

MINISTRY OF EDUCATION AND SCIENCE  
OF THE RUSSIAN FEDERATION  
NATIONAL RESEARCH  
TOMSK STATE UNIVERSITY

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# **BASIC PHYSICAL METEOROLOGY**

**Textbook**

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Tomsk  
TSU Press  
2023

UDC 551.51 (075.8)  
BBC 26.23ya73  
R93

**Rybakova, Zh. V.**  
R93 Basic Physical Meteorology : textbook. – Tomsk : TSU Press,  
2023. – 184 p.  
ISBN 978-5-907572-05-8

This textbook includes the main sections of physical meteorology, which is a basic subject in the system of training specialists in the field of meteorology. The language of instruction is English. This makes it possible for students who do not speak Russian well enough, but speak English, to study the main concepts of physical meteorology.

The textbook is for students of meteorology and students of related specialties.

UDC 551.51 (075.8)  
BBC 26.23ya73

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ISBN 978-5-907572-05-8

Zh. V. Rybakova, 2023

This textbook is dedicated  
to the blessed memory of  
***Nina Polikarpovna Fast,***  
Associate Professor of Tomsk State University,  
my first teacher in meteorology

and

***Lyudmila Nikolaevna Bimberekova (Rybakova),***  
Senior Lecturer at Novosibirsk Aerospace Lyceum,  
my mother and my first English teacher

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# **1. ELEMENTARY METEOROLOGY**

## **1.1. Meteorological quantities and atmospheric phenomena. Meteorological gradients**

Meteorology is the science of the atmosphere, the outer shell of our planet, its composition and structure. Physical meteorology is the physics of the atmosphere. Atmospheric physics is the science of the physical state of the atmosphere, of the physical processes occurring in it. The task of physical meteorology is to establish a diagnosis of the atmosphere from the available data. This diagnosis is connected with the definition of its condition. Without such a diagnosis, it is impossible to make a forecast of the state of the atmosphere, that is, to predict its state in the future.

There are two important concepts underlying the physical state of the atmosphere. They are meteorological quantities and atmospheric phenomena. These concepts represent the quantitative and qualitative characteristics of the atmosphere.

Meteorological quantities include atmospheric pressure, air and underlying surface temperature, air humidity, wind speed and direction, cloud amount, precipitation amounts and other quantitative characteristics of the state of the atmosphere.

Atmospheric phenomena include clouds, fogs, precipitation, as well as squalls, snowstorms, thunderstorms, lightnings, rainbows, haloes, crowns, mirages and other qualitative characteristics of the state of the atmosphere.

Atmospheric phenomena are the result of a sharp change in at least one meteorological quantity during a certain process. Atmospheric phe-

nomena are associated with a number of meteorological quantities that characterize these phenomena. For example, in clouds (atmospheric phenomena), such meteorological quantities are the height of their lower boundary, water content, and power. Another example: when we say “hail”, we mean an atmospheric phenomenon. If we are talking about a measured amount of hail, we mean a meteorological quantity. Of course, both meteorological quantities and atmospheric phenomena are determined by the nature of atmospheric processes.

The totality of meteorological quantities and atmospheric phenomena at a certain point in time or for a short period of time is called weather. The characteristic weather pattern in a certain area is called climate.

Meteorological quantities and atmospheric phenomena change in time and space.

The determination of changes in meteorological quantities over time serves to establish the prognostic characteristics of these quantities (their values for a certain time ahead).

Changes in meteorological quantities in space create fields of these quantities. A detailed presentation on this subject is given in [2]. In this textbook, we will consider only the concept of the gradient of meteorological quantities. The gradients of meteorological quantities determine their fields. The field of a meteorological quantity is created by a set of its values in space. In such fields, the gradients of meteorological quantities are a measure of the variability of these quantities in space.

The gradient of meteorological quantities can be decomposed into their changes in different directions. The change in meteorological quantities in all directions is determined by the total gradient. The total gradient of a meteorological quantity is the total derivative of a meteorological quantity along the normal to the surface of equal values of this quantity in its field. This total derivative is taken with a minus sign:

$-\frac{dF}{dN}$ , where  $F$  is an arbitrary meteorological quantity, and  $N$  is the normal to the surface of equal values of the meteorological quantity.



In physical meteorology, the total gradients of meteorological quantities are usually made by vertical gradients and horizontal gradients. Such gradients are partial derivatives of meteorological quantities in vertical and horizontal directions with a minus sign.

It is important to note that there is always a minus sign in the recording of gradients of meteorological values. This is due to the fact that in cases of a full gradient, the change in meteorological quantities is taken in the direction of their decrease. In this direction, the changes in these quantities have negative values. The distance increment along the normal  $N$  is always a positive value. As a result, a negative change in the value of a meteorological quantity, in combination with a minus sign in front of the derivative, is a positive value.

Thus, the total gradients of meteorological quantities are always positive.

As for horizontal gradients of meteorological quantities, they are also always positive. This is due to the fact that changes in these quantities in the horizontal direction are always taken in the direction of their decrease.

Vertical gradients can be both positive and negative, and equal to zero. In the latter case, the meteorological value does not change with altitude. With a positive value of the vertical gradient of a meteorological quantity, its values decrease with height. With a negative value of the vertical gradient of a meteorological quantity, its values increase.

Gradients of meteorological quantities help to judge the processes taking place in the atmosphere, the possibility and nature of the formation of certain atmospheric phenomena. The fields of meteorological quantities can determine the state of the atmosphere (whether it is stable or unstable and to what extent, how high the wind speeds are). This is a decisive factor for representing the direction and rate of development of the atmospheric process.

The plan for considering meteorological quantities and atmospheric phenomena is given in [2].

***Test questions:***

1. What is weather?
2. What is climate?
3. What are the main differences between these concepts?

## **1.2. Composition of the atmosphere**

In its chemical composition, the atmosphere has numerous gas components and various aerosols. Aerosols are solid and liquid particles suspended in the air. In other words, air is a mechanical mixture of gases and aerosols.

The share of all components in the total composition of the atmosphere is different. The properties of these components are also different. Various numerous components in the lower layers of the atmosphere can be divided into four groups. The first of these are the main, almost constant components. These include molecular nitrogen, molecular oxygen and argon. In total, the volume content of these gases accounts for 99.964% (nitrogen 78.084%, oxygen 20.946%, argon 0.934%). The gases of this group in the lower layers practically do not absorb the radiant energy of the sun. In the high layers of the atmosphere, oxygen and nitrogen absorb the radiant energy of the sun. The gases of this group do not absorb the long-wave radiation energy of the underlying surface. So, the general properties of the gases of the first group are: 1) a large proportion in the total mixture of gases; 2) a constancy of their ratios up to high altitudes.

The gases of the second group have a high absorbing capacity for radiant energy. In this they differ from the gases of the first group. The second group includes important variable components — water vapor, carbon dioxide and ozone. Carbon dioxide and water vapor absorb longwave radiation. The main absorber of longwave radiation is water vapor. It has the largest number of absorption bands in the indicated range, a large width of these bands, and a high absorption coefficient in

these bands. In addition, an amount of water vapor at high air temperatures can be quite noticeable. The volume content of water vapor varies from almost zero to values of approximately 4%. An increase in the content of water vapor occurs due to a decrease in gases from the group of constant components (the first group).

The second gas in this group is carbon dioxide. This gas accounts for 0.033% of the total volume (in 1900, the volume content of this gas was 0.029%). It should not be thought that this increase is insignificant. Even the smallest increase in the content of this gas leads to a noticeable increase in the greenhouse effect. As a result, the temperature of the air and the underlying surface rises, and the areas occupied by snow and ice are reduced. Ultimately, the area of water on Earth increases and the surface of land shrinks.

The third gas of the second group is ozone. In the lowest layers of the atmosphere, ozone is scarce, and in these layers, ozone is an air pollutant. From about 10 km, the ozone content increases with height. In the 20–50 km layer, ozone is an important protector of all life from short-wave radiation, a “shield” for life on Earth. The flora and fauna of our planet are not ready for the effects of penetrating (biologically active) radiation. Ozone absorbs this radiation. If such radiation reached living objects, their cells would either die or become transformed. It is important to note that ozone can absorb radiant energy in other parts of the spectrum as well. So ozone absorbs in the region of visible and longwave radiation. As for shortwave radiation, it is “cut off” by ozone in the layer of its highest content in the range of 0.22–0.29 microns. For this reason, such penetrating radiation does not reach the troposphere at the current ozone content and does not affect us. Ozone depletion is caused by nitrogen oxides and chlorides.

The third group in the atmosphere consists of numerous gases that do not have a large share in the total composition of the lower layers. In other words, the gases of this group are impurities in the lower atmosphere. It should not be thought that these gases do not play a large

role in life on Earth. Let us take methane for example. This gas absorbs the longwave radiation of the underlying surface, therefore it is one of the components of the greenhouse effect. Another example is radon and iodine. Radon is an air ionizer. With an insufficient amount of radon, the formation of the number of ion pairs in a volume unit per unit of time decreases. Depending on the concentration, radon and iodine can be both beneficial and harmful to humans. Radon affects immune, reproductive, hematopoietic cells. A decrease in iodine content leads to dysfunction of the thyroid gland.

The fourth group is not gases, but aerosols. They can be airborne liquid or solid particles of various sizes, shapes, colors, odors, chemical compositions and physical properties. Aerosols have a wide variety of origins. They can be particles of soil, sea salt, weathered rocks, volcanic dust, smoke, clouds and fogs, microorganisms, plant pollen and other airborne particles.

The composition of the air changes with height. With the distance from the underlying surface (which is the source of aerosols) the number of suspended air particles decreases. With height, the content of impurities and even gases of the second group also decreases. At the same time, with a decrease in air density with height and, consequently, a decrease even in the content of gases of the first group, the quantitative ratios between the main gases (nitrogen, oxygen, argon) remain almost constant. This constancy can be traced to high altitudes — up to about 100 km. It is the layer from the underlying surface to the level of 100 km above it that is considered the lower layer when dividing the atmosphere according to the composition of the air. The reason for the preservation of the main composition of the air in such a thick layer is the intensive vertical mixing of air layers.

Above the level of 100 km, there is a significant change in the composition of the air with height. In layers so high from the underlying surface, two reasons for this change can be traced. The main reason is the active dissociation of molecular oxygen above 100 km. From a

height of 250–300 km, dissociation of molecular nitrogen is also traced. Another reason is the manifestation at high altitudes of the process of gravitational separation of gases in height (heavier gases are located lower than lighter ones).

Under the influence of these two factors, in the 300–1000 km layer, the main composition of the air is atomic oxygen with a small admixture of atomic nitrogen. Above the level of 1000 km the atmosphere consists mainly of the lightest gases. These gases are helium and hydrogen.

Thus, from the underlying surface to a level of 100 km, the composition of the air, without taking into account impurities and variable components, is practically the same. The composition of the air above 100 km changes with altitude.

The lower layer of the atmosphere (up to the level of 100 km) is called the homosphere. The layer above 100 km is called the heterosphere.

Light gases in the uppermost atmosphere have weak connections of a gravitational nature with the center of mass for the Earth. For this reason, they can leave the limits of the atmosphere and go into outer space. There they become part of the interplanetary gas.

In the highest layers of the atmosphere, particles are strongly influenced by the fluxes of radiant energy from the sun. Under this influence, molecular particles decay into atomic ionized particles. Ionization consists in the fact that the outer valence electron absorbs some of the energy and leaves its neutral system. As a result, a positive ion and a free electron are formed. An electron can join a neutral particle and form a negative ion. Thus, in the layers of the atmosphere above about 60 km there is a significant amount of ions of both signs and free electrons.

***Test questions:***

1. What gases are the main components of the atmosphere in the lower layers?

2. What gases and why play the greatest role in the thermal regime of the underlying surface and atmosphere?
3. What is the trend of change in the composition of air in the lower layers of the atmosphere? The content of what gases has changed significantly in recent decades?
4. What are the possible consequences of these changes?
5. What are aerosols?
6. What causes changes in air composition with altitude?
7. What is the composition of gases in the highest layers of the earth's atmosphere?

### **1.3. The structure of the atmosphere**

The atmosphere is heterogeneous in all directions. Especially heterogeneity can be traced in the vertical direction. The heterogeneity of the atmosphere in horizontal directions and those close to them is manifested in the presence of heterogeneous air masses.

Air masses are formed over such types of the underlying surface that are not identical in terms of thermo-hygrometry (temperature and humidity). Air masses can stay above these surfaces for a certain time and receive various properties from them.

Air masses are large volumes of air with almost the same properties. The contact between heterogeneous air masses leads to the formation of transition zones between them. In these zones, from one border to another, meteorological quantities change very quickly. In other words, when moving from one air mass to another, the horizontal gradients of meteorological quantities increase quite noticeably. Within the same air mass, these horizontal gradients are small. These transitional zones between two air masses are called frontal zones. If such zones are crossed by any altitude levels, for example, sea level or a level of 5 km, there will be surfaces at the intersection. These surfaces are much narrower than the vast air masses. So they are called atmospheric fronts, that is, lines between air masses.

Air masses determine the weather conditions of the area over which they are or pass. The passage of an atmospheric front indicates a change in air masses. A change in the air mass leads to a change in weather conditions.

Of course, weather conditions can change gradually, as a result of the transformation of the air mass. The transformation of the air mass is a change in its properties under the influence of the underlying surface. For such a change to take place, the air mass has to be above the underlying surface for some time or move above it at a low speed. However, during the passage of an atmospheric front, weather conditions change quite quickly, since one air mass is replaced by another.

The atmosphere, as already mentioned, is most heterogeneous along the vertical. That is why the atmosphere is divided into layers in the vertical direction. There are several principles for such division. The first and main of them is the principle of dividing the atmosphere according to the nature of the change in temperature with height. According to this principle, five layers are distinguished: troposphere, stratosphere, mesosphere, thermosphere and exosphere. These layers are separated by narrow intermediate layers called pauses. The pauses are named after the underlying large layers. Such pauses are tropopause, stratopause, mesopause, thermopause. Above the exosphere is the interplanetary space.

All pauses have a common characteristic property: the temperature in them practically does not change with height. In other words, the temperature profile in the pauses is isothermal.

In the main layers, the temperature changes with height. This change occurs in different directions from the underlying layer. So in the troposphere, the temperature generally decreases with height, in the stratosphere it rises, in the mesosphere it decreases, in the thermosphere it rises. The density in the exosphere is very low; therefore, it is assumed that the air temperature in it increases with height, but more slowly than in the thermosphere.

The drop in temperature with height is understandable. This is due to the absorption by the atmosphere of mainly longwave radiation from the underlying surface. For this reason, the largest part of this radiation is absorbed in the lowest layer due to its closer location to the radiation surface and its greater density. In other words, each subsequent overlying layer receives the remainder of the heat flux, which has been partially absorbed by the underlying layer. In addition, absorption by each higher layer occurs with an ever smaller number of air particles, since its density decreases with height. As a result, the air temperature drops with height. The specified distribution of temperature with height is valid, first of all, for the troposphere. There, the indicated drop is significant and is a characteristic property of this layer.

In the higher mesosphere, the decrease in temperature with height is slower than in the troposphere. In the stratosphere and thermosphere, the temperature rises with height. In the stratosphere, this is due to the absorption of radiant energy by ozone. The main layer of ozone is located in the stratosphere. In the thermosphere, short-wave radiation is absorbed mainly by oxygen. In this case, oxygen undergoes a process of dissociation.

The thickness of the main layers of the atmosphere is different. The narrowest layer, the troposphere, extends from the underlying surface to about 11 km in the middle latitudes and to about 17 km in the low latitudes. In general, the power (thickness) of the troposphere increases with the growth of solar radiation fluxes to the underlying surface. The greater the specified flux from the sun, the higher the heated underlying surface will transfer the received heat. As a result, the troposphere is more powerful at lower latitudes than at middle latitudes, and especially high latitudes. The troposphere is higher in summer than in winter. It is higher during the day than at night.

Of course, in the lowest part of the troposphere, the nature of the change in temperature with height may differ from that described above. In the lower troposphere, the temperature profile is determined



by the thermal regime of the underlying surface. If it heats up, then the air temperature drops with height. If the underlying surface cools, then the atmosphere gives it its heat. In this case, the greatest heat flux comes from the lowest layers. They are closer to the cooling underlying surface and have a higher density. It is clear that in these layers there are more particles that give off heat to the underlying surface. As a result, the lower layers of the atmosphere are colder. Higher layers give off less heat and are warmer. This distribution of temperature with height is called inversion. In addition to the usual distribution of temperature with height (falling along the vertical) and the inverted distribution (temperature increase with height), there is, as mentioned above, an isothermal distribution (temperature constancy along the vertical). Isothermal distribution occurs not only in intermediate layers — pauses — but also in other cases.

Such cases occur when changing from the usual distribution to the inverted distribution or vice versa. Indeed, when the underlying surface warms up in the morning, a drop in temperature with height is established. This fall intensifies towards the noon hours. The largest positive vertical temperature gradient is observed at a constant atmospheric process during the greatest heating of the underlying surface. This happens at the highest position of the sun. In the evening, the heating of the underlying surface is weakened. As a result, the temperature drop decreases with height. At some point, the temperature of the underlying surface and the layer of the air adjacent to it becomes approximately equal to the temperature in the higher layers of the atmosphere. This is the temporary establishment of an isothermal distribution of temperature with height.

Subsequently, the temperature of the underlying surface becomes lower. The layer of the air adjacent to it is also cooled. This layer gives off its heat to the underlying surface. In such cases, the higher layers are warmer than the lower layers. Thus, an inverted temperature distribution with height is established. It is typical for the night hours.

In the cold period of the year, the inverted distribution can also be observed at other hours (sometimes throughout the whole day and night). In the morning, when the sun rises and the underlying surface warms up, the inverted temperature distribution with height changes first to isothermal, and then to the usual temperature distribution with height (temperature drop along the vertical).

In this way, the “controller” of temperature variation with height in the lower troposphere is the underlying surface, its thermal regime. A characteristic distribution during the day is a drop in temperature with height. In the evening and early morning, the temperature is constant with height, and at night the temperature rises with height.

Above the lower troposphere (above 2 kilometers), the temperature usually drops with height.

Sometimes processes occurring in the atmosphere can lead to a change in the temperature profile. For example, cases of formation of overcloud and undercloud inversions are well known. So stratus and stratocumulus clouds are often sub-inversion type clouds. We will talk about this in the corresponding section.

Of the other properties of different layers of the atmosphere, it is important to note that the lower (0–2 km) and middle troposphere (2–6 km) is the main moisture-bearing layer. The most watery clouds form in this layer. In general, all clouds of the main types are observed precisely in the troposphere.

Very thin iridescent clouds can form in the stratosphere at a height of 22–25 km. They are called mother-of-pearl (or nacreous) clouds [4]. Towards the upper boundary of the stratosphere, the temperature rises to almost zero degrees Celsius. In the lowest stratosphere, the temperature increase may be less pronounced [1, 2]. Mother-of-pearl clouds form under the lower boundary of the stratospheric inversion.

In the mesosphere, the temperature drops with height to its uppermost boundary almost to  $-100\text{ }^{\circ}\text{C}$ . At such a low air temperature, it is possible to achieve the state of saturation with water vapor (when it

condenses into clouds) even with a small amount of it. These low temperatures are observed at the farthest position of our planet from the sun (at aphelion). In the northern hemisphere, this situation develops in summer. In the southern hemisphere, it is observed in winter. It is at the position of our planet at aphelion that the highest clouds appear in the earth's atmosphere. They are called noctilucent or mesospheric clouds. Stars and other celestial objects are visible through them. These clouds are visible in the dark, and in the daytime their brightness is blocked out by the brightness of the sky.

Compared to the troposphere, other atmospheric layers are more powerful. The stratosphere (11–50 km) and mesosphere (50–90 km) are close to each other in terms of vertical extent. The thermosphere is a much more extended vertical layer (90–450 km), and the exosphere includes all other layers of the air above 450 km. The total thickness of the atmosphere is 60–70 thousand km. However, most of the atmospheric mass is concentrated in the troposphere. It ranges from 75% at high latitudes to 90% at low latitudes of the total mass of the atmosphere. This is due to the high density of tropospheric air compared to the density of the air in higher layers. The above applies, most of all, to the exosphere with its low density. In the exosphere, the process of dissipation is observed (the escape of particles from the Earth's atmosphere into space).

The atmosphere can also be divided in the vertical direction according to other principles of division. So the atmosphere can be divided according to the composition of the air and the measured ionic composition, according to the strength of the interaction of the atmospheric layers with the underlying surface, according to its effect on spacecraft.

***Test questions:***

1. Why is the atmosphere divided into layers in the vertical direction?
2. Which of the principles of dividing the atmosphere into layers is the main one?
3. What properties do the layers allocated according to each of these principles have?

## 1.4. Air state equations

The state of the air is determined by three physical parameters. These are pressure, volume (or density) and temperature. The ratio of these parameters using the gas constant is the equation of state for any gas. In these equations, some quantities can be expressed in terms of others, but the equation of state should be recognized in any form. Let us consider the equations of state for an ideal gas, dry air, water vapor, and moist air. After that, we will compare them with each other.

The equation of state for an ideal gas is:

$$PV = RT, \quad (0.1)$$

where  $R$  is the universal gas constant,  $P$  is pressure,  $T$  is temperature,  $V$  is volume.

The equation of state for dry air is<sup>1</sup>:

$$P\nu = R_c T, \quad (0.2)$$

where  $\nu$  is specific volume; all equations in physical meteorology are written for a unit mass; specific volume is the volume of a unit mass;  $R_c$  is the specific gas constant for dry air.

$$R_c = \frac{R}{\mu_c}, \quad (0.3)$$

where  $\mu_c$  is the molar mass of dry air.

The equation of state for water vapor is:

$$e\nu = R_w T, \quad (0.4)$$

where  $e$  is the partial (its own) pressure of water vapor.

The total atmospheric pressure is the sum of the partial pressures of all gases in the atmosphere.

In (1.4)  $R_w$  is the specific gas constant of water vapor:

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<sup>1</sup> Here and throughout the book we will use some Russian letters in formulas to conform to [2].

$$R_{\text{H}} = \frac{R}{\mu_{\text{H}}}, \quad (0.5)$$

where  $\mu_{\text{H}}$  is the molar mass of water vapor.

The equation of state for moist air is obtained by adding the equations of state for dry air and water vapor. It looks as follows:

$$P\nu = R_c T_v, \quad (0.6)$$

where  $T_v$  is the virtual temperature:

$$T_v = T(1 + 0,608 \cdot s) = T\left(1 + 0,378 \cdot \frac{e}{P}\right).$$

The virtual temperature is related to the presence of water vapor in the atmosphere. In (1.6)  $s$  is the specific humidity,  $e$  is the elasticity of water vapor (partial pressure of water vapor),  $s = 0.622 \cdot \frac{e}{P}$ .

Thus, the main types of equations of state for an ideal gas, dry air, water vapor and moist air have the same form. Other types of these equations and derivations of all equations of state are given in [2]. Note that in physical meteorology the equations are written for a unit mass ( $\rho = \frac{1}{\nu}$ ,  $\nu = \frac{1}{\rho}$ ), so, for example, (1.2) can be written as:

$$P = \rho R_c T. \quad (0.7)$$

**Test questions:**

1. What is the equation of state for an ideal gas?
2. What kinds of the equation of state for dry air are there?
3. What are the equation of state for water vapor and the equation of state for moist air?
4. What is virtual temperature?

## 1.5. Basic equation of atmospheric statics. Barometric formulas. Atmosphere models

Atmospheric statics establishes the laws of the structure of the atmosphere and determines the nature of atmospheric processes under the condition that there is no air movement relative to the underlying surface. Interestingly, the laws established in this section are also well confirmed in the motion of the atmosphere.

The main law of this section is expressed by the basic equation of atmospheric statics. This law, despite its simplicity and conciseness, is a very informative equation.

The derivation of the basic equation of atmospheric statics is quite simple. Let us assume that the allocated unit volume of air does not move horizontally. The fulfillment of this condition is required by the principles of atmospheric statics. In this case, horizontal pressure gradients must be equal to zero. Otherwise, the separated volume of air will move towards a lower pressure. Then only pressure forces from the surrounding air and gravity will act on the unit volume of the air. The pressure forces are directed inside this volume. The force of gravity is directed downward.

In physical meteorology, forces are considered to be positive if they are directed upwards (along the positive  $y$ -axis). Forces are negative if they are directed along the negative  $y$ -axis (downwards). In our case, the force of gravity and the force of pressure directed from above into the volume are negative. The pressure force directed from below into the volume is positive. The sum of all three forces acting on the given volume of air is equal to zero. This is dictated by the condition of statics — the volume is at rest, that is, it does not move if no forces act on it or the sum of the forces acting on it is equal to zero.

The force of gravity is  $-mgdz$ , where  $m$  is the mass of air,  $g$  is the acceleration of gravity,  $dz$  is the height of the allocated unit volume of air. The pressure force acting on the lower face of the volume from the

bottom up is equal to the value of atmospheric pressure at the level of this lower face (which is  $P$ ). The pressure force acting on the upper face of the air volume is equal to the value of atmospheric pressure at the level of the upper face and is equal to the value  $-(P + dP)$ , where  $dP$  is the negative increment of atmospheric pressure as it decreases with height. Then the sum of all forces acting on a unit volume of air is:

$$P - (P + dP) - mgdz = 0, \quad (0.8)$$

$$P - P - dP - mgdz = 0,$$

$$-dP = mgdz. \quad (0.9)$$

Because  $m = \rho v$ , and the volume of the air is equal to 1 (unit volume), the mass is numerically equal to the density of the air with different units of measurement of these quantities. Then

$$-dP = \rho g dz. \quad (0.10)$$

Equation (1.10) is the basic equation of atmospheric statics.

From this small equation, the following four conclusions can be drawn:

1. Atmospheric pressure drops with height. This conclusion follows from the fact that on its right side there are positive factors (air density, acceleration of gravity and a positive increment in height when moving up); then the left side must also contain a positive value: the pressure increment must be negative so that its value with a minus sign in front of it ( $-dP$ ) turns out to be positive.

2. The second conclusion determines the reason for the pressure drop with height in the atmosphere. Atmospheric pressure is numerically equal to the weight of an air column of a single cross-section with a height from the upper boundary of the atmosphere to the measurement level. With height, the column of air that exerts this pressure contracts, so the pressure decreases with height.

3. The third conclusion concerns the rate of pressure drop with altitude in the atmosphere; this becomes obvious if we write (1.10) in the

form:  $-\frac{dP}{dz} = \rho g$ , where  $-\frac{dP}{dz}$  is the rate of pressure drop with height.

It is determined by the values of the acceleration due to gravity and air density. The acceleration of gravity varies with altitude within very narrow limits, so the rate of pressure drop with altitude depends mainly on the value of air density. At high density, the rate of pressure drop with height is greater than at lower density. As a result, in low layers, the pressure drops rapidly with height, while in high layers, pressure decreases more and more slowly with height. This affects the power of the atmosphere. Indeed, in high layers it is necessary to travel for very long distances vertically in order to lower the atmospheric pressure even by a very small value.

4. Air density is inversely related to air temperature. At low temperatures, the density has the highest values, and the rate of atmospheric pressure drop increases with height; in warm air at a lower density, the rate of pressure drop slows down with height.

It is possible to obtain integral forms of the basic equation of statics [2] by integrating the right and left parts of the basic equation of statics (1.10). If various assumptions are made in the equations obtained regarding the change in meteorological quantities with height, then these equations are called barometric formulas.

Barometric formulas are the basis of atmospheric models. The following atmospheric models are used in physical meteorology: homogeneous, isothermal, polytropic atmospheres. They differ in the nature of the fall of atmospheric pressure with height (barometric formulas), the nature of the change in density and air temperature with height, the values of the vertical air temperature gradient, power, and the scope of use.

In a homogeneous atmosphere, density does not change with height. In such an atmosphere, pressure and temperature change linearly with altitude. This determines the rapid fall of the given values with height. The vertical temperature gradient in a homogeneous atmosphere is of great importance. It significantly exceeds the average vertical gradient in



the real atmosphere. The rapid drop in pressure with height leads to the rapid completion of a homogeneous atmosphere. Its thickness is about 8 km. Thus, a homogeneous atmosphere is thin. If the entire real atmosphere is considered to be completely homogeneous, then it includes only the troposphere, and not at all latitudes. At low latitudes, the thickness of the troposphere reaches about 17 km. Even at higher latitudes during warm times, the thickness of the troposphere exceeds the thickness of a homogeneous atmosphere. Only at the highest latitudes in winter can this value be comparable with the power values of a homogeneous atmosphere.

Thus, the real atmosphere as a whole cannot be calculated from a homogeneous atmosphere model. However, this model can be not only theoretical. It can be quite real in certain layers and under conditions of a large influx of solar radiation. These conditions are created at low latitudes, in summer, during the daytime, with a cloudless sky, above the underlying surface with low heat capacity. Such surfaces heat up quickly and significantly. In this case, the homogeneous atmosphere model is applicable only to the air layer closest to the strongly heated underlying surface. Only in this layer can there be a very large vertical temperature gradient and the characteristic properties of a homogeneous atmosphere.

An isothermal atmosphere is an atmosphere with a constant vertical temperature. In it, atmospheric pressure and air density decrease unevenly with height. In such an atmosphere, pressure and density change fairly quickly in low layers and much more slowly in high layers. If we imagine a real atmosphere according to the model of an isothermal atmosphere, then such an atmosphere will turn out to be practically infinite.

Where can you use the isothermal atmosphere model? Of course, it's applicable for all layers with an isothermal distribution of temperature with height. In addition, this model can be used for the atmosphere as a whole. Indeed, in the troposphere the temperature drops with height, in the stratosphere it rises, in the mesosphere it falls, in the thermosphere it rises. Thus, in the atmosphere as a whole, the value of the vertical temperature gradient can be considered equal to zero.

A polytropic atmosphere is an atmosphere with a constant value of the vertical temperature gradient. In such an atmosphere, the pressure and density of air decrease unevenly with height (faster in the lower layers and slower in the high layers). The fall with height of the values of these meteorological quantities in a polytropic atmosphere occurs faster than in an isothermal atmosphere. For this reason, the polytropic atmosphere is not infinite. At standard parameters, it is about 42 km. Of course, such an atmosphere would include the troposphere and a significant part of the stratosphere, but not the entire real atmosphere. Thus, one polytropic atmosphere model cannot be used for the entire real atmosphere. At the same time, the real atmosphere can be represented by a set of polytropic atmospheres with different constant values of vertical temperature gradients. To this end, in the temperature profile, it is necessary to single out layers within which the vertical temperature gradient does not change or almost does not change. These layers will of course be different in their thickness.

Thus, all the mentioned atmospheric models can be used for calculations in the real atmosphere. At the same time, the scope of use for each of the mentioned models is different.

If we talk about the barometric formulas of the real atmosphere, then there are several of them. Let us mention the complete barometric formula of Laplace (1.11), and the incomplete barometric formula of Babinet (1.12). For a real atmosphere, formulas (1.13) and (1.14) can also be used. They are very similar to the barometric formula for an isothermal atmosphere (1.15).

$$z_2 - z_1 = B(1 + \alpha t_{cp})(1 + 0,608s_{cp})(1 + a_1 \cos 2\varphi)(1 + a_2 z_{cp}) \lg \frac{P_1}{P_2}, \quad (0.11)$$

$$z_2 - z_1 = B(1 + \alpha t_{cp}) \lg \frac{P_1}{P_2}, \quad (0.12)$$

$$P_2 = P_1 e^{-\frac{g(z_2 - z_1)}{R_c \bar{T}}}, \quad (0.13)$$

and taking into account the humidity of the air:

$$P_2 = P_1 e^{-\frac{g(z_2 - z_1)}{R_c T_v}} \quad (0.14)$$

The barometric formula of an isothermal atmosphere is known:

$$P = P_0 e^{-\frac{gz}{R_c T_0}} \quad (0.15)$$

where  $(z_2 - z_1)$  is the layer for which the calculation is made;  $z_{cp}$ ,  $t_{cp}$ ,  $s_{cp}$  are the average values of the level height, temperature, specific humidity within the layer  $(z_2 - z_1)$ ;  $B$  is the barometric constant;  $\bar{T}$  is the average layer temperature;  $\bar{T}_v$  is the average virtual temperature;  $\varphi$  is the latitude of the place.

**Test questions:**

1. What are the main forces acting in the atmosphere? What is the nature of these forces?
2. What is the basic equation of atmospheric statics?
3. What conclusions can be drawn from the analysis of this equation?
4. What are barometric formulas?
5. In what cases (under what conditions) can one use the homogeneous atmosphere model?
6. How close is the homogeneous atmosphere model to the real atmosphere?
7. When can the isothermal atmosphere model be used?
8. How close is the isothermal atmosphere model to the real atmosphere?
9. What is the scope of use for the polytropic atmosphere model?
10. What is the difference between the complete Laplace barometric formula and the incomplete Babinet barometric formula?

## 2. RADIANT ENERGY IN THE ATMOSPHERE

Various fluxes of radiant energy are observed in the atmosphere in energetic (quantitative) and qualitative terms (in different regions of wavelengths). Indeed, radiation fluxes from the sun and radiation fluxes from the underlying surface of our planet enter the atmosphere. In addition, radiation fluxes from other volumes of the atmosphere pass through each volume of the air. All these flows are not only absorbed by the underlying surface and the atmosphere, but are also reflected from these media. In this case, the radiation fluxes from the sun are short-wave radiation. The radiation fluxes of the underlying surface and the atmosphere are long-wave radiation. The radiation of any surface of an object or medium falls mainly on a certain region of the spectrum. This is easy to understand based on Wien's laws for a black body:

$$\lambda_{\max} = \frac{c'}{T}, \quad (1.1)$$

$$B(T, \lambda_{\max}) = c'' T^5. \quad (1.2)$$

The first of these laws (2.1) is called Wien's displacement law. It shows the shift of the wavelength, which accounts for the maximum radiation of a black body with a change in its temperature. According to (2.1), with an increase in the temperature of the radiating body, the wavelength that accounts for the maximum radiation of an absolutely black body ( $\lambda_{\max}$ ) is shifted towards shorter wavelengths.

In (2.1)  $c'$  is a constant equal to the value of  $0.28976 \cdot 10^{-2} \text{ m} \cdot \text{K}$ ,  $T$  is the temperature on the Kelvin scale. Wien's second law (2.2) determines the radiation flux of a black body in the region of its maximum radiation depending on temperature. In (2.2)  $B(T, \lambda_{\max})$  is the flux in

the region of maximum radiation,  $c''$  is a constant equal to the value of  $1.301 \cdot 10^{-5} \text{ W/m}^3 \cdot \text{K}^5$ .

From analysis of (2.1) it follows that the hot sun sends out short-wave radiation. The underlying surface and the atmosphere of our planet with their temperature emit long-wave radiation.

The total radiation flux of an absolutely black body is determined by the Stefan–Boltzmann law:

$$B = \sigma T^4, \quad (1.3)$$

where  $\sigma$  is Stefan–Boltzmann constant equal to  $5.67 \cdot 10^{-8} \text{ W/m}^2$ . It follows from a comparison of (2.2) and (2.3) that as the temperature of the radiating body increases, the flux in the region of maximum radiation increases much faster than the total radiation flux at all wavelengths.

## 2.1. Fluxes of solar radiation in the atmosphere

Let the flux of solar radiation reach the upper boundary of the atmosphere and enter it. At the same time, it immediately splits into two streams. They are the flux of direct solar radiation and the flux of diffuse solar radiation. The flow of direct solar radiation comes in the form of a parallel beam of rays directly from the sun. The actinometer device measures the direct solar radiation ( $I$ ) coming to the receiving surface — a blackened silver plate which is perpendicular to the sun's rays. To obtain a direct solar radiation flux onto a horizontal (underlying) surface ( $I'$ ), you need to know the height of the sun above this surface ( $h_{\odot}$ ):

$$I' = I \sin h_{\odot}. \quad (1.4)$$

To measure the flux  $I$ , the actinometer tube is rotated as the sun's altitude changes. The need for rotation is due to the fact that the receiving surface must always take a perpendicular position relative to the flow of sunlight.

The flow of scattered solar radiation comes to the underlying surface from various points of the celestial sphere. It is formed from that part of direct solar radiation, which goes around various inhomogeneities in the atmosphere. These heterogeneities can be gas particles and aerosols. As a result of such a process, some of the scattered radiation flux comes to the underlying surface from points very far from the direction of the initial influx of solar radiation. The flow of scattered radiation is measured by a pyranometer device. When measuring the flux of scattered radiation, the solar disk is covered with a special screen. As a result, only scattered radiation enters the receiving surface of the pyranometer, and no direct radiation enters it. The sum of direct solar radiation arriving on a horizontal surface ( $I'$ ), and scattered solar radiation ( $i$ ) forms the flux of total solar radiation ( $Q$ ):

$$Q = I' + i . \quad (1.5)$$

The flux of direct solar radiation increases with the height of the sun and decreases with the sun descending. This is due to the fact that at high altitudes of the sun, the fluxes of its radiation travel a short path to the underlying surface. At low altitudes of the sun, due to the oblate shape of the Earth, the fluxes of solar radiation travel a long way. At large distances, there is a significant scattering of these fluxes. In such cases, the diffuse radiation flux may be greater than the direct radiation flux. At very low altitudes of the sun, the flux of total solar radiation can be represented only by the flux of scattered radiation. Note that we are talking about the ratio of these flows. At the same time, both the direct flux and the diffuse solar radiation flux are greater at the highest position of the sun. The fact is that with an increase in the flow of direct solar radiation, its share, which becomes a flow of diffuse radiation, also grows. The ratio of these streams changes because each of these streams reacts differently to changes in the height of the sun. So at low altitudes of the sun, the flux of direct solar radiation decreases significantly. As a result, in winter, as well as in the morning and in the even-

ing (in any season of the year), the scattered radiation flux often turns out to be greater than the direct radiation flux.

It is useful to recall one more factor in the increase in the flux of scattered radiation in cases of low solar altitude. When rays fall on any surface at a large angle, they are weakly reflected from this surface. In cases where the rays fall at a small angle to any surface, they are significantly reflected from this surface. The same thing happens in the atmosphere. Incident fluxes of solar radiation at low altitudes of the sun are largely reflected from the underlying surface. In winter, this is also facilitated by the snow surface, which has a high reflectivity. The reflected streams are again dissipated in the atmosphere and increase the flux of scattered solar radiation. Under conditions of snow cover and low sun altitude, there is a very noticeable increase in the flux of scattered radiation.

## **2.2. Attenuation of solar radiation in the atmosphere**

In the atmosphere, solar radiation fluxes are weakened by two processes. They are scattering and absorption by gases and aerosols of the earth's atmosphere.

Molecular scattering (by gas particles) and aerosol scattering have their own characteristics. So for molecular scattering:

a) the scattering intensity is greatest in directions parallel to the incident beam (in the direction of beam incidence and in the opposite direction); in directions perpendicular to the incident beam, the scattering intensity is minimal;

b) rays of shorter wavelengths are scattered more; in the visible part of the spectrum, blue rays are scattered more; rays of shorter wavelengths are rather weak in the stream of the visible part of the spectrum; they are purple;

long-wave rays are scattered least of all (in the visible part of the spectrum they fall in the red color); it is for this reason that a clear sky is associated with a deep blue tone.

Aerosol scattering also has two features:

a) the scattering intensity is greatest in the direction of the incident beam and directions close to it; in other directions, the scattering intensity is low;

b) the intensity of scattering of rays of different wavelengths is approximately the same; for this reason, in the visible part of the spectrum, aerosol scattering is associated with white and whitish color (when combining all colors, white is obtained).

In the real atmosphere, the scattering of solar radiation fluxes is mixed. With an increase in the amount of aerosols, the blue color is “diluted” with white. This is the color commonly called pale blue; the more aerosols in the atmosphere, the more the color of the sky loses its brightness; clouds (water aerosol media) are white or whitish in color.

When assessing the attenuation of solar radiation fluxes in the Earth’s atmosphere, calculations are carried out using the general attenuation parameters for direct and diffuse radiation. There are several such parameters. The most commonly used ones are the following: mass attenuation coefficient ( $\alpha$ ), optical mass of the atmosphere ( $m$ ), optical thickness of the atmosphere ( $\tau$ ), atmospheric transparency coefficient ( $p$ ), attenuation parameter from the formula of V. G. Kastrov ( $c$ ) and the atmospheric turbidity factor ( $T$ ). Each of these parameters is used to calculate the solar radiation flux near the underlying surface after it has passed from the upper boundary of the atmosphere.

Suppose that  $I_0$  is the radiant energy of the sun at the top of the atmosphere, and  $I$  is the solar radiation flux at the underlying surface,  $S$  is the path of the flux from the upper boundary of the atmosphere to the underlying surface. Then, when using formulas that include the above parameters of the attenuation of solar radiation in the Earth’s atmosphere, we obtain:

$$I = I_0 \exp\left(-\alpha \int_0^S \rho dS\right), \quad (1.6)$$



$$I = I_0 \exp(-\tau m), \quad (1.7)$$

$$I = I_0 p^m, \quad (1.8)$$

$$I = \frac{I_0}{1 + cm}, \quad (1.9)$$

$$I = I_0 \exp(-\tau_c Tm), \quad (1.10)$$

$$I = I_0 p_c^{Tm}. \quad (1.11)$$

Formulas (2.6–2.11) make it possible to calculate the solar radiation flux near the underlying surface ( $I$ ). The complete derivation of these formulas is given in [2]. In this textbook, we will only explain the physical meaning of each of them, as well as the parameters of the attenuation of solar radiation in the atmosphere.

Formula (2.6) is obtained by integrating the right and left parts of Buger's (or Bouguer's) law. This law reflects the dependence of the attenuation ( $dI$ ) of the solar radiation flux ( $I$ ) as it travels a unit distance ( $dS$ ) on its way to the underlying surface. This attenuation depends on the number of scattering and absorbing particles. The number of such particles characterizes the product  $\rho dS$ . As a result, the attenuation of the solar radiation flux is proportional to the product  $\rho dS$  and the value of the attenuated flow. The greater the attenuated flow, the greater is its part consisting of the loss of the flow due to its weakening in the atmosphere. In addition, the right side of (2.6) includes a coefficient called the mass attenuation coefficient ( $\alpha$ ).

The mass attenuation coefficient is a relative value of the attenuation of the solar radiation flux when it passes through an air column of unit mass. This conclusion can be drawn by multiplying the product  $\rho dS$  by a square meter. In this case, we get the density multiplied by the volume. The resulting product is the mass of the air column. If we now consider this mass equal to 1, then we will get the physical meaning of the mass attenuation coefficient. The minus sign on the right side of the equation indicates the weakening of the solar radiation flux as it passes

from the upper boundary of the atmosphere to the underlying surface. The minus sign in (2.7) and (2.10) has the same meaning.

Equation (2.7) includes the optical thickness of the atmosphere ( $\tau$ ) and the optical mass of the atmosphere ( $m$ ). The optical mass of the atmosphere is the ratio of the masses of two columns of air of a unit cross section — inclined and vertical ( $M/M_0$ ). The mass of the inclined air column ( $M$ ) is determined by the position of the sun. The column goes from the area of the upper boundary of the atmosphere, corresponding to the position of the sun, to the underlying surface. The mass of the vertical column of air ( $M_0$ ) is the mass of the column above the observation point on the underlying surface. This column also extends from the area of the upper boundary of the atmosphere to the underlying surface. If the sun is at its zenith (directly above the point of observation), then the vertical and inclined columns coincide. In this case  $M_0 = M$ , that is why  $m = 1$ . At other heights of the sun  $M > M_0$ , so the optical mass ( $m$ ) is greater than 1. The mass of an inclined column of air depends on the height of the sun, so the optical mass of the atmosphere ( $m$ ) also depends on the height of the sun:

$$m = \frac{1}{\sin h_{\odot}} . \quad (1.12)$$

The optical mass of the atmosphere varies unevenly with the height of the sun. When the sun is high, this change is small. At positions of the sun close to the horizon (during sunrise and sunset), the change in optical mass is fast and significant. When the sun is strictly on the horizon, the optical mass of the atmosphere is about 34 times greater than when the sun is at its zenith. This dependence of the optical mass on the height of the sun has a great influence on the change in temperature, air humidity and other meteorological quantities during the day. At first, at sunrise, these quantities change their values fast. In the daytime, changes occur smoothly and slightly. In the evening, at sunset, the meteoro-

logical quantities change fast again, as does the optical mass of the atmosphere.

The optical thickness of the atmosphere ( $\tau$ ) is the product of the mass attenuation coefficient of the radiation flux ( $\alpha$ ) and the mass of the vertical air column ( $M_0$ ).

We have already seen that when multiplying the product  $\rho dz$  by a unit cross section  $1 \text{ m}^2$ , we get the density multiplied by the elementary volume. This product is the mass of an elementary column of air. If we integrate this mass within the range from the level of the underlying surface to the level of the upper boundary of the atmosphere, then we obtain the mass of the entire vertical air column above the observation point:

$$M_0 = \int_0^z \rho dz . \quad (1.13)$$

If the resulting value ( $M_0$ ) is multiplied by  $\alpha$ , then we get the weakening of the solar radiation flux not in a unit mass of air, but in the entire vertical column of air:

$$\tau = \alpha M_0 = \alpha \int_0^z \rho dz . \quad (1.14)$$

This value is used in (2.7).

Formula (2.6) can be written as:

$$I = I_0 \exp(-\alpha M) = I_0 \exp(-\alpha M_0 m) , \quad (1.15)$$

because  $m = M/M_0$  ,  $M = M_0 m$  .

The transparency coefficient of the atmosphere is:

$$p = \exp(-\tau) . \quad (1.16)$$

Taking (2.16) into account, formula (2.8) is written. From analysis of (2.8), the physical meaning of the transparency coefficient ( $p$ ) is clear. If the sun is at its zenith ( $m = 1$ ), then  $p = I/I_0$  . Thus, the transparency coefficient is the fraction of the solar radiation flux that reaches

the underlying surface at the position of the sun at its zenith. The flux of solar radiation at the upper boundary of the atmosphere is of course greater than the flux that has come to the underlying surface. As a result,  $p < 1$ .

The turbidity factor of the atmosphere ( $T$ ) is determined by the ratio of the optical thickness of the real atmosphere and the optical thickness of the dry and clean (ideal) atmosphere:

$$T = \frac{\tau}{\tau_c}. \quad (1.17)$$

It is clear that  $T > 1$ . Note that the turbidity factor ( $T$ ) depends less on the height of the sun and on the optical mass of the atmosphere ( $m$ ) than the transparency coefficient ( $p$ ).

The dependence of the transparency coefficient on the optical mass of the atmosphere for the flux of solar radiation from rays of different wavelengths is very complex. Usually in calculations we use the average value of the transparency coefficients for rays of all wavelengths in the radiant energy flux at certain values of the optical mass. Most often, calculations are carried out at  $m = 1$  and  $m = 2$ .

There is a well-known effect of Forbes. It lies in the fact that during the passage of radiant energy through the atmosphere, a shift occurs in the solar flux towards longer waves. This is because shorter wavelengths are more attenuated in the atmosphere. As a result, a flux of solar radiation with a lower content of short-wave rays enters each subsequent volume along its way in the atmosphere. Of course, the flux of radiant energy passing through the atmosphere is attenuated at all wavelengths, but to a different extent. In each subsequent volume of the air on the way to the underlying surface, the proportion of long-wave radiation in the flow increases.

The described effect is clearly seen in the example of rays in the visible part of the spectrum. So during sunrise and sunset, the sun can turn red. This is due to the fact that our atmosphere is somewhat flattened,

so when the sun is at its zenith, its rays travel the shortest path, and when the sun is at the horizon, they travel the longest path. On a long path, as the flow of radiant energy in the atmosphere moves from its upper boundary, rays of short wavelengths are more and more scattered. Such scattering occurs already at the initial stages of this path. As a result, only rays of longer wavelengths remain in the beam of rays. In the visible part of the spectrum these are red rays.

The solar radiation flux near the underlying surface can be calculated by transforming formulas (2.7) and (2.8) and taking into account (2.17):

$$I = I_0 \exp(-\tau_c T m), \quad (1.18)$$

$$I = I_0 p_c^{Tm}. \quad (1.19)$$

In (2.18)  $\tau_c$  is the optical depth of dry and clean (ideal) atmosphere. In (2.19)  $p_c$  is the coefficient of transparency of dry and clean atmosphere. The derivation of these formulas is given in [2].

In some cases, the flux of solar radiation near the underlying surface is determined by the formula of V. G. Kastrov (2.9). This formula uses the attenuation parameter ( $c$ ), which, in contrast to the transparency coefficient ( $p$ ), increases with cloudiness of the atmosphere. This is understandable, since in (2.9) this parameter is in the denominator of the right side of this formula.

### ***Test questions:***

1. What streams of radiant energy are found in the atmosphere? Describe each of them both quantitatively and qualitatively (spectral side).
2. What dependencies reflect the law of Wien, and Stefan–Boltzmann formula?
3. Under the influence of what processes solar radiation is weakened in the Earth's atmosphere?
4. What types of scattering are observed in the atmosphere? What are the characteristics of each of these types? What examples characterize each type of scattering?
5. What parameters can characterize the attenuation of radiant energy in the Earth's atmosphere?

6. What is the mass attenuation coefficient, optical mass and optical thickness of the atmosphere?
7. What is the transparency coefficient and haze factor of the atmosphere?
8. What is the Forbes effect?
9. What dependence does the Kastrov's formula reflect?

### **2.3. The radiation balance of the underlying surface, atmosphere and the planet as a whole**

Any balance is the income and expenditure of something relative to any surface or volume. In physical meteorology, the radiation balance of the underlying surface, the atmosphere, and the planet as a whole is considered. To obtain any of these radiation balances, it is necessary to determine their incoming and outgoing parts.

#### **2.3.1. Radiation balance of the underlying surface ( $R$ )**

The incoming items are direct ( $I'$ ) and scattered ( $i$ ) solar radiation, as well as the radiation flux from the atmosphere, absorbed by the underlying surface. The atmosphere radiates in all directions, including the underlying surface. The radiation of the atmosphere towards the underlying surface is called the counter-radiation of the atmosphere ( $B_A$ ). This name is due to the fact that this radiation is directed towards the radiation of the underlying surface (against the radiation of the underlying surface). Part of this flux is absorbed by the underlying surface ( $\delta B_A$ ), the other part of the flux is reflected and is not an incoming item in this radiation balance.

The expenditure items are the intrinsic radiation of the underlying surface ( $B_0$ ) and those parts of direct and diffuse solar radiation that are reflected from the underlying surface.

The incoming items are recorded with a “plus” sign, the expenditure items — with a “minus” sign. As a result, the equation of the radiation balance of the underlying surface has the following forms:

$$R = (I' - rI') + (i - ri) + \delta B_A - B_0, \quad (1.20)$$

$$R = I'(1 - r) + i(1 - r) - (B_0 - \delta B_A), \quad (1.21)$$

$$R = (I' + i)(1 - r) - B_0^*, \quad (1.22)$$

where  $r$  is the relative reflectance or reflectivity of the underlying surface;  $rI'$  and  $ri$  are parts of the fluxes of direct and diffuse solar radiation reflected from the underlying surface;  $B_0^*$  is the effective radiation of the underlying surface; it is equal to the difference between the natural radiation of the atmosphere ( $B_0$ ) and that part of the counter radiation of the atmosphere which the underlying atmosphere absorbed ( $\delta B_A$ ).

If  $R > 0$ , then more flow comes to the underlying surface than leaves it, so the underlying surface heats up. If a larger flux of radiation leaves the underlying surface than arrives at it ( $R < 0$ ), then this underlying surface cools down. An option is also possible when the underlying surface maintains its temperature ( $R = 0$ ).

### 2.3.2. Radiation balance of the atmosphere

The atmosphere, as mentioned above, absorbs mainly long-wave radiation from the underlying surface ( $U_{\Pi}$ ). The atmosphere absorbs much less radiant energy from the sun ( $Q_{\Pi}$ ). Expenditure items are made up by the radiation which radiates out to space ( $B_{\infty}$ ) and that part of the counter radiation of the atmosphere which was absorbed by the underlying surface ( $\delta B_A$ ).

The general form of the atmospheric radiation balance equation is:

$$R_A = U_{\Pi} + Q_{\Pi} - B_{\infty} - \delta B_A. \quad (1.23)$$

The annual radiation balance of the atmosphere is  $R_A < 0$ . This is due to the fact that  $Q_{\Pi}$  is of little importance, and part of  $U_{\Pi}$  radiates into space ( $B_{\infty}$ ). In addition, when the radiation of the underlying surface is absorbed, the atmosphere also radiates towards it. The atmos-

phere loses this radiation when part of this radiation is absorbed by the underlying surface.

### **2.3.3. Radiation balance of the underlying surface and atmosphere system**

In this consideration of the radiation balance, we will not take into account the exchange of radiant energy fluxes between the underlying surface and the atmosphere. Let us take into account only those flows that enter through the upper boundary of the atmosphere towards our planet and leave it into the outer space. The simplest is the equation of the radiation balance of our planet in the form:

$$R_s = I_0' (1 - r_s) - U_\infty, \quad (1.24)$$

where  $R_s$  is the radiation balance of the planet,  $I_0'$  is the flux of radiant energy from the sun to the upper boundary of the atmosphere,  $r_s$  is the albedo of the planet (reflectivity of the upper boundary of the atmosphere, which is the outer boundary of the Earth);  $U_\infty$  is the outgoing radiation of the underlying surface and atmosphere (of the planet as a whole).

You can write the equation of the planet's radiation balance in other forms. In such cases one should sum up all the radiation fluxes coming separately to the underlying surface and to the atmosphere, and subtract the radiation fluxes that leave the underlying surface and the atmosphere. The equation of the planet's radiation balance in other forms is given in [2].

#### **Test questions:**

1. What is the radiation balance?
2. What components make up the radiation balance of the underlying surface?
3. What processes and phenomena are influenced by the radiation balance of the underlying surface?
4. What are the components of the radiation balance of the atmosphere?
5. What are the characteristics of this balance?
6. What components make up the radiation balance of the Earth as a planet (earth-atmosphere system)?



### 3. THERMODYNAMICS OF THE ATMOSPHERE

The basis of thermodynamics is its two principles — two basic laws. In atmospheric thermodynamics we will mostly use the first law of thermodynamics.

#### 3.1. Equation of the first law of thermodynamics for dry and humid unsaturated air

We use the following formulation of the first law of thermodynamics: the amount of heat ( $dq$ ) transferred to the system is converted to its internal energy ( $du_i$ ) and work against external forces ( $dw_i$ ):

$$dq = du_i + dw_i . \quad (3.1)$$

The index  $i$  shows that we are talking about the internal energy of the system and the work done by it. All system parameters are denoted by the index  $i$ . All parameters of the environment surrounding this system are denoted by the index  $e$ .

Equation (3.1) expresses the first law of thermodynamics in general terms. In a course of physics, this equation is written for an ideal gas. In physical meteorology, the equations of the first law of thermodynamics are written separately for dry air, for humid air not saturated with water vapor, and for humid air saturated with water vapor.

A system in physical meteorology is a unit mass of air, which can make independent movements relative to the air surrounding it. The parameters of this unit air mass are denoted by the index  $i$ :  $P_i, \rho_i, T_i, m_i$ , and the parameters of the air surrounding this particle are denoted by the index  $e$ :  $P_e, \rho_e, T_e, m_e$ .

In physical meteorology, we accept a quasi-static condition. It presupposes that the pressure inside the particle is equal to the pressure in its environment:  $P_i = P_e = P$ . The quasi-static condition is satisfied most accurately in cases where the speed of particles is much less than the speed of sound. This condition is not met only in the high layers of the atmosphere and at very high speeds in the lower layers.

If an air particle is given a certain amount of heat ( $dq$ ), the internal energy of the given air particle ( $u_i$ ) will increase by the value  $du_i$ . In addition, the particle will expand, doing work against external pressure forces. This work will be  $dW_i$ . In the atmosphere, this work is done against external pressure forces.

Internal energy is the energy of thermal chaotic motion. In general, internal energy is a state function depending on temperature and volume. For an ideal gas, the internal energy depends only on temperature, but does not depend on volume. The fact is that the volume characterizes the degree of interaction of particles. However, in an ideal gas, particles do not interact with each other most of the time. For this reason, if in general:

$$u = f(T) + F(v), \text{ t.e. } du = \left( \frac{\partial u}{\partial T} \right)_v dT + \left( \frac{\partial u}{\partial v} \right)_T dv, \quad (3.2)$$

then for an ideal gas:

$$du = \left( \frac{\partial u}{\partial T} \right)_v dT, \quad (3.3)$$

where  $\frac{\partial u}{\partial T} = c_v$ .

Dry and humid unsaturated air is similar in its properties to the properties of an ideal gas. A gas is the closer in its properties to an ideal gas, the lower its critical temperature relative to the actual air temperature is, and the lower the partial pressure relative to the saturation pressure of the given gas is.

The critical temperature is the temperature above which, at any pressure, a gas cannot be converted to a liquid state. It is known that the gaseous and liquid states of matter differ in the degree of particle interaction. The smaller the degree of interaction between the particles of a gas is, the closer this gas is to the properties of an ideal gas. A decrease in the partial pressure of a gas also indicates a weakening of the interaction of its particles. Almost all gas components of a real gas have critical temperatures much lower than real temperatures. In those high layers where the temperatures are very low (in the upper mesosphere) the air density is also low. Under these conditions the interactions between the particles of a real gas are very weak. In the lower layers, the only exceptions are two gases — carbon dioxide and water vapor. However, the share of carbon dioxide in the total gas mixture is small. Carbon dioxide's partial pressure is far from saturation pressure. Water vapor, even when air is saturated with it, can also be considered a gas to a certain extent, not very far in its properties from the properties of an ideal gas.

Thus, for dry and moist unsaturated air, the equation of the first law of thermodynamics has the same form as for an ideal gas:

$$dq = c_v dT_i + P dV_i, \quad (3.4)$$

where

$$c_v dT_i = du_i. \quad (3.5)$$

The work done by a unit mass of air when absorbing heat is the work of expansion against external pressure forces. It depends on this pressure and volume change:

$$dw_i = P dV_i. \quad (3.6)$$

Equation (3.4) is the first form of the equation of the first law of thermodynamics. Its right side includes two expenditure items, which absorb the heat received by the system. However, this equation contains an unmeasured and unknown quantity  $dV_i$ . For this reason, (3.4) should be transformed in such a way as to express  $dV_i$  through known

or measurable quantities. To this end, we use the equation of state for dry air and take the derivative of the left and right parts of this equation:

$$Pv_i = R_c T_i, \quad (3.7)$$

$$Pdv_i + v_i dP = R_c dT_i, \quad (3.8)$$

$$Pdv_i = R_c dT_i - v_i dP. \quad (3.9)$$

Let us substitute the value  $Pdv_i$  into equation (3.4):

$$dq = c_v dT_i + R_c dT_i - v_i dP. \quad (3.10)$$

In thermodynamics, there are useful equations of the German scientist J. R. von Mayer. According to Mayer's research, the quantitative relationship between the values of specific heat capacity at constant pressure ( $c_p$ ) and specific heat capacity at constant volume is set using the gas constant:

$$c_v + R_c = c_p. \quad (3.11)$$

Then in (3.10):

$$c_v dT_i + R_c dT_i = c_p dT_i, \quad (3.12)$$

therefore (3.10) takes the form:

$$dq = c_p dT_i - v_i dP. \quad (3.13)$$

However, in (3.13) there is an unknown and unmeasurable quantity  $v_i$ . It can be obtained from the equation of state (3.7):

$$v_i = \frac{R_c T_i}{P}.$$

Then (3.13) takes the form:

$$dq = c_p dT_i - R_c T_i \frac{dP}{P}. \quad (3.14)$$

Equation (3.14) is the second form of the equation of the first law of thermodynamics. In this equation, the terms on its right side are not divided into two expenditure items when a certain amount of heat is

transferred to the system. However, this equation is often used for various calculations in physical meteorology. In (3.14) all quantities on the right side of the equation are known or can be measured.

### 3.2. Adiabatic processes in dry and wet unsaturated air

Adiabatic processes proceed without heat exchange between the system and its environment. For adiabatic processes  $dq = 0$ . In the atmosphere, many processes can be considered close to adiabatic processes. Vertical air movements can be considered adiabatic. In these cases, the change in heat content  $c_p dT_i$  due to the work of expansion against external forces of pressure far exceeds the heat influx from the surrounding air. For this reason, the influx of heat from the outside can be neglected.

Vertical movements in the atmosphere are ascending movements of air in cyclone systems, in particular in the zones of atmospheric fronts. In these zones, warm air masses move along the wedge of cold masses. The same movements include descending air movements in anticyclones. In addition, ascending movements of air masses heated more than the air surrounding these masses also belong to vertical movements. Such conditions are created above an inhomogeneous underlying surface, which itself heats up unevenly. If in equations (3.1), (3.4) and (3.14)  $dq$  is considered equal to zero, then these equations will take the form:

$$dw_i = -du_i, \quad (3.15)$$

$$Pdv_i = -c_v dT_i, \quad (3.16)$$

$$c_p dT_i = R_c T_i \frac{dP}{P}. \quad (3.17)$$

The minus sign on the right side of (3.15) and (3.16) indicates that in an adiabatic process, work against external pressure forces is per-

formed solely at the expense of the internal energy of the system. Such a system in physical meteorology is a unit mass of air. If there is an extension ( $d\nu_i > 0$ ), then the internal energy of the particle decreases ( $dT_i < 0$ ). This situation occurs when the air particle moves upward. When an air particle is compressed ( $d\nu_i < 0$ ), its internal energy grows ( $dT_i > 0$ ). This happens when the particle moves down. Such a conclusion follows from the analysis of (3.16). The same conclusion follows from the analysis of (3.17). When an air particle rises, its pressure decreases ( $dP < 0$ ), therefore  $dT_i < 0$ .

Thus, from the analysis of both equations (3.16) and (3.17) it follows that when a particle rises adiabatically, its temperature always decreases, and when it descends adiabatically, its temperature rises.

We can now obtain the equation of the French scientist S. D. Poisson with which, in a dry adiabatic process, it is possible to calculate the temperature or pressure in one state if they are known for another state. To this end, we will integrate (3.17) dividing the variables first:

$$\int_{T_{i0}}^{T_i} c_p \frac{dT_i}{T_i} = \int_{P_0}^P R_c \frac{dP}{P}, \quad (3.18)$$

$$\frac{T_i}{T_{i0}} = \left( \frac{P}{P_0} \right)^{\frac{R_c}{c_p}}. \quad (3.19)$$

Taking into account the Mayer equation, we get:

$$\frac{T_i}{T_{i0}} = \left( \frac{P}{P_0} \right)^{\frac{c_p - c_v}{c_p}}. \quad (3.19')$$

Let us divide each parameter of the exponent on the right side of (3.19') by  $c_v$  :

$$\frac{T_i}{T_{i0}} = \left(\frac{P}{P_0}\right)^{\frac{\kappa-1}{\kappa}}, \quad (3.19'')$$

where  $\kappa = c_p/c_v = 1,4$ ; and  $(\kappa-1)/\kappa = 0,286$ . Then the equation (3.19'') can be given the form:

$$\frac{T_i}{T_{i0}} = \left(\frac{P}{P_0}\right)^{0,286}. \quad (3.19''')$$

Equations (3.19)–(3.19''') are called the equations of the adiabatic process in integral form, or the dry adiabatic equations, or the Poisson equations. Using equation (3.14) and assuming that  $dq = 0$  in it, one can obtain the vertical temperature gradient in the dry adiabatic process. Such a gradient characterizes the adiabatic process in dry and moist air unsaturated with water vapor. It is called the dry adiabatic gradient. The term “vertical” is not applied to such a gradient, since it is known that the adiabatic processes occur in vertical motions.

$$c_p dT_i - R_c T_i \frac{dP}{P} = 0. \quad (3.20)$$

Since, according to the quasi-static condition, the pressure inside the particle  $P_i$  is equal to the pressure in the surrounding air  $P_e$ , then  $P = \rho_e R_c T_e$  and, according to the basic equation of statics,  $dP = -g \rho_e dz$ , that's why:

$$\frac{dP}{P} = -g \frac{dz}{R_c T_e}. \quad (3.21)$$

Substitute the value  $\frac{dP}{P}$  from (3.21) to (3.20):

$$c_p dT_i + g \frac{T_i}{T_e} dz = 0. \quad (3.22)$$

Divide the terms on the left side of (3.22) by  $c_p dz$  :

$$\left(\frac{dT_i}{dz}\right)_\alpha = -\frac{g}{c_p} \cdot \frac{T_i}{T_e}, \quad (3.23)$$

where  $\left(-\frac{dT_i}{dz}\right)_\alpha$  is a dry adiabatic gradient. It determines the change in temperature of an air particle with a change in altitude:

$$\gamma_a = \frac{g}{c_p} \cdot \frac{T_i}{T_e}. \quad (3.24)$$

In (3.24) the value  $T_i/T_e$  is close to 1, since in the real atmosphere the particle temperature difference between  $T_i$  and the temperature of its environment  $T_e$  does not exceed 5–10°C. With this approach, the dry adiabatic gradient  $\gamma_a$  becomes a constant:

$$\gamma_a = \frac{g}{c_p}. \quad (3.25)$$

Let us get the numerical value of the dry adiabatic gradient:

$$\begin{aligned} \gamma_a &= \frac{9,8 \text{ m/sec}^2}{1005 \text{ J/kg} \cdot \text{K}} = 0,0098 \frac{\text{m}}{\text{sec}^2} \frac{\text{sec}^2}{\text{kg} \cdot \text{m}^2} \text{kg} \cdot \text{K} = \\ &= 0,0098 \text{ K/m} = 0,98 \text{ K/100 m} = \\ &= 0,98^\circ\text{C/100 m} \approx 1^\circ\text{C/100 m} \approx 1 \text{ K/100 m} \end{aligned}$$

When sizing, keep in mind that

$$\text{J} = \text{N} \cdot \text{m} = \text{m/sec}^2 \cdot \text{kg} \cdot \text{m} = \text{kg} \cdot \text{m}^2/\text{sec}^2.$$

Thus, it is believed that the temperature of an adiabatically rising dry air particle drops by about one degree Celsius for every 100 m of altitude. Once again, we note that all vertical gradients are calculated in units of the given meteorological quantity per 100 m of height.

We now obtain the approximate equation of the dry adiabat. By definition of the dry adiabatic gradient:



$$\gamma_a = -\frac{dT_i}{dz}. \quad (3.26)$$

Let us integrate (3.26) within the limits of  $z_0$ , where the particle temperature is  $T_{i0}$  to  $z$ , where the temperature of the particle is  $T_i$ :

$$\int_{T_{i0}}^{T_i} dT_i = -\gamma_a \int_{z_0}^z dz; \quad T_i = T_{i0} - \gamma_a (z - z_0). \quad (3.27)$$

Equation (3.27) is the approximate dry adiabatic equation. The equation is approximate because  $\gamma_a$  in (3.27) is considered a constant value. In this case, the change in temperature with height is represented by a straight line on the graph. On it the height is plotted on the ordinate axis, and the temperature is plotted on the abscissa axis. The line of change in temperature with height is called the dry adiabat or the state curve of a dry air particle.

The adiabatic process is a special case of the polytropic process. This process is described by the general equation in the form:

$$dq = cdT, \quad (3.28)$$

where  $c$  is the heat capacity. In each version of the polytropic process, the quantities included in (3.28) have their own values. In the case of an adiabatic process,  $c = 0$ , since  $dq = 0$ , but  $dT$  does not take zero values.

### 3.3. Particle method for estimating the stability of the atmosphere

We will analyze the state of the atmosphere. For this purpose, at the level near which the state of the atmosphere is analyzed, let us single out an air particle (also called parcel). Let us imagine its movement along the vertical, as it happens in the atmospheric situations listed above. Let this particle move up and down from the initial level, i.e. without heat input. Such a change in the position of the particle occurs according to adiabatic laws. First, we will consider a moving particle of

dry air. The temperature of this particle at the initial level  $z_0$  is considered to be equal to  $T_{i0}$ . The particle method compares the values of vertical temperature gradients ( $\gamma$ ) with an adiabatic gradient. In the case of a dry air particle, the adiabatic gradient is the dry adiabatic gradient  $\gamma_a$ .

These gradients differ both in essence and notation. The vertical temperature gradient is written as the partial derivative of temperature with respect to height:

$$\gamma = -\frac{\partial T_e}{\partial z}. \quad (3.29)$$

It characterizes the change in temperature in the vertical direction in the air surrounding the particle. However, the temperature changes in all directions in this air, in particular in the vertical direction. Equation (3.26) contains the total derivative of the change in particle temperature with height. This is due to the fact that the adiabatic change in the atmosphere is associated only with the vertical motion of the particle.

The particle method considers three fundamentally different cases of temperature distribution in a dry atmosphere:

$$\gamma > \gamma_a, \quad \gamma = \gamma_a, \quad \gamma < \gamma_a.$$

1)  $\gamma > \gamma_a$ , that is, the temperature of the air falls with height rapidly, faster than  $1^\circ\text{C}/100\text{ m}$ . Graphically, this situation is depicted as follows. The faster the temperature drops with height, the smaller is the angle of the temperature change curve to the abscissa axis (temperature scale). This curve is called the stratification curve. As stated above, the curve of temperature change with height for a particle is called the state curve of that particle. In the case of a dry air particle, this curve is called the dry adiabat. Under the above conditions, the stratification curve is inclined to the abscissa axis more than the dry adiabat.

On the baseline ( $z_0$ ) the temperature of the particle is equal to the temperature of the air surrounding it; since the particle is separated from this air,  $T_{i0} = T_{e0}$ . Let us move the particle dry-adiabatically to a

higher level  $z_2$  relative to the baseline  $z_0$ . At the level  $z_2$  the particle temperature is  $T_{i2}$ , and in the air around it, the temperature is  $T_{e2}$ . Since  $T_{i2} > T_{e2}$ ,  $\rho_{i2} < \rho_{e2}$  according to the equation of state and the quasi-static condition ( $P_i = P_e$ ).

For simplicity, we will assume that the air particle has a unit volume. Any air particle at each level in the atmosphere is affected by two forces that determine the direction and speed of movement of this particle vertically:

a) the force of gravity directed downward and equal to the value  $g\rho_i$ ; in the general case the force of gravity is equal to  $mg$ , but in the case of a unit volume, the mass is numerically equal to the density ( $m = \rho$ );

b) the force of Archimedes, directed upwards and equal to the value  $g\rho_e$ .

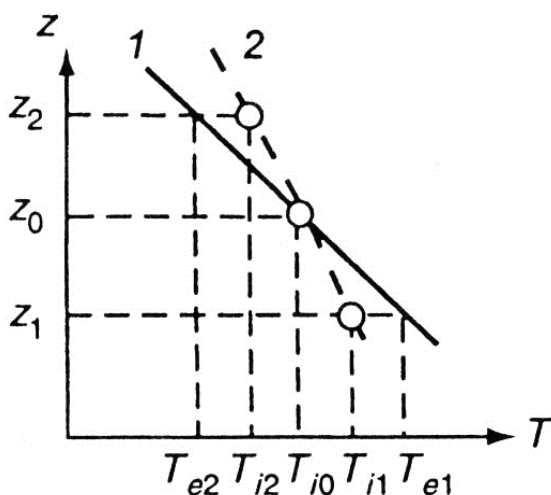
The resultant of these two forces is called the buoyancy force and is  $g(\rho_e - \rho_i)$ . This force is directed upward at  $\rho_e > \rho_i$  and downward at  $\rho_e < \rho_i$ . Under the influence of the buoyancy force, the particle receives an acceleration:

$$\frac{d^2z}{dt^2} = g \frac{\rho_e - \rho_i}{\rho_i} = g \frac{T_i - T_e}{T_e}. \quad (3.30)$$

Let the particle move from the initial level up to the level  $z_2$ :  $\gamma > \gamma_a$ ,  $T_{i2} > T_{e2}$ ,  $\rho_{i2} < \rho_{e2}$ . This determines the positive acceleration of the particle at a given level. As a result, the particle accelerates upward.

Let now a particle of dry air move from the initial level  $z_0$  to the lower level  $z_1$ . At this level  $T_{i1} < T_{e1}$  and  $\rho_{i1} > \rho_{e1}$ , so the particle will move downward with increasing speed.

Thus, provided  $\gamma > \gamma_a$ , a particle displaced from the initial level in any direction to any distance acquires the speed and acceleration in the direction in which it received a momentum at the initial moment. In this case, the absolute value of the velocity becomes the greater, the farther the particle is from the initial level. On Fig. 3.1 it can be seen that in this case the temperature difference between the particle and the air surrounding it increases.



**Fig. 3.1.** Dry unstable stratification ( $\gamma > \gamma_a$ ): 1 – stratification curve, 2 – dry adiabat

Analysis of the situation leads to the conclusion that under the condition that  $\gamma > \gamma_a$ , the equilibrium of an air particle is already unstable in its initial position, therefore, the stratification of the atmosphere at  $\gamma > \gamma_a$  is called dry unstable stratification.

2)  $\gamma = \gamma_a$ , i. e. when rising up, the temperature in the atmosphere loses one degree for every 100 m of altitude. It is clear that the tempera-

ture values of the air particle and in its environment at the initial level are equal. At the level  $z_1$  the ratios of the temperature values and its environment are as follows:  $T_{i1} = T_{e1}$ ,  $\rho_{i1} = \rho_{e1}$ , and at the level  $z_2$  :  
 $T_{i2} = T_{e2}$ ,  $\rho_{i2} = \rho_{e2}$ .

Thus, provided  $\gamma = \gamma_a$ , whatever level we displace the allocated volume of air to, the acceleration of its movement, according to (3.30), is always equal to zero. The state of the atmosphere provided  $\gamma = \gamma_a$  is called neutral (indifferent) dry stratification or equilibrium stratification.

3)  $\gamma < \gamma_a$ , i. e. the temperature in the atmosphere decreases with altitude more slowly than  $1^\circ\text{C}/100\text{ m}$ . In this case, at the level  $z_2$  the temperature of the particle is below the surrounding air temperature:  $T_{i2} < T_{e2}$ , that is why  $\rho_{i2} > \rho_{e2}$ . In such a situation, the particle acquires, according to (3.30), a negative acceleration. When moving down, the air particle acquires a positive acceleration. This means that the particle will return to its original position wherever it is moved vertically. So, after the termination of the action of external forces, the particle returns to its original level. Strictly speaking, it will oscillate around the initial level from where it was previously displaced.

It is clear that the state of the atmosphere under the condition when  $\gamma < \gamma_a$  is characterized by stability and dry stratification. In the real atmosphere, dry stable stratification is observed at isothermal ( $\gamma = 0$ ) and inverted ( $\gamma < 0$ ) temperature distribution with altitude. With a slight drop in temperature with height, the stratification of the atmosphere is close to or is dry and stable. With a stable stratification, especially during inversion, the vertical movement of air particles is difficult, and in cases of strong inversion it is completely excluded.

In cases of unstable stratification, favorable conditions are formed for the emergence and development of vertical motions of air particles.

If the air is moist but not saturated with water vapor, then the thermodynamic processes in it, in particular adiabatic processes, do not noticeably differ from adiabatic processes in dry air. In this case, with a small error, the adiabatic gradient is the dry adiabatic gradient.

If the air is humid and saturated with water vapor, then using the particle method it is considered that: if  $\gamma > \gamma_a'$ , the stratification is moist unstable; if  $\gamma = \gamma_a'$ , the stratification is moist neutral (indifferent); if  $\gamma < \gamma_a'$ , the stratification is moist stable. In the given examples  $\gamma_a'$  is a wet adiabatic gradient, the meaning of which will be discussed below.

The described method of analyzing the state of the atmosphere is called the particle method, since when analyzing the thermodynamic stability of air layers, the indicator of their state is the behavior of an air particle. In our example, the particle is a unit volume of air. It is assumed that its vertical motions occur according to the adiabatic model. In the real atmosphere, this assumption is fulfilled only to a certain extent. At the same time, the qualitative conclusions drawn under this assumption are quite consistent with what is observed in reality.

This method is convenient, quite reliable, and is often used to analyze the state of the atmosphere.

### 3.4. Potential temperature

In physical meteorology, in particular in the thermodynamics of the atmosphere, many temperature characteristics are used. A very special place in this series is occupied by the potential temperature. In this section, we will consider options for calculating this meteorological quantity and its properties. In addition, in this section and subsequent sections, we will consider the application area of potential temperature in theory and practice.

The potential temperature is the temperature that an air particle will take if it is lowered or lifted dry-adiabatically from the initial level (at which the potential temperature is being calculated) to the level where the atmospheric pressure is equal to the value of 1000 hPa.

Why does this definition refer to a level with a specified pressure value? The fact is that the average atmospheric pressure at the underlying surface is 1013.25 hPa. If we round this value, which is more convenient both for calculations and for graphical solutions of various problems, then we will get the value of 1000 hPa.

### 3.4.1. Potential temperature calculation

Potential temperature is indicated by the symbol  $\Theta$ . Let us try to calculate this meteorological value. Let the air particle have two states: initial ( $T_i, P$ ) and final ( $\Theta, 1000$  hPa). The process in which the potential temperature is determined is dry adiabatic. The Poisson equation is applicable to such a process, in which the parameters of the final state are indicated above. Then the equation (3.19'') will look like:

$$\frac{\Theta}{T_i} = \left(\frac{1000}{P}\right)^{\frac{\kappa-1}{\kappa}}, \quad (3.31)$$

$$\Theta = T_i \left(\frac{1000}{P}\right)^{\frac{\kappa-1}{\kappa}}. \quad (3.32)$$

Equation (3.32) can be used to calculate the potential temperature quite accurately. At the same time, some computer technology is required to calculate the second factor on the right side of (3.32). Of course, if (3.32) is included as a fragment in a large calculation using a computer, then the problem of calculating the potential temperature can be considered solved. However, in a number of cases it is required to quickly estimate the value of the potential temperature, albeit with some small error. The solution to this problem is the use of approximate expressions for calculating the potential temperature. The form of these expressions depends on the relative position of three levels — the level

of the underlying surface, the level with an atmospheric pressure of 1000 hPa, and the level for which the potential temperature is calculated and at which an air particle is singled out.

Here are some approximate expressions for calculating the potential temperature.

1. Let the level of the underlying surface and the level at which the value of the atmospheric pressure is equal to the value of 1000 hPa coincide. Let the potential temperature be calculated at the same level. In this case, the air particle isolated at the underlying surface does not need to be shifted anywhere to obtain the potential temperature:  $\Theta = T_i$ , therefore, the potential temperature is equal to the air temperature observed directly at the underlying surface.

2. Let the level of the underlying surface still coincide with the level where the atmospheric pressure is 1000 hPa. The level for which it is required to calculate the potential temperature is above the underlying surface. In this case, the air particle should be lowered dry-adiabatically to the level with an atmospheric pressure of 1000 hPa. It is clear that the air particle temperature increases by 1 °C when lowered for every 100 m of height. As a result:

$$\Theta = T_i + z/100 . \quad (3.33)$$

In other words, in this case, as many degrees are added to the measured air temperature at the level for which the potential temperature is calculated, as many 100-meter segments are between this level and the underlying surface.

3. Let the level of the underlying surface not coincide with the level where the atmospheric pressure is 1000 hPa. So, for example, at the underlying surface, the pressure exceeds 1000 hPa. Let it be required to calculate the potential temperature at the level located between the underlying surface and the level with a pressure of 1000 hPa. In this case, the calculation is performed in two steps. First, the particle must reach the underlying surface in a dry adiabatic manner. That is, to the temper-



ature of the level for which the calculation is performed you should add a value equal to  $z/100$ . After that, it should rise dry-adiabatically from the underlying surface to the level with the pressure of  $P = 1000$  hPa. The second part of the calculation requires special explanation. The thing is that we do not always know at what height relative to the underlying surface the level with a pressure of 1000 hPa is located. At the same time, we know that this level is located near the underlying surface, since the average surface atmospheric pressure is 1013.25 hPa. For this reason, the difference in atmospheric pressure at the underlying surface and at the level where the pressure is 1000 hPa can be converted into a height difference between these levels. This is done using the pressure step ( $h$ ). Near the underlying surface, this meteorological value is about 8 hPa/m. As a result, the specified height difference will be equal to the value  $(1000 - P_0) \cdot h$ . The resulting height excess above the underlying surface should then be divided by 100 m. This action is due to the fact that during the dry adiabatic movement of a particle, its temperature changes for every 100 m by  $1^\circ\text{C}$ .

Ultimately, the approximate expression for calculating the potential temperature takes the form:

$$\Theta = T_i + \frac{z}{100} + \frac{(1000 - P_0)8}{100},$$

$$\Theta = T_i + \frac{z}{100} + \frac{(1000 - P_0)}{12,5}. \quad (3.34)$$

In this case, the difference  $(1000 - P_0)$  will be negative, since the pressure at the underlying surface is higher than the value of 1000 hPa. Interestingly, (3.34) can be used to calculate the potential temperature in any case. Let us demonstrate it.

4. Now let the value of the atmospheric pressure at the underlying surface be below 1000 hPa. Then the level with a value of 1000 hPa is theoretically below the underlying surface. Let it be required to calcu-

late the potential temperature at the level located above the underlying surface. Let us use expression (3.34). To the temperature at the calculation level we add the second term equal to  $z/100 \cdot 1^\circ\text{C}$ . Thus, we will obtain the value of the temperature of the particle during its dry adiabatic lowering to the level of the underlying surface. At the same time, in this case, the particle should albeit only theoretically fall even lower — to the theoretical level with a pressure of 1000 hPa. To this end, we add to the temperature value obtained by summing the first two terms in (3.34) also the third term. In the case under consideration, the difference  $(1000 - P_0)$  will be positive. Thus, the first two terms on the right side of (3.34) make the main contribution to the value of the calculated potential temperature. The third term corrects the obtained value since the difference  $(1000 - P_0)$  is small.

With height, the potential temperature decreases under the condition of dry unstable stratification, does not change under the condition of dry indifferent stratification, and increases under dry stable stratification. The proof of this proposition will be given below. The nature of the change in potential temperature with height gives the right to consider it as one of the methods for assessing the stratification of the atmosphere in the layer from the underlying surface to the altitude level, where moist unsaturated air becomes saturated with water vapor.

### **3.4.2. Potential temperature properties**

Why did it become necessary to introduce in addition to other temperature characteristics also the potential temperature? The point is that potential temperature has two important properties.

The first of them is related to the constancy of the potential temperature in the dry adiabatic process for any displacement of the particle. Let us show that this is so. To this end, we will take the logarithm and differentiate (3.32):

$$\frac{d\Theta}{\Theta} = \frac{dT_i}{T_i} - \frac{R_c}{c_p} \cdot \frac{dP}{P}. \quad (3.35)$$

If we multiply the right and left sides of this equation by the product  $c_p T_i$ , then we get:

$$c_p T_i \frac{d\Theta}{\Theta} = c_p dT_i - R_c T_i \frac{dP}{P}. \quad (3.36)$$

According to (3.20), in the dry adiabatic process, the right side of (3.36) is equal to zero, therefore, the left side of this equation is also equal to zero. Since the first two factors of the left side of (3.36) are not equal to zero, then the third factor is equal to zero:

$$\frac{d\Theta}{\Theta} = 0, \quad d\Theta = 0, \quad \Theta = \text{const}. \quad (3.37)$$

Thus, when a particle of dry air moves adiabatically, its temperature ( $T_i$ ) changes, but the potential temperature remains constant. This property is also applicable to the air mass as a whole: if during the movement of the air mass its potential temperature has changed, then this means that there has been an inflow to or outflow of heat from this air mass.

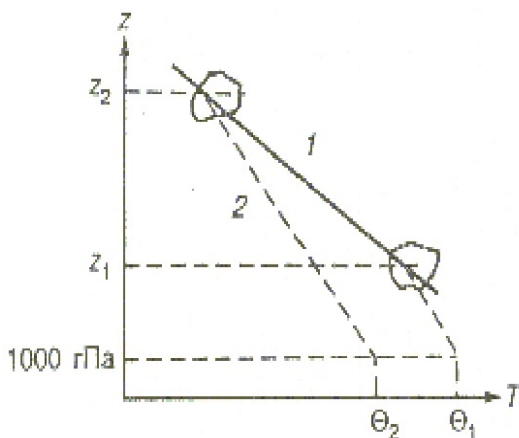
Another important property of potential temperature is the conservation of the total energy of a dry air particle at a constant value of potential temperature. This energy consists of three terms: enthalpy, potential energy of interaction between particles, and energy of particle instability. The complete proof of this property is given in [2].

### **3.4.3. Change of potential temperature with height**

The significance of potential temperature in the theory and practice of physical meteorology is also manifested in the nature of the change in potential temperature with height. This nature depends on the type of stratification in dry and moist unsaturated air. Let us consider the nature

of potential temperature change in three variants of atmospheric stratification.

1. Let  $\gamma > \gamma_a$ , i. e. the stratification is dry unstable. In this case, the stratification curve is inclined to the abscissa axis more, at a smaller angle, than dry adiabats. Let us separate particles of dry and moist unsaturated air at two levels  $z_2$  and  $z_1$ . Let these particles move dry-adiabatically to a level with an atmospheric pressure of 1000 hPa. Let us fix the potential temperature on the abscissa axis (the temperature scale). In this case, at a higher level ( $z_2$ ), the value of the potential temperature will be lower than the value at a lower level ( $z_1$ ):  $\Theta_2 < \Theta_1$ . Thus, with dry unstable stratification, the potential temperature decreases with height (Fig. 3.2).



**Fig. 3.2.** Variation of potential temperature with height at dry unstable stratification ( $\Theta_2 < \Theta_1$ ). 1 – stratification curve, 2 – dry adiabat [12]

2. Let  $\gamma = \gamma_a$ , i. e. the stratification is dry indifferent. With such stratification the potential temperature has the same value at any height. The potential temperature values are the same at the levels  $z_2$  and  $z_1$ .

This is due to the fact that in this case, at the same temperature of dry or moist unsaturated air, the stratification curve and the particle state curve (dry adiabat) coincide. For this reason, from any level, the particle will move to the level with an atmospheric pressure of 1000 hPa along the same dry adiabat. Thus, with dry indifferent stratification, the potential temperature does not change with height.

3. Let now  $\gamma < \gamma_a$ , i. e. the stratification is dry stable. In this case, the stratification curve is inclined less to the abscissa axis, at a larger angle to it, than the dry adiabat. Then  $\Theta_2 > \Theta_1$ , i. e. with dry stable stratification, the potential temperature increases with height.

So, we have considered the direction of change in potential temperature with height for different types of atmospheric stratification in cases of dry or moist air which is not saturated with water vapor.

We have considered the issue from a qualitative point of view. Let us now consider it from the quantitative side. To do this, we need to obtain a vertical potential temperature gradient. Let us take the logarithm and the height derivative of the right and left sides of equation (3.32), obtaining (3.38):

$$\frac{\partial \Theta}{\partial z} = \frac{\Theta}{T} (\gamma_a - \gamma). \quad (3.38)$$

It follows from analysis of (3.38) that with dry unstable stratification ( $\gamma > \gamma_a$ ):

$$\frac{\partial \Theta}{\partial z} < 0,$$

i. e. the potential temperature decreases with height.

With dry indifference stratification ( $\gamma = \gamma_a$ ):

$$\frac{\partial \Theta}{\partial z} = 0,$$

therefore, the potential temperature does not change with height.

And with dry stable stratification ( $\gamma < \gamma_a$ ):

$$\frac{\partial \Theta}{\partial z} > 0,$$

which means that the potential temperature increases with height.

Thus, the conclusions obtained with the help of graphs and by mathematical inferences are completely the same.

### **3.5. Thermodynamic processes in humid saturated air**

In humid saturated air, thermodynamic processes proceed differently than in dry and humid unsaturated air. At the same time, these differences are significant and exist not only quantitatively, but also qualitatively. This is confirmed by the form of the equations of the first law of thermodynamics for humid saturated air, in particular for the humid adiabatic process, as well as the wet adiabatic gradient, which differs from the dry adiabatic gradient.

#### ***3.5.1. Equation of the first law of thermodynamics for humid saturated air***

Let us isolate a particle of unit mass in the air saturated with water vapor and give this particle some amount of heat  $dq$ . The particle will spend the received heat on three processes:

- a) to change its internal energy  $c_v dT_i$ ;
- b) to do the work of expansion against external pressure forces  $Pdv_i$ ;
- c) to evaporate a certain amount of water.

The third process does not take place in cases of imparting heat to particles of dry and moist unsaturated air. Why does water evaporate in this case?

When communicating heat  $dq$  particle temperature rises. At the same time, the saturation elasticity  $E$  also increases, and when it is reached, the air is saturated. Thus, if heat is imparted to a particle of humid saturated air and the amount of water vapor remains unchanged, this particle will lose its state of saturation with water vapor. Only the evaporation process helps the particle to maintain its state of saturation with water vapor. So, it is necessary to increase the specific humidity by the value of  $dS_m$  by evaporating the water. This process consumes the amount of heat  $LdS_m$ . In this expression,  $L$  is the specific heat of vaporization (the amount of heat required to evaporate a unit mass of water). The mass of water vapor increases by  $dS_m$ , so the heat of evaporation will be  $LdS_m$ . Evaporation of water will compensate for the lack of water vapor with an increase in  $E$  by increasing the temperature of the particle. Then the equation of the first law of thermodynamics for moist saturated air will have the form:

$$dq = c_v dT_i + Pdv_i + LdS_m . \quad (3.39)$$

Equation (3.39) is the first form of the equation of the first law of thermodynamics for humid saturated air. This equation differs from equation (3.4) for dry and moist unsaturated air only by the presence of a third term on the right side. Like (3.4), equation (3.39) has a second term on the right-hand side, which includes an unknown and unmeasured factor  $dv_i$ . As in the case of the equation of the first law of thermodynamics for dry and moist unsaturated air, equation (3.39) can be easily reduced to the form:

$$dq = c_p dT_i - RT_i \frac{dP}{P} + LdS_m . \quad (3.40)$$

In (3.40)  $R$  is the specific gas constant of humid air.

Like equation (3.4) for dry and moist unsaturated air, equation (3.39) for air saturated with water vapor contains on the right side those consumption items for which the air particle will spend the received

amount of heat  $dq$ . Equation (3.40), like equation (3.14), is convenient for calculations, since it does not contain unknown and unmeasured quantities. At the same time, on the right side of (3.40), as in (3.14), the first two terms taken separately no longer represent an expense item when it receives heat from outside.

### 3.5.2. Wet adiabatic processes. Wet adiabatic gradient

If the equations of the first law of thermodynamics for moist saturated air are written for an adiabatic process, then it is possible to obtain a vertical temperature gradient in this process. This gradient is called the wet adiabatic gradient and is denoted  $\gamma'_a$ .

Thus, let  $dq = 0$  and let us write down equations (3.39) and (3.40) under this condition. As a result, we get the equations:

$$c_v dT_i + P dv_i + L dS_m = 0, \quad (3.41)$$

$$c_p dT_i - RT_i \frac{dP}{P} + L dS_m = 0. \quad (3.42)$$

We transform (3.42) and obtain an expression for the wet adiabatic gradient:

$$\gamma'_a = \gamma_a \frac{P + 0,622 \frac{LE}{RT_i}}{P + 0,622 \frac{L^2 E}{c_p R_n T_i^2}}. \quad (3.43)$$

On the right side of (3.43), the second factor is a proper fraction, that is, the denominator is greater than the numerator, since the denominator includes the value  $L$  with a large value, and even squared. However, since the second factor on the right side of (3.43) is a proper fraction, then  $\gamma'_a < \gamma_a$ . From analysis of (3.43) it follows that at the same pressure  $P$  the value of the wet adiabatic gradient  $\gamma'_a$  decreases with



increasing temperature. At low temperatures  $\gamma'_a$  approaches the value  $\gamma_a$ , and at high temperatures the wet adiabatic gradient moves away from the value  $\gamma_a$ . At constant temperature, with increasing pressure  $P$ , the wet adiabatic gradient  $\gamma'_a$  increases. The derivation of (3.43) is given in [2].

Table 3.1

**Wet adiabatic gradient ( $\gamma'_a$ ) at certain values of temperature ( $t$ ) and pressure ( $P$ )**

$P$ (hPa)	$t$ °C	$\gamma'_a$ '(°C/100 m)	$t$ °C	$\gamma'_a$ '(°C/100 m)	$t$ °C	$\gamma'_a$ '(°C/100 m)
1000	0	0.658	- 60	0.973	40	0.315
800	0	0.614	- 60	0.972	40	0.294

The reasons for the difference between wet adiabatic processes and adiabatic processes in dry and moist unsaturated air, not only quantitatively, but also qualitatively, lie in the fact that wet adiabatic processes cause phase transitions of water. Let us show that this is so.

Let a particle of moist unsaturated air rise. When moving vertically for some time, the state of the particle changes according to the dry adiabatic law. In this case, the particle temperature drops by about 1 °C when climbing every 100 m of height. In this process, the specific humidity does not change its value, since phase transitions have not yet occurred. However, the relative humidity of an adiabatically rising particle increases with decreasing temperature. This is due to the fact that saturation elasticity  $E$  is a function of temperature. More precisely, in the relative humidity formula, the saturation elasticity is in the denominator. This suggests that indirectly the temperature is in the denominator of the relative humidity formula.

Eventually, as the relative humidity increases with height, the particle comes to a level where the relative humidity reaches 100%. This

indicates that the air particle has reached a state of saturation. If the particle crosses the specified level, which is called the condensation level ( $z_{\kappa}$ ), the parameters of its state will change according to the laws of the wet adiabatic process. Above the condensation level, water vapor condenses. This is due to the drop in temperature in conditions where the relative humidity has already reached 100%. The relative humidity cannot increase above this value. The content of water vapor limits the elasticity of saturation corresponding to the temperature of the particle. In such a situation, as the particle rises further, an excess of water vapor is formed. How does the air react to it? It will transfer this “extra” water vapor to a different phase of water. In the lower layers of the atmosphere, this is the transition from vapor to water (the process of condensation). When condensing each kilogram of water vapor, a large amount of heat is released (latent heat of condensation). It is about 2500 kJ.

So, due to the release of heat in the process of condensation, the temperature of the particle above the condensation level decreases more slowly with height. This is due to the fact that part of the expansion work during the wet adiabatic process is performed due to the latent heat of condensation released during the rise of an air particle.

Below the level of condensation (during the dry adiabatic process), the work of expansion is performed only at the expense of the internal energy of the particle. As a result, the dry adiabatic gradient ( $\gamma_a$ ) is always greater than the wet adiabatic gradient ( $\gamma'_a$ ).

The curves characterizing the change in the state of an air particle during the dry adiabatic process and the wet adiabatic process (dry and wet adiabats) are also different. On the form of the aerological diagram, it can be seen that the dry adiabat is a straight line, a line that does not change its slope relative to the abscissa (temperature scale). This is due to the fact that the dry adiabatic gradient can be considered a quantity with a constant value. The wet adiabat changes its slope with respect to

the temperature scale, since the wet adiabatic gradient does not have a constant value. This value depends on atmospheric pressure and air temperature, therefore it changes with altitude. In the lower layers, the wet adiabat goes at a large angle to the abscissa axis, which coincides with the temperature scale.

Let us once again emphasize the differences between dry adiabatic and wet adiabatic processes, expressed in the change of temperature and humidity characteristics with height.

In a dry adiabatic process:

1) the temperature decreases rapidly with height, so the dry adiabatic gradient is approximately  $1^{\circ}\text{C}$  for every 100 m of height ( $\gamma_a \approx 1^{\circ}\text{C}/100\text{m}$ );

2) specific humidity does not change with height, since phase transitions do not occur in such processes, therefore, the mass of water vapor remains constant;

3) relative humidity increases with height, since the air temperature during this process drops significantly with height.

In a wet adiabatic process:

1) the temperature decreases with height more slowly than in the dry adiabatic process ( $\gamma'_a < \gamma_a$ );

2) specific humidity decreases with height due to the phase transition, since part of the water vapor passes into another phase state;

3) relative humidity does not change with height; it is always 100%, otherwise the air does not reach saturation with water vapor.

### **3.5.3. Pseudoadiabatic process**

Pseudoadiabatic processes take place in moist saturated air. However, they differ from wet adiabatic processes. This difference is not the same at different stages of these adiabatic processes.

Let us consider the rise and fall of a particle isolated in moist saturated air (above the level of condensation).

The particle rises to the level of condensation changing the temperature according to the dry adiabatic law. In other words, it loses  $1^\circ\text{C}$  when climbing for every 100 m of height.

Above the level of condensation, the temperature of a rising particle falls more slowly with height due to the release of latent heat of condensation.

At the ascent stage, a distinction is already seen between wet adiabatic and pseudoadiabatic processes. The fact is that when the particle rises during the wet adiabatic process, all the condensed moisture remains inside the particle. In the course of the pseudoadiabatic process, this moisture falls out of the particle. As a result, at a certain height, the particle is already devoid of moisture. Such a particle will descend as a particle of moist unsaturated air. In some cases, this particle is a particle of almost dry air. For this reason, when lowering, it changes its temperature increasing it according to the dry adiabatic law.

Ultimately, in a pseudoadiabatic process, the temperature values of a particle before ascent and after subsidence, even to the same height level, will be different. This is due to the fact that when the particle rises, the decrease in the temperature of the particle with height occurs more slowly than its growth when descending ( $\gamma'_a < \gamma_a$ ).

From the point of view of the physical process, this is understandable. When a particle rises, the released latent heat of condensation is not spent by it on the evaporation of moisture when lowering (there is no moisture in it any longer).

Nothing of the kind happens to a particle in the course of a wet adiabatic process. The moisture formed in the process of condensation of water vapor remains in it when the particle rises. It will evaporate when lowered. In this case, the temperature values of the particle before its ascent and after its descent at the same height level will be equal. The descent of a particle as well as its rise occurs according to the same wet adiabatic law.

Let us imagine a pseudoadiabatic process when an air mass passes over a mountain obstacle. Clouds form on the windward side of a mountain above the condensation level. The thickness of the clouds will be determined by the height of this obstacle and the height of the condensation level. The air mass will descend along the leeward slope. During the lowering, the air mass will heat up and move away from the saturation state. Of course, after lowering, this air will be warm and dry.

In areas with sufficiently high mountain obstacles, this effect is well known. This is the local wind called “foehn”. In the case of foehns, there are no clouds over the lee sides of mountains. Sometimes foehn winds are so intense that they dry out the mucous membranes of humans and animals. Foehns are associated with the dynamics of snow and ice.

In some cases, processes close to adiabatic processes take place in the atmosphere. So, for example, not all moisture falls out of the particle when it rises. Then, when the particle is lowered, it will not immediately increase its temperature according to the dry adiabatic law, but from a certain level. However, even in this case, after reaching the height level from which it rose, the particle will be warmer than before the rise. This is due to the fact that during a significant part of the path, when lowering, the particle will heat up according to the dry-adiabatic law.

The pseudoadiabatic process is characterized by the pseudopotential ( $\Theta_p$ ) and pseudo-equivalent ( $T_p$ ) temperature. Both temperatures fully reflect the pseudoadiabatic process. Some difference in their values lies in the fact that in the case of calculating the pseudopotential temperature, the particle must descend dry adiabatically to a level with a pressure of 1000 hPa. This is how any potential temperature is calculated. When calculating the pseudo-equivalent temperature, the particle should drop dry-adiabatically to the initial level. The initial level is considered to be the altitude level from which the particle rose. Which temperature will have the higher value? It depends on the ratio of the

heights of the level where the atmospheric pressure is 1000 hPa and the level from which the particle began to rise.

Pseudoequivalent is the temperature that a particle will take if it is dry-adiabatically raised to the level of condensation, pseudo-adiabatically above the level of condensation until complete condensation of water vapor in it and the subsequent complete loss of all liquid moisture from it, and then dry-adiabatically lowered to the initial level. Pseudopotential is the temperature that a particle will take if it is raised dry-adiabatically to the level of condensation, pseudo-adiabatically above the level of condensation until complete condensation of water vapor in it and complete loss of all liquid moisture from it, and then dry-adiabatically lowered to the level with a pressure of 1000 hPa.

Thus, the pseudoadiabatic process is an irreversible process, while the wet adiabatic process is a reversible process. The values of both temperatures ( $\Theta_p$  and  $T_p$ ) can be found both analytically and graphically (using an aerological diagram).

The graphical definition of pseudopotential and pseudo-equivalent temperatures is a convenient and simple matter. First, you need to find a point according to the values of temperature and pressure, from which the particle or air mass will rise along the dry adiabat to the level of condensation. The dry adiabat is the yellow or brown line on an aerological diagram that is tilted to the left (toward the y-axis).

It is easy to determine the level of condensation on an aerological diagram. This level coincides with the level of intersection of the dry adiabat and the isogram. The dry adiabat passes through the starting point plotted at the values of pressure and temperature. The isogram passes through the point plotted at the value of the dew point and the same pressure value at the initial level.

Thus, the particle rises to the level of condensation along the dry adiabatic curve. Above the level of condensation, the particle rises along

the wet adiabat. The wet adiabat is a dotted green curved line on an aerological diagram.

To obtain pseudopotential and pseudoequivalent temperatures, it is necessary to rise above the level of condensation to the level where dry and wet adiabats are parallel to each other or merge. At this level, the particle has already freed itself from the moisture present in it and has become a particle of dry air. This corresponds to the definition of a pseudoadiabatic process.

Further, the particle will descend along the dry adiabat. If the particle descends to the initial level from which it previously rose, then we will get a pseudo-equivalent temperature ( $T_p$ ). If the particle descends to the level with an atmospheric pressure of 1000 hPa, then we will get a pseudopotential temperature ( $\Theta_p$ ).

The rise along the wet adiabat above the level of condensation during the adiabatic process is marked on the aerological diagram, because the wet adiabat and the pseudoadiabat differ very little. The placement of both curves on the aerological diagram would clutter the graph. This would not lead to a great refinement of the result, but it would complicate the work. Indeed, in pseudoadiabatic and wet adiabatic processes, the rise of a particle differs only with an insignificant quantitative correction relative to each other. The descent of a particle in these processes differs significantly (not only quantitatively, but also qualitatively). Between the values of  $P$  and  $T_p$  on the one hand, and the values of 1000 hPa and  $\Theta_p$  on the other hand, there is a dry adiabat. For this reason, to calculate any of the above quantities, you can use the Poisson equation for a dry adiabatic process:

$$\Theta_p = T_p (1000/P)^{\kappa-1/\kappa}, \quad (3.44)$$

where  $P$  and  $T_p$  are parameters of the initial state of a particle of dry air, and  $\Theta_p$  and 1000 hPa are the parameters of its final state.

The pseudopotential temperature, like the pseudo-equivalent temperature, is a conservative temperature characteristic of adiabatic processes in moist saturated air. It determines the nature of the movement of particles (and air masses in general) above the level of condensation. If the value  $\Theta_p$  is changed, then there was an influx of heat to the particle or air mass (the process did not proceed according to the adiabatic law). The change of  $\Theta_p$  and  $T_p$ , serves as a quantitative criterion for the influx of heat to an air particle or air mass.

The value of  $T_p$  in (3.44) is not difficult to find if we reason as follows. Why is the temperature of the particle after its lowering higher than before it was raised? Because during the rise of the particle and the subsequent condensation of water vapor, heat is released, which is not consumed when the particle is lowered (there is nothing to evaporate—all the condensed moisture has fallen out of the particle). Then the pseudo-equivalent temperature ( $T_p$ ) can be represented as the sum of the temperature of the particle before the rise ( $T$ ) plus additions ( $\Delta T_p$ ) due to the condensation of water vapor:

$$T_p = T + \Delta T_p . \quad (3.45)$$

During the condensation process, there condenses a mass of water vapor equal to  $s$ , where  $s$  is the specific humidity. In this case, the amount of heat released is equal to  $Ls$ , where  $L$  is the latent heat of condensation. We write equations for a unit of mass, so this quantity is the specific heat of condensation. Then:

$$Ls \approx c_p \Delta T_p . \quad (3.46)$$

The released heat is used to increase the temperature of the particle. It also depends on its heat capacity, since at the same temperature objects and media are heated differently. From (3.46):  $\Delta T_p = \frac{Ls}{c_p}$ . Let us

calculate the numerical value  $\Delta T_p$ :



$$L = 2.5 \cdot 10^6 \text{ J/kg};$$

$$c_p \approx 10^3 \text{ J/kg} \cdot \text{K};$$

$$\Delta T_p = \frac{2.5 \cdot 10^6 \text{ J/kg} \cdot \text{s}}{10^3 \text{ J/kg} \cdot \text{K}} = 2.5 \cdot 10^3 \text{ s} \cdot \text{K},$$

if  $[s] = \text{kg/kg}$ . If  $[s] = \text{g/kg}$ , then

$$\Delta T_p = 2.5 \text{ s}, \quad (3.47)$$

then

$$T_p = T + 2.5 \text{ s}. \quad (3.48)$$

We substitute into (3.44) the expression for  $T_p$  from (3.48) and obtain  $\Theta_p$ .

### **3.6. Adiabatic processes in the atmosphere, including both moist unsaturated and moist saturated air**

In the real atmosphere, adjacent layers or columns of air can contain both the air not saturated with water vapor and the air saturated with it. As a result, it is necessary to evaluate the stratification of the atmosphere taking into account this factor.

#### ***3.6.1. Estimation of atmospheric stratification relative to dry adiabatic and wet adiabatic motions simultaneously***

With respect to the dry adiabatic motion, the stratification is estimated by comparing the values of the vertical temperature gradient with the dry adiabatic gradient ( $\gamma > \gamma_a$ ,  $\gamma = \gamma_a$  and  $\gamma < \gamma_a$ ). In assessing the stratification with respect to the wet adiabatic motion, the vertical temperature gradient and the wet adiabatic gradient are compared ( $\gamma > \gamma'_a$ ,  $\gamma = \gamma'_a$  and  $\gamma < \gamma'_a$ ). However, a more complete assessment of stratification is carried out by comparing the values of the vertical temperature gradient with the values of both adiabatic gradi-

ents — dry adiabatic and wet adiabatic. There are five cases of atmospheric stratification:

1.  $\gamma > \gamma_a > \gamma_a'$ . This is an absolutely unstable stratification. In this case, instability is noted up to the level of condensation and above it;

2.  $\gamma_a = \gamma > \gamma_a'$ . This is a dry-indifferent and moisture-unstable stratification. In this case, the instability manifests itself after crossing the condensation level; up to this level, an indifferent stratification develops;

3.  $\gamma_a > \gamma > \gamma_a'$ . This is a dry-stable and moisture-unstable stratification. It is similar to the previous stratification, but the vertical temperature gradient is smaller than in the previous case; up to the level of condensation, there is stability, and after crossing the level of condensation, stability is replaced by instability. The stratification is called conditionally stable. This stratification is often observed at the beginning of the development of cumulus clouds;

4.  $\gamma_a > \gamma = \gamma_a'$ . This is a dry-stable and wet-indifferent stratification. The vertical temperature gradient is even less than in the above cases; instability does not manifest itself either before the level of condensation or even after crossing it.

5.  $\gamma < \gamma_a' < \gamma_a$ . This is an absolutely stable stratification. The atmosphere is stable. The situation does not change at any position of the condensation level.

At the same time, in the real atmosphere, there are other situations when the air humidity is high at high temperatures. If the air has not reached the state of saturation with water vapor, then the situation is quite interesting. On the one hand, the calculation using the model of an atmosphere saturated with water vapor is impossible, since phase transitions do not occur in such an atmosphere. On the other hand, calculations based on a model of a dry atmosphere or an atmosphere not saturated with water vapor will lead to a noticeable error, since the amount of water vapor is quite large. The adiabatic gradient in this situation will differ from the dry adiabatic gradient. This difference can be sig-

nificant, depending on the temperature and humidity values. The only possible option is to replace the air temperature with a virtual temperature ( $T_v$ ), the vertical temperature gradient — with the vertical gradient of the virtual temperature, and the dry-adiabatic gradient ( $\gamma_a$ ) — with the adiabatic gradient of the virtual temperature ( $\gamma_{vi}$ ).

Let us get the above mentioned values. It is known that the virtual temperature is defined as

$$T_v = (1 + 0,608s)T . \quad (3.49)$$

To obtain the vertical gradient of the virtual temperature, you need to take the derivative with respect to the height of the right and left sides (3.49):

$$\frac{\partial T_v}{\partial z} = (1 + 0,608s) \frac{\partial T}{\partial z} + 0,608T \frac{\partial s}{\partial z} . \quad (3.50)$$

Let us change the signs in the right and left sides of equation (3.50); since the vertical gradient of the virtual temperature  $\gamma_v = -\frac{\partial T_v}{\partial z}$ , then:

$$-\frac{\partial T_v}{\partial z} = (1 + 0,608s) \left( -\frac{\partial T}{\partial z} \right) - 0,608T \frac{\partial s}{\partial z} , \quad (3.51)$$

$$\gamma_v = (1 + 0,608s) \gamma - 0,608T \frac{\partial s}{\partial z} , \quad (3.52)$$

where  $\gamma = -\frac{\partial T}{\partial z}$ . In most cases  $\frac{\partial s}{\partial z} < 0$ , so in (3.52) the right side is the sum, not the difference. Then  $\gamma_v > \gamma$ . So, we took into account the influence of humidity on the density of the air surrounding the particle. As for a particle:

$$\gamma_{vi} = -\frac{dT_{vi}}{dz}, \gamma_{vi} = (1 + 0,608s_i) \gamma_a - 0,608T_i \frac{ds_i}{dz} . \quad (3.53)$$

Equations (3.52) and (3.53) are written for the air with high relative humidity, but under the condition that both the particle and the surrounding air have not reached saturation. If the air is moist, unsaturated, and its humidity is low, then the criteria for its stability are the same as for dry air; if the air is moist saturated, then the stability criteria for it are defined in section 3.5. In this section, we are talking about cases when the air has not reached the state of saturation, but is close to it. At the same time, the air temperature is quite high. It is in such cases that the temperature of the particle and the air surrounding it should be replaced by the virtual temperature, the vertical temperature gradient  $\gamma$  — by the vertical virtual temperature gradient, and the dry adiabatic gradient  $\gamma_a$  — by the adiabatic virtual temperature gradient. In this case, the criteria for the stability of the atmosphere are as follows:

- 1)  $\gamma_v > \gamma_{vi}$ ;
- 2)  $\gamma_v = \gamma_{vi}$ ;
- 3)  $\gamma_v < \gamma_{vi}$ .

The differences between the densities of a particle and the air surrounding it can be replaced (in the formulas for the buoyancy force and its acceleration, as well as the instability energy) by temperature differences. In this case, the equation of state for moist air is used. As a result, differences in virtual temperatures of the air particle and the surrounding air will appear in the formulas of these quantities.

### **3.6.2. Potentially unstable and potentially stable stratification**

If we follow the vertical movement not only of the particle, but also of a sufficiently large layer of air, in which, moreover, the humidity changes strongly with height, then one can detect a noticeable change in stability or instability during the rise of this layer. It is very important to know about such a change in the stratification of air layers. These

layers move as a whole in complex terrain, in frontal zones, and in a number of other cases. This has an impact on the development of clouds of certain shapes.

Let us consider the cases of changes in stratification during the rise of air layers of a certain thickness.

1. Let in the layer  $AB$  (Fig. 3.3) the relative humidity strongly decreases from the bottom of the layer to the top; in this case, the stratification of the layer before the rise is isothermal, that is, absolutely stable ( $\gamma < \gamma_a$ ). The lower part of the layer, when it rises, will reach its condensation level much earlier than the top of this layer. Particles from the bottom rise for a short time along the dry adiabat, and then they rise along the wet adiabat; particles from the top ( $B$ ) rise along the dry adiabat for a long time to reach their level of condensation at point  $B'$ . It is clear that the stratification in the  $AB$  layer differs noticeably from the stratification in the  $A'B'$  layer. It has changed towards instability. Depending on the moisture profile in the rising layer, and depending on the thickness of the layer, the change in the stratification of the rising air can be greater or lesser. In particular, an absolutely stable layer can also turn into an absolutely unstable layer. In other words, over the same period of time, the particles from the level  $B$  lose temperature faster than the particles from the level  $A$ , since the air temperature drops more along the dry adiabat than along the wet adiabat ( $\gamma_a > \gamma_a'$ ) (Fig. 3.3).

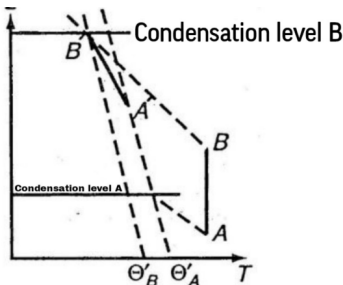


Fig. 3.3. Potentially unstable stratification

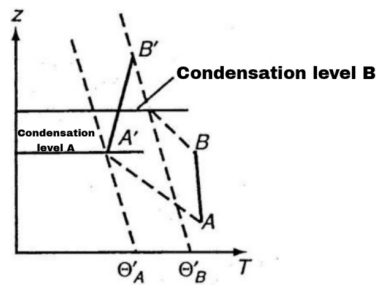


Fig. 3.4. Potentially stable stratification

2. Now let the relative humidity in the layer under consideration increase noticeably with height. In this case, as in the first case, the vertical temperature gradient  $\gamma < \gamma_a'$  (the stratification is isothermal, that is, the layer is absolutely stable). Particles from the level  $A$  move for a long time along the dry adiabat, cooling noticeably. During the same period of time, particles from the level  $B$  move for a short time to their condensation level along the dry adiabat, but mostly they move along the wet adiabat. As a result, after a certain period of time, an inversive stratification is formed in the layer  $A'B'$ . So the layer becomes even more stable (Fig. 3.4). A situation is also possible when a layer having an unstable stratification becomes stable.

In the first case, the stratification is potentially unstable; in the second case, the stratification is potentially stable. The term “potentially” is used because the change in the stratification of the layer is associated with the possible rise of the air layer. If it rises under the given conditions, there will be a change in its stratification. This change in the stratification of air layers during their rise can significantly change the nature and strength of vertical movements. Such a change can take place even within the already formed clouds, leading to their development or dissipation.

### **3.6.3. Layer method for assessing atmospheric stratification**

The layer method is more complex to use than the particle method described above. At the same time, using the layer method, in some cases, it is possible to obtain more complete information than using the particle method.

Each of these methods for assessing the stratification of the atmosphere is based on certain conditions (assumptions). Thus, the particle method assumes that the particle moves vertically in a stationary environment. In a real atmosphere, this condition is not met. Each movement of air particles is always compensated by movements of particles in the opposite direction. This is due to the fact that the atmosphere is

an elastic and inseparable medium. The effect of these movements is taken into account by the layer method.

This method was named so because it assumes that there is a layer in the atmosphere that is fairly widespread in the vertical and horizontal directions. In this layer, vertical movements in both directions are carried out. The stratification of the atmosphere using the layer method is established by observing the behavior of a layer filled with upward motions. When using the particle method, conclusions about the stratification of the atmosphere were made on the basis of observations of the behavior of individual air particles.

Let a vertical column (layer) of air with a cross section  $S'$  be filled with upward movements with positive velocity values  $w'$ . Around this column of air there is a layer in the form of a ring, filled with downward movements with negative velocity values  $w''$ . Recall that all vector quantities associated with movements along the negative semi-axis of ordinates have negative values. Downstream velocity is just such a quantity.

The layer method assumes:

1) all changes in meteorological quantities within the selected layer occur only according to the adiabatic law (no advective changes in these quantities occur);

2) in a selected column of air with upward movements and in an air ring with downward movements around this column, the mass of air does not change with height; we are talking about the fact that in the same time, the same masses of air pass through the sections of columns with ascending and descending movements.

Strictly vertical motions are associated with the formation of clouds of vertical development in the layer above the condensation level. Descending movements lead to the destruction of clouds and, ultimately, to their absence. Thus, areas of downward movement correspond to cloudless gaps between clouds. They are called intercloud gaps.

Let the speeds of movement in the air column and in the ring around this column not be equal in their values. Then, according to the second condition of the layer method, the equality of air masses with height is established by means of the corresponding ratio of the values of their cross sections. In other words, higher velocities are observed in the air column or the ring around it with the smallest horizontal sections.

In the layer method, stability criteria can be derived with respect to the vertical temperature gradient ( $\gamma$ ) and with respect to  $S'/S''$ . The derivations of these criteria are given in [2]. This textbook provides the criteria themselves:

1.  $\gamma > \gamma_{kp}$  — the atmosphere is unstable;

$\gamma = \gamma_{kp}$  — the atmosphere is neutral;

$\gamma < \gamma_{kp}$  — the atmosphere is stable.

$\gamma_{kp}$  is the so-called “critical” vertical temperature gradient, which is used in the condition when the temperature of the ascending and descending air masses is the same at any fixed level  $z$ .

In these inequalities, the critical vertical temperature gradient is determined by the formula:

$$\gamma_{kp} = \frac{\gamma_a' + \gamma_a \frac{S'}{S''}}{1 + \frac{S'}{S''}}, \quad (3.54)$$

where  $\gamma_a$  is the dry adiabatic gradient;  $\gamma_a'$  is the wet adiabatic gradient.

2.

$$\frac{S'}{S''} < \frac{\gamma - \gamma_a'}{\gamma_a - \gamma}, \quad (3.55)$$



$$\frac{S'}{S''} > \frac{\gamma - \gamma_a'}{\gamma_a - \gamma}. \quad (3.56)$$

Inequality (3.55) is the condition for instability of the atmosphere. Inequality (3.56) is the condition for stability of the atmosphere.

Let us analyze both given sets of criteria. In both cases, the stability or instability of the atmosphere also depends on the value of the vertical temperature gradient ( $\gamma$ ), and on the ratio of horizontal sections  $S'/S''$ .

The layer method gives the same results as the particle method in almost all cases. There is only one case where the layer method provides additional information regarding the stratification of the atmosphere compared to the particle method. This happens when the ascending flow is saturated with water vapor, the descending flow is not saturated with it, and the stratification of the atmosphere is conditionally stable. In this case, the value of the vertical gradient is between the values of the two adiabatic gradients ( $\gamma_a' < \gamma < \gamma_a$ ).

It follows from the analysis of (3.54) that for small values of  $S'/S''$  (the sum of the cross sections of the flows of ascending movements is very small)  $\gamma_{kp}$  approaches the value  $\gamma_a'$ . This is due to the fact that  $\gamma_a \frac{S'}{S''} \rightarrow 0$  and  $\frac{S'}{S''} \rightarrow 0$ . If  $\frac{S'}{S''}$  has a greater value,  $\gamma_a \frac{S'}{S''} \gg \gamma_a'$  and  $\frac{S'}{S''} \gg 1$ , that is why  $\gamma_{kp}$  is approaching  $\gamma_a$ .

Let us now analyze (3.55) and (3.56). The analysis leads to an interesting conclusion: in the case of an ascending flow saturated with water vapor, and a descending flow unsaturated with water vapor, and in the case of conditionally stable stratification, the atmosphere unequally manifests its resistance to flows of different horizontal sizes. The atmosphere is unstable ( $\Delta T > 0$ ) for small ascending flows. It is for them that  $S'/S''$

satisfies the condition (3.55). The same atmosphere will be stable for large updrafts. For them  $S'/S''$  satisfies the condition (3.56). In such cases, the atmosphere is said to be selectively unstable.

Let us give an example. Let the cumulus cloud develop. The beginning of its development falls on a small horizontal section. It satisfies (3.55). At this time, the conditions for cloud development are favorable. When  $\Delta T > 0$ , the buoyancy force is positive and the updraft accelerates. Inequality (3.55) is satisfied up to a certain moment, since with the development of the cloud its cross section also increases. At the moment when the inequality (3.56) is fulfilled, favorable conditions for the development of the cloud will come to an end. At  $\Delta T < 0$ , the buoyancy force will be directed downward. Under the influence of the negative buoyancy force, ascending movements in the cloud will be replaced by descending movements. If the cloud by this time has become a cumulonimbus cloud, then the downward movement will coincide with the time of precipitation. If the cloud is cumulonimbus before the sign of the buoyancy force changes, and precipitation is already falling, then the establishment of a downward movement will coincide with the time of an increase in the intensity of precipitation. If the cloud is still cumulus by this moment, then the downward movement will lead to its destruction.

The ratio  $S'/S''$  can be determined by assessing the sum of horizontal dimensions of clouds and the gaps between them.

The dimensions of the streams of ascending motions, for which the atmosphere is unstable, depend, as follows from (3.55) and (3.56), also on the vertical temperature gradient  $\gamma$ . If  $\gamma$  is close in value to  $\gamma'_a$ , then the atmosphere is unstable with respect to only very small fluxes (with a small cross section  $S'$ ). They will rapidly move vertically. In this case, cumulus clouds are poorly developed vertically and do not occupy any significant part of the sky. If the vertical temperature gradient is large (close to the values of  $\gamma_a$ ), the picture is quite different. The at-

mosphere in this case is unstable with respect to flows of ascending motions in a fairly wide range of horizontal dimensions. Then we can observe both clouds with little and considerable vertical development. In general, such clouds occupy a significant part of the sky.

Thus, we obtain more complete information using the layer method in estimating the stability of the atmosphere in comparison with the particle method in only one case. In this case, the ascending flow is saturated with water vapor, the descending flow is not saturated with it, and the atmospheric stratification is conditionally stable.

***Test questions:***

1. What is the first law of thermodynamics?
2. What is the quasi-static condition?
3. How is the equation of the first law of thermodynamics written for dry air?
4. What is an adiabatic process?
5. What dependence reflects the Poisson equation?
6. What is a dry adiabatic gradient?
7. How is the approximate dry adiabatic equation written? What is its difference from the Poisson equation?
8. What are the criteria for the stability and instability of the atmosphere used in the particle method?
9. What is the buoyancy force and the acceleration generated by this force?
10. What is potential temperature and what are its properties?
11. How to calculate potential temperature?
12. What does the direction of change in potential temperature with height indicate?
13. What is the difference between adiabatic processes in dry and humid unsaturated air?
14. How is the equation of the first law of thermodynamics written for moist unsaturated and moist saturated air? What is the difference?
15. What gradients characterize the change in particle temperature with height in the case of an adiabatic process in humid unsaturated and humid saturated air?
16. How to determine the values of these gradients?

17. How is the equation of the first law of thermodynamics written for a wet adiabatic process?
18. What is a pseudoadiabatic process? What is its difference from other adiabatic processes?
19. What are pseudopotential and pseudoequivalent temperatures? What are their properties?
20. How to calculate these temperatures?
21. How to assess the state of the atmosphere with respect to dry adiabatic and wet adiabatic motions?
22. What is absolutely unstable, absolutely stable and conditionally stable stratification?
23. How does the humidity of a particle and its surrounding air affect the buoyancy force and the acceleration created by this force?
24. What is potentially stable and potentially unstable stratification?
25. What is the layer method for assessing atmospheric stratification?
26. What are the criteria for the stability of the atmosphere according to this method?
27. In which cases is it advisable to use the criteria of the particle method to assess the stratification of the atmosphere, and in which cases, the criteria of the layer method?

## 4. TURBULENT NATURE OF ATMOSPHERIC MOTION

The continuous movement of air in horizontal layers and vertical columns of the atmosphere under conditions of an inhomogeneous pressure field can have a different character. A feature of the atmosphere is a disordered movement of individual air particles relative to general flows. This character of movement plays an important role in all atmospheric processes, in establishing and changing the state of the atmosphere. It is called the turbulent nature of the movement. At the same time, orderly movements also manifest themselves in the atmosphere. In most cases, they are “overlapped” by turbulent movements.

Why is the atmosphere characterized by turbulent motion of its particles? The degree of motion disorder increases with increasing particle velocity. The atmosphere is very light and therefore mobile.

With an ordered nature of motion in a medium, its particles move along smooth trajectories. They change little over time. Such a regime is observed at low particle velocities and is called laminar.

During turbulent motions, the particles of the medium move across the general flow or even against it. At the same time, not only the speed and direction of movement, but also other meteorological quantities experience sharp changes in time. In such cases, the meteorological quantities are said to pulsate.

### 4.1. Some concepts and characteristics associated with turbulent motion

**1. Turbulent mole.** “Mole” in physical meteorology has a completely different format and meaning than in other sciences, such as

chemistry. In atmospheric physics, moles are air particles with independent movements. These movements are not related to the speeds and directions of the general flows. In other words, turbulent moles disrupt the mode of motion in the general flow. These particles create mixing of air layers in all directions. This results in the alignment of all environment properties. The composition of the air is equalized. This can be seen in the homosphere (in a large layer from the underlying surface to a level of approximately 100 km). Thermal, humidity and other properties of the atmosphere are also aligned. The more turbulent moles there are, and the higher the degree of all manifestations of atmospheric turbulence is, the faster its properties are equalized.

**2. Mixing path ( $l$ ).** In the beginning a turbulent mole separates from the general flow. After that, it moves on its way. The turbulent mole ceases to exist at the moment of mixing with the general flow, and then it moves in this flow. The path of a turbulent mole for all the time, from the moment of its origin to the moment of completion of its existence, is called the path of mixing. The longer the mixing path is, the higher the degree of development of turbulent motions is.

**3. Specific impurity content ( $s$ ).** This is a very important concept, since the alignment of the values of the specific impurity content during turbulent mixing in some cases can even be observed visually. The specific content of impurities is the mass content of impurities in a unit mass of air. If the impurity is water vapor, then  $s$  is the specific humidity.

**4. Pulsations of meteorological quantities.** Pulsations were mentioned above, since they are included in the very concept of turbulent motion. Consider fluctuations in the velocity of particles in a turbulent atmosphere. In this atmosphere:

$$c^* = c + c', \quad (4.1)$$

where  $c^*$  is the instantaneous velocity of the air particle,  $c$  is the average velocity of the total flow; it is determined by averaging the speed over a sufficiently large period of time or in a sufficiently large volume

of air;  $c'$  is the speed of pulsation (deviation of the instantaneous speed from the average speed). In the form of the same sums, but scalar, it is possible to represent the instantaneous values of other meteorological quantities in a turbulent atmosphere. Pulsation values are a diagnostic sign of the degree of turbulence development. The higher the degree of development of turbulence, the greater the pulsations experienced by meteorological quantities. It is clear that velocity fluctuations create separation from the general flow of motion of individual particles. They are turbulent moles.

### 5. Reynolds number (parameter) ( $Re$ ).

$$Re = \frac{\rho u L}{\eta}. \quad (4.2)$$

In (4.2)  $\rho$  is the density of the air,  $u$  is the speed of movement,  $L$  is the characteristic scale of movement ( $L$  is the horizontal size of baric systems, processes or phenomena),  $\eta$  is the dynamic coefficient of molecular viscosity.

With an increase in the Reynolds number, there occurs a transition from the laminar to the turbulent nature of motion. This is facilitated by an increase in the speed of motion and a characteristic scale, but is hindered by an increase in viscosity. Indeed, it is much more difficult to bring a viscous medium into a turbulent state than a less viscous one.

**6. Richardson number ( $Ri$ ).** A very inhomogeneous medium of the atmosphere is characterized by a noticeable change in meteorological quantities with height. This cannot but affect the degree of mixing of air layers horizontally and especially vertically (the change in all meteorological quantities is greater vertically). The Reynolds number does not take this vertical variability into account. The Reynolds number may well characterize the degree of turbulence without involving other parameters in the media with a slight change in characteristics along the vertical. Such an example is a weak change in density in the hydro-

sphere. In the atmosphere, vertical variability is taken into account using the dimensionless Richardson number ( $Ri$ ):

$$Ri = \frac{g}{\Theta} \frac{\partial \Theta / \partial z}{(\partial c / \partial z)^2} = \frac{g}{T} \frac{(\gamma_a - \gamma)}{(\partial c / \partial z)^2}, \quad (4.3)$$

because  $\frac{\partial \Theta}{\partial z} = \frac{\Theta}{T}(\gamma_a - \gamma)$ .

In (4.3)  $g$  is the acceleration of gravity,  $\Theta$  is the potential temperature,  $c$  is the horizontal component of the wind speed.

$$\frac{\partial c}{\partial z} = \sqrt{\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2}, \quad (4.4)$$

where  $u$  and  $v$  are the velocity projections on the x- and y-axes.

The Richardson number can serve as a criterion for assessing the stratification of the atmosphere. Indeed, if the potential temperature decreases with height ( $\frac{\partial \Theta}{\partial z} < 0$ ), then the atmosphere is dry-unstable ( $Ri < 0$ ); if the potential temperature does not change with height (dry-adiabatic process), then the atmosphere is neutral, and  $Ri = 0$ ; if the potential temperature increases with height (dry-stable stratification), then  $\frac{\partial \Theta}{\partial z} > 0$ , and  $Ri > 0$ . So, up to the level of condensation, one can estimate the stability of the atmosphere using the Richardson number. It takes into account not only thermal but also wind stratification. This parameter quite fully reflects the stratification of the atmosphere along the vertical.

**7. Turbulent impurity flow ( $Q$ ).** Let the specific impurity content in the atmosphere at two closely spaced levels  $z$  and  $z+\Delta z$  be  $s$  and  $s+\Delta s$ , respectively. We will assume that the given values of the specific impurity content were obtained by averaging the values over a sufficiently large area at the indicated levels. Let there be turbulent mixing.



Under the influence of this process, the specific content of impurities in the air will equalize within the selected levels. If  $\Delta s < 0$  (in real conditions, the specific impurity content usually decreases with height), then at the upper level  $z + \Delta z$  this value will increase with time, and at the lower level  $z$  it will decrease. Such a change occurs because a turbulent flow of the specific impurity content occurs. If  $Q$  is the impurity mass that is transferred in the process of turbulent mixing through a unit area of the horizontal surface between the levels  $z$  and  $z + \Delta z$  per unit of time, then  $Q$  is a turbulent flow of an impurity (or other substance, or some property). This impurity is carried by turbulent moles. According to the definition of flow, its dimension is  $\text{kg}/\text{m}^2\text{s}$ . This flux is the greater, the more inhomogeneous the atmosphere in the field of a given value is, i. e., the greater the difference in the values of the specific impurity content at closer levels is:

$$Q = -A \frac{\Delta s}{\Delta z}, \quad (4.5)$$

where  $A$  is the coefficient of proportionality. Formula (4.5) in the differential form looks as follows:

$$Q = -A \frac{\partial s}{\partial z}, \quad (4.6)$$

where  $-\frac{\partial s}{\partial z}$  is the vertical gradient of the specific impurity content. It is positive when the specific impurity content decreases with height. The dimension of this gradient is

$$\left[ \frac{\partial s}{\partial z} \right] = \text{kg}/\text{kg} \cdot 1/\text{m} = \text{m}^{-1}.$$

The greater the turbulent flow  $Q$ , the higher the degree of development of turbulent motions.

**8. Turbulent exchange coefficient (mixing coefficient).** In (4.5) and (4.6)  $A$  plays the role of a proportionality coefficient. It is called the

turbulent exchange coefficient or mixing coefficient. The unit of measurement for  $A$  can be derived from (4.6):

$$A = -Q \left/ \frac{\partial s}{\partial z} \right., \quad (4.7)$$

then:

$$[A] = \frac{\text{kg}}{\text{m}^2 \text{sec}} : \frac{1}{\text{m}} = \text{kg} / \text{m} \cdot \text{sec} .$$

The higher the degree of development of turbulent motion is, the greater the coefficient of turbulent exchange  $A$  is.

**9. Turbulence coefficient ( $k$ ).** The turbulence coefficient has a very wide application in atmospheric physics. Numerically it is equal to:

$$k = \frac{A}{\rho}, \quad (4.8)$$

so its dimension is:

$$[k] = \left[ \frac{A}{\rho} \right] = \frac{\text{kg} / \text{m} \cdot \text{sec}}{\text{kg} / \text{m}^3} = \text{m}^2 / \text{sec} .$$

Turbulent motion is described by various parameters. They are definitely related. So, for example, wind speed fluctuations ( $c'$ ) determine the mixing path ( $l$ ), as well as the coefficients  $A$  and  $k$ . In turn, these coefficients depend on many factors: on the stratification of the atmosphere (whether it is stable or unstable and to what extent), on the vertical gradient of wind speed, on the properties of the underlying surface (its roughness and thermal inhomogeneity). The mixing path  $l$ , the coefficients  $A$  and  $k$ , and, consequently, the degree of development of the turbulent nature of motion increase with height. Thus, the mixing path  $l$  near the underlying surface cannot be large. Indeed, the underlying surface prevents the movement of a turbulent mole down from the level of its origin. However, the mass of air which is carried in the process of turbulent mixing, from top to bottom and from bottom to top, should be the same. Therefore, the displacement of turbulent particles (moles)

upwards is possible only at the same distance as downwards, that is, it cannot be large. As a result, the mixing path  $l$ , the coefficients  $A$  and  $k$ , cannot be large near the underlying surface. As altitude increases, turbulent moles can travel much greater distances than at low levels. The change in the height of the mixing path and the turbulence coefficient was expressed by the German scientist Prandtl as a linear relationship:

$$l = \kappa (z+z_{00}), \quad (4.9)$$

$$k = k_0 + az. \quad (4.10)$$

In (4.9)  $z_{00}$  is the roughness parameter of the underlying surface. It represents the thickness of the layer in which the speed of the average translational motion is equal to zero. In this case, the speeds of individual particles can be significant, but during turbulent motion they are directed in different directions. As a result, the algebraic sum of multidirectional motions is equal to zero. Most often, the roughness parameter has values from fractions of a centimeter to several centimeters. In (4.9)  $\kappa = 0.38$  is the dimensionless constant of Theodore von Kármán. It was determined by him empirically. In (4.10)  $k_0$  is the value of the turbulence coefficient at the height  $z = 0$ ;  $a$  is the value that determines the growth rate of  $k$  with height. In practice, it is equal to the value of  $k$  at a height of  $z_1 = 1$  m.

Since  $k_0$  is small compared to the parameter  $a$ , then  $k_l = k_0 + a \approx a$ .

The layer of the atmosphere in which the turbulence coefficient  $k$  under any conditions increases with height is called the surface layer. There are other definitions of the surface layer, for example, according to temperature characteristics. The surface layer is the lowest part of the atmosphere (on average, from the underlying surface to 100–150 m). At the upper boundary of the surface layer, the values of the turbulence coefficient vary from fractions of  $\text{m}^2/\text{sec}$  to tens of  $\text{m}^2/\text{sec}$ , which depends on the stability of the atmosphere, wind speed and its change with height, and the roughness of the underlying surface.

**10. Molecular impurity flux ( $Q_m$ ).** Molecular flows of impurities are observed in any layer of the atmosphere and at any time, but these

flows are blocked by turbulent flows. The only exception is a viscous sublayer a few millimeters thick. In all other layers, molecular fluxes in the atmosphere are neglected. Molecular flows of the specific impurity content have the same units of measurement as turbulent flows. They are also proportional to the gradient of the specific impurity content:

$$Q_M = -\rho D \frac{\partial s}{\partial z}, \quad (4.11)$$

where  $D$  is the molecular diffusion coefficient, which is measured in the same units as the turbulence coefficient ( $\text{m}^2/\text{s}$ ). The molecular diffusion coefficient depends on the air temperature. Due to the ratio of the turbulence coefficient and the molecular diffusion coefficient, the ratio of the turbulent and molecular impurity flows is  $\frac{Q}{Q_M} \approx 10^4\text{--}10^6$ .

Turbulent flows of the specific impurity content can be written, like the molecular flow, in the form (4.11), that is, using the air density value:

$$Q_x = -A' \frac{\partial s}{\partial x} = -\rho k' \frac{\partial s}{\partial x}, \quad (4.12)$$

$$Q_y = -A' \frac{\partial s}{\partial y} = -\rho k' \frac{\partial s}{\partial y}, \quad (4.13)$$

$$Q_z = -A \frac{\partial s}{\partial z} = -\rho k \frac{\partial s}{\partial z}. \quad (4.14)$$

In (4.12–4.14)  $A'$  and  $k'$  are the coefficients of turbulent exchange and turbulence in the horizontal direction, and  $A$  and  $k$  — in the vertical direction. In this case,  $A'$  and  $k'$  are one or two orders of magnitude larger than  $A$  and  $k$ .

## 4.2. Heat fluxes in a turbulent atmosphere

Heat enters the troposphere mainly from the underlying surface, since the atmosphere is mainly a transmitting medium for solar radia-

tion. The process of heat exchange between the underlying surface and the atmosphere is quite multifaceted and occurs due to:

- 1) turbulent heat transfer;
- 2) convective heat transfer;
- 3) radiant heat transfer;
- 4) phase transitions of water (condensation and sublimation of water vapor, evaporation and freezing of water);
- 5) molecular heat transfer.

Radiant heat transfer was considered in Section 2. The effect of phase transitions will be considered in Section 5. Molecular fluxes are small, so the contribution of molecular heat transfer to the overall process of heat transfer between the underlying surface and the atmosphere is negligible in all layers. The only exception is the viscous sublayer.

In this section, we will consider convective and turbulent heat fluxes. Convective flow takes place in a moving medium, which is the atmosphere. Moving air particles transfer the heat content  $c_p T$  through a single area along the normal to it per unit of time. This is how a convective heat flux ( $Q_\kappa$ ) is formed. This flow is due to the ordered movement of air at an average speed  $c$ . By definition, the flow has the dimension:  $[Q_\kappa] = \text{J/m}^2\text{sec}$ . So, in the formula for the convective flow, there should be heat content (this is what is transferred), and air mass (this is what heat content is transferred by); the transfer is carried out through a unit area per unit of time, therefore:

$$Q_\kappa = c_p T \rho c. \quad (4.15)$$

The correctness of the record is determined by the units of measurement:

$[c_p T] = \text{J/kg K} \cdot \text{K} = \text{J/kg}$ ;  $[\rho c] = \text{kg/m}^3 \cdot \text{m/sec} = \text{kg/m}^2\text{sec}$ ,  
hence:  $[Q_\kappa] = \text{J/kg} \cdot \text{kg/m}^2\text{sec} = \text{J/m}^2\text{sec}$ .

The horizontal component of the average wind speed is hundreds of times greater than the vertical one, so the convective heat flux ( $Q_\kappa$ ) is

mainly a horizontal component, which is called the advective heat flux. The vertical component is called the actual convective heat flux.

In contrast to the convective heat flux, the turbulent heat flux is due to fluctuations in the wind speed. Turbulent heat flow is different from other turbulent flows and from other heat flows. The fact is that a substance or property (what is transferred in the process of turbulent mixing) must satisfy three general conditions:

1) the substance must be constant in a turbulent mole as it moves until it mixes with the general flow; this condition is called the indestructibility of the substance;

2) when a turbulent mole is mixed with the general flow, the amount of substance carried by the turbulent mole must be preserved; this condition is called the conservation of quantity of substance;

3) the transported substance should not affect the movement of turbulent particles; this property is called the passivity of the substance.

In the case of transfer of the specific impurity content in the process of turbulent mixing, all three conditions are met, so the turbulent flow of the specific impurity content can be written in the form (4.6). As for the air heat content  $c_p T$ , it does not satisfy the first and third conditions. Indeed, in the process of motion, in particular in the vertical direction, the temperature of the particles changes. It, moreover, has a very strong effect on the movement of these particles. However, there are conservative temperature characteristics. For example, up to the level of condensation during adiabatic movements of particles, their potential temperature remains constant. The role of an indestructible and passive substance in this case is played by the potential heat content  $c_p \Theta$ . This is what will be transferred in the process of turbulent mixing,

$$Q_z = -c_p A \frac{\partial \Theta}{\partial z} = -c_p \rho k \frac{\partial \Theta}{\partial z}, \quad (4.16)$$

and in horizontal directions:

$$Q_x = -c_p A' \frac{\partial \Theta}{\partial x} = -c_p \rho k' \frac{\partial \Theta}{\partial x}, \quad (4.17)$$

$$Q_y = -c_p A' \frac{\partial \Theta}{\partial y} = -c_p \rho k' \frac{\partial \Theta}{\partial y}. \quad (4.18)$$

The values of  $A'$  and  $k'$  are much larger than  $A$  and  $k$ , so the horizontal components of the heat flux  $Q_x$  and  $Q_y$  are certainly greater than its vertical component  $Q_z$ . However, horizontally, heat transfer is carried out mainly due to advection, and vertically, just due to turbulent mixing. As a result, when simplifying the equations containing turbulent heat fluxes, the horizontal components  $Q_x$  and  $Q_y$  are omitted. The vertical component  $Q_z$  remains in these equations.

### 4.3. Heat and moisture inflows in a turbulent atmosphere

The influx of heat (moisture) is the difference between the heat (moisture) flows entering any volume or mass of air and the heat (moisture) flows leaving this volume or mass over a certain time. The sign and value of the influx of heat (moisture) determine the temperature (humidity) of the volume or mass of air.

The complete derivation of the heat (moisture) influx equation in a turbulent atmosphere is given in [2].

The general form of the heat inflow equation in a turbulent atmosphere can be represented as follows:

$$\frac{\partial T}{\partial t} = - \left( u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} \right) + w(\gamma - \gamma_a) + \frac{\partial}{\partial z} k \frac{\partial \Theta}{\partial z} + \frac{\varepsilon_a + \varepsilon_n}{c_p}, \quad (4.19)$$

where  $\frac{\partial T}{\partial t}$  is the change of temperature ( $T$ ) in time ( $t$ ) at a fixed point in space;  $u$ ,  $v$ ,  $w$  are projections of the air velocity on the  $x$ -,  $y$ -, and  $z$ -axes;  $k$  is the turbulence coefficient,  $\Theta$  is the potential temperature,  $c_p$

is the specific heat capacity at constant pressure,  $\varepsilon_r$  and  $\varepsilon_n$  are heat inflows due to radiant heat transfer and phase transitions of water in the atmosphere, respectively. Thus, according to (4.19), the change in temperature with time at any point depends on the values of the four terms on the right side of the equation.

The first term (together with the minus sign) is called the advective term. It determines the change in temperature under the influence of air transfer in the horizontal direction. Recall that advection is the movement of air horizontally. This term includes the horizontal components of the speed of movement and temperature changes in horizontal directions.

The second term is called the convective term. It determines the change in air temperature under the influence of vertical movements (convection). The second term includes the vertical velocity, the vertical temperature gradient, and the dry adiabatic gradient.

The third term on the right side of (4.19) is the turbulent term. It includes the turbulence factor.

The fourth term on the right side of (4.19) is the non-stationary term. It is associated with heat inflows during radiant heat transfer and phase transitions of water.

In a number of cases, equation (4.19) can be simplified. For this purpose, in (4.19) one should take into account only those terms that in a particular process make the greatest contribution to the change in temperature with time at a fixed point in space. Let's take a few examples.

1. Let non-periodic temperature changes in the free atmosphere, that is, above the boundary layer, be calculated. In this case, on the right side of (4.19) we will take into account only two terms. First of all, temperature changes occur due to the influence of advection of heat or cold. Note that in a short period of time, advection processes of different nature will not compensate for each other. As a rule, either heat ad-



vection or cold advection is possible in a short period of time. In addition, non-periodic manifestations include vertical air movements due to baric formations. Thus, ascending movements are observed in the cyclone as a whole, and descending movements are observed in the anticyclone. Vertical movements in both directions lead to a change in air temperature. When moving up (along the positive semi-axis of ordinates), the vertical speed will have positive values. When moving down, the vertical speed has negative values. Note that during vertical motions in moist saturated air, for example, in a cloud, the dry adiabatic gradient  $\gamma_a$  should be replaced by a wet adiabatic gradient  $\gamma'_a$ .

2. In cases where daily fluctuations in the boundary layer are determined, only the third term can be taken into account on the right side of (4.19). This is due to the fact that in the system “the underlying surface–the atmosphere” it is the underlying surface that experiences the greatest diurnal temperature fluctuations. If it heats up, then a significant part of the heat is transferred from it to the atmosphere. If the underlying surface cools, then the atmosphere transfers heat to it. The turbulent exchange in the vertical direction is responsible for the daily temperature variation in the lower layers of the atmosphere. Note that (4.19) can contain turbulent terms not only in the vertical direction, but also in the horizontal directions. It depends on the task setting.

3. Let us now be interested in the change in temperature at a fixed point. This point can be any area or point. Let this fixed point be located under the conditions of a non-uniform underlying surface. The air moving above it will be under the influence of such a surface. The air in this case will have different properties in its different parts in the temperature field. The system will be in a non-equilibrium state. Then, due to its desire for balance, this air system will equalize its properties. As a result, the greatest contribution to the sum of the right-hand side of (4.19) will be made by those terms that are associated with bringing the system to an equilibrium state. The vertical equalization of air proper-

ties will occur primarily through turbulent mixing, the horizontal equalization will occur due to advective process.

Ultimately, the right side of (4.19) will contain only two terms: the first (advective) and the third (turbulent).

4. Let it be required to obtain the average temperature for a year or another long period of time, for example, for a season. Then (4.19) will have a very simple form. When averaged over a long period of time, both the advective term and the convective term will be close to zero. Indeed, both heat advection and cold advection are observed in a fixed point over a long period of time. In addition, both ascending and descending air movements will occur. The change in temperature over time for such a problem and with such an approach will also be considered equal to zero.

As a result, (4.19) will be the sum of heat inflows due to turbulent heat inflows in all directions, radiant heat transfer, and phase transitions of water in the atmosphere. The sum of all these inflows will be equal to zero. The considered simplified versions of the general equation of heat influx in a turbulent atmosphere are called particular equations of heat inflow and give a fairly good approximation when considering various processes and solving a number of problems.

Now consider the influx of moisture in a turbulent atmosphere. The equation for the influx of water vapor under conditions of a turbulent atmosphere is very similar to the equation for the influx of heat under the same conditions. It looks like:

$$\frac{\partial s}{\partial t} = - \left( u \frac{\partial s}{\partial x} + v \frac{\partial s}{\partial y} \right) - w \frac{\partial s}{\partial z} + \frac{\partial}{\partial z} k \frac{\partial s}{\partial z} + \varepsilon_n, \quad (4.20)$$

where  $s$  is specific humidity,  $\varepsilon_n$  is change in specific humidity under the influence of phase transitions of water; the value and sign of the last quantity depends on the type of phase transition — the evaporation of water or the condensation of water vapor contained in the released mass or the released volume of air.

How do equations (4.19) and (4.20) differ? First of all, the differences are related to the fact that in (4.19) the potential temperature ( $\Theta$ ) appears in the turbulent term, and in (4.20) the specific humidity ( $s$ ) appears in it. Specific humidity is also present on the left side of (4.20). In addition, specific humidity is included in the advective term and in the convective term (4.20). In (4.19) in all these cases, the air temperature ( $T$ ) appears. This is understandable, since in one case the equation describes the heat gain, and in the other case the equation describes the transfer of water vapor (moisture) in a turbulent atmosphere.

As for differences in turbulent terms, it is known that in the process of turbulent mixing, the transferred substance (or property) must be preserved. In this process, the given substance must be conservative.

A conservative temperature characteristic in a dry adiabatic process is the potential temperature. In adiabatic processes in humid saturated air, conservative temperature characteristics are pseudo-equivalent and pseudo-potential temperatures.

Conservative moisture characteristics are specific humidity and mixture ratio. Specific humidity is preferred in calculations and in large equations. Specific humidity is connected with other characteristics of humidity and atmospheric pressure. If we consider the change in humidity characteristics in different directions, then these changes, like changes in other meteorological quantities, occur most rapidly in the vertical direction. Specific humidity varies much less with height than absolute humidity or water vapor pressure. This is because air density decreases with height. As a result, in a unit volume, with which absolute humidity is associated, the mass of air becomes less and less. The mass of water vapor included in the total mass of air also becomes smaller. For this reason, the mass of water vapor in a unit volume (absolute humidity) decreases with height due to a general decrease in air density. Nothing of the kind happens with specific humidity values. Although air density decreases with height, specific humidity is the mass of water vapor in a unit mass of moist air. As the height increases

with decreasing air density, more and more volumes of air should be taken to obtain a unit mass. In it, the mass of water vapor (specific humidity) will be determined.

As for the elasticity of water vapor, it is the partial pressure of water vapor, therefore, it decreases with height even faster than the total atmospheric pressure.

It makes no sense even to consider all other characteristics of humidity from the point of view of conservatism. They are either functions of a very variable quantity — air temperature (as saturation elasticity  $E$ ), or differences of non-conservative quantities (deficit of elasticity  $d = E - e$ , and dew point deficit  $\Delta = T - \tau$ ). Ultimately, it is most correct in the equation for the influx of water vapor in a turbulent atmosphere, among other characteristics of humidity, to choose specific humidity, as is done in (4.20). It is specific humidity that is both the most conservative characteristic of humidity and the most convenient to use. Analysis of (4.20) leads to the same conclusions as analysis of (4.19), but only with respect to specific humidity. So, the change in specific humidity in time at a fixed point in space depends on four factors: the advection of water vapor, its convection, turbulent mixing and phase transitions of water.

Water vapor advection is associated with horizontal transport of water vapor by an ordered flow at an average speed. Water vapor convection is caused by the transfer of water vapor by ordered flows with an average speed in the vertical direction. Turbulent mixing is carried out in various directions, mainly in the vertical direction. During phase transitions, the amount of water vapor can both decrease and increase.

***Test questions:***

1. What are the characteristics of motion in the atmosphere?
2. What is a turbulent mole?
3. How is the mixing path related to the degree of turbulence development?
4. How does the blending path change with height?

5. What are meteorological pulsations?
6. How do Reynolds and Richardson numbers characterize atmospheric turbulence?
7. What does the sign of the Richardson number mean?
8. What are the coefficients of turbulent exchange and turbulence?
9. How do these coefficients change with height?
10. What is the turbulent flow of the specific impurity content and the molecular flow of the impurity?
11. How are these streams recorded and what is their difference?
12. What are convective and turbulent heat flows? How are they recorded?
13. What is the influx of heat and the influx of water vapor?
14. What is the equation for heat gain in a turbulent atmosphere?
15. What are the particular types of this equation?
16. What is the equation for the influx of water vapor in a turbulent atmosphere?
17. What are the similarities and differences between the equations of heat influx and water vapor influx in a turbulent atmosphere?

## **5. WATER REGIME OF THE ATMOSPHERE**

Water in all its states plays a very important role in the ongoing atmospheric processes. The source of water in the atmosphere is the underlying surface. Moisture evaporates from this surface and is carried by ordered and disordered air movements to higher layers. Water vapor trapped in these layers is sometimes transferred horizontally over very long distances from the source of moisture. Calculations have shown that the moisture in the atmosphere of our planet is largely oceanic in nature.

Vertical motion and the associated adiabatic cooling of the air lead to the formation of clouds. In addition, fogs and haze can form near the underlying surface.

### **5.1. Evaporation. Humidity distribution in time and space**

Evaporation is determined by the mass of water that evaporates from a unit area per unit of time. In other words, evaporation is the flow of water vapor. Evaporation depends on many factors, the main of which are:

- 1) temperature of the evaporating surface;
  - 2) stratification of the surface layer of the air;
  - 3) wind speed;
  - 4) roughness of the underlying surface;
  - 5) dimensions of the evaporating surface;
  - 6) the nature of the vegetation cover of the evaporating surface;
  - 7) the presence and nature of the snow cover of the evaporating surface.
- Evaporation calculation formulas are based on its dependence on one or more of the above factors.

We should not forget the fact that the atmosphere cannot take an infinite amount of water vapor. Its content in the atmosphere is limited by the value of saturation elasticity. The value of the saturation elasticity depends mainly on the air temperature.

Evaporation from water surface is different from evaporation from land. In meteorology there is a concept of evaporation from water-saturated land which is determined by the mass of water that would evaporate from a unit area of the land surface per unit of time if there was an unlimited supply of moisture in the soil. This concept is applicable only to the soil, because in the case of water, the moisture reserves are not limited.

It is interesting that the difference between the values of the would-be evaporation from land with a high content of moisture and evaporation from water surface can serve as a climatic characteristic of the area. If the difference between these values is small, then the area has highly moistened or waterlogged (marshy) soil. If these values differ significantly, then the area has semi-desert or even desert climate features.

The content of water vapor in the atmosphere varies in space and time. If we talk about the change in the content of water vapor and, consequently, the change in the characteristics of humidity (hygrometric values) in space, then they are not the same in different directions.

In horizontal and close to horizontal directions, the content of water vapor depends on the nature of the underlying surface and on the nature of atmospheric processes. Of course, air humidity is higher above water bodies, especially large ones. In addition, high humidity is observed over the coasts and over wetlands. As for atmospheric processes, the highest humidity is noted in the areas of cyclonic formations, the lowest — in the areas of anticyclonic formations. The combination of atmospheric and physical-geographical conditions creates a rather “variegated” field of humidity.

In the vertical direction, the moisture characteristics change with height rather quickly. The main moisture-bearing layer is the lower five-kilometer layer. To understand how the content of water vapor changes with height, researchers climbed mountains before using aircraft. There, at different heights, the researchers carried out their measurements, then, according to the data obtained, they proposed formulas.

For the troposphere as a whole, if the lowest layer is not distinguished in it, the formulas of the Austrian scientist Julius Gunn and the German scientist Reinhard Sührling are usually used. The indicated lower layer is a boundary layer approximately 1.5–2.0 km thick. For the troposphere as a whole, the formulas are based on actual measurements in the Alps at the end of the 19th century.

In 1889, Gann presented the change in water vapor elasticity with height in the form:

$$e = e_0 \cdot 10^{-z/6.3}, \quad (5.1)$$

where  $e$  is the elasticity of water vapor at a height  $z$ ,  $e_0$  is the elasticity of water vapor at the underlying surface.

The analysis of (5.1) leads to the conclusion that the content of water vapor, judging by the change in its partial pressure ( $e$ ), decreases very quickly with height. Indeed, according to (5.1), the elasticity of water vapor at a height of 6.3 km is 10 times less than at the underlying surface, and at a height of 12.6 km it decreases by a factor of 100. Thus, analysis of (5.1) confirms the conclusion that the main moisture in the atmosphere is concentrated in the lower and partly in the middle atmosphere.

In 1900, Zührling largely confirmed the results of Gann's research. At the same time, Zührling proposed not only a formula for changing with the height of water vapor elasticity, but also a formula for changing with the height of specific humidity:

$$e = e_0 \cdot 10^{-z/6 - z^2/120}, \quad (5.2)$$



$$s = s_0 \cdot 10^{-z/9 - z^2/120}, \quad (5.3)$$

where  $s$  is the specific humidity at the height  $z$ , and  $s_0$  is the specific humidity at the underlying surface. Comparison of (5.1), (5.2), and (5.3) led to the following conclusions:

1) according to the Züring equation (5.2), the elasticity of water vapor decreases with height even faster than according to the Gunn equation (5.1);

2) specific humidity decreases with height more slowly than the elasticity of water vapor; this confirms the conclusion made in section 4 that specific humidity is a more conservative characteristic than other hygrometric quantities.

These findings were later confirmed by other studies. Some details have come to light. For example, it turned out that an increase in temperature with height is accompanied by an increase in specific humidity in the same direction. This feature is most clearly observed in winter in the lower troposphere.

The change in humidity over time can be periodic and non-periodic.

Non-periodic changes are associated with the change of baric formations, the passage of atmospheric fronts. When atmospheric fronts pass through cyclone systems, humidity increases. Anticyclones due to downward air movements are associated with lower humidity values compared to cyclones.

If the atmospheric situation does not change, which happens most often when the weather is in anticyclones, then the daily variation of each of the hygrometric values is traced. In cyclones there are heterogeneous air masses, therefore, during their passage, non-periodic changes in humidity are observed. This reduces the possibility of manifestation of the diurnal variation of hygrometric values.

The well-defined diurnal variation of humidity characteristics is due to the diurnal change in the temperature of the underlying surface and the air layer adjacent to it, as well as the change in the intensity of tur-

bulent exchange during the day. Especially significant daily fluctuations in humidity characteristics (hygrometric values) are demonstrated over the continents in the warm half of the year. Recall that all hygrometric quantities are divided into two groups. The first group includes vapor pressure ( $e$ ), absolute humidity ( $a$ ), specific humidity ( $s$ ), mixture ratio ( $r$ ), that is, those characteristics that indicate the quantitative content of water vapor in the atmosphere.

During the warm period, the humidity characteristics of the first group over the continents experience complex daily fluctuations. It traces two maxima (at 7–10 o'clock and at 19–22 o'clock) and two minima (before sunrise and at 15–17 o'clock). This diurnal variation is explained as follows.

In the morning, after sunrise, the temperature of the underlying surface increases, because as the sun rises, the angle of incidence of the sun's rays becomes more and more favorable for the absorption of solar radiation by the underlying surface. Together with the temperature of the underlying surface and the layer of the air adjacent to it, the rate of evaporation of moisture from the earth's surface also increases. At the same time, the intensity of different ascending movements with the help of which water vapor enters the atmosphere also increases. As a result, the air humidity in the boundary layer increases. This growth is observed up to 7–10 hours. In the near-noon hours, with increasing temperature, the intensity of turbulent exchange increases so much that water vapor is also carried to higher layers — beyond the lower part of the boundary layer (up to 200–250 m). As a result, water vapor spreads to much larger volumes than in the morning hours. Therefore the humidity in the layer from the underlying surface to the levels of 200–250m decreases to the minimum values. The partial outflow of water vapor beyond the lower part of the boundary layer is not compensated by additional evaporation from the land surface during the warm period, since it is not sufficiently moistened. At the same time, in higher layers (above 200–250 m) there is a maximum in the content of water vapor,

which can get here precisely under the conditions of a well-developed turbulent exchange (in the daytime).

In the evening, the turbulent exchange weakens. For this reason, although the rate of evaporation decreases with air temperature, humidity increases again by the time of the second maximum at 19–22 o'clock. After this time, the air temperature, following the temperature of the underlying surface, decreases so much that the rate of evaporation and the intensity of turbulent mixing weaken. As a result, humidity drops and reaches its minimum before sunrise. At this time, the lowest air temperatures are observed. The processes of water vapor condensation also contribute to the formation of this minimum. The most favorable conditions for them are created before sunrise. During this process, part of the water vapor changes into a liquid phase throughout the night, which leads, for example, to the formation of dew or fog.

The main features of the diurnal variation of moisture characteristics are similar both at the land surface and at higher levels within the boundary layer. However, with height, the time of occurrence of maximum and minimum values shifts to later dates (within the time intervals of the onset of maximum and minimum daily values indicated above). In addition, with height, the amplitudes of the diurnal variation of humidity characteristics decrease.

The type of daily variation of these humidity characteristics is different over the land in the cold period, and over the water surface, as well as over the heavily moistened land, throughout the year. It becomes simple, that is, with one maximum and one minimum. The maximum occurs around noon at 10–12 a. m. The minimum is before sunrise. This is due to the fact that in winter over land, and in any time over a highly moistened surface and water, the intensity of turbulent exchange does not increase in the near-noon hours as much as in warm time over dry soil. As a result, water vapor as a rule is not carried from the lower part of the boundary layer to higher layers. If this happens,

for example, in the warm season, then the removal of water vapor to higher layers is compensated by an additional influx of water vapor.

The characteristics of the hygrometric values of the second group, which show how close the air is to the state of saturation, have other types of diurnal variation. Let us consider the diurnal behavior of the brightest hygrometric quantity of the second group, the relative humidity ( $f$ ). The analysis of the relative humidity formula ( $f = \frac{e}{E} \cdot 100\%$ )

leads to the conclusion that this value has a pronounced diurnal variation. This is due to the fact that the numerator and denominator in the relative humidity formula have well-defined daily fluctuations, especially in summer. Saturation elasticity ( $E$ ) is a function of temperature, so the relative humidity reaches its maximum values during the lowest temperatures of the underlying surface and air — before sunrise. After that, the relative humidity decreases with rising temperatures until the mid-afternoon hours. During the onset of the highest temperatures, it reaches its minimum values (approximately at 15–17 o'clock). Thus, the daily course of relative humidity is simple — with one maximum and one minimum.

## 5.2. Phase transitions of water in the atmosphere

Water vapor is the only gas in the Earth's atmosphere that under certain conditions changes its phase state.

The following phase transitions of water in the atmosphere are possible: water–water vapor, water vapor–water, ice–water vapor, water vapor–ice, water–ice and ice–water.

When determining the conditions for phase transitions, it is necessary to determine the dependence of saturation elasticity on various factors. Chief among them is air temperature. Saturation elasticity ( $E$ ) is the maximum value of water vapor pressure ( $e$ ) at each temperature.

The dependence of saturation elasticity values on air temperature is reflected in the Clausius–Clapeyron equation:

$$\frac{dE}{dT} = \frac{L_{1,2}}{T(v_2 - v_1)}, \quad (5.4)$$

where  $L_{1,2}$  is the latent heat of the phase transition from phase state 1 to phase state 2 at temperature  $T$ ;  $(v_2 - v_1)$  is the difference in volumes in two phase states. Magnus proposed a formula for the dependence of saturation elasticity on temperature in the form:

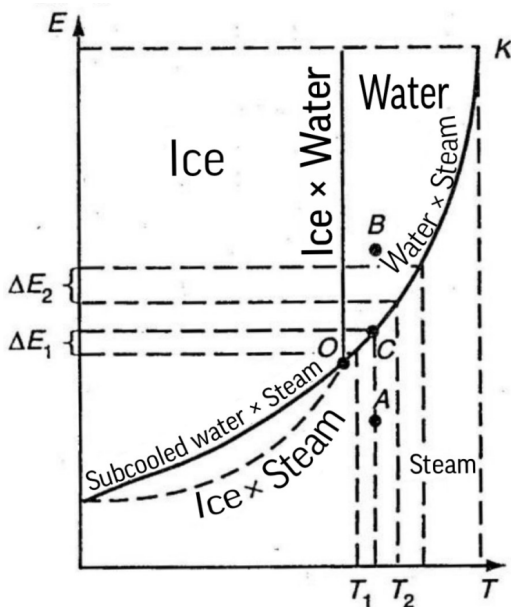
$$E = E_0 \cdot 10^{at/b + t}, \quad (5.5)$$

where  $a$  and  $b$  depend on the phase state: for water  $a = 7.63$  and  $b = 241.9$ ; for ice  $a = 9.5$  and  $b = 265.5$ . The values of  $E$  for water and  $E_i$  for ice at various temperatures are given in the “Psychrometric Tables”.

The graphic representation of the dependence of saturation elasticity ( $E$ ) on temperature ( $T$ ) allows us to consider the dependence of  $E$  on the state of aggregation of the evaporating surface. The graph of  $E$  versus  $T$  consists of three curves: the evaporation curve, the sublimation curve and the melting curve (Fig. 5.1).

Let us consider the evaporation curve first. This curve passes through several known points. One of them is the so-called triple point. On our graph, it is indicated by the point  $O$  ( $t = 0.01^\circ \text{C}$  or  $T = 273.16 \text{ K}$ ,  $E = 6.1114 \text{ hPa}$ ). In it, all three phases of water (liquid, solid and gaseous) are in a state of equilibrium. At temperatures above the triple point, water can be in liquid or gaseous states. At temperatures below the triple point, the saturation elasticity increases very slowly with increasing temperature; at temperatures above the triple point, with its increase, the saturation elasticity increases more and more, at first slowly, then faster and faster. In other words, the dependence of  $E$  on  $T$  is non-linear and is determined by the law of the exponential function.

The next point through which the evaporation curve passes is called the boiling point of water at normal pressure ( $t = 100^{\circ}\text{C}$ ,  $E = 1013.25 \text{ hPa}$ ). The third point through which the evaporation curve passes is called the critical point ( $t_{kp} = 374^{\circ}\text{C}$ ,  $E_{kp} = 221,000 \text{ hPa}$ ). The critical point is characterized by the absence of a distinction between the liquid and gaseous states of water. There is a constant transition from a liquid state to a gaseous state in it, and vice versa. The latent heat of vaporization (or condensation) is zero. D. I. Mendeleev called the critical temperature the temperature at which the cohesive forces between the molecules of the liquid disappear. If the temperature is higher than the critical temperature, then the gas under no pressure can change its state of aggregation.



**Fig. 5.1.** Dependence of saturation elasticity of water vapor on temperature and phase state of the evaporating surface [1]

If the temperature is below the triple point, then water can be not only in a solid, but also in a supercooled liquid state. In the atmosphere, the state of supercooled water is characteristic, fairly stable, and is called metastable. As a result, at temperatures below the triple point, two curves are located on the graph. They are part of the evaporation curve and the sublimation curve (dashed line). These two curves are close to each other because the values of saturation elasticity over water and ice do not differ greatly, but even these small differences are very important for atmospheric processes.

So, the evaporation curve at temperatures below the triple point is a curve of dynamic equilibrium of two phases — supercooled water and steam. The sublimation curve is a curve of dynamic equilibrium between ice and steam. On the graph (Fig. 5.1.) the sublimation curve passes below the evaporation curve. This is due to the fact that the saturation elasticity over ice is less than over water at the same temperatures. Of course, the cohesive force between the molecules of a solid body or medium is greater than in a liquid. As a result, more evaporates from the water surface than from the ice surface at the same temperature. Ultimately, there are more water vapor molecules above supercooled water than over ice. For this reason, the saturation elasticity over supercooled water ( $E$ ) has higher values than over the ice surface ( $E_n$ ) at the same temperature. The saturation elasticity difference ( $\Delta E = E - E_n$ ) has the greatest value at the temperature of about  $-12^\circ\text{C}$  and amounts to 0.269 hPa. A small value of  $\Delta E$  plays a very important role in the processes of the atmosphere.

Let there be both drops of supercooled water and ice crystals nearby in the cloud, and the vapor pressure in the air above the drop and the crystal is  $e$ . Then, at the same humidity, the air can be supersaturated for a crystal, while for a drop it will be unsaturated. Under these conditions, evaporation will occur from the surface of the drop, and sublimation will occur on the surface of the ice crystal. Thus, the process of “distillation” of water from a drop to a crystal takes place. It plays an

important role in the enlargement of cloud particles (crystals due to droplets) and in the formation of precipitation. This process reaches its highest intensity at temperatures close to  $-12^{\circ}\text{C}$ . At such temperatures, under the conditions of a mixed cloud, the drops acquire the smallest dimensions, and the crystals become the largest.

There is another curve on the graph of saturation elasticity ( $E$ ) versus temperature ( $T$ ) (Fig. 5.1). It is called the melting curve or the curve of dynamic equilibrium of two phases — solid (ice) and liquid (water). This curve is almost parallel to the  $y$ -axis, but somewhat inclined towards it, which is important under atmospheric conditions. The water vapor pressure  $e = E$  depends very weakly on the melting point. When the saturation elasticity changes from 6.11 hPa to 1013.25 hPa, the melting point decreases from  $0.01^{\circ}\text{C}$  to  $0^{\circ}\text{C}$ .

So, according to the curves of evaporation and sublimation, the saturation elasticity  $E$  increases with increasing temperature. According to the melting curve however, the elasticity of saturation decreases with increasing temperature. This character of the dependence of  $E$  on  $T$  follows from the analysis of the Clausius–Clapeyron equation (5.4), i. e.

$$\frac{dE}{dT} = \frac{L_{1,2}}{T(v_2 - v_1)}.$$

Let us consider several phase transitions.

1. During the phase transition water–steam, the system absorbs heat, that is,  $L_{1,2} > 0$  and  $(v_2 - v_1) > 0$ , as  $(v_2 \gg v_1)$ , therefore  $\frac{dE}{dT} > 0$ . This means that a positive increase in temperature on the evaporation curve corresponds to a positive increase in saturation elasticity.

2. During the phase transition steam–water, when  $L_{1,2} < 0$ , the system releases heat, that is, it loses it, and  $(v_2 - v_1) < 0$ , since  $(v_2 \ll v_1)$ ,  $\frac{dE}{dT} > 0$ . In this case, again, a positive increase in temperature on the



evaporation curve corresponds to a positive increase in saturation elasticity.

3. During the phase transition water–ice  $L_{1,2} < 0$ , since the system releases heat, and  $(v_2 - v_1) > 0$ ; since  $(v_2 > v_1)$ ,  $\frac{dE}{dT} < 0$ , that is, a positive increment in temperature on the melting curve corresponds to a negative increment in saturation elasticity.

4. At the phase transition ice–water  $L_{1,2} > 0$  (the system absorbs heat when ice melts), and  $(v_2 - v_1) < 0$ , because  $(v_2 < v_1)$ , then  $\frac{dE}{dT} < 0$ , that is, a positive increment in temperature corresponds to a negative increment in saturation elasticity on the melting curve.

In addition to temperature and state of aggregation, other factors also affect saturation elasticity.

The saturation elasticity also depends on: the shape of the evaporating surface, the presence, quantity and chemical composition of condensation nuclei, and the electric charge on the surface of the drop.

The saturation elasticity to a certain extent depends on the shape of the evaporating surface. Indeed, if we consider the escape of a water vapor molecule from a convex, flat, and concave surface, then more liquid particles will interact with it from a concave evaporating surface and less from a convex surface compared to a flat one. Evaporation from the concave surface is difficult. The more water vapor molecules are above the surface, the more pressure they exert. As a result, at the moment of reaching saturation with water vapor, the elasticity of saturation over a convex surface will be greater than over a flat one, and even more so over a concave evaporating surface.

There are droplets of various sizes in the atmosphere. Very small droplets exhibit the properties of a convex surface more, while for large droplets these properties are weakened. Separate areas of the surface of large drops can be considered as flat surfaces, that is, a large drop is

closer in properties to a flat evaporating surface. The largest drops under conditions of turbulent atmosphere and inevitable deformation have areas of both flat and concave surfaces.

Let the drop have a radius  $r$ , and the saturation elasticity above it is  $E_r$ . Then, according to the Thomson formula, the ratio of the values of saturation elasticity above a drop of radius  $r$  and above the surface of flat water has the form:

$$\ln \frac{E_r}{E} = \frac{2\sigma}{R_{\text{II}} \rho_{\text{K}} T r}, \quad (5.6)$$

where  $\sigma$  is the coefficient of surface tension at the boundary of two media (water–water vapor),  $\rho_{\text{K}}$  is the density of the drop, that is, the density of water,  $R_{\text{II}}$  is the specific gas constant of water vapor. In (5.6)  $E_r > E$ . Calculations according to formula (5.6) lead to the conclusion that when the drop radius is more than  $10^{-6}$  m, the ratio  $E_r/E$  tends to 1. Thus, the curvature in the case of a large drop has no effect on the saturation elasticity.

In stable clouds and fogs, the radius of the predominant number of drops is greater than  $10^{-6}$  m = 1  $\mu$ m. For this reason, such drops are similar in properties to a flat evaporating surface. This indicates that the relative humidity over large drops is close in value to the relative humidity over a flat surface. For a saturation state, this value is approximately 100%. So, in clouds and fogs, relative humidity is close to 100%. For the condensation of water vapor on small particles (with a radius of about  $10^{-9}$  m), a large supersaturation with water vapor is required. In this case, the  $E_r/E$  ratio as a percentage is more than 300% [1, 12]. As the particles grow, the supersaturation decreases and reaches 100% with radii of the order of  $10^{-6}$  m.

The effect of condensation nuclei on saturation elasticity is multifaceted. It is affected by the very presence of the nuclei, their number, size, chemical composition and other characteristics. In perfectly clean air in the real atmosphere, water vapor condensation cannot occur; it

needs too much supersaturation in perfectly clean air. In this situation, the nuclei of a new, liquid, phase are complexes of gas particles that have formed due to random convergence of water vapor molecules during their thermal motion. There should be many such complexes of water vapor molecules. When comparing their number with the number of drops in the clouds and mists of the earth's atmosphere, a 12-fold supersaturation of the air with water vapor is required. This number is reduced by charged complexes of molecules. However, in this case too, supersaturation which is impossible in the real atmosphere is needed (6-fold in the case of positive ions and 4-fold in the case of negative ions). The presence of nuclei is a necessary condition for condensation processes and even more so for the formation of clouds and fogs.

Condensation of water vapor in the atmosphere is a fairly frequent phenomenon, since it always contains nuclei. These nuclei are different: some of them are active, others are inactive. There are also those who do not take part in the condensation process at all. The condensation process depends on the size of the nuclei. With their radii of the order of  $10^{-7}$ – $10^{-5}$ m, condensation is possible already at a relative humidity of about 100% or even at lower values. At the same time, there are always much smaller nuclei in the atmosphere, which in most cases are not active (during condensation on them, supersaturation of the air with water vapor is required).

The size of the nuclei in the earth's atmosphere varies within a fairly wide range — from  $10^{-9}$  m to  $10^{-5}$  m. According to the size of the kernel, they are conditionally divided into three groups:

1. Aitken nuclei with a radius from  $5 \cdot 10^{-9}$  to  $2 \cdot 10^{-7}$  m;
2. large nuclei with a radius of  $2 \cdot 10^{-7}$  to  $10^{-6}$  m;
3. giant nuclei with a radius of more than  $10^{-6}$  m.

Of course, the nuclei of the second group are usually involved in the process of condensation [13]. A very important principle of nuclear fission is the separation of soluble and insoluble nuclei among them. This largely determines the degree of their activity in the process of

condensation. Soluble nuclei are primarily products of combustion and nuclei of marine origin. Combustion products are continental; they are formed as a result of industrial combustion of fuel, in the process of forest and peat fires, and in some other cases. Nuclei of marine origin are particles of sea salt or small drops of sea salt solution. They consist mainly of sodium chloride and magnesium chloride ( $\text{NaCl}$  — 77.8%,  $\text{MgCl}_2$  — 10.9%).

Insoluble nuclei, for example, soil particles, can take a very limited part in the condensation process, but only if they are large enough. There are many insoluble nuclei in the atmosphere. They, in addition to soil particles, can be particles of rocks, organic substances, microorganisms and other particles. The number of nuclei over the continents varies on average from a few hundred per cubic centimeter (in mountainous regions) to several hundred thousand per cubic centimeter (in large cities and industrial centers). The number of active nuclei is only 102–103 per cubic centimeter. Thus, most nuclei over the continents are not active and do not take part in condensation processes.

The chemical composition of the nuclei has a great influence on the elasticity of saturation and on all characteristics of moisture in the state of saturation. Condensation on the core may form a drop of salt or acid solution. Let us consider the effect of saturation of nuclei of marine origin on the elasticity.  $\text{NaCl}$  is the main component of sea salt. Let us consider the dynamics of a drop of sodium chloride from its formation to its participation in the composition of a cloudy aerosol medium.

The saturation elasticity over a salt solution is less than over pure water as during evaporation from the surface of the solution, the forces of chemical interaction make the process difficult. As a result, at the same temperature, more particles evaporate from the surface of pure water than from the surface of the solution. This will determine the difference in the values of saturation elasticities over these media.

It is interesting to consider the change in the values of relative air humidity during the growth of a drop that is a solution of salt ( $\text{NaCl}$ ).

Condensation of water vapor on various nuclei begins at a certain value of relative humidity. It depends on the nature of the nucleus. In the case of NaCl nuclei, this value is 78%. So, in the case under consideration, the relative humidity has not yet reached 100%, and condensation has already begun. As the drop grows, the saturation elasticity in different directions is affected by two factors:

a) with a decrease in the concentration of the solution, the elasticity of saturation increases, since the forces of chemical interaction that impede evaporation from the surface of the salt solution are weakened;

b) even a small increase in the droplet radius leads to a weakening of the properties of the convex evaporating surface, which reduces the saturation elasticity.

These two factors affect the saturation elasticity differently at different stages in the process of droplet growth. At the very beginning of the condensation process, the droplet size increases insignificantly. However, with any increase in the drop, the salt solution in it is diluted. This leads to an increase in saturation elasticity. To increase the size of a drop, it is necessary that with its growth the relative humidity of the air around it also increases. Let the relative humidity already reach 100%. Then a further increase in the drop is possible only when the air is supersaturated with water vapor. In this case, the relative humidity should become somewhat higher than 100% (in the case of NaCl nuclei, supersaturation does not exceed 0.4–0.5%, which is possible under atmospheric conditions). Finally, there comes a point when the droplet solution is already highly diluted and does not affect the saturation elasticity. This moment is called the critical point.

Up to the critical point, the difference between the elasticity of water vapor ( $e$ ) and the saturation elasticity  $E_{r,m}$  above the droplet surface decreases (the indices indicate the dependence of the saturation elasticity on the shape of the evaporating surface with the radius  $r$  and on the mass of salt in the solution droplet  $m$ ). This is an unfavorable condition for droplet growth, so up to the critical point, the droplet grows slowly.

The growth time of a drop from its formation to the critical point is called the pre-cloud stage.

After the critical point, the droplet growth conditions change significantly. The solution of the drop by this moment is already significantly diluted, so its growth only leads to a decrease in the saturation elasticity above it. Under these conditions, the drop approaches more and more the properties of a flat evaporating surface. Over such a surface, the saturation elasticity value is less than its value over a convex surface. As a result, the difference  $e - E_{r,m}$  (more precisely,  $e - E_r$ , since the substance dissolved in the drop practically no longer affects the saturation elasticity) increases. This condition is favorable for droplet growth. There are enough water vapor molecules around the drop. They can take part in its growth. As a result, after the critical point, the rate of condensation increases. The droplet grows rapidly under such conditions, and the relative humidity approaches 100%. This indicates that the drop is already large enough. Its surface can be considered as a flat surface of pure water. The relative humidity above it is 100%.

The influence of electric charges on a drop can be traced only in the case of very small drops with a radius of  $10^{-8}$ – $10^{-9}$  m. In this case, charges of any sign lead to a decrease in saturation elasticity. Under such conditions, electrical forces of interaction manifest themselves. They hinder the evaporation of charged droplets from the surface.

In the atmosphere, the water can freeze. The achievement of the solid phase of water (ice) can occur in the process of two completely different phase transitions. These transitions are vapor-ice, bypassing the liquid phase (sublimation) and water-ice (water freezing or crystallization). The transition from the liquid state to the crystalline state can also occur in two ways. It can occur during homogeneous or heterogeneous phase transitions.

These two types of phase transitions from water to ice differ in the nature of the nucleus. The formation of a new (ice) phase begins with the formation of the nucleus. In the case of a homogeneous phase tran-

sition, a new phase nucleus is formed inside the drop due to random motions of the liquid molecules. In this case, accumulations with the density of ice can accidentally form. The lower the temperature is, the more likely such a random event is.

Further, the process depends on the stability of the nuclei formed. Stable nuclei should be energetically favorable. Among others, large nuclei are the most stable ones because energy expended on the formation of a new phase nucleus is proportional to the surface of the nucleus. When an ice phase nucleus appears, energy is released, but it is proportional to the volume of the nucleus. If the expended energy is greater than the released energy, then the resulting nucleus is energetically unfavorable — it is unstable and falls apart. With an increase in the nucleus, its volume increases faster than its surface. As a result, the probability of preserving a larger nucleus compared to a smaller one increases.

The stability of the nucleus of the ice phase is also affected by the air temperature. In colder air around the drop, less energy is spent on the formation of a new phase nucleus. Such nuclei are more likely to be stable. Thus, the process of transition of water into ice is observed at lower temperatures.

Experimental data confirm the above theoretical reasoning. Thus, small supercooled droplets ( $r < 5$  microns) can stay liquid down to very low temperatures (to  $-40$  °C). If the temperatures are higher than this value, then only individual crystals are found among many small droplets. If the temperature has values below  $-41$  °C and is decreasing, then there is a rapid increase in the number of crystals. The freezing point of droplets is largely determined by their size. The smallest drops do not freeze to lower temperatures. Inside a large drop, an accidental formation of ice density in some part of it due to the movement of molecules is a more likely event.

For a heterogeneous transition from the liquid phase to the crystalline phase, crystallization nuclei are required. They will serve as the

beginning of a new (solid) phase. If there are crystallization nuclei in the atmosphere, the freezing of droplets can begin already at relatively high temperatures. Such temperatures can be only slightly below 0 °C.

It is believed that the main type of phase transition (water–ice) is a homogeneous transition, since stable supercooled clouds are often observed at temperatures in the range of  $-12 \div -15$  °C. In a heterogeneous transition, such clouds at given temperatures would already become crystalline. It is possible that there are not so many crystallization nuclei in the atmosphere, with the help of which a heterogeneous phase transition could occur. In any case, when certain nuclei are introduced into a supercooled cloud, crystallization is sharply accelerated. This has long been used in artificial impacts on clouds. So, for example, when silver iodide is introduced into a liquid-drop cloud, crystallization can begin already at a temperature of  $-4$  °C.

Crystallization nuclei can be different. Some of them are mixed nuclei. In such cases, they are condensation nuclei, and at low temperatures they can act as crystallization nuclei. So, for example, when the temperature drops, salt particles fall out of the solution. In the future, they can become crystallization nuclei.

***Test questions:***

1. What can be judged by the value of the difference between the values of the would-be evaporation from land with a high content of moisture and actual evaporation from water surface?
2. On what factors does evaporation from the underlying surface depend?
3. How do humidity values change with height?
4. What are the features of the daily variation of moisture characteristics over different types of underlying surface in different seasons of the year?
5. What is the dependence of saturation elasticity on air temperature?
6. What is the form of the Clausius–Clapeyron equation and the Magnus formula?
7. What does a graphic representation of the dependence of saturation elasticity on air temperature look like?



8. What is the dependence of saturation elasticity on the state of aggregation of the evaporating surface?
9. What is the dependence of saturation elasticity on the shape (curvature) of the evaporating surface?
10. What are the consequences of this dependence in the real atmosphere?
11. What is the dependence of saturation elasticity on the properties of condensation nuclei?
12. In what cases does the presence of a charge on a drop affect the saturation elasticity?
13. How does water freeze in the atmosphere?
14. Which type of water-ice phase transition is the main one in the atmosphere?

## 6. FOGS AND CLOUDS

Clouds and fogs are water aerosol media. The definitions of these atmospheric phenomena are similar.

Fog is a visible collection of water droplets or ice crystals, or both at the same time, directly at the underlying surface, when the visibility range deteriorates to less than 1 km. When defining a cloud, the second part of the above definition sounds different. First, the indicated visible collection in the case of clouds is located above the underlying surface. Second, the definition of clouds does not specify a reduction in visibility value.

As a result, the definition of clouds is as follows: clouds are a visible collection of water droplets or ice crystals, or both at a certain height above the underlying surface.

The deterioration of visibility in the definition of clouds is not specified, not because clouds do not reduce visibility. The fact is that clouds differ from each other more than fogs differ from each other. This also applies to reduced visibility. In addition, fogs directly affect various aspects of people's lives, in particular, the operation of vehicles, water transport, the takeoff and landing of aircraft. In all these cases, indication of visibility directly at the underlying surface is necessary. The main difference between clouds and fogs is that fogs form in a homogeneous, therefore, equilibrium atmosphere. Clouds form in a variety of ways. So some of them are formed in a homogeneous stable atmosphere, other clouds are associated with extremely non-equilibrium systems with high air instability.

The water aerosol media in the case of clouds and fogs are visible because all aerosol media scatter rays of different wavelengths almost

equally. In the visible part of the spectrum, the addition of different colors leads to a white or whitish color of clouds and fogs.

## 6.1. Fogs

Fogs can form under the influence of one or more major factors. In addition, some secondary factors are usually involved in the formation of fogs.

The main factors in the formation of fogs can be water evaporation, mixing of heterogeneous air masses, and air cooling.

Other fogging factors may be secondary to those factors that play a major role in the formation of this atmospheric phenomenon.

An atmospheric phenomenon close to fog is mist. Fog and mist vary in intensity. Thus, mists are characterized by a deterioration in visibility to values of less than 10 km. Fogs reduce visibility to values less than 1 km. Within fogs and mists, there is also a difference in the visibility range. This determines the intensity of fogs and mists (Table 6.1).

Table 6.1

**Visibility range in fogs and mists of varying intensity**

The intensity of fogs and mists	Visibility range
Heavy fog	< 50 m
Moderate fog	50–500 m
Weak fog	500–1000 m
Heavy mist	1–2 km
Moderate mist	2–4 km
Weak mist	4–10 km

Let us consider the formation of fogs under the influence of the main factors.

### 6.1.1. Evaporation of water as a factor in the formation of fog

Relative humidity ( $f = \frac{e}{E}100\%$ ) can increase and reach 100% in

two ways: by increasing the amount of water vapor or by lowering the temperature. In the first case, the partial pressure ( $e$ ) of water vapor increases, that is, the numerator in the relative humidity formula increases. In the second case, the saturation elasticity ( $E$ ) decreases, that is, the denominator in the relative humidity formula decreases.

If the fog is formed under the influence of evaporation as the main factor, then the fog is called evaporation fog.

The mechanism of evaporation fog formation is associated with the evaporation of a warmer underlying surface into colder air. The point is that the underlying surface evaporates until the value of saturation elasticity at the temperature of this surface is reached ( $E_j$ ). However, the saturation elasticity value of the air above the underlying surface ( $E$ ) in cold air is reached before the value  $E_j$ . Recall that the elasticity of saturation is the greater, the higher the temperature is. In other words, the air has already reached saturation at its lower temperature. At the same time, the warmer underlying surface continues to evaporate into colder air. However, air cannot contain more water vapor than the amount that creates the saturation elasticity value at a given temperature.

What should the air do with the excess water vapor supplied by the evaporating surface when this air has already reached the state of saturation with water vapor? Air will transfer this “surplus” of water vapor to another phase state of water. Drops of water or ice crystals will appear in the air. In cases where their number is sufficient, evaporation fog may form.

If the air temperature is higher than the temperature of the evaporating surface, then evaporation stops before air saturation ( $E$ ) is reached. In this case, reaching the state of saturation with air is impossible.

When can vapor fog form? Let us name the favorable conditions for the formation of such a fog.

1. An elevated inversion at a height of 60–100 m. This is due to the fact that the evaporation of water is not a powerful fogging factor. With a high elevated inversion, water vapor is propagated into large volumes of air. The vapor concentration may not be sufficient to form fog. In cases of fog formation, the penetration of water vapor into higher layers must be limited. This function can be performed by elevated inversion. It is desirable that the lower limit of this inversion is not very high. In any case, during inversion at altitudes of 200–300 m, evaporation fog is not formed.

2. Sufficiently large temperature difference between warmer water and colder air above it.

3. Sufficiently high relative humidity of the air before it reaches the surface of warmer water.

4. Average speeds of air movement over the water surface. At high wind speeds, turbulent mixing increases, which increases the elevated inversion. This does not favor the formation of fog. In addition, at high wind speeds, the water evaporating surface does not have time to saturate the air with water vapor, since more and more air volumes approach the water surface. On the other hand, very low speeds are not a favorable condition for the formation of evaporation fog either. This is due to the need to move the air to the surface of open water.

These favorable conditions for the formation of evaporation fog often develop at high latitudes, in particular, over the waters of the Arctic Ocean and its coasts. Under these physical and geographical conditions, air can first move over fields of snow and ice, and then move to the surface of open water. This water is warmer than the fields of ice and snow.

Evaporation of water may not be the main factor in fogging, but a factor that enhances the effect of the main factor. For example, before evaporating into the air from the surface of the water, this air has

cooled strongly over the surface of the land. Due to cooling, the air has largely approached a state of saturation. In this case, the main factor in the formation of fog is the radiation factor, and evaporation is an important additional factor.

### **6.1.2. *Mixing of air masses with different thermohygro-metric properties as a factor in fog formation***

When masses of air with different values of temperature and humidity mix together, fog can form in the resulting mixture.

Let warmer and colder air masses mix. The temperature of the first one goes down. As a result, the warmer mass approaches saturation. If this mass of air has reached the state of saturation and continues to cool, then excess moisture will lead to the need for a phase transition.

Let the process of condensation take place, i. e. drops be formed. In the process of mixing, the drops will spread to the entire volume of the mixing masses.

At the same time, the colder mass of air is heated by mixing with the warmer mass. This will cause the colder mass to move away from saturation. For this reason, some of the droplets formed will certainly evaporate. The rest of the droplets, under favorable conditions, can form a fog. Fogs formed by mixing are not the same as fogs formed by cooling, since one of the mixing air masses is heated in this process.

Let us determine the favorable conditions for the formation of fog when mixing of two air masses is the main factor.

1. High values of relative humidity of air masses before their mixing. If they are close to 100%, then fog is formed.

2. Sufficiently high temperature difference ( $T_2 - T_1$ ) of the mixing masses.

3. Sufficiently high values of the temperatures  $T_1$  and  $T_2$  of the mixing masses themselves. At sufficiently high temperatures, a decrease in the temperature of the warm mass will lead to a larger drop in  $E_2$  than an increase in the saturation elasticity  $E_1$  associated with an increase in

the temperature of the cold mass. The possibility of fog formation in this case depends on the difference ( $\Delta E_2 - \Delta E_1$ ).

If the mixing of air masses occurs at low temperatures in each of the mixing air masses, then the decrease in saturation elasticity when warm air is cooled will be approximately equal to the increase in saturation elasticity when cold air is heated. In this case, the excess moisture when mixing air masses is too small to form fog.

Sometimes researchers do not distinguish mixing fogs as an independent type of fog. However, such fogs form in areas where water is adjacent to land. So these fogs are observed near the coastline, partly above the water, partly above the coast. Sometimes they do not have a significant horizontal extent in the direction from land to water. It should be noted that the mixing of air masses leads to the formation of fog only when their relative humidity is not lower than 95%, and the temperature difference is not less than 10 °C. For this reason, mixing fogs most often do not have high repeatability and significant intensity.

In cases of frontal fogs, not just air masses are mixed, but two air masses, which is not the same thing in physical meteorology. In the formation of frontal fogs, mixing is one of the main fogging factors. In the formation of some other types of fogs, the mixing of air masses plays an important auxiliary role and enhances the main factor. Studies have shown that the factor under consideration plays an important role not only in the formation of fogs, but also clouds, primarily in frontal zones.

### **6.1.3. Cooling of the air as a factor in the formation of fog**

One of the main reasons of the formation of fogs is a decrease in air temperature, which is associated with the most intense fogs. Depending on the air cooling mechanism, radiation, advection, advection-radiation and ascension fogs are distinguished.

*Radiation fogs* are caused by the cooling of the Earth's surface by means of radiation that escapes into world space. Such fogs are formed

under conditions of weak counterradiation of the atmosphere. The air, especially adjacent to the cooling Earth's surface, gives it heat in the process of turbulent heat transfer. The air is also cooled in this way. As a result of the cooling of the lower layer of the air, excess water vapor appears in it, which passes into a different phase state. If drops of water or ice crystals are sufficient to reduce the visibility range to less than one kilometer, then fog is formed.

Favorable conditions for the formation of radiation fogs are:

1. The absence of clouds or the presence of only high-level clouds (only in the upper tier); sometimes it is enough to disperse the clouds or at least raise the level of their lower boundary. In any case, conditions must be created in the atmosphere to reduce the counter radiation of the atmosphere and, consequently, to increase the effective radiation of the earth's surface.

2. High relative humidity before cooling.

3. Weak wind or calm in the lower layer, so that the air is above the radiating earth's surface for a sufficiently long time.

It is clear that if any condition for the development of fogging processes is not fully met, then another favorable condition should compensate for it. For example, there are high- and mid-level clouds present in the atmosphere. This cloud cover erodes very slowly. Then the relative humidity should be so high that it does not require much cooling of the air to form fog.

Radiation fogs are formed when the radiation balance of the Earth's surface is negative, that is, in the cold season or at other times of the year, but only at night.

*Advection fogs* are formed during advection, that is, during horizontal movement of a warm mass of air to a colder underlying surface. Advection fogs occupy large areas, since the conditions for air movement are created by the features of the atmospheric pressure field, which cannot have a small-scale manifestation. In addition, advection fogs are characterized by high intensity and relatively high power. This is due to



the way the air is cooled at an average wind speed near the underlying surface (the wind speed noticeably increases with height). In such fogs, turbulent mixing is more developed than in other fogs. As a result, condensation and sublimation products are transported to higher levels than with other fogs.

The underlying surface over which advection fog is formed may be both land and water. Over land, such fog is more often observed in autumn and especially in winter. At this time, the land becomes very cold. Over water, advection fog is usually observed in spring and especially in summer, when the water surface is colder than land.

However, there are many situations in which advection fog is formed. In addition to the more well-known variants of the transition of air from the surface of the water to the surface of cooled land and the less well-known variants of the transition of air over a warmer continent to the surface of colder water, quite different cases are often observed. For example, advection fog can form when air moves from lower latitudes to higher latitudes (both over land and over water) and when air moves from a warm ocean current to a cold one. In such cases, the land may not participate in the formation of fog at all. In addition, advection fog can form even when air moves from wetlands to places with much less soil moisture.

In the latter case, fog formation is associated with differences in soil properties. Waterlogged soils are characterized by high evaporation and heat loss for this process. Waterlogged soils do not warm up enough. This is due to the high values of heat capacity and thermal conductivity of such soils. On the other hand, dry soils heat up quickly but also cool down quickly.

Variants are possible when dry soils are colder than waterlogged soils. In such cases, the air moves from waterlogged soils to dry soils and cools down. When is it possible? Let the air above the waterlogged soil be close to the state of saturation with water vapor. Then there will be no intense evaporation. The heat consumption for evaporation will not be large either. However, at night or during transitional seasons, dry

soil cools down quite quickly. Waterlogged soil cools down slowly. As a result, dry soil may be colder than waterlogged soil for a period of time. Humid air does not need long and strong cooling over dry soil. In this case, the formation of fog is possible. This variant of the formation of advection fog is less common than others.

There are more favorable conditions for the formation of advection fog than for other cooling fogs. This is due to the fact that advection fogs are very powerful and occupy fairly large areas. For the formation of such fogs, special conditions must develop.

So, favorable conditions for the formation of advection fogs are:

1. High relative humidity of moving warmer air even before it enters the cold underlying surface (this condition is especially important when fogs form over land).

2. Large temperature difference between the underlying surface and the shifting air mass.

3. Average wind speeds near the underlying surface (about 2–5 m/sec); at high speeds, a strong turbulent exchange develops, which prevents the formation of fog, spreading the products of condensation and sublimation to large volumes of air; at very weak winds, the main factor of fog formation (horizontal displacement) is less pronounced; then the slowly moving air cools down from the underlying surface too slowly, which also does not favor the formation of advection fog.

4. Increase or at least the same specific humidity with height; this condition follows from the fact that in an advection fog, that is, a sufficiently powerful fog, there occurs turbulent mixing; in the course of such a process, the specific humidity decreases with height, and the amount of water vapor at the underlying surface also decreases. Indeed, during turbulent mixing, an equalization of all the properties of the atmosphere is observed. In this case, there is an outflow of water vapor from the air near the underlying surface to higher layers. If the specific humidity increases with height, then in the process of turbulent mixing, an influx of water vapor to the underlying surface from higher layers is

observed. This condition refers to the formation of advection fogs over land. Above the water, the outflow of water vapor is easily compensated by evaporation from the water surface.

5. Moderately stable stratification and relatively weak turbulent exchange. With a very stable stratification, the turbulent exchange is greatly weakened, so the cooling of the air over the underlying surface occurs slowly. In this case, fog can form only in a very narrow layer above the underlying surface.

*Advection-radiation fogs* are formed under the influence of two factors — advection and radiative cooling. Each of these factors is the main one, and it is impossible to determine the predominant influence of one of them. In other words, first there is a horizontal movement of the air to a colder underlying surface, then the air is cooled above the radiant underlying surface. The most common form of such fog is the movement of air from the underlying water surface with a higher temperature at night to colder land. Subsequently, the land and the air above it continue to cool down by radiation.

Favorable conditions for the formation of such a fog are made up of favorable conditions for the formation of advection and radiation fogs.

*Ascension fogs*, which are otherwise called upslope fogs, are formed when air rises along the slopes of any hills and mountains. The reason for the formation of such fogs is adiabatic cooling, approaching the saturation state of the air, reaching this state, and then water vapor condensing. It is known that during the upward movement, the air expands, uses energy on the work of expansion and, as a result, cools down.

Favorable conditions for the formation of ascension fogs are:

1. The stratification of the air rising along the slope must be stable. Otherwise, cumulus clouds form rather than fog;
2. The relative humidity of the air before it rises must be sufficiently high;
3. The height of the slope must be above the level of condensation.

In the case of ascension fogs, it is difficult to give any average parameters. The power of the fog and its length along the slope depend on the height of the obstacle, on the steepness of the slope, on the vertical air temperature gradient, on the speed of the oncoming flow, and, in addition, on those favorable conditions on which cooling fogs depend (on the relative humidity of the rising air, and on the difference between the temperature of the slope surface and the air).

#### **6.1.4. Frontal fogs**

Usually, fogs are intra-mass, but sometimes fogs are associated with atmospheric fronts, that is, they are observed when air masses change. These fogs are of two types:

- a) frontal fog,
- b) fog behind a warm front.

Prefrontal fog occurs ahead of a warm or occluded front. In essence, this fog is a type of evaporation fog, but the falling and fallen frontal precipitation evaporates. Because of evaporation from the surfaces of falling drops and soaked soil, the air reaches the state of saturation. As a result, fog may form. With this fog, precipitation falls from prefrontal cloudiness, so fog is observed as a continuous band in front of the front (together with rain). The formation of this fog is also facilitated by the pressure drop ahead of the front, which causes the adiabatic expansion of the air and its adiabatic cooling.

Frontal fog can be observed behind a warm front. In this case, fog is formed by three main factors at once: cooling (advection of warm air behind the front onto a colder underlying surface), mixing of air masses in the frontal zone, and evaporation of falling and fallen frontal precipitation.

The characteristic features of this fog formation, except that there are three factors (and all of them are major), are the mixing factor and the evaporation factor. So, not just two masses of air are mixed, but whole air masses. This increases the scale of the process. The evapora-

tion factor is also special: fallen and falling precipitation evaporates not into cold air, as happens during the formation of evaporation fogs, but into warm air. So, this factor alone will not lead to the formation of fog. But all three factors lead to the formation of frontal fogs that are the most powerful ones among fogs of different types and spread significantly along the front line.

All the types of fog described above can be enhanced by the influence of the city, in which numerous industrial enterprises and heating systems operate, and vehicles are concentrated. As a result, various active condensation nuclei enter the atmosphere of the city, and in large quantities. Such fogs can persist for quite a long time.

Smogs occupy a special place among fogs. They are fogs containing products of incomplete combustion or chemical production waste. London smogs are well known; they are observed for a long time and are harmful to human health.

### **6.1.5. Physical characteristics of fogs**

This section will show common features of fogs of various types, as well as differences between them in terms of the values of physical characteristics.

1. The water content of fogs varies over a wide range — from fractions to  $1.5\text{--}2.0\text{ g/m}^3$ . Any, even a small change in the values of water content significantly affects visibility in the fog. The greater its water content, the higher the intensity. In other words, the water content of a strong fog is greater than that of a moderate or a weak one.

2. The water content profile in fogs is closely related to the temperature profile. The beginning of fog development falls on the time of surface inversion. As a result, the water content has the highest values in the layer of lower temperatures, that is, near the underlying surface. Over time, the inversion becomes elevated. In the layer below the lower boundary of the inversion, the temperature decreases with height. The level of the highest water content rises with height along with the in-

crease in the level of the lowest temperatures. In the last stage of fog development, its upper boundary most often coincides with the lower boundary of an elevated inversion. This is where the highest water content is observed.

3. The distribution of temperature with height in fogs depends on the stage of their development. In cases of cooling fogs, the ground inversion is established first. As the fog develops, the surface inversion collapses in its lower part and turns into an elevated inversion. This rearrangement of the air temperature profile occurs gradually. At first, the lowest temperature level is the underlying surface and the layer of air adjacent to it.

In fogs of different types, the cooling of the underlying surface occurs in its own way. In cases of radiation fogs, the underlying surface is cooled due to its own radiation. The lower layer of air gives it its heat and also cools down. During advection fogs, warmer air enters the cold underlying surface and also gives it its heat. In both cases, cooling occurs most in the lower part of the air layer. It is this part, that due to its close position to the underlying surface and the greater air density, directs the largest heat flux down. As a result, the lower air layer becomes the coldest one. In other words, a surface inversion has formed (an increase in temperature with height).

As for the evaporation fogs, the open water surface meets the air that already contains surface inversion.

Further, the situation for any fog develops as follows. In cooling fogs, the surface inversion is initially preserved over time. At the same time, two heat fluxes are directed to the underlying surface. One of them is the molecular heat flux from deeper soil layers. These layers are warmer than the cooling underlying surface. Another heat flow is the turbulent flow from warmer layers of air. It is clear that the turbulent heat flux from the atmosphere significantly exceeds the molecular flux from the soil. However, in sum, these two heat fluxes do not compensate for the heat loss of the underlying surface. It continues to chill.

The situation begins to change when the thickness of the developing fog reaches approximately 200–300 m. In this case, the counterradiation of the atmosphere increases so much that the heat fluxes arriving at the underlying surface sometimes not only compensate for its heat losses, but even exceed them. Either way, the ground inversion starts to break down at the bottom of the fog.

The remainder of the former surface inversion is retained only from a certain level. Below this level, the temperature decreases with height. This is due to the fact that the air directly above the underlying surface is slightly heated. For heating, it uses the heat received by the underlying surface from two media. One of them is the soil; another medium is fog that has increased its power.

The preserved part of the inversion is no longer surface inversion and is called an elevated inversion. In the case of evaporation fogs, the surface inversion will begin to break down as very cold air from the fields of snow and ice enters the surface of the water. Already in the process of fog formation, cold air above the warmer surface of open water will somewhat heat up. As a result, the lowest temperature level will rise. If this increase is not large, evaporation fog may form.

An elevated inversion has also been found in the mixing fogs. Thus, regardless of the type of fog, the temperature profile has two important parts: a layer of temperature drop with height, and a layer of temperature increase with height. They are separated by the level of the lowest temperature, which is the lower boundary of the elevated inversion.

The difference in the time of occurrence of this inversion in different fogs is that in the cases of evaporation fogs, the raised inversion appears already during the formation of fog, and in the cases of cooling fogs it appears later, when the developing fog increases its power. Sometimes, with a high thermal conductivity of the soil and a relatively developed turbulent exchange in the air, large heat fluxes from the soil and air add up. This leads to a strong increase in the temperature of the underlying surface and the layer of air closest to it. The existing fog in

the specified layer can completely dissipate. The remaining part of it is no longer fog, since it does not come into contact with the underlying surface. This atmospheric phenomenon is a cloudiness of raised fog (*Stratus nebulosus*).

4. Relative humidity in fogs depends on temperature values. At positive and small negative temperatures, relative humidity has values close to 100% (95–100%). At low temperatures ( $-30 \div -40$  °C), the relative humidity in the fog decreases to 80–70%. However, whatever the relative humidity is, it practically does not change with height.

5. The aggregate state of fogs depends on the air temperature and the presence of sufficiently active centers of phase transitions. According to the state of aggregation, fogs are drop-liquid, crystalline and mixed. It depends on the air temperature and the presence of sufficiently active condensation centers. At positive temperatures, fogs are drop-liquid, and at negative temperatures, fog can remain as drop-liquid even down to  $-28$  °C, if the air is clean enough. If there are many active aerosols in the atmosphere, then the fog remains as a drop-liquid only up to the smallest negative values of air temperature (of the order of  $-3 \div -4$  °C). In mixed fogs, the temperature is  $-11 \div -26$  °C, and in crystalline fogs, the air temperature is  $-16 \div -58$  °C.

6. Sizes of particles and their distribution in fogs. Fog as an aerosol medium is polydisperse, that is, it has particles of various sizes — from fractions of a micrometer to tens of micrometers in drop fogs and up to hundreds of micrometers in the case of crystalline fogs. Most droplets in fogs have a radius in the range of 2–18  $\mu\text{m}$ , with a maximum repeatability of droplets with a radius of 12  $\mu\text{m}$ . In crystalline fogs (according to observations of evaporation fogs on the Angara), the sizes of crystals are related to the intensity of fogs and are: from 3  $\mu\text{m}$  to 125  $\mu\text{m}$  in weak fogs, from 9  $\mu\text{m}$  to 355  $\mu\text{m}$  in moderate fogs, and from 9  $\mu\text{m}$  to 475  $\mu\text{m}$  in strong evaporation fogs. Of course, these values vary within fairly wide limits depending on the type of fog, on the place of its formation, and on atmospheric conditions.



In a polydisperse medium, which is fog, the size distribution of droplets is characterized by the distribution curve that always has a clearly defined maximum corresponding to the greatest frequency of drops of a certain radius  $r_m$  (modal radius). From this value, the decrease in repeatability occurs faster in the direction of smaller sizes than in the direction of large ones. This characteristic feature of the frequency distribution of various fog particles can also be traced in the cases of crystalline fogs. Since the frequency of smaller fog particles decreases faster from the value  $r_m$ , the modal radius has lower values than the arithmetic mean radius ( $r_{cp}$ ).

7. The power of the fogs. This characteristic is determined by the height position of the upper boundary of fogs, since their lower boundary always coincides (by definition) with the underlying surface. As several studies have shown, the highest frequency of the location of the upper boundary of fogs falls on the layer of 200–300 m. According to the data of Moscow region, in the cold season this layer accounts for 16 cases out of 45, 12 other cases were in the layer of 300–600 m, and 10 cases in 100–200 m layer [12]. In total, these gradations account for 38 cases out of 45, i.e. 84.4%.

The average fog power was obtained in Ukraine from aircraft sounding data. This thickness is 155 m for radiation fogs, 260 m for advection-radiation fogs, 320 m for advection fogs, and 400 m for frontal fogs. In general, the thickness of fogs increases with increasing wind speed. The lowest wind speeds (practically calm near the underlying surface) are observed with radiation fogs, the highest — with advection and frontal fogs. The connection between the wind speed and the position of the upper boundary of the fogs is understandable, given that in most cases the upper boundary of the fogs coincides with the lower boundary of the elevated inversion. The higher the wind speed is, the more developed turbulent mixing is. In this case, the surface inversion becomes an elevated inversion.

The average thickness of evaporation fogs is about 80 m.

8. Wind speed in fogs. It is known that fogs form in a stable atmosphere in which high wind speeds are not observed. For the formation of each type of fog a certain gradation of speed is characteristic, which most favorably affects the process of fogging. For radiation fogs, calm conditions near the underlying surface or very weak winds are most favorable.

Advection fogs are associated with the movement of warmer air to a colder underlying surface, and the movement occurs at a certain speed, most often not exceeding 6 m/s; in frontal fogs, wind speeds also rarely exceed these values near the underlying surface.

With height, the wind speed increases, therefore, at higher levels, calm in fog is considered a fairly rare occurrence. So, at a level of 100 m, even with radiation fogs, calm almost never occurs. According to Moscow data, in cases of fog at a height of 100 m, the wind speed is most often 1–3 m/sec, in the layer of 200–300 m it is 2–5 m/sec, in the layer of 400–600 m the wind speed is 3–6 m/sec. The type of fog imposes some corrections to the given values.

#### ***6.1.6. Features of the formation and preservation of fogs over a snowy surface***

The underlying snow surface creates unfavorable conditions for the formation of a drop-liquid fog above it. Unfavorable conditions are created by the snow surface even for the preservation of droplet-liquid fog in the air mass if it enters the snow surface.

The fact is that the air reaches the state of saturation over ice earlier than over water ( $E_i < E$ ). For this reason, when  $e = E_i$  is reached, that is, when the value of saturation elasticity relative to the underlying snow surface is reached, sublimation of the water vapor contained in the air will begin on it. This process can be traced both in cases where the fog has not yet formed, and in cases when an air mass comes to the snow surface, which already contains a drop-liquid fog. In the second case, the process of “distillation of water from a drop to a crystal” will

take place, which is considered in Section 5.2. The only exceptions are those cases when the temperature of the air entering the snow surface drops very quickly and strongly. In these cases, the droplet-liquid fog can even intensify, that is, increase the intensity. But even in these cases, the dissipation of fog is only delayed. In the end, the droplet-liquid fog still dissipates over the snow surface.

The snow surface has the greatest scattering effect on droplet-liquid fog at temperatures close to  $-12\text{ }^{\circ}\text{C}$  ( $-8 \div -15\text{ }^{\circ}\text{C}$ ), when the difference in saturation elasticity over water and ice is the largest. If at these temperatures droplet-liquid fog is formed due to the large temperature difference between the air and the snow surface, then such fogs are quite rare. Their intensity is low compared to fogs over other types of underlying surface at the same temperature contrasts (over water, over soil not covered with snow).

The most favorable conditions for the formation and preservation of droplet-liquid fogs over the snow surface are formed at temperatures close to  $0\text{ }^{\circ}\text{C}$ . In these cases, the difference between the values of saturation elasticity over water and ice is minimal. That is why the snow surface is even a strong fogging factor, since it cools the air. Droplet-liquid fogs can form above the snow surface (or remain in the air that has come to the snow surface if it has already formed before that time) at temperatures of  $5\text{ }^{\circ}\text{C} \div -5\text{ }^{\circ}\text{C}$ .

If the air temperature is below  $-15\text{ }^{\circ}\text{C}$ , then the formed fogs are, as a rule, crystalline. The conditions for the formation and preservation of such fogs over the snowy surface are favorable. The fogs are in good balance with the underlying snow surface. Under these conditions, the snow surface is a strong factor in lowering the temperature for the fog. The frequency of occurrence of such fogs at temperatures below  $-15\text{ }^{\circ}\text{C}$  increases.

### **6.1.7. Daily and annual variations in the frequency of fogs**

The main reason for the formation of fogs is the cooling of the air, so the frequency of fogs has a pronounced diurnal variation. Over most of the underlying surfaces, the annual variation can also be traced.

It is clear that the daily course of fogs is more pronounced over land. In the diurnal course, the maximum values of the frequency of fogs fall on the time before sunrise, the minimum values fall on the near-noon hours. This is due to the diurnal variation of the temperature of the underlying surface and the air layer adjacent to it.

Radiation fogs demonstrate the most pronounced diurnal variation. The maximum frequency of formation of radiation fogs is observed 1–2 hours before sunrise, and the maximum frequency of fog dissipation occurs 1.5–2.5 hours after sunrise. The diurnal pattern of the frequency of fogs is more pronounced at higher temperatures, when the saturation elasticity increases faster with increasing temperature. In summer, after sunrise, fogs dissipate quickly and are rarely observed during the daytime. In winter, at low temperatures and their weak growth during the day, fogs dissipate slowly. At this time, the elasticity of water vapor in the fog slowly and weakly moves away from the values of saturation elasticity. At  $e \approx E$ , the air is in a state of saturation with water vapor, so fog can persist for a day and night.

Two main types can be traced in the annual course of fog frequency. The first type is typical for moderate and low latitudes. In this type, the maximum frequency of fogs over land occurs in the autumn-winter period, and the minimum frequency occurs in summer. Over water, fogs are more often observed in spring and especially in summer, when the surface of the water and the air above it have lower temperatures than over land.

The second type of annual variation in the frequency of fogs can be traced on the coast and islands of the Arctic Ocean. The type has two subtypes with a maximum frequency of fogs in summer and with a uniform distribution of frequency throughout the year.

***Test questions:***

1. What is fog?
2. What is the difference between fogs and mists?
3. What are the similarities between fogs and mists?
4. How is the intensity of fog and mist determined?
5. What are the main factors in the formation of fogs?
6. What are the mechanisms of fog formation under the influence of each of the main factors?
  7. What are the favorable conditions for the formation of each type of fog?
  8. What are the values of meteorological quantities in fogs (water content, relative humidity, wind speed, air temperature) and how do these meteorological quantities change with height in fogs?
  9. What are fogs according to their state of aggregation?
  10. What is the power of different types of fogs?
  11. What are the features of the formation and preservation of drop-liquid fogs over the underlying snow surface?
  12. What effect does the underlying snow surface have on crystalline fogs?
  13. Are there periodic changes in the frequency of fogs?
  14. In which types of fogs are periodic changes in frequency more pronounced, and in which are they more weakly or not pronounced?
  15. What are the features of the diurnal variation in the frequency of fogs?
  16. What are the features of the annual variation in the frequency of fogs?

## **6.2. Clouds**

The visible characteristics of clouds are their number, shape, height of the lower boundary of the clouds when it is located within a two-kilometer layer. These characteristics are determined at each meteorological period. This is due to the fact that clouds or their absence uniquely determine the nature of the atmospheric process. Any cloud is a visible collection of water droplets or ice crystals, or both at a certain height above the underlying surface. There are no fundamental differences in microphysical structure between clouds and fogs. There are fundamental differences between clouds and fogs in vertical thickness,

horizontal scales of manifestation, in the strength of upward movements, the degree of turbulent mixing, and in the conditions of formation. Clouds, like fogs, are formed as a result of the increase in moisture content and decrease in the temperature of the air, or under the influence of one of these factors.

There are three valid and officially recognized cloud classifications. The first principle is based on the nature of cloud formation processes, that is, the origin of clouds. This principle of classification is called genetic and is very important. Clouds selected according to this principle make it possible to assess the state of the atmosphere, many atmospheric parameters, and even their change with height. According to the genetic classification, there are cumulus, undulating and stratus clouds.

The second principle is based on the appearance of clouds, including the height of their location. The classification of clouds by appearance is called morphological. This classification is adopted for meteorological observations at hydrometeorological stations. Morphological classification does not provide direct information about the origin of clouds and their microphysical structure. However, when performing qualitative observations, this classification makes it possible to judge the processes occurring in the atmosphere by indirect signs. The convenience of this classification lies in the fact that the observer, not knowing the features of cloud formation processes and the conditions of phase transitions of water in each specific case, has the opportunity to perform high-quality observations of clouds. With the results of these qualitative observations, it is then possible to proceed to the processes of cloud formation.

The third principle of cloud classification is based on the state of aggregation, the size of cloud elements, the number of drops and crystals per unit volume, the shape of crystals, and other microphysical characteristics. The classification is called classification according to the microphysical structure (microstructure) of clouds. This classification makes it possible to divide clouds according to their internal struc-

ture (by geometric structure and physical properties). According to this classification, first of all, there can be distinguished crystalline clouds, clouds of a drop-liquid and mixed structure. Sometimes, among drop-liquid clouds, there are distinguished those that do not give precipitation, clouds with drizzle, and clouds with rain.

### **6.2.1. Cumulus (convective) clouds**

Outwardly, cumulus clouds differ from all other clouds because they look like isolated cloud masses. These clouds develop vertically. Development in this direction occurs to the extent that the main factors of their formation allow them. Such factors are thermal convection and pronounced turbulent exchange.

Cumulus clouds form when the air is highly unstable. Let the instability be observed in a thin layer. Then in this layer clouds of “shallow” convection are formed. They are flat cumulus (Cu hum.) and medium cumulus (Cu med.) clouds. In case of instability in a thick layer, powerful cumulus clouds (Cu cong.) and even cumulonimbus clouds (Cb) can develop. The most striking manifestation of the instability of the atmosphere are the most powerful cumulonimbus clouds and precipitation associated with them (shower precipitation, including hail).

Cumulonimbus clouds are different from other clouds even when their lower parts join each other. Such cases are observed quite often. Inexperienced observers may even confuse cumulonimbus clouds with nimbostratus clouds. However, these clouds differ not only in appearance, but also in the nature of the precipitation falling from them. In addition, which is most important, these clouds differ in the conditions of their formation. The meteorological situation with these clouds also has some differences.

So, let the lower parts of different cumulonimbus clouds join together. What do they have in common with nimbostratus clouds? A solid covering of the firmament, a whole set of gray shades on their bases. Both types of clouds have their base in the lower troposphere (up

to the level of 2 km). Both stratocumulus and cumulonimbus clouds do not allow the sun and moon to be seen. At the same time, cumulonimbus clouds often have darker areas against the general background of a gray field. The most important thing is that in the case of cumulonimbus clouds, much more noticeable inhomogeneities are traced compared to stratus clouds. Nimbostratus clouds have more homogeneous areas in the region of their base. This base is not clearly expressed due to precipitation.

Showers fall from cumulonimbus clouds. This precipitation does not fall from every part of the cloud. At the same time, the nature of heavy rainfall is not associated with constant precipitation over a long period of time. As a result, the base of cumulonimbus clouds is seen most of the time.

Cumulus clouds do not occupy the sky completely. Through the gaps between the clouds, the side parts of these clouds are visible.

Cumulus and cumulonimbus clouds differ in their microphysical structure. Cumulus clouds (flat, medium, powerful) have a drop structure. Cumulonimbus clouds have a mixed structure, that is, they contain both drops and crystals.

Cumulus clouds are not associated with precipitation. The exception is powerful cumulus clouds in tropical latitudes. There, precipitation can fall from these clouds in the form of showers. Cumulonimbus clouds are associated with heavy precipitation in the form of heavy rain, heavy snow, heavy sleet, graupel, and hail. Most of the developed cumulus clouds (powerful cumulus and cumulonimbus clouds) are in the area of negative temperatures. This does not mean at all that these clouds contain mostly crystals. This is not true. First, at negative temperatures in the atmosphere there can be not only crystals, but also supercooled droplets. All cumulus clouds have water in their systems only in two phase states (gaseous and liquid). Cumulonimbus clouds contain water in three phase states (gaseous, liquid and solid).



In general, cumulus clouds are associated with ascending movements. However, in developed cumulus clouds, vertical movements can be not only ascending, but also descending. Downward movements in the clouds and beyond are compensating for upward movements. Ascending motions in cumulus clouds predominate. However, downward movements prevail around the cloud and above it. In developed clouds, ascending movements acquire high speeds, especially in the upper parts. Downward movements outside of such clouds cannot fully compensate for upward movements. As a result, downdrafts also appear inside the clouds.

Ascending air movements lead to a decrease in its temperature. The downward movement of air increases its temperature. Multidirectional vertical air movements lead to interesting effects. A developed cumulus cloud lives, as it were, in a warm air shell. The width of this shell and the temperature difference between the near-cloud space and the cloud itself increase from the base of the cloud to its top. Near the top of the cloud (above the cloud and around it), the warm shell is more pronounced. Sometimes a cold layer first surrounds the cloud, but a warm shell is found behind it. The cold interlayer is associated with evaporation of cloud particles into the near-cloud space. In the cloud itself, air movements of the same direction and approximately the same vertical speed are combined into separate convective flows. These flows may be in the form of jets or bubbles. The jet has a diameter, which varies little in the vertical direction in the cloud. The bubble has a hemispherical shape of the front part and a plume of relatively colder air. Unlike a jet, the bubbles have an unequal cross section in their front and rear parts.

Convective elements in the form of jets and bubbles in and below cumulus clouds are called thermals.

The direction and speed in convective flows depend on the altitude level in the cloud and on the stage of development of this cloud. With height, the speed of vertical movements increases. The degree of development of turbulent motions also increases with height. In the upper-

most part of the cumulonimbus cloud, the movement is so disordered that it is impossible to determine the direction of the vertical movements.

As for the direction of the vertical flows, in rising clouds and at the very beginning of their mature stage, ascending movements predominate. At the end of the mature stage and in the disintegrating clouds, descending movements predominate.

The height of the lower boundary and the thickness of cumulus clouds depend on the fluxes of radiant energy. Solar energy is absorbed mainly by the underlying surface. The heated underlying surface gives a flux of long-wave radiation into the atmosphere. The greater the fluxes of solar radiation, the greater the fluxes from the underlying surface into the atmosphere. The temperature of the underlying surface and the atmosphere have a great influence on the position of the lower boundary of the clouds and their power.

The lower boundary of the cloud practically coincides with the level of condensation. At the condensation level, the relative humidity is equal to the saturation value (in most cases it is 100%). Relative humidity depends on air temperature. The higher the temperature is, the higher the air must rise to cool and reach the state of saturation with water vapor.

Thus, where there is a large flow of solar radiation, the level of condensation is higher. In such areas, a higher position of the lower boundary of clouds is also noted. In these areas, the values of vertical temperature gradients are high. With such vertical gradients, air particles acquire positive acceleration values and travel large vertical distances. As a result, clouds are characterized by high power values. The position of the lower boundary of cumulus clouds coincides with the height of the condensation level and can be calculated using the formula of the American meteorologist William Ferrel:

$$z_k = 122 (T_0 - \tau_0), \quad (6.1)$$

where  $T_0$  and  $\tau_0$  are the values of temperature and dew point near the underlying surface. The derivation of (6.1.) is given in [2]. According to (6.1), the level of condensation is higher in more southern latitudes, in summer, and during the day. So the height of the lower boundary of cumulus clouds varies within a fairly wide range — from 300 m to 2500–3000 m. However, in summer, during the daytime in southern latitudes, cumulus clouds are located higher, and during transitional periods at night, these clouds are located lower. In winter, cumulus clouds can appear only in the variant of the initial and rapidly flowing stage of development of cumulonimbus clouds. Cumulonimbus clouds are observed not only in the warm period, but also in the cold period. In the latter case, cumulonimbus clouds form in cloud systems of cold atmospheric fronts.

The height of the base of cumulonimbus clouds is 0.6–1.0 km. At the same time, cumulonimbus clouds are higher in summer than in other seasons of the year, especially in winter.

The average thickness of these clouds is 1.5 km in winter, 3 km in spring, 4.6 km in summer, and 2.6 km in autumn. The given values demonstrate quite well the dependence of the power of cumulonimbus clouds on the influx of radiant energy from the sun.

An interesting feature of cumulus clouds is the process of involvement of air from the near-cloud space into the cloud system. At the same time, thermals of smaller sizes attach to themselves a smaller mass of air from the near-cloud space at a higher speed. With the development of a thermal, the intensity of its exchange with the environment weakens.

Cumulus clouds have a pronounced daily and annual variation in frequency. The annual and especially the daily course of recurrence is shown by clouds that are not associated with the development of non-periodic processes.

The daily variation in frequency of cumulus clouds is especially pronounced over land in summer. This is due to the nature of cumulus clouds. They are intramass clouds and form in warm air currents.

Annual fluctuations in the frequency of cumulus clouds are more pronounced over land that changes the nature of the underlying surface during the year. Such land is covered with snow during the cold period. The ascending motion streams over such a surface are not formed at this time. Everywhere, the development of cumulus clouds over land begins after the snow cover has melted.

At low latitudes over land, the annual variation in the frequency of cumulus clouds is not pronounced. This is due to the absence of a pronounced annual variation in the temperature regime and in the change of the nature of the underlying surface.

Cumulus clouds form over water during the cold season. Water in the cold period is warmer than land. This medium has a large heat capacity, so it cools down more slowly and less than land. The lower layers of the air above the water warm up in the cold period. This leads to the formation of cumulus clouds.

Cumulonimbus clouds are clouds of unstable air masses and cold fronts. Most often, such clouds have a frontal origin, so they are observed not only in the warm, but also in the cold period. In cold front systems, cumulonimbus clouds are observed in all seasons of the year. In warm times, they can also be present in the warm front cloud system.

Intramass cumulonimbus clouds develop during the warm period under the influence of overheating of the underlying surface and the air layer close to it. Such clouds reveal not only annual, but also daily frequency fluctuations. The maximum frequency of these clouds occurs in the evening hours. This is due to the fact that cumulonimbus clouds pass part of their development as cumulus clouds [3].

### **6.2.2. Wave clouds**

For the formation of undulating clouds, wave motions are necessary in the atmosphere above the level of condensation. In the crests of waves, the air makes upward movements, cools down, approaches the state of saturation and reaches it. As a result, clouds form on the crests

of the waves. In the troughs of the waves, the air makes downward movements and moves away from the state of saturation. In the troughs of the waves, gaps in the cloudy environment can form. Cases are also possible when, instead of gaps in the cloud layer, a decrease in thickness is observed in the troughs compared to the thickness in the crests of the same waves. In any case, such clouds are visually perceived as a layer of cloud waves.

With the manifestation of wave motions in a cloudy medium, the observed clouds are defined by morphological classification as wave clouds (undulates).

Wave clouds are formed mainly in two processes:

- 1) wave motions at the boundaries of two media;
- 2) overcoming obstacles by air.

In the first case, the lower layer of the air has a lower temperature, and the upper layer has a higher temperature. Let the air temperature decrease with height in the lower layer, and the air move upward. The rising air enters the warmer upper layer and turns out to be denser and colder than the surrounding environment. Then the rising air descends. As a result of adiabatic heating during the downward movement, this air becomes less dense and warmer in the cold air surrounding it. Then air rises, and the process repeats.

This is how wave motion is formed. During this movement, wavy clouds can form or transform in the cloud environment from other clouds into wavy clouds. If the waves are stable, then their amplitudes are conserved in time. In this case, the appearance of undulating clouds is preserved in time.

Among the various waves, the longer waves are the most stable. The greater the temperature difference in adjacent layers and the smaller the difference in wind speeds in the same layers, the shorter waves can also become stable [2].

In the second case, the air flow overcomes an obstacle. Then the top of the obstacle is the creator of the wave crest in the atmosphere. There is

a wave motion. This movement can be traced to great heights and at a sufficiently large distance from the obstacle that gave rise to it.

Wave clouds are observed in all three tiers (Cc und., Ac und., Sc und., St und., As und.). Even cumulus clouds can form on the crests of waves with a large amplitude. They usually don't have much power.

The length and amplitude of the obstacle waves depends on the shape of this obstacle (height and width of the mountain, the angle of incidence of the slopes), as well as on the parameters of the atmosphere. First of all, these parameters are the vertical temperature gradient and the speed of the flow running on the obstacle. Of course, the formation of obstacle clouds depends on the humidity of the air before it rises.

The wave structure of stratus clouds is best seen from above (from an airplane or from a mountain top). This is due to the fact that the wavelength can be large, so it is difficult to see the wave from below, from the underlying surface.

It is important to note that undulating clouds are most often the result of the transformation of other clouds. If frontal clouds Ns – As are transferred from areas of low or lower pressure to areas of high or higher pressure, then under the influence of descending movements these clouds are transformed into wavy clouds (St, Sc or Ac und). This is due to the fact that subsidence inversions are formed in anticyclones under the influence of downward movements. In them, warmer air is above the colder air. Further, the process proceeds as described above — wave motions are formed at the interface between two layers with different temperatures. In a downward movement, the upper part of the cloud, transferred from areas of lower pressure, dissipates, moving away from the saturation state. The rest of the cloud layer is affected by wave motion.

There are other options for transforming clouds. For example, when the troughs are filled, and the upward movement changes to the downward movement, the Ns clouds can transform into St and Sc. In addition, when an atmospheric front passes over a region with a complex relief, the frontal clouds As transform into Ac und.

Most often, the transformation into wave clouds occurs from frontal clouds of the lower layer Ns. From the frontal clouds of the middle tier As, the transformation occurs less frequently. This is due to the higher power of Ns compared to As. With the destruction of the upper part, high-stratus clouds have less opportunity to remain in their lower part for two reasons:

- 1) clouds As are less powerful compared to clouds Ns;
- 2) clouds As have a lower water content due to their position compared to clouds Ns.

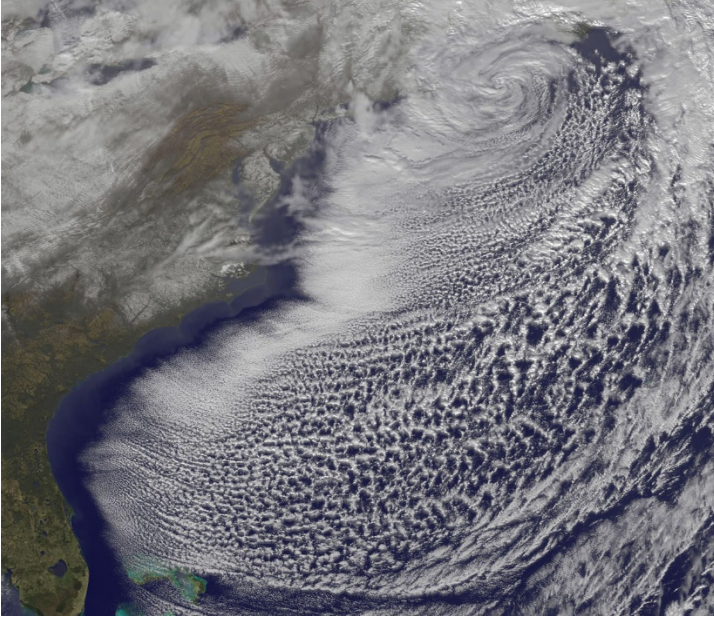
Sometimes wave motions lead to the formation of convective cells. These cells are of two types — open and closed. The type of cells is determined by the presence or absence of clouds in the central parts of these cloud elements.

In open cells, upward movements are created at the cell peripheries. These upward movements favor cloud formation. There are no clouds in the central part of open cells.

In closed cells, ascending movements and associated clouds are observed in the central part of the cells. At the periphery of closed cells, downward air movements are created, so there are no clouds there.

The horizontal diameter of open and closed cells is approximately the same and has a maximum repeatability in the range of 31–40 km. Open cells are formed in those layers where the vertical temperature gradient decreases with height. Closed cells are traced in layers associated with an increase in vertical temperature gradients with height.

Closed cells often form over cold ocean currents in warmer air. In this case, the temperature difference between water and air is negative. With distance from the underlying surface, this difference acquires a zero value, and then positive values are traced. These positive values increase with height. An increase in the vertical temperature gradient with height is a favorable condition for the formation of clouds. As a result, clouds are traced in the centers of closed cells. In the case of open cells, the opposite is observed.

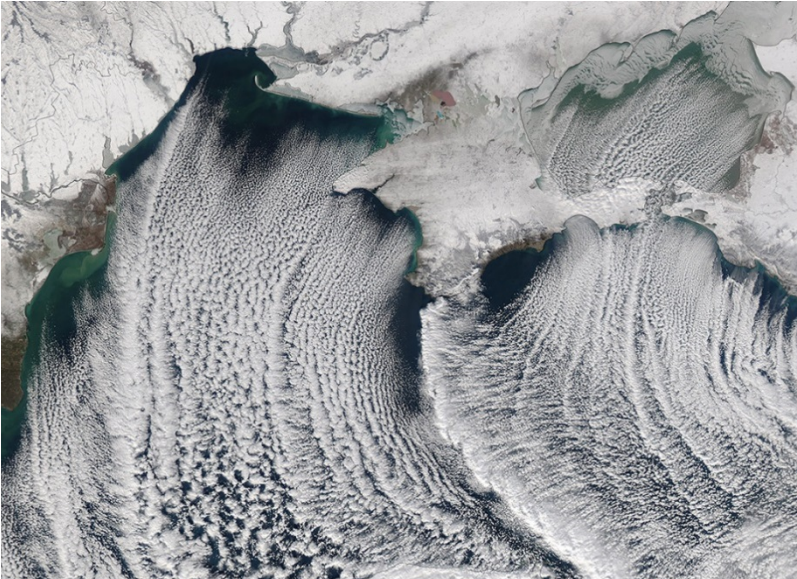


**Fig. 6.1.** Rear part of an occluded cyclone with rows of cells. The Florida peninsula is visible at the bottom left of the image. December 27, 2010



**Fig. 6.2.** Ridges of cells over the Bering Sea on April 7, 2013





**Fig. 6.3.** Ridges of cells over the Black and Azov Seas on January 8, 2015.

The Crimean Peninsula and the Kerch Strait are visible in the upper part of the image

The reason for the formation of cells is related to the difference in wind speeds in two adjacent layers. Wave motions are created at their border. In the upper layer, the speeds are usually higher. For this reason, the crests of the waves in the upper layer overtake the hollows of the same waves in the lower layer. At the same time, the elements of the waves quickly “overwhelm” and regenerate into rows of closed cells.

Cells are also observed over land, but less frequently than over water surfaces. In cases where cells are formed over land, the temperature difference between the underlying surface and the air must be significant. In any case, the areas where open cells are observed are filled with cold air masses, and the underlying surface there is warmer. In the areas of observation of closed cells, the underlying surface has a lower temperature than the air moving above it.

Sometimes individual cells are combined into ridges. This happens when the wind speed increases with altitude while the direction of the wind stays the same or nearly the same. In such cases, the cloud ridges are oriented approximately in the direction of the wind. Above the ridges is a capping inversion (an elevated inversion layer).

### **6.2.3. Stratiform clouds**

Stratiform clouds develop more horizontally than vertically. Based on this, some researchers attribute all tiered clouds to this class of clouds. The only exceptions are cumulus clouds (Cu and Cb). With another approach, stratus clouds (St) are not included into it. Our view on this approach is presented in [2]. In our opinion, stratiform clouds should be divided into two groups:

- 1) frontal stratiform clouds;
- 2) intramass stratiform clouds.

The main mechanisms for the formation of stratiform clouds are as follows:

- a) adiabatic cooling of air during upward movements, in particular, movements of a large scale;
- b) turbulent air mixing;
- c) radiative cooling due to radiation from the underlying surface and air layers, especially in the lower troposphere.

In the case of frontal stratiform clouds, we are dealing with a large-scale rise of a warm air mass along a wedge of a cold air mass. This movement has a small vertical component. This is due to the slight inclination of the frontal zone separating warm and cold air masses. As a result, the movement of a warm air mass upwards is rather slow. However, even such a small component of the vertical movement of a warm air mass is very important for its adiabatic cooling. In this case, the air approaches the state of saturation with water vapor. With a further rise in the air mass, a phase transition of water in the atmosphere takes place. In the lower part of the rising mass, this transition may be the condensation of water vapor.

As a result of this large-scale rise of large air masses, a sufficiently powerful frontal cloud system is formed in the warm air mass above the cold air mass wedge. The sequence of clouds of various shapes, the width of a multilayer cloud array, the thickness of clouds, and other characteristics of frontal cloud fields depend on many factors. They are: the type of the front, the distance of its individual sections from the center of the cyclone, as well as the distance from the line of the atmospheric front [5].

In cases of a warm front, cirrus clouds (Ci) are traced ahead of its surface line. Cirrostratus clouds (Cs) pass behind and below them. Altostratus clouds (As) are observed even closer to the surface front line and even lower. Last and below other frontal clouds are the most powerful stratiform clouds — nimbostratus clouds (Ns). The base of these clouds is located within the lowest part of the troposphere (within a kilometer layer from the underlying surface).

The cloud system of the slowly shifting cold front is to a large extent a “mirror image” of the cloud system of the warm front. Nimbostratus clouds (Ns) appear first, and cirrus clouds (Ci) are the last to appear. At the same time, this cloud system is both “less correct” and less wide compared to the warm front cloud system. So, cumulonimbus clouds join the tiered clouds here. They form in cloud systems of a cold slowly moving front. This is due to the rise of a warm air mass at a higher speed. At high speeds, the degree of turbulence in the rising air mass increases.

Any cold front (even a slowly moving one) has higher speeds than a warm front. For this reason, cumulonimbus clouds in cold front systems tend to be combined with layered frontal clouds at any time of the year, but more often during the warm period. In summer, cumulonimbus clouds can also appear in cloud systems of warm fronts. In such cases, cumulus clouds introduce various inhomogeneities into the cloud frontal systems. This is manifested in the appearance of the cloud system (especially in the form of cloud bases) and in the nature of precipitation. So, constant precipitation associated with nimbostratus clouds

throughout the year, and additionally with altostratus clouds in winter, is joined by heavy rainfall (or snowfall) from cumulonimbus clouds.

Radiative cooling is an important factor in the formation of intramass stratus clouds *St nebulosus*. Lowering the temperature can bring the cooling air to a state of saturation with water vapor. The lower part of the air gives off its heat to the underlying surface and at the same time cools down. The higher layers give off less heat to the underlying surface than the lower layers. As a result, the air in the higher layers may be warmer than the lower layers. This distribution of temperature with height is an inversion. Subsequently, with some heating of the lower layer, the surface inversion in its lower part is destroyed. So the surface inversion becomes elevated. Under such an inversion, water vapor and condensation nuclei are concentrated. Fogs and stratus clouds can form here *St neb*. Such clouds are often referred to as sub-inversion clouds. In the cases of these clouds, the elevated inversion is located higher than in fogs. Often, this situation is also noted in the cases of some stratocumulus clouds (*Sc*).

Interestingly, the already formed *St* and *Sc* clouds can favor the formation of an over cloud inversion. This explains the presence of inversion in the absence of strong radiative cooling. The mechanism of formation of a over cloud inversion under the influence of the above clouds is considered in [2].

#### **6.2.4. Some characteristics of clouds**

1. For all clouds, their lower boundary rises with an increase in the fluxes of solar radiant energy coming to the underlying surface. Clouds are higher at lower latitudes, in summer, and during the day.

2. Water content and thickness of clouds increase with increasing temperature at the lower boundary of the cloud. At high latitudes, clouds are less watery and powerful than at middle and especially at low latitudes. During the warm period, the clouds are more watery and powerful than during the cold period.

The change in water content with height depends on the shape of the cloud. In the St and Sc clouds, the highest water content is between the middle and upper parts of these clouds. Several levels of the highest water content can be traced in the Ns and As clouds. These levels are located near the base, in the middle of the cloud, and another one with a cloud thickness of at least 2–3 km.

3. The amount of clouds is related to their shape. Clouds Ns, As, St, and often Sc have the form of a continuous layer. It most often covers the entire firmament (10 points). These forms are usually observed in winter. However, they dominate over other forms of clouds. In summer, the frequency of cumulus and cumulonimbus clouds increases. This leads to a decrease in cloudiness in the warm period. So the amount of cloud cover equal to 7–9 and 4–6 points is observed more often in summer than in winter. Clear skies are not typical for summer. It occurs more frequently in winter than in summer. In areas dominated by the Asian anticyclone in winter, the frequency of clear skies increases, and continuous cloudiness decreases compared to neighboring areas.

4. The phase state of clouds depends on their thermal state. All clouds of the upper tier have a crystalline structure due to their location in the field of low temperatures. Clouds of the middle tier (Ac) are more often droplets. Frontal clouds of the same middle tier (As) are more powerful, so they are more often mixed or crystalline. Low-level clouds of intramass origin (St and Sc) usually have a droplet structure. Powerful frontal nimbostratus clouds of the lower layer (Ns) have a mixed structure. Cumulus clouds (Cu) are droplets, and cumulonimbus clouds (Cb) have a mixed structure.

Features of the phase structure of clouds have a great influence on many atmospheric phenomena and processes. So the phase state has a great influence on the growth of cloud particles, precipitation, on visibility in clouds, on the icing of aircraft in them, and on the formation of a number of optical phenomena [9].

5. Size distribution of cloud particles. Knowing the size of cloud particles is important in solving many problems related to atmospheric processes. To obtain the necessary information, droplet size distribution curves are built for clouds. With their help, formulas are created to get an idea about these aqueous aerosol media. To build the above distribution curves for the actual material, the values of the distribution function are plotted on the y-axis. The distribution function is  $f(r) = n_r/n$ , where  $n_r$  is the number of drops with a radius from  $r$  to  $r+l$ , and  $n$  is the total number of drops in the same volume of air. The radius or diameter of the droplets is plotted along the abscissa axis.

There are some characteristic features of droplet size distribution curves in drop clouds and fogs. These curves show the maximum repeatability of a certain radius ( $r_m$ ). This radius is called the modal radius. It is easily determined based on actual data. In addition, this radius can be calculated using the value of the arithmetic mean radius [2, 12]:

$$r_m = \frac{2}{3} r_{cp}. \quad (6.2)$$

The main maximum and the radius on which it falls ( $r_m$ ) depends on the shape and type of the cloud, on the stage of its development. The modal radius  $r_m$  is the top of the distribution curve. From the values of the arithmetic mean and modal radii, one more important radius for understanding atmospheric processes can be calculated. It is called the surface-equivalent radius ( $r^*$ ):

$$r^* = \frac{5}{2} r_m, \quad (6.3)$$

$$r^* = \frac{5}{3} r_{cp}. \quad (6.4)$$

The surface-equivalent radius ( $r^*$ ) of drops of fog or clouds makes the greatest contribution to water content.

In clouds and fogs, the modal radius is less important than the arithmetic mean radius. This is manifested in the features of droplet size distribution curves in these aqueous aerosol media. From the top of the curves where the value is marked as  $r_m$ , the frequency decreases faster towards smaller droplets than towards larger drops.

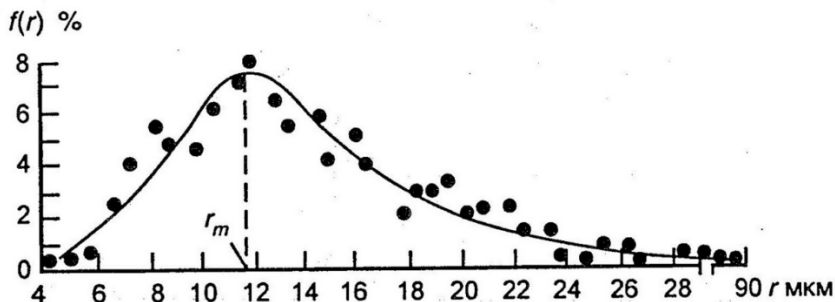


Fig. 6.4. Size distribution of droplets in droplet-liquid fogs at air temperature from  $-9$  to  $-10$  °C

The difference between the values of the arithmetic mean and modal radii depends on the shape and type of the cloud and on the stage of its development.

In the course of various experiments, the values of the radii in the clouds were determined. Let us present the results obtained by A. M. Borovikov and I. P. Mazin. According to these scientists, the average radius is  $4.5$   $\mu\text{m}$  in stratus (St) and  $4.7$   $\mu\text{m}$  in stratocumulus (Sc) and altocumulus (Ac) clouds. In nimbostratus clouds (Ns) the average radius is within  $6$ – $7$   $\mu\text{m}$ , in cumulus flat clouds (Cu hum.) it is within  $2.5$ – $3.5$   $\mu\text{m}$ , in cumulus medium clouds (Cu med.) it is  $3.5$ – $4.5$   $\mu\text{m}$ . In powerful cumulus clouds (Cu cong.) the average radius is  $4.5$ – $7.0$   $\mu\text{m}$ . As the cloud develops, the droplet size increases from  $2.5$ – $25$   $\mu\text{m}$  at the stage of cloud formation to  $25$ – $100$   $\mu\text{m}$ .

At the initial stage, cloud drops are the same as fog drops. Further, the cloud drops grow to the size of a drizzle. Under conditions of weak upward motions, such small droplets can slowly fall. Such conditions are created in stratus clouds (St).

Any cloud, and likewise fog, is a polydisperse medium, that is, it contains particles of various sizes. The range of droplet sizes in clouds is larger than in fogs. This is due to the fact that clouds are more powerful than fogs. For this reason, clouds are in the fields of meteorological quantities (including temperatures) with a large range of values.

If we consider the entire range of droplet sizes in clouds, then the smallest of them have a radius of 1–2  $\mu\text{m}$  or less. The largest cloud drops have a radius of more than 100  $\mu\text{m}$ . These cloud particles overcome the force of air resistance and fall out of the cloud, that is, they become precipitation. Such large drops are present in Cb and Ns clouds.

If the clouds have a crystalline structure, or such a structure has a part of the cloud mixed in terms of microphysical structure, then the shape of the cloud particles is also determined. Crystalline cloud particles have predominantly the form of a hexagonal prism. If such a prism has a large base area at a low height, then it is called a plate. If the height of the prism is significant, and the base has a small area, then such a prism is called a column.

With intense sublimation, rapid growth occurs at the corners of the ice plates. As a result of this process, stars of various shapes are obtained. All stars have six rays, since the plate has six corners. Such plates are called snowflakes.

Cirrus clouds consist mainly of crystals in the form of columns or hexagonal prisms 0.1–0.3 mm long. Cirrostratus clouds are composed of columns and hexagonal plates.

In the upper part of the Ns clouds of mixed microphysical structure, there are long columns. They are needles 2–3 mm long. Needles are observed in Ns clouds, since these clouds are relatively warm. In addition, Ns has hexagonal plates.



**Test questions:**

1. What is a cloud?
2. What are the differences between clouds and fogs?
3. What are the similarities between clouds and fogs?
4. What are the classifications of clouds?
5. Characterize the classes according to the developed classifications of clouds.
6. What movements are typical for a cumulus cloud and for the near-cloud space?
7. How does the stage of development of cumulus clouds affect the nature of their movements?
8. What dependence does the Ferrel formula show?
9. What factors influence the thickness of cumulus clouds?
10. What is a thermal?
11. What is the engagement process?
12. Do cumulus and cumulonimbus clouds differ in terms of formation?
13. When do wave clouds develop?
14. What are convective cells?
15. What are the types of convective cells?
16. What information do different types of convective cells contain?
17. Under what conditions are ridges of cells formed?
18. What mechanisms lead to the formation of stratus clouds? On what factors does the water content in the cloud and its distribution with height depend?
19. How are the number and shape of clouds related?
20. What is the characteristic phase state of various clouds?
21. What are the features of the size distribution of cloud droplets?

## 7. ATMOSPHERIC PRECIPITATION

Atmospheric precipitation is water in liquid and solid state when it is deposited on the underlying surface from the atmosphere. All atmospheric precipitation is divided into two groups.

1. Precipitation appearing on the underlying surface and its objects as a result of direct deposition of moisture from the atmosphere. The mechanism for the appearance of this type of precipitation consists in the phase transition from vapor to water (condensation) or from vapor to ice (sublimation). As a result, moisture is deposited from the lower part of the atmosphere when it comes into contact with the underlying surface. This type of precipitation constitutes a minor part of all atmospheric precipitation. Such precipitation is not associated with clouds and is called ground precipitation.

2. Precipitation falling from clouds. This type of precipitation makes up the bulk of atmospheric precipitation.

### 7.1. Ground hydrometeors

Ground hydrometeors, or ground precipitation, are atmospheric precipitation of the first group described above. These include dew, hoarfrost, liquid and solid deposits, rime ice, and glazed frost.

Let us consider each of these atmospheric phenomena, as well as the conditions for their formation.

1. Dew and hoarfrost are formed as a result of a noticeable decrease in the temperature of the underlying surface and the air adjacent to it.

In the diurnal course, the formation of this precipitation occurs at night and early in the morning, sometimes in the evening. Most often, ground-based hydrometeors form before sunrise.

The appearance of this atmospheric phenomenon is favored by a weak wind or calm, a clear sky or a slight cloudiness. It is under such conditions that the radiative cooling of the soil and objects on its surface occurs.

In the annual course, dew is formed at a time when there is still enough water vapor in the atmosphere, but significant cooling is possible in the future. Such conditions are created with a sufficiently large evaporation from the underlying surface and if the atmosphere is able to accept water vapor. If there is then radiative cooling of the underlying surface and the air above it, then dew may form.

Evaporation and the ability of the atmosphere to receive water vapor depend on the temperature of the soil and the lower air layer. As a result, dew forms when these temperatures are not very low. Subsequent cooling occurs during long nights, when heat losses due to radiation are significant, and there is no compensation for these losses. All of the above leads to the conclusion that dew is typical for the end of summer. It is formed when the temperature drops to values not lower than 0 °C. Dew is a symbol of the coming autumn.

An atmospheric phenomenon close to dew is hoarfrost. Hoarfrost is solid precipitation with a delicate structure. The difference between dew and hoarfrost is manifested in the processes of different phase transitions. Dew is formed in the process of condensation, that is, during the phase transition from vapor to water. Hoarfrost is formed as a result of sublimation of water vapor, that is, in the process of phase transition from vapor to ice, bypassing the liquid phase.

The differences in phase transitions are related to the difference in the temperature conditions of these phase transitions. In contrast to dew, hoarfrost forms at negative values of the temperature of the underlying surface. As a result, hoarfrost is a sign of the coming winter.

2. Liquid and solid deposits, unlike dew and hoarfrost, form on vertical surfaces. These surfaces can be walls and columns of houses, fences and gates, or tree trunks. Between themselves, liquid and solid

deposits differ in the same way as dew and hoarfrost — in their phase transition.

Liquid deposits in the form of individual drops or a continuous film of water are formed during the condensation process.

Hard deposits are formed in the process of sublimation. They look like densely packed small ice crystals.

The formation of liquid and solid deposits is associated with cooling of vertical surfaces and subsequent advection of heat. Warm air comes into contact with cold vertical surfaces on their windward side. Further, this air cools down and approaches the state of saturation. With further cooling an “excess” of moisture in the air is deposited on these surfaces at a new, lower temperature.

Moisture precipitation in the form of film or plaque can occur at any time of the day when conditions for heat advection are formed. Liquid and solid deposits are formed during the thawing season.

3. Rime ice is a very beautiful type of solid precipitation. It is formed on the branches of trees and shrubs, on wires and poles, on ledges and corners of buildings. Rime ice is different from hoarfrost and solid deposits. It differs from solid deposits in its structure, and it differs from hoarfrost in the time and conditions of formation. Rime ice often forms in cold, foggy or cloudy weather.

Rime ice can be of two types — granular and crystalline. Granular rime ice is snow-like precipitation, and crystalline rime ice consists of very small crystals. Granular rime ice practically does not glitter. This is due to the fact that the deposits in the form of granular rime ice do not have crystalline structure; they are amorphous. In cases of crystalline rime ice, the faces of ice crystals glitter.

Granular rime ice is formed in the water-to-ice phase transition (crystallization process). Crystalline rime ice is formed in another phase transition — from water vapor to ice (sublimation process).

Let the air mass contain some fog. Then granular rime ice is formed when fog drops come into contact with objects and freeze on these ob-

jects. In this case, small supercooled fog droplets freeze quickly; they do not flow over the surface of the objects. Crystalline rime ice is formed by the evaporation of fog or haze droplets and the sublimation of water vapor in the form of small crystals. In severe frosts, crystalline rime ice can form in the absence of fog or haze. At very low temperatures, sublimation of water vapor can occur on various fibers, branches, needles, and wires. Any rime ice forms on the windward sides of objects.

Crystalline and granular rime ice differ in the thickness of deposition. So, crystalline rime ice is a deposit of the order of 1 cm. Sometimes rime ice deposits can be up to several centimeters thick. Granular rime ice gives more powerful deposits, which can even reach a meter.

4. Glazed frost is a dense layer of ice formed on horizontal and vertical surfaces due to the freezing of supercooled drops of fog, drizzle, and rain. Thus, in cases of glazed frost, we have a water-to-ice phase transition (crystallization).

Glazed frost differs from solid deposits in greater density and intensity. The fact is that before freezing, large drops spread over the surface. Glazed frost forms during mild frosts.

## **7.2. Atmospheric precipitation falling from clouds**

Precipitation falling from clouds differs in appearance and in the nature of the fallout. Precipitation with different patterns of fallout is formed in certain clouds, therefore, they have their own formation conditions. The nature of precipitation is a diagnostic feature of the atmospheric process. For this reason, among the three types of classification, the main one is the genetic classification of precipitation. It allows you to classify precipitation according to its origin. According to the genetic classification, there are:

- 1) extensive precipitation;
- 2) heavy rainfall (showers);

### 3) drizzling precipitation.

Extensive precipitation falls from clouds of frontal origin; therefore, it is frontal precipitation. Throughout the year, extensive precipitation falls from nimbostratus clouds (Ns), and during cold seasons from altostratus clouds (As). Extensive rain, snow, and sleet fall from Ns. Extensive snow or sleet falls from As. It is exceptionally rare during transitional periods that light extensive rainfall falls from As.

Extensive precipitation has the following characteristics:

- a) it falls out for a long time without interruption or with short interruptions;
- b) in the process of precipitation, there are no sharp fluctuations in intensity (the amount of precipitation per unit time);
- c) it falls out over a large area.

Showers are precipitation of unstable air masses and cold, rapidly moving fronts. Such precipitation falls only from cumulonimbus clouds (Cb) at any time of the year. Heavy precipitation can be in the form of heavy rain, heavy snow, heavy sleet, graupel (snow pellets), and hail.

Shower precipitation falls not only from frontal Cb, but also from intramass Cb. Frontal showers differ from intramass showers in the amount of precipitation, change in intensity during precipitation, the duration of precipitation, and other characteristics.

Heavy rainfall has the following features:

- a) it starts and ends abruptly;
- b) it has sharp changes in intensity;
- c) it has a certain locality of fallout; even in nearby areas, showers differ in the very fact of falling, intensity, and time of falling.

Showers can fall against the background of extensive precipitation in cold front zones throughout the year and in warm front zones over land in summer.

Drizzling precipitation is very fine intramass precipitation. It is not connected with the passage of atmospheric fronts and changing weather conditions.

Drizzling precipitation falls from the densest and lowest stratus clouds (St). Sometimes drizzling precipitation falls from stratocumulus clouds (Sc). This type of precipitation has the following features:

a) the particles are homogeneous and so small that they are not visible to the eye;

b) the precipitation has very low intensity;

c) it does not have a directed fall;

d) it does not occupy large areas, being precipitation from intramass clouds.

The second classification of atmospheric precipitation is the morphological classification. It is a classification according to the appearance of precipitation. According to the morphological classification, they distinguish: drizzle, rain, snow, sleet, graupel (snow pellets), hail, freezing rain, ice needles.

The third classification of atmospheric precipitation is the classification according to their phase state. According to this classification, we distinguish the following groups of precipitation: solid precipitation (snow, graupel, hail, freezing rain, ice needles); liquid precipitation (drizzle, rain); and mixed precipitation (sleet).

Of course, it is possible to combine data for all three classifications. However, it is possible to use each of them separately. For example, the morphological classification is used by meteorological observers at meteorological stations. This classification allows qualitative observations of precipitation in the absence of knowledge about the state of the atmosphere. The genetic classification is used by specialists in the study of atmospheric physics. This classification is uniquely related to the state of the atmosphere and the processes taking place in it. The classification according to the aggregate (phase) state of precipitation provides important information for many branches of the national economy.

### **7.3. Processes of cloud particle enlargement and precipitation formation**

Precipitation falling from clouds consists of cloud particles. They went through a process of enlargement and overcame the force of air resistance. A large role in the process of falling of cloud particles belongs to the force of ascending motions.

The processes of enlargement of cloud particles are not the same at different stages of cloud development. In different clouds and at different stages of development, cloud particles acquire certain sizes. The fallout of such particles from the cloud essentially depends on the speed of ascending motions. Ascending motions in the subcloud layer may prevent a cloud particle from falling out of the cloud. Depending on the size of the particles and the speed of the upward motions, some particles will not leave the cloud. Other particles will not be delayed by the upward motions, but the speed of their fall will be reduced. Still other particles will quickly pass through the subcloud layer to the underlying surface and will not significantly reduce their mass and fall velocity.

If the upward motions are very weak, then even very small cloud particles can fall out of the cloud and become precipitation. Strong upward motions support the cloud particle in the cloud system. This makes it possible for the particle to grow larger if the process of increasing the size of its particles is well-expressed in the given cloud.

When considering the formation of precipitation, it is necessary to study the features of the cloud, the possibility of enlargement of cloud particles in it, the nature and strength of ascending motions in the cloud and subcloud layer. Regarding the problem of the processes of cloud particle enlargement, it is necessary to take into account the combined influence of several mechanisms of enlargement and identify the main ones at each stage of cloud development and at different stages of this process of enlargement. It follows from the above that the process of precipitation formation is complex and depends on many factors.



The possibility of precipitation from the cloud and the intensity of precipitation largely depends on the power of the cloud. The power of clouds determines the volume of the medium in which cloud particles can grow larger. With altitude, the speed of vertical movements increases, so in such clouds the particles are supported by strong upward movements and grow. More powerful clouds are in a larger temperature range in the vertical direction. As a result, in such clouds there is a higher probability of changing the aggregate state of cloud particles than in clouds of lower power. The aggregate state of the cloud and its vertical change determine the mechanisms of enlargement of cloud particles, the formation of precipitation and its fallout.

If a cloud has water in only two phase states (water vapor and water, or water vapor and ice), then the cloud particles are enlarged within their own phase (water or ice).

If the cloud has water in all three phase states (water vapor, water, and ice), then the crystals acquire an additional opportunity to grow due to drops. In this case, the process of “distillation” of water from a drop to a crystal is carried out. This additional mechanism for the growth of cloud particles is very important.

According to observations of cloudiness of the frontal type, precipitation falls from clouds of a mixed structure in about 90% of cases. From crystalline clouds and clouds with separated phases in height, precipitation falls in 70% of cases. Precipitation falls from drop clouds only in 9% of cases. Thus, it is most difficult for a cloud particle to grow in a liquid-droplet cloud. This is due to the fact that the saturation elasticity over drops is greater than over crystals. To achieve the value of saturation elasticity, more water vapor is required above the drops than above the crystals. A drop helps a crystal grow if the crystal is in a cloud next to a drop of water. This happens in clouds of mixed structure. The drop evaporates into their common cloudy space, and the crystal accumulates this moisture on its surface.

Precipitation falls from the cloud and moves in the subcloud space to the underlying surface. In this case, precipitation may not reach the underlying surface if it has completely evaporated in the subcloud layer. Precipitation can reach the underlying surface having evaporated to a significant extent or hardly changing the size of cloud particles. It depends on many reasons — on the temperature and humidity of the air in the subcloud layer, on the phase state of precipitation, on the speed of precipitation, on the thickness of the subcloud layer.

The probability of reaching the underlying surface is also higher for precipitation from clouds of mixed structure (78–83%). For precipitation from clouds of crystalline structure, this probability is approximately 50%. For precipitation from liquid-drop clouds, this probability is only 6%. At the same time, precipitation from drop clouds was usually noted as weak. Heavy precipitation often falls from clouds of mixed structure.

At different stages of cloud development, the main processes are different processes of cloud particle enlargement. Thus, at the beginning of cloud development, the enlargement of cloud particles is due to the process of water vapor condensation. This process occurs in ascending movements during adiabatic air cooling. As a result, supersaturation of air with water vapor around cloud particles is created and they grow.

When crystals appear among supercooled droplets, the growth of cloud particles is accelerated due to the process of “distillation” of water from droplets to crystals. When cloud particles increase to the size of 20-60 microns, the second stage of development of the cloud and its individual particles begins. This stage, when the particle sizes reach the specified range of values, begins depending on the strength of the upward movements. During strong upward motions, cloud particles are retained by them and continue to grow larger in the cloud.

At some point, particles of a certain size are no longer held by the updrafts and begin to move down. At this second stage, the falling particles grow in the cloud especially rapidly. At this stage of cloud parti-

cle enlargement, the main role belongs to the coagulation process. Coagulation is the coalescence of drops or the adhesion of crystals. In other words, there is a union of cloud particles.

Coagulation can be caused by various reasons. The main type of coagulation is gravitational coagulation. This type of coagulation is created under the influence of gravity. Large particles fall faster, catch up with small particles and combine with them. Other types of coagulation are described in [2].

Of course, more than one factor acts at each stage of the growth of a cloud particle. We are talking only about the main mechanism of enlargement, which is characteristic of each stage of the growth of these particles. Thus, at the beginning of the growth of a cloud drop or crystal, condensation (sublimation) growth proceeds actively. Under the influence of this mechanism, the droplet can increase in size by 2–3 times. Further, the efficiency of this mechanism drops sharply. At this time, the process of distillation of water from smaller drops to larger ones manifests itself. The result of this process is cloudy environment with droplets of approximately the same size. Such droplets usually cannot fall out of the cloud and become precipitation.

The same process proceeds if crystals appear in the cloud.

Among all the processes of precipitation formation, coagulation plays the greatest role. At the same time, with an increase in the size of the drops, the speed of coagulation grows more and more. At this stage, cloud particles make downward movements, that is, they fall in the cloud. During the downward movement of particles with sufficiently high velocities, neither condensation growth nor the processes of “distillation” of water can proceed actively.

With a sufficiently large cloud power, a large cloud particle can meet many small drops on its way down when falling and capture them. During this process, cloud particles increase in size very noticeably. As a result, large drops or flakes of snow fall out of the cloud.

Let us consider the possibility of precipitation from clouds of various forms. Precipitation does not form in cirrocumulus clouds. These clouds have a crystalline structure without large differences in the size of cloud particles. Such clouds are homogeneous. They do not undergo processes of “distillation” of water. In addition, Cc clouds have a low power, so coagulation processes are not typical for these clouds. In other words, cloud particles do not grow in Cc clouds in such a way that precipitation becomes possible.

Stratus, stratocumulus, cumulus and altocumulus clouds, with all the differences in the height of their base, in thickness, and formation conditions, have a similar phase state. All of them usually have a liquid-drop structure.

Of course, these clouds have different cloud particle size distribution curves. As for their power, in stratus and stratocumulus clouds the thickness is low (up to 1 km). The velocities of ascending movements in the cases of these clouds are also small, especially in stratus clouds (up to 1 m/sec). Large drops of course do not form in such clouds.

In stratus clouds, condensation growth of droplets occurs. However, these clouds exist for a long time — a day or more. During this time, in the densest and most powerful St, coagulation can occur and drizzle can form. In conditions of weak upward movements, drizzle can fall out of the cloud. In cases of the lowest stratus clouds, drizzle may reach the underlying surface. This is favored by high humidity in a small sub-cloud layer.

In the cases of stratocumulus clouds, the ascending movements are stronger than in the cases of stratus clouds. As a result, larger droplets are formed in Sc. Sometimes they fall as precipitation. This precipitation usually falls in the form of drizzle-like droplets, or in the form of small and rare snowflakes. Sc clouds are located higher than St. As a result, when precipitation falls from stratocumulus clouds, they evaporate in a larger subcloud layer. Precipitation from Sc is less frequent than from St.

In the case of cumulus clouds, ascending movements in some cases have high velocities (up to 10 m/sec). The thickness of these clouds can be several kilometers. At the same time, cumulus clouds are a liquid-drop medium. In it at first the growth of cloud drops occurs due to the condensation factor. This is followed by the process of “distillation” of water from small drops to large drops. A stronger mechanism of “distillation” of water (from droplets to crystals) is absent in this cloudy medium due to the absence of crystals.

Unlike long-lived St and Sc clouds, which are observed during relatively weak upward movements, cumulus clouds “live” for a short time during strong upward movements. These upward movements trap cloud particles within the cloud. The processes of coagulation during ascending movements increasing with height are not well-pronounced. This is due to the fact that the process of falling of drops in the cloud is weakened. Subsequently, cumulus clouds either break up or transform into Cb. At the same time, there are cases of droplets falling out of powerful cumulus clouds. Such precipitation may occur in tropical latitudes. With high water content and cloud thickness in the field of high temperatures, the change in the values of saturation elasticity ( $E$ ) is significant even with small changes in the air temperature. As a result, in tropical latitudes, another mechanism of “distillation” of water is carried out — from a warm drop to a cold one. This additional mechanism of cloud particle enlargement plays a large role in the formation of precipitation. Thus, in tropical latitudes, showers can fall from powerful cumulus clouds. Sometimes these showers can even be heavy.

Alto cumulus clouds, with the exception of castellanus (Ac cast.) [3, 4], are not powerful. They do not have the processes of enlargement of cloud particles to such an extent that precipitation falls out them. Only in the case of one type of Alto cumulus clouds (Ac virga) precipitation falls, but it does not reach the underlying surface. With such clouds, precipitation bands can be traced to a certain height from the base of the cloud. However, Ac vir. clouds are not observed frequently. In connec-

tion with the above, it can be concluded that Ac clouds as a rule do not give precipitation.

In crystalline clouds, where generally more favorable conditions for the growth of cloud particles are created, especially in frontal clouds, precipitation is often formed. Frontal clouds of crystalline structure are Ci, Cs and part of winter As. The enlargement of crystalline cloud particles does not require such a large amount of water vapor as the enlargement of particles in liquid-drop clouds. On the other hand, during the formation of frontal crystalline clouds, a slow large-scale rise of a warm air mass along a cold mass wedge occurs. As a result, precipitation falls from Ci and Cs clouds but does not reach the underlying surface due to the high position of these clouds.

Clouds of the frontal type (As) have rather large values of water content and power. Precipitation in them is formed and falls out. In winter, even from thin As, precipitation reaches the underlying surface. This is due to the fact that the bases of the clouds, including As, are lower in winter than in other seasons. As a result, in winter, precipitation passes a shorter distance between the cloud and the underlying surface. In the subcloud layer, low temperatures are observed in winter. At such temperatures the evaporation rate is not high. Extensive snow from As in winter falls in the form of not only light, but also moderate snowfall.

During the development of the above clouds (Ci, Cs, parts of As), both stages of cloud particle enlargement are well-expressed — sublimation and coagulation.

In mixed clouds Ns, Cb, and some As, crystals grow during sublimation of water vapor from droplets. This process continues until all drops have evaporated. From this moment the As cloud can become completely crystalline.

The above does not apply to Ns and Cb. In these clouds, the drop medium is large and is always preserved. Most often, larger solid particles fall out of mixed clouds. In the process of falling out, the crystals

can keep their phase state in the cloud or outside it, or they can change it and fall out as rain.

Of the less powerful As mixed clouds, snow reaches the underlying surface only in cold weather. The As clouds have the highest position of their base among the clouds of mixed structure. It is clear that precipitation from As clouds has to travel a longer distance than precipitation from other clouds of mixed structure. Precipitation particles from As are smaller than those from Ns and Cb.

At the same time, precipitation always falls from altostratus clouds, as from clouds of frontal origin. However, it does not always reach the underlying surface. If the precipitation is crystals, they evaporate less on the way to the underlying surface. If drops move to the underlying surface in the subcloud layer, they evaporate. In summer, precipitation from As clouds never reaches the underlying surface.

Nimbostratus clouds of mixed structure (Ns) are the most powerful among tiered clouds. The thickness of these clouds is several kilometers. In their upper part, Ns clouds consist of crystals; in their lower part, the clouds consist of droplets. Nimbostratus clouds exist for a long time — hours, days or more. In these clouds, in addition to extensive rain, extensive snow, and extensive sleet, freezing rain can also form. It is formed when cloud particles fall in the cloud and enter layers with different temperatures. In such cases, a change in the phase state of cloud particles is possible. So, when falling, drops can fall into a layer with negative temperatures and freeze. In this case, precipitation in the form of freezing rain will come to the underlying surface.

Cumulonimbus clouds are the most powerful of all clouds in existence. These clouds are developing rapidly. However, such clouds do not exist for long — from tens of minutes to several hours. Cumulonimbus clouds develop in an unstable atmosphere. In such an atmosphere, wind speed and upward movements are of great importance.

With an intensive rise of cloud particles, first condensation and sublimation growth occurs, then the process of “distillation” of water from

a drop to a crystal takes place. The next stage of growth is the process of coagulation. It goes very quickly and productively in Cb clouds. This is due to the specific conditions of particle growth in these clouds. With strong upward motions (10 m/sec and more), Cb cloud particles remain in suspension for a long time during their growth. When the coagulation process begins, and the particles overcome the forces of air resistance, they reach a sufficiently large size. Very large cloud particles fall in the cloud layer of high power. There they attach to themselves a large number of small particles. First, the heaviest large particles fall out of these clouds. They create a downward flow. Smaller precipitation may then fall in this stream. It is known that Cb are characterized by a change in the intensity of precipitation in the process of precipitation.

Graupel and hail can fall from cumulonimbus clouds. Graupel is formed when a crystal collides with a drop in a mixed cloud. If a crystal and small drops meet, then the small drops quickly freeze onto the crystal. This process is called accretion (or riming). During this process, small water grains appear on the crystal. Under the influence of this process, flat ice crystals (plates, snowflakes) become spherical crystals of white or dull white color. This is graupel (or snow pellets). Snow pellets are heavier than snowflakes and fall out of the cloud before them. Later, snow pellets may continue to fall along with snowflakes (snow showers).

An ice pellet is formed when a crystal collides with a large drop. Large drops spread over the surface of the crystals, and only then freeze. In this case, a transparent grain is formed with a dull core in the center. Ice pellets can fall along with heavy rain.

Ice pellets are different in appearance from freezing rain. There is a visible white core in the ice pellet. There is no visible core in freezing rain particles. The white core of an ice pellet is the result of riming of a snowflake. The snowflake then turns into a snow pellet. If further on the surface of this snow pellet a large drop spreads, then an ice pellet



with a white core of a snow pellet is obtained. Graupel forms only in cumulonimbus clouds.

Hail also forms only in cumulonimbus clouds. This precipitation is due to the features of Cb with their high power, high speeds of updrafts and developed turbulence. In the growth of hailstones, dry and wet growths are distinguished. The nuclei of hailstones are large drops that freeze in the upper part of the cloud. Due to coagulation with supercooled drops, hail nuclei grow rapidly. If the water content is low, then accretion of the nucleus occurs. As a result, a layer of opaque ice forms on its surface. This growth is called dry hail growth. If the nuclei of hailstones are in a layer of high water content, then the drops spread over their surface. In this case, a film of translucent ice forms on the surface of the hailstones. This growth of hailstones is called wet growth.

The water content in different parts of an inhomogeneous cumulonimbus cloud is not the same. In such a cloud, dry and wet growths of hailstones alternate. As a result, the hailstone has a layered structure. This can be seen if the hailstone is cut vertically.

The contribution of the processes of dry and wet growth of hailstones is not the same. Large hailstones are formed mainly in the process of wet growth. During the coagulation process, the hailstones freeze together.

The size of the hailstones depends on the speed of the updraft. In addition, the size of the hailstones depends on how many times the particles that have fallen out of the cloud are again thrown into the cloud by the updraft. In such cases, the hailstone growth cycle is repeated. Hail is formed during the warm period, when the most powerful Cb are formed.

## **7.4. Precipitation and evaporation**

All clouds are divided into clouds that give precipitation, clouds that do not form precipitation, and clouds from which precipitation falls under certain conditions. Let us summarize. Precipitation is formed in frontal layer

clouds (Ci, Cs, As, Ns), as well as in clouds of vertical development Cb (both in frontal and intramass clouds). Precipitation falls from these clouds throughout the year.

Precipitation never falls from Cc clouds. In most cases, precipitation does not fall from Ac. In our latitudes, precipitation does not fall from Cu.

Precipitation occasionally falls out of Ac in streaks under the cloud (Ac virga). In the tropics, precipitation can fall from powerful cumulus clouds (Cu cong.). Fine precipitation can fall from the densest and lowest clouds (St and Sc).

The fate of precipitation in the subcloud layer depends on the size of precipitation particles. Their dimensions determine the speed and the direction of their fall. In addition, the time of their passage from the base of the cloud to the underlying surface and the possibility of evaporation also depend on their size.

If drops with a radius of more than  $100\ \mu\text{m}$  fall out of the cloud, they reach the underlying surface. However, their size does not change significantly. The size of a drop is determined by the lifetime of a drop on its way from the cloud to the underlying surface and the distance that this drop travels in the subcloud layer. So the distance traveled by a drop in humid unsaturated air is proportional to the fourth power of its initial radius. The lifetime of a drop is proportional to the square of its initial radius.

Formulas for determining the falling velocity of droplets of different sizes are given in [10].

Solid precipitation passes the layer between the base of the cloud and the underlying surface more easily. Such precipitation is observed during the cold period, when the clouds take a lower position. This shortens the path to the underlying surface. In addition, at this time, lower values are observed in the subcloud temperature field. In the field of relative humidity, higher values are recorded in the same subcloud layer and at the same time of the year. This reduces the evaporation of precipitation. In addition, the evaporation of solid precipitates has a lower rate compared to the evaporation of droplets.

Larger cloud particles have a directed fall, which determines a shorter path to the underlying surface. This is due to the influence of gravitational forces on cloud particles.

Small drops (drizzle) can reach the underlying surface only when they fall out of low clouds with weak upward movements. However, even under these conditions, drizzle does not always reach the underlying surface. This is due to the fact that very small precipitation particles are strongly affected by even the finest turbulent motions. As a result, fine precipitation moves along very complex and rather long trajectories. It is clear that the fact of reaching the underlying surface depends on the size of small droplets.

***Test questions:***

1. What is atmospheric precipitation?
2. What groups is atmospheric precipitation divided into?
3. What types of precipitation are ground hydrometeors?
4. What are the conditions for their formation?
5. What are the classifications of precipitation falling from clouds?
6. Give a description of the precipitation allocated for each classification.
7. What does the precipitation of each of the groups of the genetic classification indicate?
8. Why does drizzle form only in stratus and partly in stratocumulus clouds?
9. Why do graupel and hail only form in clouds of the same type? Name it.
10. What is the difference between the formation of ice pellets and freezing rain?
11. What are the mechanisms of cloud particle enlargement in clouds with different microphysical structures?
12. When can precipitation fall from cumulus clouds?
13. What are the main mechanisms of cloud particle enlargement at different stages of cloud development?
14. What role do ascending motions play in the enlargement of cloud particles?
15. What conditions are favorable for the precipitation falling from the cloud to reach the underlying surface?
16. Name the clouds from which precipitation 1) always falls; 2) never falls; 3) falls out, but does not reach the underlying surface; 4) falls out only under certain conditions; 5) reaches the underlying surface only under certain conditions.

## CONCLUSION

The last page of this manual has been turned over. We emphasize once again that it does not include all, even important details that make a student prepared in the field of Physical Meteorology. Physical Meteorology is the most important subject for our specialty.

Of course, more fully and deeper the content of this subject is presented in [1–3], as well as in [6–9].

This manual provides an opportunity to get support in the study of Physical Meteorology, especially for students who do not have a proper command of the Russian language.

The author hopes that such students will compare the content of the sections of this manual and the corresponding sections of [1–3, 6–9], published in Russian. When the same sections are studied from different sources, a more correct personal judgment is formed regarding each issue.

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*Учебное издание*

**Zh. V. Rybakova**

**BASIC PHYSICAL METEOROLOGY**

**Textbook**

*Издание подготовлено в авторской редакции*

Подписано к печати 23.06.2023 г. Формат 60×84<sup>1</sup>/<sub>16</sub>.

Бумага для офисной техники. Гарнитура Times.

Печ. л. 11,5. Усл. печ. л. 10,7.

Тираж 50 экз. Заказ № 5503.

Отпечатано на оборудовании  
Издательства Томского государственного университета  
634050, г. Томск, пр. Ленина, 36  
Тел. 8+(382-2)–52-98-49  
Сайт: <http://publish.tsu.ru>  
E-mail: [rio.tsu@mail.ru](mailto:rio.tsu@mail.ru)

ISBN 978-5-907572-05-8

