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Boris M. Smirnov

Microphysics of Atmospheric Phenomena

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Microphysics of Atmospheric Phenomena

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Boris M. Smirnov
Institute for High Temperatures
Russian Academy of Sciences
Moscow
Russia

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Preface

The goal of this book is to give a reliable description of various global atmospheric properties and some phenomena including photoprocesses, processes of atmospheric electricity with participation of aerosols, the atmospheric chemistry with processes involving ions, oxygen atoms, excited atomic particles and ozone molecules, as well as processes in an ionospheric plasma. In the course of this consideration, we are based on observational data, existing atmospheric concepts, and measured rate constants of elementary processes in atmospheric air. Fundamentals of global atmospheric phenomena are based on the standard atmosphere model with averaged atmospheric parameters over the globe and time, the global electric circuit, the energetic balance of the Earth, and other global concepts. As a result, one can give a qualitative description of atmospheric phenomena and estimate their parameters. This analysis is based on processes involving elementary atmospheric particles and gives a simple and reliable understanding of the Earth processes and phenomena. In addition, this analysis allows one to glance at future trends of topical atmospheric problems, such as ozone problem and climate change.

Moscow, Russia

Boris M. Smirnov

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Chapter 1

Introduction

The goal of this book is the analysis of some aspects of atmospheric physics and chemistry on the basis of elementary processes in the Earth's atmosphere, as well as global atmospheric properties and phenomena that are based on these processes. This book has common features with [1] for the kinetics of excited air. Here we consider mostly other aspects of this general problem including radiative and aerosol processes, and also respective atmospheric phenomena. In consideration of atmospheric phenomena, we are grounded on observational and measured data using simple reliable concepts and models. Taking elementary processes in excited air as a basis, we use appropriate global models and concepts including the global circuit model [2] for electric processes in the atmosphere and the model of standard atmosphere [3–6]; that is, we use an average of atmospheric parameters over the globe. This allows us to obtain detailed information about atmospheric properties or phenomena in a simple form. Following is a list of such problems to deepen understanding on the basis of this analysis.

1. Collisions of neutral aerosols in different aggregate states leads to their charging, and the subsequent fall of positive and negative aerosols with different falling velocities leads to charge separation in the atmosphere and the creation of atmospheric electric fields as a source of atmospheric electrical phenomena. Charged aerosols constitute cumulus clouds, and due to a charge, growth of aerosols in cumulus clouds proceeds with a lower rate compared with that for neutral aerosols. Then from rates of the coagulation process and the lifetime of a cumulus cloud it follows that a typical charge of aerosols in cumulus clouds includes $Z = 25 - 30$ electron charges. Ionization of atmospheric air is realized by secondary Mev-energy particles which are formed in nuclear reactions involving Gev-energy protons or neutrons.

2. Note a significant role of cosmic rays in atmospheric processes. Atmospheric ionization by cosmic rays is important for aerosol discharging and creating a plasma that does not allow for clouds to expand in a surrounding space. In addition, showers of cosmic rays initiate the beginning of lightning in a thunderstorm.

3. The atmospheric optical depth u for infrared radiation is approximately equal to $u = 2.7$, as follows from the analysis of the energetic balance of the Earth and its atmosphere. The greenhouse atmospheric effect is determined mostly by atmospheric

water, and, as follows from the subsequent analysis, the above optical depth may be created by atmospheric aerosols if (1–2) % of atmospheric water is converted in aerosols. This may cause anxiety because of an atmospheric instability for the global climate that can be changed by a weak action. Next, according to NASA data [7], the average Earth temperature has increased by $(0.8 \pm 0.1)^\circ\text{C}$ from 1880 up to now. Keeping in mind the huge effort to obtain this value which requires monitoring by thousands of meteorological stations because temperature fluctuations are tens of degrees, one can note that this change corresponds to an increase of the total mass of atmospheric water by approximately 10 %. The press and TV convince us that the main reason for the above temperature increase is an accumulation of atmospheric CO_2 because its concentration in the atmosphere has increased by about 30 % during the twentieth century. In this book we analyze the contribution of atmospheric CO_2 in the greenhouse effect step by step and find that doubling of the CO_2 concentration in the atmosphere leads to an increase of the Earth's global temperature by $(0.4 \pm 0.2)^\circ\text{C}$ and at the contemporary rate of change of the CO_2 concentration the doubling of its concentration will take 130 years. Thus, the contribution of atmospheric CO_2 to the greenhouse effect of the atmosphere is several times less than that from atmospheric water. This contribution also depends on the manner of CO_2 generation, so that deforestation acts more strongly on the balance of atmospheric CO_2 than that due to combustion of fossil fuels.

4. In spite of a small ozone concentration (<0.01 %) in the stratosphere, it is important both for absorption of ultraviolet solar radiation and for thermal balance of the stratosphere. But seasonal and daily fluctuations of the ozone concentration are compared with its concentration, as well as its average concentration in various parts of the globe. The latter is reflected by the term “ozone hole” for poles where the ozone concentration at the poles is less than approximately twice that at equatorial regions, but this does not characterize an absorption of ultraviolet radiation in atmospheric air. For example, the probability of reaching the Earth's surface for ultraviolet solar photons, which mostly are absorbed by ozone molecules, is $\exp(-200)$ in an equatorial zone and $\exp(-100)$ at the poles. The same changes relate to daily and seasonal ozone concentrations, so that the term “ozone hole” does not correspond to a meaning of these words.

Although above we enumerate some atmospheric problems that are distorted in the press or are not taken into account in atmospheric science, it is not the main goal of this book, which consists of an academic description of elementary atmospheric processes and their kinetics. This, together with other atmospheric reviews and books, allows us to formulate the physical picture of atmospheric processes and to correct some positions of this description. Let us demonstrate this for electrical atmospheric processes. One can consider processes of atmospheric electricity as a secondary phenomenon with respect to water circulation through the atmosphere [8]. The central process of atmospheric electricity is the charging process as a result of collisions of aerosols that are located in different aggregate states. This characteristic of the charging process follows from the experiment [9] which, in the author's opinion, is underestimated.

Cosmic rays are of importance for atmospheric electricity and electrical atmospheric processes. Air ionization by cosmic rays in the troposphere and above the tropopause is important for the Earth's discharge, which is charged negatively [10, 11]. In reality, the study of cosmic rays starts from investigation of air ionization in the troposphere [12]. But the role of cosmic rays in processes of atmospheric electricity is not restricted by ion formation in the troposphere. One of the problems of atmospheric electricity is the origin of lightning. Because a typical electrical field strength in a thunderstorm due to a cloud charge is two orders of magnitude lower than the breakdown of electrical field strength in air, the development of lightning is impossible from one seed electron. It is possible as a result of streamer propagation, but approximately 10^8 electrons must be gathered in a small volume for streamer realization [13] in order to create electrical field strengths above the breakdown. The analysis [14] shows that known mechanisms do not explain this stage of lightning evolution. Evidently, the answer for this follows from the specific investigations in the Tien Shan station of cosmic rays [15] where the correlation was observed between showers of cosmic rays and radio emission from lightning that is evidence of lightning origin as a result of atmospheric ionization by cosmic showers.

Lightning or electrical breakdown of the atmosphere is a stage of atmospheric electricity that leads to negative charging of the Earth. The contemporary state of lightning physics is represented in Uman's books [16–19], where experimental methods are given and observational parameters of lightning are obtained in detail. The theoretical analysis of lightning development is contained in the Bazelyan and Raizer books [20–23] where models are used on the basis of electric schemes and a uniform plasma. Therefore we describe this phenomenon briefly, referring to the above books and mentioning only some aspects of this phenomenon where physics of elementary and other processes are able to give a new standpoint.

However, we correct the appropriate atmospheric electricity parameters if they contradict the results of the special analysis. In particular, the first stage of lightning, the stepwise leader, is an ionization wave that forms a conductive channel for subsequent propagation of a lightning electric current. In order to conserve a plasma in this channel during leader propagation, it is necessary to suppress attachment of electrons to oxygen molecules, and this is realized due to a heightened channel temperature. This temperature is expected above 5000 K according to [20–23]. On the basis of rate constants of the electron attachment process in the following we prove that this boundary temperature is about 1200 K.

Next, it is supposed on the basis of measurements that the lightning leader transfers on average a charge of 5 C that corresponds to a negative charge gradient along the leader channel $\sim 10^{-3} \text{C/m}$ [14]. One can see that at a channel radius of 10 cm this corresponds to the number density of excess electrons on the order of 10^{11}cm^{-3} . At this excess number density of electrons the electric potential of 50 MV, a typical potential of a thundercloud, arises at a length of 1 m; that is, the ionization wave is locked at such distances. This means that the statement about a charge transfer at the stepwise leader stage is wrong, and an indicated transferred charge of 5 C relates to other stages of lightning evolution.

The above examples for lightning demonstrate the possibilities of using rates of microphysical processes in the analysis of atmospheric processes and phenomena. Returning to lightning, note that such an approach allows us to compare the contribution of various mechanisms to energy release in the lightning channel during propagation of an electric current through it, the lifetime of the conductive channel in the absence of an electric current in it, the character of expansion of the conductive channel in time, and so on. All this may be added to a contemporary understanding of the lightning physics.

The analysis of the Earth's ionosphere as one of the objectives of this book is based, in the first place, on elementary processes involving its atomic particles. According to H.S.W. Massey, the classic atmospheric physics and physics of atomic collisions, "A detailed understanding of the ionospheres of the Earth and planets requires knowledge of the rates of many ionic reactions as well as of electron recombination coefficients and photoionization cross sections" [24]. Indeed, the Earth's ionosphere contains nonequilibrium excited air; that is, its properties are determined by processes involving its atomic particles, and this air nonequilibrium starts from photoionization and photodissociation of atmospheric air under the action of solar radiation. Study of the ionosphere develops in this manner (e.g., [25–28]). Below we collect information about elementary processes that is useful for ionospheric properties.

Thus, the goal of this book is to give a simple and reliable analysis of atmospheric processes and corresponding properties of the Earth's atmosphere on the basis of physical laws and rate constants of elementary processes involving atmospheric particles. This analysis allows us to ground some existing concepts of atmospheric physics and chemistry with additional information. As a result, the analysis gives a new perspective on some aspects of processes and phenomena under consideration, and some additions to the contemporary physical picture of the Earth's atmosphere. Briefly, we consider the Earth's atmosphere as a physical object, and this approach allows us to determine definitely the mechanisms of some atmospheric phenomena with estimations for their parameters.

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Part I
Fundamentals of Atmosphere
and Atmospheric Processes

Chapter 2

Properties of Neutral Atmosphere

Abstract Some atmospheric parameters as a function of altitude are represented for the atmospheric standard model that operates with averaging over time and season parameters at the latitude of $45^\circ N$. As a result of air circulation due to air convective motion, molecules and small aerosols are captured by vortices, and their concentration in air does not vary with altitude, whereas micron-size aerosols do not penetrate altitudes above those of their formation. Therefore concentration of water molecules in the stratosphere is small compared to that in the troposphere. The water and carbon balances in the atmosphere are analyzed, and accumulation of CO_2 in the atmosphere is considered from various standpoints. The atmospheric energetic balance of the Earth and its atmosphere is represented and is compared with energetics of other processes.

2.1 Standard Atmosphere

In order to analyze global properties of the Earth's atmosphere, it is convenient to deal with average atmospheric parameters. A widespread average over time and the globe leads to the so-called standard atmosphere model. The standard atmosphere parameters are averaged over longitude, season, and time, so that they depend on the altitude only. Instead of averaging over the latitudes, these data are taken for the latitude of $45^\circ N$. Atmospheric parameters depend on radiative fluxes from the Sun, and we use a basis of atmospheric parameters at moderate solar activity. Below we use atmospheric parameters taken from [1–4]. In this case basic atmospheric parameters at sea level are equal as follows. The air pressure is $p = 101.3 \text{ kPa}$, the air temperature is $T = 288.2 \text{ K}$, the air density is $\rho = 1.225 \text{ kg/m}^3$, and the number density of air molecules is $2.547 \times 10^{19} \text{ cm}^{-3}$. In addition, the mean free path of air molecules is $\lambda = 66.33 \text{ nm}$, the mean molecular weight in atomic units of masses is $M = 28.96$, the sound speed is $u = 340.3 \text{ m/s}$, the freefall acceleration is $g = 980.7 \text{ cm/s}^2$, the heat capacity at constant pressure is $C_p = 1.005 \text{ J/(g} \cdot \text{K)}$, the heat capacity at constant volume $C_v = 0.718 \text{ J/(g} \cdot \text{K)}$, the dynamic viscosity coefficient of air is $\eta = 1.789 \text{ P}$ (the dynamic viscosity unit P , Poiseuille, is equal to $1P = 0.1 \text{ Pa} \cdot \text{s}$) and corresponds to the kinematic viscosity coefficient of air $\nu = 0.146 \text{ cm}^2/\text{s}$,

and the thermal conductivity coefficient of air is $\kappa = 2.596 \times 10^{-4} \text{ W}/(\text{cm} \cdot \text{K})$ corresponds to the thermal diffusivity coefficient $\chi = \kappa/(C_p \cdot \rho) = 0.21 \text{ cm}^2/\text{s}$. Note that we express the temperature in energetic units, and a transfer from Kelvin to other energetic units is given in Appendix A.3.

Figure 2.1 represents the temperature distributions over altitudes. As shown, the atmosphere temperature has nonmonotonic dependence on the altitude. At sea level the US standard atmosphere has a temperature of 288 K [1]. The first temperature minimum $T = 217 \text{ K}$ is located above the tropopause in the altitude range $h = 11\text{--}20 \text{ km}$. Another temperature minimum $T = 187 \text{ K}$ is found in a range of the mesopause at altitudes $h = 86\text{--}93 \text{ km}$. One can repeat that current average values of the atmospheric temperature may differ from a given value; that is, the data represented have a qualitative character.

Figures 2.2, 2.3, and 2.4 give the altitude dependencies for the pressure, air density, and density of air molecules, respectively, under conditions of standard atmosphere in accordance with [1–3]. It is convenient to introduce the typical size (scale) with respect to pressure Λ_p , density Λ_ρ , and number density of air molecules Λ_N on the basis of relations

$$\Lambda_p = \left(\frac{d \ln p}{dh} \right)^{-1}, \quad \Lambda_\rho = \left(\frac{d \ln \rho}{dh} \right)^{-1}, \quad \Lambda_N = \left(\frac{d \ln N}{dh} \right)^{-1}, \quad (2.1)$$

Fig. 2.1 Temperature distribution over altitude in the atmosphere with average parameters and notations of low atmosphere regions

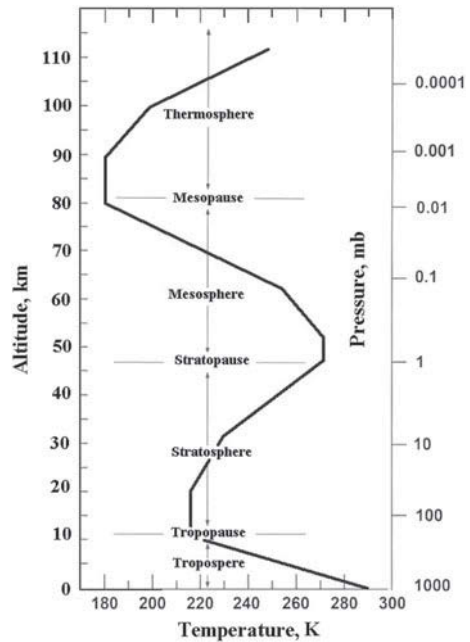


Fig. 2.2 Dependence of the air pressure p on the altitude for standard atmosphere

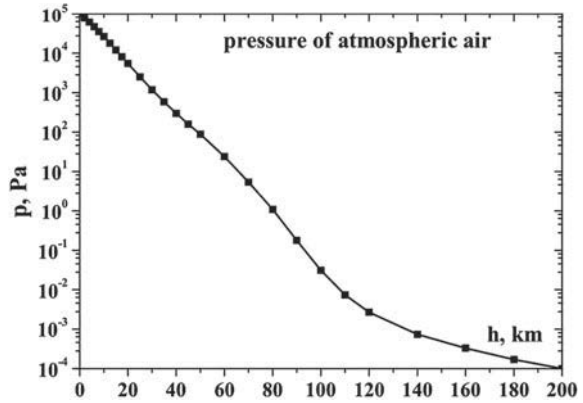


Fig. 2.3 Dependence of the air density ρ on the altitude h for standard atmosphere

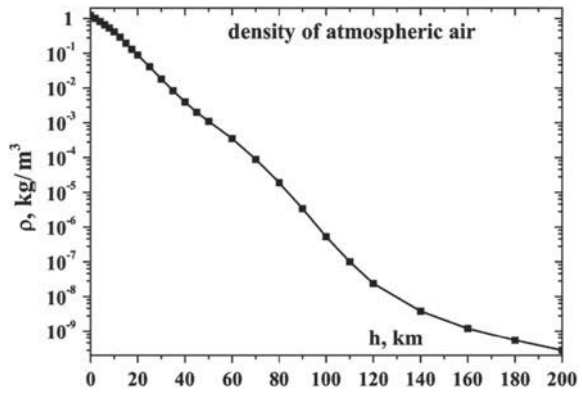


Fig. 2.4 Dependence of the number density of air molecules N on the altitude h for standard atmosphere

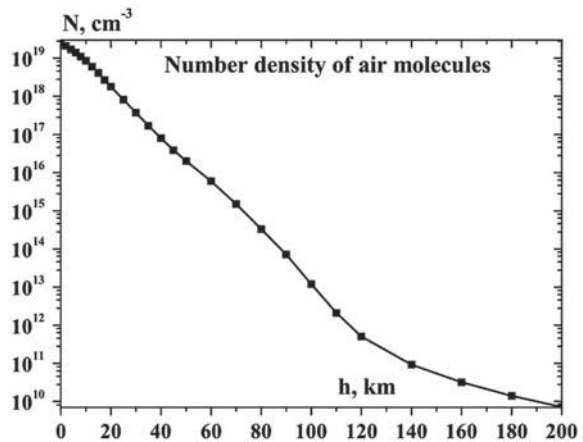
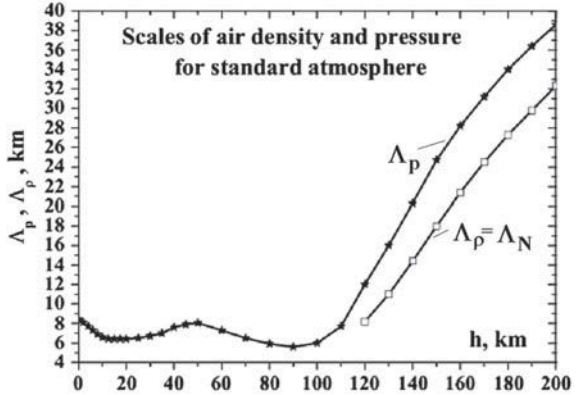


Fig. 2.5 Dependence of scales of the air pressure $\Lambda_p = (d \ln p/dh)^{-1}$ and the air density $\Lambda_N = (d \ln N/dh)^{-1}$ on the altitude h for standard atmosphere



These quantities describe the character of variation of corresponding parameters, in particular, for the air pressure near the altitude h_o , where it is $p(h_o)$, we have

$$p(h) = p(h_o) \cdot \exp\left(-\frac{h - h_o}{\Lambda_p}\right) \quad (2.2)$$

The same relation corresponds to the number density of molecules or atoms $N(h)$ at an altitude h

$$N(h) = N(h_o) \cdot \exp\left(-\frac{h - h_o}{\Lambda_N}\right) \quad (2.3)$$

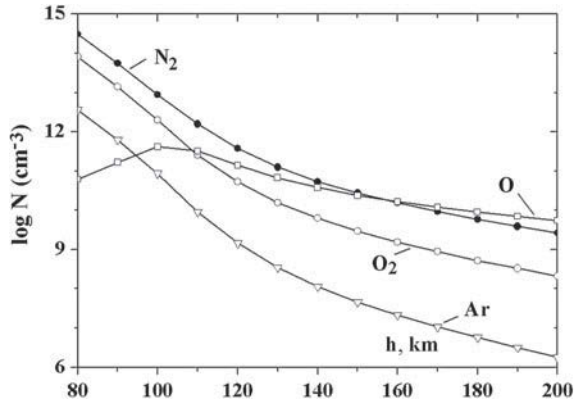
In addition, one can use the definition of the mass density $\rho = Nm$, where m is the air molecule mass, and we have $\Lambda_p = \Lambda_N$. The state equation for an ideal gas $p = NT$ given in appropriate units leads to the relation:

$$\frac{1}{\Lambda_p} = \frac{1}{\Lambda_N} + \frac{1}{\Lambda_T} \quad (2.4)$$

Figure 2.5 contains the altitude dependence for the parameter Λ_p and also for the parameter $\Lambda_N = \Lambda_\rho$ in accordance with data [1-3].

In the above consideration we define air molecules that have the average parameters of nitrogen and oxygen molecules. We now take into account that basic components of the Earth's atmosphere within the framework of the standard atmosphere include at sea level 78.08 % nitrogen molecules, 20.95 % oxygen molecules, 0.93 % argon atoms, 0.04 % CO_2 molecules, and 0.0018 % neon atoms. These concentrations of atoms and molecules in air are conserved at not high altitudes because of mixing of air components due to convection. The atmospheric composition varies at altitudes above 100km, in the first place, due to dissociation of oxygen molecules under the action of solar radiation. Figure 2.6 represents the altitude dependencies for the number densities of nitrogen molecules $N(N_2)$, oxygen molecules

Fig. 2.6 Altitude dependence for the density logarithm ($\log_{10} N$) of the number densities of nitrogen and oxygen molecules, as well as the number densities of oxygen and argon atoms



$N(O_2)$, oxygen atoms $N(O)$, and argon atoms $N(Ar)$ in the upper atmosphere. As shown, the number density of oxygen atoms becomes equal to the number density of oxygen molecules at an altitude of about 100 km and to the number density of nitrogen molecules at an altitude of 155 km.

2.2 Air Transport in the Atmosphere

Heat transport in the lower atmospheric layers proceeds at altitudes 3–10 km in the form of air convection which is accompanied by processes of water condensation with formation of water aerosols [5]. Simultaneously, micron-size liquid aerosols and solid aerosols-ice particles-may be formed at such altitudes. In addition, atmospheric convection is of importance for these altitudes because of large atmospheric sizes and high atmospheric pressure. Convective motion is important for transport of air and water in the atmosphere and for atmospheric electric phenomena.

In the first place, transport of atmospheric air and water is important for the weather. Describing these processes by corresponding equations, one can expect the possibility to obtain atmospheric parameters at each point and a given time on the basis of values of these parameters at the beginning. Then one can predict the weather by solving the corresponding transport equations, given initial conditions on the basis of data at the beginning for a wide region of the globe. This concept of weather prediction at a given locality and in a given time as a result of numerical solution of transport equations has existed for almost a century (e.g., [6]), and the contemporary forecast is based on power computer codes. The authenticity of such a forecast decreases with increasing prediction time, and the maximum time is accepted as one week, a time during which airstreams on the troposphere boundary encircle the globe.

The uncertainty in the local weather forecast follows from the random character of atmospheric processes due to transport of heat and moisture. Indeed, dynamics

of the Earth's atmosphere is determined by mass and energy transport [7–9], and a turbulent character of motion results from collision of air streams [10, 11], creating a random factor in the atmospheric development at a given locality. Such a character of air evolution in the atmosphere determines the climate [12, 13] and its change in time, as well as other atmospheric phenomena including the electric ones [14–20].

Thus, in predicting the weather and climate we are based on the values of initial atmospheric parameters and on the deterministic character of atmospheric processes. An error in such a description of atmospheric parameters due to random processes in the atmosphere is accumulated in time. As a result, deterministic weather prediction is not reliable for large periods of time. Empiric methods may be used for longer times. As for electric phenomena, their predictions may be more rough. For example, it is impossible to predict at what time a thunderstorm will be observed in a certain locality, and to which house a thunderbolt will come. Nevertheless, the above problems of variation of atmospheric parameters, including electric ones, are the object of meteorological investigations. Our analysis in this book related to global atmospheric phenomena is based on the concepts of standard atmosphere and the global electric circuit. This means that we use averaged parameters over the globe and time. In this consideration the information which is of interest for meteorology is lost. (In particular, the Earth's temperature equals 288 K and does not vary in time and over the globe). But such a description is convenient in the analysis of global processes that are the object of this book. Such a simplification allows us to extract the processes and mechanisms that determine a certain atmospheric property. In addition, analysis on the basis of averaged parameters has a qualitative character. Therefore in this analysis one can use estimations for some atmospheric parameters.

Convective air motion may be represented as a sum of individual vortices, and below we estimate the size of an elementary vortex in a middle troposphere assuming that convection results from the Rayleigh–Taylor instability [21–25] under the action of the gravitational force and temperature gradient. If conditions of the convective instability are fulfilled in a region between two heated plates, structures such as Bennard cells [26–28] are formed near the threshold of this instability. If a convection motion proceeds in an infinite space, such structures are mixed, leading to a random convective air motion in the form of vortices of different sizes. Let us estimate the minimal vortex size on the basis of the criterion that the Rayleigh–Taylor instability arises and develops, if the Rayleigh number is $Ra \sim 1000$ [21, 22, 29].

The Rayleigh number for a vortex of size l , that is, a size of the elementary air cell, has the form [21]

$$Ra = \frac{dT}{Tdh} \frac{gl^4}{\nu\chi} = \frac{gl^4}{\Lambda\nu\chi} \quad (2.5)$$

Here $g = 980 \text{ cm/s}^2$ is the freefall acceleration at the Earth's surface, $\nu = \eta/\rho = 0.3 \text{ cm}^2/\text{s}$ is the air kinematic viscosity, $\rho = 0.6 \text{ kg/m}^3$ is the air density, $\Lambda = Tdh/dT \approx 44 \text{ km}$ is a distance that characterizes a temperature variation in the tro-

posphere, and we are guided by altitudes where the air pressure is $p = 0.5$ atm. Next, $\chi = \kappa/Nc_p = 0.4 \text{ cm}^2/\text{s}$ is the air thermal diffusivity, so that N is the number density of air molecules, and $c_p = 7/2$ is the heat capacity per molecule. This gives for a typical vortex size

$$l \approx \left(\frac{Ra \cdot \nu \chi \Lambda}{g} \right)^{1/4}, \quad (2.6)$$

From this one can estimate the minimal vortex size in the atmosphere that is equal to $l \sim 20$ cm under the above conditions at altitudes where the pressure is $p = 0.5$ atm. This quantity depends on the number density of air molecules N as $l \sim 1/\sqrt{N}$, so that near the Earth's surface $l \sim 10$ cm.

One can estimate the velocity v_l of motion for a vortex of a size l from the Navier–Stokes equation

$$\eta \frac{\partial^2 v}{\partial h^2} \sim \eta \frac{v_l}{l^2} \sim PN', \quad (2.7)$$

where η is the air viscosity, $P = mg$ is the molecule weight (m is the molecule mass, g is the free fall acceleration), $N' \sim Nl/\Lambda$ is the difference of the molecule number density at a distance l between two air layers, and N is the number density of air molecules ($N' \ll N$). From this we have for the vortex velocity v_l of a size l

$$v_l \sim \frac{mgNl^3}{\Lambda\eta}, \quad (2.8)$$

and the diffusion coefficient for transfer of such vortices is estimated as

$$D_l \approx \frac{v_l l}{3} \sim \frac{mgNl^4}{3\Lambda\eta} \quad (2.9)$$

Applying these results for the troposphere at altitudes approximately 5 km, where the air pressure is $p = 0.5$ atm, one can obtain there $v_l \sim 20$ cm/s, and the diffusion coefficient for air transport through vortices of a minimal size equals $D_l \sim 100 \text{ cm}^2/\text{s}$ for the troposphere. The Reynolds number is equal for this vortex size $Re = v_l l/\nu \sim 10^3$ that has the same order of magnitude for gasses as the Rayleigh number Ra [25].

Air transport over large distances that significantly exceed a size L of the largest elementary vortex proceeds through turbulent motion. Transition to the turbulent character of motion corresponds to the critical Reynolds number; that is, $Re_{cr} = v_L L/\nu = 2 \times 10^5$ [21]. From this one can obtain for the diffusion coefficient of atmospheric air through large vortices

$$D_L \approx \frac{Re_{cr}\nu}{3} \sim 2 \times 10^4 \text{ cm}^2/\text{s} \quad (2.10)$$

In addition, Formulas (2.8) and (2.9) allow one to estimate the size of large vortices in accordance with

$$L \approx \left(\frac{3\Lambda\eta D_L}{mgN} \right)^{1/4} \quad (2.11)$$

Applying these formulas to the troposphere, as before, we obtain $L \sim 60\text{cm}$, $v_L \sim 10^3\text{ cm/s}$. Thus, we obtain roughly for the ratio of sizes for large and small vortices in the course of air convection in the troposphere $L/l \approx 3$. One can see that transport of atmospheric air and its components over large distances is realized through vortices of a maximum size, and the appropriate diffusion coefficient for tropospheric air is estimated as $D_L \sim 2 \times 10^4\text{ cm}^2/\text{s}$.

Thus, mass and energy transport in atmospheric air over large distances has a random character [7–9] that is created by collisions of air fluxes resulting from the mixing of air layers [10, 11]. Because of the random character of the transport process, one can characterize this transport by the diffusion coefficient (2.10). This diffusion coefficient varies with altitude, and Fig. 2.7 gives the turbulent diffusion coefficient for air of the standard atmosphere, where observational data for air transport are treated on the basis of the diffusion character of its motion.

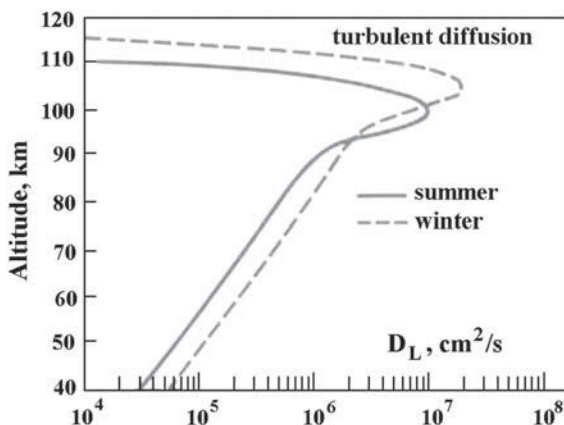
In the regime realized in the troposphere we have the following dependence of parameters of the largest vortices on the number density N of air molecules

$$L \sim \frac{1}{\sqrt{N}}, \quad D_L \sim \frac{1}{N}, \quad v_L \sim \frac{D_L}{L} \sim \frac{1}{\sqrt{N}} \quad (2.12)$$

This similarity law holds true until the velocity v_L of vortices is small compared to the thermal velocity of air molecules or the sound speed c_s . This takes place in the low atmosphere

$$N \gg N_o, \quad (2.13)$$

Fig. 2.7 Altitude dependence for the coefficient of turbulent diffusion [30, 31]



where $N_o \sim 10^{16} \text{cm}^{-3}$ which corresponds to altitudes $h < 55 \text{ km}$; that is, the regime under consideration is fulfilled in the mesosphere and lower layers.

In considering transport of air at higher altitudes, one can assume the velocity of the largest vortices to be equal to the sound speed $v_L \approx c_s \approx 3 \times 10^4 \text{ cm/s}$. Because the critical value of the Reynolds number $Re_{cr} = v_L L / \nu = 2 \times 10^5$ [21] corresponds to the largest vortices, one can obtain the similarity law instead of (2.12)

$$L \sim \frac{1}{N}, \quad D_L \sim \frac{1}{N}, \quad v_L = c_s, \quad (2.14)$$

and the largest vortex size is

$$L = \frac{Re_{cr} \nu}{c_s} \quad (2.15)$$

In particular, for the altitude $h = 100 \text{ km}$ this formula gives ($\nu = 30 \text{ m}^2/\text{s}$ [1]), which corresponds to $L \sim 10 \text{ km}$ as it follows from treatment of observational data (see Fig. 2.7). Next, Formula (2.10)

$$D_L \approx \frac{Re_{cr} \nu}{3} \quad (2.16)$$

gives at the altitude $h = 100 \text{ km}$ $D_L \sim 2 \times 10^6 \text{ cm}^2/\text{s}$. This value is less than that following from treatment of observational data $D_L \sim 1 \times 10^7 \text{ cm}^2/\text{s}$ [32]. Note that in this case $L \sim \Lambda$; that is, the diffusion character of air transport for large distances is violated.

In the above consideration convection is created by air molecules, that is, nitrogen and oxygen molecules. Therefore the average mass of these molecules ($m = 29$ in units of atomic mass) is included in Formulas (2.8) and (2.9). Other components and admixtures do not contribute to convective air motion. Collisions between molecules are important; for example, the mean free path of air molecules at the altitude 100 Km is 0.14 m [1], that is, small compared with values L and Λ responsible for convection. Therefore components of air are captured by its motion and are mixed with air. As a result, the concentration of air admixture does not depend on the altitude h up to $h \sim 100 \text{ km}$. One can add to this that transport of air in this consideration consists of two parts, namely, diffusion motion of air under the action of convection and drift motion of molecules due to gravitational forces. In this consideration displacement of molecules owing to gravitational forces is weak, so that one can neglect the action of the molecule weight on the convection process.

In order to give the criterion of the convection regime under consideration, where air vortices lead to mixing of molecules of air and its components, let us introduce an effective drift velocity w_L of a given component due to convection, which follows from the expression for a flux j of this component:

$$j = -D_L \nabla N = w_L N, \quad (2.17)$$

where N is the number density of particles for this component, and its drift velocity is

$$w_L \sim \frac{D_L}{\Lambda}, \quad (2.18)$$

For mixing of this component with air as a result of convective air motion, this drift velocity w_L must be large compared to the drift velocity w_g under the action of the gravitational force (particle weight). The latter is equal according to the Einstein relation [33–35]

$$w_g = \frac{Dm g}{T} \quad (2.19)$$

Here m is the mass of an admixture particle, $g \approx 980 \text{ cm/s}^2$ is the freefall acceleration, T is the air temperature given in energetic units, and D is the diffusion coefficient of a given atomic particle in air. Within the framework of the Chapman–Enskog approximation [36–38], the hard-sphere model for colliding atoms and molecules is given by [38, 39]

$$D = \frac{3\sqrt{\pi T}}{8\sqrt{2\mu N_a \sigma_g}}, \quad (2.20)$$

where $\mu = m_1 m_2 / (m_1 + m_2)$ is the reduced mass of colliding particles of masses m_1 and m_2 , N_a is the number density of air molecules, and σ_g is the gas-kinetic cross section for colliding atoms or molecules. Let us calculate the drift velocity for oxygen atoms located in the atmosphere. In this case from Formulas (2.13) and (2.14) we have (the gas-kinetic cross section at room temperature is $\sigma_g = 25 \text{ \AA}^2$ [39])

$$N_a w_g = 1.5 \times 10^{13} \text{ cm}^{-2} \text{ s}^{-1}; \quad (2.21)$$

From this we have for the troposphere $w_L \sim 0.03 \text{ cm/s}$, and $w_g \sim 10^{-6} \text{ cm/s}$, whereas at the altitude $\sim 100 \text{ km}$ which is roughly the boundary of the convective air motion, we obtain for air molecules $w_L \sim 3 \times 10^4 \text{ cm/s}$, and $w_g \sim 1 \text{ cm/s}$. Hence, the criterion

$$w_L \gg w_g, \quad (2.22)$$

that allows us to neglect the action of gravitational forces on the convective character of motion of air molecules in the Earth’s atmosphere, holds true for air molecules.

2.3 Atmospheric Water

Water is an important component of the Earth’s atmosphere, in the first place, because electrical atmospheric processes accompany water circulation in the Earth’s atmosphere [40]. Within the framework of the global circuit model for atmospheric electricity [41], where the Earth and its atmosphere are considered as a whole in some electric scheme, we represent parameters of water and transport through the atmosphere to be averaged over time and to be summed over the globe. The rate of

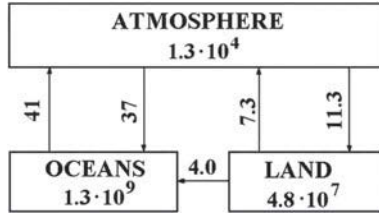


Fig. 2.8 Rates of water transitions between oceans, Earth’s land, and the atmosphere expressed in units $10^4 \text{ km}^3/\text{year}$ are indicated at corresponding *arrows*. An amount of water in these species is given in km^3 inside *rectangles*

water evaporation from the Earth’s surface and its passage through the atmosphere is on average $4.8 \times 10^5 \text{ km}^3/\text{year}$ [13, 42, 43] and the same water amount returns to the Earth in the form of precipitation. Here we take the volume of liquid water as a unit, as usual. Transferring to standard units, we obtain for the rate of water evaporation from the Earth’s surface to the atmosphere as well as the rate of water precipitation on the Earth’s surface from the atmosphere as $4.8 \times 10^{20} \text{ g}/\text{year}$ or $1.5 \times 10^{13} \text{ g/s}$. An average content of water in the atmosphere is $1.3 \times 10^{19} \text{ g}$ [44–47], compared to the total air mass $5.1 \times 10^{21} \text{ g}$ which is located in atmospheric air on average. From this one can find the residence time of water in the atmosphere which is approximately 9 days. If atmospheric water in the liquid state is precipitated uniformly over the Earth’s surface, it forms a layer of the depth of 2.5 cm [48]. In addition, the Earth contains approximately $1.4 \times 10^{24} \text{ g}$ of water, and 96 % of this water is saline. This includes approximately $1.0 \times 10^{22} \text{ g}$ of fresh groundwater. Figure 2.8 shows the rates of exchange by water between the atmosphere, oceans, and land, as well as the water amount inside each component. Note also that the rate of snow precipitation on the Earth’s surface is $1.0 \times 10^3 \text{ km}^3/\text{year}$ [49] compared to $4.8 \times 10^5 \text{ km}^3/\text{year}$ for liquid water. The accuracy of the above data is 20 %, which follows from comparison of data from different sources.

As follows from the above parameters, the average water vapor mass density in the atmosphere near the Earth’s surface is $3 \text{ g}/\text{m}^3$ in comparison with the air mass density $1.2 \text{ kg}/\text{m}^3$ for dry air. An indicated water-specific density in air corresponds to the partial water pressure 2 Torr near the Earth’s surface. The saturated vapor pressure of water at the temperature 0°C is 4.7 Torr [4, 50]. One can conclude from this that water is located in the Earth’s atmosphere in the form of molecules, and condensation of water vapor in regions with a low temperature leads to formation of aerosols, micron-size water particles. Because the condensation process proceeds at temperatures near the water melting point, water may be present in the atmosphere in the form of aerosols where water is found in any aggregate state.

In considering the atmospheric electric current as a secondary phenomenon of water circulation in the atmosphere, we find the ratio of the atmospheric electric current due to lightning ($I = 1700 \text{ A}$ [17, 51]) to the transferring mass through the atmosphere per unit time $dM/dt = 1.5 \times 10^{13} \text{ g/s}$, that gives the following specific transferring mass [52, 53]

$$\frac{I}{dM/dt} = 1.2 \times 10^{-10} \frac{C}{g} \quad (2.23)$$

We are guided by this value in the analysis of atmospheric electric processes.

The temperature of atmospheric air decreases with increasing altitude up to the tropopause [47] that is located at the altitude 9 km for poles and 17 km for the equator where the temperature has a minimum. The average temperature of the Earth's surface is $15^\circ C$, the tropopause temperature is $-55^\circ C$, and the temperature gradient is -6.5 K/km [54]. From this one can see that water condensation proceeds at altitudes, at least, above 2–3 km. Forming micron-size aerosols are liquid, snow, or ice microparticles, and, possibly, a mixture of these forms.

Table 2.1 [4] contains the temperature dependence for the water-saturated vapor pressure and other water parameters, so that water microparticles are formed at pressures which exceed the saturated vapor pressure at a given temperature. Because only water aerosols are responsible for electrical atmospheric phenomena, these processes take place if the water content exceeds a certain threshold. In particular, it is usually accepted that thunderstorm weather is possible if the water vapor mass mixing ratio exceeds 7 g/kg (grams of water per kilogram of air) [55]. This water content corresponds to the saturated vapor pressure at the temperature $8^\circ C$.

If the partial water pressure exceeds the saturated vapor pressure, its excess is transformed in a condensed phase. This process in a space starts from formation of nano-size water clusters as a result of attachment of water molecules to condensation nuclei, that is, to ions or chemically active atomic particles, in particular, to radicals. Subsequent growth of small water particles by attachment of water molecules leads to formation of water aerosols, that is, micron-size water particles. Water aerosols may be found in different aggregate states depending on the air temperature where their growth takes place. Namely, water aerosols may be liquid drops, ice particles, snow particles, and a mixture of snow and water, that is, graupels. Aerosols in atmospheric air constitute clouds of various types. In considering processes involving aerosols, we are guided by aerosols in cumulus clouds which take part in atmospheric electricity and finally cause the origination of lightning. A mature cumulus cloud contains aerosols with the following averaged parameters [56–59].

$$r = 8\mu m, \quad N_p = 10^3 \text{ cm}^{-3}, \quad (2.24)$$

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 is $2g/m^3$, and this corresponds to the specific water mass 1.6 g/kg under standard conditions (the temperature $0^\circ C$ and pressure 1 atm). If we take these aerosols from the atmosphere and transform them in liquid water, the ratio of the thickness of the water layer h to the cloud thickness L is $h/L = 2\text{mm/km}$.