SPECIFIC FEATURES OF THE STRATOSPHERIC AEROSOL CORRELATIONS

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This paper presents some results of the investigation of inter level correlation of the stratospheric aerosol scattering ratio as well as of the correlations between the scattering ratio and meteorological parameters of the atmosphere. Temporal series formed of lidar and radiosonde data compiled during the observational period from 1989 have been used in this study. Vertical profiles of the corresponding correlation coefficients are presented. Using these profiles certain peculiarities in the inter level interaction of stratospheric aerosol as well as of its interaction with meteorological parameters were revealed.

The stratospheric aerosol is an optically active component of the atmosphere and consequently one of the factors affecting weather formation. Therefore an extensive and detailed studies of its behavior are important and promising. The routine high altitude lidar observation of vertical stratification of the stratospheric aerosol scattering ratio performed during 1986–1989 above the Western Siberia[1] have revealed noticeable variability of this aerosol characteristic which is mainly stochastic but has certain elements of definiteness and randomness. The studies of the stratospheric aerosol variability from the stand–point of interrelations between its optical characteristics and meteorological parameters of the atmosphere enable one to fully understand its dynamics. The statistical analysis of the problem presented in this paper is based on studying behavior of the pair correlation coefficient K_{xy} which finally makes it possible to judge on the stratospheric aerosol correlations. Calculations of K_{xy} for two sequences X and Y of the length N are performed by the formula from Ref. 2

$$K_{xy}=\frac{\sum\limits_{i=1}^{N} X_i\cdot Y_i-\left[\sum\limits_{i=1}^{N} X_i\cdot\sum\limits_{i=1}^{N} Y_i\right]\Big/ N}{\sqrt{\sum\limits_{i=1}^{N} X_i^2-\dfrac{\left[\sum\limits_{i=1}^{N} X_i\right]^2}{N}}\cdot\sqrt{\sum\limits_{i=1}^{N} Y_i^2-\dfrac{\left[\sum\limits_{i=1}^{N} Y_i\right]^2}{N}}}. \quad (1)$$

The value of this coefficient determines the degree of deviation of the relation between X_i and Y_i from a linear one.

In this paper the scattering ratio R is used as the parameter describing vertical stratification of aerosol. This quantity is the ratio of the total backscattering coefficient $\beta_m^\pi(H) + \beta_a^\pi(H)$ to the molecular backscattering coefficient $\beta_m^\pi(H)$:

$$R(H) = \frac{\beta_m^\pi(H) + \beta_a^\pi(H)}{\beta_m^\pi(H)} = 1 + \frac{\beta_a^\pi(H)}{\beta_m^\pi(H)}, \quad (2)$$

where H is the height, is the aerosol backscattering coefficient.

INTERLEVEL CORRELATION OF THE SCATTERING RATIO

To study the interlevel correlation of the scattering ratio $R(H)$ the time series of the scattering ratio spatially averaged over the 0.3744 km altitude intervals are formed in one kilometer altitude step of the available experimental measurements and employed as sequences X_i and Y_i entering into Eq. (1). The calculational results for two principal seasons are shown in Fig. 1*a* (winter) and Fig. 1*b* (summer) , respectively , based on 64 and 35 night–averaged vertical profiles of the scattering ratio. The solid lines show the correlation coefficients (interlevel correlation coefficients) obtained at the same altitude with the relative error of calculations lower than 5%. These lines can be treated as an altitude function of the correlation coefficients of the scattering ratio with respect to its time behavior at certain altitude level. For example, $C_m(H)$ stands for the function of the correlation coefficients with respect to the level m ($m < H$, for $m = H$ and $C_m(H) = 1$). At the bottom of the figure a horizontal line denotes a tropopause level.

Since the vertical profile of the winter correlation function in the stratospheric region from 14 to 24 km approaches sufficiently close limiting values (0.05–0.2) achieved in a narrow range of altitudes 26–27 km there appears the difference in the decrease of $C_m(H)$ (the higher the level m (closer to 26 km) the larger is the decrease in $C_m(H)$). The functions $C_m(H)$ for the higher layers (above 24 km) do not approach any definite limiting value. This should be accounted for by limitedness of the given altitude range. Thus, as it follows from the functions $C_m(H)$ in the range of altitudes between 14 and 24 km, their fall off to the limiting value would occurs within the altitude interval of 7 to 10 km.

The altitude behaviors of the correlation coefficients presented in the lower portion of Fig. 1*a* (10–12 km) completely differ from $C_m(H)$ for $m > 14$. This part of the atmosphere is characterized by a sharp decrease of the correlation coefficient down to 0.2–0.3 and for $C_{10}(H)$ to –0.03 as low as at the first two adjacent levels. For more distant levels the correlation coefficient also decreases, but smoothly, to the values –0.2 to –0.1 at an altitude of

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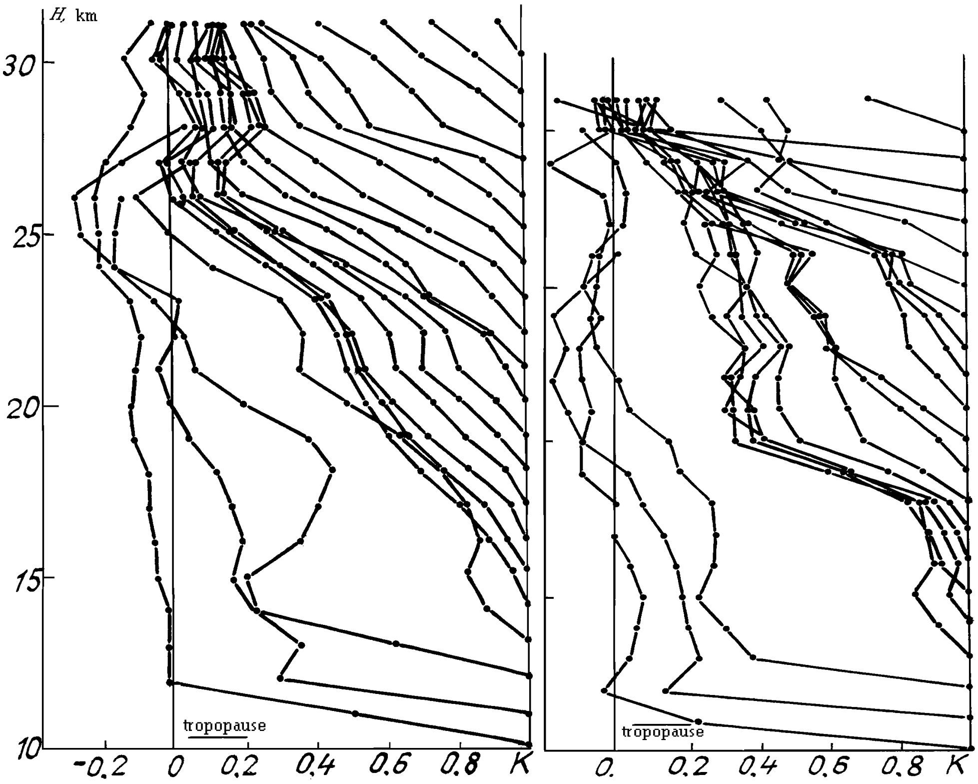

a b

FIG. 1.

from 25 to 26 km. The maxima of the functions $C_{11}(H)$ and $C_{12}(H)$ are at 16 and 18 km, respectively. Although the function $C_{13}(H)$ has the maximum also at 16 km and the minimum value –0.1 at 26 km, its vertical profile is on the whole the same as that of the function $C_m(H)$ for $m > 14$ km. By comparing the profiles of correlation coefficients at altitudes of from 11 to 13 km one can see a specific tendency in the vertical profile transformations of these functions which is probably related to the distance of the corresponding atmospheric layers from the tropopause and, finally, by their physical properties. Similar situation (different vertical profiles of $C_{10}(H)$ – $C_{12}(H)$ and $C_m(H)$ for $m > 12$) is also observed in summer. However, for the summer correlation functions $C_{10}(H)$ – $C_{12}(H)$ (Fig. 1*b*) the transformation to the functions of the overlying layers are not so pronounced. As a result they are separated out in the figure (to the left) in the form of a more isolated and dense group compared with the same functions for winter. Such a behavior of $C_{10}(H)$ – $C_{12}(H)$ allows one to assume that in summer the interaction is not so strong as in winter, and hence the air mass exchange between the troposphere and stratosphere is less intense.

Moreover, in Fig. 1*b* the correlation functions for altitudes of from 13 to 17 km with a form of an isolated group of profiles. They are characterized by a large correlation coefficient at the level > 0.8 which, in fact, is independent of the distance between layers. And by a sharp decrease down to 0.35 in the limited range of altitudes between 18 and 20 km, and a weak decrease down to 0.1 at altitudes of 20 to 29 km.

For the atmospheric layers above 17 km the identity of vertical profiles $C_m(H)$ occurs but partially and their successive transformation is observed.

The decrease in $C_m(H)$ at altitudes of 18 to 20 km for summer is more obvious than that in winter which is probably accounted for by a velopause (section of the stratosphere where the wind changes its direction at midlatitudes in summer). As can be seen from the analysis of aerosol properties, this range of altitudes is characterized by localization of the scattering ratio maximum in the Younge aerosol layer located in the lower stratosphere. As follows from Fig. 1*b* the lower part of the layer (13–18 km) is described by a strong interlevel interaction which is independent of distance between the layers. This allows one to consider it as a unique stratospheric aerosol formation with its own specific properties as compared with the overlying and underlying atmospheric layers.

CORRELATIONS BETWEEN THE SCATTERING RATIO AND METEOROLOGICAL PARAMETERS

The study of such correlations in stratospheric aerosol is based on the same time series of the scattering ratio, as those used for studying the interlevel correlations, and the meteorological parameters formed in the same manner using the radiosonde data obtained at the nearest meteorological stations in Novosibirsk and Kolpashevo. Since the aerosol content and temperature distribution are determined by the general atmospheric circulation, their characteristics depend on a synoptic situation which is formed by a random processes on a local scale. Therefore to decrease the effect of random variations and to increase the statistical reliability of the values entering into Eq. (1) they were monthly averaged. The results of calculations of the coefficients of pair correlation between the scattering ratio and the meteorological parameters obtained at 32 monthly averaged points and having the standard deviation lower than 5% are presented in Fig. 2.

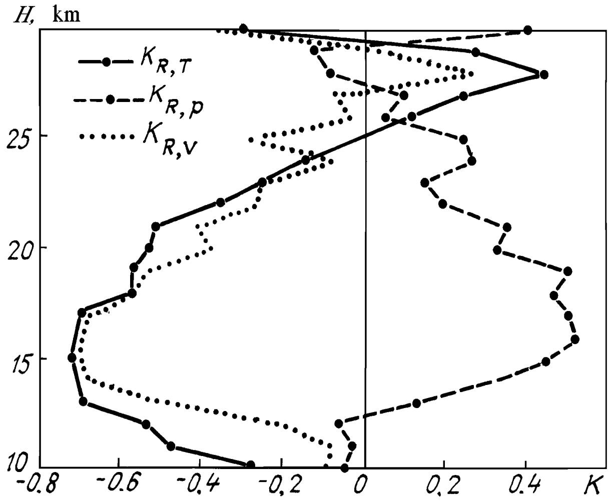

FIG. 2.

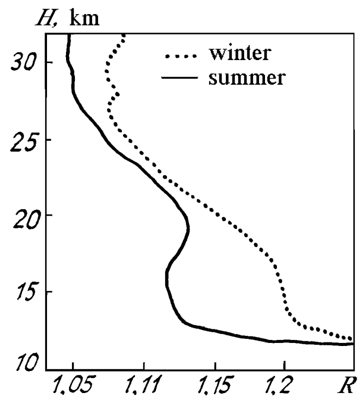

FIG. 3.

The coefficient of correlation between the scattering ratio and temperature $|K_{R,T}|$ in the altitude range from 10 and 25 km is negative which is indicative of the inverse interrelation between the parameters under study (i.e., $R \sim 1/T$). Its highest value (by absolute value) is observed at altitudes 13–17 km. The decrease in $|K_{R,T}|$ is observed above and below this atmospheric layer. The altitude higher than the velopause ~ 20–21 km and up to 28 km is characterized by a monotonic change in $K_{R,T}$ from –0.5 to +0.4. The increase in $|K_{R,T}|$ clearly observed at altitudes from 10 to 13 km is probably determined by the decreased effect of the tropopause and by increased distance between the stratospheric layers and the tropopause where cloud formation often occurs. If the correlation coefficient higher than 0.5 is taken as a criterion for the existence of correlation, then it becomes clear that the statistically significant negative correlation between the temperature and the scattering ratio is manifested only in a range of altitudes between 12 and 21 km. It is this correlation which determines a particularly pronounced seasonal behavior of the scattering ratio in the given range of altitudes. As an illustration of this statement, Fig. 3 presents the mean summer and winter profiles of the scattering ratio obtained during 1986–1988.

The dotted and dashed lines in Fig. 2 show the coefficients of correlation between the scattering ratio and pressure $K_{R,P}$ and the wind velocity $K_{R,V}$, respectively. The vertical profiles of $K_{R,T}$ and $K_{R,P}$ are on the whole similar. Positive values of $K_{R,V}$ with the maximum of 0.5 at altitudes of 16–19 km are observed over the whole altitude range.

Temperature is one of the primary meteorological parameters which finally affects the formation of spatio–temporal fields of pressure and wind.[3] Hence, the values $K_{R,P}$ and $K_{R,V}$ which are higher than 0.5 most probably reflect the interrelationship between R and T related through the pressure and the wind velocity, respectively. Their strongest variability at the altitudes higher than 20 km is probably determined by a nonlinear dependence between $P(T)$ and $V(T)$. For example, following the equation for the state an ideal gas the temperature T and the pressure P are related to by the ratio

$$P = \rho R T ,$$

where ρ is the air density, R is the universal gas constant for air. But in a real atmosphere the air density is subject to temperature–induced variations and wave processes.[3,4]

In this paper the studies have been performed of the interlevel correlation of the scattering ratio and the correlation between the scattering ratio and the meteorological parameters. The closest interlevel correlation is observed in summer in the range of altitudes between 13 and 18 km where the main portion of stratospheric aerosol is concentrated. The correlation between the scattering ratio and the temperature is observed in the stratospheric layer at altitudes from 12 to 21 km. Negative values of the thusly obtained correlation coefficient testify to the inverse functional dependence between the scattering ratio and temperature.

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