LIDAR STUDIES OF FLUCTUATION SPECTRA OF THE COEFFICIENT OF BACKSCATTERING IN THE ATMOSPHERIC SURFACE LAYER

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Results are presented from experimental lidar studies of fluctuation spectra of the coefficient of backscattering in the atmospheric surface layer. Significant differences are found between the fluctuation spectra of scattering and of wind speed, presumably explained away as the effects of temperature stratification and of convective turbulence. When stratification of the surface layer is stable there appears a "sub-area of buoyancy" (the Boljiano-Monin spectrum) in fluctuation spectra of the coefficient of backscattering.

Fluctuations of laser echo–signals from aerosol inhomogeneities in the atmosphere are of interest both for studies of the origin and properties of those inhomogeneities themselves, and for various applications, in particular for developing laser meters of wind speed.[1]

Entrainment of particulate matter by air flows[2] and the presence of vertical gradient of particle number density in the atmosphere[3] both make it possible to envisage aerosol as a passive conservative admixture in the atmosphere. References. 1 and 4 cite study results testifying to applicability of the hypothesis of a passive conservative admixture and of respective model power spectrum to statistical description of fluctuations of atmospheric aerosol number density. It follows from applicability of that hypothesis and of the Kolmogorov model spectrum to statistical description of the atmospheric aerosol number density[4] that a power range, $f^{-5/3}$, should be present in the frequency spectrum of fluctuations of scattered radiative power.

In addition we present simple qualitative considerations to explain the process of entrainment of an aerosol particle by eddy flow. Atmospheric aerosol matter, mainly consisting of particles in the micron range and slowly sedimenting gravitationally, features a short relaxation time, which may be estimated to within its order of magnitude, as a ratio of sedimentation velocity to acceleration of free fall: $\tau_p \approx 10^{-4}$ (m/s)/9.81 (m/s^2) $\approx 10^{-5}$ s (see Ref. 5). The time interval a particle takes to fall down through an eddy may be approximately estimated dividing the eddy cross–section by the falling velocity of the particle: $\tau_{np} \approx 1$ m/10^{-4} (m/s) $\approx 10^{-4}$ s. A particle will be entrained by an eddy completely or almost completely in case $\tau_r \ll \tau_{ed} \ll \tau_{in}$, where τ_b is the lifetime of an eddy, given, to its order of magnitude, as the ratio of the size of that eddy to pulsation of wind speed within it, $\tau_{ed} \approx 1$ m/10^{-1} (m/s) ≈ 10 s. The relative volume occupied by particulate matter in a highly transparent atmosphere is comparatively small. The particle size is also small when compared to the sizes of eddies in the flow. These considerations make it possible to regard aerosol as passive, not affecting the dynamics of the flow itself.

According to the above qualitative considerations aerosol is a passive conservative admixture completely entrained by air eddies. A serious argument in favor of that statement is also the fact that the spectrum of fluctuations of backscattering is similar to that of wind speed. However field measurements indicate certain differences between the two spectra, which may be due to a whole number of reasons calling for dedicated theoretical treatment and experimental studies.

The present article cites results from lidar observations in the atmospheric surface layer. The study aims at estimating limits of applicability of the Kolmogorov model spectrum to fluctuation spectra of the power of backscattering. Also some systematic are presented on the fluctuation spectra retrieved from lidar signals, which also present the spectra of fluctuations of the coefficient of backscattering, β_π. In its turn that coefficient depends on the particle number density in the sounded volume.[4] In what follows we speak about the "backscattering spectrum" for brevity reasons, understanding it as the spectrum of the power of fluctuations of the coefficient of backscattering at a given elevation above ground.

Classical micrometeorology treats the surface layer as such a layer several tens of meters thick in which the vertical flux of pulse and the vertical eddy fluxes of passive admixtures, such as those of heat, moisture, particulate matter, may be considered constant with height.

The scheme of experiment is shown in Fig. 1. The laser aerosol sensor used to sound the atmosphere was described elsewhere.[6] To sound the atmosphere short laser pulses were emitted at sending frequency of 12.5 Hz, their beampath aimed at 3.2° to the horizon. The lidar spatial resolution was 10 m. The observed lidar echo–signal was only recorded for the 16 strobes covering heights from 10 to 30 m, and the statistics on fluctuations of the backscattering coefficient was accumulated for 30 min to one hour. To eliminate interference due to fluctuations of individual pulses energy each set of 6 sequential echo–signals was summated. Thus the temporal lapse in a set reached 0.48 s (≈ 2 Hz).

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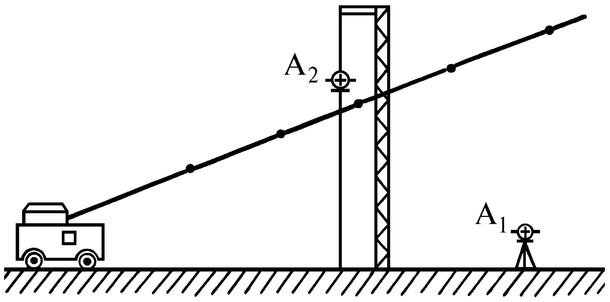

FIG. 1. Experiment scheme. Anemometers A1 and A2 are positioned 2 *and* 20 m *above ground.*

Data on fluctuations of absolute wind speed were obtained and recorded by an acoustic anemometer,[7] placed 20 m above ground on a weather mast, that is at lidar mid–range. Anemometer recording frequency was 4 Hz. Lidar echo–signals, proportional to β_π, and anemometer wind speed, v, were recorded in parallel.

After the field phase the accumulated temporal sets were statistically processed by way of high frequency polynomial filtering, so as to exclude trends. Applying the spectral window[8] reduced the effect of "leaking", and the following fast Fourier transform made it possible to retrieve the power spectrum. The number of the degrees of freedom, n_d in spectral estimates was increased averaging over frequencies and additionally averaging lidar data over the ensemble of spectra from different distances (16 strobes).

Observations were taken at the "Zarechny" field site of the Institute of Atmospheric Optics during summers and autumns of 1987 and 1988. The site was a potato field of approximately 2 km^2.

Measurements yielded about 100 spectral profiles. Analysing the total set of data resulted in identifying five basic types of fluctuation spectra for the backscattering coefficient, β_π. Note that all the spectra were dispersion normalized.

Typically, all the wind speed fluctuation spectra feature a power range of an incline of $f^{-5/3}$. Accounting for wind speed, v, temporal frequencies covered the range from 0.005 to 1 Hz, corresponding to spatial wavelengths, $\lambda = v/f$, from approximately 500 to 1 m. The top limit in our case is given by the length of the sensing pulse, hence by the irradiated atmospheric volume. The lower limit results from the overall length of the set. Its extension beyond the chosen length is unreasonable, since the probability to obtain sets of non–stationary processes generated by the duirnal trends of meteorologic parameters becomes high then. Another reason to limit processing to 30 minute averaging is the shape itself of the spectra of fluctuations of meteorologic parameters (such as wind speed, temperature, pressure), which have deep mesoclimatic minima around the period of 1 hour, those minima separating the ranges of small–scale turbulence from large–scale synoptic oscillations.[9]

The studied interval of frequencies included the classical inertia range. The rate of energy transfer through the inertia range is constant, equal to the rate of dissipation of energy of eddies into heat.[10] All the joint observations yielded an incline in the spectra of wind speed fluctuations coinciding with the Kolmogorov–Obukhov "–5/3" theoretical incline to the accuracy of 20%. The only exclusion was type in Fig. 2, both spectra there having a wavy shape, apparently due to intermittent turbulence. However, even then the average spectral incline was equal to "–5/3".

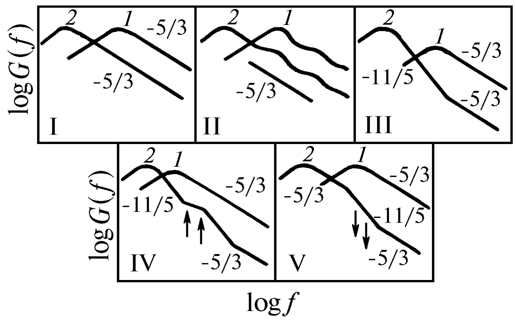

FIG. 2. Schematic presentation of the obtained spectra of fluctuations of the backscattering coefficient. Curves 1: wind speed fluctuation spectrum; curves 2: β_π *fluctuation spectrum.*

To give an illustration, Fig. 3 shows spectra of wind speed fluctuations normalized to dispersion, $G_v(f)/\sigma_v^2$ (dots) and of $\beta_\pi - G_{\beta_\pi}(f)/\sigma_{\beta_\pi}^2$, normalized as well (crosses). A typical feature of all the pairs of spectra obtained is that the maximum in the spectrum of β_π is always to the left of the maximum in the spectrum of wind speed, the latter position controlled by the external scale of turbulence, L_0, which is close to the scale of the largest eddies in its order of magnitude. A certain scale, β_π, corresponds to the position of the maximum in the L_a spectrum, which we may call the "external scale of aerosol inhomogeneities."

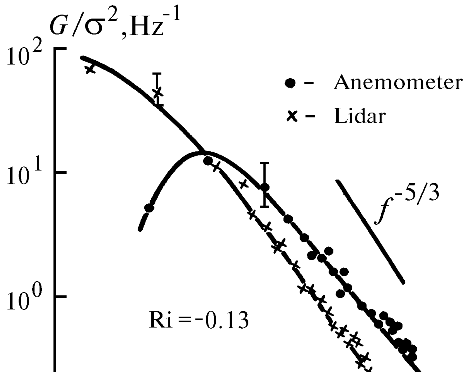

*FIG. 3. Normalized fluctuation spectra of wind speed (dots) and of the coefficient of backscattering (crosses) in case of weak instability (*Ri = – 0.16*).*

The fact that $L_0 < L_a$ indicates presence of slow oscillations in the spectrum of fluctuations of number density, which are either absent or too weak in the spectra of wind speed. The question then arises on the possible nature of those slow oscillations in the β_π spectra. It is, first, related to local orography, that is to possible influence of the underlying surface. Indeed, the temporal and the corresponding spatial periods of averaging considerably exceed the size of the field over which observations were taken. One may, therefore, assume that the air mass cloud, before approaching the site, pass over strips of land, radiatively patchy enough. In combination with convective and dynamic turbulence that could result in the appearance of large aerosol inhomogeneities, equal to inhomogeneities in the flow in their size. Further on the parameters of the turbulent flow could "adapt" above the site itself, while large–scale aerosol inhomogeneities could still exist for a while. Second, that effect could stem from the possible influence of convective turbulence. Note that the vertical component of wind was not measured during the experiment.

Below we cite results from a different experiment that we believe interesting (Fig. 4). It went as following: a distributed "dust generator" was active during our measurements, when a long column of agricultural machines proceeded at a low speed along the field edge. The dust it raised into the air was entrained by eddies of a size not larger than the height of observation point, that size mainly controlled by the wind shear in the surface layer. Aerosol inhomogeneities of that size were of higher contrast, than they would have been were there no such dust. That is why respective spectra of β_π feature uplifts at corresponding frequencies. Wind speed at 20 m above ground was approximately 5 m/s. Three local maxima may be identified in the spectrum of β_π, Fig. 4. Reducing them according to $\lambda = v/f$ we find respective wavelengths of 20, 10, and 5 m. Most probably divisible frequencies result from some periodic process, e.g., trucks passing at intervals.

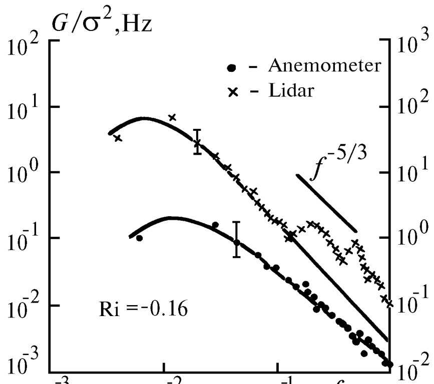

*FIG. 4. Spectra of fluctuations of wind speed (dots) and of the backscattering coefficient (crosses) for weak instability (*Ri = – 0.13*) in case of a distributed source of dust aerosol in the surface air layer.*

Three other types of spectra β_π, shown in Fig. 2, are all peculiar in having a power range with an incline noticeable in excess of –5/3. It is closer actually to –11/5, that spectrum described elsewhere and known as the Boljiano–Monin spectrum. The Boljiano–Monin spectrum was theoretically predicted for an atmosphere of stable stratification. Its steeper incline is explained by the loss of part of the turbulent energy from the flow due to work against the Archimedes forces, hampering vertical movements of parcels of air. The range of wavenumbers in which the shape of the spectrum is controlled by the thermal stratification, is called the "buoyancy sub–range". It includes those scales which enter the inertia range for an indifferent temperature stratification. However, experimental data from studies of fluctuations of wind speed and temperature by several authors did not support the existence of any buoyancy sub–range in the surface atmospheric layer. Estimates from Ref. 11 indicated that a buoyancy range cannot be found at heights below 100 m. Note that our observations also yielded a spectrum of fluctuations of wind speed close to the Kolmogorov one. The well–known study by N.Z. Pinus et al.[12] notes the appearance of the buoyancy sub–range in aircraft measured spectra of fluctuations of wind speed at 400 m or higher. The power, m, in the f^{-m} spectrum varied within 2–3.5.

S.J. Koffee[13] described the spectra of the horizontal components of wind speed, obtained from tethered ballons in stable air above the mixing layer (that is above the convective boundary layer), which had inclines close to 11/5.

During experiments the degree of temperature stability was estimated by the data from a pair of acoustic anemometers at 2 and 20 m, which could yield temperatures to the accuracy of 0.1°. The Richardson gradient number was found from the formula.

$$\mathrm{Ri} = \frac{g}{T}\left(\frac{\partial T}{\partial z} - g_a\right) \Big/ \left(\frac{\partial v}{\partial z}\right)^2,$$

where g is the free fall acceleration; τ is the absolute temperature; z is the height,; γ_a is the adiabatic temperature lapse rate.

Note that it was not always that the Ri number remained positive (thus corresponding to stable stratification) if the buoyancy sub–range appeared in the β_π spectra. Such a situation may be explained by the presence of stable air above the 20–30 m surface layer in which our measurements were taken.

To test that statement we staged experiments to retrieve the power m in the f^{-m} spectrum. During those we sounded the atmosphere up to 1 km in conditions of summertime elevated temperature inversion. The inversion level was lidar monitored as the maximum of the aerosol layer, usually observed during the early morning hours immediately below the inversion itself.[14] Figure 5 gives, by way of illustration, results from sounding the summertime inversion before sunrise. The profile shown is that of total scattering β_π, retrieved by the lidar. The inversion level was found around 300 m.

The same plot carries four values of power m for the respective spectra β_π. One finds $m > 11/5$ below the inversion, while above it that power equals 5/3, to within the experiment accuracy. When two such ranges, 11/5 and 5/3 could be identified in the β_π spectra, we retrieved the spatial size of inhomogeneities corresponding to the knee in the spectral curve. That size either coincided with or differed insignificantly from observation height, staying close to 20 m.

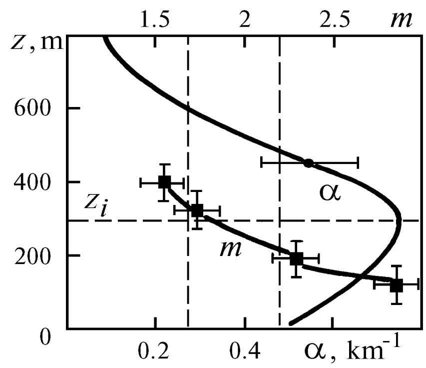

FIG. 5. Vertical profile of the coefficient of total scattering, α, (solid curve) and the profile of the power index, m, for the $G_{\beta_\pi}(f) \approx f^{-m}$ spectrum (dots) from lidar sounding of elevated temperature inversion.

The appearance of the buoyancy sub–range in the spectra of β_π alone is probably due to processes taking place above the surface air layer. Besides, we already mentioned the effect of convective turbulence. Moreover, aerosol inhomogeneities of the order of tens and hundreds meters may exist at heights comparable to their size. Naturally, the anisotropy of such aerosol formations is quite high.[15]

We now proceed to the two last types of spectra β_π, presented in Fig. 2. These seem to be quite complex. Arrows in Fig. 2 are used to show conventionally the ranges of energy "inflow" and "sink" in those spectra. Type 4 there is similar to type 3, although the presence of a "source" of eddy pulsations within the buoyancy sub–range resulted in uplifting the spectral curve at that "source" frequency. Ref. 10 commented on the possibility of existence of eddy flows having two spectral ranges or more, characterized by an influx of external energy into them. For example, the spectrum of oceanic eddies has three different ranges of energy influx.[10] In our case we may assume an influx of energy into the β_π spectrum due to dynamic turbulence produced by wind shear in the surface air layer. Fig. 6*a* shows the spectrum of wind speed and that of β_π presented in fully logarithmic coordinates. One clearly sees ranges in the β_π spectrum then, which correspond to the buoyancy sub–range ($m = 11/5$), the influx of energy to the spectrum, the extension of the buoyancy sub–range, and the inertia range ($m = 5/3$). Fig. 6*b* shows the same spectra in semi–logarithmic scale vs. the spatial frequencies $k = 2\pi/\lambda = 2\pi f/v$. Analising the fifth type of spectrum of β_π, shown in Fig.2, one may assume the existence of a thin stable layer within the surface air layer.

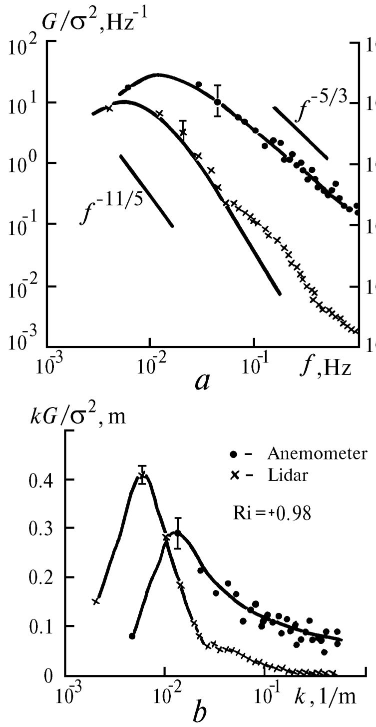

FIG. 6. Spectra fluctuations of wind speed (dots) and of the backscattering coefficient (crosses) for a stable stratification (Ri = + 1)*: a) full logarithmic scale; b) semi–logarithmic scale.*

Note two more results obtained in the course of analysing experimental data. We attempted to plot correlations of the external scale of aerosol inhomogeneities, L_a, vs. the external scale of turbulence, L_0, and the Richardson number, Ri. The scales were found from the half–width of the autocorrelation function, $\tau_{0.5}$, following the expression $L = 2\pi v \tau_{0.5}$ (see Ref. 16). The value of L_0 changed within 13–118 m across the whole range, while the same span for L_a was 38–590 m. Moreover, there was no mutual correlation found between L_0 and L_a. That fact indicates that the low frequency ranges in the spectra of β_π and of wind speed are formed by different factors, as commented above.

Another comment due is that the lidar makes it possible to record fluctuations of β_π at varying distances, so we used this opportunity to compute the coherence spectrum for β_π as a function of frequency f for variously distanced points in space, ρ = 10, 20, ..., 160 m (see Ref. 8). This test resulted in obtaining a positive correlation between the influx of energy to the spectrum of β_π (see Fig. 7*a*) and the uplift of the spectrum of coherence, particularly in the range of energy influx (see Fig. 7*b*). In other words, the lidar may be used as an instrument to study "fine" effects of the atmospheric turbulence, as long as it has certain advantages, which the traditional in situ instruments (such as anemometers, temperature sensors, etc) do not feature.

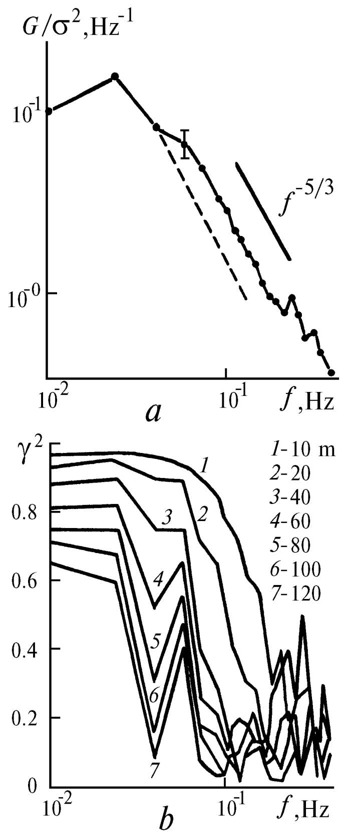

FIG. 7. Spectra of fluctuations of the backscattering coefficient, (a), and of respective coherence, (b) for various distances (Ri = + 1) *between the measurement volumes.*

Summing the above we may conclude that:

1) High–frequency part of the spectrum of fluctuations of the coefficient of backscattering, β_π, is controlled by the dynamic turbulence in the surface air layer, following of "five thirds", while its low–frequency range is possibly controlled by the convective eddies and the terrain relief.

2) Stable temperature stratification in the surface layer or immediately above it results in the appearance of the buoyancy sub–range in the spectra of β_π (of the Boljiano–Monin spectra).

3) A range of energy influx (sink) may appear in the spectra of fluctuations of the backscattering coefficient produced by the dynamic turbulence (by thermal stratification of the atmospheric surface layer into sub–layers).

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