Physics – Uspekhi **57** (11) 1041 – 1062 (2014)

REVIEWS OF TOPICAL PROBLEMS

PACS numbers: 92.60.Bh, 92.60.Mt, 92.60.Pw

Electrical cycle in the Earth's atmosphere

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DOI: 10.3367/UFNe.0184.201411a.1153

Contents

Abstract. **A qualitative physical picture of atmospheric electricity as a secondary phenomenon of atmospheric water circulation is presented using the key electrical atmospheric processes, their average observed parameters, and a detailed analysis of separate processes as the basis. The electrical processes begin with the charging of aerosols at kilometer altitudes due to aerosols of different aggregate states colliding with one another. Atmospheric electric currents arise from the fall speed difference (in cumulus clouds) between mass-different positively and negatively charged aerosols, resulting in aerosols with a negative charge of, on average, (25–30)*e* at the lower edge of a cloud. This creates an electric field between Earth and the cloud, and the subsequent penetration of streams of warm wet air into the cloud causes the atmosphere to electrically break down, thus producing lightning flashes. At the same time, these processes cause aerosols to grow and to fall as rain onto Earth. Processes in atmospheric air, including those involving aerosols, electrons, and ions, provide a unified physical picture of electric phenomena in the terrestrial atmosphere.**

1. Introduction

Electric phenomena in the Earth's atmosphere are the subject of this paper; they include the formation processes of electric charges in the atmosphere and the passage of electric currents through the Earth's atmosphere. Atmospheric electrical processes are secondary phenomena with respect to water circulation [1]; they are governed by known physical laws that allow one to study them from the standpoint of general physical concepts. Investigations of electrical processes in the Earth's atmosphere and on the Earth's surface, which give our contemporary understanding of them, have been carried out for a few centuries. In particular, the study by Benjamin Franklin (USA) in the middle of the 18th century, flying a kite on a metal wire in a thundercloud, exhibited the identity of atmospheric electricity with that obtained by friction on a revolving circle and proved the electrical nature of lightning. Franklin established two forms of electric currents, positive and negative, and demonstrated the possibility of taking away electric charges from the atmosphere by using metal wires erected over buildings and directed underground.

The understanding of the electrical nature of thunderstorm phenomena allowed the construction of the lightning rod, which appeared in the USA in 1752 and at approximately the same time in what is now the Czech Republic [2], and this was the first practical step toward the protection from atmospheric electricity. On the other hand, investigations by M V Lomonosov (theoretically) and G V Richter, who was killed by a lightning discharge during measurements of atmospheric electric currents, allowed us to understand some aspects of atmospheric electricity. In particular, Lomonosov assumed that atmospheric electricity results from friction of dust particles and other weighted air particles against water droplets. This corresponds, in principle, to the contemporary concept of the origin of atmospheric electricity, with the key process resulting from collisions of snowlike or liquid water aerosols with micron-sized ice particles, which leads to particle charging [3–11]. The experience obtained in the study of atmospheric electrical processes, in the understanding of the nature of these processes, and in the creation of new measuring instruments, provided a basis for the development of some modern areas of

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Received 20 January 2014, revised 28 June 2014
Uspekhi Fizicheskikh Nauk **184** (11) 1153 – 1176 (2014)
DOI: 10.3367/UFNr.0184.201411a.1153
Translated by B M Smirnov

physics, such as plasma physics and the physics of high energies.

The large contribution of C T R Wilson to developing and understanding the physics of atmospheric electricity at the beginning of 20 century should be noted. He constructed and modified a condensation chamber for charged particle detection, which is known as the Wilson chamber, and awarded the Nobel Prize in Physics 1927. Measurements on the basis of this chamber, which were subsequently coupled with balloon measurements and the analysis of results of such measurements, allowed Wilson to formulate and understand a series of problems of atmospheric electricity, whereas the subsequent development of his concepts occurred with a substantial delay. Namely Wilson [12, 13] demonstrated the production of ions in the atmosphere. He carried out the first measurements of electric currents flowing in the atmosphere and the electric field strength near the Earth's surface [14, 15]. Wilson proved that the Earth is charged negatively [16–18], and cosmic rays (X-ray radiation and fast particles penetrating the Earth's lower atmosphere from the Sun) cause discharging of the Earth. Because of a low air density at high altitudes, charged clouds may cause electric breakdown in upper layers of the atmosphere [19, 20]. Glowing objects resulting from this breakdown are now called sprites. Next, Wilson revealed the possibility of formation of fast electrons at high altitudes in the Earth's atmosphere [21]. This atmospheric phenomenon is now called running electrons [22].

Progress in the study of thunderstorm phenomena proceeded in the first half of the 20th century on the basis of photographing lightning with the Boys camera and measuring the electric and magnetic fields of lightning, atmospheric electric currents, and the spectroscopic analysis of lightning radiation that allow one to construct a physical picture of thunderstorms in the atmosphere. Figure 1 gives the distribution of electric charges in a thundercloud [23] that follows from measurements of the parameters of thunderstorms in South Africa. These data give the physical picture of lightning as a physical phenomenon [24–29, 31]. One can add to this radar diagnostics of clouds, recording of various waves in a thunderstorm, laboratory modeling and the theoretical analysis of electrical processes in the atmosphere. Methods of registration of thunderstorm processes in the atmosphere and the physical picture of the electric breakdown of the atmosphere resulting from such measurements are given in books [32–35].

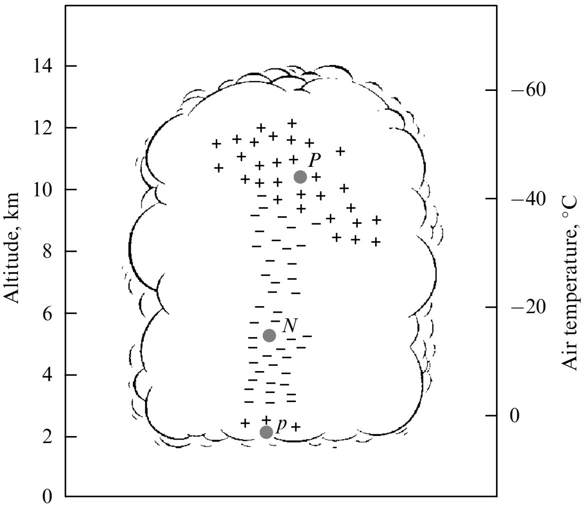

Figure 1. Typical distribution of charges in a thundercloud [23] which is constructed on the basis of measurements of cloud electric fields in South Africa. In this case, the positive charges are $P = 40$ C, $p = 10$ C; the negative charge is $N = -40$ C.

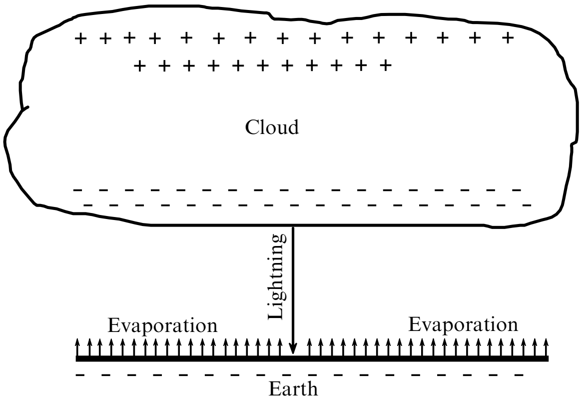

Figure 2. Simplified diagram of electrical processes which accompany water circulation in the atmosphere. Earth charging is realized by lightning [36], corresponding to breakdown between clouds and the Earth's surface [37, 38]. Clouds are formed and decay as a result of moisture circulation through the atmosphere due to water evaporation from Earth's surface and its returning in the form of rain and precipitated aerogels.

As follows from Fig. 1, basic electrical processes proceed in a relatively thin layer of the atmosphere with a depth of approximately 10 km. Typical electric field strengths due to the separation of charges in a cloud are on the order of 1 kV cm^{-1}, and the voltage for a charged cloud layer is on the order of tens of MV. But lightning is only one phenomenon of atmospheric electricity, though the most remarkable. Just lightning transfers a negative charge to Earth's surface and provides the negative Earth charge (Fig. 2) [17, 20, 36–38]. This charge creates an electric field in the atmosphere, and weak electric currents flow under the action of this electric field and lead to Earth's discharging. As follows from Figs 1, 2, atmospheric electrical properties are created in clouds. Hence, the properties and dynamics of atmospheric clouds [37, 39–45] are an important element of atmospheric electricity.

Though atmospheric electric currents are distributed nonuniformly both over the Earth's surface and over time, it is convenient to use for the analysis of atmospheric processes the average parameters of atmospheric electricity [46–48], which are weighted over the surface of the globe and time. The averaged values of atmospheric electrical parameters are checked in time that determines the reliability of these data and allows the construction on their basis of a simple model of atmospheric electricity. Of course, Nature is richer and more multifarious than a simple model which, nevertheless, allows imparting a qualitative character to the physical phenomena of atmospheric electricity.

The goal of this review is the analysis of a totality of key atmospheric electrical processes, the understanding of which has existed for half a century and is based on both measurements of atmospheric electrical processes and laboratory investigations. Adding to this the analysis of individual physical processes related to atmospheric electricity and based on information about some plasma processes, we present a logically closed qualitative physical picture of atmospheric electric phenomena.

2. Earth's atmosphere as an electrical system

2.1 Peculiarities of atmospheric electrical processes

Let us formulate a general problem of this review and the method of its analysis. A diagram of corresponding atmospheric processes is given in Fig. 3, which represents a generalization of Fig. 2. These processes form a closed chain, with some of them being rather natural, while the analysis of other ones gives important information. Our goal is to use this information for a qualitative analysis of atmospheric electrical processes that finally leads to a detailed description of atmospheric electricity.

In the analysis of electrical processes in the Earth's atmosphere, we operate with electrical parameters averaged over time and the surface of the globe. Let us clarify an error in this analysis. In the first place, one can find the result of this approximation if we use it for the analysis of meteorological parameters. In this case, we operate with power rates of some processes for the globe as a whole [49–51]. In particular, the total power of Sun's radiation that reaches the Earth's atmosphere is 1.7×10^{17} W. If this power is further emitted by the Earth as a whole, its temperature will be equal to 293 K. This information would correspond to a weather analysis for the method under consideration. On the contrary, if we consider atmospheric electrical processes from the standpoint of requirements for weather forecasting, in the analysis of electric phenomena it is necessary to predict at what time a thunderstorm would be observed in a given locality and which house would be struck by lightning.

Let us consider the distribution of thunderbolts over the globe's surface. Approximately 100 lightning flashes per second are observed, on average, over the globe according to registered data [52–54]. More precise observations of flashes are made by NASA satellites [55], according to which 44 lightning flashes are observed per second over all the globe. A degree of coincidence of the above results testifies to their accuracy. Next, the most frequent observations of thunderstorm phenomena relate to continental and coastal areas, whereas over oceans and in the polar regions they are rarely observed. Approximately 70% of thunderstorms correspond to tropical land, in accordance with the concept that thunderstorms are observed in warm seasons.

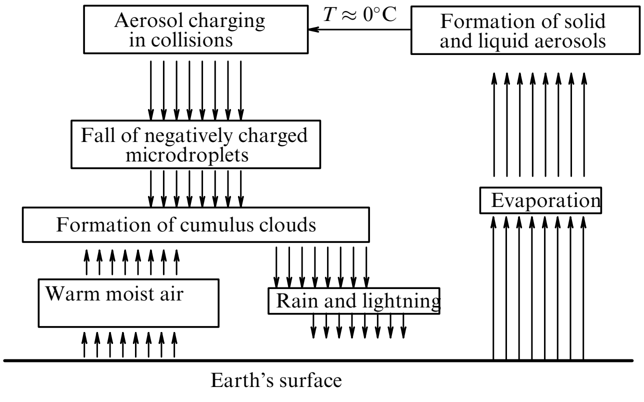

Figure 3. General diagram of water circulation and electrical processes in the Earth's atmosphere constructed on the understanding reached in Refs [36–38] concerning the principles of atmospheric electricity and processes of cloud evolution.

Figure 4. Typical Catatumbo lightning [57].

A champion for rates of flashes is the village of Kifuka in the Democratic Republic of Congo. This village is located at an altitude of 975 m above sea level, and the intensity of lightning flashes is 160 km^{-2} per year, whereas the average of this quantity over the globe is 3 km^{-2} per year. Another example of a region with a higher probability of thunderstorms is the mouth of the Catatumbo river, which flows into Lake Maracaibo in Venezuela, where often thunderstorms are called the Catatumbo phenomenon (Fig. 4) [56–58]. In this region, thunderstorms are observed at night and last approximately 10 hours for 150 days per year. The total number of lightning flashes reaches 1.2×10^6 per year. The Catatumbo region is surrounded on three sides by chains of the Andes and Cordilleras 3700 meters in height above sea level, which causes a specific character of air fluxes. In addition, surrounding marshes produce methane, whose jets facilitate formation of lightning flashes between clouds with lengths of up to 10 km. Assuming that each flash transfers a charge of 5 C, on average [32], it can be concluded that the contribution of the Catatumbo region is 0.2 A to the total electric current of Earth's surface, which is 1700 A, on average. As is seen, thunderstorms are distributed nonuniformly over the Earth's surface, so that the subsequent usage of averaged parameters for atmospheric electrical processes in the analysis of atmospheric electricity leads to a qualitative picture of this phenomenon.

2.2 Electrical processes in the Earth's atmosphere

Let us discuss the average electrical parameters of the Earth's atmosphere and the Earth as a whole. Earth is charged negatively, and the average electric field strength near the Earth's surface is maintained at a level of $E_0 = 130$ V m^{-1}, which corresponds to the Earth's negative charge $Q = E_0 R_{\oplus}^2 = 5.8 \times 10^5$ C [47, 59] ($R_{\oplus} = 6300$ km is the Earth's radius). The Earth's voltage with respect to the surrounding space is approximately $U_0 \simeq 240-300$ kV [60, 61]. Let us model the Earth by a spherical capacitor, with the Earth's surface as the lower electrode, if the electric field strength decreases linearly with altitude, and the voltage is zero at the upper electrode of this capacitor. From this picture, we obtain the distance between electrodes $L = 2U_0/E_0 \approx 5-6$ km. It follows from these estimates that electrical processes in the Earth's atmosphere proceed at altitudes of several kilometers.

We use an appropriate model for the Earth's electrical system, assuming that the Earth's charge is screened by an excess positive charge at atmosphere altitudes up to L, so that the total atmospheric positive charge compensates for the Earth charge. This takes place if positive and negative molecular ions are present in the atmosphere with the difference ΔN in their number densities. Then, the Poisson equation for the electric field strength $E(h)$ as a function of the altitude h has the form

$$\frac{\mathrm{d}E}{\mathrm{d}h} = 4\pi e \Delta N\,. \tag{2.1}$$

Assuming the difference ΔN between the number densities of positive and negative ions to be independent of the altitude, we obtain the solution of this equation as

$$E = E_0\left(1 - \frac{h}{L}\right), \quad L = \frac{E_0}{4\pi e \Delta N}\,. \tag{2.2}$$

It is apparent that

$$L = \frac{2U_0}{E_0}\,, \quad \Delta N = \frac{E_0}{4\pi e L} = \frac{E_0^2}{8\pi e U_0}\,.$$

On the basis of the above values, we have $L \approx 5$ km and $\Delta N \approx 3\ \mathrm{cm}^{-3}$. These estimates of parameters have a qualitative character, because atmospheric parameters vary in time and space. The above value of the charged layer thickness L testifies to the fact that electric phenomena which are responsible for the Earth's charging proceed at altitudes of a few kilometers.

In addition, note that according to the measurements carried out, the average current density discharged by Earth equals 2.4×10^{-16} A cm^{-2} over land, and 3.7×10^{-16} A m^{-2} over oceans [47, 30]. This corresponds approximately to the total discharging current for the Earth's atmosphere $I = 1700$ A [36]. Molecular ions located in the Earth's atmosphere have nearby values of mobilities equal approximately to 2 cm^2 V^{-1} s^{-1} [62]. This mobility leads to a typical time $\tau = Q/I \approx 6$ min of Earth's discharging. As is seen, a mechanism responsible for charging Earth must exist to maintain the negative Earth charge. This process results from electric breakdown between clouds and the Earth's surface through lightning flashes [36]. Exactly lightning flashes transfer a negative charge to the Earth's surface and provide the charging of the Earth [37, 38]. In turn, it is necessary to induce a high voltage between clouds and Earth's surface for lightning formation [33, 35] that determines the lightning specifics. Though we consider negative charging of the Earth by lightning flashes, positive charging of the Earth as a result of electric breakdown between clouds and the Earth's surface is also not prohibited. According to measurements [63], the ratio of the numbers of negative and positive lightning flashes is 2.1 ± 0.5, and the ratio of currents for transfer of the negative and positive charges is 3.2 ± 1.2.

Lightning is a most important stage of electrical processes in the Earth's atmosphere. This electric breakdown proceeds due to a cloud electric charge and may be both between a cloud and Earth's surface and between two clouds. A basic stage of lightning is an electric current which flows along a conductive lightning channel. This is of interest as the stage of transfer of the basic electric charge to the Earth's surface. Below, we give some average lightning parameters by which we will be guided in the subsequent analysis.

An electric field between clouds and the Earth results from charge separation in the atmosphere due to a fall of charged aerosols, and a relatively high electric charge is accumulated at the lower edge of a cumulus cloud, which is more often negative. The electric potential of the lower edge of a cumulus cloud is about $20 - 100$ MV [53] with respect to the Earth's surface, and lightning passes a distance of several kilometers (see, for example, Fig. 5) at a cloud–Earth separation of about $1 - 5$ km. Lightning transfers a part of the cloud charge to Earth. An average charge transferred by a typical lightning is approximately 20 C, and each lightning occurrence consists of 3–4 flashes [33, 35], i.e., lightning includes several current pulses which propagate along almost the same channel. Therefore, each flash, i.e., each current pulse, together with a subsequent phase of a weaker continuous current which follows after a flash, transfers a charge of approximately 5 C [32–35]. Air is heated in the process and expands as a result of the propagation of the electric current, so that the air pressure in the lightning conductive channel is equalized by the pressure of the surrounding air. If the rate of air expansion exceeds the speed of sound in cold air, 3.3×10^4 cm s^{-1} [64], a shock wave arises.

Figure 5. Lightning discharge propagating along one channel [65].

Let us consider the character of propagation of the electric current in lightning. The first stage in the evolution of lightning consists in the formation of the conductive channel in air as a result of the propagation of a stepwise leader [32–35]. A stepwise leader represents a weakly glowing phenomenon which propagates along a broken line with a segment length of tens of meters. The first stage of propagation of a signal along each segment has a streamer nature. Next, a plasma case is formed near the streamer head, providing partial air ionization in this domain [66–69]. A typical velocity of propagation of the stepwise leader is on the order of 1×10^7 cm s^{-1}, which corresponds to the electron drift velocity in these electric fields. When a conductive channel is created, an electric current flows along it and causes intense glowing of the channel plasma. This stage of lightning is called the recurrent stroke and proceeds with a high velocity of up to 5×10^9 cm s^{-1}, which corresponds to propagation of the electric signal front in conductors. The recurrent stroke is relatively short, on the order of a few μs, so that the lightning discharge includes all the phases of channel creation, and propagation of the electric current of an individual flash lasts less than 10^{-3} s. Despite the fact that during this time the conductive channel expands, energy inserted into the lightning channel under the action of the electric current is partially conserved and is consumed on air heating, dissociation, and partial ionization, both in a conducting channel and in adjoining regions.

It should be noted that the lightning electric energy is relatively high and is concentrated in a restricted space area. In order to demonstrate the scale of this energy, let us compare the energy of a lightning flash carrying a charge of 5 C at a voltage of 50 MV with the energy of a gun shell which is moving with a velocity of 760 m s^{-1}. Equalizing the lightning energy and kinetic energy of the gun shell, the shell mass is determined to be 3 t. This estimation demonstrates the danger of lightning for houses and buildings without a lightning protector that allows carrying a lightning charge into the Earth via an artificial conducting channel. From this, it also follows that a carried charge might be dangerous for neighboring objects.

One can add to this that the processes of water evaporation from the Earth's surface and discharging processes of the Earth proceed more or less uniformly over its surface, as do charging processes in cumulus clouds, whereas processes involving lightning proceed in small areas on the Earth's surface. Therefore, local energy fluxes accompanying the charge transfer from a cloud to the Earth's surface are quite high. Indeed, the flux of energy consumed for water evaporation (and, correspondingly, subsequently released upon water condensation) amounts to 6×10^{-3} W cm^{-2}, on average, and the average energy flux for the Earth's discharging is 1×10^{-10} W cm^{-2}. The local energy fluxes in lightning are higher, because they relate to a low area of the Earth's surface. Let us take according to the above data the rate of lightning discharges to be 100 flashes per second [52–54], and that the energy is released in an area of 1 m^2 in a time on the order of 1×10^{-4} s. Assuming that the development of lightning proceeds under the action of a voltage of 50 MV, we obtain about 1×10^{-9} W cm^{-2} for the local power flux. This flux characterizes power transfer, rather than power deposition, but this energy release exhibits a high power in comparison with the initial process of water transport.

In order to estimate the scale of the thunderstorm phenomenon, let us use a simple model of a cumulus cloud in the form of a metal disk with a charge $Q = 20$ C that is transferred by a typical lightning. Requiring the electric field strength near the disk surface E to be below the breakdown electric field strength $E = 30$ kV cm^{-1} for dry air, we arrive at the charge density $\sigma = E/(2\pi)$ at the disk below about 5×10^{-9} C cm^{-2}, which corresponds to a disk radius above $R = \sqrt{Q/(\pi\sigma)} = 400$ m. This testifies to the threshold character of thunderstorm phenomena, so that the cloud size must exceed a certain threshold value. Indeed, thunderstorm phenomena are developed in volcanic dust if the cloud size exceeds 500 m, and the size of a usual cloud must exceed 3 km if the thunderstorm proceeds in air [59].

2.3 Water in the Earth's atmosphere

Since atmospheric electrical processes accompany water circulation in the Earth's atmosphere [1] and electrical processes are considered for the Earth as a whole, let us present parameters of water transport through the atmosphere, which are averaged over time and are summed over the globe surface. The rate of water evaporation from the Earth's surface and its passage through the atmosphere reaches 4×10^{20} g $year^{-1}$ or 1.3×10^{13} g s^{-1}, as is the rate of water precipitation on the Earth's surface. The average content of water in the atmosphere amounts to 1.3×10^{16} kg [70], in addition to the water mass of 5.1×10^{18} kg located on the Earth's surface. Correspondingly, the average specific water density near the Earth's surface equals 3 g m^{-3}, in comparison with the specific density of 1.2 kg m^{-3} for dry air. The indicated specific density of water in air corresponds to the partial water pressure of 2 Torr near the Earth's surface. The saturated vapor pressure of water at a temperature of 0 °C is 4.7 Torr [71]. One can conclude from these estimates that water is located in the Earth's atmosphere mostly in the form of molecules, and the condensation of water vapor in regions with a low temperature leads to the formation of aerosols — micron-sized water particles. Since the condensation proceeds at temperatures near the melting point of water, water may be found in aerosols in different aggregate states.

One can consider the atmospheric electric current as a secondary phenomenon with respect to water circulation in the atmosphere. The ratio of the net atmospheric electric current ($I = 1700$ A) to the water mass transferred through the atmosphere per unit time, $\mathrm{d}M/\mathrm{d}t = 1.3 \times 10^{13}$ g s^{-1}, gives the following specific transferred charge:

$$\frac{I}{\mathrm{d}M/\mathrm{d}t} = 1.4 \times 10^{-10}\ \mathrm{C\,g^{-1}}\,. \tag{2.3}$$

We will be guided by this magnitude in the analysis of atmospheric electrical processes.

The temperature of atmospheric air decreases with increasing the altitude up to the tropopause [70] (the layer of the atmosphere with the minimal temperature) that is located at the altitude of 9 km for the poles, and 17 km for the equator, where the temperature has a minimum. The average temperature on the Earth's surface is 15 °C, the tropopause temperature −55 °C, and the temperature gradient −6.5 K km^{-1} [72]. From this it follows that the water condensation proceeds at altitudes above 2–3 km. The micron-sized aerosols formed are liquid, snow, or ice microparticles, or possibly a mixture of these forms.

Table 1 [73] contains the temperature dependences for the saturated vapor pressure for water, and water microparticles are formed at pressures which exceed the saturated vapor pressure at a given temperature. Because just water aerosols

Table 1. Saturated water vapor in atmospheric air [73]: p_{sat} is the saturated vapor pressure for water at an indicated temperature, N_{sat} is the number density of water molecules in a saturated vapor, and C is the ratio of the mass of saturated water vapor to the air mass located in a given volume.

T, °C	T, K	p_{sat}, kPa	N_{sat}, 10^{17} cm^{-3}	C, g kg^{-1}
−20	253	0.103	0.295	0.631
−15	258	0.165	0.463	1.01
−10	263	0.260	0.716	1.59
−5	268	0.402	1.09	2.46
0	273	0.611	1.62	3.74
5	278	0.873	2.27	5.35
10	283	1.23	3.14	7.52
15	288	1.71	4.29	10.4
20	293	2.34	5.78	14.3
25	298	3.17	7.70	19.4
30	303	4.25	10.2	26.0

are responsible for atmospheric electric phenomena, these processes take place if the water content exceeds a threshold value. In particular, it is usually assumed that thunderstorm weather is attained if the water content exceeds 7 g kg^{-1} (grams of water per kilogram of air) [59]. This water content corresponds to the saturated vapor pressure at the temperature 8 °C.

2.4 Key processes of atmospheric electricity

A diagram illustrating the chain of processes of water transport through the Earth's atmosphere and accompanying electrical processes is given in Fig. 3. Nevertheless, the key processes of atmospheric electricity are the formation of charged aerosols and the subsequent charge separation, which leads to the creation of atmospheric electric fields. By aerosols are implied water particles and water droplets of micron sizes. Let us consider briefly the key processes of atmospheric electricity. One can expect that molecular ions determining an electric current in a dry atmosphere give rise to atmospheric electrical processes which subsequently lead to electric breakdown of the atmosphere. But, in reality, molecular ions are not connected with the creation of electric fields in the atmosphere, while the beginning of electrical processes follow from collisions of aerosols in different aggregate states, meaning that water aerosols may be found in liquid, solid (ice), and amorphous (snow) states, i.e., consist of snow as graupel. On the other hand, in accordance with its chemical structure, water is an electrolyte which contains ions H_3O^+ and OH^-. Upon contact of two aerosols in different aggregate states, the transition of ions is possible through the interface, and after aerosol separation one of those may contain an excess of H_3O^+ ions, whereas another aerosol may contain an excess of OH^-. The charging of aerosols as a result of their collisions proceeds in this manner.

The above character of charge creation in the atmosphere was demonstrated in the experiment by Reynolds et al. [3] in 1957, which, in the author's opinion, cannot be underestimated. One can expect that in the past one could precisely determine which charge is established in aerosols depending on their aggregate states, size, temperature, and collision velocity. Unfortunately, we have now only fragmentary information on this process. This probably results from the sensitivity of the character and rate of this process to external conditions. Indeed, an attempt to repeat this experiment in 1961 [74, 75] gave low charging rates which are several orders of magnitude below those from previous experiment [3]. Evidently, the large difference in results follows from different experimental conditions [7], so that aerosol growth takes place in the first case, while in the second case, on the contrary, water molecules evaporate from the aerosol surface. Because of the high sensitivity of the aerosol charging rate to the aerosol temperature and size, we will not analyze experiments of this type. One more peculiarity of this process lies in the fact that the Coulomb interaction energy of charged aerosols significantly exceeds their thermal energy. Therefore, if aerosols do not acquire a high speed in the experiment, their charge in air is greatly below the optimal one.

Thus, we assume below that aerosols in different aggregate states can coexist in atmospheric regions with a temperature of about 0 °C, and they get a charge as a result of collisions. Aerosol charging is an important stage of atmospheric electricity, and the coexistence of aerosols in different aggregate states is necessary for this. From this it follows that thunderstorms are absent in winter at middle latitudes, and the polar regions do not contribute to Earth's charging. Subsequently, charged aerosols are included in cumulus clouds, which are a source of electric breakdown in the atmosphere.

Further charge separation proceeds due to a gravity fall of aerosols in the atmosphere. In considering this process, we assume that aerosols of different signs have different masses, so that the fall velocities for positively and negatively charged aerosols are different. Moreover, since the Earth is charged negatively, one can assume that liquid aerosols are a constituent of clouds. Since solid aerosols fall more slowly, one can assume for simplicity that they remain in the charging region. This charging behavior leads to the dipole structure of a cumulus cloud.

Of course, this assumption for the charge structure of a cumulus cloud is rough, as follows from Fig. 1. One can add to this that air convective motion leads to mixing of atmospheric layers, which in turn gives rise to charge redistribution over the altitude. Nevertheless, since convective motion of air has a random character, this mechanism of charge separation is conserved in the convective motion of air with aerosols.

The above peculiarities of charge separation in the atmosphere allow the scale of aerosol sizes in this process to be determined [62]. This process proceeds in an electric field which is created by both the charged surface of the Earth and a charged layer of a cloud being formed. In this process, the aerosol weight exceeds the force acting on an aerosol from an electric field. Assuming an aerosol to be singly charged, we determine that the force from an electric field is proportional to an aerosol radius, whereas its weight is proportional to the radius cubed, i.e., small aerosols move up and this mechanism does not work for them. The lower limit for a radius of such aerosols is equal to 0.4 μm [62]. In particular, from this it follows that molecular ions cannot partake in the above mechanism of charge separation. Next, the efficiency of the charge transport, i.e., the ratio of the water aerosol charge to the mass of transfer, drops as the radius of the aerosol increases. Then, the maximum aerosol size ranges 20–30 μm. For this size, the ratio of the aerosol charge to mass corresponds to the ratio of the average charge transferred to the Earth's surface to the mass of transferred water per unit time. We will be guided below by a typical aerosol radius $r = 8-10$ μm, which is found in the indicated range from 0.4 μm to $20-30$ μm.

A cloud with a charged lower layer results from the above processes. The ensuing process leads to partial discharging of this layer, which consists in the charge transfer to the Earth's surface. This occurs in an electric breakdown to earth in the form of lightning. Note that the lower charged layer of a cumulus cloud contains charged aerosols, i.e., dielectric matter whose charge is determined by aerosols. Charge transfer to Earth does not mean that these charges go to the Earth. In order to model a partial transfer of a cumulus cloud charge to the Earth's surface, let us join earth and the cloud by a metal rod. Since a lower charged layer creates an electric field, an electric current will flow under the action of this field and transfer a positive charge up, while a negative charge will go to earth and proceed along the rod under the action of an electric field. The electric current flows until an electric field exists in the rod region. In this manner, a subsequent neutralization of positive ions from the rod and a negative charge of aerosols leads to a partial charge transfer to earth.

The above model describes the mechanism of charge transfer from aerosols of a cumulus cloud to earth, where the analog of a metal rod is a lightning conductive channel, and the formation of this channel is a specific stage of the creation and propagation of lightning in atmospheric air. Lightning as electric breakdown of the atmosphere is a multistage process [32–35], and its first stage concerns the propagation of a stepwise leader as was described in Section 2.2, which prepares the conductive channel and is an ionization wave [66–69]. Leader propagation is a complex physical process which provides a high electric field strength near the front of this ionization wave at a sufficiently lower average electric field strength. As a result, a channel with a partially ionized air is prepared. When a conducting channel has been created, an electric current passes through it. The channel lifetime resulting from the passage of an electric current of an individual lightning flash is determined by the heating of air in the conductive channel and its expansion. When a separate flash finishes, redistribution of lightning charges takes place, and the next charge portion is possibly transferred after some time along the conductive channel if it did not decay to completion. This schematic description exhibits the creation of an electric field in the atmosphere and propagation of an electric current through it proceed.

Experimental study of lightning as an electric breakdown of the atmosphere [32–35] allowed one to construct a theoretical lightning model comprising the analysis of some aspects of this phenomenon [68, 69]. Unfortunately, this model is mainly based on electrotechnical properties of this phenomenon and uses to a small degree the physical processes in ionized air. This model ignores the mechanism of charge transport at the lightning front due to the radiation of excited nitrogen molecules in the wavelength range $980\ \text{Å} < \lambda < 1025\ \text{Å}$ [76], since the threshold wavelength for ionization of oxygen molecules, 1025 Å [77], exceeds the threshold wavelength of 980 Å [78] for ionization of nitrogen molecules. The cross section of the oxygen molecule photoionization is $\sigma_{\text{ph}} = 0.035\ \text{cm}^{-1}\,\text{Torr}^{-1} \approx 1 \times 10^{-18}\ \text{cm}^2$ [79]. This corresponds to the photon mean free path of about 2 mm in atmospheric air that determines the character of streamer propagation in air. Next, it is argued in the lightning model [68, 69] that air in the lightning leader must be heated to temperatures of about 5000–6000 K in order to destruct negative ions in the leader channel. But in accordance with negative ion parameters, as well as with the rates of processes involving negative ions, this temperature must be 4–5 times less than the indicated one. Let us mark other principal lightning peculiarities. Each occurrence of lightning contains several flashes during which a large electric current moves along the conductive channel. Evidently, if we model lightning as the electric breakdown of air located between capacitor plates, the plate discharge occurs in one pulse. We also note that the lightning leader is labelled stepwise, because it stops after passing some distance and then continues on its way. The above facts testify to the complex nature of lightning and do not find an explanation in model [68, 69].

Aiming to represent a logically close physical picture of electrical processes in the Earth's atmosphere, we focus our attention on the most probable scenario of development of atmospheric electric phenomena and simplify each stage of subsequent processes of atmospheric electricity. As a result, we leave aside other possible scenarios of atmospheric electricity, among which are the evolution of a positively charged cumulus cloud, electric breakdown between clouds, propagation of lightning along forked lightning channels, electric breakdown between a cloud and ionosphere, and sprites. In addition, the basic physical models of the evolution of atmospheric electricity are simplified, and some of these will be considered below. Nevertheless, the simple scheme under consideration allows one to follow the whole chain of electric atmospheric processes and gives a physical picture of atmospheric electricity.

3. Atmospheric phenomena involving aerosols

3.1 Aerosol association in the Earth's atmosphere

The key processes of the general chain of the evolution of atmospheric electrical processes were given in Section 2 (see Fig. 2). They include the formation processes of charged aerosols as a result of collisions of neutral aerosols in different aggregate states, the creation of electric fields in the atmosphere due to different velocities of gravity fall for positive and negative aerosols, and electric breakdown of the atmosphere. We consider these processes in detail below. Aerosols are formed from evaporated water at certain altitudes of the atmosphere with a not high temperature, where the partial water pressure exceeds the saturated vapor pressure. Though water evaporates from the Earth's surface in the form of free molecules, at altitudes with low atmospheric temperatures water forms aerosols, i.e., liquid, snow, or ice micron-sized particles. Since processes involving aerosols determine the character of charge formation and its separation in the atmosphere, which lead subsequently to atmospheric electric breakdown in the form of lightning, below we will study the processes of aerosol growth in the Earth's atmosphere.

We base our consideration of the character of aerosol formation and growth on the general principles of nucleation processes in gases [80]. Two mechanisms of the growth of liquid aerosols in the atmosphere are given in Fig. 6 [62, 81]. The first one is coagulation, where two liquid droplets are joined after contact, and the second nucleation mechanism is coalescence or Ostwald ripening [82, 83] and results from aerosol equilibrium with its vapor, so that processes of molecule attachment to aerosols and evaporation of molecules from the aerosol surface lead to the growth of large aerosols and vaporization of small ones. As a result, this gives a continuous increase of the average aerosol size.

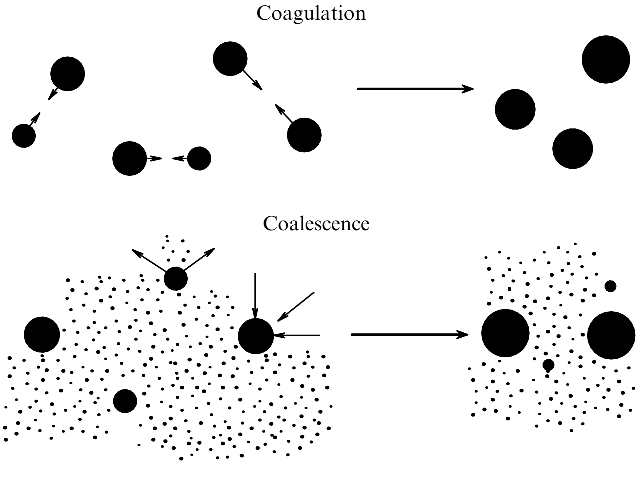

Figure 6. Mechanisms of aerosol growth in atmospheric air. Coagulation results from association of two aerosols in one drop, and coalescence corresponds to attachment of individual molecules to large aerosols and evaporation of small aerosols [62, 81].

In considering nucleation processes involving aerosols, we are guided by aerosols in a cumulus cloud which partakes in atmospheric electric phenomena and finally lead to the origin of lightning. A mature cumulus cloud contains aerosols with the following averaged parameters [37, 84–86]:

$$r = 8\ \mu\text{m}, \quad N_{\text{p}} = 10^3\ \text{cm}^{-3}, \tag{3.1}$$

where r is the average aerosol radius, and N_{p} is the average number density of aerosols in the cloud. Note that the average density of water in the atmosphere equals 2 g m^{-3}, which under standard conditions (temperature 0 °C and pressure 1 atm) corresponds to the specific water mass of 1.6 g kg^{-1}. If these cloud aerosols are placed on the Earth's surface and transformed there in liquid water, the ratio of the thickness h of the water layer to the cloud thickness L is $h/L = 2$ mm km^{-1}. It is convenient to introduce a number of molecules n in a spherical aerosol according to formula

$$n = \left(\frac{r}{r_{\text{W}}}\right)^3, \tag{3.2}$$

where r_{W} is the Wigner-Seitz radius [87, 88], which is equal to $r_{\text{W}} = 1.92$ Å in the case of water [89]. Correspondingly, a number of molecules in a typical cumulus cloud (3.1) is $n = 7 \times 10^{13}$.

The diffusion mechanism of aerosol growth is realized under the given conditions in accordance with the criterion

$$r \gg \lambda. \tag{3.3}$$

Here, λ is the mean free path of air molecules in air that is equal to $\approx 0.1\ \mu$m [90] under atmospheric pressure, and criterion (3.3) for the diffusion mechanism of aerosol growth holds true. The rate constant of association of aerosols with radii r_1 and r_2 is [89]

$$k_{\text{as}} = 4\pi(D_1 + D_2)(r_1 + r_2),$$

where D_1 and D_2 are the diffusion coefficients for colliding aerosols in air. Since the diffusion coefficient of aerosols in air is proportional to their radius under given conditions, the rate constant of aerosol association is independent of their radius, if these radii are of the same order of magnitude. Therefore, averaged over aerosol sizes, the rate constant of aerosol association is expressed as [62]

$$k_{\text{as}} = \frac{8T}{3\eta}, \tag{3.4}$$

where η is the viscosity coefficient of air. Formula (3.4) gives $k_{\text{as}} = 5.8 \times 10^{-10}$ cm^3 s^{-1} at room temperature.

The equation describing the evolution of the average aerosol size $\bar{n}$ for this growth mechanism is given by [62]

$$\frac{\mathrm{d}\bar{n}}{\mathrm{d}t} = \frac{1}{2}\,k_{\text{as}}N_{\text{b}} = \frac{1}{2}\,k_{\text{as}}N_{\text{p}}\bar{n}, \tag{3.5}$$

where $N_{\text{b}} = N_{\text{p}}\bar{n}$ is the total number density of molecules in aerosols, and N_{p} is the aerosol number density. For the parameters of a typical cumulus cloud (3.1), this quantity is $N_{\text{b}} = 7 \times 10^{16}$ cm^{-3}. The solution of equation (3.5), namely

$$\bar{n} = \frac{1}{2}\,k_{\text{as}}N_{\text{b}}t, \tag{3.6}$$

shows an increase in the average aerosol size with time. Let us introduce the aerosol doubling time τ_{d} during which the number of aerosol molecules is doubled, and the average aerosol radius becomes $2^{1/3}r$ with respect to the initial average aerosol radius r. Then, the aerosol number density reduces to $N_{\text{p}}/2$. In the case under consideration, the doubling time is

$$\tau_{\text{d}} = \frac{2\bar{n}}{k_{\text{as}}N_{\text{b}}} = \frac{2}{k_{\text{as}}N_{\text{p}}}. \tag{3.7}$$

Assuming that a cumulus cloud contains neutral aerosols, we find the doubling time for an aerosol size of a typical cumulus cloud: $\tau_{\text{d}} = 2/(k_{\text{as}}N_{\text{p}}) = 40$ days. Since this time significantly exceeds the lifetime of a cumulus cloud, one can conclude that the diffusion mechanism of aerosol coagulation growth is not of interest for the evolution of aerosols in a cumulus cloud.

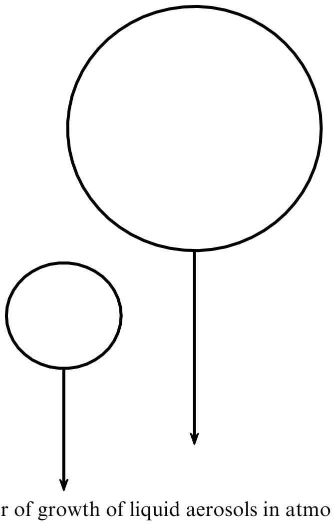

Figure 7. Character of growth of liquid aerosols in atmospheric air under the action of the Earth's gravitation field. Heavier droplets overtake lighter ones and join with them in one droplet as a result of their contact [81, 173].

Another mechanism of aerosol association is given in Fig. 7 and consists of joining two colliding aerosols at their contact if this collision is determined by a different velocity of aerosol fall in the gravitation field of the Earth. The rate constant of aerosol association is estimated as

$$k_{\text{as}} \sim \Delta w\,\sigma \sim \frac{\rho g r^4}{\eta},$$

where Δw is the difference between the velocities of fall of two aerosols in atmospheric air, $\sigma \sim r^2$ is a typical area of the projection of contacted aerosols on the Earth's surface, r is a typical aerosol radius, ρ is the density of an aerosol material (for water aerosols $\rho = 1$ g cm^{-3}), g is the acceleration of free fall, and η is the air viscosity. The velocity of free fall for a micron-sized spherical particle in the Earth's gravitation field is [89]

$$w = \frac{2\rho g r^2}{9\eta}. \tag{3.8}$$

Averaging over sizes and taking into account the size distribution function of aerosols, one can obtain for the rate constant of aerosol growth for this association mechanism the following expression [81]

$$k_{\text{as}} = \frac{2r^4\rho g}{\eta}. \tag{3.9}$$

Comparing the rate constants of aerosol association in accordance with formulas (3.4) and (3.9), one can conclude that the rates of coagulation and gravitation association are comparable at $r \approx 1$ μm, so that the gravitation mechanism of association dominates at large aerosol sizes. In particular, for an average aerosol radius $r = 8$ μm typical for a cumulus cloud, we have according to formula (3.9) $k_{\mathrm{as}} = 4.3 \times 10^{-6}$ cm^3 s^{-1}. Assuming aerosols to be neutral, we obtain the lifetime $\tau \approx 8$ min of a cumulus cloud with respect to the doubling of an aerosol size, which is less significantly than the lifetime of a cumulus cloud measured by hours. From this it follows that aerosols of a cumulus cloud are likely charged, and this prevents aerosols from association. Assuming aerosol charges Z to be identical in a cumulus cloud, we find the rate constant of aerosol association in the form [81]

$$k_{\mathrm{as}} = \frac{2r^4 \rho g}{\eta} \exp\left(-\frac{Z^2 e^2}{2rT}\right) \qquad (3.10)$$

instead of formula (3.9) for the association of neutral aerosols. In this case, electrostatic repulsion of charged aerosols prevents them from the contact. One can represent formula (3.10) for liquid water aerosols in atmospheric air as

$$k_{\mathrm{as}} = r^4 \exp\left(-\frac{0.06 Z^2}{r}\right), \qquad (3.11)$$

where the rate constant k_{as} of aerosol association is measured in the units of 10^{-9} cm^3 s^{-1}, and the average aerosol radius is expressed in μm. Note that an aerosol charge $Z \approx 27$ creates a weak field on its surface that, under given conditions, is $E = Ze/r^2 \approx 6$ V cm^{-1}, which corresponds to the electric potential of the aerosol with respect to surrounding air: $U = Ze/r \approx 5$ mV.

Let us determine the aerosol doubling time in a cumulus cloud for the gravitation association mechanism (Fig. 7), where the rate constant of aerosol association is given by formula (3.10). Let us assume the number density N_{b} of bound water molecules to be constant in time, as is the aerosol charge Z. We then obtain the following doubling time for the number of aerosol molecules on the basis of formula (3.10) for the rate constant of aerosol association:

$$\tau_{\mathrm{d}} = \int_{n_1}^{n_2} \frac{2\,\mathrm{d}n}{k_{\mathrm{as}} N_{\mathrm{b}}} = \frac{3\eta}{\rho g N_{\mathrm{b}} r_{\mathrm{W}}^3} \int_{r_2}^{r_1} \mathrm{d}\left(\frac{1}{r}\right) \exp\left(\frac{Z^2}{2rT}\right) = \tau_0 F\left(\frac{Z^2}{2rT}\right), \qquad (3.12)$$

where n_1, n_2 and r_1, r_2 are the initial and final average numbers of aerosol molecules and their average radii, respectively. The parameters of the final state are given by

$$\tau_0 = \frac{3\eta}{\rho g N_{\mathrm{b}} r_{\mathrm{W}}^3 r}, \quad F(x) = x\left[\exp(x) - \exp\left(\frac{x}{2^{1/3}}\right)\right]. \qquad (3.13)$$

For typical parameters of aerosols (3.1) in a cumulus cloud, we have for a doubling time $\tau_0 = 24$ min, and the function $F(x)$ is plotted in Fig. 8. If we take the typical lifetime of a cumulus cloud to be between 3 and 12 hours, for the corresponding aerosol charge one can obtain on the basis of formula (3.12):

$$Z = 27 \pm 2\,. \qquad (3.14)$$

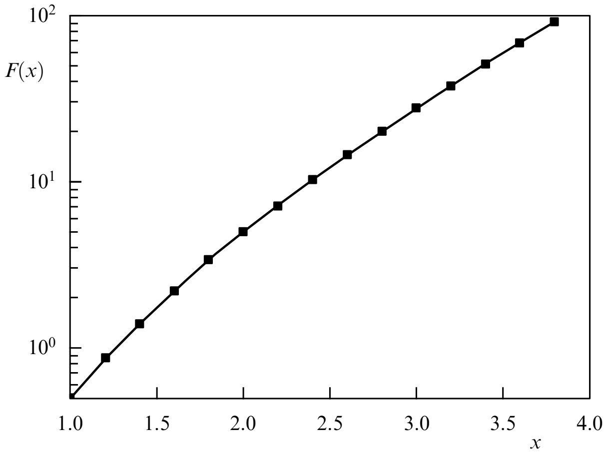

Figure 8. Function $F(x)$ plotted in accordance with formula (3.13).

Thus, the aerosol charge weakly depends on the typical time of aerosol association, and the repulsion electrostatic potential for aerosols with a charge $Z = 27$ and radius $r = 8$ μm is $Z^2 e^2/(2r) = 760$ K, which exceeds a thermal aerosol energy. At room temperature, in particular, we have for limiting values of the aerosol charge according to formula (3.14): $Z^2 e^2/(2rT) = 2.6$ and $\exp\left[Z^2 e^2/(2rT)\right] = 13$. The results presented in Fig. 9 confirm the weak dependence of the aerosol charge Z on growth time, and an increase in the cloud lifetime by an order of magnitude leads to an increase in the aerosol charge by 15%, which provides this growth time. This fact characterizes the reliability of determining the aerosol charge and, together with observational parameters of cloud evolution, allows us to analyze the aerosol stage of water circulation in the atmosphere.

Let us return to the concept for electricity transport in the atmosphere being a secondary phenomenon with respect to water transport, and the ratio of the transferred charge to the mass of transferred water equal to 1.4×10^{-10} C g^{-1}. If the channel of water transport in the atmospheric water circulation corresponds to the transport of charged aerosols of a cumulus cloud, we obtain the following value expressed in the units of C g^{-1} for the ratio of a transferred aerosol charge Ze

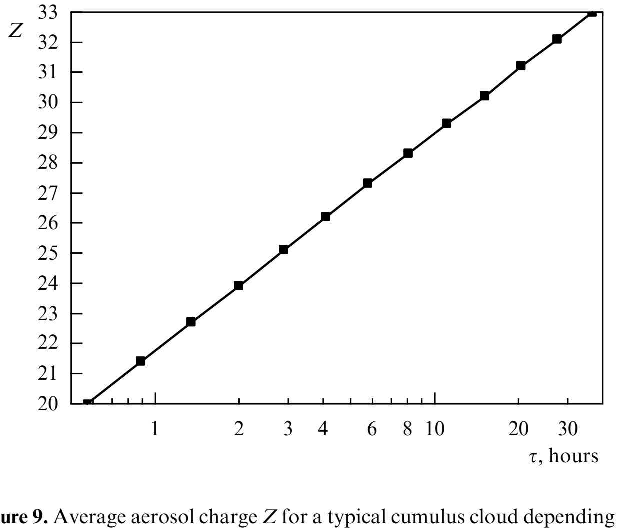

Figure 9. Average aerosol charge Z for a typical cumulus cloud depending on the doubling time for a number of water molecules in an aerosol [91].

to water mass m in aerosols for this channel:

$$\frac{Ze}{m} = 2 \times 10^{-9}\,, \tag{3.15}$$

which is several times larger than that for the atmosphere as a whole. From this it follows that in order to provide the observed amount of charge transferred to the Earth's surface, it is necessary that several percent of the water evaporated in the atmosphere be connected with the formation and growth of aerosol particles. This result also confirms the possibility of transporting the charge from the atmosphere to the Earth's surface through aerosols.

Let us make one more estimate on the basis of the results obtained. One can estimate a typical strength of the electric field created by a lower edge of a cumulus cloud, which determines the cloud charge per unit area of the cloud, On the other hand, formulas (3.1) and (3.14) give the charge number density for a cloud. The combination of these quantities allows one to determine the thickness of the cloud lower edge where the charge is mainly located. Indeed, based on the observed values of the electric voltage of the lower cloud edge with respect to Earth, $U = 20-100$ MV [53], we take for estimations the average voltage $U = 60$ MV. For a typical distance $L = 3$ km of the lower cloud edge from Earth, this corresponds to the average electric field strength $E = 200$ V cm^{-1}, which gives the following charge density of the lower edge of a cumulus cloud:

$$\sigma = \frac{E}{4\pi} \approx 1 \times 10^8 \, \frac{e}{\text{cm}^2} \approx 2 \times 10^{-11} \ \text{C}\,\text{cm}^{-2}\,. \tag{3.16}$$

Referring to formulas (3.1) and (3.14), the charge density inside the cloud is $ZeN_{\text{p}} = 4 \times 10^{-15}$ C cm^{-3}. Note that the charge density in a cumulus cloud according to airplane measurements [92, 93] is on the order of 1 C km^{-3} or 1×10^{-15} C cm^{-3}. From the ratio of the surface charge density to its volume density, we obtain $l \sim 100$ m for the thickness of a charged cloud layer [81], which is much less than the cloud size.

Taking further the charge of the lower cloud edge to be equal to the charge of typical lightning $Q = 20$ C [32, 35, 53], one can determine an area from which this charge is collected:

$$S = \frac{Q}{\sigma} \sim 100 \ \text{km}^2\,, \tag{3.17}$$

which corresponds to observational data [59]. Note the threshold character of thunderstorm phenomena, so that the minimum size of a thunderstorm cloud is $L \approx 3$ km [53]. Note that as follows from the above estimation the charge transferred by lightning is collected from a large area of the lower cloud edge. The cloud is characterized by a low conductivity and has no capacity to provide charge transport to the lightning conductive channel during the time allotted for the basic current passage. This means that lightning development due to the creation of a conductive channel from the lower cloud edge to the Earth's surface is accompanied by a change in the electrical properties of the lower cloud edge as a result of passing the positive charge from the lightning channel to it.

3.2 Coalescence in aerosol plasma

As follows from the above analysis, the rate of association for likely charged aerosol droplets sharply decreases with increasing charge. For another mechanism of aerosol growth, coalescence or Ostwald ripening [82, 83], this process is independent of the aerosol charge. In this case, an ensemble of aerosols is located in a buffer gas, and the character of aerosol growth is determined by processes of aerosol evaporation and the attachment of atoms or molecules from a vapor found in equilibrium with aerosols to their surfaces. Such a system consisting of aerosols and their vapor departs from equilibrium, because the equilibrium between aerosols and vapor ultimately leads to an increase in the average aerosol size due to the growth of large aerosols and the vaporization of small aerosols. As a result, the average aerosol size increases with time.

The automodel character of the size distribution function for the coalescence process [94, 96, 97] is the principal peculiarity of the process considered. This means that the size distribution function of aerosols depends on a dimensionless parameter — that is, the ratio of an aerosol size to its critical value — and does not depend on time directly, whereas the aerosol critical radius increases with time. The classical theory of coalescence [95, 98] relates to the diffusion mechanism of this process, while the above fundamental property also corresponds to the kinetic regime of coalescence [81, 91]. These two regimes differ by the concentration of molecules of the parent vapor in a buffer gas, and the approach of an attaching molecule and aerosol in the diffusion regime proceeds as a result of diffusion in a buffer gas, while in the kinetic regime of coalescence the molecule approach occurs from a straightforward molecular trajectory. Below, we consider the diffusion mechanism of coalescence being guided by growth processes for water drops in air.

According to the automodel character of aerosol growth, the size distribution function of aerosols in the form $f(u)$ does not vary in time (Fig. 10). Here, $u = n/n_{\text{cr}}$, n is the current number of aerosol molecules, and n_{cr} is the critical number of aerosol molecules at which the rates of molecular attachment and evaporation of aerosols per unit time are equal. In analyzing the equilibrium between free and bound water molecules, we use an expansion of the total binding energy E_n of bound molecules in an aerosol consisting of n molecules in terms of a small parameter $n^{-1/3}$ that has the form [99]

$$E_n = \varepsilon_0 n - An^{2/3}\,. \tag{3.18}$$

Here, ε_0 is the average binding energy per water molecule in a macroscopic droplet, A is the specific surface energy, and these parameters for water are equal on the basis of handbook [73] to $\varepsilon_0 = 0.43$ eV, $A = 0.21$ eV (these parameters relate to the liquid aggregate state of aerosols at room temperature).

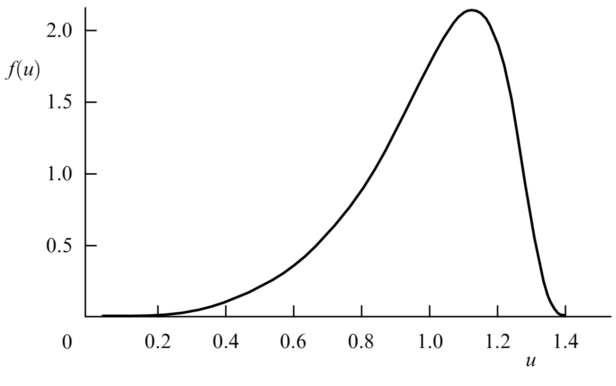

Figure 10. Size distribution function $f(u)$ of aerosols for the diffusion mechanism of coagulation in accordance with formula (3.3) [94, 95].

From this we obtain the binding energy change $\Delta\varepsilon(n)$ for a water molecules in an aerosol consisting of n bound molecules compared to the limit of a large droplet:

$$\Delta\varepsilon(n) = \frac{\mathrm{d}E_n}{\mathrm{d}n} - \varepsilon_0 = \frac{2A}{3n^{1/3}}\,. \tag{3.19}$$

Let us consider the equilibrium between aerosols and a vapor of free molecules for large aerosols according to the criterion $A \ll Tn^{1/3}$. Taking into account that the equilibrium between the number densities N_w of free molecules and the number density $N_\mathrm{sat}(T)$ in a saturated vapor for large aerosols has the form $N_\mathrm{w} = N_\mathrm{sat}(T)$, we use the temperature dependence for the molecule number density in the saturated vapor in the form $N_\mathrm{sat}(T) \sim \exp(-\varepsilon_0/T)$, where ε_0 is the molecule binding energy for bulk water. From this, we have for the critical aerosol size the following expression [89]

$$N_\mathrm{w} = N_\mathrm{sat}(T)\exp\left[\frac{\Delta\varepsilon(n_\mathrm{cr})}{T}\right]. \tag{3.20}$$

Assuming for simplicity that each contact of a free molecule with the aerosol surface leads to the molecule attachment, we obtain on the basis of the Smoluchowski formula [100] for the number of molecules attached to an aerosol per unit time:

$$J_\mathrm{at} = 4\pi D r N_\mathrm{w}\,, \tag{3.21}$$

where D is the diffusion coefficient for water molecules in air. Let us take into consideration that the total change in the number of bound molecules for an aerosol of a given size per unit time is

$$J = J_\mathrm{ev} - J_\mathrm{at} = J_0 j(u)\,, \quad J_0 = \frac{8\pi A r_\mathrm{W} D N_\mathrm{m}}{3T}\,, \qquad u = \left(\frac{n}{n_\mathrm{cr}}\right)^{1/3}, \quad n < n_\mathrm{cr}\,, \tag{3.22}$$

where $J_\mathrm{ev}(n,T)$ is the rate of evaporation, and N_m is the number density of free molecules. This formula relates to aerosols whose size is below the critical one. Figure 11 depicts the partial flux of molecules falling on the aerosol surface as a function of the parameter u:

$$j(u) = (1-u)f(u)\,, \tag{3.23}$$

and it follows at the equilibrium that the numbers of evaporated and attached molecules are equal, i.e.

$$\int_0^1 (1-u)f(u)\,\mathrm{d}u = \int_1^{1.5}(u-1)f(u)\,\mathrm{d}u\,. \tag{3.24}$$

On the basis of the data given in Fig. 11, we obtain the total rate of molecule release from aerosols whose size is below the critical one or the total growth rate of aerosols whose size exceeds the critical one:

$$J = J_0\int_0^1 j(u)\,\mathrm{d}u = -J_0\int_1^{1.5} j(u)\,\mathrm{d}u = 0.085J_0\,. \tag{3.25}$$

From this, the equation follows for an increase in the average aerosol size $\bar{n}$ that is connected with the critical aerosol size by the relation $\bar{n} = 1.13n_\mathrm{cr}$:

$$\frac{\mathrm{d}\bar{n}}{\mathrm{d}t} = \frac{\bar{n}}{n_\mathrm{cr}}\,\frac{J}{f(1)} = 0.056J_0\,. \tag{3.26}$$

This gives for the doubling time of an aerosol size, when the average number of aerosol molecules varies from $\bar{n}$ to $2\bar{n}$ [81, 91]:

$$\tau_\mathrm{d} = \frac{36\bar{n}}{J_0} = \frac{4.3\bar{n}}{D r_\mathrm{W} N_\mathrm{m}}\,\frac{T}{A}\,. \tag{3.27}$$

It should be noted that molecule evaporation from the aerosol surface or molecule attachment to the aerosol surface are accompanied by heat processes that create a temperature difference between the aerosol and surrounding air, and this in turn influences the total rate I of aerosol evaporation or molecule attachment to aerosols. This effect is characterized by the parameter Φ, which is the ratio of the partial growth rates of aerosols with sizes exceeding critical ones without accounting for the heat effect (J) and with its account (I):

$$\Phi = \frac{J}{I}\,, \tag{3.28}$$

where J is given by formulas (3.25) and (3.22). The factor Φ for the growth of water aerosols in air is represented in Fig. 12 using the parameters of water and its vapor [73], and the diffusion coefficient of water molecules in air at the temperature $T = 300$ K and atmospheric pressure is given as $D = 0.2\ \mathrm{cm^2\,s^{-1}}$ [101]. The above factor influences the doubling time for the aerosol size, increasing it by the factor Φ compared to that according to formula (3.27). Figure 13 depicts the temperature dependence for the doubling time of aerosol size in a cumulus cloud, which is given according to

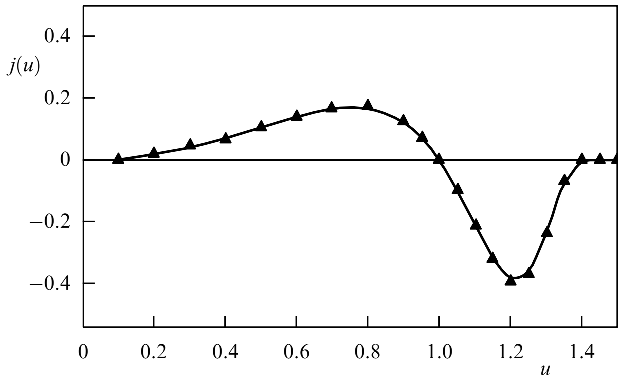

Figure 11. Partial rate $j(u)$ of molecule attachment to aerosols [94, 95].

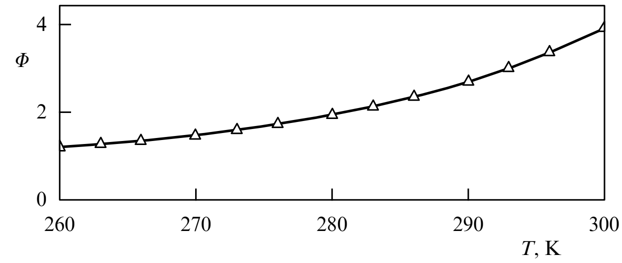

Figure 12. Thermal factor $\Phi(T)$ characterizing the rate of growth of water aerosols whose size exceeds the critical one. Aerosols are found at equilibrium with the water vapor in atmospheric air, and aerosol growth proceeds in the diffusion regime of aerosol growth as a result of coalescence [81, 91].

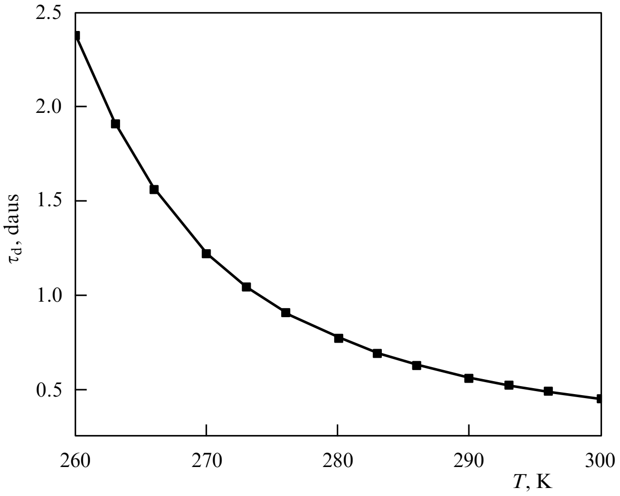

Figure 13. Doubling time for an average number of aerosol molecules in aerosols of a cumulus cloud with parameters (3.1) in atmospheric air if aerosol growth is determined by the diffusion regime of coalescence [81, 91].

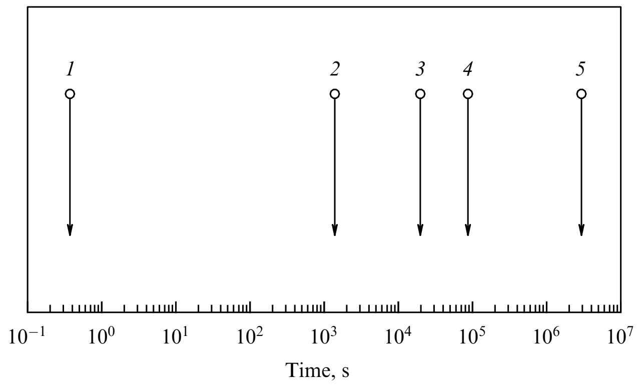

Figure 14. Hierarchy of typical times in a system consisting of water aerosols and water vapor located in atmospheric air at temperature $T = 0\,^\circ$C and a pressure of 1 atm, if typical aerosol parameters correspond to a cumulus cloud (3.1): *1* — equilibrium between free water molecules and aerosols, *2* — gravitation growth of neutral aerosols, *3* — gravitation growth of charged aerosols, *4* — coalescence, and *5* — coagulation of neutral aerosols.

formula (3.27) by

$$\tau_d = \frac{4.3\bar{n}}{\Phi D r_W N_m} \frac{T}{A} \,. \tag{3.29}$$

In analyzing the coalescence process, we have assumed that an equilibrium between the aerosols and vapor is established fast compared to the coalescence time. Let us determine the time of establishing this equilibrium in the limit of large aerosols, where the equilibrium number density N_0 of molecules near an aerosol is equal to their number density N_{sat} in the saturated vapor. For this purpose we introduce the number density N_b of bound water molecules in an aerosol:

$$N_b = N_p \left(\frac{r}{r_W} \right)^3 , \tag{3.30}$$

where N_p is the aerosol number density, and r is the average aerosol radius. Assuming $N_b \gg N_{sat}$, from the equilibrium condition we obtain $\delta N_{sat} = -\delta N_b$. Therefore, violation of the equilibrium does not influence the aerosol parameters. Vapor relaxation is thus described by the equation

$$\frac{dN_w}{dt} = 4\pi D r (N_{sat} - N_w) \,, \tag{3.31}$$

and the solution of this equation gives for the number density of free molecules the following expression

$$N_w = N_{sat} - (N_{sat} - N_0) \exp\left(-\frac{t}{\tau} \right) . \tag{3.32}$$

As is seen, the relaxation time or time of establishment of the equilibrium between free molecules and aerosols in atmospheric air is described as

$$\tau = \frac{1}{4\pi D r N_p} \,. \tag{3.33}$$

This time depends weakly on the temperature of air where aerosols are located, and in the above consideration it must be much lower compared to a growth time for aerosols. In particular, according to formula (3.33) and for parameters of a typical cumulus cloud (3.1) the time of equilibrium establishment ranges from 0.5 to 0.4 s at atmospheric pressure and the air temperature in the range 260–300 K.

Figure 14 displays typical times of the processes of aerosol growth under consideration and a time of equilibrium establishment between aerosols and saturated water vapor in atmospheric air for parameters of a typical cumulus cloud (3.1) and an air temperature of $0\,^\circ$C. Note that the number densities of free and bound water molecules are comparable under these conditions. Thus, at an air pressure of 0.5 atm, which corresponds to an altitude of approximately 6 km, the specific water content in aerosols equals 3.3 g kg^{-1}, while the content of free water molecules in the saturated vapor amounts to 7.5 g kg^{-1}.

3.3 Processes in clouds

Basic atmospheric electrical processes proceed in clouds and involve aerosols. Some examples of clouds are given in Figs 15, 16, and they clearly demonstrate a nonuniform structure of clouds with mixing of their elements. Cumulus clouds are of interest for the analysis of atmospheric electric phenomena only with typical aerosol parameters according to formula (3.1). These clouds are formed in a region of

Figure 15. View of clouds from an airplane [102].

Figure 16. Clouds lower than the mountains [103].

condensation of a water vapor up to altitudes of 10 km (see Fig. 1).

Since water holds fundamental significance for atmospheric electricity, and solid and liquid water microparticles are responsible for charge separation in clouds, let us analyze in detail atmospheric processes involving aerosols. The dry air mass reaches 5.1×10^{18} kg in the Earth's atmosphere, its specific weight near the air surface is 1.2 kg m^{-3}, on average, and the average water mass in the atmosphere amounts to 1.3×10^{16} kg [70]. This corresponds to an average specific weight of atmospheric water of 3 g m^{-3} and practically coincides with the specific water weight in aerosols if one takes into account a decrease in this quantity with increasing altitude above the ground. This corresponds to a partial water pressure of about 2 Torr, whereas even at the temperature 0 °C the saturated vapor pressure is 4.7 Torr [71]. One can conclude from this that water is found in the atmosphere mostly in the form of free molecules, and condensation of a water vapor proceeds in regions with a heightened water pressure and low temperature.

As was noted in Section 2.3, the atmosphere temperature decreases with increasing altitude up to the tropopause, i.e., the atmospheric layer with a minimal temperature. For an average atmosphere, the tropopause is found at an altitude of 9 km for the poles, and 17 km for the equator [70]. The average atmospheric temperature on the Earth's surface is 15 °C, the temperature of tropopause equals −55 °C, and the temperature gradient reaches -6.5 K km^{-1} [72]. On the basis of these numerical estimates, one can introduce a characteristic size

$$\Lambda = \left| \frac{\mathrm{d}T}{T\,\mathrm{d}h} \right|^{-1}, \tag{3.34}$$

which takes the value $\approx 2 \times 10^6$ cm.

As follows from the above discussion, water condensation with the formation of water aerosols proceeds at altitudes of about 3–10 km over the Earth's surface [104]. At such altitudes, micron-sized liquid aerosols and solid aerosols — ice particles — may be formed simultaneously, and this is necessary for charge formation and separation in the atmosphere. Because of the large atmosphere scales and high atmospheric pressure, convective motion is of importance for both the transport of air and water and the formation and development of atmospheric electric phenomena, significantly accelerating them. Let us estimate a size of an elementary vortex in the middle troposphere accounting for its typical size (3.34). We assume that convection results from Rayleigh–Taylor instability [105–109] under the action of the gravitation force and temperature gradient. If conditions of the convective instability are fulfilled in a region between two heated plates, structures like Benard cells are formed [110–112]. In the transition to a uniform infinite system these structures are mixed and lead to a random convective air motion in the form of vortices of different sizes. Let us determine the minimal vortex size on the basis of the criterion establishing that the Rayleigh–Taylor instability arises and develops if the Rayleigh number is $\mathrm{Ra} \sim 1000$ [105, 106, 113].

Let us represent the Rayleigh number for a vortex of size l, i.e., the size of the elementary air cell, in the form [105]

$$\mathrm{Ra} = \frac{\mathrm{d}T}{T\,\mathrm{d}h}\,\frac{gl^4}{\nu\chi} = \frac{gl^4}{\Lambda\nu\chi}\,. \tag{3.35}$$

Here, $g = 980$ cm s^{-2} is the free fall acceleration on the surface of Earth, $\nu = \eta/\rho = 0.3\ \mathrm{cm}^2\,\mathrm{s}^{-1}$ is the air kinematic viscosity, $\rho = 0.6\ \mathrm{g\,m}^{-3}$ is the air density, η is the air viscosity, and we are guided by altitudes where the air pressure is $p = 0.5$ atm. Next, $\chi = \kappa/Nc_\mathrm{p} = 0.4\ \mathrm{cm}^2\,\mathrm{s}^{-1}$ is the air thermal diffusivity, so that N is the number density of air molecules, and $c_\mathrm{p} = 7/2$ is the air heat capacity per molecule. This gives the following estimate for a typical vortex size

$$l \approx \left(\frac{\mathrm{Ra}\,\nu\chi\Lambda}{g} \right)^{1/4}. \tag{3.36}$$

From this we find the minimal vortex size near Earth's surface to be $l \sim 30$ cm. This value increases with distance from the Earth's surface as $l \sim 1/\sqrt{N}$. Mass transport on large distances which significantly exceed the size of the elementary vortex is perceived as a turbulent atmospheric motion which leads to mixing of air and its components.

The velocity of motion for a vortex of size l follows from the definition of the gas viscosity:

$$\frac{\Delta P}{l^2} \sim \eta\,\frac{v_l}{l}\,, \tag{3.37}$$

where ΔP is the weight difference for two atmospheric layers of a thickness l spaced at a distance on the order of l from each other:

$$v_l \sim \frac{mgNl^3}{\Lambda\eta}\,. \tag{3.38}$$

The diffusion coefficient D_l due to transfer in vortices of size l is given by

$$D_l \approx \frac{v_l l}{3} \sim \frac{mgNl^4}{3\Lambda\eta}\,. \tag{3.39}$$

In particular, for a minimal vortex size $l \sim 30$ cm, these parameters are equal to $v_l \sim 10$ cm s^{-1}, and $D_l \sim 1 \times 10^2\ \mathrm{cm}^2\,\mathrm{s}^{-1}$. The Reynolds number for this vortex size is $\mathrm{Re} = v_l\,l\nu \sim 1 \times 10^2$. The maximum vortex size for a turbulent air motion follows from the condition for the critical Reynolds number, viz. $\mathrm{Re}_\mathrm{cr} = v_l L\nu = 2 \times 10^5$ [105], which corresponds to the origin of turbulence, and L is the maximum vortex size. We have $L/l_\mathrm{min} \approx 4$, so that vortex sizes in convective motion of atmospheric air range from l to $L \sim 1$ m. Note that the Reynolds and Rayleigh numbers have the same order of magnitude for a gas [109]. As is seen, the transport of atmospheric air and its components over long distances is realized in vortices of a maximum size, and the corresponding diffusion coefficient for air transport in the atmosphere is estimated as

$$D_L \approx \frac{\mathrm{Re}_{\mathrm{cr}}\,\eta}{3} \sim 2\times 10^4\ \mathrm{cm}^2\ \mathrm{s}^{-1}\,. \tag{3.40}$$

Thus, transport of mass and energy in atmospheric air over large distances has a random character [114–116], where collisions of air fluxes in the atmosphere lead to a turbulent character of motion for mixing air layers [117, 118].

3.4 Rain formation from clouds

The last stage in the development of a cumulus cloud with charged aerosols is rain generation under the action of layers of a cumulus cloud with the fluxes of warm and wet air, and the mixing of this air with aerosols leads simultaneously to lightning formation. It is accepted in Ref. [59] that, as a result of this mixing, water molecules attach to aerosols which grow fast and are released from the charge. The fast growth of neutral aerosols proceeds in accordance with the gravitation mechanism (see Fig. 7), leading to an increase in the velocity of their fall, and aerosols become large near earth. Below, we consider the kinetics of this process and estimate a size of falling droplets depending on external conditions.

On the basis of formulas (3.5) and (3.9), we have

$$k_{\mathrm{as}} = k_0\left(\frac{r}{a}\right)^4,$$

where $k_0 = 1.1\times 10^{-5}\ \mathrm{cm}^3\ \mathrm{s}^{-1}$, and $a = 10\ \mu\mathrm{m}$. Along with equation of aerosol growth (3.5), we use the equation of aerosol motion under the action of the gravity force:

$$\frac{\mathrm{d}h}{\mathrm{d}t} = w\,, \tag{3.41}$$

where h is the path passed by the aerosol during its fall, and w is the fall velocity given by formula (3.8). Excluding time from these equations and assuming the total number density of bound molecules N_{b} to be independent of time, we obtain

$$\frac{\mathrm{d}h}{\mathrm{d}r} = \mathrm{const} = 1.4\times 10^6\,. \tag{3.42}$$

The constant in the last equation is determined by initial conditions in accordance with formula (3.1). In particular, if a height of fall is $h = 2$ km, the final droplet radius is $r = 1.4$ mm, and a typical time of aerosol fall is $\tau_{\mathrm{f}} \approx 25$ min.

Note that according to observations [119] the radius of rain droplets near the Earth's surface is found in the range 0.1–9 mm. Formula (3.8) is violated at large drop sizes, so that the velocities of fall for droplets of radii 0.5 mm and 5 mm are $2\ \mathrm{m\ s}^{-1}$ and $9\ \mathrm{m\ s}^{-1}$, respectively, whereas according to formula (3.8) these velocities are $2\ \mathrm{m\ s}^{-1}$ and $200\ \mathrm{m\ s}^{-1}$, respectively. The reason for the discrepancy at large droplet sizes is that at large Reynolds numbers the Stokes formula (3.8) is violated, and the drop spherical shape is lost [119]. Nevertheless, the Stokes formula (3.8) may be used for an estimate. In addition, if aerosols of a cumulus cloud with parameters (3.1) are transformed into water on the Earth's surface, the thickness of the water layer $\varDelta$ is proportional to the thickness L of a cumulus cloud, and

$$\frac{\varDelta}{L} = 2\ \mathrm{mm\ km}^{-1}\,. \tag{3.43}$$

If we compare this value with the velocity of water precipitation during rain, which ranges from $2.5\ \mathrm{m\ h}^{-1}$ to $7.6\ \mathrm{m\ h}^{-1}$ [119], one may conclude that a certain contribution to this velocity comes from moisture which is found in fluxes of wet air and is mixed with a cumulus cloud [120]. Therefore, the convection of wet air and the interaction of jets of wet air with aerosols of a cumulus cloud is of importance for rain generation [121–123].

It is usually assumed that the thunderstorm weather requires a high moisture of surface air that must exceed $7\ \mathrm{g\ kg}^{-1}$ [59], which corresponds to the saturated vapor pressure at the temperature 8 °C (see Table 1). At lower temperatures humid air condenses and water aerosols are formed; according to formula (3.1), the water content in aerosols of a typical cumulus cloud is $3\ \mathrm{g\ kg}^{-1}$, and their formation leads to an increase in the air temperature by 16 K. Below, we estimate the rate of moisture conversion in rain droplets in the thunderstorm weather, starting from the aerosol parameters in a cumulus cloud (3.1) and assuming that the air penetrates fast inside a cumulus cloud. The rate constant of attachment of water molecules to an aerosol of radius r is determined by the Smoluchowski formula (3.21) [100]:

$$k_{\mathrm{at}} = 4\pi D r\,, \tag{3.44}$$

where the diffusion coefficient of water molecules in air is $D \approx 0.2\ \mathrm{cm}^2\ \mathrm{s}^{-1}$ [101], and the equation of aerosol growth has the form

$$\frac{\mathrm{d}n}{\mathrm{d}t} = k_{\mathrm{at}} N_{\mathrm{w}}\,, \tag{3.45}$$

where n is the typical number of water molecules in a droplet-aerosol, and $N_{\mathrm{w}} = 3\times 10^{17}\ \mathrm{cm}^{-3}$ is the number density of water molecules, which corresponds to the threshold content of water in air, $7\ \mathrm{g\ kg}^{-1}$, for the thunderstorm weather [59]. This gives the typical time of aerosol growth:

$$\tau = \frac{3r^2}{8\pi D N_{\mathrm{w}} r_{\mathrm{W}}^3}\,, \tag{3.46}$$

where $r = 8\ \mu\mathrm{m}$ is the aerosol radius, and $r_{\mathrm{W}} = 0.192$ nm is the Wigner–Seitz radius for water. According to formula (3.46), a typical time of aerosol growth is $\tau \approx 0.2$ s. This also shows that aerosol growth as a result of attachment of free water molecules to an aerosol proceeds faster than that for the gravitation mechanism of aerosol growth. Indeed, a comparison of the rate constants for gravitation growth (3.9) and the rate constant of attachment of free water molecules to aerosols (3.44) demonstrates that they become equal at aerosol radius $r = 0.1$ mm, while mechanism (3.44) dominates at lower sizes. In addition, the attachment of free water molecules to aerosols proceeds fast, i.e. water molecules penetrating in a cumulus cloud interior transform there into the bound molecules in aerosols, if this process is permitted by thermal conditions.

It should be noted that rain rises after a lightning discharge in the thunderstorm weather according to the version under consideration. Indeed, after a lightning flash a positive charge is led to charging of aerosols, and at the beginning it is found in the form of positive molecular ions. Collisions of positive molecular ions with negatively charged aerosols neutralize the aerosols and cause fast growth of water droplets, that leads to rain production. Evidently, convective jets of warm wet air, as a thunderstorm precursor, delay the attachment of positive molecular ions to aerosols due to thermal effects, so that they neutralize charged aerosols, and the subsequent growth of water droplets proceeds according to the above scenario. It is difficult to estimate a typical time of neutralization of negatively charged aerosols, because charge transport as a result of convective processes plays an important role. Nevertheless, because a delay of rain production after the first thunderbolt is not usually observed, one can expect that a typical time of aerosol neutralization does not exceed 10 min, i.e., a typical time of rain production.

4. Ions and electric fields in the Earth's atmosphere

4.1 Ions in terrestrial atmosphere

One of the stages of atmospheric electricity is transport of ions in the atmosphere. Electrons are usually absent in a quiet atmosphere because of attachment of electrons to oxygen molecules. For thermal electrons, such process proceeds in three body collisions with oxygen and nitrogen molecules:

$$\mathrm{e} + \mathrm{O_2} + \mathrm{O_2(N_2)} \to \mathrm{O_2^-} + \mathrm{O_2(N_2)}\,. \tag{4.1}$$

A typical time of electron attachment to oxygen molecules in atmospheric air is $\tau \approx 10^{-8}$ s at room temperature. Electrons accelerated by a strong electric field may attach to oxygen molecules as a result of the process

$$\mathrm{e} + \mathrm{O_2} \to \mathrm{O^-} + \mathrm{O}\,. \tag{4.2}$$

The cross section of this process as a function of the electron energy has a maximum at the electron energy of about 7 eV [124], and the rate constant of process (4.1) at the electron energy of $\varepsilon = 6.5$ eV is $k = 2 \times 10^{-10}$ cm^3 s^{-1} [90]. Hence, electrons are absent in a quiet atmosphere, and negative ions partake in ion-molecular reactions with radicals (see, for example, Refs [124, 125]), and this finally leads to the formation of the most stable negative ions. It should be noted that the basic sort of positive and negative ions in a quiet atmosphere may be different depending on admixtures in the atmosphere, but a restricted number of stable ions may be present in the atmosphere. Then, negative ions are characterized by a large electron binding energy; in air, such ions are $\mathrm{NO_2^-}$ (the electron binding energy is 2.3 eV), $\mathrm{NO_3^-}$ (the electron binding energy is 3.7 eV), and $\mathrm{CO_3^-}$ (the electron binding energy is 2.8 eV) [126]. For comparison, the electron binding energy in the negative ion $\mathrm{O_2^-}$ equals 0.43 eV, and the electron binding energy in the negative ion $\mathrm{O^-}$ is approximately 1.5 eV. The last two sorts of negative ions are formed in air at the first stage of the ionization process, and subsequently they are converted into stable negative ions as a result of corresponding ion-molecular reactions. Table 2 contains the mobilities of likely negative and positive ions in nitrogen, measured at low electric field strengths. As is seen, the mobilities of different ions differ slightly from each other, and we use below as the characteristic ion mobility the value of $K \approx 2$ cm^2 V^{-1} s^{-1} [62]. Let us use the Einstein relation [127–129]

$$K_{\mathrm{i}} = \frac{eD_{\mathrm{i}}}{T} \tag{4.3}$$

between the mobility K_{i} and the diffusion coefficient D_{i} of a charged particle in a buffer gas. On the basis of formula (4.3), we have $D_{\mathrm{i}} \approx 0.4$ cm^2 s^{-1} for the ion diffusion coefficient in air at the air pressure of 0.5 atm and room temperature, which determines ion transport at not large distances in a motionless atmosphere.

One can estimate a typical ion density in a dry quiet atmosphere on the basis of the current densities in the atmosphere, which are, on average, equal to 2.4×10^{-16} A cm^{-2} over a land, and 3.7×10^{-16} A cm^{-2} over the ocean [47], or $1500e$ cm^{-2} s^{-1} and $2300e$ cm^{-2} s^{-1}, respectively. Based on the data listed in Table 2, we take a typical ion mobility of 2 cm^2 V^{-1} s^{-1}. Being guided by the average electric field strength $E = 1.3$ V cm^{-1} that corresponds to the ion drift velocity $w = 2.6$ cm s^{-1} in this field, one can obtain the ion number density $N_{\mathrm{i}} \approx 300$ cm^{-3} over land, and $N_{\mathrm{i}} \approx 400$ cm^{-3} over the ocean, taking into account that an atmospheric current is created by both positive and negative ions. Note that this number density relates to molecular ions, whereas macroscopic charged particles may also be present in the atmosphere, and they do not make a contribution to an atmospheric current.

There are two mechanisms of air ionization in a low atmosphere. Near the Earth's surface, it is determined by the radioactivity of the surface if it contains granite or similar components. But ions formed near the Earth's surface do not influence atmospheric electrical processes, because they recombine near the surface and cannot be removed from it. The same relates to ions which are formed at altitudes of 100–200 km as a result of air photoionization under the action of ultraviolet solar radiation. These ions recombine at high altitudes and do not penetrate into the troposphere where atmospheric electricity occurs. Ionization in the troposphere is determined by galactic and solar rays [19, 134, 135]. The maximum ionization proceeds at altitudes from 11 up to 15 km, where the maximum rate of ionization ranges 30–40 cm^{-3} s^{-1} [135, 136]. In accordance with Fig. 17 data, we take the recombination coefficient as 10^{-6} cm^3 s^{-1}. Indeed, according to recombination mechanisms for positive and negative ions, the recombination coefficient depends weakly on the pressure in its range under the consideration. Assuming that ion formation in air proceeds under the action of fast solar particles and ion decay results from the recombination of positive and negative ions, we find that the number density of ions in a region of the maximum ionization of air is equal approximately to 6×10^3 cm^{-3}. This considerably exceeds the ion number density in lower atmospheric layers. Correspondingly, the time of establishment of ion equilibrium is $t \sim 1/(\alpha N_{\mathrm{i}}) \approx 3$ min.

Based on the above ion number density $N_{\mathrm{i}} \approx 300-400$ cm^{-3} in the troposphere, we find a typical recombina-

Table 2. Mobilities of negative and positive ions in nitrogen at room temperature [130–133], reduced to the normal number density of molecules $N = 2.69 \times 10^{19}$ cm^{-3}.

Ion	$\mathrm{NO_2^-}$	$\mathrm{NO_3^-}$	$\mathrm{CO_3^-}$	$\mathrm{NO_2^- \cdot H_2O}$	$\mathrm{NO_3^- \cdot H_2O}$	$\mathrm{CO_3^- \cdot H_2O}$	$\mathrm{N_2^+}$	$\mathrm{CO_2^+}$	$\mathrm{N_2H^+}$	$\mathrm{N_3^+}$	$\mathrm{N_2O^+}$	$\mathrm{N_4^+}$	$\mathrm{H^+ \cdot H_2O}$	$\mathrm{H^+ \cdot (H_2O)_2}$	$\mathrm{H^+ \cdot (H_2O)_3}$
K, cm^2 V^{-1} s^{-1}	2.5	2.3	2.4	2.4	2.2	2.1	1.9	2.2	2.1	2.3	2.3	2.3	2.8	2.3	2.1

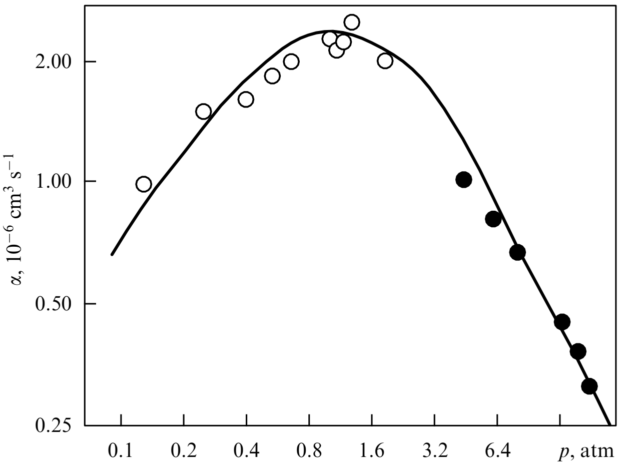

Figure 17. Recombination coefficient of positive and negative ions in air. The solid curve is an approximation of experimental data, which are represented by light [137] and dark [138] dots. The left part of the solid curve corresponds to three-body recombination of positive and negative ions in accordance with the Thomson model [139], and recombination at large pressures (the right part of the curve) is guided by the Langevin formula [140].

tion time $t \sim (\alpha N_i)^{-1} \sim 1$ h. For this time, ions pass $\sim$ 100 m under the action of the Earth's electric field. From this it follows that ions cannot penetrate into the low atmospheric layers from altitudes above 10 km and must be formed just at a given altitude. Let us check the possibility of ion transport from high altitudes as a result of convection. Assuming that the ion number density at a given altitude is determined by ion diffusion in air and ion recombination along their path, we have the following balance equation for the ion number density $N_i(z)$ at a distance z from the ion source:

$$D_i \frac{d^2 N_i}{dz^2} - \alpha N_i^2 = 0, \tag{4.4}$$

where D_i is the diffusion coefficient for ions in air. For simplicity, we take the diffusion coefficient to be independent of the altitude: $D_i(z) = \text{const}$. Then, multiplying equation (4.4) by dN_i/dz and integrating the equation obtained, we reduce it to the form

$$\frac{1}{2}\left(\frac{dN_i}{dz}\right)^2 - \frac{\alpha}{3D_i} N_i^3 = 0.$$

In this equation, we assume that the ion number density N_i and its derivation dN_i/dz tend to zero at the infinite distance from the ion source, i.e., at large z. The solution of this equation with the indicated boundary condition takes the form

$$\frac{1}{\sqrt{N_i}} = \frac{1}{\sqrt{N_0}} + \sqrt{\frac{\alpha}{6D_i}}\, z = \frac{1}{\sqrt{N_0}}\left(1 + \frac{z}{z_0}\right), \quad z_0 = \sqrt{\frac{6D_i}{\alpha N_0}}. \tag{4.5}$$

Here, $N_0 = 6 \times 10^3$ cm^{-3} is the number density of ions at the ion source, the recombination coefficient is $\alpha \approx 2 \times 10^{-6}$ cm^3 s^{-1} as follows from Fig. 17, and $D_i \sim 2 \times 10^4$ cm^2 s^{-1} according to formula (3.39), since we assume that diffusion is governed by air convection. This gives $z_0 \sim 30$ m. According to formula (4.5), we obtain for the number density of ions far from the source

$$N_i(z) = N_0 \frac{z_0^2}{z^2}, \quad z \gg z_0. \tag{4.6}$$

As follows from the last formula, the ion number density is $N_i \approx 10$ cm^{-3} at a distance $z = 1$ km from the ion source, whereas a typical number density of ions in the troposphere is $N_i \approx 300$ cm^{-3}. One can conclude from this that, along with intense air ionization at altitudes of 10–15 km, ionization due to hard radiation of the Sun also proceeds at lower altitudes. This is confirmed by the total rate of ionization of atmospheric air: 4.5×10^7 cm^{-2} s^{-1} [141]. Indeed, the ratio of this value to the maximum specific rate of ionization, reaching approximately 30–40 cm^{-3} s^{-1} [135, 136, 141], gives the layer thickness $h \sim 10$ km, where the ionization proceeds, which corresponds to the conclusion based on equation (4.6) that air ionization takes place over the entire thickness of the lower atmosphere. From this, one can also determine the typical absorption cross section of cosmic rays, which is on the order of 10^{-24} cm^2. It should also be noted that the energy spent on the formation of one ion pair in air amounts to 30 eV. Based on the total rate of air ionization under the action of cosmic rays, 4.5×10^7 cm^{-2} s^{-1}, we obtain the energy flux 2.2×10^{-10} W cm^{-2} due to cosmic rays, which gives the power of cosmic rays for the total globe, $P_{\cos} = 3 \times 10^8$ W. One can compare this power with the total power of solar radiation, 1.7×10^{17} W, that penetrates the Earth's atmosphere.

It should be added to the above that the used diffusion coefficient $D_L \sim 2 \times 10^4$ cm^2 s^{-1} due to air convection relates also to the particles located in air. For water molecules, in particular, this value exceeds by several orders of magnitude the diffusion coefficient of water molecules, $D = 0.2$ cm^2 s^{-1} [101], which was used above. In reality, this small diffusion coefficient is responsible for air transport over small distances below $l \sim 30$ m, where convection is absent. At larger distances, the convective diffusion coefficients D_l or D_L must be applied.

The same relates to the molecular ions whose mobility of 2 cm^2 V^{-1} s^{-1} corresponds to the ion diffusion coefficient $D_i \sim 0.1$ cm^2 s^{-1} at room temperature and is responsible for ion transport over small distances, whereas the diffusion coefficients $D_l \sim 100$ cm^2 s^{-1} and $D_L \sim 2 \times 10^4$ cm^2 s^{-1} may be responsible for ion transport over large distances with corresponding sizes of vortices $l \sim 30$ m and $L \sim 100$ m. Taking into consideration the drift of ions under the action of the Earth's electric field with the drift velocity $w = 2.6$ cm s^{-1}, one can introduce a characteristic size, $r_0 = D_i/w = 0.04$ cm, so that an ion that moves over distances shorter r_0 is determined by ion diffusion with the diffusion coefficient D_i, while ion motion over large distances $z \gg r_0$ proceeds under the action of the Earth's electric field, until convection begins to work, i.e., up to distances $l \sim 30$ m. Let us introduce a typical distance $R_0 = D_L/w = 80$ m, so that ion propagation over larger distances $z \gg R_0$ has a drift character. Thus, the drift character of ion motion takes place for distances $z \ll r_0$, $z \gg R_0$, excluding a narrow range of distances with transport driven by convective diffusion. This analysis leads to the conclusion that molecular ions are formed in the Earth's lower atmosphere under the action of solar or cosmic rays. Ion traveling in the atmosphere results;from both the diffusion in air and the drift in the atmosphere under the action of the Earth's electric field.

4.2 Charge penetration into clouds

The above analysis relates to a dry atmosphere, while electrical processes in the Earth's atmosphere are determined by the aerosols forming clouds. In this case, an additional channel of decay of molecular ions arises due to ion attachment to aerosols. This process is identical for positive and negative ions. Subsequent collisions of positive and negative ions lead to a charge loss in the atmosphere, where ions are formed under the action of cosmic rays. Being guided by aerosol parameters in a typical cumulus cloud (3.1), let us determine first the rate constant $k_{\rm at}$ of ion attachment to neutral aerosols of a cumulus cloud that is determined by the Smoluchowski formula (3.44) [100]:

$$k_{\rm at} = 4\pi D_{\rm i} r \,, \tag{4.7}$$

where $D_{\rm i}$ is the ion diffusion coefficient in air. Referring to the above estimate, $D_{\rm i} = 0.1\ {\rm cm^2\,s^{-1}}$ for the ion diffusion coefficient in motionless air, given in Section 4.1, we obtain at the rate constant of ion attachment to the aerosol surface, $k_{\rm at} \sim 1 \times 10^{-3}\ {\rm cm^3\,s^{-1}}$. Note that the rate of attachment of positive and negative ions to aerosols exceeds the rate of their recombination if the following criterion holds true:

$$k_{\rm at} N_{\rm p} \gg \alpha N_{\rm i} \,, \tag{4.8}$$

where $N_{\rm p}$ is the aerosol number density. This criterion gives $N_{\rm i}/N_{\rm p} \ll 10^3$ if the ion attachment process proceeds in a cumulus cloud, whereas in reality we have $N_{\rm p} \sim N_{\rm i} \sim 10^3\ {\rm cm^{-3}}$, so that criterion (4.8) is satisfied in a cumulus cloud, and a loss of molecular ions in the cloud results from attachment of ions to aerosols and the subsequent recombination of positively and negatively charged aerosols in their collisions. One can find from this a typical lifetime of molecular ions penetrating inside the cumulus cloud.

A typical time of a molecular ion attachment to aerosols is estimated as $\tau = (Nv\sigma)^{-1} \sim 0.06$ s, where $v \sim 3 \times 10^4\ {\rm cm\,s^{-1}}$ is the thermal velocity of molecular ions, and $\sigma \approx 2 \times 10^{-6}\ {\rm cm^2}$ is the cross section of an aerosol or the cross section of ion attachment to an aerosol. As is seen, a typical time of ion attachment to aerosols in a cumulus cloud is much less than that of mutual neutralization of positive and negative molecular ions in their collisions in dry air, which is estimated as $\tau_{\rm rec} \sim 10^3$ s. From this it follows that a thickness of penetration of molecular ions inside a cumulus cloud, $l \sim (2D_{\rm i}\tau_{\rm at})^{1/2} \sim 10$ cm, is much larger than the mean free path $\lambda \sim 0.06$ μm of ions in atmospheric air, but is significantly less than a typical cloud size.

4.3 Processes of aerosol charging

Cumulus clouds play a fundamental role in atmospheric electric phenomena [37, 39–42, 45] and allow focusing the atmospheric electric energy in a small region, which may cause the development of thunderstorms. Clouds consist of micron-size water aerosols, and precisely these aerosols are responsible for charge separation in the Earth's atmosphere. Clouds are formed as a result of convective transport in the atmosphere [142–144], so that atmospheric moisture is condensed in cold atmospheric regions, and, in spite of energy release in the course of water condensation, this process intensifies as a result of transport phenomena, and bound water is concentrated in a restricted atmospheric region. Since this process is determined by convective transport in the atmosphere, the creation of cumulus clouds is a random process which is accompanied by large gradients in aerosol distribution.

According to existing models of transport processes in clouds (see, for example, papers [142, 145–148]), convective transport promotes both the formation and smearing of cumulus clouds. Below, leaving aside processes of cloud development and cloud influence on the Earth's climate, we consider cumulus clouds as a medium in which electric currents are formed and lead to charge separation. In considering cumulus clouds as sources of atmospheric electricity, we will be guided by aerosol parameters in a typical cumulus cloud (3.1) in accordance with available data [45, 84–86], which include the average aerosol radius $r_0 = 8$ μm and water density of 2 g ${\rm m^{-3}}$ in aerosols in atmospheric air, though it may reach the value of 20 g ${\rm m^{-3}}$ [149] at the last stage of development of cumulus clouds before water precipitation. The average water content in aerosols corresponds to the number density $N_{\rm p} = 1 \times 10^3\ {\rm cm^{-3}}$ of aerosols. Based on average parameters of aerosols in a cumulus cloud, we will analyze below the character of formation of charged aerosols and charge separation in the atmosphere. A charge is formed at altitudes where water may be found in different aggregate states. Along with liquid and solid (ice) phases, aerosols may contain amorphous water (hailstones) or a mixture of snow and liquid water (graupel). These types of aerosols may coexist at altitudes of several kilometers with temperatures close to 0 °C. For this reason thunderstorms are observed in warm seasons when a zero atmospheric temperature is attained at altitudes of several kilometers.

Charging in collisions of water and ice microparticles was first observed in 1957 [3], where micron-size droplets acquired a negative charge, while microparticles of ice were charged positively. Various mechanisms of interaction may lead to this charging, but in any case the charging process results from the contact voltage at the interface between different aggregate states of water aerosols [150, 151], which leads to charging of colliding particles [4, 8, 152, 153, 155]. Just this process leads to aerosol charging in clouds [6, 7, 45, 152–157]. In spite of different standpoints [4, 8, 152, 153, 155] about the rates of charging and the existence of empiric relations [6, 158, 159] for a certain range of charging parameters, the description of this process has a qualitative character and the focus is on the collisions of ice particles and graupel consisting of a snow and water mixture [157, 160–162]. Therefore, we consider below the nature of the process of aerosol charging in collisions, and the aerosol charge after charge separation in the atmosphere as a result of aerosol falling we determine on the basis of the growth rate of aerosols in the course of their gravitational association.

If we consider aerosol water as a weak electrolyte containing positive H_3O^+ ions and negative OH^- ions, one can describe aerosol charging as ion transfer through the interface during aerosol contact, so that one aerosol contains an excess of H_3O^+ ions, while another aerosol comprises an excess of OH^- ions. If aerosols are in the solid state, one can connect the aerosol charging with the production of charged defects. Then, the aerosol charging proceeds not only in collision processes, but also in processes of aerosol growth and evaporation. According to experiments [4, 163, 164], an excess of OH^- ions is observed in aerosol growth via the addition of water molecules, whereas an excess of H_3O^+ ions takes place in the sublimation process. The electrolyte

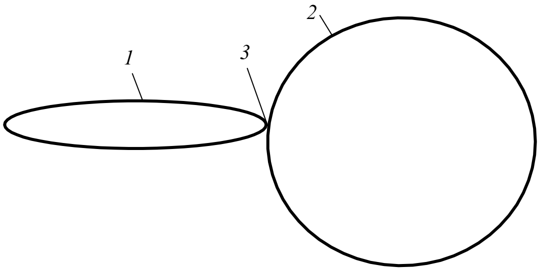

Figure 18. Aerosol charging in collisions of ice microparticles (*1*) and liquid water microparticles (*2*). Upon their contact (*3*), an electric voltage arises giving rise to charging of particles after their separation.

properties of water lead to aerosol charging in collisions with a surface. For example, electric discharges are observed in collisions of snowflakes with aircraft antennas [165], and charges of clouds influence flight electrification [166].

A diagram of aerosol charging is given in Fig. 18. Leaving aside the mechanism of this process, note that a interaction energy at the interface exceeds significantly a thermal energy of micron-sized aerosols. In particular, the specific energy of ice melting is 334 J g^{-1} [73] or 0.06 eV per water molecule, which exceeds a thermal aerosol energy (1/40 eV). Hence, adhesion of charged aerosols takes place at their contact, and their separation is possible if the potential of the Coulomb attraction of aerosols does not exceed their thermal energy $\sim T$, so that a typical charge Z_0 of originally formed aerosols is connected with their radius r_0 in accordance with the following estimate [62, 81]:

$$\frac{Z_0^2 e^2}{r_0} \sim T\,.$$

Hence, a typical average charge Z_0 of forming aerosols is given by formula

$$Z_0 = b\sqrt{\frac{r_0 T}{e^2}}\,, \qquad (4.9)$$

where the numerical factor b is on the order of unity. Let us assume that at subsequent aerosol joining in collisions a forming aerosol acquires a summary charge, i.e., the ratio of the aerosol charge to mass does not vary in the course of aerosol association. Then, based on formula (3.15) and assuming that the initial charging conforms to the aerosol charge $Z = 1$, one can find the initial aerosol mass $m = 8 \times 10^{-11}$ g that corresponds to an initial aerosol radius $r_0 = 2.7$ μm. Subsequent aerosol growth in a cumulus cloud leads to an increase in the aerosol charge [formula (3.14)] and mass by approximately a factor of 30.

We assume for simplicity that the initial aerosol charging results from collisions of liquid water aerosols and ice microparticles, leading to the charging of heavier liquid aerosols negatively. A faster fall of these aerosols leads finally to negative charging of the Earth's surface, as is observed. But laboratory study [10, 11] reveals that collisions of neutral ice microparticles and graupel (a mixture of snow and liquid) bring about a negative charging of ice microparticles. In some observations, a subsequent gravity fall of ice microparticles leads to a positive charging of the low cloud edge [167]. Nevertheless, the mechanism of aerosol charging in the clouds under consideration [3] as a result of collisions of aerosols in different aggregate states has no alternative, though it requires a detailed study.

Let us consider the character of aerosol charging according to the history of the study of this process. It is clear that the charged aerosols after their formation follow a charge separation due to the gravity fall of the aerosols and the induction of the Earth's electric field in this manner, and nanometer-size particles cannot create high electric fields. In this consideration, it is necessary that aerosols of different sizes acquire charges of different signs. On the basis of ideas developed in the first half of the 20th century, Frenkel [168] assumed a selective character of formation of charged aerosols, where large aerosols acquire a negative charge. But this contradicts the contemporary understanding of aerosol processes, and the above mechanism of formation of the electric charge results in collisions between aerosols in different aggregate states [3]. This charging process depends on many factors, including the character of formation of the aerosol surface. This led to contradictions in the rates of the process of aerosol charging (see, for example, Refs [7, 74, 75]). It is probably for this reason that paper [3] did not occupy such a place in the physics of atmospheric electricity that it merits.

4.4 Charge separation in clouds

The separation of positive and negative charges is an important stage of atmospheric electrical processes, and for simplicity we will consider it as a result of the gravity fall of ice and water aerosols. According to the Stokes formula, the velocity of fall of aerosols in the atmosphere is determined from the equality between their weight and the friction force of their motion in air. From this condition, we may find the fall velocity w of micron-sized water droplets in atmospheric air at the temperature $T = 300$ K [62]:

$$w = 0.012\left(\frac{r_0}{a}\right)^2, \qquad (4.10)$$

where the velocity of aerosol fall is expressed in cm s^{-1}, the aerosol radius r_0 is given in μm, and $a = 1$ μm.

Aerosol parameters in a cumulus cloud according to formulas (3.1) and (3.14) are as follows: the average aerosol radius $r_0 = 8$ μm, and the average charge $Z = 27$. Then, referring to formula (4.10) allows arriving at the fall velocity of heavier aerosols, $w = 0.7$ cm s^{-1} (we assume that these aerosols are liquid), which leads to the following current density as a result of aerosol falling:

$$i = ZeN_{\mathrm{p}}w\,. \qquad (4.11)$$

This gives $i = 3 \times 10^{-15}$ A cm^{-2} for the parameters indicated in formula (4.11). From this it follows that the total electric current through the Earth's atmosphere is $I = 1700$ A, and it may be reached if cumulus clouds cover approximately 10% of the sky area.

One can add to this that, though we are using aerosol motion to the Earth's surface under the action of the gravitational field as our basis, convection may lead to a stronger charge transfer. But since convection is a random process, it does not lead to charge separation, and charge transfer over a long time is determined by the gravity fall of aerosols. Therefore, charge transfer through long time intervals may be connected with the gravity falling of aerosols. Note that the average time of water location in the

Figure 19. Eruption of the Puyehue volcano in Chile in June 2011 with lightning flashes inside eruption jets [170, 171].

atmosphere is 9 days [169], and during this time an aerosol passes 6 km, which is comparable to or exceeds the height of a cumulus cloud. Note that the values obtained are estimates, rather than average values of parameters. Nevertheless, they give a noncontradicted physical picture for processes of aerosol charging and charge separation in the atmosphere, which are the basis of the physics of atmospheric electricity.

As is seen, the separation of charges and resulting electric breakdown in the Earth's atmosphere in the form of thunderbolt are determined by collisions between aerosols in different aggregate states. In other media, say, in the atmospheres of other planets, particles of other types may be responsible for electrical processes. As for the Earth's atmosphere, dust microparticles of any type may participate in electrical processes, so that lightning may be observed in sandstorms, in coal mines with coal dust, in jets of volcano eruptions, and in an explosion of an atomic bomb. In particular, volcano eruptions are often accompanied by lightning flashes inside them, as demonstrated in Fig. 19, where the eruption of the Puyehue volcano in Chile is represented. The minimum size for charge separation in volcano dust that leads to lightning creation is less than that in thunderclouds, so that lightning flashes are developed in volcanic dust at approximately 500 m, whereas the minimum cloud size for thunderstorms is 3 km [59]. In all these cases, electricity arises in a gas as a result of collisions between microparticles of various compounds. This mechanism is also realized in the atmospheres of other planets. Note that, though electrical processes in various cases have identical natures, the parameters of these processes may be different.

Earth's charging as a result of the falling of charged aerosols may proceed without lightning flashes if charged aerosols can reach the Earth's surface. Evidently, air upstreams from the Earth's surface do not allow clouds to come to the Earth, and then their charge can reach the Earth's surface in the form of lightning. Probably, the direct channel of charge transfer through charged aerosols exists and is realized in mountains and oceans. Then, high electric fields occur near the surface and create corona discharges near conducting objects, causing the air to glow at their location. According to my mountaineer experience, it is dangerous to reside on a ridge or high open places in prethundery weather. There, ice axes, climbing ions, and other metallic objects jingle, testifying to the propagation of electric currents through them. The glowing of oblong objects in prethundery mist weather has been known for thousands of years under different names. This phenomenon was often observed by sailors at sea, and it is known in this case as St. Elmo's fire [172], with St. Elmo being a protector of sailors. As follows from the above consideration, the character of electric phenomena in Earth's atmosphere is more complex and richer than what follows from simple patterns.

5. Conclusions

Summing up the above analysis, one can conclude that four basic processes are of fundamental importance for atmospheric electricity, namely, aerosol charging, charge separation in the atmosphere, charge transfer from cumulus clouds to the Earth's surface through lightning flashes, and discharging of the Earth by currents of molecular ions in a dry atmosphere under the action of cosmic rays. These processes have become known over at last fifty years, and our task was to join them in a united physical picture. Let us briefly outline these processes.

The charging of micron-size aerosols proceeds in atmospheric layers where water aerosols may exist in different aggregate states, i.e., liquid drops, ice particles, snow particles, hailstones, and graupel. Collisions between two aerosols in different aggregate states lead to their charging. A subsequent process of charge separation results in a gravity fall of charged aerosols, and if a size and number density of positively and negatively charged aerosols are different, they fall with different velocities, creating an electric current in the atmosphere. Assuming this current to be determined by the falling of one singly charged component only, we obtain in modelling an aerosol by a spherical particle the following estimate for an aerosol radius r [62]:

$$0.1\ \mu\text{m} < r < 17\ \mu\text{m}\,. \tag{5.1}$$

If an aerosol radius is below the lower limit of formula (5.1), negatively charged aerosols will rise under the action of the Earth's electric field, rather than fall. For this very reason, molecular ions do not partake in Earth's charging. On the other hand, the transfer of an electric charge from the atmosphere towards the Earth's surface is a secondary phenomenon with respect to water circulation in the atmosphere. This is characterized by a certain ratio of a transferred charge to the transferred water mass in accordance with formula (2.3). Thus, atmospheric electricity is created by micron-sized aerosols.

As a result of the gravity fall of charged aerosols, the lower cloud edge becomes charged, and more often negatively. But air with charged aerosols is a dielectric medium, and therefore charge transfer to the Earth's surface results from thunderstorm phenomena, if positive ions of the upper part of the conductive lightning channel travel to negatively charged aerosols and partially neutralize them. Finally, the origin for the last key process of atmospheric electricity makes hard cosmic rays of the Sun which ionize the air over all altitudes, thus leading to the formation of molecular ions and subsequent discharging of the Earth.

For simplicity, we used in the above analysis averaged values of observational data and a simple system for atmospheric processes. For example, the frequency of the appearance of lightning flashes according to NASA satellite observations [55] is distributed nonuniformly over the globe surface. It is high for land and coastal regions of the subtropics, while over the oceans lightning flashes are seldom observed. Another example is given in Fig. 1, which shows that a real charge space distribution is more complex than the dipole charge distribution in a cloud that is formed in the course of charge separation in the gravity fall of charged aerosols. These examples testify to a qualitative description of atmospheric electricity on the basis of simple models.

It should be noted that, although concepts which are the basis of each stage of the electrical cycle in the atmosphere were formulated at least half a century ago, we cannot now answer some principal questions for each key process. In particular, correct information is absent for aerosol charging depending on the aggregate state and the parameters of atmospheric layers with a temperature between 0 °C and −40 °C, where water aerosols may coexist in different aggregate states. Without this information, we cannot answer the question of why the basic part of a cumulus cloud is neutral and only a small part is charged. Next, according to our experience, lightning flashes arise after penetration of fluxes of a warm wet air into charged layers of a cumulus cloud, and air convection plays a fundamental role in the initialization of thunderstorms, but a simple model for describing this process is absent.

In this context, one might expect a deep understanding of the nature of lightning because of its long study, since as early as the middle of the 18th century the lightning rod was created for building protection, which testifies to the understanding of this problem at that time. If lightning flash is considered as the electric breakdown between a charged cloud and the Earth's surface, one can divide this phenomenon into several processes. The concept of a streamer, the first stage of these processes, was formulated about 80 years ago and considers the streamer as an ionization wave that is not varied during this time. The streamer's main features are an intensification of the electric field near the wave front due to charge redistribution and the transport of ionization by photons which are emitted at the ionization wave front. Another type of ionization wave, the lightning leader, is understood to a less degree. The main unsolved questions are why a stepwise leader moves with stops, rather than monotonically, and what instabilities cause such a character of this process. It is surprising that long-term investigations of a long spark which, as might be expected, is a lightning model, could not answer this question.

The basic stage of the transport of the electric charge in lightning was considered in this review, where a charged cloud and the Earth's surface are modelled by conducting electrodes. If a conducting plasma channel is created, the rate of charge transport along this channel is determined by the plasma conductivity, and the conducting channel is heated and expanded in the course of propagation of the electric current. In this consideration, charge transport proceeds in a continuous manner, so that the electric current and parameters of the plasma conducting channel vary slightly in time. In spite of an expected analogy between lightning and this model, lightning flash demonstrates a complex character of electric current propagation, with current containing repeated phases of strong and weak electric currents, as is detected in observations of different thunderbolts. It is clear that this may be connected with the convective motion of a charged cloud and charge transfer across the Earth's surface, but a simple model is necessary for the observational charge transfer in a lightning flash. Thus, though the physics of atmospheric electricity is understood in general, some details of this phenomenon require yet further study.

The author thanks Prof. V P Krainov for valuable comments.

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