SEMI–EMPIRICAL MODELS OF THE UPPER ATMOSPHERIC AEROSOL COMPOSITION. III. COAGULATION MODEL

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A general detailed formulation of the problem of calculating the equilibrium concentration of primary cosmic dust particles at altitudes from 30 *to* 110 m *is presented. The flux of micrometeorites in the mass range from* 10^{-17} *to* 10^{-8} *g is preset at* $z_{\max} = 110$ km *while the source of vapors of the micrometeoric matter is assumed to obey the Gaussian vertical distribution with the maximum at* $z_0 = 95$ km *and with a standard deviation of* ± 5 km *. The sink for this cosmic dust aerosol is the Junge layer of the sulphate aerosol at* $z_{\min} = 20$ km *. The micrometeoric matter influx is assumed to be 45 tons per day.*

As numerical calculations have shown, the condensation of vapors of the meteoric matter takes place at the altitudes of from 18 *to* 100 km. *Below* 50 km *the coagulation process leads to the quick formation of aerosol particles. Thus formed aerosol particles with masses* $m = 10$ g *scatter light most effectively. As a result, the atmospheric turbidity increases with z decreasing. Further increase of the aerosol particle size leads to the increase of the sedimentation rate and, as a consequence, the atmospheric turbidity at altitudes below* 30 km *decreases. This model is the first one which gives the mechanism for the formation of the upper stratospheric light scattering layer of aerosol observed during dusk (see Refs.* 8 , 11 , 13 , 14 , *and* 21*).*

INTRODUCTION

According to the modern conception, the aerosols in the middle atmosphere above 30 km are almost exclusively of the cosmic origin.[9,11,14] The condensation of the water vapor in the process of formation of noctilucent clouds[23] can occur sometimes only within the thin layer in the mesopause. However, even in this case, it is quite probable that the vapor is brought to the upper atmosphere mainly by mini–comets from space[17] while the micrometeorites form the condensation nuclei.[20]

The aerosol in the middle atmosphere is formed due to the inflow of meteoric matter and involves a number of physical processes:[11]

1. Deceleration of micrometeorites with masses $m < 10^{-8}$ g in the upper atmosphere with insignificant mass losses.[22]

2. Almost complete vaporization, due to the deceleration, of larger cosmic dust particles with masses 10^{-8}–10^{-3} g known as meteoroids that produce meteors and bolids.[10]

3. Rapid condensation of vapors of the meteoric matter at the initial stage of the relatively dense trails of bright meteors and bolids before they spread due to the molecular and turbulent diffusion and mixed with the background, and rapid coagulation of aerosol particles formed in the trails.[18]

4. Slower background condensation and coagulation of the meteoric matter vapor and aerosols outside the meteor and bolid trails.

5. Gravitational sedimentation of the micrometeorites and condensed vapors and coagulated meteoric matter through the entire thickness of the atmosphere.

6. Turbulent mixing of the atmosphere that can influence the sedimentation rate of aerosols.

The problem of taking into account all these processes simultaneously is extremely difficult to solve and in fact it is practically insoluble even with the use of modern computers, so nobody have even tried to solve it. Various authors solved some simpler problems such as constructing sedimentation[1,11] and diffusion–sedimentation[2,16] models of micrometeorites, modeling the processes of condensation and coagulation in the trails of meteors and bolids[3,18], and so on. In addition to certain independent significance these particular models enable one to assess the contribution of different factors and to select those, which must be taken into account when describing the aerosols of different masses within different altitude ranges. Finally, this should allow us to obtain a solution of the general problem without significant losses in accuracy.

In this paper we deal with the development of a coagulation model of the background aerosol of cosmic origin in the middle atmosphere. The model is based on the following assumptions:

1. The processes of rapid condensation and coagulation in the trails of meteors and bolids are neglected. The validity of this simplification has to be proved.

2. The latitudinal and longitudinal variations of the influx of the meteoric matter and of the atmospheric parameters are also neglected. This means that only a one–dimensional problem is studied.

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3. Temporal variations of the influx of the meteoric matter and of the atmospheric parameters are neglected, i.e., only the stationary solution of the problem is sought.

4. The sources of vapor of the meteoric matter and of micrometeorites are preset but approximately, and this requires additional foundation.

FOUNDATION OF THE INITIAL DATA

The most detailed model of the influx of primary cosmic dust particles of different masses has been developed in Ref. 11. According to this model total influx of the meteoric matter into the Earth's atmosphere is about 45 tons per day. At present this magnitude of the meteoric matter influx is considered as the minimum one, while estimates of its maximum value exceed it by about an order of magnitude.

According to this model the particles with masses of 10^{-8}–10^{-2} g are responsible for more than 90% of the total influx of vapors of the meteoric matter into the atmosphere. Such particles evaporate predominantly within the relatively thin atmospheric layer with the maximum at the altitude of 95 km (Ref. 10). In this connection we assume that the power of the source of vapor of the meteoric matter $q(z)$ has a Gaussian vertical distribution with the maximum at $z = 95$ km and the variance $\sigma_0 = 5$ km

$$q(z) = \frac{Q}{m_1} \frac{1}{\sqrt{2\pi} s_0} \exp\left[-\frac{(z - z_0)^2}{2\sigma_0} \right], \tag{1}$$

where $Q = 7 \cdot 10^{-17}$ g·cm^{-2}·s^{-1} is the mass flux density of the particles completely evaporating in the upper atmosphere and $m_1 = 10^{-22}$ g is the mean mass of the molecule of the meteoric matter vapor.

Rapid condensation and coagulation are quite efficient only in the trails of meteors and bolids produced by meteoric bodies with masses $m = 1$ g (Ref. 3). Such large bodies make a contribution to the total mass of the meteoric matter which is an order of magnitude lower than that micrometeors. For this reason the contribution of large bodies to the background aerosol can be neglected.

As our calculations have shown, an account of the background condensation and coagulation occurring with the participation of micrometeorites is needed only at altitudes $z < 100$ km. Therefore, we ignore the fact that micrometeorites of different masses moving with different velocities loose their initial cosmic velocity within a wide altitude range from 100 to 150 km and assume that all the micrometeorites have a stationary rate of gravitational sedimentation already at the top $z_{\max} = 110$ km of the altitude range of interest. In addition as follows from our data, below 30 km the sedimentation rate of particles increases rapidly compared to the diffusion rate as a result of the growth of aerosol particles. This effect should become even more rapid in the Junge layer so a sink of aerosol at the altitude $z_{\min} = 20$ km is introduced.

According to the data from Ref. 12 the mean cosmic dust particle density is $\delta = 1.5$ g·cm$^{-3.}$ The most widespread component of the meteoric matter is, on the whole, SiO_2 and the mean molecular mass of the vapors of meteor matter is close in value to the mass of SiO_2 molecule.[19] In this connection we will assume the meteoric matter to be composed of SiO_2 when calculating the background condensation and coagulation of meteoric matter in the atmosphere.

Our model of the background aerosol in the middle atmosphere is minimal since we have accepted minimum estimation of the influx of meteoric matter into the atmosphere. A considerable increase of the influx and a change of the distribution of cosmic dust particles over masses will require certain modification of the model on the whole.

Within the scope of this model of the minimum influx[3] a mathematical description of the aerosol mass distribution is given with an account of sedimentation, diffusion, and background condensation and coagulation processes.

RESULTS OF NUMERICAL CALCULATIONS

The system of coagulation equations given in Ref. 3 has been solved using an interactive scheme with the fixed altitude Δz and automatically choosen time step Δt. This results from the fact that characteristic times 2 of different processes vary in a very wide range from several seconds (formation of the condensation nuclei) to 10^6 s (diffusion and sedimentation). The computation time increases rapidly with broadening of the mass spectrum of aerosols, and therefore our computational scheme is inapplicable if the influx of the meteoric matter increases markedly. In this case the entire range of altitudes (20 – 100 km) should be divided into several subranges where different approximation have to be used.

The profile of the vertical component of the coefficient of the eddy diffusion D_T (Ref. 4) taken for our calculations is shown in Fig. 1. The values of $\tilde{r}_k$ and $\tilde{m}$ calculated for different ranges of variation of masses of aerosols chosen by us are given in Table I.

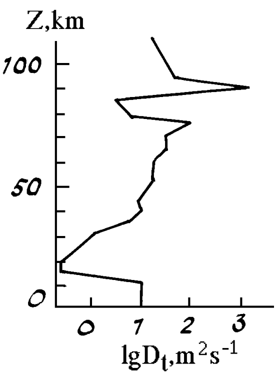

FIG. 1. Vertical profile of the eddy diffusion coefficient vertical component (Ref. 4).

Figure 2 shows the results of calculating the concentrations C_k of the particles of different masses at the altitudes of 30, 40, 60, 80, 100, and 110 km. The vertical profile of the relative turbidity of the atmosphere $s(z)$ is shown in Fig. 3 (curve *3*). For comparison, curves *1* and *2* in this figure show the vertical profiles $S(z)$ calculated using the sedimentation and diffusion–sedimentation models of the micrometeorites in the atmosphere, respectively.[1,2]

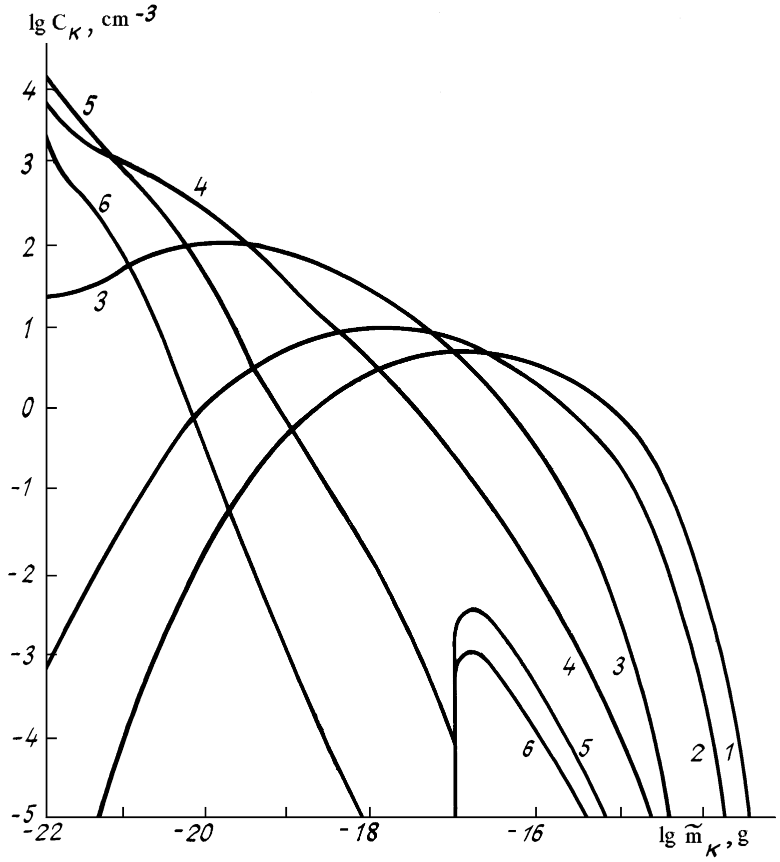

FIG. 2. Absolute concentration of particles of different masses at the altitudes: 1) 30, *2)* 40, *3)* 60, *4)* 80, *5)* 100, *and 6)* 110 km.

TABLE I. Mean radii and masses of particles for ranges of mass variations chosen by us.

k	$\tilde{r}_k$, μm	$\tilde{m}_k$, g	k	$\tilde{r}_k$, μm	$\tilde{m}_k$, g	k	$\tilde{r}_k$, μm	$\tilde{m}_k$, g
1	0.00021	$1.0{\cdot}10^{-22}$	15	0.0062	$2.4{\cdot}10^{-18}$	29	0.16	$4.0{\cdot}10^{-14}$
2	0.00029	$2.5{\cdot}10^{-22}$	16	0.0078	$4.9{\cdot}10^{-18}$	30	0.20	$8.0{\cdot}10^{-14}$
3	0.00037	$5.5{\cdot}10^{-22}$	17	0.0098	$9.8{\cdot}10^{-18}$	31	0.25	$1.6{\cdot}10^{-13}$
4	0.00048	$1.1{\cdot}10^{-21}$	18	0.012	$2.0{\cdot}10^{-17}$	32	0.31	$3.2{\cdot}10^{-13}$
5	0.00061	$2.3{\cdot}10^{-21}$	19	0.016	$3.9{\cdot}10^{-17}$	33	0.39	$6.4{\cdot}10^{-13}$
6	0.00077	$4.7{\cdot}10^{-21}$	20	0.020	$7.9{\cdot}10^{-17}$	34	0.50	$1.3{\cdot}10^{-12}$
7	0.00097	$9.5{\cdot}10^{-21}$	21	0.025	$1.6{\cdot}10^{-16}$	35	0.63	$2.6{\cdot}10^{-12}$
8	0.0012	$1.9{\cdot}10^{-20}$	22	0.031	$3.1{\cdot}10^{-16}$	36	0.79	$5.1{\cdot}10^{-12}$
9	0.0015	$3.9{\cdot}10^{-20}$	23	0.039	$6.3{\cdot}10^{-16}$	37	1.0	$1.0{\cdot}10^{-11}$
10	0.0019	$7.7{\cdot}10^{-20}$	24	0.049	$1.3{\cdot}10^{-15}$	38	1.3	$2.0{\cdot}10^{-11}$
11	0.0024	$1.5{\cdot}10^{-19}$	25	0.062	$2.5{\cdot}10^{-15}$	39	1.6	$4.1{\cdot}10^{-11}$
12	0.0031	$3.1{\cdot}10^{-10}$	26	0.078	$5.0{\cdot}10^{-15}$	40	2.0	$8.2{\cdot}10^{-11}$
13	0.0039	$6.1{\cdot}10^{-19}$	27	0.099	$1.0{\cdot}10^{-14}$	41	2.5	$1.6{\cdot}10^{-10}$
14	0.0049	$1.2{\cdot}10^{-18}$	28	0.12	$2.0{\cdot}10^{-14}$	42	3.1	$3.3{\cdot}10^{-10}$

DISCUSSION OF RESULTS

As can be seen from the data shown in Fig. 2, the vapor of the meteoric matter are condensated mainly within the 80–100 km altitude range. Below 70 km not only the mass of the vapor is much lower than the total mass of aerosols but even the number density of molecules of the meteoric matter is smaller than the number density of aerosol particles. Below 100 km there is an intense coagulation of the smallest aerosol particles (condensation nuclei). Thus, in the curves plotted for altitudes of 100 and 110 km a dip in the mass distribution of aerosols with $m \approx 5{\cdot}10^{-18}$ g (can be distinctly seen caused by the fact that our model did not take into account influx of micrometeorites with masses less than 10^{-17} g) while at the curves plotted for lower altitudes, starting from an altitude of 80 km, this dip is no longer seen.

Taking into account the dependence of the sedimentation rate and of the scattering efficiency factor on the mass of aerosol particles and assuming a fixed total mass influx of the meteoric matter and monodisperse aerosols, one can find that the light of the visible range is most efficiently scattered by aerosol particles with the mass $m \approx 10^{-14}$g. According to our model such aerosols are formed most intensively within 30–50 km altitude range. Just this process gives rise to the increase of the coefficient of relative turbidity of the atmosphere $s = \sigma_a/\sigma_m$ with decrease of altitude from 50 to 30 km (see Fig. 3). Further increase of the aerosol particle size results in an increase in their sedimentation rate and in a less light scattering efficiency of unit mass of the aerosol. This in turn results in lower atmospheric turbidity below 30 km.

This result of our calculations can be regarded as the first description of a physical mechanism of forming the light scattering layer of aerosols in the upper stratosphere that is readily observed during dusk.[8,11,13,14,21] For an adequate description of the characteristics of this layer employing our model, an optimization of the influx of the meteoric matter into the atmosphere is needed.

The data shown in Fig. 3 as well as the results of our previous investigations (Refs. 1, 2) show that assumptions of some authors[6,7] that main contribution to light scattering at an altitude of 100 km comes from aerosol particles with radii of ~ 1 μm unsubstantiated. The matter is that such particles would very rapidly settle at a rate of hundred meters per second and can not be formed, due to coagulation, above 60 km.

As intakes of the micrometeorites in the atmosphere have shown almost half of the particles are friable agglomerations of smaller particles. It is obvious that such friable particles are formed due to coagulation of aerosol in the mesosphere and stratosphere and it is unnecessary that they come from the space.

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