DISTRIBUTION OF CLOUD CONDENSATION NUCLEI WITH HEIGHT IN THE FREE ATMOSPHERE

GEOPHYSICS

1967

SovietRxiv

Abstract

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UDC 551.510.04

GEOPHYSICS

A. G. LAKTIONOV

DISTRIBUTION OF CLOUD CONDENSATION NUCLEI WITH HEIGHT IN THE FREE ATMOSPHERE

(Presented by Academician E. K. Fedorov on 8 XII 1966)

In the atmosphere, at sufficiently small supersaturations, water vapor condenses on cloud condensation nuclei, forming clouds and fogs. As shown in work (ˆ1), in order to measure the concentration of cloud condensation nuclei active at supersaturation S, it is necessary to determine the concentration of droplets that have grown at this supersaturation to sizes exceeding the critical radius r_v of droplets of pure water. The value r_v is determined from the Kelvin equations. Distributions of the concentrations of cloud condensation nuclei with respect to supersaturations that satisfy the above condition for measurements have been published in works (ˆ2–4), where cloud condensation nuclei located in the near-ground layer of the atmosphere were investigated.

The present article gives the results of investigations of cloud condensation nuclei in the free atmosphere. The measurements were made from an aircraft with a flow-through photoelectric apparatus described in work (ˆ4).

Air containing cloud condensation nuclei was taken in through a tube with an inlet aperture 5 mm in diameter, placed outside the aircraft strictly facing the flow. Cloud condensation nuclei active at supersaturations greater than 0.1% are mainly particles less than 0.5μ in diameter. The aspiration coefficients of such particles from the oncoming flow into the tube (the Stokes numbers for such particles are less than 10^{-2}) are practically equal to 1.

The apparatus consisted of a vertically arranged temperature-gradient diffusion chamber, through which the air under investigation was continuously drawn. The time of droplet growth in the diffusion chamber was 250 sec. From the chamber, the air with the grown droplets was drawn through a photoelectric meter of droplet concentration and sizes, by means of which the integral concentration of droplets of radius $r \geqslant r_v$ was determined. The temperature of the inner cylinder surface of the diffusion chamber, continuously wetted with water, was kept constantly equal to 25°, while the temperature of the outer cylinder surface of the diffusion chamber, continuously wetted with water, was set equal

Fig. 1

Figure 1: Fig. 1

to one of the following values: 1) 26.7° (with the maximum supersaturation created in the chamber $S = 0.1\%$); 2) 27.3° ($S = 0.2\%$); 3) 28.4° ($S = 0.3\%$); 4) 28.3° ($S = 0.4\%$); 5) 29.3° ($S = 0.65\%$). The temperature of both inner surfaces of the chamber was maintained constant by means of two thermostats.

In 1966, 9 soundings of the atmosphere were carried out: on 19 and 23 IV in the Bukhara region, on 26 and 27 IV in the Tashkent region, on 18 V, 9 VI, and 21 VII in the Riga region, on 4 VIII in the Daugavpils region, and on 8 VIII in the Smolensk region. All atmospheric soundings were carried out during the daytime, with air temperature at the ground from +15 to +25°, in clear weather (no more than 3 tenths of lower- and middle-tier cloudiness). The soundings were carried out over territories at least 50 km distant from large cities. The distributions of cloud condensation nuclei concentrations with respect to supersaturations were measured at levels of 200, 1000, 2000, 3000, 4000, and 5000 m.

Analysis of the material obtained showed that in the lower layer of the troposphere (200–1000 m) the relative integral distribution of the concentrations of cloud condensation nuclei with respect to supersaturations (the cumulative function $N, \%$) in the region $S \leqslant 0.6\%$ is satisfactorily approximated by the incomplete gamma function (1) with index α, varying from 1 to 5:

$$N(\%) = \int_0^{S/\beta} e^{-S/\beta} \left(\frac{S}{\beta}\right)^{\alpha-1} d\left(\frac{S}{\beta}\right). \qquad (1)$$

The dependence of N on S, averaged from the measurement results in the lower layer of the troposphere (curve 1 in Fig. 1), is described by an incomplete gamma function with parameters $\alpha = 2$ and $\beta = 0.07\%$. In this layer the mean integral concentration of cloud nuclei active at supersaturations $S \leqslant 0.6\%$ is 300 cm^{-3}. The distributions found for the concentrations of cloud condensation nuclei with respect to supersaturations in the lower layer of the troposphere agree with the distributions in the near-ground layer ([4]).

Fig. 1. Distribution of the integral concentrations of cloud nuclei with respect to supersaturations in the layer 200–1000 m (1) and in the layer 4000–5000 m (2)

In each of the soundings, in the layer from 4000 to 5000 m, the distributions of the integral concentrations of cloud condensation nuclei with respect to supersaturations remained constant. In this layer it proved possible to approximate the dependence $N_{S \leqslant S_i}$ on S by a one-parameter function:

$$N_{S \leqslant S_i} = A(S)^{\chi} \tag{2}$$

with exponent χ, varying from 1 to 2. In the 4000–5000 m layer, the mean integral distribution of the concentrations of cloud condensation nuclei with respect to supersaturations is described by the power function (2) with exponent $\chi = 1.5$ and $N_{S \leqslant 0.6\%} = 60$ cm^{-3} (curve 2 in Fig. 1).

The measurement data also make it possible to analyze changes with height in the concentrations of cloud condensation nuclei. In all atmospheric soundings the concentration of cloud condensation nuclei changes monotonically with height. Despite the quantitatively different rate of decrease of cloud-nuclei concentration with height, in all soundings a qualitatively identical character of their change was recorded. In the lower layer, from 200 m to 1.5 km, a slow decrease in the concentration of cloud condensation nuclei is observed; in the layer from 2 to 3 km there is a more rapid decrease in the concentration of cloud nuclei; in the layer from 3.5 to 5 km the concentration of cloud nuclei changes little. The mean change with height in the concentration of cloud nuclei $N_{S \leqslant 0.3\%}$ is shown in Fig. 2 (curve 1).

In the atmospheric soundings indicated above, the concentrations of giant aerosol particles of diameter $d_2 \sim 20\mu$ were determined with a photoelectric cloud-droplet meter described in ([5]).

In the photoelectric cloud-droplet meter, the air under investigation is blown through the open shaft of the instrument, 25 mm in diameter. The instrument shaft is positioned strictly in the direction of the oncoming flow. Particle concentrations are determined from the number of particles that have crossed the illuminated volume of the jet, which is located on the axis of the shaft. The velocities on the axis of the shaft and in the oncoming flow are equal. Under such isokinetic sampling conditions, the aspiration coefficients of the investigated particles from the oncoming···

of the stream entering the central part of the shaft are equal to 1. In Fig. 2 (curve 2) is shown the decrease with height, averaged from the results of the atmospheric soundings carried out, in the concentration of the fraction of aerosol particles with diameters of 19–21 μ.

From the course of curves 1 and 2 in Fig. 2 it is seen that the nature of the change with height in the concentration of giant aerosol particles and cloud condensation nuclei is the same. Three layers are clearly distinguished with height, in each of which the decrease in the concentrations of cloud nuclei and giant aerosol particles follows an exponential law.

In the absence of significant horizontal gradients in particle concentration, the equation of stationary vertical transport of aerosol particles in the atmosphere is written as

Fig. 2. Changes with height in the concentration of: cloud condensation nuclei active at supersaturation $S \sim 0.3\%$ (1), aerosol particles 20 μ in diameter (2), and Aitken nuclei (3)

Figure 2: Fig. 2. Changes with height in the concentration of: cloud condensation nuclei active at supersaturation $S \sim 0.3\%$ (1), aerosol particles 20 μ in diameter (2), and Aitken nuclei (3)

$$K\, d^2N/dz^2 + w\, dN/dz - aN^2 - \psi = 0, \tag{3}$$

where K is the mixing coefficient; w is the Stokes settling velocity of the particles; a is the coagulation constant, and ψ is the function of aerosol removal.

Removal of aerosols in the atmosphere occurs both through the descending motions of air masses observed in regions of increased pressure, and through the washout of aerosols by precipitation. In the first case the function ψ is written as

$$\psi_1 = -c\, dN/dz, \tag{4}$$

where c is the velocity of the descending motion of the air. The rate of washout of particles may be considered proportional to their concentration,

$$\psi_2 = bN, \tag{5}$$

where b is the washout coefficient.

Fig. 2. Changes with height in the concentration of: cloud condensation nuclei active at supersaturation $S \sim 0.3\%$ (1), aerosol particles 20 μ in diameter (2), and Aitken nuclei (3)

The influence of particle coagulation on the vertical transport of cloud condensation nuclei and larger aerosol particles, as will be shown below, is small. In carrying out the calculations we shall neglect the term aN^2 in equation (3).

The solution of equation (3) for the i-th fraction of aerosol particles, taking into account descending motions in an atmospheric layer of height Δz and the boundary conditions: at $\Delta z = 0$, $N_i = N_i(0)$, and at $\Delta z = \infty$, $N_i = 0$, is written as

$$N_i(\Delta z) = N_i(0) \exp[-(c + w_i)/k]\Delta z. \tag{6}$$

When the washout of particles by clouds and precipitation is taken into account, the solution of equation (3) under the same boundary conditions is represented as

$$N_i(\Delta z) = N_i(0) \exp\left[-w_i/2K - \sqrt{(w_i/2K)^2 + b/K}\right] \Delta z. \quad (7)$$

The exponential character obtained for the decrease with height of the concentration of cloud condensation nuclei (particle diameter $d_1 \sim 0.1\ \mu$ and $w_1 \sim 0$ cm·sec^{-1}) and of giant aerosol particles ($d_2 = 20\ \mu$ and $w_2 \sim 2$ cm·sec^{-1}) agrees with the solutions of equation (3).

From the averaged data (curves 1 and 2 in Fig. 2), estimates were obtained for the quantities c, K, and b in two layers: in the lower layer of the troposphere, 200-1200 m, and in the layer 1800-2800 m. The velocity of the descending motion of the air, c, in both layers proved to be equal to 1 cm $\cdot$ sec^{-1}, which agrees with the estimate of c made earlier [6].

The values of the coefficients of vertical mixing, calculated from formulas (6) and (7), practically coincide: in the lower layer of the tropo-

of the sphere $K = 1 \cdot 10^6$ cm$^2 \cdot$ sec^{-1}, and in the layer 1800-2800 m, $K = 2 \cdot 10^5$ cm$^2 \cdot$ sec^{-1}. The high coefficient of mixing in the lower layer of the troposphere is explained by thermal convection, which develops intensively in summer during the daytime. The values of K are consistent with data obtained earlier ([6]).

On the basis of the data obtained, the values of the washout coefficients were calculated under the assumption that these coefficients are equal for giant aerosol particles and cloud condensation nuclei. The washout coefficient of particles in the lower layer of the troposphere is 10^{-6} sec^{-1}, and in the layer 1800-2800 m $b = 3 \cdot 10^{-6}$ sec^{-1}. At a concentration of cloud condensation nuclei in the lower layer of the troposphere $N_1 \sim 300$ cm^{-3}, the rate of their washout is $bN_1 \sim 3 \cdot 10^{-4}$ cm$^{-3} \cdot$ sec^{-1}. In the lower layer of the troposphere, the rate of coagulation decrease in the concentration of cloud condensation nuclei ($d_1 \sim 0.1\ \mu$, $a_1 \sim 5 \cdot 10^{-10}$ cm$^3 \cdot$ sec^{-1}), $a_1 N_1^2 \sim 4 \cdot 10^{-5}$ cm$^{-3} \cdot$ sec^{-1}, is an order of magnitude less than the washout rate. The coagulation rate of giant aerosol particles ($d_2 = 20\ \mu$, $a_2 = 3 \cdot 10^{-10}$ cm$^3 \cdot$ sec^{-1}, $N_2 \sim 6 \cdot 10^{-3}$ cm^{-3}), equal to $a_2 N_2^2 \sim 1 \cdot 10^{-14}$ cm$^{-3} \cdot$ sec^{-1}, is negligible. Consequently, neglect of the coagulation term in equation (3) is justified.

With a decrease in particle size and an increase in their concentration, the role of the coagulation term in equation (3) increases. The decrease with height of the concentration of Aitken nuclei, as shown in ([7]), is determined mainly by particle coagulation and is described by the parabolic formula:

$$N_0(z) = N_0(0)\,[D/(z + D)]^2, \quad (8)$$

in which z is the height in meters, and D is a coefficient whose mean value for the warm half-year is 1200. In Fig. 2, curve 3 shows the character of the change in the concentration of Aitken nuclei with height ($\bar{D} = 1200$). Comparison of curves 1 and 3 in Fig. 2 confirms that the most distinct difference in the decrease

with height of the concentrations of cloud condensation nuclei and Aitken nuclei is manifested in the lower layer of the atmosphere, where, owing to the maximum concentration of Aitken nuclei ($N_0 \sim 10^4$ cm^{-3}), the principal mechanism of the decrease of their concentration with height is particle coagulation.

Institute of Applied Geophysics

Received
24 XI 1966

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Note: Figure translations are in progress. See original paper for figures.