over one day the vertical distributions of backscattering coefficient of contrary behavior, that have different values of lapse can be observed. This lapse value as well as the values of pulsations of transport velocity, principally determine the fluctuations of the aerosol particles concentration in the atmospheric layer near the ground.

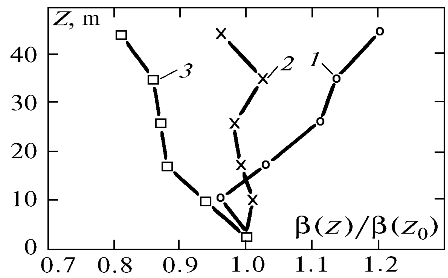

FIG. 2. Distribution of the normalized typical profiles of the backscattering coefficient at following values of the Richardson number: 1.3 *(1),* 0.01 *(2), and* – 0.53 *(3).*

The results of experimental investigations of interrelations between these physical parameters are shown in Figs. 3–5. Figure 3 presents the comparison of the data on the absolute value of the fluctuation component of wind velocity obtained with an anemometer and the variations δβ of the backscattering coefficient.

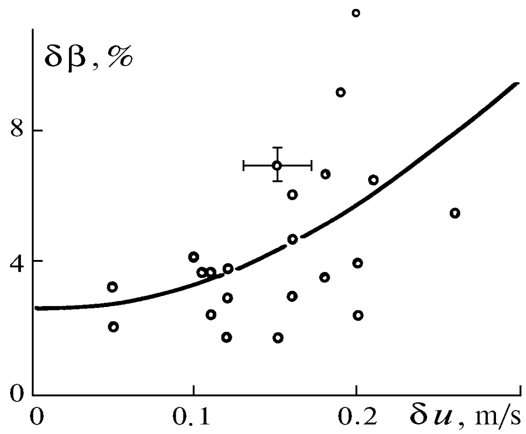

FIG. 3. Variations of the backscattering coefficient as function of the fluctuations of wind speed.

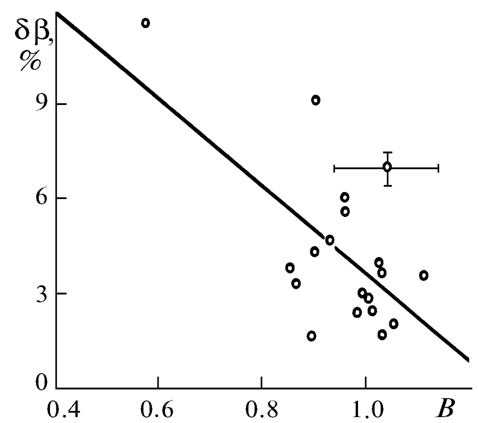

FIG. 4. Variations of the backscattering coefficient as function of its rate.

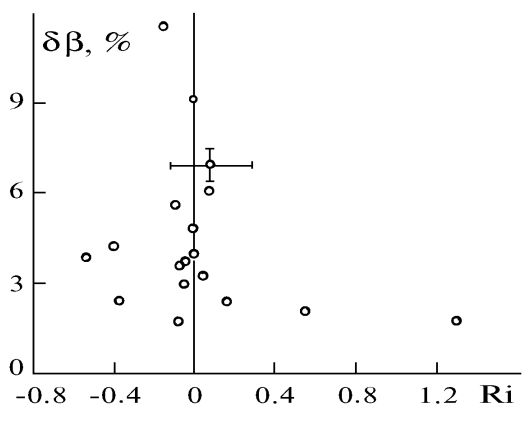

FIG. 5. Variations of the backscattering coefficient as function of the Richardson number Ri.

The scatter of points on the plot makes it possible to reveal, in principal, the tendency for an increase of δβ as the fluctuations δ increase, the form of this dependence is parabolic. The parabola ($\delta\beta = 76.3\delta u^2 + 2.6$) in the figure represents the regression curve calculated by the least squares method. Since the scatter of points essentially exceeds the error in measuring shown in the figure, it can not be explained only by the turbulent regime. As it has been already noted, the second essential moment is the effect of the lapse of aerosol particles concentration. One can judge about it based on the ratio $B = \beta(z_1)/\beta(z_0)$. These data are shown in Fig. 4. The regression here is also calculated by the least square method for the linear function $\delta\beta = 13.1B + 17.1$. It is seen from the figure that the faster is decrease of the particles concentration with height (Ri < 0, Fig. 2), the higher are the fluctuations δβ.

General analysis of discussed experimental material indicates that there is no a universal dependence between these physical parameters of atmosphere. Direct analysis of interrelation between the backscattering coefficient fluctuations and the number Ri shown in Fig. 5 leads us to the same conclusion. The general tendency of this dependence is increasing δβ as the nonstability of the atmosphere increases (Ri < 0). The peculiarity of the data presented is the sharp increase of δβ near Ri | 0, i.e in the case of the indifferent atmospheric stratification.

Let us consider now the behavior of fluctuations δβ when normalizing to the scale β*. The versatile function was calculated by Eq. (6) using all the data obtained. Calculational results given in Fig. 6 show that one can approximately describe the behavior of the dependence $\delta\beta^* = f(\mathrm{Ri})$ by hyperbolic function $\delta\beta^* = 0.027/\mathrm{Ri}$, selecting its asymptotes as the coordinate axes. As is seen from the figure, one portion of the experimental points is placed on other side of the ordinate axis that is most probably explained by great error in measuring the number Ri. Nevertheless, the analysis of the plot as a whole makes it possible to judge about more regular placement of the experimental points in comparison with all dependences discussed above, including the dependence given in Fig. 5.

Another characteristic peculiarity of the considered versatile function is the discontinuity point at Ri = 0. Formally (according to calculational formulas) the break of the function means sharp vanishing the scale β*. In fact it is corresponds to the decrease of values of turbulent flux of particles to minimum ones, with change of sign of the particles concentration lapse and the motion direction of turbulent flux in passing through the point Ri = 0. This peculiarity makes it possible to judge about temperature stratification of the atmosphere quite simply on the base of the lidar data solely.

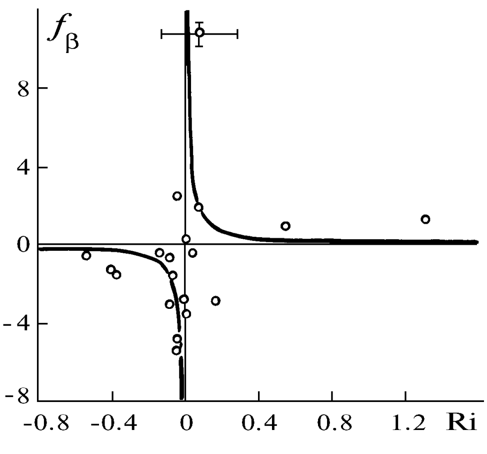

FIG. 6. The shape of the versatile function of the variance of the backscattering coefficient fluctuations.

Let us note for conclusion that the attempt at obtaining the versatile function, that connects the variance of the aerosol concentration fluctuations with temperature stratification of the atmosphere, is only the indicative of the main direction of such a kind of investigations. At a later time authors plan to return to the problem under consideration on the base of the more extended experimental data.

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