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The Heat Balance of the Earth's Surface

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by M. I. Budyko

Translated by Nina A. Stepanova

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by M.I. BUDYKO

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Office of Climatology

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Translator's Remarks

It was intended to keep this translation as close as possible to the author's style and avoid rephrasing.

Consecutive numbers were added to the list of references to Russian literature to facilitate the finding of transliterated Russian names which are listed according to the Cyrillic alphabet in the original text.

There are several names mentioned by the author in the text but not included in the list of references.

M. I. Budyko
The Heat Balance of the Earth's Surface

Errata

	Instead of:	Should be:
p. 34, Table 3, heading	"... kg-cal/cm^2/hour"	"... cal/cm^2/hour"
p. 42, Table 7, heading	"... kg-cal/cm^2/min"	"... cal/cm^2/min"
p. 98, §7. Radiation balance, 5th paragraph, 2nd line.	"... 220 cal/cm^2/year"	"... 220 kg-cal/cm^2/year"
p. 136, Figure 45,	"... kg-cal/cm^2/min"	"... cal/cm^2/min"
p. 137, Figure 46,	"... kg-cal/cm^2/min"	"... cal/cm^2/min"

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THE HEAT BALANCE OF THE EARTH'S SURFACE

Annotation

This monograph summarizes the results of investigations in heat balance climatology of the earth's surface. Various methods for determining the components of the heat balance are analyzed and systematized. Data on geographical distribution of all components of the heat balance and of their annual and diurnal variations are presented. Applications of the heat balance climatology to various problems of physical geography, agro-meteorology, and hydrology are interpreted. The utilization of heat balance data for the analysis of meteorological effectiveness of ameliorative measures is investigated.

This monograph can be used by scientists, aspirants and students, who are working in fields of climatology, meteorology, land geography, and oceanography, and also, by scientists and practitioners in other professions who might be interested in problems of the transformation of solar energy on the earth's surface.

Preface

Investigations of heat balance on the earth's surface are now occupying an important place in all hydrometeorological disciplines, including meteorology, climatology, land hydrology, and oceanography.

The main purpose of these investigations is the study of the causal principles which determine the meteorological and hydrological regimes in various geographical regions and could be used for prognostication and calculation of important hydrometeorological processes and phenomena. In fact, a whole series of investigations on heat balance has been made in order to: evaluate the effect of meliorative measures on climatic conditions near the ground; calculate the evaporation from the reservoirs which have been planned for construction; develop methods for forecasting the reservoirs' freezing dates; and solve many other practical problems.

In recent years, due to the initiative of the academician A.A. Grigor'ev, the data on heat balance have largely been used in studies of the general problems in physical geography including the problem of geographical zonality.

The rapid growth in the number of requests for material on heat balance in various regions has stimulated a considerable progress in climatological investigations of this subject, especially concerning the geographical distribution of the heat balance components.

As a result of work along these lines, that has been accomplished in the Central Geophysical Observatory (Voeikov's), and also by some other groups of scientists, the climatology of heat balance has been established by now as one of the branches of general climatology.

This monograph is devoted to the interpretation of the problems of heat balance climatology.

The author expresses his sincere gratitude to all persons who have read this work in manuscript and offered their comments.

Chapter I

Introduction

Solar radiation is the main source of heat energy for almost all the natural processes developing in the atmosphere, hydrosphere, and in the upper layers of the lithosphere.

On the other hand, the utilization of solar energy is of paramount importance in economical activities, and particularly valuable for agricultural production.

Consequently, the problem of the transformation of solar energy in the atmosphere, in the hydrosphere, and in the upper layers of the lithosphere (i. e., in the outer geographical medium), is very important for development of a large scope of problems in the practical, as well as in the theoretical field of knowledge.

The general aspect of the basic transformations of solar energy in the outer geographical medium could be interpreted, according to the most recent conceptions in the following way:

The flux of the solar radiation at the average distance of the earth from the sun is approximately equal to 1000 kg-cal/cm^2 per year. Because of the spherical shape of the earth, a unit of the surface on the outer boundary of the atmosphere receives, on the average, 1/4 of the total flux; i.e., about 250 kg-cal/cm^2 per year, and about 150 kg-cal/cm^2 per year is absorbed by the earth as a planet.

It is very significant that the main portion of the absorbed solar radiation - about 3/4 of the total amount - is absorbed by the earth's surface, whereas the atmosphere absorbs only 1/4 of it.

The surface of the earth, when heated as a result of solar radiation absorption, becomes a source of long-wave radiation which, in turn, heats the atmosphere. The presence of water vapor in the atmosphere, and also of some gases and dust particles that absorb the long-wave radiation, reduces considerably the effective radiation [1] of the surface as compared with that which would have been observed if the atmosphere would be perfectly translucent (the green house effect). As a result of the green house effect, the radiation balance of the earth's surface (i.e., the difference between the absorbed radiation and the effective outgoing radiation) is rendered positive.

Since the radiation balance of the earth as a planet is close to zero, on the average for the year, the radiation balance of the atmosphere, which is equal to the difference between the radiation balance of the earth as a whole and the radiation balance of its surface, is rendered negative.

Besides radiational transformations, a considerable redistribution of heat in the atmosphere in a vertical direction is accomplished by the

---

1) The difference between the long-wave radiation from the earth's surface and the opposite long-wave radiation from the atmosphere.

processes of moisture exchange which are connected with the expenditure of heat for evaporation at the level of the underlying surface and the release of latent heat of condensation in the atmosphere, as well as by processes of vertical turbulent heat exchange.

Along with the processes of vertical redistribution of solar energy, vigorous processes of horizontal redistribution of heat are developed in the outer geographical medium. Among them, of special importance, is the exchange of heat energy in the atmosphere and hydrosphere which takes place between the higher and lower latitudes. This exchange is induced by substantial differences in radiational heating rates over the spherical surface of the earth. It is accomplished by the macroturbulent heat exchange, by transfer of heat by sea currents, and also, by the redistribution of condensation heat in the atmosphere.

All these processes of solar energy transformation are induced by radiational factors and affect,enormously, the energy regime in the outer geographical medium. In particular, they modify considerably the radiation regime on the surface of the earth, which depends a lot on the circulation of the atmosphere and hydrosphere, on condensation and evaporation processes, etc.

In association with processes of the "first order" transformation of solar energy, which greatly affect the radiational and thermal regime, other processes of solar energy transformation are also developed in the outer geographical medium. These processes involve some comparatively small quantities of heat, and therefore they do not exert any noticeable or direct influence on the radiation and heat regime. They are usually of lesser significance in meteorological investigations, but some of them are of exceptional interest for some other branches of natural sciences, as for instance, the process of photosynthesis, which involves the transformation of radiation energy into a relatively stable form of chemical energy creating organic matter.

The basic data, from which the study of all forms of transformation of solar energy in the outer geographical medium proceeds now, are the data on radiation and heat balance. Among them, data on radiation and heat balance of the earth's surface are especially valuable, since the surface absorbs 75% of the total amount of solar energy absorbed by the earth, and consequently, it presents the main source of energy for the outer geographical medium.

Because of the fact that just at the earth's surface the greatest intensity of the most important natural processes is observed, like the biological, hydrological, exogenous geomorphological, soil formation processes, and others, it is obvious that data on the heat balance of the surface will be of an essential significance for the study of causal relationships of natural processes in the outer geographical medium.

In this monograph, the basic laws governing the radiation and heat balance of the earth's surface are analyzed in a geographical aspect; i.e., the climatology of the heat balance is interpreted.

The climatology of the heat balance includes, first of all, the methods for processing results of hydrometeorological observations which permit the calculation of the principal components of the balance.

The method for determining the components (terms) of the heat balance by proceeding from data of ordinary hydrometeorological observations is described in chapter 2.

The application of methods for calculating the balance components permitted us to develop the climatography of heat balance which includes, by now, the data for almost the whole surface of the terrestrial globe.

The fundamentals of the heat balance climatography are presented briefly in chapter 3.

As it turned out, it was possible to use the data on geographical distribution of the heat balance in solving various climatological problems, and also in studying some general problems of physical geography.

Thus, the utilization of heat balance data allows us to derive many conclusions about the regularities in heat exchange and the turn-over of moisture in the atmosphere. The results of these investigations are given in chapters 3, 4, 5, and 6. They complete the explanation of the causes of some climatic phenomena and provide for a quantitative interpretation of those processes which formerly were only qualitatively studied.

Among studies based on application of heat balance data,a special place is occupied by investigations of changes in the climatic regime as effected by amelioration (Ch. 5). Taking the data on heat balance into account it was possible to make quite a few conclusions of a definite practical importance.

As far as the transformation of solar energy on the earth's surface exerts a tremendous influence on the dynamics of all exogenous natural processes, it is obvious that data on heat balance can be successfully used for studying many geographical regularities. Accordingly,in chapter 4 an analysis is given of those relationships which connect the conditions of heat energy balance with hydrological processes and with the development of natural vegetation. It also turned out that it was possible to clarify some causal relationships which determine the phenomenon of geographical zonality that was first discovered by V. V. Dokuchaev.

The data presented in this monograph contributed to an improvement of existing concepts of the general balance of energy on the earth and of its water balance, and have also thrown more light on some other questions connected with the process of heat and moisture exchange.

§1. Heat balance equations

Equations of the heat balance represent some particular cases of the basic law in physics - the law of energy conservation. These equations can be derived for various volumes and surfaces in the outer geographical medium. In recent investigations, equations of the balance for the earth's surface and the equation of the balance for the system "earth - atmosphere," i.e., for the vertical column that extends through the whole geographical medium, are most frequently used.

To obtain the equation of the heat balance of the earth's surface all the streams of heat energy flowing between the element of the surface and the ambient space must be summarized.

We will designate the value of the radiational flux of heat by R';

the turbulent flux of heat between the underlying surface and the atmosphere by P' ; the heat flux between the underlying surface and the lower layers by A' ; and the expenditure of heat for evaporation (or the emission of heat from condensation) by LE' (L — the latent heat of evaporation, E' - the speed of evaporation or condensation). Since all other components of heat balance are usually much smaller than the here cited fluxes of heat, we may write, in the first approximation, the heat balance equation in the following form:

$$R' = LE' + P' + A'. \qquad (1)$$

The value of R' is considered to be positive if it designates an inflow of heat to the underlying surface, and all other values are positive if they designate the expenditure of heat.[2)]

The scheme of heat streams included in the equation of the heat balance is shown in fig. 1.

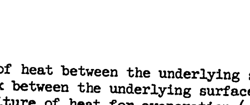

Figure 1

Scheme of the heat balance of the earth's surface.

Regarding those components of heat balance which have not been included into equation (1), a more considerable value could be attributed to the expenditure of heat for melting of snow or ice on the earth's surface (or to the gain of heat from water freezing processes). However, for longer

2) It is necessary to mention that in many papers a different system of signs (+ and -) is used for the heat balance, due to which all terms of heat balance equations have the same sign according to gain or loss of heat. Such a system of designating, although more logical, can lead, however, to some inconveniences. According to the system of determining the signs, the loss of heat by evaporation and turbulence from the earth's surface to the atmosphere is negative. But this contradicts the usual practices.

period averages (for a year or so), the latter value, as a rule, is considerably smaller than the main components of the balance, and only for some cases (for instance, the periods of snow melt in the middle and upper latitudes) should this value be included into equation (1) as an additional term.

The other components of the heat balance - heat streams originated by dissipation of mechanical energy of wind, of wind waves, tides, and currents, the heat flux (positive or negative) transferred by fall of precipitation which has a different temperature than that of the underlying surface, as well as the expenditure of heat for photosynthesis, and the gain from oxidation of biological substances are usually considerably smaller than the main components of the balance, and this is true for averages obtained for periods of any length.

Exceptions from this rule are possible, (as for instance in case of a forest fire, when great quantities of heat accumulated formerly by the process of photosynthesis are rapidly discharged) but are relatively rare.

The problem of accounting for the effect of heat advection must be considered separately. In some investigations, suggestions have been made to introduce into equation (1) an additional component representing the advective inflow of heat to the underlying surface. Therefore, we would like to give some simple considerations which will illustrate how wrong this point of view is, and also explain the relation between the horizontal transfer of heat and the heat balance components (Budyko, 1949b [42]).

The equation of heat exchange in the lower layer of the atmosphere, in the presence of a horizontal transfer of heat, will be:

$$\frac{\partial\theta}{\partial t} - u\frac{\partial\theta}{\partial x} = \frac{\partial}{\partial z}\left(k\frac{\partial\theta}{\partial z}\right), \qquad (2)$$

where θ — is air temperature, x — horizontal ordinate, directed according to wind direction in the lower layer of the atmosphere, z — vertical ordinate, u — wind speed in the lower layer of the atmosphere, k — coefficient of turbulent exchange, t — time.

Integrating equation (2) by z we obtain the following equation:

$$\int_0^z \frac{\partial\theta}{\partial t}dz + \int_0^z u\frac{\partial\theta}{\partial x}dz - \frac{P'}{\rho c_p} = k\frac{\partial\theta}{\partial z}, \qquad (3)$$

where $\frac{P'}{\rho c_p}$ - is the constant of integration, equal to the heat flux between the underlying surface and the atmosphere divided by the value of air density and heat capacity.

The direct effect of horizontal heat transfer on the heat balance of an air layer is represented by the term $\int_0^z u\frac{\partial\theta}{\partial x}dz$, the magnitude of this member depends, to a considerable degree, on z ; i.e., on the height of the analyzed layer. Computing the balance for the earth's surface, z must be approaching zero, and consequently the term $\int_0^z u\frac{\partial\theta}{\partial x}dz$ will become zero (u and $\frac{\partial\theta}{\partial z}$ are finite values). Since the term $\int_0^z \frac{\partial\theta}{\partial t}dz$ will also be equal to zero, the heat exchange between the earth's surface and the atmosphere at $z \to 0$ will be determined only by term $k\frac{\partial\theta}{\partial z}$, which shows the vertical heat flux. Estimating the order of magnitude of the components of equation (3), it could be established that, even for the lower layer of the atmosphere, 10-100 m thick, the terms $\int_0^z u\frac{\partial\theta}{\partial x}dz$ and $\int_0^z \frac{\partial\theta}{\partial t}dz$ are usually so much smaller than the term $k\frac{\partial\theta}{\partial z}$, that they could be neglected. Thus, the horizontal transfer of heat has no direct effect, either on the heat balance of the earth's surface, or on the heat balance of the air layer near the ground.

This statement does not contradict the fact that a considerable effect is exerted by the horizontal heat transfer on the heat balance of the earth's surface by changing the values of the components of the balance, such as: the radiational flux, the turbulent flux of heat, the expenditure of heat for evaporation, etc.

The effect of horizontal heat transfer in the hydrosphere on the heat balance could be explained in a similar way. In this case, the effect of horizontal transfer produces only some changes in the vertical flux of heat A', and also, in the other components of the balance.

In the equation of heat balance (1), the components of the balance representing the heat streams could be substituted by their sums for the period of time t. Then, we will obtain an equation which will coincide with equation (1):

$$R = LE + P + A, \tag{4}$$

where the values without primes show the sums of heat for the analyzed period of time.

The sum of the radiational flux of heat at the level of the earth's surface (be it positive or negative) is usually called - the radiation balance.

The radiation balance value is equal to the difference between the amount of radiation absorbed by the earth's surface and the amount of effective outgoing radiation,

$$R = (Q + q)(1 - \alpha) - I, \tag{5}$$

where: Q - is the sum of direct radiation, q - the sum of diffused radiation, α - albedo, I - the effective outgoing radiation (the difference between incoming and outgoing heat amounts on the earth's surface, which is determined by radiation from the earth's surface and counter radiation of the atmosphere).

Many authors have already pointed out, and rightly so, that the term "radiation balance" is not very good, since the word "balance" in this case does not have its usual meaning, and accounts for only one category of energy - the radiation energy. Even so, the application of the expression "radiation balance" is especially inconvenient to use in the study of heat balance, since these similar terms have an entirely different physical meaning (for instance, the radiation balance usually is not equal to zero, whereas the sum of the components of heat balance always equals zero, etc.); however, at the present time, it would be very difficult to

discard this term, since it has been so largely used in all hydrometeorological studies.

The amount of heat exchange between the active surface and the lower layers A can be determined by the other heat balance components of the upper layers of the lithosphere or hydrosphere. The heat balance equation for a vertical column with the upper limit on the earth's surface, and the base at a depth where the annual variations of temperature cease, will be:

$$A = B + F, \tag{6}$$

where A - is the heat exchange between the column and the active surface, F - the heat exchange between the column and the ambient space of the litho or hydrosphere in a horizontal direction, B - the change in heat amount inside the column during the given period of time, (fig. 2).

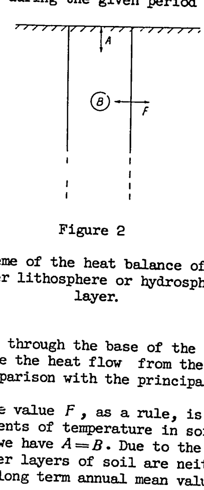

Figure 2

Scheme of the heat balance of the upper lithosphere or hydrosphere layer.

Vertical heat exchange through the base of the column could be assumed to be equal to zero, since the heat flow from the depth of the earth is usually negligible in comparison with the principal components of the heat balance.

In the lithosphere, the value F, as a rule, is insignificant, since the mean horizontal gradients of temperature in soil are very small. Therefore, for the land, we have $A = B$. Due to the fact that, on the average for the year, the upper layers of soil are neither warmed up nor cooled off, we must assume that long term annual mean values for the land are $A = B = 0$.

In the heat balance of more or less significant water reservoirs, analyzed as a whole body, the values of A are also very close to B, since the heat exchange between the reservoir and ground is usually insignificant in comparison with the principal components of heat balance.

However, for some portions of the oceans (lakes and seas), the values A and B might be very different, since in this case a redistribution of rather large quantities of heat in a horizontal direction might be affected by currents and by macroturbulent exchange.

Consequently, under such conditions, the mean annual value of heat exchange between the active surface and the lower layers will not equal zero but will be equal to the sum of heat which is gained or lost through currents and macroturbulence by the vertical column extending through the hydrosphere (i.e.,$A=F$).

Thus, the equation of heat balance for the land, for the mean annual period, will be:

$$R=LE+P, \qquad (7)$$

and for the ocean:

$$R=LE+P+F. \qquad (8)$$

In some cases, equations (7) and (8) could be simplified. Thus in deserts, where the amount of evaporation is close to zero, equation (7) will simply be:

$$R=P.$$

For the world's ocean as a whole, where the general redistribution of heat by sea currents, due to recompensation, is zero, equation (8) is transformed into:

$$R=LE+P.$$

Concluding our analysis of the problem of heat balance equations for the earth's surface it should be noted that when these equations are applied one must keep in mind that the conception of the "earth's surface" is somewhat conventional (sometimes it is called the "active surface," or the"underlying surface"). Actually, the "surface" processes of solar energy transformation are developed not on a two dimensional surface, but inside a layer of some thickness, as for example, when the processes of heat expenditure for evaporation on the mainland take place, or when water reservoirs are absorbing the solar energy, etc. The "active layer" reaches a considerable thickness in places with a high vegetation (especially so in forests).

However, even when dealing with an active layer of considerable thickness we can use the concept of the active surface and it will not lead us into any noticeable inaccuracies, especially in studies of the components of balance where longer periods of time are averaged. But in single cases (studying the rapid changes of the components, etc.) it would be more expedient to use the concept of an active layer instead of the active surface.

To derive a heat balance equation for the system earth-atmosphere (i.e., the balance for the whole outer geographical medium) we must analyze the gains and losses of heat energy in the vertical column that extends through the whole atmosphere and upper layers of the hydro and lithosphere down to those levels where seasonal and diurnal variations in temperature cease. In our computations we will use the sums of heat streams for a certain period of time t.

The exchange of heat between the column and the outer space will be characterized by its radiation balance R_s, which actually equals the difference between solar radiation absorbed by the whole column and the total long-wave outgoing radiation from this column during the analyzed

period of time (fig. 3). The radiation balance of the system earth- atmosphere may have a different sign, and we will consider the value R_s as being positive, when it shows that heat is gained by this system.

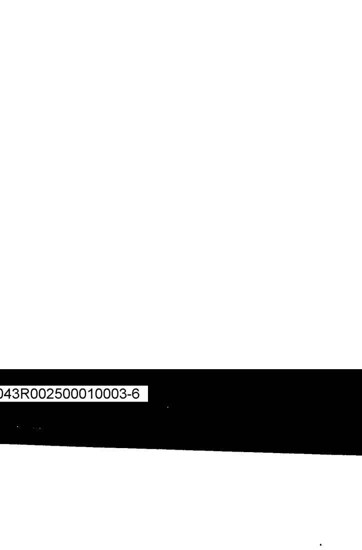

Figure 3

Scheme of the heat balance of the earth - atmosphere system.

Extending this column deeper into the lithosphere or hydrosphere down to those layers where the thermal regime is no longer affected by variations of meteorological factors, we may assume that the inflow of heat through the base of the column is practically absent, it equals zero.

The inflow of heat through the lateral surface of the column is determined by the horizontal transfer of heat in the atmosphere and hydrosphere. The difference between the gain and loss of heat due to the transfer of heat in the atmosphere is presented in fig. 3 by the arrow C, and the similar characteristic for the hydrosphere - by arrow F.

Besides the heat exchange that occurs through the surface or the column, there are some other factors that also affect the heat exchange, namely some sources of heat (positive or negative) located inside the column. Among them, a greater significance is attributed to the surplus or deficiency of heat, which is associated with the changes of water phases and especially by the process of evaporation and condensation.

The gain of heat from condensation processes in the atmosphere (the difference between the gain of heat from condensation of water vapor, and the loss of it for evaporation of water droplets in the atmosphere) over a sufficiently homogeneous surface is approximately equal to the product of the latent heat of evaporation L and the sum of precipitation r.[3] The expenditure of heat for evaporation (the difference between the loss of heat for evaporation from the surface of reservoirs, from vegetation, from soil, and the gain of heat from condensation on these objects) is equal to LE. The general influence of condensation and evaporation on the heat balance of the column could be approximately expressed by the value $L(r-E)$.

Among the other heat balance components of the column, the change in the heat content inside the column, that occurred during the period for which the values B_s have been summarized, should be taken into account. The other components of the balance (the gain of heat from dissipation of mechanical energy, the difference between the amount spent for ice melting and the gain from ice formation processes, the difference between the loss of heat for photosynthesis processes and the gain of it from oxidation of the organic matters, etc.) are usually insignificant in the heat balance of the earth-atmosphere system and could be omitted.

The equation of the heat balance of the earth-atmosphere system will be:

$$R_s = C + F + L(E-r) + B_s, \qquad (9)$$

in this instance, we will assume that all the terms on the right side are positive in case they show the expenditure of heat. For the average annual period the value B_s will, approximately, be close to zero, and equation (9) will be transformed into:

$$R_s = C + F + L(E-r). \qquad (10)$$

For the mainland this equation will have a simpler form:

$$R_s = C + L(E-r). \qquad (11)$$

Since, for the earth as a whole, $E=r$, and the horizontal transfer of heat in the atmosphere and hydrosphere in total is, approximately, equal to zero, the heat balance equation for the whole outer geographical medium will assume a simple form of:

$$R_s = 0. \qquad (12)$$

The radiation balance equation of the earth-atmosphere system R_s is similar to the radiation balance equation for the earth's surface (5):

$$R_s = Q_s(1-\alpha_s) - I_s, \qquad (13)$$

where Q_s — is the short-wave solar radiation received on the outer boundary of the atmosphere, α_s — albedo of the earth-atmosphere system, I_s — total long-wave outgoing radiation into outer space.

The heat balance equation of the atmosphere could be obtained by summing up the relevant heat streams, or simply, by taking the difference between the heat balance of the earth-atmosphere system and the earth's surface.

Assuming that the radiation balance of the atmosphere is:

$$R_a = R_s - R,$$

and the change in heat content of the atmosphere is:

$$B_a = B_s - B,$$

we will find, that

$$R_a = C - Lr - P + B_a, \qquad (14)$$

and for the average annual period

$$R_a = C - Lr - P. \qquad (15)$$

In many calculations, along with the heat balance equations, we will have to use the equation of water balance also.

The equation of water balance for the land surface is the expression of the condition when the algebraic sum of all forms of gain and loss of solid, fluid, and gaseous water received by a horizontal surface from the ambient space during certain time intervals is equal to zero.

This equation will be:

$$r = E + f_w + m, \qquad (16)$$

where r — is precipitation, E — the difference between the evaporation and condensation on the earth's surface (usually called just evaporation), f_w — the surface runoff, m — moisture exchange between the earth's surface and the lower layers. The value m - is the algebraic sum of the gravitational flux of fluid moisture from the soil surface into the deeper layers, the vertical flux of the film-moisture between soil layers of various moistening, the vertical flux of water vapor, flux of water which is raised by the roots of plants, etc., obtained for the analyzed period of time.

Equation (16) is more often used in some modified form, which could be derived by considering the fact that the vertical flux of moisture m is equal to the sum of the underground runoff f_p and the change of water content in the upper layers of the lithosphere b (this equality corresponds to the equation of water balance of the vertical column that extends through the upper layers of the lithosphere down to depths where moisture exchange is practically absent).

Keeping in mind the fact that the sum of the surface runoff f_w and underground runoff f_p is equal to the full runoff f, we find:

$$r = E + f + b. \qquad (17)$$

Equation (17) can also be used for calculating the water balance in water reservoirs or in some portions of them. In this case, the value f will characterize the redistribution of water in a horizontal direction for the analyzed period in the reservoir itself and in the lower layers of the ground (if there is any noticeable redistribution of moisture). Similarly, value b, if taken for the reservoir, would have to determine the total

3) The gain of heat from condensation in the atmosphere is equal to the difference between the gain and loss of heat which is associated with condensation and evaporation of water drops in clouds and fogs. Upon a more or less homogeneous surface the difference between the sums of condensation and evaporation in the atmosphere, obtained as the averages from long periods, is equal to the amount of precipitation. But this might not hold for regions of a dissected terrain and also for some short time intervals.

change in water quantity in the reservoir itself, and in the lower layers as well, if there was a noticeable change in moisture content. Practically, in many cases value b is determined for the water reservoirs by the change of water level. For the average annual value b is often very small, and therefore, the equation of the water balance will be:

$$r = E + f. \tag{18}$$

For the whole globe the horizontal redistribution of moisture equals zero, and therefore the equation of water balance will have the simple form:

$$r = E. \tag{19}$$

The equation of water balance has the same form, for the year, in those portions of land where no runoff is observed, including deserts.

In closing, we will give the equation of water balance in the atmosphere. By adding up all categories of the gain and loss of moisture in the vertical column which extends through the atmosphere, we offer the following equation:

$$E = r + C + b_a, \tag{20}$$

where C — is the amount of moisture which is gained or lost by the vertical column as effected by air currents and by the horizontal turbulent exchange; b_a — is the change in water content of the column during the analyzed period of time.

Since the atmosphere can contain only relatively small quantities of water in any of its phases, the value of b_a is usually much smaller than the other components of the balance. The average annual value is close to zero.

The equations of heat and water balance, which have been given here actually represent the basis for all constructions and derivations outlined in this monograph.

§ 2. General review of investigations on heat balance of the earth's surface

The formulation of the heat balance investigations' problem belongs to the outstanding climatologist and geographer, A.I. Voeĭkov. In recent literature we frequently find the citation of this remarkably deep statement by A.I. Voeĭkov, which concludes the first chapter of his monograph - The Climates of the Terrestrial Globe, (Voeĭkov, 1884 [68]): "*I think that one of the most important problems of physical sciences at the present time is - the bookkeeping of solar heat amounts received by the earth with its gaseous and fluid envelopments.* [4]

"*We have to know: how much of solar heat is received at the upper boundaries of the atmosphere, how much is spent, for heating of the atmosphere,*

4) Italics by A.I. Voeĭkov.

for the water vapor change of state contained in the atmosphere; and further on, what is the quantity that reaches the land surface or water surface, what quantity is spent for heating of various bodies, and what is spent for the change of their state (from the solid phase into fluid, and from fluid into gaseous), for the chemical reactions, especially those connected with organic life; further on, we have to know how much heat is spent by the earth through radiation into outer space, and how it happens; i.e., how much of it occurs on account of a decrease in temperature and how much on account of changes in the state of bodies, especially of water.

The difficulties in achieving this goal cannot scare the scholars, who understand the wide problems of science. It cannot be done in one century. Therefore, I assume it is useful to state this very comprehensive problem in full size, hiding nothing of the huge difficulties involved in obtaining not only the complete solution but even in finding some approximate answer."

To estimate the importance of these thoughts of A.I. Voeĭkov, it must be remembered, that in those times when The Climates of the Terrestrial Globe was being prepared, (the eighteen eighties) the problem of the solar energy transformations occurring in the outer geographical medium was almost completely untouched. More or less systematic actinometric observations were started only at the end of the 19th century, and the earliest attempts of calculating the incoming solar energy to the earth's surface were not known by Voeĭkov when he was writing this monograph. [5] Nevertheless, Voeĭkov has not only correctly formulated the main problems in the study of heat balance, but with a scientific optimism, so typical of him, he predicted with confidence that the huge difficulties in solving these problems would be successfully overcome.

In many investigations done by A.I. Voeĭkov, various concrete questions associated with the study of heat balance have been analyzed. For instance, in The Climates of the Terrestrial Globe ... Voeĭkov has paid much attention to the calculation of the annual variations in heat content of lakes. These calculations permitted him to draw some conclusions about the influence of water reservoirs on climatic conditions in various regions. In his work The Heat Exchange in the Outer Coating of the Terrestrial Globe (1904 [69]) the question about the climate-forming effect of the heat exchange in soil and water reservoirs is analyzed in detail. Many ideas of Voeĭkov, outlined in this work, have not lost their scientific importance, even up to now. It is enough to remember, for one thing, Voeĭkov's statement about the exceptionally important conception of the outer active surface and his deep analysis of the relationship between the heat-exchange and the annual and diurnal variations in temperature.

However, it is obvious that any, more or less comprehensive, investigations of the heat balance on the earth's surface could be started only

5) The work of Angot, 1883, the first investigation of the laws governing the incoming solar energy to the earth's surface could be utilized by A.I. Voeĭkov only when he was preparing the second edition of The Climates of the Terrestrial Globe published in 1887.

after an effective method for determining its principal components had been suggested.

The development of methods for determining the components of the balance has been started in two principal directions: 1) designing of special instruments for measuring the separate components of the balance, and 2) development of methods for calculating the components proceeding from theoretical conceptions and using ample data of regular hydrometeorological observations.

The first stage in the development of methods for determining the components of the balance, based on application of special instruments, was closely connected with the development of actinometric investigations.

Works of our scientists were of great importance in creating the scientific actinometry. So, in particular, a great advance in development of actinometric observations was made when O.D. KHvolson invented his actinometer. Systematic observations of direct solar radiation were started with this instrument at Pavlovsk in 1890. The possibilities for measuring the solar short-wave radiation were greatly expanded when K.D. Ångström invented the pyrheliometer in 1895; V.A. Mikhel'son designed a perfect actinometer in 1906; in 1910-1911 S.I. Savinov improved the actinograph of Crova (Kalitin, 1950 /122/) and a series of other actinometric instruments were developed. Much later, at the end of the nineteen thirties, the more or less accurate measurements of scattered radiation could be secured after the IU.D. IAnishevskii pyranometer was introduced in 1934. (It is described in IAnishevskii's paper 1951 /248/).

The progress achieved in improvements of actinometric instruments contributed to a rather rapid expansion of the actinometric network observing the flux of short-wave radiation.

Before the Great October Social Revolution in Russia, actinometric observations were being taken only at five points, but afterwards, at the end of the nineteen thirties, these observations were carried on at several scores of stations (Kalitin, 1947 /121/). The actinometric network of the world has grown especially fast during the last decade.

The development of the actinometric network could be characterized by comparison of the consolidated tables of actinometric observations that have been published annually. So, for instance, the paper by Kimball that was published in 1927 and 1930, contains the mean values of total radiation obtained at only 32 stations, located in various regions of the world. In Gorczinsky's paper of 1945, observations of 58 stations are given. In the first summarizing paper by T.G. Berliand (1949 /267/) the mean values of total radiation are given for 85 stations, and in the second paper (Berliand, 1954 /287/) - for 139 stations. Even though, the last number is many times less than the number of meteorological stations taking ordinary observations, nevertheless, at the present time, a direct generalization of available data on actinometric observations could be done and could be utilized for obtaining some climatological conclusions.

However, it must be stated, that up to the present time, actinometric observations for most of the stations have been limited to measurements of the flux of short-wave radiation.

The measurements of long-wave radiation, and especially that of the effective outgoing radiation, were associated with considerable methodical difficulties and therefore they were started much later than the measurements of short-wave radiation.

The first instrument that was used for more or less systematic measurements of the effective outgoing radiation, was the pyrgeometer of K. Ångström designed in 1905 (Ångström, 1916). Later on, many investigations established that the observations made with this instrument usually contained fairly large errors. Many authors have tried to improve the construction of instruments measuring the effective outgoing radiation, but up to recent times, the results have not been quite satisfactory. Only recently, some instruments have been constructed, although not without defects, but still permitting measurements of effective outgoing radiation, without any great principal errors, and not only at night but also during daylight hours. The existing instruments for measuring long-wave radiation were used for observations by stations and by expeditions, but results obtained are still too scarce in volume to warrant any climatological generalizations. Still, these data are valuable in verifying various calculation methods for determining the effective outgoing radiation. More details on this are given on pp. 43-46(pp.40-44 of this translation).

More or less reliable instruments for direct measurements of radiation balance were constructed not very long ago. The first steps in this direction were made in the nineteen twenties by V.A. Mikhel'son, and later I.G. Liutershtein and A.A. Skvortsov. Later, IU.D. IAnishevskii and F. Albrecht devoted much effort for the development of a balance-counter. As a result of long years of investigations, IU.D. IAnishevskii has designed a rather simple construction of a balance-meter, which permits the measurement of radiation balance values without any large principal errors (IAnishevskii, 1949 /247/). During recent years, Albrecht suggested several improvements of this instrument (Albrecht 1933 and others). Most recently a new construction of a balance-meter has been suggested by D.L. Laikhtman and N.B. Kucherov (1952 /155 & 151/). We will not cite further the constructions of balance-meters, but will mention the fact that instruments for direct measurement of radiation balance are now being extensively used by expeditions investigating the amelioration of climate in the lower layers of the atmosphere, and for other problems in physics of the air layer near the ground. For stationary observations, the application of the balance-meters has not been very wide as yet. Consequently, the observational material obtained through balance-meters is, at the present time, very scarce, and at any rate, it is considerably smaller than the amount of observations on the effective outgoing radiation. For this reason, the data of direct radiation balance measurements could be used mainly for investigation of physicometeorological regularities and in solving various methodical problems. The application of these data for comparison of climatic conditions in various regions still involves considerable difficulties.

Methods for direct measurements of other components of the heat balance of the earth's surface, and especially of the expenditures of heat for evaporation and for turbulent heat exchange, are much less developed than those for measurements of the components of radiation balance.

Since the latent heat of evaporation represents a well known physical value, there is always a possibility to find the amount of heat lost on evaporation by measuring the evaporation from the earth's surface. The instruments for determining evaporation from land surfaces (evaporimeters of various makes) were being developed by numerous authors during a long period of time. Some of the soil evaporimeters have been applied for observation by some hydrometeorological stations. 6)

Data on evaporation obtained by evaporimeters are relatively scarce and, as it has been pointed out by many authors, they are not free from considerable systematic errors. Therefore, the soil evaporimeters could not be accepted as a universal method for determining the evaporation from the soil surface. Various kinds of evaporimeters have been used to determine the evaporation from water surfaces. However, these observations also contain some errors, and are insufficient to warrant any larger climatological generalizations. 7)

In recent years, many authors have used gradient methods for determining evaporation and the loss of heat associated with it.

These methods involve the calculating of evaporation proceeding from measurements of vertical moisture gradients, and simultaneously taking into account the values of the coefficient of turbulent exchange. Another variation of the gradient method for determining evaporation is the so called balance-method; this determines the evaporation or loss of heat for evaporation by measuring the vertical gradients of temperature and humidity in the air layer near the ground, and by measuring the radiation balance and heat exchange in soil.

The gradient method also permits us to determine the magnitude of the turbulent heat flux - one of the most difficult components of the heat balance for direct measurement.

Numerous works with gradient methods made in the USSR (Budyko, 1946a [35],1948a [39]; Timofeev, 1951 [219]; Budyko and Timofeev, 1952 [59]; Methodical Instructions, edited by Rusin, 1954 [171], and many others) and also abroad (Thornthwaite and Holzman, 1942; Holzman, 1943, and many others) have corroborated the great importance of these methods. However, the observations needed for determining the components of heat balance by gradient methods are still very scarce; since the network of hydrometeorological stations has not been taking them as yet.

Among other methods for a direct determination of heat balance components, the idea of B.A. Aizenshtat (1948, 1951 [2 & 3] and others) should be pointed out. This author suggested designs of several instruments which measure the components of the heat balance (including heat exchange between the active surface and atmosphere) by the method of compensation.

Aizenshtat's instruments have been used by several expeditions, and interesting results were obtained. But regrettably, the Aizenshtat method is best fitted for determining the components of the balance of the active surface without vegetation. Quite recently, a new instrument for measuring turbulent heat flow has been suggested by N.V. Kucherov (1952 [151]). In many respects it is similar to those suggested by B.A. Aizenshtat.

The great progress in experimental meteorology has made it possible, at the present time, to measure all the principal components of heat balance in various physiographical regions. However, these observations are usually insufficient for making any larger climatological generalizations, since these special observations were taken mainly during the course of some special investigations and are not included in the program of observations taken by the hydrometeorological network except for short-wave radiation measurements.

Therefore, at the present time, methods for determining the components of heat balance from data in standard meteorological observations are of a great importance.

The first calculations of the components of heat balance determined the changes of heat content in limited water reservoirs and in the upper layers of soil. Similar calculations, which are relatively simple, have been made in the last century by A.I. Voeikov, Ferrel and others. Among the works in this direction, the investigations by Homen (1897) must be pointed out. He was the first to compare the diurnal heat exchange of a granite rock with those of a peat meadow and of a sandy soil. The results obtained by Homen have been cited many times in various meteorological textbooks.

The first fundamental investigations concerning the calculations of the transformation of solar energy in the atmosphere were published at the end of the 19th century. They include the above mentioned investigation by Angot on determining the amount of short-wave radiation which reaches the earth's surface in various latitudinal zones of the globe.

However, the first calculations of the components of heat balance on the earth's surface have only been accomplished in the first few years of the 20th century. Of great importance, for the investigation of heat balance, was the work of W. Schmidt (1915). Using the calculation methods, he determined the mean annual values of heat balance components for the latitudinal zones of the ocean in the Northern and Eastern hemispheres, including calculation of the mean quantities of heat, for each latitude, which are transferred inside the ocean in a horizontal direction as a result of sea currents and macroturbulence. Though the calculation methods used by Schmidt were rather rough, (especially that for determining the loss of heat for evaporation and turbulent exchange) nevertheless he managed to obtain a proper order of magnitude for the principal components of heat balance.

It must be mentioned here that, Schmidt was the first who tied up the calculations of the components of heat balance with determination of water balance in the world oceans.

Among all subsequent investigations of heat balance it is proper to mention the works of A. Ångström. In a paper published in 1920, Ångström determined all the components of heat balance for a limited reservoir - the Lake Vassi Jaure in Sweden. Simultaneously, he essentially improved

6) So in some periods, the evaporimeter designed by M.A. Rykachev (1898 [199] was used by several stations. Data of observations obtained by these instruments and also by some other evaporimeters have been subsequently published, (Data of observation on evaporation ... 1939 and others).

7) Summaries of observations with water evaporimeters and evaporation reservoirs are available in papers published by B.D. Zaikov (1949 [102]), Follansby (1933) and others.

the methods for calculating the components of radiation and heat balances, though the problem of calculating the amounts of heat lost for evaporation and turbulent exchange has not been successfully solved by Ångström without using a rather conventional hypothesis.

In 1925, Ångström published a paper (1925b) containing the results of calculations of all components of heat balance in the Stockholm region for all months of the year. The principal defect of this, in many respects a very valuable work, was the neglect of the use of reflected radiation in determining the radiation balance for the warm season. Annual and monthly values of radiation balance without this principal error have been calculated for the first time by Savinov for Pavlovsk. These calculations are published in the Course of Geophysics by P.N. Tverskoi (1934 /216/).

For development of investigations on heat balance of the seas, the paper contributed by V.V. SHuleikin (1935 /240/) was of great importance. In this work, for the first time, the components of radiation balance of a single sea (the Kara Sea) were calculated by using the results of special observations and of some calculations.

Having determined the components of the heat balance, SHuleikin also demonstrated the great importance of the warm current affecting the thermal regime of the Kara Sea. This conclusion was confirmed later by direct observations (SHuleikin, 1941 /241/).

When these works were published, many authors started investigations of heat balance, and the amount of calculations of the radiation and heat balance at various points on land and for water reservoirs grew rapidly.

In the work by F. Albrecht (1940) the values of the components of radiation and heat balance were determined for 12 points, 6 of these were located in various regions on land, 5 in various regions of the ocean and 1 on a small lake. Using largely the calculation methods for determining the components of the balances, Albrecht also worked up some data from special observations. Along with the values of the components of balances for a month and for an annual period, Albrecht has obtained some data (though very limited) on diurnal variations of the components of radiation and heat balance.

Attention must be paid especially to some conclusions in this paper concerning the interrelations between the climatic conditions and the regime of heat balance components.

Among the works done for determining the components of heat balance for single points on land, the investigation by S.A. Sapozhnikova (1948b /207/) should be considered.

S. A. Sapozhnikova accomplished the calculations of the annual and seasonal values of heat balance components for 8 points, in various geographical zones of the USSR. The analysis of these materials on heat balance has enabled Sapozhnikova to explain some physicogeographical regularities (for instance, the factors determining the northern boundary of the forest zone).

Calculations of monthly values of radiation balance at some points in the Lower Volga region have been carried out by B.M. Gal'perin (1949b /76/). It must be also mentioned that the calculations of radiation balance for some regions in the Arctic have been accomplished by A.S. Kaledkina (1939 /116/)and R.N. SHpakovskaia (1940 /239/). A detailed investigation of the

radiation regime and radiation balance in the Moscow region has been published by M.S. Averkiev (1947 /1/). There are also some works containing the results of calculations of the components of radiation and heat balance for a series of points in Western Siberia, (Orlova, 1954 /182/) for the Yakutsk region (Gavrilova, 1954 /72/)and some other points in the USSR.

The calculations of radiation and heat balances of various water bodies are methodically much simpler than the computations of heat balance for the land, and they have been very much in use for the last 20-30 years.

In developing these investigations it was important that, in the calculations of the relationship between the loss of heat for evaporation and turbulent heat exchange, the so called "Bowen relationship" could be used. This relationship ties up the evaporation and turbulent heat exchange on one side with the difference between the temperature of the water surface and air, and the corresponding difference in specific humidity, on the other. The application of this formula /by Cummings and Richardson, (1927), and others/ has greatly facilitated the determination of heat balance components from data of ordinary meteorological observations.

Among the works on radiation balance and on the components of heat balance for water reservoirs, the investigations of the following authors, on the heat balance of various seas, must be mentioned: KH.K. Ulanov - for the Black Sea (1938 /222/);O. Mertsalova - for the Barents Sea (1938 /170/);V.V.Timonov and P.P. Kuz'min - for the White Sea (1939 /217/); N.T. CHernigovskii - for the Arctic Seas (1940a, 1940b /230 & 231/); L.F. Rudovits (1927 /197/); and I.A. Benashvili (1941 /20/) - for the Caspian Sea; B.D. Zaikov - for the Aral Sea (1946 /101/); B.A. SHliamin - for the Azov Sea (1947 /238/); A.F. SHishko - for the White Sea (1948 /237/); N.I. Egorov - for the Red Sea (1950 /98/) and others.

In modern literature on the heat balance there are also some investigations concerning the heat balance of the ocean's surface. Mosby (1936) has compiled the radiation balance for the latitudinal zones of the world oceans, and determined the amount of evaporation from the oceans. At the beginning of the nineteen forties the first attempt was made to construct maps for the components of heat balance in some regions of the world oceans. Jacobs (1943) and Sverdrup (1945) designed schematic maps of the components of heat balance for the northern portions of the Atlantic and Pacific Oceans. Albrecht (1949, 1951) computed the components of heat balance for the Pacific and Indian Oceans and constructed a series of maps showing the distribution of the balance components for single months and for the year.

The computations of the components of water balance for the region of the Gulf Stream were published by Konoplev (1953 /138/). Sauberer and Dirmhirn (1954) made computations and constructed maps of radiation balance for the oceans of the Northern Hemisphere for four single months - March, June, September, and December.

The heat balance of lakes and evaporation basins has been studied by Cummings and Richardson (1927), Richardson (1931), Cummings (1936), L.N. Demchenko (1952 /90/),Sauberer (1953) and others. Investigations of the heat balance in artificial reservoirs have been carried out by A.P. Braslavskii and Z.A. Vikulina (1954 /33/).

Some works, which have contributed some improvements to the calculation methods for determining the components of heat balance, must be mentioned here too. Among those works are the investigations by A. Ångström (1922) and S.I. Savinov (1933 [202 &203]) dealing with methods for calculating the total short-wave radiation. Methods for calculating the long-wave radiation that first were suggested by Ångström (1916) and Brunt, (1934) were developed later in the theoretical investigations by K. ÎA. Kondrat'ev (1949a, 1949b [133 & 134] etc.), by M.E. Berliand and T.G. Berliand (1952 [24]), by T.V. Kirillova (1951 [126]), in an experimental work by Bolz and Falkenberg (1949) and in a series of other investigations.

During the last 10-15 years the calculations and experimental investigations of heat balance on the earth's surface have been widely extended, stimulated by rapidly growing demands and requests for this type of data to meet the needs of all hydrometeorological sciences.

In meteorological researches the data on heat balance are now used in calculating the transformation of air masses. Works along these lines include the investigation of V.G. Kastrov (1938 [124]) dealing with the physical mechanism of drought development, and also some works by M.E. Berliand (1952, 1953 etc. [22 & 23]) and M.V. Zavarina (1953 [100]) in which the calculations of the transformation processes are associated with development of methods for the forecasting of thermal regimes.

Among the agrometeorological investigations in which data on heat balance have been utilized are the works by Sapozhnikova. In one of her researches (1948a [206]) the calculations of heat balance for a cultivated field made it possible to obtain valuable practical conclusions about the reasons that prevent the extension of grain crops in a northerly direction.

In works accomplished under the supervision of S.A. Sapozhnikova (as, for instance - Climatological data for the region between the Volga and Ural rivers, 1951), and also in investigations of V.V. Orlova (1954 [182]) the data on heat balance are used in explaining the laws governing the agrometeorological regime in various regions of the USSR.

The most recent investigations of amelioration of climate have also been based mainly on heat balance data. For instance, in works by M.I. ÎUdin (Budyko, Drozdov, L'vovich, Pogosian, Sapozhnikova and ÎUdin, 1952 [56]), D.L. Laĭkhtman (1953 [156]) and other authors, the data on heat balance are used for calculating changes produced by irrigation in the hydrometeorological regime.

The calculations of heat balance components have also been used by M.I. ÎUdin and by the author in works dealing with methods for evaluating the hydrometeorological effect of the forest shelter belts (Budyko and ÎUdin, 1951, 1952 [62 & 63]) and in other works by the author which will be mentioned in chapter V.

It must be noted, that considerable progress in the study of hydrometeorological effectiveness of ameliorative measures, made on the basis of heat balance investigations, has been achieved by the complex expeditions sponsored by the Central Geophysical Observatory and, particularly, by the expedition to Kamennaîa Step' in 1951 (headed by Prof. Drozdov) and by the expedition to the irrigated oasis of Pakhta-Aral in 1952 (headed by Prof. D.L. Laĭkhtman). Results of observations gathered by these expeditions (Trudy GGO, no. 39, 1953 and no. 40, 1953) and stationary observations organized and carried out by special agrometeorological stations have shown that the accounting of the heat balance regime is of paramount importance for estimating the effectiveness of the ameliorative measures.

Accordingly, in numerous modern investigations, the data on heat balance have been used in selecting the most effective construction of field protective shelter belts; in estimating the effect of irrigation on the climate of the layer near the ground; in studying the dependence of water amounts, needed for irrigation, upon weather factors; etc.

Among the theoretical problems in modern meteorology that could be solved by using data on heat balance of the earth's surface, we will point out the works on the theory of climate (L.R. Rakipova, 1952, 1953, [194 & 195] etc.).

The utilization of data on heat balance permitted us to verify the initial hypotheses and improve some of the principal statements in the theory of climate origin.

Also, heat balance data are used in solving many questions of dynamic meteorology, where the equation of heat balance is applied as showing the boundary conditions. Among these investigations are works on the theory of diurnal variations of meteorological elements, on the theory of local circulations, etc. Also, data on heat balance are now used extensively in investigations of land and sea hydrology as well. In particular, the computations of heat balance are now one of the principal methods for forecasting the snow melt regime. A series of important investigations of this problem has been made by P.P. Kuz'min (1947, 1950, 1951 [147, 149 & 150] etc.). Among other methods of hydrological forecasting that are based mainly on data of heat balance of water reservoirs, the forecasting of the thermal regime in reservoirs including the forecast of freezing and melting of water surfaces and the process of ice melting should be mentioned.

Data on heat balance are also used in investigations of the hydrological regime of swamps, including computations of the amounts of evaporation from swamps; in calculations of the amounts of evaporation and runoff for various regions; and in studies of climatic conditions of moistening. Works along this latter line will be treated in more detail in chapter IV.

Calculations of heat balance are also of great importance in determining the amounts of evaporation from the existing and from planned reservoirs. This problem, elaborated in a particular case by A.P. Braslavskiĭ and Z.A. Vikulina (1954 [33]), is of a great practical importance.

The heat balance calculations of the existing reservoirs that have been made in order to allow an evaluation of changes in the hydrometeorological regime, when accomplishing larger ameliorative measures, must also be mentioned here. These works are directly associated with investigations of the heat balance of the seas and oceans which, primarily have been made for clarifying the regularities in the hydrometeorological regime of the analyzed reservoirs. In developing these investigations the works of V.V. SHuleĭkin were of great importance.

Data on heat balance have been used not only in hydrometeorological studies but also in investigations of general problems in physical geography. It is obvious, that in an investigation of the mechanism of natural processes and their interaction, it would be very important to have certain concepts about the geographical regularities in transformation of solar radiation on the earth's surface, which actually is the main basis and source of energy for all natural exogenous processes. Accordingly, A.A. Grigor'ev has established many relationships connecting the characteristics of radiation and heat balance of the earth's surface and atmosphere with the intensity of major physicogeographical processes (his papers of 1937, 1946, 1948, 1951, 1954 [80, 81, 82, 83 & 84] etc.). Among the studies in this direction we will also note the investigations done by D.L. Armand (1949, 1950 [13 & 14]) and by author (these will be reviewed later).

A rapid growth of requests for data on heat balance stimulated a considerable expansion of climatological investigations concerning the distribution of the components of radiation and heat balance.

Up to the middle forties, the climatological regularities of the components of heat balance had not been sufficiently studied. Data on the mean values of the components of heat balance on land were available in literature for only a few points, and the accuracy of their calculation was completely unknown. For water reservoirs, there was more data available (calculations for some seas and lakes and schematic maps for some portions of the ocean surface); however, for the oceans, there were no world maps of the distribution of balance components, and the degree of accuracy of accomplished calculations was oftentimes debatable.

Taking all this into account, a group of scientists in the Central Geophysical Observatory has undertaken detailed studies of the climatological regularities of radiation and heat balance. In these studies, attention has been paid to development of methods for an independent determination of all components of heat balance, which would permit an objective verification of the degree of accuracy of the accomplished computations of the balance equation. This question was first solved for single point conditions (Budyko, 1946c [37]), and later for a large land region--the southern areas of the European USSR (Budyko 1947 [38]). Having confirmed, in this way, a sufficient reliability of developed methods for determining the components of the balance, the authors was able, for the first time, to construct maps of the distribution of heat balance components for the mainland's surface.

In ensuing works of the Central Observatory many other maps have been designed: seasonal and annual maps of the components of heat balance for the European part of the USSR (T.G. Berliand, 1948 [25]); annual maps of the balance components for the extratropical portion of the northern hemisphere (T.G. Berliand, 1949 [26]); annual maps of radiation balance for West Europe and for the eastern part of North America (Zubenok, 1949a [105]); and some calculations of the latitudinal values of the components in the Northern Hemisphere have been accomplished (Budyko, 1949b [42]).

Having made all these investigations, which were to a certain extent of a preparatory nature, the authors, T.G. Berliand, and L.I. Zubenok have carried out a work on designing the world maps of the balance components for single months and for the year. This series of maps contains indices of the total short-wave radiation, of radiation balance on the earth's surface, on evaporation and heat losses for evaporation, on turbulent heat exchange between the underlying surface and the atmosphere, and also, a map (for the year) of the heat amounts gained or lost by the ocean surface as affected by the sea currents. In total, this series contains 66 maps which show the general geographical regularities in the transformation of solar energy on the earth's surface.

Annual maps of this series have been published in the Marine Atlas, Vol. 2 (Budyko, Berliand, Zubenok, 1953 [50]) together with the explanatory text in an article (Budyko, Berliand, Zubenok, 1954a [51]). The whole series of maps has been published in 1955 in The Atlas of Heat Balance [15].

Since the world maps of heat balance are necessarily of a schematic nature, more detailed maps have been designed by T.G. Berliand and M.A. Efimova (1955 [29]) for the USSR. They show the total radiation and radiation balance (monthly and annual). The latter series contains a total of 26 maps.

In addition, detailed maps of evaporation and amount of heat spent for evaporation have been constructed in the Central Observatory for a portion of the USSR for single months.

Along with this work, the calculations of diurnal variations of the radiation balance components in various climatic zones (Biriukova, 1955 [31]) have been made, and some features in the variations of the components have been studied.

Evaluating the whole scope of data on the climatography of heat balance that have been obtained during the recent years, we can say that, at the present time, the knowledge of the distribution of some balance components in time and space is not less than that about the basic meteorological elements (for instance, we now have world maps of total radiation and radiation balance for all months of the year, whereas no similar maps exist for such important meteorological elements as precipitation). However, it must be noted that the majority of heat balance maps are of a schematic nature, and their reliability is not always adequate, especially for regions with scarce data on the hydrometeorological regime - in higher latitudes, in the oceans of the southern hemisphere, etc.

When these new data on the climatography of heat balance were obtained they permitted a considerable expansion for the study of climatological regularities in heat balance and provided new conclusions of a general nature. Thus, these data, in conjunction with some special studies of heat exchange in the atmosphere, provided a basis to refute the earlier ideas that the atmosphere, by and large, transfers the heat to the earth's surface through a trubulent heat exchange (Budyko and IUdin, 1946, 1948 [60 & 61]). The accomplished calculations of heat balance have promoted, for the first time, a comparison of quantities of heat carried from the lower to the upper latitudes in the atmosphere and hydrosphere (Budyko, 1949b [42]).

Without the mention of other climatological regularities, which have been found by analyzing maps of heat balance components, we will note only that, these data have been used in works on the theory of climate mentioned above; in investigations of the general circulation of the atmosphere (Usmanov, 1953 [224] etc.); in works on description of climates in various regions (as, for instance, in works of Fedorov and Baranov, 1949; /225/ Orlova, 1954 /182/, and others); in synoptic climatology; and also in many investigations of the hydrometeorological effects of ameliorative measures, which have already been mentioned above.

The new data on the climatograpy of heat balance have been used extensively in various physicogeographical investigations. Among these are recent works by A.A. Grigor'ev (1951, 1954 /83 & 84/) on the theory of geographical zonality, and also some works by the author (1948b, 1949a, 1950a, 1951b /40, 41, 43, & 46/ and others).

In these works the author has tried to apply the data of the energy balance to the study of physicogeographical regularities by a deductive method, on the basis of general physical laws. This method combined with utilization of empirical data and generalizations, greatly expands the possibilities of geographical investigations.

The utilization of data on heat balance has made it possible to establish certain regularities and relationships that exist between the climatic factors on one side, and the hydrometeorological regime, the geobotanical zonality, soil zones, and some indices of the productiveness of natural vegetation on the other.

The development of investigations in this direction reflects the recent tendency to tie up more closely the sphere of physical geography with geophysics, on the basis of more extensive use of the quantitative methods and physical methods of analysis in tackling the problems of physical geography. It is conceivable that, a combined use of the physicogeographical and geophysical method of investigation will expand the possibilities of solving many practical and theoretical problems, related to physical geography as well as to geophysics.

Chapter II

Methods for the Climatological Calculation of the Components of Heat Balance

In the preceding chapter it is stated that the existing data of direct observations of the components of heat balance are very scarce and, as a rule, are not sufficient to warrant more extensive climatological generalizations. Therefore, the studies of spatial distribution of the balance components are based, at the present time, principally on indirect calculation methods, making use of ordinary meteorological observations of temperature, humidity, cloud amounts, precipitation, wind, soil and water temperatures, etc.

The methods of climatological calculations of the heat balance components may be more or less complicated, depending on what kind of meteorological data could be used in these calculations (whether, for instance, there are available data on clouds for various heights or only on total amounts, etc.). On the other hand, the degree of details in the method used must also depend on the nature of the problem to be tackled - so, a calculation of schematic maps of some mean values of the heat balance components for the land and oceans could be accomplished by using a less differentiated method, as compared with that used in calculating these data for small regions in microclimatic investigations, or for separate short periods of time.

The methods for climatological calculations of the balance components for the long period averages (long-period mean values, monthly and annual) have already been developed to a more or less higher degree by now. The problem of developing methods for calculating the components of heat balance for shorter periods, by using the basic meteorological observations, has not been solved in its entire scope as yet.

§ 3. Radiation Balance

In climatological calculations of radiation balance, its value is usually determined by formula (5) as the difference between the absorbed radiation $(Q+q)$ $(1-\alpha)$ (where Q — is the direct radiation, q —the scattered radiation, α — the albedo) and the effective outgoing radiation I. When using this method for calculating the radiation balance, the amount of the total radiation $(Q+q)$ must be determined first.

As has been said in chapter I, only total radiation is the one single component of the radiation and heat balance which is measured by comparatively numerous actinometric stations (at present, totaling about 200 stations on all continents).

However, it must be indicated, that the network measuring the total radiation is of a very irregular density. Most of the stations are located in the European part of the USSR and in North America. Due to this fact, N.N. Kalitin was able to construct a map of annual amounts of total radiation for the European USSR as early as 1945, Kalitin (1945 /120/). A little later, Hand constructed a similar map for the United States, Hand (1953).

In other regions the net of actinometric stations is very sparse and does not permit any climatological generalization or any, more or less, detailed construction of maps for total radiation. On the oceans there are practically no observations of this type of any, more or less, systematic nature.

Therefore, it is impossible to make a comprehensive investigation of the distribution of total radiation over the world without using the methods of climatological calculations.

The first investigations in calculating the amounts of short-wave radiation received by the earth's surface were confined to the determination of the direct radiation only. In the work by Angot, cited above, and also in ensuing investigations by S.I. Savinov (1925, 1928 /200 & 201/), M. Milankovich (1939 /173/), V.G. Kastrov (1928 /123/), B.M. Gal'perin (1949a /75/) and other authors, certain methods have been developed for calculating the amounts of direct radiation received by the earth's surface as dependent on the degree of transparency of the atmosphere.

Later some attempts were made to also determine the amount of diffused radiation by the theoretical or empirical methods, and to evaluate the effect of cloudiness on the total radiation.

The results of most of these investigations have not been used very much for calculating total radiation because of the cumbersome formulas, their insufficient accuracy, and the necessity to take into account many parameters that are quite variable and have not been studied sufficiently.

A simple and sufficient method for determining the total radiation has been suggested by A. Ångström (1922), Kimball (1928) and Savinov (1933a, 1933b /202 & 203/).

In an investigation published in 1922 Ångström suggested the following formula for determining the total radiation:

$$(Q+q)=(Q+q)_0\,[k+(1-k)\,S]\,, \qquad (21)$$

where $(Q+q)$ and $(Q+q)_0$ — the total radiation with natural conditions and a clear sky (no clouds), S — the ratio of observed sunshine hours to the possible amount for the given period, k — a coefficient, determining what portion of the possible radiation consists of actual radiation with overcast sky conditions. Using the observational data from Stockholm, Ångström found that coefficient k was equal to 0.235.

Kimball (1928) found a similar relationship using the data of several American stations:

$$(Q+q)=(Q+q)_0\,[0{,}29+0{,}71\,(1-n)]\,, \qquad (22)$$

where n — is the mean cloud amount in tenths.

S.I. Savinov (1933a, 1933b /202 & 203/) investigated in detail the interrelations between the values S and $(1-n)$, using observations taken in Pavlovsk, and he found that these values usually differed considerably from each other. Savinov came to the conclusion that the best agreement with the true value of the ratio of the actual amount of solar radiation to the possible radiation could be obtained by using the mean arithmetic value of S and $(1-n)$.

For calculations of the direct and total radiation Savinov suggested the following formulas:

$$Q=Q_0\,(1-\bar{n}) \qquad (23)$$

and

$$(Q+q)=(Q+q)_0\,(1-c\bar{n}), \qquad (24)$$

where

$$\bar{n}=1-\frac{1-n+S}{2}\,,$$

c — coefficient showing the effect of clouds on radiation.

According to the conclusions reached by Savinov, and also by B.M. Gal'perin (1949a /75/) and other authors, in the calculations of total radiation the utilization of factor $\bar{n}$ gives better results in comparison with those obtained by using the characteristics of the general cloudiness or sunshine hours. However, for many regions there are no reliable data on sunshine hours and therefore this compels us to use the data on clouds when determining the amount of total radiation.

The works of S.I. Savinov have contributed to the popularity of the formulas given above and they have been largely used in climatological calculations of the amounts of short-wave radiation. Of great importance is the fact that the amount of possible radiation $(Q+q)_0$, that is included in formulas (21), (22), & (24) appears to be a rather stable factor, which depends mostly on the latitude and season. This facilitated the use of these formulas for calculating the total radiation.

Among other empirical methods for determining total radiation, the formulas suggested by V.N. Ukraintsev, (1939 /221/) and Albrecht (1940) must also be mentioned.

Using observations of stations located inside the zone between 35° and 70°N, Ukraintsev has set up the following formula:

$$(Q+q)=m\sum S+n\,, \qquad (25)$$

where ΣS — is the total amount of sunshine hours, m and n — are the coefficients, which depend on the season and latitude.

Having computed these coefficients, Ukraintsev presented them in tables for each month and for various latitudes.

The testing of Ukraintsev's method on results of recent observations has shown that oftentimes it gives extenuated values of total radiation. It is possible, that this is the result of using the observations on scattered radiation which have been taken by obsolete instruments that give extenuated values.

The formula, suggested by Albrecht is:

$$(Q+q)=a\sin h_{\odot}\left(b-\frac{1}{\sqrt{\sin h_{\odot}}}\right)[1+(1-\eta)\,n]\ \ \mathrm{cal/cm^2/min.}\,, \qquad (26)$$

where $h_{\odot}$ —is the altitude of the sun, a, b and η —are numerical coefficients.

When calculating the amount of total radiation for more or less longer periods of time, the data obtained by this formula must be summed up according to diurnal variations of the sun's altitude.

Having determined the values of coefficients a and b by using the scarce observational data, Albrecht found $a = 0.31$-0.34 cal/cm^2/min, $b = 5.0$-7.4.

Comparing the results obtained by Albrecht's formula with observational data it could be found that the formula usually shows rather large systematic errors, mainly extenuating the values of total radiation.

This is apparently connected with the fact that Albrecht started with very insufficient observational data when deriving formula (26) and determining its coefficients. Some better results, as it seems, could be obtained by using the other Albrecht formula which has, with some simplifications, the following form:

$$(Q+q) = (a_0 \sin h_{\odot} - b_0 \sqrt{\sin h_{\odot}})\,[1-(1-\eta)\,n]\ \text{cal/cm}^2\text{/min} \qquad (26a)$$

The coefficient a_0, according to observational data, varies in limits 1.7-2.4; the coefficient b_0 approximately equals 0.32.

The physical meaning of this relationship was explained by K. ÎA. Kondrat'ev (1954 [137]).

In investigations on the climatology of heat balance, made by the Central Geophysical Observatory, the following equation has been used for calculating total radiation:

$$(Q+q) = (Q+q)_0\,[1-(1-k)\,n], \qquad (27)$$

In modern literature this formula is usually called the Savino-Ångström formula.

The parameters, included in the formula, have been determined by T.G. Berlîand from data of actinometric observations (Budyko, Berlîand, Zubenok, 1954a, 1954b [51 & 52]).

The mean monthly values of possible radiation $(Q+q)_0$ have been found for various latitudes and for all months of the year by the suggested method of V.N. Ukraintsev (1939 [221]). Using this method, graphs have been constructed for stations located in various latitudes, showing, by the abscissa, the day of the year, and by the ordinate, the corresponding daily amounts of total radiation derived from several years of observations. The points on the graphs were located inside certain regions with a very definite upper boundary. Since the upper points on these graphs apparently show the clear days conditions, we can draw a curve through these points and obtain the annual march of the daily values of total radiation under cloudless sky conditions. The data which have been determined from this graph are presented in table 1.

It must be said here that the values of possible radiation, shown by

this table, are a little larger than those found in most of the preceding investigations. This difference is apparently connected, to a certain degree, with the utilization of the values of possible radiation from comparatively limited observational data, oftentimes obtained by obsolete instruments, which extenuated the values of the diffused radiation. On the other hand, some features of Ukraintsev's method (which gives the possible radiation of a highly transparent atmosphere, [1]) rather than conditions of average transparency) could contribute to the fact that the values of possible radiation given in table 1 could be a little bit exaggerated. However, as will be explained later, this fact should not lead to any noticeable systematic errors in calculations of total radiation by formula (27).

For determining the coefficient k, which accounts for the effect of cloudiness on the total radiation, data of observations from various latitudinal zones have also been used. The coefficient k presents the ratio between the actual radiation under overcast sky conditions and the possible radiation. It must depend on the mean altitude of the sun, on the properties of clouds and on the conditions of reflection of short-wave radiation (the value of the albedo).

Consequently, the mean values of the coefficient k, will be different for different regions, and also, this coefficient will change according to diurnal and annual variations.

The mean annual values of coefficient k, averaged for various latitudes are presented in table 2.

The values of coefficient k given in table 2 have been computed from data of actinometric observations at 62 locations. Since this coefficient was computed by formula (27) with the actually observed mean values $(Q+q)$ and calculated the values $(Q+q)_0$, by the above cited method, it is quite clear, that a small systematic error in values $(Q+q)_0$ will accordingly change the values of the coefficient k. This provides for some compensation of the effect of errors which arose by determining the parameters in formula (27) for the evaluation of total radiation.

In determining the total radiation by formula (27) and from tables 1 and 2, the effect of changes in the transparency of the atmosphere and the effect of changes in the mean heights and forms of clouds are counted only as the mean factors of the latitude [through the latitudinal changes of values $(Q+q)_0$ and k]. Besides, these calculations do not account for the annual variations of coefficient k, which, according to many authors, could be quite significant.

Consequently, the analyzed method for calculating total radiation must be regarded as a rather schematic one and mainly for use in calculating the distribution of radiation over vast areas of a continent and over the whole globe. An important advantage of this method is seen in the pos-

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1) Besides, in determining the possible radiation by Ukraintsev's method, in some cases, its values could be exaggerated because of the insignificant cloudiness (less than 2-3 tenths), which sometimes does not reduce but increases the total radiation in comparison with a cloudless sky.

sibility of using only the most readily available data on general cloudiness (since climatological data on frequency of various forms of clouds are missing or are not sufficiently reliable for many foreign countries and many portions of oceans). The question of the accuracy of total radiation amounts calculated by using formula (27) and tables 1 and 2 will be analyzed in § 6.

Table 1.

Latitude	J	F	M	A	M	J	J	A	S	O	N	D
80° N	0,0	0,0	2,5	9,6	17,9	20,3	18,9	10,8	3,6	0,4	0,0	0,0
75	0,1	0,6	4,0	11,2	18,7	20,9	19,7	12,3	5,3	1,7	0,2	0,0
70	0,2	1,4	5,8	12,7	19,4	21,4	20,3	13,7	7,0	3,0	0,7	0,1
65	0,8	2,5	7,6	14,1	20,1	21,9	21,0	15,1	8,8	4,5	1,5	0,4
60	1,7	3,9	9,6	15,4	20,8	22,3	21,6	16,4	10,5	6,1	2,6	1,2
55	3,0	5,6	11,5	16,6	21,5	22,7	22,1	17,7	12,3	7,7	4,1	2,3
50	4,7	7,5	13,5	17,8	22,1	23,0	22,5	18,8	14,2	9,6	5,8	3,8
45	6,6	9,4	15,4	19,0	22,6	23,3	22,9	20,1	16,0	11,6	7,7	5,7
40	8,7	11,5	17,0	20,0	22,9	23,5	23,2	21,1	17,6	13,4	9,7	7,7
35	10,8	13,6	18,5	21,0	23,0	23,5	23,3	21,8	18,8	15,1	11,8	9,6
30	12,7	15,2	19,5	21,6	23,0	23,5	23,3	22,2	19,8	16,5	13,6	11,4
25	14,3	16,5	20,3	21,8	22,9	23,4	23,1	22,3	20,5	17,6	15,0	13,1
20	15,5	17,5	20,8	21,8	22,6	22,9	22,7	22,2	21,0	18,5	16,3	14,5
15	16,6	18,3	21,0	21,6	22,0	22,2	22,1	21,8	21,1	19,2	17,3	15,7
10	17,4	19,0	21,0	21,3	21,2	21,2	21,2	21,2	21,1	19,6	18,0	16,6
5	18,0	19,5	20,8	20,8	20,4	19,8	20,1	20,5	20,8	19,9	18,6	17,3
0	18,5	19,8	20,4	20,2	19,2	18,0	18,7	19,6	20,4	20,0	19,0	18,0

Total radiation with a cloudless sky $(Q+q)_0$ kg-cal/cm^2/month.

Table 2.

$\varphi°$	75	70	65	60	55	50	45	40
k	0,55	0,50	0,45	0,40	0,38	0,36	0,34	0,33
$\varphi°$	35	30	25	20	15	10	5	0
k	0,32	0,3[illegible]	0,32	0,33	0,33	0,34	0,34	0,35

Mean latitudinal values of the coefficient k .

A more differentiated method of calculating total radiation should take into account the effect of forms and heights of clouds on total radiation in each location, and should also account for the effect of changes in the transparency of the atmosphere.

The effect of changes in properties of clouds on the annual march of radiation could be approximately accounted for by changes of the values of coefficient k. .

The annual variations of the correponding coefficients in the formulas of Savinov and Ångström are analyzed in the paper by B.M. Gal'perin (1949a [75]). In this investigation it is pointed out that coefficient c of Savinov's formula (24), in some regions, changes considerably during the year.

Our calculations have proven that, by utilization of formula (27) the changes of coefficient k during the annual period are also perceivable, but usually the neglect of these changes will not produce any considerable errors in the calculations of total radiation amounts.

Another way of estimating the effect of cloud properties on the total radiation is the inclusion of indices in the calculation formulas, showing cloud quantities at various cloud heights. So, for instance, P.P. Kuz'min (1950 [149]) assumed that the ratio of the actual amount of total radiation to the possible is:

$$1 - c_1(n_0 - n_н) - c_2 n_н ,$$

where n_0 — is the total amount of clouds; $n_н$ — the amount of lower clouds; c_1 and c_2 — are the coefficients; the first is equal to 0.14, the second to 0.67. This method of calculation could be applied to those cases when data on lower clouds are available.

In the investigation by A.P. Braslavskiĭ (Braslavskiĭ and Vikulina, 1954 [33]), the problem of estimating the effect of some additional factors in determining the amount of total radiation has been studied. Braslavskiĭ indicated, and rightly so, that in calculations of the possible radiation by theoretical methods, the effect of the albedo of the surface on the diffused radiation (this means, on the total radiation as well) must be directly estimated. The effect of the albedo on the total radiation is, to a certain degree, taken automatically into account when used in calculating the amounts of possible radiation, that have been found by observations, for the actual state of the surface.

The calculations made by Braslavskiĭ showed that changes in the atmospheric transparency which are associated with changes in humidity of the air, and also changes in heights of places up to a level of several kilometers, exert only an inconsiderable influence on the amount of total radiation.

The problem of methods for climatological calculations of diurnal variations in the total radiation has been elaborated by L.A. Biriukova (1955 [31]). For this purpose she has used formula (27) and has determined the parameters of this formula from observational data.

Diurnal variations of the amounts of possible radiation have been computed by Biriukova, using the idea of Ukraintsev's method. The graphs of diurnal variations have been constructed for all months of the year for 7 locations in the USSR: Pavlovsk, Riga, Sverdlovsk, Irkutsk, Odessa, Vladivostok, Tbilisi. The calculations showed that, the values of the possible radiation depend basically on the latitude, season and hour of the day. The computed values of the possible radiation (i.e., the total radiation of a cloudless sky) at various latitudes are given in table 3.

Table 3

Total radiation with a cloudless sky $(Q+q)_0$ kg-cal/cm^2/hour (according to observations made in the U.S.S.R.).

Lat.	Mo.	Hours 21 / 3	20 / 4	19 / 5	18 / 6	17 / 7	16 / 8	15 / 9	14 / 10	13 / 11	12
60°N	I							2	6	10	12
	II					1	5	13	19	24	29
	III				2	6	15	25	31	49	37
	IV		1	3	10	21	31	40	49	53	56
	V		4	10	19	29	39	47	55	58	60
	VI	2	5	13	22	31	40	48	55	60	62
	VII	1	5	12	19	29	40	47	55	59	61
	VIII		2	7	14	24	34	40	47	51	53
	IX			1	4	12	21	30	39	43	44
	X					3	11	19	27	31	34
	XI						2	5	10	15	18
	XII							2	3	6	8
55	I						2	7	13	18	20
	II					2	8	18	38	33	36
	III				3	13	21	33	42	46	49
	IV			3	10	20	31	42	50	56	58
	V		2	6	17	28	39	50	57	60	62
	VI	1	4	11	20	30	41	50	58	62	66
	VII		3	8	15	27	37	47	56	62	63
	VIII		2	5	13	24	34	42	50	55	58
	IX			1	6	14	24	36	45	50	53
	X				1	5	13	23	32	39	40
	XI						3	10	18	24	27
	XII							2	6	11	13
50	I						3	11	19	25	27
	II				1	4	11	21	34	42	45
	III			1	4	12	24	38	48	52	54
	IV			3	10	21	33	44	54	61	63
	V			5	15	28	41	53	60	65	68
	VI		2	9	19	30	44	54	62	68	70
	VII		2	6	14	27	40	52	61	66	68
	VIII		1	4	12	25	36	46	54	61	63
	IX				6	15	27	40	51	57	59
	X				2	7	16	26	36	43	46
	XI					1	5	13	23	29	31
	XII						1	6	12	18	20
45	I					1	5	15	24	32	35
	II				1	4	14	25	39	46	50
	III				4	13	25	42	53	59	61
	IV			2	9	21	35	48	59	66	68
	V			4	14	30	45	58	66	70	73
	VI		2	7	19	31	46	58	69	75	77
	VII		1	5	15	30	45	58	66	71	74
	VIII			3	11	26	39	50	61	67	70
	IX				5	16	30	45	57	64	66
	X				2	9	19	30	43	49	51
	XI					2	5	17	28	34	37
	XII						3	10	20	26	29

Having at hand the data on diurnal variations of the possible radiation and comparing them, according to formula (27), with data of observations on total radiation, it is easy to compute the diurnal variations of coefficient k and to determine its average relationship to the altitude of the sun.

This relationship, determined from calculation data for several points in the USSR, is presented in fig. 4.

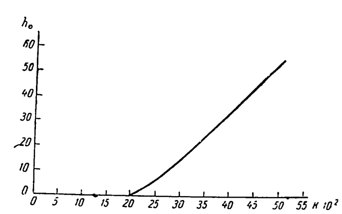

Figure 4

Relationship of coefficient k with the sun's altitude.

As can be seen from this graph, the magnitude of coefficient k decreases with the diminishing height of the sun. The reason for such changes is apparently found in the fact that, in the presence of clouds the increasing length of the sunbeam path in the atmosphere diminishes considerably the amount of radiation that reaches the earth's surface, as compared with the amount during cloudless conditions.

Using formula (27), table 3 and fig. 4, the diurnal variations of total radiation for average conditions could be computed. The question of the possibility of using this method for calculating the total radiation for short periods of time is still not clear.

Having at hand the data of total radiation, the value of the albedo of the underlying surface for short-wave radiation must be computed to enable us to determine the amount of absorbed radiation.

At present, ample data from observations on the mean values of the albedo for various underlying surfaces are available. Among the numerous works of determining the albedo we will mention the papers by A. Ångström

(1925a), A.A. Skvortsov (1928 [211]), N.N. Kalitin (1929 [118]), B.M. Gal'perin (1938 [74]), P.P. Kuz'min (1939 [146]), I.N. ÎAroslavtsev (1952 [249]), T.V. Kirillova (1952 [127]), which included surface observations of the albedo; and works by L.I. Zubenok (1949b [106]), Fritz (1949), and V.L. Gaevskiĭ (1953 [73]) where the albedo of the underlying surface was determined from an airplane.

The consolidated results of albedo measurements for various types of surfaces have been given in works by Budyko (1948a [39]), Berlíand (1948 [25]), Gaevskiĭ (1953 [73]), Kondrat'ev (1954 [137]) and others. The mean values of the albedo, obtained by the most reliable measurements in various physical geographical conditions, are presented in table 4.

Table 4

Albedo of the natural surfaces.

Types of surface	Albedo	Types of surface	Albedo
Snow and ice		Fields, meadows, tundra	
Fresh, dry snow	0.80-0.95	Rye and wheat fields	0.10-0.25
Pure, white snow	0.60-0.70	Potato plantations	0.15-0.25
Polluted snow	0.40-0.50	Cotton plantations	0.20-0.25
Sea ice	0.30-0.40	Meadows	0.15-0.25
		Dry steppe	0.20-0.30
		Tundra	0.15-0.20
Bare soil		Forests	
Dark soils	0.05-0.15		
Moist grey soils	0.10-0.20	Coniferous forests	0.10-0.15
Dry, clay or grey soils	0.20-0.35	Deciduous forests	0.15-0.20
Dry, light, sandy soils	0.25-0.45		

For better schematic climatological computations of the absorbed radiation it is more convenient to use the more generalized average values of the albedo, which are shown in table 5 (Budyko, Berlíand, Zubenok (1954b [52]).

From data in tables 4 and 5 we can see that, in moderate altitudes the values of the albedo from the land surface will change considerably in the annual course, reaching the maximum in winter months when a stable snow cover is observed.

Table 5

Mean values of the albedo for the main types of natural land surfaces.

Stable snow cover of higher latitudes (60° and higher)	0.80
Stable snow cover of the middle latitudes (below 60°)	0.70
Unstable snow cover	0.45
Coniferous forest	0.14
Tundra, steppe, deciduous forest, savanna in the moist season	0.18
Savanna in the dry season and semideserts	0.25
Deserts	0.30

The highest values of the albedo are observed in winter in the high latitudes, where the surface of snow is preserved pure and is not polluted, since the air contains only insignificant amounts of dust. Very large values of the albedo for fresh snow are also noticed in moderate latitudes.

Compared with that in moist areas, a higher albedo has been observed in dry regions and especially deserts. Even so, the albedoes measured in deserts are subject to wide fluctuations [2], however, as a rule, they are still larger than that of the vegetation covered surface.

The modern investigations by I.N. ÎAroslavtsev (1952 [249]), V.L. Gaevskiĭ (1953 [73]), K. ÎA. Kondrat'ev and N.E. Ter-Markariants (Kondrat'ev, 1954 [137]), and other authors, have shown that the values of the land surface albedo often change considerably during the day. With the lower altitudes of the sun (in the morning and evening hours), the albedo is

2) The color of soil surfaces in deserts is very variable, which provides for a corresponding variability of albedoes and this fact is oftentime encountered even within a limited geographical region.

usually considerably larger. The reason for this is seen in the different reflective capacity of the rough underlying surfaces, for sun rays falling at different angles (at high sun, its rays penetrate deep into the vegetation layer and are absorbed there, whereas at low sun the rays do not penetrate as much into the vegetation layer and a larger portion is reflected by the surface.).

Besides, the diurnal variations of the albedo are sometimes also affected by the spectral composition of short-wave radiation at different heights of the sun.

In climatological calculations of diurnal variations of the amount of absorbed radiation, the relationship between the value of the albedo and height of the sun, as found by L.A. Biriukova (1955 /31/), from observational data, could be applied. Biriukova has noticed that, for the snowless period the albedo in its diurnal variations changes diurnally, depending on the height of the sun and cloud amounts. It has also been found that an increase in cloudiness diminished the degree of dependence of the albedo on the height of the sun, because the increase in cloud amounts reduced the direct sun radiation and augmented the diffused one, the absorption of which does not depend directly on the height of the sun.

The relationship between $\Delta\alpha$ (difference between the value of the albedo at a given hour and at noon) and the height of the sun under average conditions of cloudiness is presented in fig. 5.

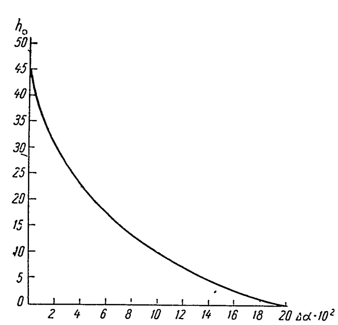

Figure 5

Dependence of the albedo on the sun's altitude.

This graph can be used for an approximate estimation of the diurnal variations of absorbed radiation during the snowless periods. When there is a snow cover, according to Biriukova, there is no need of estimating the albedo variations in determining the amount of absorbed radiation.

The albedo of water surfaces is, on the average, less than that of most of the natural land surfaces. A relatively great absorption of short-wave radiation in water reservoirs is explained by the fact that the sun rays penetrate the upper translucent water layers, where they are scattered and almost completely absorbed. This is why the albedo of muddy water reservoirs is considerably higher.

For direct radiation the albedo of a water surface depends greatly on the altitude of the sun, and varies from a few per cent at high sun to almost 100% for the sun near the horizon.

The dependence of the albedo on the angle of the sun rays could be calculated theoretically by Fresnel's formula. Many authors have shown that theoretical calculations of the albedo for direct radiation comply fairly well with observational data (Kondrat'ev, 1954 /137/).

The albedo of a water surface for diffused radiation varies in much closer limits, and on the average it is on the order of 8-10%. The estimate of albedo variations for diffused radiation, as dependent on cloudiness and on other factors, might be of some importance in calculating amounts of radiation reflected by the water surface. However, for those climatological calculations of amounts of absorbed radiation that are of interest to us (the absorbed portion is usually larger than the reflected one), the possible variations of the albedo for diffused radiation are insignificant and do not affect the results of the computations very much.

Because of a very close dependence of the albedo of water reservoirs on the sun's altitude, the albedo of the total radiation shows a definite annual and diurnal course.

To determine the average values of the albedo of water reservoirs, in the investigations made by Budyko, Berliand, Zubenok (1954a, 1954b /51 & 52/), the data of S.I. Sivkov (1952 /210/) have been utilized. On the basis of experimental data, and also using some results obtained by theoretical calculations, Sivkov found a relationship between the albedo of the water surface for direct radiation and the altitudes of the sun. Assuming the albedo for diffused radiation as being, on the average, 0.10 and estimating the mean relationship between direct and diffused radiation at various latitudes, we found the values of the albedo of water surfaces for the total radiation and they are given in table 6.

These data can be also used in calculations for the Southern Hemisphere, taking into account the proper changes of seasons.

In climatological calculations of solar radiation absorbed by water reservoirs, it must be kept in mind that, the change in the state of water surfaces produced by the rise of waves, exerts a definite effect on the albedo value. However, it must be pointed out that, these changes cannot affect, to any substantial degree, the values of absorbed radiation. Since the mean values of the albedo of water surfaces usually do not exceed 0.10, it must be clear that, the comparatively large changes in this value will affect the values of absorbed radiation only to a relatively insignificant

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degree. This permits a neglect of the effect of waves on the albedo when calculating the sums of absorbed radiation for periods on the order of a month, of ten days, etc.

Table 6

Water surface albedo for total radiation.

Lat.	J	F	M	A	M	J	J	A	S	O	N	D
70° N	—	0,23	0,16	0,11	0,09	0,09	0,09	0,10	0,13	0,15	—	—
60	0,20	0,16	0,11	0,08	0,08	0.07	0,08	0,09	0,10	0,14	0,19	0,21
50	0,16	0,12	0,09	0,07	0,07	0,06	0,07	0,07	0,08	0,11	0,14	0,16
40	0,11	0,09	0,08	0,07	0,06	0,06	0,06	0,06	0,07	0,08	0,11	0,12
30	0,09	0,08	0,07	0,06	0,06	0,06	0,06	0,06	0,06	0,07	0,08	0,09
20	0,07	0,07	0,06	0,06	0,06	0,06	0,06	0,06	0,06	0,06	0,07	0,07
10	0,06	0,06	0,06	0,06	0,06	0,06	0,06	0,06	0,06	0,06	0,06	0,07
0	0,06	0,06	0,06	0,06	0,06	0,06	0,06	0,06	0,06	0,06	0,06	0,06

In the calculations of radiation balance it is necessary to take into account not only the amounts of short-wave radiation lost by reflection but also the loss of radiation heat through the effective outgoing radiation.

The radiation of the underlying surface follows the Stefan law and is equal to $s\sigma\theta_w^4$ cal/cm²/min., where θ_w —is the temperature of the surface, σ — -the Stefan-Boltzmann's constant, which according to recently obtained data is equal to $8.14 \cdot 10^{-11}$; s — is the coefficient which characterizes the deviation of radiation of the given surface from that of a black body.

According to measurements made by Aleksandrov and Kurtener (1941 [7]), Falkenberg (1928) and other authors, the values of coefficient s for the most natural surfaces are equal to 0.85-1.00.

A considerable portion of the flux of long-wave radiation that is radiated by the underlying surface is compensated by counter radiation from the atmosphere, which depends mainly on the content of water vapor, air temperature, and cloud conditions.

The methods for measuring counter radiation from the atmosphere and also for the effective outgoing radiation have been under development for a long time, but only recently the instruments for measuring outgoing radiation at various hours without sizeable errors have been contructed. The instruments that were employed earlier had some constructive defects and oftentimes had a faulty calibration, which was the reason for the exaggerated values of the effective outgoing radiation [this is explained by the fact that the calibration was usually done with the pyrgeometer of

Ångström or with some instruments that were previously calibrated with it. This pyrgeometer, as was discovered later, gave substantially exaggerated readings (M.E. Berlîand and T.G. Berlîand, 1952 [24])].

The most satisfactory, of all modern instruments for measuring outgoing radiation, is the effective pyranometer by ÎAnishevskiĭ and the vibrational pyranometer by Falkenberg, though these instruments are not without fault either. Observational data, obtained by various instruments measuring the effective outgoing radiation, have often been used for determining the rate of dependence of the effective outgoing radiation on meteorological factors.

Most of the formulas connecting the value of the effective outgoing radiation under a cloudless sky with the temperature and humidity of the air, have the following form:

$$I_0 = \sigma\theta^4 \left(a_1 + b_1 \cdot 10^{-c_1 e}\right), \qquad (28)$$

or

$$I_0 = \sigma\theta^4 \left(a_2 + b_2 \sqrt{e}\right), \qquad (29)$$

where I_0 — is the effective outgoing radiation, θ — air temperature, e —vapor pressure, a_1, b_1, c_1, a_2 and b_2 — coefficients.

The first of these equations was suggested by Ångström, the second by Brunt.

The coefficients in formula (28) and (29) were determined, in some investigations, from observational data.

In climatological calculations of effective outgoing radiation, the Ångström formula had earlier been used with the coefficients given in Linke's meteorological textbook (Linke, 1934), $a_1 = 0.194$, $b_1 = 0.236$, $c_1 = 0.069$, for determining the effective outgoing radiation in cal/cm²/min. and air humidity in mm.

In researches of recent years the theoretical methods for determining values of effective outgoing radiation have been developed. In this field the papers by K. ÎA. Kondrat'ev (1949a, 1949b [133 & 134], etc.) were of great significance. He established and put in use the scheme of a differentiated accounting of the spectrum of absorption coefficients of long-wave radiation in the atmosphere.

Utilizing the results obtained by R. ÎA. Kondrat'ev, M.E. Berlîand has established a theoretical relation of the effective outgoing radiation in a cloudless sky to air temperature and humidity (M.E. Berlîand and T.G. Berlîand, 1952 [24]).

This relation can be approximately expressed in the following analytical form:

$$I_0 = s\sigma\theta^4 \left(0,39 - 0,058 \sqrt{e}\right), \qquad (30)$$

where e —is in mm., I_0 — in cal/cm²/min.

For practical computations it is convenient to use table 7, which has been computed according to Berlîand's calculations.

It should be noted that the relation that has been established by M.E. Berlîand, theoretically, turned out to be very close to the empirical relationship which was found by Bolz and Falkenberg (1949) from 1320 observations taken with a vibrational pyranometer under a cloudless sky.

Besides the air temperature and humidity, some other factors exert a substantial influence on the effective outgoing radiation. These are: amount of clouds and the difference in temperatures between the soil surface and the air.

Table 7

Effective outgoing radiation with a cloudless sky in kg-cal/cm^2/min.

Temperature	Humidity of the air mm										
	1	2	3	4	5	6	7	8	10	12	15
−20°	0,11										
−15	0,12										
−10	0,13	0,12									
−5	0,14	0,13	0,12								
0	0,15	0,14	0,13	0,12							
5	0,16	0,15	0,14	0,13	0,13	0,12					
10	0,17	0,16	0,15	0,14	0,14	0,13	0,12	0,11			
15		0,17	0,16	0,15	0,15	0,14	0,13	0,12	0,11	0,10	
20			0,17	0,16	0,16	0,15	0,14	0,13	0,12	0,11	
25				0,17	0,17	0,16	0,15	0,14	0,13	0,12	0,10
30					0,18	0,17	0,16	0,15	0,14	0,13	0,11

The estimate of the effect of clouds on effective outgoing radiation was done earlier by formula:

$$I = I_0(1 - cn), \qquad (31)$$

where I — effective outgoing radiation at the existent cloud amount, n — - cloud amount in tenths, c — coefficient.

Ångström found the average value of c to be 0.75, Asklöf, Dorno and other authors found the average value of c varied for clouds of different heights - for high clouds the magnitude of this coefficient turned out to be much smaller than that of lower ones. Considering this fact, N.G. Efimov (1939 [97]) suggested the following formula for calculating the effective outgoing radiation dependent on clouds

$$I = I_0[1 - (c_в n_в + c_с n_с + c_н n_н)], \qquad (32)$$

where $n_н$, $n_с$ and $n_в$ — are amounts of clouds for the higher, middle, and lower layers, $c_в$, $c_с$ and $c_н$ — the corresponding coefficients. Efimov estimated these coefficients as being: $c_в = 0.15$-0.20; $c_с = 0.5$-0.6; $c_н = 0.7$-0.8.

In the work by I.G. Liutershtein and A.F. CHudnovskii (1946 [164]) some larger values of these coefficients are given:

$$c_в = 0.20, \quad c_с = 0.6 - 0.7, \quad c_н = 0.8 - 0.9.$$

Recent investigations show that according to observational data the effective outgoing radiation decreases with increasing cloudiness not linearly but noticeably faster. Therefore, the following formula has been developed for determining the effective outgoing radiation:

$$I = I_0(1 - cn^m), \qquad (33)$$

where $m = 1.5$-2.0

Obviously, when determining the value of c under overcast sky conditions, this value will be the same whether derived by formula (31) or (33).

The theoretical calculations of the mean values of coefficient c for various latitudes has been done by M.E. Berliand. In these calculations he has taken into account the mean frequency of clouds at various heights in various latitudes. The obtained values of coefficient c are presented in table 8.

Table 8

The mean values of coefficient c.

$\varphi°$	75	70	65	60	55	50	45	40
c	0.82	0.80	0.78	0.76	0.74	0.72	0.70	0.68
$\varphi°$	35	30	25	20	15	10	5	0
c	0.65	0.63	0.61	0.59	0.57	0.55	0.52	0.50

Smaller values of this coefficient in lower latitudes are explained mainly by greater heights of middle layer clouds in these regions.

In some works (Kuz'min, 1948 [148]; Bolz, 1949) the effect of clouds on outgoing radiation is taken into account by introducing a correction, but this correction does not refer to the effective outgoing radiation, but to the value of the counter radiation from the atmosphere. As has been shown by K. IA. Kondrat'ev (1951 [135]), such a method for calculating the effect of clouds on outgoing radiation has no substantial advantages in comparison with the use of formulas (31) and (33).

In the latest works of the Central Geophysical Observatory on construction of radiation balance maps, the effective outgoing radiation has been calculated by the following formula:

$$I = I_0(1 - cn^2) + 4s\sigma\theta^3(\theta_w - \theta). \qquad (34)$$

The second term of this formula permits an estimate of the effect of temperature differences between the underlying surface and the air on the effective outgoing radiation.

When there is a difference in temperatures, the effective outgoing radiation changes, and this change is represented by $s\sigma\theta_w^4 - s\sigma\theta^4$, which is approximately equal to:

$$4\,s\sigma\theta^3(\theta_w - \theta).$$

The theoretical explanation of this correction can be found in several works (Kondrat'ev, 1951 /135/; M.E. Berliand and T.G. Berliand, 1952 /24/; and others).

When using formula (34) coefficient s was on the average taken as 0.9, values c and I_0 were computed from tables 7 and 8. The temperature of θ_w, the active surface for water reservoirs, can be determined from observational data. Since there are usually no reliable data on temperatures of the underlying surface on the mainland it is feasible to use an indirect method.

In the author's works (Budyko 1949a /41/, 1950b /44/ and others), the turbulent stream of heat has been determined by formula: $P = b(\theta_w - \theta)$, where b — is the coefficient of proportionality (more details about this relationship can be found in §4). Taking into account this relationship and also formulas (4), (5) and (34) the following equation may be set up:

$$4s\sigma\theta^3(\theta_w - \theta) = \frac{(Q+q)(1-\alpha) - I_0(1-cn^2) - LE - A}{1 + \frac{b}{4s\sigma\theta^3}}. \qquad (35)$$

The term $\frac{b}{4s\sigma\theta^3}$ is variable and, in particular, it depends substantially on the intensity of the turbulent exchange in the air layer near the ground. However, considering the fact that the term $4s\sigma\theta^3(\theta_w - \theta)$ usually presents a comparatively small correction to the radiation balance value (except for the cold season in temperate and higher latitudes), we can use, for an approximate calculation of radiation balance, just the mean value of the mentioned term.

The results of the computations permitted the following conclusions about the average climatic values of the term $\frac{b}{4s\sigma\theta^3}$:

1) For conditions when $(Q+q)(1-\alpha) - I_0(1-cn^2) - LE - A > 0$,

the value of the ratio $\frac{b}{4s\sigma\theta^3}$ is, on the average, 3;

2) For conditions when $(Q+q)(1-\alpha) - \bar{I}_0(1-cn^2) - LE - A < 0$,

the value of the ratio $\frac{b}{4s\sigma\theta^3}$ is, on the average, 1.

The different values of the analyzed ratio in these two cases are the result of the fact that in the first case, in the air layer near the ground a superadiabatic stratification prevailed and the turbulent exchange was reinforced, whereas in the second case an inversion took place that diminished the intensity of turbulent exchange.

Utilization of the given mean values permits an approximate determination of the effect exerted by the term $4s\sigma\theta^3(\theta_w - \theta)$ upon the effective outgoing radiation and the radiation balance on land (the question about the calculation of values LE and A, included in the formulas will be analyzed later).

The outlined method for the climatological calculation of effective outgoing radiation permits one to determine its magnitude from ordinary meteorological observations of temperature, humidity, cloudiness and (for reservoirs) water surface temperature. In cases when data on heights and forms of clouds are available, as well as data on vertical distribution of temperature and humidity in the troposphere, more differentiated methods for determining the effective outgoing radiation can be employed. So, for instance, in the investigation made by M.E. Berliand and T.G. Berliand (1952 /24/), and in some other works, the average values of coefficient c, for various forms of clouds, are presented. The application of these coefficients in those cases when data on cloud forms are available can make the calculations of the effective outgoing radiation more precise. The question of the necessity of accounting for the vertical distribution of temperature and humidity when calculating the outgoing radiation has been analyzed in many researches. The computations accomplished in recent years by V.V. Mukhenberg have shown that, when long period means of the effective outgoing radiation are determined for single months, the estimate of the really observed vertical gradients of temperature does not effect any larger changes in those values of the effective outgoing radiation which have been determined according to the scheme described above, proceeding only from data of surface observations.

However, it must be pointed out that, for calculations of the effective outgoing radiation during shorter periods, vertical gradients of temperature and humidity are of substantial importance. In such calculations it would be expedient to use methods that have been elaborately developed in the latest works of F.N. SHekhter (1950 /236/), T.V. Kirillova and E.D. Kovaleva (1951 /128/) and other authors.

§ 4. Turbulent heat exchange between the underlying surface and the atmosphere

The temperature of the underlying land surface, as well as that of water surfaces, is usually different from the temperature of the lower layer of the atmosphere. Consequently, a vertical flux of heat arises between the underlying surface and the atmosphere, effected by turbulent heat conduction in the air layer near the ground.

The evaluation of the vertical turbulent heat flux usually presents the greatest difficulties, in comparison with evaluating the other components of heat balance, regardless how it is done, either by direct measurements or by indirect climatological calculations.

Some methods for calculating the turbulent flux of heat are based on the equation of Fick, which was used in calculating the flow of heat and moisture in the atmosphere, by Taylor (1915) and Schmidt (1917).

This equation is now widely used in studies of heat exchange and exchange of moisture in the air layer near the ground.

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Assuming that, according to the idea of Taylor and Schmidt, the process of turbulent diffusion is similar to that of molecular diffusion, we will obtain the following formula for the vertical turbulent heat flux in the air layer near the ground:

$$P = -\rho c_p k \frac{\partial \theta}{\partial z}, \qquad (36)$$

where ρ — is air density, c_p — heat capacity of the air at constant pressure, k — coefficient of the turbulent exchange, $\frac{\partial \theta}{\partial z}$ — vertical gradient of the absolute temperature.

In equation (36), there is no need to take into account the value of the equilibrium gradient 3) since in the layer near the ground there usually are gradients exceeding the equilibrium gradients by 10-100 times.

Integration of formula (36) by z will result in the equation:

$$P = \rho c_p D(\theta_w - \theta), \qquad (37)$$

where θ_w is the temperature of the active surface, θ — air temperature at some height, D — integral characteristic of the conditions of vertical turbulent transfer that goes on between the underlying surface and the atmosphere, which will further be called - the coefficient of external diffusion.

The last equation actually presents the well-known Newton law that established the relationship between the heat exchange and difference in temperatures of the surface of a heated (or cooled) body and the air.

Equations (36) and (37) present the basis for a series of methods for determining the turbulent heat exchange. It is usually possible to apply the first of these equation when some special observations of the gradient are available, i.e., when we have on hand measurements of the vertical temperature gradient in the layer near the ground, and also measurements of the gradients of those meteorological elements which are necessary in calculating the turbulent exchange coefficient k.

Let us analyze briefly the methods for determining the coefficient of exchange.

Some of these methods are based on the utilization of measurements of vertical streams of heat and moisture.

In modern researches, the method frequently used in determining the coefficient of exchange from measurements of evaporation was the one initially suggested in 1946 (Budyko 1946a [35]).

The essential features of this method follow. By analogy with equation (36) for the conditions of the air layer near the ground we may write the formula:

$$E = -\rho k \frac{\partial q}{\partial z}, \qquad (38)$$

3) The equilibrium gradient represents a vertical gradient of temperature, which brings the turbulent flux of heat down to zero. According to researches made by the author and M.I. Ĭudin (1946, 1948 [60 & 61]), the value of the equilibrium gradient is, on the average, about 0.6° per 100m.

where E — is the speed of evaporation, $\frac{\partial q}{\partial z}$ — the vertical gradient of specific humidity.

Numerous experimental investigations have proven that the coefficient of turbulent exchange in the air layer near the ground increases with height according to a law that is close to the following equation:

$$k = k_1 z, \qquad (39)$$

where k_1 — is the coefficient of exchange at a unit of height.

Integrating equation (38) by z and taking formula (39) into account, we will obtain:

$$E = \rho k_1 \frac{q_1 - q_2}{\ln \frac{z_2}{z_1}}, \qquad (40)$$

where q_1 and q_2 — are the specific humidities at heights z_1 and z_2.

Hence we can see that the value of exchange coefficient k_1 could be computed by formula:

$$k_1 = \frac{E \ln \frac{z_2}{z_1}}{\rho (q_1 - q_2)}. \qquad (41)$$

This equation has been used many times in determining the coefficient of exchange, by expeditions, when measuring evaporation and air moisture at two heights.

Another method for determining the coefficient of exchange (the so-called method of heat balance) is based on the utilization of a formula that is derived from equation (36), integrating it by z and taking into account equality (39). This formula is:

$$P = \rho c_p k_1 \frac{\theta_1 - \theta_2}{\ln \frac{z_2}{z_1}}. \qquad (42)$$

Eliminating the values of P and E, from formulas (40) and (42) and from the equation of heat balance we will obtain the following formula for the exchange coefficient:

$$k_1 = \frac{(R - A) \ln \frac{z_2}{z_1}}{\rho [c_p(\theta_1 - \theta_2) + L(q_1 - q_2)]}. \qquad (43)$$

Now the method of heat balance is one of the main ways in determining the coefficient of exchange, derived from observations taken by expeditions (Rusin, 1952 [198]; Ogneva, 1955 [179] and others). It must be noted that the principal formula of this method (43) is based on the assumption that the values of the exchange coefficient for heat and moisture in the air layer near the ground are very similar. This assumption, which used to be very debatable (see Sverdrup, 1935b; Pasquill, 1949), was later confirmed by a series of recent investigations (Budyko, 1948a [39]; Timofeev 1951 [219], and others).

In recent years many authors have paid a great deal of attention to the development of such methods, for determining the values of exchange coefficients in the air layer near the ground, which would not be based on the use of data on evaporation or radiation balance measurements.

In researches of Rossby (1932) and Rossby and Montgomery (1935) a semi-empirical theory of the boundary layer was used in determining the values of turbulent exchange in the air layer near the ground. In accordance with Prandtl's (1932) point of view, Rossby and Montgomery assumed that, with an adiabatic distribution of temperature, the mixing length in the lower air layer increases lineary with height, and at the level where the wind speed equals zero (at z), it reaches a value which is proportional to the roughness of the underlying surface, i.e.,

$$l=\varkappa(z+z_0), \tag{44}$$

where l — is the mixing length; z_0 —roughness; $\varkappa$ — a constant without dimensions, approximately equal (according to experiments in tubes) to 0.38.

According to Prandtl, the magnitude of turbulent friction τ in the boundary layer is:

$$\tau=\rho l^2\left(\frac{\partial u}{\partial z}\right)^2=\rho\varkappa^2(z+z_0)^2\left(\frac{\partial u}{\partial z}\right)^2, \tag{45}$$

where $\frac{\partial u}{\partial z}$ — is the vertical gradient of wind speed. Having assumed the hypothesis that, in the lower layer of the atmosphere τ is constant with height, we can obtain the following equation of the wind profile, by integrating equation (45) by z:

$$u=\frac{1}{\varkappa}\sqrt{\frac{\tau}{\rho}}\ln\frac{z+z_0}{z_0} \tag{46}$$

and for the coefficient of turbulent exchange:

$$k=l^2\left(\frac{\partial u}{\partial z}\right)=\varkappa^2(z+z_0)\frac{u_1}{\ln\frac{z_1+z_0}{z_0}}, \tag{47}$$

where u_1 — is wind speed measured at the height of z_1.

It can be seen from formula (47) that, in the air layer near the ground the coefficient of exchange is proportional to z and to wind speed u , and that from formulas (45) and (46), the vertical gradient of wind speed is also proportional to the speed.

Numerous observations on wind profiles in the lower layers of the atmosphere have shown, however, that the magnitude of the vertical gradient of wind speed over a given underlying surface depends substantially not only on wind speed, but also on the vertical distribution of temperature.

So, for instance, in our works (Budyko, 1945, 1946b [34 & 36]) it was established that the ratio of the wind speed at 5m to that of 1m depends on the wind speed at 2m and the difference in temperatures between 20 and 150cm. The relationship presented in fig. 6 shows how substantial the effect of thermal stratification is on wind profiles in the air layer near the ground. On temperature decrease with height (which is shown in fig. 6 by the positive values of $\Delta\theta$) the ratio of wind speeds at two heights decreases considerably as compared to that under conditions of an inversion.

In accordance with this conclusion it becomes clear that the effect of vertical temperature distribution on the coefficient of exchange must necessarily be taken into account.

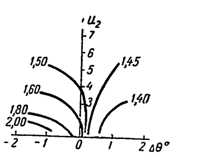

Figure 6

Dependence of the wind speed ratio at 5m and 1m levels on wind speed at 2m (u_2) and on temperature difference between the heights of 20 and 150cm ($\Delta\theta$) according to observational data.

The question about the effect of the vertical temperature profile on the turbulent exchange in the air layer near the ground was analyzed in a series of researches, including theoretical as well as experimental investigations.

A wide range of investigations on this problem have been accomplished, particularly at the Central Geophysical Observatory (Budyko, 1946b, 1948a [36 & 39]; Laikhtman, 1944, 1947 [152, 153 & 154]; Timofeev, 1951 [219] and many other).

The first works of this series have already established the fact that during daylight hours the values of the exchange coefficient in the air layer near the ground are considerably higher than those determined by formula (47) for the equilibrium state of thermal stratification. For night hours a reverse relationship was found. Consequently, the use of the above given formula (47), obtained in a semiempirical way, for determining the turbulent exchange coefficient, will inevitably result in large errors in computing the exchange for daylight hours, which are the most important for determining the turbulent heat exchange.

This conclusion has been corroborated by results of a verification of evaporation calculations that were done earlier by Thornthwaite and Holzman (1942), who computed the evaporation by using the coefficient of exchange obtained by a formula that did not take into account the effect of thermal stratification. The verification of these computations was made by methods of heat and water balance, and it showed (Budyko, 1946b [36]) that, the failure to account for the equilibrium effect on the exchange may lead to very great errors in determining evaporation and turbulent heat exchange.

In my works (1946, 1948a [35, 36, 37 & 39]) the following formula was suggested for computing the effect of equilibrium stratification on the turbulent exchange:

$$k=k_p f, \tag{48}$$

[where k_p —is the coefficient of exchange at the equilibrium state, f—

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function depending on characteristics of the Richardson's number in the air layer near the ground $\frac{\Delta\theta}{\Delta u^2}$ or $\frac{\Delta\theta}{u^2}$, but independent of height ($\Delta\theta$ — difference of the absolue temperatures at two levels, Δu – difference in wind speed at two levels, u — wind speed)].

From equation (48) and from some additional assumptions the following formula for the coefficient of exchange was obtained:

$$k = \frac{0{,}144\Delta u}{\ln \frac{z_2}{z_1}} \left[1 + \ln \frac{z_2}{z_1} \frac{\Delta\theta}{(\Delta u)^2} \right] z, \qquad (49)$$

where Δu — in m/sec for the heights z_2 and z_1; $\Delta\theta$ — difference in temperature at same heights. 4)

This formula permits the calculation of the exchange coefficient without using any data on the roughness of the underlying surface z_0. On the other hand, the necessity of taking into account the values of wind speed differences at two heights $\Delta\bar{u}$, which are usually measured with a considerable error, lowers the accuracy of the exchange coefficient computation when it is done by formula (49).

More precise results have been obtained by calculating the coefficient of exchange by using formulae, similar to formula (49), but involving data on wind speed at one height only. In this case, however, it is necessary to utilize data on the roughness of the underlying surface.

Fig. 7 shows the dependence of the exchange coefficient, at the height of 1m, on wind speed at this level (u_1) and on the temperature difference at heights of 55 and 150 cm, $\Delta\theta$ (here the difference of temperature is assumed as being positive even when the temperature decreases with altitude). The isolines of the exchange coefficients (in cm^2/sec) have here been computed for the roughness value equal to 2 cm, which approximately represents average conditions for the warm season on land.

Some other semiempirical formulas for determining the coefficient of exchange, that take into account the effect of stability, have been suggested by M.P. Timofeev (1951 [219]) and D.L. Laĭkhtman (1944, 1947 [152, 153 & 154], and others). Fig. 8 shows the comparison of the exchange coefficient relationships with the ratio of temperature differences at two heights to the second power of wind speed, i.e., with the parameter that characterizes stability according to formulae by Timofeev (curve 2), by Laĭkhtman (curve 3), and the author (curve 1).

This graph distinctly shows the sufficiently close similarity of all three formulae (Budyko, Laĭkhtman, Timofeev, 1953 [58]). The good agreement of different methods of computing the exchange coefficient is one of the indirect proofs of their sufficient reliability.

These formulae were tested on ample experimental material. For this purpose the computed coefficients of exchange were partly used, according to the evaporation and vertical gradient of moisture and according to

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4) It must be kept in mind that, at $\ln \frac{z_2}{z_1} \frac{\Delta\theta}{(\Delta u)^2}$ a factor is attached which is, for the order of magnitude used here, very close to one (Budyko, 1948a [39]).

measurements of radiation balance by earlier mentioned methods.

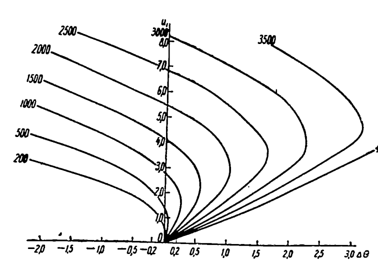

Figure 7

Dependence of the exchange coefficient at the 1 m level (in cm^2/sec) on thermal stratification $\overline{\Delta\theta}$ and on wind speed at 1 m.

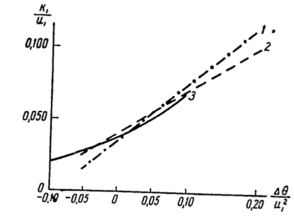

Figure 8

Dependence of the coefficient of exchange on the stability according to formulas developed in the Central Geophysical Observatory.

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The results of this verification showed (Budyko, 1948a /39/; Timofeev, 1951 /219/; Budyko, Laikhtman, Timofeev, 1953 /58/, Ogneva, 1955 /179/ and others) that the semiempirical formulae obtained in the Central Geophysical Observatory for computing the coefficient of exchange from gradient observations can render quite a satisfactory rate of accuracy 5). It should be mentioned here, by the way, that the use of these formulae permitted a quantitative interpretation of the empirically established relationship of vertical wind speed and temperature gradients with wind velocity, which is presented in fig. 6. The same kind of relationship. obtained by calculation, and presented in fig. 9, shows a great similarity with the empirical regularity. Such congruence serves as a confirmation of correctness for the scheme used in estimating the effect of thermal stratification on turbulent exchange. Among other works dedicated to analysis and development of methods for determining the coefficient of exchange in the air layer near the ground, those made by scientists of the Geophysical Institute A.M. Obukhov 1946 /177/; Monin and Obukhov, 1954 /175/ and others, should also be mentioned. The results of these works, which contain a series of interesting theoretical considerations, have not as yet been compared with empirical materials.

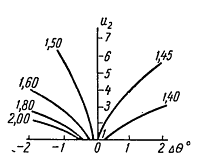

Figure 9

Dependence of the ratio of wind speed at the 5 m and 1 m levels on wind speed at the 2 m level (u_2) and on temperature differences between the levels of 20 and 150 cm ($\Delta\theta$) according to calculation data.

As a result of the application of methods described above, for determining the coefficient of the turbulent exchange in the air layer near the ground, a series of principles concerning this coefficient were found and investigated.

5) The method for determining the coefficient of exchange, by the formulas mentioned which employ the measurements of gradients made by hydrometeorological stations, has been developed by N.P. Rusin (Methodical Instructions, /171/ 1954).

The available data permits the following conclusions to be made concerning the basic principles underlying the variations of the exchange coefficient: 1. The coefficient of exchange in the air layer near the ground increases with height and is approximately proportional to the elevation up to a level of several scores of meters. During daylight hours (with a superadiabatic stratification) the coefficient of exchange usually increases somewhat faster than the height; during night hours (with inversions) the increase of the exchange coefficient occurs at a somewhat slower rate. Therefore, variations of the exchange coefficient with height could be calculated by formulae using the generalized expotential law, which was suggested by D.L. Laikhtman:

$$k = k_1 z^{1-\varepsilon}, \tag{50}$$

where ε — is a parameter, less than zero under superadiabatic stratification conditions, and greater than zero under inversion conditions in the air layer near the ground. It should be pointed out that, under average conditions during daylight hours (including noontime), as well as at nighttime, the parameter ε is very small - usually it is not greater than $\pm$ (0.10-0.15) (Ogneva, 1955 /179/).

This conclusion shows that in climatological calculations of the exchange coefficient diurnal variations with height, a practically sufficient accuracy could be obtained by using formula $k = k_1 z$. The vertical distribution of temperature, wind velocity and air humidity, resulting from the formulas given above, will be presented in the form of well-known logarithmic laws.

2. The mean values of the exchange coefficient at a level of 1 m, during daylight hours of the warmer season on land, are on the order of 1500-2000 cm^2/sec. Daily variations of this coefficient in the warmer season on land are very large, reaching a maximum in the afternoon and a minimum at night and in the early morning hours. Under clear weather conditions the coefficient of exchange may vary at least 10-100 times its value during 24 hours. In cloudy weather and with high winds the diurnal variations of the coefficient of exchange are smaller.

The annual variations of the coefficient in moderate latitudes reach the greatest values in summer because of the increasing roughness and great superadiabatic gradient of temperature during daylight hours.

In winter the coefficient of exchange diminishes considerably because of the insignificant roughness of the snow surface and frequent inversions during daylight hours.

3. On large water reservoirs (especially on oceans) the coefficient of exchange in the lower layer of the atmosphere depends mainly upon wind speed, since the vertical gradients of temperature are usually relatively small. Besides the wind speed, the coefficient of exchange, in this case, is also affected by the form of the underlying surface (waves), but, this influence is comparatively insignificant.

Considerable diurnal and annual variations of the coefficient, and also some changes caused by a series of variable factors, make it very difficult to use the mean values of the coefficient in calculations of turbulent heat exchange by formula (36), and especially so for shorter periods.

In those cases where it is possible to determine the value of the exchange coefficient, and the vertical gradient of temperature is known, the value of the turbulent heat flux can be calculated by formula (42). However, since available results of the gradient measurements are very scarce, use of formula (42) in determining the turbulent heat flux cannot be of any great importance in climatological calculations.

It is much easier to use formula (37) for determining the turbulent heat exchange from data of numerous observations, since this formula contains a characteristic integral of the vertical turbulent transfer between the underlying surface and the atmosphere D (the coefficient of the outer diffusion) instead of the coefficient of exchange k.

The coefficients D and k are directly related to each other. After integration of equation (36) by z, between the level of $z=0$ (where $\theta=\theta_w$) and the level z with temperature θ, we will obtain the formula:

$$P=\frac{\rho c_p(\theta_w-\theta)}{\int\limits_0^z \frac{dz}{k}}, \qquad (51)$$

and resulting from this:

$$D=\frac{1}{\int\limits_0^z \frac{dz}{k}}. \qquad (52)$$

But it should be indicated here that value k differs from coefficient D by its comparatively small dependence on height. Elementary estimates have shown that beginning with the height, on the order of one meter, the change of level z by several times would change coefficient D only by a few per cent.

The diffusion method for calculating the turbulent heat exchange, which is connected with formula (37), is used in modern investigations for water reservoirs, and for land.

Let us examine first a somewhat more complicated problem of determining the turbulent heat exchange on land by the diffusion method.

In this case, considerable difficulties arise in connection with the necessity to estimate two parameters in formula (37) - the coefficient of outer diffusion D and the temperature of the underlying surface θ_w.

To determine coefficient D for land, two basic methods could be used. The first is based on equation (52), i.e., on utilization of the relationship of the outer diffusion coefficient and the exchange coefficient.

When performing the integration of the denominator in formula (52) it should be kept in mind that, the turbulent exchange regularities near the underlying surface, i.e., with very small values of z, are very complicated ones.

The authors of the majority of earlier theoretical researches on heat exchange and evaporation have used different hypotheses, about the form of function $k(z)$ with small z, for determining the integral $\int\limits_0^z \frac{dz}{k}$. This was done without any substantiative proof.

So, for instance, in the work of M.E. SHvets (1941 [234]) it was assumed that the coefficient of diffusion for heat and moisture transfer is determined by formula:

$$k=k_1(z+z_0), \qquad (53)$$

where z_0 — is the aerodynamic roughness.

In other investigations (Dorodnitsyn, 1941 [94]; SHvets 1943 [235]; Liûtershtein and CHudnoskiĭ 1946 [164], and others) the following model was used:

$$k=k_1z+k_0 \qquad (54)$$

(k_0 — coefficient of molecular diffusion), i.e., it was assumed that, at the surface the value of the exchange coefficient reaches that of the coefficient of molecular diffusion, which is usually considerably smaller than value k_1z_0.

In addition, some authors, when using models (53) or (54), assumed that, at the surface a thin laminar sublayer exists in which the diffusion has a molecular character; Millar (1937) and Montgomery (1940) also assumed that, over the sea surface the height of the laminar sublayer is determined by the Karman laws.

The results of numerous observations of the vertical distribution of wind speed in the air layer near the ground showed that, the prevailing majority of natural surfaces on land are "rough," which means that the transfer of turbulent friction to the underlying surface occurs not through the sublayer by molecular viscosity (as in smooth tubs), but directly by the elements of roughness (irregularities of soil surfaces and plants), upon which local gradients of pressure are dependent.

Consequently, it might be assumed that, in natural conditions, an aerodynamical laminar sublayer with molecular conductivity for heat and moisture diffusion, practically does not exist.

In order to solve the problem about diffusion in the proximity of natural rough surfaces an experimental investigation was carried out (1947). The temperature of a bare underlying surface, and the air temperature at several levels was measured by a thin resistence thermometer.

The ratio of differences in temperature between the surface and a height of 150 cm ($\theta_w-\theta_{150}$) and differences in temperature between 55 and 150 cm ($\theta_{55}-\theta_{150}$), obtained by measurements taken during daylight hours, are shown in fig. 10 by dots.

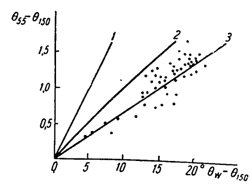

Figure 10

The relationship between the difference in temperature at two levels and the difference in temperature between soil surfaces and the air.

Lines 1 and 2 show the relationship between values $(\theta_w - \theta_{150})$ and $(\theta_{55} - \theta_{150})$ according to formulas (53) and (54). As can be seen in this case, from the location of the dots, both formulas do not fit, since jumps of temperature between the underlying surface and the air, that have been predicted by these schemes, turned out to be considerably extenuated. The distribution of dots in fig. 10, could be described, on the average, by line 3, which fits the value of relationship:

$$\frac{\theta_w - \theta_{150}}{\theta_{55} - \theta_{150}} = 15. \qquad (55)$$

Hence it should be concluded that, the use of formulas (53) and (54) for determining the relationship between values of the exchange coefficient and the coefficient of diffusion may lead to considerable errors. Instead of using these equations it is expedient to utilize the empirical relationships, which are analogous to formula (55).

Considering the fact that the following relationship could be derived from formulas (37) and (42):

$$D = \frac{\theta_1 - \theta_2}{\theta_w - \theta_2} \frac{k_1}{\ln \frac{z_2}{z_1}} \qquad (56)$$

(θ_1, and θ_2 — air temperatures at heights of z_1 and z_2). We will assume that:

$$D = ak_1, \qquad (57)$$

where a — is the coefficient which depends on the boundary conditions of heat exchange of the underlying surface.

Proceeding from the experimental data given above, we find that the value of $a = \frac{1}{15}$. Other experimental results, that have been obtained during daylight hours over various kinds of surfaces without vegetation, usually provide values of a that vary between the limits of $\frac{1}{10} - \frac{1}{20}$.6)

For night periods with temperature inversions, in the air layer near the ground, coefficient a grows considerably in comparison with the daylight hours' values and, on the average, it exceeds these values by 3 - 5 times.

Although, determining coefficient a in each individual case is usually a difficult matter, nevertheless, formula (57) could still be used for estimating the mean values of coefficient D.

Other methods for determining the coefficient of outer diffusion, by analogy with methods for calculating the coefficient of exchange, are based on measurements of heat and moisture streams. So, if data of evaporation rates are available, coefficient D could be determined by the following formula:

$$D = \frac{E}{\rho} \frac{(\theta_1 - \theta_2)}{(\theta_w - \theta_2)} \frac{1}{(q_1 - q_2)}, \qquad (58)$$

6) Determination of coefficient a is connected with considerable methodical difficulties because of the necessity to measure the temperature of the underlying surface θ_w. Utilization of mercury thermometers or various kinds of electric thermometers with sensing parts that were not thin enough, resulted in considerable errors in determining θ_w, and subsequently, in the calculations of value a. But the most complicated problem is the measurement of temperature of underlying surfaces with vegetation.

and, if data on the measurement of radiation balance and on heat exchange in soil are on hand, the following equation can be used:

$$D = \frac{R - A}{\rho c_p (\theta_w - \theta) \left[1 + \frac{L}{c_p} \frac{(q_1 - q_2)}{(\theta_1 - \theta_2)} \right]}. \qquad (59)$$

By application of these relationships, a series of data on the coefficient of external diffusion on land has been obtained.

Proceeding from available data the following conclusions concerning the principal properties of the coefficient of external diffusion can be derived

1. The external diffusion coefficient varies only slightly with the change of the level at which the temperature (or moisture) are measured but, only when this level is higher than 1 m.

2. The mean values of the external diffusion coefficient, during daylight hours on land, are on the order of 1.0-1.5 cm/sec. With inversions the external diffusion coefficient decreases, in comparison with its values at superadiabatic gradients of temperature, but the dependence of the coefficient of external diffusion upon thermal stratification is markedly weaker than the analogous dependence of the exchange coefficient (this is effected by certain changes in the conditions of heat exchange on the underlying surface, and reflected in corresponding changes of coefficient a).

The mean daily values of coefficient D are markedly smaller than those for daylight hours, and for the warm season on land they are about 0.6-0.7 cm/sec.

In dry regions the coefficients of external diffusion are usually somewhat higher than those in moist areas.

3. On large water bodies the coefficient of external diffusion changes considerably, depending on changes in wind velocity. On land, the mean values of this coefficient depend on wind speed to a much lesser degree since, in the first place, the variability of the mean wind speed on a great portion of land is relatively low, and in the second place, a reduced wind velocity observed at superadiabatic gradients results in a greater effect of the thermal factors on turbulent exchange, which partly compensates the reduction of the external diffusion coefficient (Budyko, 1947, 1948a /38 & 39/ and others).

A lesser variability of the external diffusion coefficient on land, as compared with the variability of the exchange coefficient, facilitates the utilization of its mean values in climatological calculations. However, in calculations of heat exchange in water reservoirs for a long period, it is necessary to even take into account the dependence of coefficient D on wind speed.

Besides difficulties encountered in determining coefficient D, when calculating the heat exchange on land by formula (37), there will also be some difficulties in determining the temperature of the underlying surface θ_w. The concept of the underlying (or active) surface, which was established much earlier by A.I. Voeikov (1904 /69/), appears to be very complicated for land surfaces. In this case, especially so for vegetation, the underlying surface actually presents a layer of considerable thickness, and inside this layer, more or less, rapid changes of temperature, and other meteorological elements, take place in a vertical direction.

The only, more or less, reliable method for determining the temperature of the underlying surface under various conditions, is the use of the radiation thermometer, i.e., a method for calculating temperature from measurements of the change in the long-wave radiation flux. This method has been successfully employed in a series of researches but, because of technical diffiuclties it is not very popular.

Among other methods for measuring the values of θ_w, the use of thin resistance thermometers has brought some satisfactory results. These thermometers were placed on a, more or less, leveled soil surface without vegetation.

A lesser degree of accuracy was obtained by measuring the temperature of the underlying surface with mercurial thermometers placed on the soil surface. Some researches (for instance that of Zubenok,/104/1947) have shown that even on the bare soil, the readings of the mercury thermometers could have considerable errors in measurements of temperature of the underlying surface. However, it must be kept in mind that those errors that arise from use of the mercury thermometers are to a certain extent of a systematic nature, which facilitates the estimation of these errors in determining the heat exchange values.

In one of the investigations on heat balance (Budyko, 1947 /38/), an approximate method was suggested for calculating the sums of turbulent exchange from data of standard meteorological observations that make use of surface temperature measurements with mercurial thermometers.

This method takes into account the correlation (found by L.I. Zubenok, 1947 /104/) between the temperature gradients in the air layer near the ground and temperature differences between the soil surface (with grass) and the air, measured with mercurial thermometers.

This relationship for daylight hours is represented by formula:

$$\Delta\theta_1 = 17\Delta\theta, \qquad (60)$$

where $\Delta\theta_1$ — is the difference in temperature between the soil surface and air at a height of about 2 m; $\Delta\theta$ — the difference in air temperature at various heights, the natural logarithm of the ratio of which equals 1.

From equations (42) and (49) we find that:

$$P = 0{,}144\rho c_p \Delta\theta\Delta u\left[1 + c\frac{\Delta\theta}{(\Delta u)^2}\right], \qquad (61)$$

where $c = 1$ m²/sec² degree; $\Delta\theta$ and Δu — differences in temperature and wind speed in a vertical direction in the lower air layer at various heights, the natural logarithm of the ratio of which equals 1.

Considering the relationship between vertical gradients of temperature and wind speed, it can be concluded that for the average wind speeds at superadiabatic temperature gradients the value of the turbulent heat flux changes only slightly with the change in wind speed. This permits an approximation of formula (61):

$$P = 0{,}48(\Delta\theta)^{1,2} \quad \mathrm{cal/cm^2/min} \qquad (62)$$

which is satisfactory for use with mean values of wind speed and for a broken order of several centimeters.

Substituting for value $\Delta\theta$, the difference between the temperature taken with a mercurial thermometer (placed on the natural soil surface between the plants) and the air temperature $\Delta\theta_1$, from equations (60) and (62), we obtain an approximate equation for turbulent exchange:

$$P = 0{,}96(\Delta\theta_1)^{1,2} \quad \mathrm{cal/cm^2/hour} \qquad (63)$$

In calculations of turbulent exchange from data of standard observations, we can usually employ value $\Delta\theta_1$ only for observations taken at 1 p.m.; in calculations of the total heat exchange during all daylight hours (during the period with superadiabatic lapse rates) the regularities of diurnal variations in the turbulent heat flux should be taken into account.

According to calculations by L.I. Zubenok (1946 /103/), the diurnal variation of turbulent heat exchange during daylight hours is well described by a simple sinusoid equation.

Accordingly, the daily total of turbulent heat exchange can be determined by formula:

$$P_1 = \frac{2}{\pi} T_g P_m, \qquad (64)$$

where P_1 — is the daily total of the positive (directed upward) turbulent stream of heat, P_m — the highest daily value of the turbulent heat stream, T_g — duration of the positive heat exchange during 24 hrs. (duration of the period with a superadiabatic lapse rate).

Assuming that the highest value of the turbulent heat exchange differs only slightly from the magnitude of heat exchange at 1 p.m., we obtain the following formula for calculating the sums of the positive turbulent heat exchange from formulas (63) and (64):

$$P = 0{,}61 T_g (\Delta\theta_1)^{1,2} \quad \mathrm{cal/cm^2}, \qquad (65)$$

where T_g is in hours.

The amount of negative turbulent heat exchange during a more or less longer period, is usually considerably smaller than the amount of positive heat exchange. The cause of this phenomenon (which is usually called the ventil effect), is seen in the fact that when the stream of turbulent heat is directed downward, which happens in temperature inversions, the turbulent exchange is reduced considerably. And conversely, when the stream goes upwards, the superadiabatic stratification of temperature reinforces the turbulent mixing. This results in a considerable difference between the mean values of turbulent streams, which are directed from the earth's surface to the atmosphere and those which go from the atmosphere to the earth's surface. 7)

In calculations of the sums of turbulent heat exchange, negative values of the same can be roughly evaluated, since these values are relatively small. So, for instance, in the author's work (1947 /38/) the annual sum

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7) More details about the ventil effect and its significance for some processes in the air layer near the ground are given in § 10.

of the turbulent flux's negative amounts (observed in diurnal variations) was assumed to be approximately equal to 20% of the annual amount of heat exchange for the southern section of European Russia.

The outlined method for the climatological calculation of turbulent heat exchange sums has been verified by the equation of heat balance (see § 6), which rendered quite satisfactory results. It must be noted that the application of this method is only possible for long period averages; for short periods, however, use of the approximated formulas (62), (64) and especially (60) may lead to considerable errors. On the other hand, this method is apparently of little value for calculating heat exchange in wooded regions where the relationship shown by formula (60) is of a very conventional character. Subsequently, this method can not be regarded as basically universal for determining the sums of turbulent heat exchange between the land surface and the atmosphere.

Among other ways of climatological computations of turbulent heat exchange, the idea of M.I. I͡Udin deserves noteworthy attention. I͡Udin suggested a determination of the heat stream magnitude from the amplitude of diurnal variations in temperature (I͡Udin 1948 /243/). Developing this idea, M.I. I͡Udin obtained interesting results, however his method has not been worked out into a computation scheme which would be convenient for a larger practical utilization.

In climatological calculations, the values of turbulent heat exchange on land are mostly determined by solving the equation of heat balance.

The simplest way to do this is to determine the heat exchange as the remainder of the terms of balance, i.e., according to formula:

$$P = R - LE - A. \qquad (66)$$

This method, which already has been used in many investigations, gives satisfactory results in those cases when value P is not very small in comparison with the principal terms of heat balance (mainly as compared with radiation balance). In those cases, however, when the turbulent stream is much smaller than the radiation balance, the method of the "remainder term" may lead to large relative errors. Even errors in the sign (i.e., of direction) of the turbulent flux may arise here.

Other methods of calculating the turbulent heat exchange on land, based on the equation of heat balance, and calculations of the underlying surface temperature will be analyzed in § 5 together with the description of methods for calculating losses of heat for evaporation.

Now let us go over to the problem of determining the turbulent heat exchange between the water surface and the atmosphere.

Climatological calculations of turbulent heat exchange over water bodies are, as a rule, much simpler in comparison with those for land, because of the possibility of using standard water surface temperature observations.

The diffusion method for calculating the turbulent heat exchange, based on an application of formula (37), may be used for determining the amount of turbulent exchange from data of the many standard observations. The only prerequisite is that the magnitude of the coefficient of diffsuion D, dependent on meteorological factors, be evaluated.

The method of determining this coefficient for the water surface was examined by Sverdrup (1936a) in an investigation of methods for determining the evaporation from water surfaces.

For this purpose Sverdrup integrated formula (38) by z, assuming that in the lower turbulent air layer the coefficient of exchange depends on the height, according to formula: $k = k_1(z + z_0)$

(where z_0 — is roughness), and over the sea surface there exists a thin sublayer with thickness d, in which the coefficient of exchange is equal to the coefficient of molecular diffusion ($k = k_0$).

After integration of the indicated relationship by z from 0 to z, in the layer of turbulent diffusion, the following equation was obtained:

$$E = \frac{\rho k_1 (q_0 - q)}{\ln \frac{z_1 + z_0}{z_0}} \qquad (67)$$

(q — specific humidity at height z_1, q_0 — specific humidity at the upper boundary of the sublayer of molecular diffusion).

Integrating formula (38) by z, inside the sublayer of molecular diffusion, we obtain the equation:

$$E = \frac{\rho k_0 (q_s - q_0)}{d} \qquad (68)$$

(q_s — is the specific humidity of saturated water vapor with respect to the vaporizing surface temperature).

By eliminating q_0 from (67) and (68), and substituting k_1, according to formula (47), Sverdrup obtained the equation for evaporation from ocean surfaces:

$$E = \frac{\rho k_0 u (q_s - q)}{ud + \frac{k_0}{\varkappa^2} \ln^2 \frac{z_1 + z_0}{z_0}}, \qquad (69)$$

where u — is wind velocity at height z_1.

From these relationships the following equations for determining D and P could be obtained:

$$D = \frac{k_0 u}{ud + \frac{k_0}{\varkappa^2} \ln^2 \frac{z_1 + z_0}{z_0}} \qquad (70)$$

and

$$P = \frac{\rho c_p k_0 u (\theta_w - \theta)}{ud + \frac{k_0}{\varkappa^2} \ln^2 \frac{z_1 + z_0}{z_0}} \qquad (71)$$

Formula (71) was initially found by P.P. Kuz'min (1938 /145/). To make use of formulas (69) and (71), two parameters must be defined: roughness of the sea surface z_0 and thickness of the sublayer of molecular diffusion d.

Sverdrup assumed that in accordance with a well-known aerodynamic relationship, which was derived from tubes with rough surfaces, the magnitude of the sea surface roughness is 1/30 the mean height of the surface irregularities (waves). Sverdrup estimated the value of d by using data of his observations on Spitzbergen and observations of Wüst on sea surfaces; on the basis of these results he defined the average value of d as equal to 0.10-0.15 cm.

Afterwards, Rossby (1936) showed that, actually the clue to the roughness of the sea surface depends only slightly on wave heights, and therefore the roughness should be taken as being constant for various rates of wind speed. Furthermore, in many works (Sverdrup, 1937, 1940, 1946 etc.) this formula was improved a great deal, however, the relationships found in these works have not been used extensively in calculations of evaporation and of turbulent heat exchange in water reservoirs.

This apparently, was associated to a certain degree with the fact that the schemes of diffusion in the lowest air layer, which have been used by Sverdrup and also later by the other authors, were, to a considerable degree, of a speculative character, and it was extremely difficult to verify them by any somewhat reliable experiments.

Therefore, the majority of the subsequent authors preferred to use a simple relationship for calculations of turbulent heat exchange:

$$P = c_p a u (\theta_w - \theta), \qquad (72)$$

where a is the coefficient of proportionality independent of wind speed and defined by the equation of heat balance or in any other way.

It must be indicated that the supposition about the independence of coefficient a from the wind speed does not imply any substantial qualitative differences in formulas (71) and (72), since the relationship given in formula (71) is comparatively weak.

The magnitude of coefficient a, according to data of most recent studies, is on the average, close to $2.4 \cdot 10^{-6}$ g/cm^3 (provided that measurements of wind speed and air temperature were made at standard heights used in ship observations). This value was, for instance, obtained in the Central Geophysical Observatory (Budyko, Berliand, Zubenok, 1954a /51/), as the mean value for the world's oceans. It must be noted here that, the computation of coefficient a, by using the Sverdrup formula, gives on the average, a figure that is very close to that indicated above.

It is quite conceivable that, for separate regions of the world's oceans and for various water reservoirs, the value of coefficient a might be slightly different.

There are some data which indicate that this coefficient is somewhat smaller for the inner seas (Caspian and Aral) than its average values for open oceans.

However, the majority of scientists assumed that, the possible variations of coefficient a are comparatively small and can be neglected in approximate calculations of the turbulent heat exchange and evaporation.

Oftentimes, for the calculation of turbulent heat exchange in water reservoirs, some formulas based on the so-called "Bowen's relationship" are used, which can be obtained in the following way.

Integrating formula (38) by z, we find a relationship similar to equation (37):

$$E = \rho D (q_s - q), \qquad (73)$$

where q_s — is the specific humidity of saturated air at the water surface temperature.

From (37) and (73) we find that:

$$\frac{P}{LE} = \frac{c_p (\theta_w - \theta)}{L (q_s - q)}. \qquad (74)$$

This relationship is usually applied in the form of:

$$\frac{P}{LE} = 0{,}46 \frac{\theta_w - \theta}{e_s - e} \frac{B}{760}, \qquad (75)$$

where e_s and e — are vapor pressure; B — air pressure in mm.

This formula permits a calculation of the amount of turbulent exchange from data on evaporation. In those cases when these data are missing, but data on radiation balance and on internal heat exchange are available, the following equation obtained from formulas (4) and (74) can be used for computing the turbulent heat exchange:

$$P = \frac{R - A}{1 + \frac{L (q_s - q)}{c_p (\theta_w - \theta)}}, \qquad (76)$$

The turbulent heat exchange in water reservoirs can also be defined by formula (66); however, its application is often complicated by difficulties in obtaining a sufficiently accurate estimate of the internal heat exchange value A, for larger reservoirs or for portions thereof.

In summarizing the results we should point out that, for climatological calculations of turbulent heat exchange on water reservoirs, it is usually most expedient to utilize formula (72), which permits a determination of the heat exchange value entirely from data on water surface and air temperature and on wind speed.

For determining turbulent heat exchange on land, in climatological calculations, the equation of heat balance has been frequently used, and turbulent exchange has been computed as the remainder term of the balance. In comparison with this, another method for determining turbulent heat exchange on land by using the equation of heat balance has some advantages. This method is described in § 5 in connection with methods for determining the loss of heat for vaporization.

§ 5. The loss of heat for evaporation

The loss of heat for evaporation is equal to the product of the latent heat of vaporization times the amount of evaporation. The latent heat of vaporization under natural conditions changes somewhat with variations of temperature of the vaporizing surface according to formula:

$$L = 597 - 0{,}60\theta \text{ кал/г}, \qquad (77)$$

where θ — is the temperature in °C.

In many climatological calculations a constant value of latent heat of vaporization can be used. It is approximately equal to 0.6 kg-cal/gr.

Many different methods are in use for determining evaporation under natural conditions. Analyzing these methods we will first examine the comparatively more complicated problem of determining evaporation from land surfaces.

One of the oldest methods for determining evaporation was based on the application of evaporimeters of various construction, which can be divided into two groups: evaporimeters with maximum moistening, and those with an

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isolated monolith. As a sample of an evaporimeter with maximum moistening the evaporimeter of Dorandt can be mentioned here. This apparatus was suggested for use in the second half of the last century. It consists of a metal cyclindrical container with soil connected by a tube to a water tank which keeps the soil of the evaporimeter in a state of maximum moistening. The speed of evaporation from Dorandt's evaporimeter was determined by the descent of water in the tank.

Some authors presume that the evaporimeters with maximum moistening show the maximum speed of evaporation from the underlying surface,which is possible under given meteorological conditions.

This is not quite so, since the evaporimeter, installed in the midst of a relatively dry soil, presents a limited vaporizing surface, and the speed of evaporation from this surface under conditions of turbulent diffusion of water vapor depends on its size and cannot show the speed of evaporation from an unlimited wet surface under the same meteorological conditions. Therefore, the speed of evaporation registered by Dorandt's evaporimeter, which when installed, for instance, in a Middle Asian desert, will exceed considerably that rate of evaporation from the underlying surface which would have been observed after maximum moistening of the surrounding land.

Since the data obtained with the evaporimeters of maximum moistening could not be used for determining actual evaporation from the land surface, evaporimeters with an isolated monolith without any additional moistening replaced them at the beginning of the 20th century. The well-known evaporimeter of Rykachev (1898 /199/),which consists of an open metal box inserted tightly into a case that is installed in soil,may serve as a sample of an apparatus of this type. A soil monolith with vegetation is inserted in this box, the box is weighted and installed into the case. The summarized speed of evaporation from soil and vegetation is determined by subsequent weighings of this box at certain time intervals. For the weighings the box is taken out from the case.

At present we have several constructions of Rykachev's evaporimeter, they differ mainly in the form and size of the box.

Some authors have indicated (Oppokov, 1934 /181/; Kuzin, 1938 /141/, and others) that because of the insulation of the monolith, which is inserted in the evaporimeter's box, from the surrounding ground, substantial distortions of the evaporation conditions will arise.

When precipitation is excessive,water in the monolith cannot percolate down to the underground water and is accumulated in the evaporimeter; under droughty conditions, conversely, the insulation of the monolith from the ground water results in a rapid desiccation of soil in the evaporimeter.

Distortion of the water exchange conditions in the insulated monolith must, undoubtedly, result in noticeable errors in determining the amount of evaporation; however, as was found by M.P.Timofeev (1952), the scale evaporimeters may present some greater errors in measurements of evaporation even without these distortions.

This conclusion is derived from the principal postulates of the elementary theory of scale evaporimeters.

The scale evaporimeters of the usually applied construction, in spite of popular notions, can not be regarded as absolute instruments, i.e., instruments measuring evaporation directly. From general considerations it is seen that the scale evaporimeter measures only changes in moisture content in that monolith which is inserted in the evaporimeter. Therefore, when using this evaporimeter and weighing it at several days intervals, for determining evaporation, we must have at hand data on precipitation and on surface and underground runoff.

For those periods when infiltration and runoff are small as compared with precipitation (for instance, in climates of insufficient moistening during the warmer season), it is usually assumed that evaporation during the given period is:

$$E' = r + \delta w', \qquad (78)$$

where r — is the amount of precipitation, $\delta w'$ — change of moisture content in the soil monolith inserted in the evaporimeter.

Actually, for these conditions the amount of evaporation must be:

$$E = r + \delta w, \qquad (79)$$

where δw — is the change in water content in the whole layer of the active moisture exchange in soil.

Because of this the relative error in determining the amount of evaporation with the scale evaporimeter will be:

$$\frac{E - E'}{E} = \frac{\delta w - \delta w'}{r + \delta w}, \qquad (80)$$

and during the period without precipitation:

$$\frac{E - E'}{E} = \frac{\delta w - \delta w'}{\delta w} \qquad (81)$$

Thus, to estimate the principal error which arises when using the scale evaporimeter, it is necessary to compute the ratio of moisture content change in the layer of soil with a thickness equal to the depth of the evaporimeter, to the moisture content change in the whole layer of the active moisture exchange.

It must be emphasized that the indicated principal error is not directly related to the distortion of natural conditions for moisture exchange in a separated monolith and it would take place even with a very frequent change of the monolith. 8)

To estimate the magnitude of the principal error of Rykachev's evaporimeter, with a depth of 500 mm, we give here data of calculations made, partly, by E.H. Romanova for various natural zones of the USSR.

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8) To demonstrate the reality of this statement we give here the following sample of reasoning. Let us assume that we are using this evaporimeter during a rainless period with soil which dried up down to the level of this evaporimeter. Then, further loss of moisture would occur from the deeper layers of soil. It is obvious that in this case the weighing evaporimeter would give absurd results no matter how often we change the monolith.

A similar situation will always arise when a considerable loss of moisture from deeper soil layers (in comparison with the depth of the evaporimeter) is effected by the process of evaporation.

The relationships between changes in moisture content in soil under summer wheat in the layer of 500-1000 mm, to the change of moisture content in the layer 0-500 mm, were determined by these calculations. Average data for the steppe zone, for wooded steppe and deciduous forest, and for the forest zone (coniferous and mixed forest) are given in table 9.

Table 9

The relationship between changes in moisture content of soil under summer wheat in the layer 500-1000 mm. and the change of moisture content in the layer 0-500 mm. (in per cent).

	From the last decade of April to the last decade of August	From the last decade of April to the first decade of June	From the first decade of June to the last decade of August
Steppe Zone (21 stations)	97	89	123
	From the first decade of May to the last decade of August	From the first decade of May to the second decade of June	From the second decade of June to the third decade of August
Wooded Steppe (13 stations)	102	75	177
	From the last decade of April to the last decade of August	From the last decade of April to the second decade of June	From the second decade of June to the last decade of August
Forest Zone	28	0	58

The data in table 9 permit us to conclude that, in steppe and in wooded steppe zones, as well as in the forest zone during the second half of summer, the moisture content change in the layer 500-1000 mm is quite comparable to

the change in the layer of 0-500 mm, and is often greater. A similar deduction could be also derived from the analysis of observations on soil moisture dynamics under a field of winter wheat.

It must be particularly noted here that, under conditions of deficient moistening (steppe, wooded steppe), considerable changes of moisture content in the lower soil layer under summer wheat are also observed in May and at the beginning of June, when the root system of plants does not yet reach the deep soil layers.

Figures shown in table 9 permit us to conclude that, in steppe and wooded steppe zones, as well as in the forest zone during the second half of the summer season, the change of moisture content in the layer between 500-1000 mm is quite comparable to the change in the layer of 0-500 mm and often is even larger. A similar conclusion could also be derived from the analysis of data on observation of the soil moisture dynamics under winter wheat.

It must be noted, however, that under conditions of deficient moistening (steppe, and wooded steppe), considerable changes in moisture of the lower soil layer under summer wheat are also observed in May and at the beginning of June, when the root system of plants does not yet reach these deeper levels.

From these data it can be concluded that, the use of scale evaporimeters with a depth of 500 mm will lead to certain errors in measuring evaporation under conditions of insufficient moistening. These errors, which mainly lower the measurements of the evaporation amount during rainless periods, will, as data in table 9 show, reach a very great magnitude which will be quite comparable to the exact amount being measured. For the period with more or less abundant precipitation, the relative error of evaporation measurements will be markedly lowered, as is seen from the formulas given above. However, the error of the evaporimeter as an instrument, measuring changes of moisture in the active soil layer, will not be diminished and the accuracy in determining evaporation will be higher only because of the fact that, in the calculation of evaporation the change in moisture measured with a considerable error will be added up with the amount of precipitation, and the latter can be measured without any great principal error. In this case, the error caused by determining evaporation from measurements made with the scale evaporimeter and with a rain gage will be so much greater, since the amount of measured evaporation is greater in comparison with measurements of the rain gage.

Thus, the scale evaporimeter of limited depth (for instance of 500 mm) cannot be regarded as a universal instrument which permits measurement of total evaporation in various climatic regions. Such evaporimeters can possibly give more or less reliable results only for sufficiently moist soil and in the absence of a large root system.

To improve the method of soil evaporimeters the following suggestions have been made.

In studies of V.P. Popov (1928, 1929 [190 & 191]) and others, a design of a scale evaporimeter with a netted bottom is elaborated. The net permitted an improvement of the moisture exchange conditions between the soil monolith in the evaporimeter and the lower layers of soil.

A more radical way of improving the soil evaporimeter was used by specialists of the State Hydrological Institute (V.A. Uryvaev, 1953 [223] and others), who designed an evaporimeter containing a monolith of very large dimensions. Since it was difficult to weigh the heavy monoliths with the usual scales, this was done by means of a hydraulic transmission, which permits an accurate measurement of relatively small changes in weight of heavy bodies.

By using hydraulic evaporimeters we can obtain some comparatively accurate data on evaporation, though use of this instrument also has some limitations (especially during periods of deep moisture infiltration).

In evaluating the existing results of observations obtained with soil evaporimeters, it must be indicated that at the present time, some deductions of a climatological character can be mainly derived from data obtained with Rykachev's evaporimeters, which were used for several years at a few stations in the USSR (Results of observations on evaporation . . ., 1939, and others).

Taking into account that the accuracy of these data is not high, and also rather limited, it must be acknowledged that, for climatological computations of heat losses for evaporation, the data obtained with evaporimeters have, so far, only a limited value.

Among other methods of determining evaporation that are based on use of data obtained from special observations, the gradient methods, which are analogous to the corresponding ways of determining turbulent exchange, must be pointed out (see § 4).

The most reliable method for determining evaporation from gradient observations is the method of heat balance, which is based on the following formulae. From the equation of heat balance (4) and formulae (36) and (38) we find that:

$$R = -L\rho k \frac{\partial q}{\partial z} - \rho c_p k \frac{\partial \theta}{\partial z} + A. \qquad (82)$$

By solving equations (38) and (82) together, we obtain:

$$E = \frac{R-A}{L + c_p \dfrac{\frac{\partial \theta}{\partial z}}{\frac{\partial q}{\partial z}}}, \qquad (83)$$

or, after integration by z :

$$E = \frac{R-A}{L + c_p \dfrac{\theta_1-\theta_2}{q_1-q_2}}, \qquad (84)$$

where $\theta_1 - \theta_2$ and $q_1 - q_2$ — are differences in temperature and in specific humidities at two different heights.

The latter relationship, which is analogous to formula (76), permits us to determine with sufficient accuracy, the amount of evaporation for periods

of time with values of the principal terms of heat balance that are not too small, i.e., mainly for daytime in the warm season, and also for the mean daily (or ten-day means and monthly means) of the warm season. 9) It must also be kept in mind that, in calculations made by this formula it is often possible to either neglect value A , or take it into account in a very rough approximation, since for a more or less longer period of averaging, this value is usually considerably smaller than R. This simplifies the procedure of determining the amount of evaporation.

At the present time, formula (84) is used for determining evaporation from data obtained by expeditions and by special stationary observations (see Franssila, 1936; Albrecht, 1940; Trudy expeditsii GGO and many others). The application of equation (84) to climatological computations of evaporation is hampered by the very limited amount of existing data on gradient and balance observations.

A similar situation exists in relation to the gradient diffusion method of determining evaporation.

This method, based on formula (40), has been used lately in many investigations.

W. Schmidt suggested determining the evaporation from the vertical turbulent stream of water vapor (see Schmidt, 1917, 1925, 1935). Elaborating on this idea Thornthwaite and Holzman (1939, 1942, and others), used formula (40) and the equation of Rossby-Montgomery for the coefficient of turbulent exchange.

In order to determine differences in specific humidities ($q_1 - q_2$), Thornthwaite and Holzman constructed special instruments which were used for detailed observations on evaporation for almost a year.

As has been established in my works (1946b, 1948a [36 & 39]) these calculations of evaporation, which were accomplished by Thornthwaite and Holzman, contained substantial errors because of the fact that the formula of Rossby-Montgomery did not take into account the effect of thermal

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9) In works of many investigators a large error arose in calculation by formula (84), since they used mean daily or monthly values of the differences $\theta_1 - \theta_2$ and $q_1 - q_2$.

Because of the fact that the daily range of these values is rather large, and their mean relationships with evaporation are very different, it is obvious that, by averaging these data in this manner, large errors arise in determining the amount of evaporation.

To verify this assumption the author compared the results of annual evaporation calculations for two points, by formula (84) making use of the mean values of $\theta_1 - \theta_2$ and $q_1 - q_2$ from hourly observations, and for the mean annual values (in works of 1946b, 1948a [36 & 39]). In the second case the amount of evaporation appeared as being twice as large as that in the first case. This confirmed the impossibility of applying mean values of temperature and humidity gradients to calculations of evaporation by formula (84).

stratification on turbulent exchange, and consequently, gave much lower values of the average exchange coefficients.

Much more accurate results of evaporation calculations by the diffusive method can be obtained by using formula (49) for computing the exchange coefficient. This method has been used by many authors (Budyko, 1948a /39/; Grigor'eva, 1949 /85/; Ginsburg, 1949 /79/; Rusin, 1952 /198/; and others).

We will not go into detail in explaining this method, but one remark must be made to the effect that, at the present time it is used, as well as the method of heat balance, mainly for generalizing data of special stationary or expeditionary observations.

For determining mean amounts of evaporation from land, the method of water balance is preferred when compared with gradient methods.

For the average annual period, according to equation (18), evaporation can be determined by formula:

$$E = r - f, \tag{85}$$

i.e., as the difference between the amount of precipitation and runoff.

This way of determining mean annual amounts of evaporation has been largely used in many investigations. The caluclations of evaporation by water balance permitted the design of quite a few maps of evaporation and loss of heat for evaporation (Kuzin, 1934, 1940, 1950 /140, 142 & 143/; Budyko, 1947 /38/; Troitskii, 1948 /220/, and others) and gave extensive data of mean annual amounts of evaporation.

It must be remembered that calculations of evaporation by formula (85) give the most reliable results for comparatively large surfaces - on the order of thousands and scores of thousands of square kilometers.

For surfaces of more limited dimensions calculations by formula (85) may lead to noticeable errors because of difficulties involved in the accurate evaluation of the moisture redistribution by the underground runoff.

For shorter period averages, in calculating annual variations of evaporation, for calculating evaporation of single years and months, etc., it is feasible to use, instead of formula (85), a more general relationship:

$$E = r - f - b, \tag{86}$$

which is obtained from equation (17).

Since it is very difficult to determine the value of the complete change in moisture content of the lithosphere's upper layers b, and since this value is quite comparable to the amount of evaporation E, calculations by formula (86) will not always give sufficiently reliable results for indicated periods.

The simplest way is to use equation (86) for determining evaporation of the warm season under conditions of insufficient moistening, when changes in moisture content b are determined mainly by dynamics of moisture in the upper soil layers.

As we know, observations of moisture content in the upper layers of soil are taken regularly at many agrometeorological stations of the USSR, and therefore, for the zone of insufficient moistening, it is possible to calculate the amount of evaporation by formula (86) making use of the ample data of hydrometeorological observations.

It should be remembered however, that data on soil moisture are spatially very inhomogenious, and it is necessary to resort to considerable averaging of the initial results when calculating evaporation from climatological data.

Since, for a considerable portion of the land, no reliable data on runoff could be found, the water balance method cannot be regarded as a universal one for determining evaporation from land, not even for average annual amounts. For shorter averaged periods, the use of this method is still more limited.

Because of this, quite a few investigations have been concerned with the development of methods for determining evaporation from the land surface by merely using data on observations of ordinary meteorological elements: precipitation, temperature and humidity of the air.

Let us examine first the climatological methods of calculating evaporation for average annual conditions.

In the work of Wundt (1937), by generalizing the results of calculations of evaporation made by the water balance method, an empirical relationship between the annual amount of evaporation, and amount of precipitation, as well as the mean annual temperature, was established. The nomogram of Wundt, which was constructed on the basis of this relationship, was used in some hydrological researches, particularly by M.I. L'vovich in calculating the normals of runoff on various continents (1945 /162/).

Verification of Wundt's nomogram shows (Budyko, 1951b /46/) that, noticeable errors occur in calculations of evaporation. These errors are associated to a certain extent with the fact that Wundt used the mean annual temperature as the index of the thermal effect on evaporation (it is well known that, mean annual temperatures in moderate and higher latitudes depend a great deal on temperatures of the cold season, whereas evaporation is determined almost entirely by temperatures of the warmer season).

A somewhat more rational index of the thermal regime was chosen by B.G. Ivanov, who also constructed an empirical diagram showing the relationship between the annual amounts of evaporation and precipitation, as well as the saturation deficit (1940 /111/).

In comparison with these purely empirical methods, genetic methods based on certain physical concepts are of greater importance in determining annual amounts of evaporation.

The first genetic method for climatological calculations of evaporation was suggested by E.M. Ol'dekop, who substantiated the formula showing the dependence of evaporation on precipitation and evaporability (Ol'dekop, 1911 /180/). In my works (1948a, 1948b /39 & 40/), on the basis of a joint analysis of heat and water balance equations on the land surface, the "relation equation" was obtained - the dependence of evaporation on precipitation and on radiation balance, which presents, to a certain extent, a generalization of Ol'dekop's equation.

This equation shows the relation between the ratio of the mean annual evaporation and precipitation $\frac{E}{r}$, and the ratio of the radiation balance and heat amount that is required for the vaporization of the annual amount of precipitation $\frac{R}{Lr}$, i.e.,

$$\frac{E}{r} = \Phi\left(\frac{R}{Lr}\right), \tag{87}$$

where Φ — is a definite function.

A detailed analysis of the "relation equation" will be given in § 10.

We will only present table 10 here, which was compiled from the "relation equation" that permits computation of evaporation from data on precipitation and radiation balance.

Table 10

The relations between the precipitation, evaporation and radiation balance.

$\frac{R}{Lr}$	0.10	0.20	0.30	0.40	0.50	0.60	0.70	0.80	0.90
$\frac{E}{r}$	0.10	0.20	0.28	0.35	0.44	0.50	0.56	0.62	0.66
$\frac{R}{Lr}$	1.00	1.20	1.40	1.60	1.80	2.00	2.50	3.00	
$\frac{E}{r}$	0.70	0.76	0.82	0.86	0.88	0.90	0.94	0.97	

It should be mentioned that the value of the radiation balance R, which is included in this equation, indicates the potentially possible evaporation (evaporability) and therefore must be computed for a moist surface in the given land (see chapter IV). Value R is close to the actual values of radiation balance in a more or less humid climate, but in dry climates it might be substantially greater than the indicated values. In approximate calculations of evaporation by means of the "relation equation," the mean latitudinal values of R can be largely used, since interlatitudinal changes of these values are comparatively small in the major portion of the plain land (except in areas with monsoon climate and some other coastal regions).

The problem of calculating evaporation from data of ordinary meteorological observations for periods of a few months or scores of days is much more complicated than that of determining annual amounts of evaporation.

In one of the first investigations concerning this problem, P.S. Kuzin suggested determining the evaporation from land, under conditions of an excessive moistening, by a method that is similar to that for calculating evaporation from a water surface. In his papers (1934, 1938 [140 & 141]) Kuzin used the following formula for determining evaporation:

$$\bar{E} = ad \text{ cal-cm}^2\text{/min} \tag{88}$$

where d — is the saturation deficit in mm, a — coefficient, approximately equal to 14 - 15.

In one of the indicated investigation (1938 [141]) P.S. Kuzin also suggested an empirical formula for determining the evaporation of individual months under deficient moisture conditions.

In the development of climatological methods for calculating evaporation, the works of B.V. Poliakov (1946, 1947 [187 & 188]) are regarded as important contributions; he was the first to use the numerous data on the dynamics of moisture in the upper soil layers for determining evaporation. This permitted him to find empirical relationships between evaporation and meteorological factors under conditions of excessive moisture as well as deficient moistening.

For this computation B.V. Poliakov constructed graphs which show the amounts of evaporation dependent on mean air temperature and precipitation for each month of the warm season. For the cold period Poliakov suggested a graph which showed the evaporation for individual geographical regions to be dependent on air temperature. In a paper published in 1947, Poliakov recommended the use of corrective coefficients for some regions when determining evaporation from those graphs. It must be noted that, the graphs of Poliakov are now largely in use for determining the evaporation from land.

A verification of the evaporation amounts calculated from Poliakov's graphs, by comparing them with numerous data on water balance has shown that, annual amounts of evaporation computed for various climatic zones of the USSR from Poliakov's graphs are, as a rule, in good agreement with the annual amount of evaporation calculated from the water balance.

Some less satisfactory results have been obtained from verifying the method of Poliakov, as applied to calculating monthly amounts. Here the calculations from Poliakov's graphs were compared with calculations of water balance, by taking into account changes in the moisture content of the upper layer of soil 1m deep, in cultivated fields, from data of numerous agrometeorological stations. It should be noted that the determination of evaporation from the water balance, for a single month, is of course, not free from some errors (data on dynamics of moisture in cultivated fields cannot be quite typical for the given region as a whole. Some errors might arise because of difficulties involved in estimating the rate of infiltration, etc.). However, since the method of Poliakov has been substantiated in the same way (but only from comparatively sparse data), it is obvious that, for verifying it, use of the indicated data is quite legitimate.

As the calculations show, in most cases the annual variation of evaporation calculated from Poliakov's graphs differ from those calculated by the water balance method. The graph gives noticeably extenuated values of evaporation in summer and somewhat exaggerated values for spring and autumn. When summarized for the whole year these discrepancies are usually compensated to a certain extent, and that is why annual amounts of evaporation obtained by Poliakov's method are well in keeping with those obtained by using water balance calculations.

Among other deficiencies of Poliakov's method it must be mentioned that the principle of accounting the dependence of evaporation on time is not the most appropriate one.

If in the scheme of Poliakov, the dependence of evaporation on time would be of a physical nature, then obviously, the calculated amounts of evaporation for certain months should not change very much with the initial dates of the periods selected for calculations.

However, the verification has shown that for many points the amounts of evaporation calculated by Poliakov's method for certain months, are only slightly changed by some tenths of a per cent, depending on the selection of periods for which the initial data were averaged and the evaporation calculated. For instance, the amount of evaporation in May, as calculated from Poliakov's graphs, can differ greatly in case we calculate it for the period from May 1 to June 1, from that which was calculated by averaging the results of calculations made for the periods April 15-May 15 and May 15-June 15, taking into account the annual march.

Touching on other particular deficiencies of Poliakov's method, we only notice that this method cannot be applied to most of the tropical and subtropical regions, because of the limited range of changes which are accounted for by the scheme of calculation based on meteorological elements (temperature, precipitation).

All this provides for a conclusion about the inadequate effectiveness of the method suggested by B.V. Poliakov, although, it seems to us that in some cases it is of possible use for some approximate evaporation calculations.

Several climatological methods for calculating annual variations of evaporation from land have been suggested by F. Albrecht (1950).

Some of these methods are based on formula:

$$E=(e_s\alpha-e)f(v), \tag{89}$$

where e_s — absolute humidity of saturated air at the land surface temperature; e — absolute humidity of the air; $f(v)$ — a function depending on wind speed; α — parameter indicating the effect of the properties of an underlying surface on reduction of evaporation (Albrecht calls it "the water covered portion of land").

In formula (89) α will be equal to 1 for an absolute moist surface (and then this formula is transformed into a well-known relationship for evaporation from water surfaces), and for a partly dry surface $\alpha<1$.

Formula (89) is deemed as not quite sufficient, since it involves an obviously incorrect dependence of evaporation on air humidity. Indeed, if the soil has partly dried up and evaporation was reduced in comparison to that from the water surface, α must be less than one. Let us assume, for instance, that $\alpha=0{,}7$, then, according to Albrecht's formula, the evaporation (under conditions of an isothermal stratification of temperature in the air layer near the ground) must equal zero at the relative humidity of 70%, and when the humidity increases above 70% condensation must begin. It is obvious that, in natural conditions nothing of the kind is ever observed.

It is very characteristic that Albrecht did not succeed, as he admits, in finding a relation between the parameter α and moisture conditions of soil, although, according to the very meaning of the problem, this parameter should have taken into account the changes of soil moisture in the first

place. Instead of this Albrecht found an empirical dependence of the values α on the mean monthly relative humidity. This relationship has no physical sense and does not eliminate the controversy indicated above; when relative humidity changes during a short period (for instance in daily variations) then, according to formula (89), evaporation must be replaced by condensation, even when the relative humidity in the air layer near the ground is considerably lower than 100%.

Among other methods for the climatological calculation of evaporation that were suggested by Albrecht, we will mention only the method of heat balance, which is based on the use of equations:

$$E=\frac{1}{L}(R-P-A), \tag{90}$$

$$P=(\theta_w-\theta)f(v), \tag{91}$$

where L — latent heat of vaporization, P — turbulent heat exchange, A — heat balance in soil, θ — air temperature, θ_w — soil surface temperature.

The application of this method (which is, to a certain extent, similar to that used by the author in an investigation in 1947 [38]) is hampered by the difficulty of interpreting the available results of observations, obtained from the station network, on soil surface temperature. Moreover, when calculating evaporation by equation (90) a rather large relative error has often occurred, since the difference of values R and $(P+A)$ is in many cases not very great as compared with value R.

The third method for the climatological calculation of evaporation that has been suggested by Albrecht is based on the application of the equation of water balance. Albrecht suggests, as did the author of this work (1950b [49]), for this purpose, to solve the equation of water balance for certain periods by taking into account the dependence of the relationship of evaporation and evaporability on characteristics of soil moisture. However, the form of the latter dependence is incorrectly presented by Albrecht, according to my opinion, which results in serious defects in this method of calculation.

Not touching any details in Albrecht's reasoning, we will only notice that, under conditions of excessive soil moistening he obtains the following formula for the speed of evaporation:

$$E=r+aE_0,$$

where r — precipitation, E_0 — evaporation, a — coefficient equal to 0.5.

This relationship is physically unjustified and leads to an absurd conclusion that, in case of an excessive moistening, the evaporation will approach the evaporability only if precipitation is equal to one half of the evaporability.

Consequently, it must be assumed that calculations of monthly amounts of evaporation which were done by Albrecht with the water balance method (including the constructed maps of monthly evaporation for Australia), are not sufficiently justified and may have considerable errors.

Among other climatological methods for calculating the annual march of evaporation the proposition made by N.A. Bagrov (1954a [17]) deserves

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attention. He recommended as applicable for this purpose, a generalized formula of Ol'dekop which uses a rather complicated method of indirect accounting for moisture conditions in a given month (or decade). We must remember that, E.M. Ol'dekop himself tried to apply the formula that he had found, for calculations of the annual march of evaporation. However, the difficulties in accounting for moisture conditions of individual months did not allow him to obtain satisfactory results. Bagrov's proposition presents of course a certain step forward in comparison with Ol'dekop's calculations; however, it is desirable in order to evaluate this proposition, to verify it against empirical data on dynamics of soil moisture in its annual march.

The widest prospects for determining evaporation from land for months and decades are opened by genetic methods of calculation, which are based on a direct estimation of all components of water balance for the analyzed periods of time, also including changes of moisture content in soil. In creating a rational method for calculating evaporation, a great importance will be also attributed to a direct estimation of the influence of solar energy balance, as the most significiant factor in the evaporation process.

In the author's paper (1950b /44/) a method for calculating the annual march of evaporation was suggested. It was based on a simultaneous solution of equations of the water and heat balances. This method was used in many calculations, but its cumbrous and complicated formulas limited the possibilities of a wide application.

As it turned out later, it was possible to simplify the scheme designed for calculating evaporation, and to simultaneously provide for greater detailization.

Let us outline the fundamentals of this simplified method of calculating the annual march of evaporation in the most condensed form.

P.S. Kossovich initially established that, the process of evaporation from the soil surface is characterized by several different stages. In the first stage, when a considerable amount of moisture is available in the soil, the speed of evaporation, as has been established by numerous experimental investigations, does not depend on soil moisture and is basically determined by external meteorological factors. Results of investigations of A.M. Alpat'ev (1950, 1954 /11 & 12/) are of considerable importance. He proved that, on fields with cultivated crops, the first stage of evaporation is observed inside a rather large range of soil moisture variations. According to Alpat'ev, when soil moisture is not lower than 70-80% of field capacity, evaporation from crop fields is close to the value of evaporability, and consequently, mainly dependent on meteorological factors.

With the drying out of soil, starting with the critical moisture of soil, evaporation passes to the second stage when the speed of evaporation rapidly decreases with the diminishing of soil moisture. There are many experimental data (F.A. Kolîasev, 1939 /131/ and others) showing that in this case, the speed of evaporation from soil surfaces depends on the content of moisture in soil and this relationship could be considered as being very close to linear.

An analogous conclusion could be also derived from the analysis of experimental data obtained by S.I. Dolgov (1948 /93/) that show the evaporation from soil covered by vegetation. According to these results, in the second stage of evaporation, the speed of evaporation also diminished

approximately in a linear proportion to the decrease of soil moisture. Since evaporation from soil covered by vegetation becomes insignificant when it reaches the wilting point, then subsequently in this case, the speed of evaporation can be assumed as being proportional to the amount of productive moisture in soil.

Thus, we may conclude that, at soil moisture w, which is larger than a certain critical value $w_к$, total evaporation E depends mainly on meteorological factors and is equal to evaporability E_0. When productive moisture in soil diminished below $w_к$, evaporation becomes less than evaporability, whereupon the amount of evaporation is proportional to the quantity of productive moisture, i.e., $E = aw$. Insofar as $E = E_0$, $w = w_к$, it is obvious that $a = \frac{E_0}{w_к}$.

On the basis of the above mentioned principles, we will apply the formula for computation of evaporation, when $w \geqslant w_к$:

$$E = E_0, \tag{92}$$

and when $w < w_к$:

$$E = E_0 \frac{w}{w_к}. \tag{93}$$

To make use of the latter formula it is necessary to have at hand data on soil moisture, which can be obtained by calculating the equation of water balance:

$$r = E + f + w_2 - w_1, \tag{94}$$

where $w_2 - w_1$ are the differences in moisture of the upper soil layer, between the beginning and end of the analyzed single period.

If we assume that mean soil moisture for the analyzed period is:

$$w = \frac{w_1 + w_2}{2},$$

then, if $w < w_к$:

$$E = E_0 \frac{w_1 + w_2}{2w_к} \tag{95}$$

From (94) and (95) we obtain:

$$w_2 = \frac{1}{1 + \frac{E_0}{2w_к}} \left[w_1 \left(1 - \frac{E_0}{2w_к} \right) + r - f \right] \tag{96}$$

Using the formulae (92), (94), (95), and (96) we can calculate the annual march of evaporation for various climatic conditions.

We will give here a sample of the usual procedure of calculations. Let us assume that in a chosen region the amount of soil moisture after the spring snow melt is equal to the field capacity. Since, in this case $w > w_к$, we will calculate the evaporation for the first period (a month or decade) by using formula (92). If now, according to the water balance equation, at the end of the period the soil moisture does not diminish below $w_к$, then we may continue the calculations by formula (92) until the quantity of productive moisture diminishes below the critical point. After that, formula (95) should be used for calculating evaporation, and formula (96) for computing the moisture change, assuming that, soil moisture at the end of the preceding period w_2 is equal to soil moisture at the beginning of the following period w_1. For the season with snow cover the evaporation can be obviously determined by formula (92).

The outlined method for calculating the annual march of evaporation requires the knowledge of the following values: precipitation, runoff, evaporation, critical soil moisture, and also soil moisture at the beginning of one of the calculated periods, which is needed as the initial value to start from, since all the following amounts of moisture can be calculated by using the water balance method.

Data on precipitation and runoff are usually obtained from the numerous observations of the hydrometeorological network.

The problem of determining evaporability will be analyzed in detail in §10. Here we will examine briefly only the determining of the two last parameters of the outlined scheme of calculation.

The values of critical moisture in soil $w_к$ can be determined from results of various experimental investigations (for instance, the already mentioned work of A.M. Alpat'ev), and also by using indirect methods.

Data obtained by indirect methods 10) have shown that the mean value of $w_к$ for the upper soil layer of 1m depth varied mainly in the limits between 70 and 250 mm. These variations are associated in a certain way with geographical zonality.

In calculating the annual march of evaporation, it was possible in many cases to achieve a higher accuracy in determining parameter $w_к$ by comparing calculated annual values of evaporation with the difference between precipitation and runoff.

The last parameter of the calculation scheme - the initial soil moisture - can be determined, in some cases, from factual data (S.A. Verigo, 1948 [65], and others). It was very often possible to obtain this value by general reasoning; for instance, in many regions of temperate latitudes after the completion of snow melt, or at the end of the rainy period in the tropical and subtropical regions, soil moisture can be regarded as being close to the value determined by the field capacity of soil (data on field capacity of various soils are given in papers of A.V. Protserov, 1948 [193], and others).

In many calculations of the annual variation of evaporation the initial soil moisture conditions can be determined by the selection method, changing its value until the calculated value of moisture content, at the end of the last period calculated in the annual march, coincides with the given value of moisture at the beginning of the first calculated period.

It must be remembered that, the analyzed method (in a more complicated form) was used and verified against experimental data in the author's work (1950b [44]) when determining the mean annual march of evaporation for larger regions. This verification proved that the accuracy obtained by using this method was quite satisfactory. The work of T.G. Berliand (1952 [27]) should be also mentioned here. She used this method for calculating

10) These calculations were partly made by L.I. Zubenok and N.I. Sinitsina.

the dynamics of soil moisture in cultivated fields for single years (moist, droughty), and the results of calculations were checked against factual observations of soil moisture content.

The results obtained from this kind of verification were so good that the method described for determining evaporation can also be of importance for various hydrological, agrometeorological and other kinds of calculations pertaining not only to average conditions, but to single years as well.

The problem of determining evaporation from a water surface is simplified as compared with that from land, by the fact that for water reservoirs we usually have the surface temperature and, consequently, also the saturation point at the surface temperature. Data on water balance of limited reservoirs usually permit a more reliable determination of the annual march of evaporation as compared to land conditions.

Let us first examine the problem of applying evaporimeters to measure evaporation from water reservoirs.

The methods of determining evaporation by means of evaporimeters differ somewhat, in regards to some smaller water reservoirs, from those used for measuring evaporation from seas and oceans. Most investigations on the application of evaporimeters pertain to the solution of the first of these problems, which is of substantial practical importance.

In earlier investigations of measuring evaporation it was assumed that the speed of evaporation from a rather small open vessel, filled with water and placed in the open, is equal to the speed of evaporation from water reservoirs. Therefore, in order to obtain the evaporation from the reservoir the amount of evaporation obtained from the evaporimeter was simply multiplied by the relationship of the water reservoir surface to the surface of the vessel in the evaporimeter. However, it was noticed long ago that, such calculations resulted in exaggerated estimates of the evaporation speed from water reservoirs.

This was explained by differences in physical conditions of evaporation from an isolated evaporimeter and from a water reservoir, and research was directed towards a development and construction of floating evaporimeters, for which it was assumed, it was possible to create such conditions which would be identical with those on the surface of a water reservoir. Numerous observations were taken, and they showed that floating evaporimeters still gave incorrect results, since even with a full preservation of natural conditions of water vapor diffusion, the speed of evaporation differed because of the modification of conditions for heat exchange in the upper water layers (the heat conductivity in water contained in the vessel of an evaporimeter was different from that in the upper layers of water reservoirs, where it is affected by turbulent exchange). Besides, the utilization of floating evaporimeters was often hampered by technical difficulties, particularly so with high winds.

In 1933 Rohwer's paper (1933) was published. It presented a partial account of the Special Commission of the Society of Civil Engineers on the problem of measuring evaporation from limited water reservoirs. Rohwer compared readings of several types of coastal evaporimeters and one floating evaporimeter with evaporation from an experimental basin and established the fact that none of the evaporimeters could give the speed of evaporation identical to that of the basin.

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Another part of the account was actually an article by Follansby (1933), containing the consolidation of a vast amount of observations on evaporation made by evaporimeters from water surfaces in the U.S.A. and other countries.

Further research of methods for measuring evaporation from a water surface by means of evaporimeters was made by V.K. Davydov (1938 /87/ and others) and O.S. Poznyshev (1937 /185/ and others), and also by Hickman (1939), who studied evaporation from the Great Lakes, and Hickox (1944), who tried to utilize the theory of similarity in analyzing conditions of evaporation from evaporimeters.

In a general evaluation of the method of measuring evaporation from water reservoirs, the most important question is about the possibility of regarding the reduction coefficients 11) as consistent characteristics of the same kind of evaporimeters or if they change depending on external conditions.

Experiments made by many researchers have shown that the assumption of the persistency of reduction coefficients was in most cases unjustified, and relative changes of the mean monthly values of the reduction coefficient, for a given evaporimeter in different geographical conditions, on various water reservoirs and in different seasons, could be very great. From all this V.R. Davydov drew the conclusion that, "indeed, there is not and could never be, any constant relationship between the actual amount of evaporation and that observed by instruments..." (1938 /87/).

Apparently, the absence of any consistent relationship between evaporation from water reservoirs and readings of evaporimeters, limits to a great extent the possibility of their application.

In recent years, when difficulties in the application of evaporimeters had been clarified, special basins with diameters of scores of meters were tried for measuring evaporation. It was assumed that the speed of evaporation from such basins equals that from larger natural water bodies under similar meteorological conditions.

It must be noted here that, although the assumption about the independence of evaporation speed from the dimensions of the evaporating surface, when this surface is larger than several meters, is confirmed by some observations, nevertheless, this hypothesis contradicts theoretical deductions on this problem. It could be thought that, even by using some comparatively large basins for calculating evaporation from large natural water bodies some errors would arise, and they would be of a similar nature to those observed using small coastal evaporimeters for calculating evaporation from water reservoirs.

An augmentation of the evaporating surface of the basin would probably diminish, to a certain extent, the systematic error in such calculations, but it cannot be eliminated completely.

The measurements of the speed of evaporation from seas and oceans, by using evaporimeters, was begun in the 19th century by Leidly. After an

11) The reduction coefficient for evaporimeters is a value by which the reading has to be multiplied in order to get true values of evaporation from the natural underlying surface.

unsuccessful attempt by Mohn to use evaporimeters on the sea (they were overrun by waves), the observations were made mainly by the ship's evaporimeters, which were installed on the deck. Initially the speed of evaporation was determined here from measurements of the water volume decrease in the evaporimeter. However, because of difficulties met in preventing errors that arose from spilling water out of the evaporimeter, which occurred with a rolling sea, this method was soon replaced by calculations of the evaporation speed by measurements of changes in the concentration of salt in water contained in the evaporimeter. This method was suggested by Dieulafait (1883).

An original ship evaporimeter was constructed by V.V. SHuleĭkin (1941), who used an open calorimeter filled with sea water, where the heat exchange with the ambient air proceeds only through the evaporating surface. From readings of a thermometer which was inserted in the evaporimeter, the speed of changes in the heat content of water in the calorimeter was determined. This value characterizes the speed of evaporation in those cases, when the algebraic sum of radiation balance of the evaporating surface and the values of turbulent heat exchange of this surface with the atmosphere is small, in comparison with the total heat which is spent for evaporation during a given time interval.

It is obvious that measurements, of any kind, by ship evaporimeters do not show amounts of evaporation from the sea because of an essential difference in evaporation conditions from an isolated instrument as compared with those for water surfaces. In investigations of Wüst (1920), Cherubim (1931) and other authors, some attempts were made to evaluate quantitatively the influence of individual factors which are effecting the difference between readings of evaporimeters and actual evaporation from the sea, with the purpose of obtaining a reduction coefficient for evaporimeters.

Comparing readings of evaporimeters installed under various conditions and also using some empirical relationships, the authors mentioned calculated reduction coefficients for the ship evaporimeters by taking into account the effect of the ship's speed, the toss caused by the vibration of engines, the warming of evaporimeters through their walls, and many other factors.

Because of an essential effect exerted by the body of the ship on evaporation from ship evaporimeters, their reduction coefficients have seemingly, still less stable characteristics than reduction coefficients of floating and coastal evaporimeters.

Besides taking into account the fact that mean values of reduction coefficients are determined on the basis of rather conditional assumptions, it should be concluded that calculations of the distribution of evaporation speed on oceans, which were made by Lutgenson, Wüst and Cherubim from data of ship evaporimeters, are of a low accuracy.

Some more reliable results, in comparison with the method of evaporimeters, are usually obtained by calculating evaporation from water bodies from their water balance.

From equation (17) and (18) we find that, evaporation from water bodies can be determined from data on precipitation, on horizontal redistribution of water and on water level variations. In some cases it was necessary to also take into account the infiltration of water through the bottom of the reservoir, though, usually it is very insignificant in comparison with other terms of the water balance.

The method for determining evaporation from the water balance is mainly adapted for limited reservoirs and for more or less extended periods of averaging. To compute evaporation from the water balance for a part of the reservoir (for instance, for an individual region of the ocean) was usually quite impossible, since in this case it is very difficult to evaluate the horizontal redistribution of water.

To utilize the method of water balance for comparatively shorter periods is also difficult since the accuracy of determining the terms of water balance for shorter intervals is often insufficient.

For the more or less detailed calculations of evaporation from water reservoirs, methods of diffusion and heat balance are most frequently used.

The diffusion method for determining evaporation from water reservoirs is based on an application of formula (73), which was used for a long time in hydrometeorology as an empirical relationship, often called Dalton's Law.

Empirical formulas of this type were usually applied in the form of:

$$E=(e_s-e)f(u), \qquad (97)$$

where e_s — pressure of saturated water vapor calculated from temperatures of the evaporation surface, e — vapor pressure at certain altitudes, $f(u)$ — "wind factor" - empirically determined function of the dependence of evaporation speed on wind velocity u.

For the function $f(u)$, various authors have obtained a series of expressions, which mostly have the following form:

$$f(u)=A+Bu$$

(A and B — numerical coefficients), but sometimes they present an exponential relationship of this type:

$$f(u)\sim u^n$$

with the values of exponent n being between 0.5 and 1.0. Reviews of various experimental investigations of this problem are given in articles by M.Ю. Lur'e and N.M. Mikhailov (1935 [161]), M.K. Baranaev (1938 [19]), Rohwer (1931), Harding (1942) and others. It should be indicated here that numerous experimental investigations, accomplished under various conditions, confirmed very well the direct proportionality of the evaporation speed and saturation deficit computed from water surface temperatures. The differences in the relationship of evaporation to wind speed, which were found in some investigations, are mainly explained by the insufficient comparability of conditions of various evaporation basins, evaporimeters, and laboratory installations, which were used in studies of these relationships.

From the comparison of formulas (97) and (73) it can be seen that the problem of the form of the wind factor function coincides with the problem of the relationship between the coefficient of external diffusion D and wind speed.

As a rule, the temperature lapse rate in the lower air layer over water bodies is noticeably smaller than over land, and therefore, the relationship between the coefficient of exchange and wind speed can be adequately described by the formula of Rossby-Montgomery.

This provides a basis for the assumption that the coefficient of external diffusion over water bodies is proportional to wind speed, i.e.,

$$E=au(q_s-q). \qquad (98)$$

This relationship, which is similar to formula (72), was first obtained by V.V. SHuleĭkin and at the present time is used largely in calculations of heat expenditure for evaporation.

As has been said in § 4, results of some theoretical and experimental investigations have proven that, in determining evaporation value of coefficient a can be assumed to be almost constant and approximately equal to $2.4 \cdot 10^{-6}$ gr/cm^3 (when wind speed and temperature were measured at heights customarily used in ship observations).

In those cases when the water surface temperature of a large water body differs significantly from the air temperature, a need may arise for estimating the effect of thermal stratification on the turbulent exchange and evaporation.

From general considerations it is apparent that thermal stratification most strongly affects evaporation at lower wind speeds, when the Richardson's numbers approach the highest values because of small vertical gradients in wind speed. This conclusion was fully confirmed by results of calculations (Budyko, 1948a [39]), which established that, for medium and high velocities (which are usually observed over large water bodies, the main portion of ocean surfaces included) the effect of the difference in temperature between water and air on evaporation is insignificant.

We will not go into the details of determining evaporation from vast expanses, 12) and we will only note some features of the calculations of evaporation from the surface of relatively smaller, limited water reservoirs.

For determining evaporation from limited water reservoirs, lakes, ponds, etc., it is often necessary to utilize observational data of the principal meteorological elements obtained at coastal stations. In investigations of quite a few authors it was established that even over water reservoirs of several hundred meters in size, wind speed, humidity and air temperature often change because of the transformation of air masses over the water body. These changes usually affect evaporation substantially.

Consequently, a need arises for a quantitative account of transformation processes in determining the amounts of evaporation.

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12) Investigations of V.S. Samoĭlenko (1952 [205]) deserve special attention in reviewing the methods for determining evaporation from oceans.

In researches by Jeffreys (1918), Giblett (1921), Sutton (1934), D.L. Laikhtman (1947a [153]), M.I. IUdin (Budyko, Drozdov and others, 1952 [56]), N.I. IAkovleva (1952 [245]), A.P. Braslavskii and Z.A. Vikulina (1954 [33]) and other authors, various methods were worked out for determining evaporation from water reservoirs by taking into account the transformation of the airflow.

Results of these investigations and deductions from empirical investigations by R.R. Davydov (1944 [88]), B.D. Zaikov (1949 [102]) and others, permit, at the present time, the calculation of evaporation from various water reservoirs of limited extent.

Together with the diffusion method for determining evaporation from water reservoirs, the heat balance method has also been largely used in modern investigations.

As early as 1915, W. Schmidt calculated the radiation balance for the latitudinal zones of the ocean, and made an attempt to determine the amount of heat that is spent on evaporation in these zones. On the basis of very rough assumptions Schmidt derived that the ratio of heat expenditure for evaporation to the radiation balance in various latitudes was equal to 0.4 - 0.8. Later, some authors indicated that these estimates were not accurate, and Schmidt also partially agreed with this statement.

In investigations by Ångström (1920), who calculated the heat balance of one of the Swedish lakes, it was deduced that the ratio of the amount of heat that is spent on evaporation to the amount spent on evaporation and turbulent exchange equals 0.9. This estimate was utilized by Mosby (1936) who determined the latitudinal distribution of evaporation from oceans, using the assumption that for evaporation in all zones, an amount of heat corresponding to 91% of the radiation balance is spent.

Possibilities for a reliable computation of evaporation from water reservoirs by using the heat balance method have been expanded considerably by the introduction of the "Bowen relationship" into calculations of evaporation, i.e., formulas (74) and (75).

By using this relationship and the equation of heat balance, evaporation (or heat spent on evaporation) can be calculated from data on radiation balance, water surface temperature and humidity and temperature of the air.

The formula commonly used is:

$$LE = \frac{R - A}{1 + \frac{c_p(\theta_w - \theta)}{L(q_s - q)}}, \qquad (99)$$

where θ_w — water surface temperature, q_s — specific humidity of saturation vapor in respect to surface temperature.

This method for determining evaporation has given accurate results, as has been found by numerous verifications, and especially so for longer periods. Calculations made for shorter periods were, as a rule, less accurate, since in this case it is necessary to determine the value of the inner heat exchange A, whose determination is associated with considerable difficulties.

Summarizing this discussion on methods for determining loss of heat due

to evaporation it must be noted that, for land surfaces it is more rational to use (for climatological calculations of the annual value of evaporation) the water balance method or the "relation equation." For determining evaporation for individual months or ten-day periods it is expedient to apply the method of solving the water balance equation by taking into account the dynamics of soil moisture.

For climatological calculations of heat losses due to evaporation from water reservoirs it is usually more convenient to apply the simple equation (98), i.e., the SHuleikin formula.

Besides, for this purpose it is often possible to use those methods which are based on equations of the heat and water balances of the reservoir.

In conclusion we will briefly outline the methods for determining the other components of the heat balance.

As a rule values of these terms are considerably smaller than the principal components of heat balance - radiation balance, expenditure of heat on evaporation, and turbulent heat exchange. Therefore, in calculating these terms it is often possible to use some simplified methods, which permit an approximate estimate of their values.

We will analyze first the question of determining the heat flux between the active surface and the deeper layers of soil or water.

The heat flux (heat exchange) between the underlying surface and the deeper layers can be calculated for land conditions, from changes in heat content of the upper soil layers, whose temperature changes in the considered period of time.

Such calculations can be easily accomplished, provided that, data on soil temperature at various depths of the heat exchange layer, and the heat capacity characteristics of soil are available.

In this case, if during the analyzed period some freezing or melting of a considerable amount of soil moisture occurs, term A will be equal to the sum of the change in heat content and the gain (or loss) of heat from the freezing or melting of the soil's water. An approximation of the latter value can be determined as the product of the latent heat of freezing, soil moisture, and the difference in depths of the 0° isotherm at the beginning and end of the analyzed period.

To determine the heat exchange in soil, when data on temperature are only available for a limited depth, it is necessary to summarize the change in heat content of the upper soil layer and the value of the vertical heat flux between this and the deeper layers. The value of the indicated flux can be calculated as the product of the coefficient of temperature conductivity of soil and the vertical gradient of temperature of the corresponding level (CHudnovskii, 1948 [232]; Laikhtman and CHudnovskii, 1949 [157]; TSeitin 1951 [228] and others).

The calculations of heat exchange in soil are considerably more complicated by the presence of a snow cover, since the heat capacity and heat conductivity of snow range in very wide limits, depending on its density. Usually little data are available on vertical distribution of temperature in a snow cover. Nevertheless, at present, we have obtained rather large results on heat exchange in soil for the warmer season and partly for the period with snow cover as well. Some deductions from these data can be used for approximate evaluations of the variations in heat exchange of soil in their annual course.

From the general physical consideration it is clear that annual variations of heat exchange are closely connected with the annual range of air temperature. If the annual range of temperature is inconsiderable, then apparently, the mean monthly values of heat exchange in soil must also be close to zero.

Indeed, results of heat exchange calculations have shown that when the annual range of temperature is less than 10-15° (C) the monthly sums of heat exchange are comparatively small and could be often ignored in the approximate calculations of heat balance.

This means that heat exchange in soil is not of substantial importance in the monthly sums of heat balance for the majority of tropical regions, and also for many areas with a maritime climate in moderate latitudes.

In regions of the Northern Hemisphere with a considerable annual amplitude of temperature, the annual march of term *A* has, on the average, characteristics presented in table 11.

Table 11

Annual march of heat exchange in soil.

Jan.	Feb.	March	Apr.	May	June
-0.23	-0.15	0.08	0.15	0.23	0.23
Jul.	**Aug.**	**Sep.**	**Oct.**	**Nov.**	**Dec.**
0.19	0.12	-0.08	-0.12	-0.19	-0.23

The values of heat exchange in table 11 are given partially by the sum of the warm (or cold) period, during which the direction of heat exchange is not changed.

The sum of heat exchange for the warm (or cold) season actually represents half of the annual range. The annual range of heat exchange, as calculations have shown, depends closely on the annual range of air temperature.

According to available data, this dependence could be expressed, on the average, by the following quantitative indices.

Table 12

The relationship between the annual ranges of air temperature and heat exchange in soil.

The annual range of temperature (degrees)	10	15	20	25	30	40	50
Annual range of heat exchange in soil, cal/cm^2	1.8	2.8	3.7	4.6	5.5	7.4	9.2

Using tables 11 and 12 one can obtain, from data on the annual range of temperature alone, some approximate estimates of the annual march of heat exchange in soil. Data obtain in this manner will show average conditions in a large region (for each specific point the exchange in soil shows great microclimatic variations associated with differences in the thermal quality of soils).

The calculation of heat exchange in water reservoirs is generally much more complicated than that of heat exchange in soil because of the substantial influence exerted by the horizontal distribution of heat in various parts of the reservoir and, because of the absence of systematic observations of water temperature at various depths for the majority of reservoirs, including seas and oceans.

Because of this, for certain portions of the sea, ocean or other water bodies, it is usually possible to determine the value of heat exchange only as the remainder term of the heat balance equation. The mean annual value of this term will also show the redistribution of heat as affected by currents and horizontal heat conductivity.

For those reservoirs which have sufficiently numerous reports of temperature at various depths, it is possible to determine the change in heat content. Such calculations, in comparison to land conditions, will be simpler because of the practically constant thermal indexes of water, but on the other hand, for shallow reservoirs the calculation could be substantially complicated by the necessity to account for heat exchange between the water body and the underlying ground, which usually is a very difficult task.

The problem of determining heat exchange in designed water reservoirs will not be examined at this time. Some references to this subject are found in articles by M.M. Bernadskii and B.V. Proskuriakov (1931 /30/), B. V. Proskuriakov and D.N. Bibikov (1935 /192/) and others.

When determining the components of heat balance, we should briefly mention calculating the loss (or gain) of heat from the melting or freezing of snow and ice and estimating the redistribution of heat associated with atmospheric precipitation.

The transition of water from the solid into the liquid phase on land surfaces, as well as the converse process, usually effects a considerably smaller loss or gain of heat, as compared with annual sums of the principal components of heat balance (some exceptions could be found in regions with more or less constant snow or ice cover).

However, for monthly sums of the balance components for the corresponding periods, in many regions of moderate and higher latitudes, the estimate of this component will have an essential importance.

Methods of such calculations are based on determining the quantity of melted or frozen water at the surface. The latter value is then multiplied by the latent heat of freezing, which is approximately equal to 80 cal/gr.

In cases when no data of direct observations on the quantity of melted or frozen water are available, some schemes could be used to calculate it. Some of these schemes are based on solving the heat balance equation. A detailed description of such schemes is found, particularly in articles by Sverdrup (1936), P.P. Kuz'min (1947, 1948, 1950 /147, 148, 149/) and others. A redistribution of more or less significant quantities of heat by precipitation can be mainly noticed in tropical regions, where during the rainy period the precipitation of comparatively cold water can effect a perceivable cooling of the underlying surface. The method for calculating this component of heat balance is very simple; its value equals the product of the amount of precipitation and the difference in temperature between the precipitated water and the underlying surface. This method, however, is difficult to use because of the absence of an adequate amount of data on temperature of precipitation.

Sample calculations of this component for tropical regions are available in papers by F. Albrecht (1940 and others).

We will not examine methods for calculating other heat balance components of lesser significance, and it will only be mentioned here that in § 12 the problem of determining losses of heat used in photosynthesis by plants will be considered. Though this component is, as a rule, much smaller than the principal components of the balance equation, still, for some biological and physicogeographical problems its evaluation is of exceptional importance.

§6. The accuracy of determining the components of heat balance

The methods described above, for climatological calculations of the heat balance components have a certain degree of accuracy.

Although, the problem of the accuracy in calculating the heat balance components is very important for the climatology of heat balance, up to recent times it was only scarcely discussed in scientific literature. Concerning this, we have encountered some contradictory viewpoints regarding the probable error of the examined data. In some cases there was a tendency to extenuate the value of this error, whereas in others, its value was extremely exaggerated without any sufficient reason.

All this calls for a more or less detailed analysis of the accuracy of climatological calculations of heat balance components. It is of great interest, in this case, to compare the accuracy of heat balance data with that of climatological data of principal meteorological elements.

According to investigations by the author and N.A. Efimova (1955 [57]), it may be pointed out that, for evaluating the errors of existing methods for calculating the components of radiation and heat balance, three different ways can be used:

1) comparison of results of various independent methods for calculating the components of heat balance.
2) comparison of calculations of the balance component to direct measurements taken with special instruments.
3) evaluation of the calculation error of all balance components by closing the equation of balance with independent determination of all its components.

The third way deserves special attention. It is based on the law of preservation of energy. This method of verifying the calculations of the balance components was used by the author in (1946b[36]) and in a series of ensuing investigations.

When using the second of the above cited methods for verification, it must be remembered that modern methods of instrumental measurements of the balance components are also connected with a noticeable possible error.

The possible error of measurement of the total radiation with the

Savinov-Ianishevskii pyranometer, for more or less longer periods, may reach 5%.

The question about the error in measuring radiation balance with balance meters of various construction was investigated by T.V. Kirillova and Kucherov (1953 [129]). Taking into account their results, it can be deduced that the possible error in measuring radiation balance for more or less longer periods will be on the order of 5-10%. Measurements of other components of heat balance (in particular the expenditure of heat on evaporation and turbulent exchange) by means of special instruments, are of a smaller significance for the verification of climatological calculations of heat balance, since these measurements are either insufficiently accurate or they pertain to very short periods of time.

Let us now turn to the evaluation of the accuracy of calculations of the principal components of heat and radiation balance.

The main component of radiation balance - the total radiation - is recorded by many actinometric stations, and this considerably facilitates the verification of indirect calculations of this value.

The world maps of total radiation were constructed in the Central Geophysical Observatory by using the methods outlined above. A comparison of the calculation results with the consolidated data of mean measured values, which was compiled by T.G. Berliand (1954 [28]) showed that, discrepancies between measured and calculated data constituted, on the average, for monthly values 10% and for annual values 5% (for all available stations - Budyko, Berliand, Zubenok, 1954a [51]). It is quite evident that a certain portion of these discrepancies is due to short periods of observation. Taking this circumstance into account, it is conceivable that, even a very schematic method of calculating total radiation will permit a determination of mean monthly and annual values of this index with an error of the same order as the error of instrumental measurements. It is, however, advisable to remember that, for more or less accurate calculations of total radiation, for comparatively short periods, and particularly for calendar periods, the method of these calculations must take into account, in a direct way, the effect of clouds of various levels on radiation amounts.[13]

As a sample of total radiation verification we present in fig. 11, results of a comparison of calculations of mean monthly radiation values $(Q+q)'$, obtained by using the method outlined above, with mean monthly results of observations of these values $(Q+q)$ at 12 sites of the Soviet Union (Sverdlovsk, Riga, Minsk, Kuibyshev, Saratov, Odessa, Yakutsk, Irkutsk, Alma-Ata, Vladivostok, Tbilisy, Tashkent). The mean discrepancy between measured and

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13) In investigations of the Central Geophysical Observatory the influence of various forms of clouds on radiation was taken into account, but only indirectly, which was necessitated by the fact that only general data on cloudiness could be used in constructing world maps. As it has been shown by investigations of many authors, a direct account of the cloud distribution at various height levels permits a considerably higer accuracy in calculating total radiation, especially for short periods.

calculated values in this case was 9%.

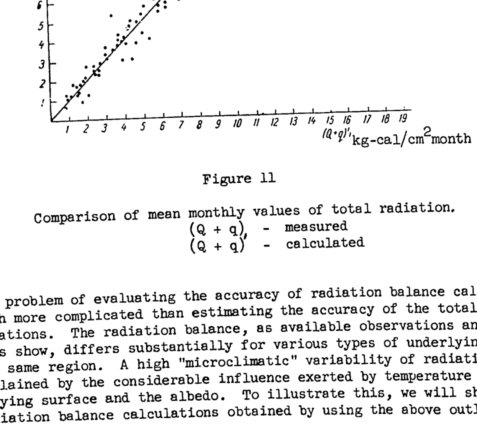

Figure 11

Comparison of mean monthly values of total radiation.
$(Q + q)$, - measured
$(Q + q)'$ - calculated

The problem of evaluating the accuracy of radiation balance calculations is much more complicated than estimating the accuracy of the total radiation calculations. The radiation balance, as available observations and calculations show, differs substantially for various types of underlying surfaces in the same region. A high "microclimatic" variability of radiation balance is explained by the considerable influence exerted by temperature in the underlying surface and the albedo. To illustrate this, we will show results of radiation balance calculations obtained by using the above outlined meth-

od for three regions: Cape Schmidt (Arctic), Sverdlovsk (forest zone) and Ashkhabad (desert zone).

In the first region, calculations of radiation balance were made for the tundra zone, with a bright graveled soil, and for the sea surface with ice floes covering a large portion of the surface, even in summertime.

In the Sverdlovsk region calculations were made for the meteorological station site, for the clover field, and for fallow soil.

In the Ashkhabad region radiation balance was calculated for the desert and for an irrigated field.

The results of calculations are given in table 13.

Table 13

Radiation balance in kg-cal/cm^2 month.

Region	Underlying surface	May	June	July	August
Cape Schmidt	Tundra	-	5.0	6.4	3.7
	Gravel soil	-	4.7	4.5	2.5
	Sea	-	2.3	1.9	0.9
Sverdlovsk	Meteorological station	5.5	7.0	6.7	4.2
	Clover field	5.0	5.9	5.6	3.5
	Fallow soil	7.2	8.5	8.2	5.5
Ashkhabad	Desert	5.4	6.7	6.4	6.1
	Irrigated field	8.2	10.4	10.5	10.0

As can be seen from the given data, in the same region values of radiation balance for various underlying surfaces can differ by scores of per cent. This fact must be taken into account in constructing maps of radiation balance, 14) and also in comparing the results obtained from measurements with those obtained by calculations.

14) Values of radiation balance, that are used in constructing maps, must be determined for a typical underlying surface in the given region. If, in the analyzed region, the factors that exert a substantial effect on radiation balance are very variable in space (as, for instance, the albedo during the snow melt period), then the characteristic value of radiation balance must be properly averaged.

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Until recently we had no data on long period observations of radiation balance. Only in very recent years, at some sites of the USSR, were systematic measurements of radiation balance carried out, for the first time, for periods of 2 to 4 years. Although values of radiation balance obtained in this way cannot be considered as showing precise long-period means, their comparison with earlier obtained results of radiation balance calculations for these sites presents a certain interest.

Results of such comparisons are given in figs. 12-15 for the region of: Leningrad (Koltushi, observations in 1947, 1949-1951 worked up by T.A. Ogneva); Tashkent (observations of 1950-1953 worked up by B.A. Aizenshtat); Sverdlovsk (observations of 1950-1952, worked up by M.V. Lileev); and of Minsk (observations of 1948-1951, worked up by L.I. Pisarchik).

Figures 12 - 15

Annual variations of radiation balance in various regions.
1 - observed values
2 - calculated values

The monthly values of radiation balance obtained from observations are shown on these graphs by the solid line 1; by line 2 - results of calculations used for constructing the climatic maps by the Central Observatory. Considering the fact that differences between the results of calculations and those of the measurements are due, not only to calculation errors, but also, to a certain degree to errors of measurements, to microclimatic variations of radiation balance, and to short periods of observation, we think that the agreement of these two independent methods for determining radiation balance values must be acknowledged as being very satisfactory. 15)

Since there were no long period observations available for the ocean surface, verification of the calculated values of radiation balance for the oceans, by using the method of comparison with measured values, was very difficult. We may indicate that one of the recent works by Albrecht (1952) tackled this problem. In this work he compared calculations of the mean values of radiation balance components for some regions of the Indian Ocean with data of short period observations. Albrecht established a good agreement between the independently determined values of radiation balance, however, this deduction can only have a comparatively limited significance.

An indirect verification of radiation balance calculations for the oceans has been done by comparing the maps constructed by individual authors who used different calculation methods. A similar comparison could be particularly done for the world maps that were constructed by the Central Observatory, with recently published maps for the northern hemisphere oceans; compiled by Sauberer and Dirmhirn, 1954. These authors constructed maps for March, June, September and December. The mean relative discrepancy between data of our maps and that of maps designed by Sauberer and Dirmhirn was 14%. The agreement of these independent calculations confirms, to a certain degree, their reliability.

The method of an independent determining of all heat balance components is more suited for the verification of the accuracy of radiation balance on land and water surfaces. It permits an evaluation of the mean error in determining all components of heat balance.

The relative value of the indicated error can be calculated as the ratio of the alegbraic sum of all heat balance components to the sum of their absolute values.

Using results of the author's works (1946b, 1947 [36 & 38]) in which he determined, by independent methods, all the components of heat balance for one point in the United States (Arlington) and for the southern part of European Russia; the mean error of determining the sum of the balance components was calculated, and it was found that, in both cases it did not exceed 5%.

The verification of the accuracy of heat balance calculations for some

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15) Apparently, the differences between measured and calculated values are of a systematic nature, only for the early spring period. It is possible that these differences are associated with different conditions of snow melt on the plots where the observations were taken, and with average conditions of the surrounding surface.

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regions of the ocean, by using this method, is more complicated than that for land surface, because one of the components of balance for oceans (the gain or loss of heat effected by sea currents), is very difficult to determine by direct methods.

Nevertheless, for annual maps of heat balance, constructed by the Central Observatory, it is possible to estimate the upper limit of the mean calculation error from the closing of the balance equation.

This value can be found in the following way.

If the mean summarized error of determining radiation balance, loss of heat for evaporation and for turbulent exchange on oceans, would be greater than the mean value of the gain or loss of heat that was effected by sea currents A, then it would be impossible to determine the latter value as the remainder term of heat balance. In other words, the map of this value, that was computed by the balance equation, would not correspond to the actual distribution of warm and cold currents.

However, since the map constructed by the Central Observatory shows that gains or losses of heat, as effected by sea currents, are in a good qualitative agreement with the distribution of most of these currents, it is clear that the summarized calculation error of the three principal components of the balance is smaller than the mean value of $\bar{A}$.

Thus, the mean relative error in the calculation of heat balance components for oceans is smaller than the relationship of the mean value of A to the sum of the radiation balance mean values, expenditures of heat for evaporation and turbulent heat exchange. Data presented by Budyko, Berliand, Zubenok (1951a) show that this relationship is equal to 23%, and consequently, the mean error of calculations of the balance components is less than 23%.

The calculations of heat expenditures for evaporation from land, as well as from water surfaces, can be verified by the water balance equation.

The method for determining the evaporation and expenditure of heat for evaporation based on the "relation equation" was verified in many investigations.

So, for instance, in the author's research (1951b [46]) it was established that the discrepancy in evaporation calculations made by using the "relation equation," with that based on water balance for river basins which have some data included in the summary by Wundt (1937), proved to be less than 10%.

The mean discrepancy of the "relation equation" calculations with measurement data turned out to be 13%, and for river basins with the runoff coefficient greater than 0.30 it was only 7%. Results of these computations are described in more detail in § 10.

The most complete verification of the evaporation calculations based on the "relation equation" has recently been accomplished by L.I. Zubenok, who made use of all available data on water balance of various continents. The calculations made by L.I. Zubenok have shown that the mean discrepancy in calculations of the annual value of evaporation based on the "relation equation" with those based on water balance is about 10%.

This value is close to the order of the magnitude of error in determining evaporation by the water balance method, and therefore shows a quite satisfactory accuracy of the evaporation calculations by the "relation equation."

Climatological methods for calculating evaporation from limited water reservoirs (particularly for the Caspian and Aral Seas), have been verified against the water balance many times, and therefore we will not touch this subject here.

Much more complicated is the problem of verifying the climatological calculations of evaporation from the ocean surface.

As it is well known, available data on precipitation over the oceans are very inaccurate, which makes it very difficult to use them in verifying the calculations of evaporation. Only the latest map compiled by O.A. Drozdov (1953 [95]) could be considered as being a more or less reliable source in regards to data of precipitation over the ocean. As it has been noted in investigations by Budyko, Berliand, Zubenok (1954a [51]), when the results of O.A. Drozdov are used, there is good agreement between values of evaporation calculated by the water balance equation and those determined in the above described method by using formula (98).

On the basis of these data (outlined earlier in the paper by the author and N.A. Efimova, 1955 [57]), and also considering other available data, we can draw the following conclusions concerning the accuracy in modern methods for calculating the components of radiation and heat balances:

The error of the mean monthly and mean annual values of total radiation, which were obtained by the climatological method calculation, is approximately equal to 5-10%.

The error in calculating radiation balance is apparently somewhat greater. Evaluating its relative value, it must be remembered that annual and monthly values of radiation balance, under certain conditions, approach zero. Therefore, it is expedient to compare the error of the radiation balance calculation for annual values with the amplitude of changes of these values, and the error of monthly values with the maximum value of radiation balance in its annual march.

Calculated thus, the relative error of the radiation balance calculations will apparently be, on the average, not more than 10%. The annual values of heat expenditures for evaporation on land are determined with a mean error on the order of 5-10%. Considering these values of error, we may evaluate the mean calculation error of turbulent heat exchange on land, which turns out to be somewhat greater than the error of the radiation balance calculation.

The calculation errors of the heat balance components over the ocean could be somewhat greater than the same errors on land, however, the order of magnitude of these errors apparently remains the same.

It should be indicated here that, for various regions of the globe, the values of the possible calculation errors of the heat balance components will differ considerably, primarily due to the different reliability of the original data used. The greatest errors will be found in higher latitudes, and also in lesser known ocean areas (especially in the Southern

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Hemisphere). It can be concluded that the possible relative error of some heat balance components, which were obtained by using modern calculation methods, will often be not greater than the possible relative error of climatological data on some principal meteorological elements (for instance, on cloudiness and precipitation). However, available detailed data on heat balance are still considerably inferior to those on such meteorological elements as air temperature, precipitation and others.

Chapter III

Geographical distribution of the components of heat balance

Until recently an investigation on the geographical distribution of the heat balance components of the earth's surface was practically impossible. This was due to the fact that data on their mean values were available for some few points only and their accuracy had not been sufficiently verified.

During the last decade, as a result of a wide application of climatological methods for determining the balance components, not only were the components for a great many points calculated, but also a series of maps was constructed.

As we have already pointed out in chapter I, among these maps are world maps of the components of heat balance constructed in the Central Observatory (Budyko, Berliand, Zubenok, 1953, 1954a [50 & 51]; Atlas of the Heat Balance, 1955 [15]) and maps for the Soviet Union (Budyko, 1947 [38]; Berliand, 1948 [25]; Berliand and Efimova, 1955 [29] and others).

Utilization of the indicated maps permitted a start on the investigation of the heat balance climatography. It must be remembered that the wide use of maps of meteorological elements obtained by indirect methods is nothing new in climatological studies.

It is already known that in the first climatological investigations on the geographical distribution of a meteorological element, together with a direct generalization of observational data, methods for calculating the values of the elements have been much in use. Particularly so when the first world maps of air temperature distribution were constructed by Humbold in 1817 (for the year) and by Dove in 1858 (for the months), with a direct use of observations. It was almost at the same time that Kämtz (in 1831) constructed maps of temperature for the months by using Meyer's formula in his calculations.

Later on, in connection with accumulated results of observations at meteorological stations, maps of principal meteorological elements were constructed, mainly on the basis of a direct generalization of observational data. However, the method of processing results of observations, that were used for the construction of maps, were gradually improved and included more and more complicated calculations (papers by G.I. Wild, E.S. Rubinshtein and other authors on climatological methods of processing temperature observations; papers by O.A. Drozdov on methods of processing data on precipitation, etc.). The application of these methods actually resulted in the fact that modern maps of distribution of principal meteorological elements - air temperature, precipitation, etc. - have acquired, to a certain extent, a calculative nature. [1)]

1) This statement is especially true in regards to maps for mountain regions.

Thus, at the present time, climatological maps of the distribution of the principal elements and maps of the heat balance components do not differ much in methods used for their construction. Both types of maps are, in substance, attained by a generalization of observations obtained from the network of hydrometeorological stations, and in designing all types of maps the calculation methods are usually applied. The difference between maps of basic meteorological elements and that of heat balance is only seen in the fact that the relative significance of calculation methods for the construction of heat balance maps is greater than for designing the majority of common climatological maps.

Let us examine the principal features in the geographical distribution of heat balance components.

§ 7. Radiation balance

Let us examine first the pattern of geographical distribution of the total sun radiation - principal component of the radiation balance. The amount of total radiation determines the amount of solar energy received by a unit of the earth's surface and utilized later by various natural processes originating near this surface.

Geographical distribution of mean annual and mean monthly amounts of total radiation for the main portion of the earth's surface is presented in series of maps included in the Atlas of the Heat Balance (1955 [15]). These maps are based on calculations of total radiation monthly values, made by using the methods described in chapter II for 1400 sites (1050 on land and 350 on the ocean).

Data obtained from these calculations permitted the construction of total radiation maps for the entire surface of the globe with the exception of the Arctic, Antarctic and high mountains, which had no adequate data, and therfore the application of the climatological calculation methods could result in considerable errors.

The map of the total radiation annual amounts is presented in fig. 16. The mountain regions, with no data, are shaded on this map, as well as on all the others (regions higher than 1.5 km above sea level between 40°N and 40°S and higher than 1 km in the remaining latitudes).

As can be seen from fig. 16, annual amounts of total radiation range from 80 kg-cal/cm^2/year to more than 220 cal/cm^2/year. In higher and moderate latitudes the distribution of total radiation shows a zonal pattern, which is substantially distorted in tropical latitudes only. In lower latitudes a remarkable decrease in total radiation occurs at the equator, which could be due to greater cloudiness.

The largest values of total radiation were observed in high pressure belts of the Northern and Southern Hemispheres and especially in deserts. The maximum value of total radiation was observed in northeastern Africa. It was associated with a very small amount of cloudiness in this place.

Smaller amounts of total radiation were observed in regions with a monsoon climate (for instance, on the eastern coast of Asia), and in some other regions of greater cloudiness.

Fig. 17 and 18 show the distribution of the total radiation amounts for December and June, i.e., for months with the greatest and least mean alti-

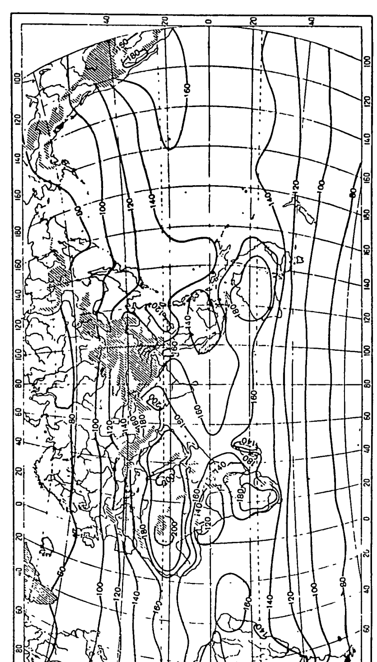

Figure 16. Total radiation, annual values in kg-cal/cm^2/year.

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tudes of the sun for the relevant hemispheres. Isolines are drawn by intervals of 2 kg-cal/cm^2/month.

The map of total radiation in December show the zero line going somewhat above the northern Polar Circle. In higher latitudes, during this month, the sun does not rise above the horizon and total radiation, obviously, equal zero.

Southward from the zero isoline total radiation grows rapidly, changing in extratropical latitudes of the Northern Hemisphere, mainly by zones. In lower latitudes the zonal pattern breaks up, and areas of higher and lower radiation are distributed according to regions with higher and lower amounts of cloudiness. In December the lower latitudes in the Northern Hemisphere and the whole Southern Hemisphere have only very slight zonal changes of total radiation. In the Southern Hemisphere, during this month, total radiation does not decrease much with higher latitudes because of the compensating effect of the increasing length of the day.

A similar pattern in distribution of total radiation is observed in June (fig. 18). In the Northern Hemisphere total radiation varies only slightly (although the areas of greater amounts in deserts are well pronounced), whereas in higher latitudes of the Southern Hemisphere total radiation decreases rapidly with increasing latitudes.

The distribution of total radiation in other months shows the following patterns. In march and September isolines of total radiation resemble qualitatively those presented in the annual map. In these cases the greatest amounts of total radiation are also observed in regions of tropical deserts. The other months (January and February, April and May, July and August, October and November) have an intermediate pattern of distribution of total radiation between those indicated above.

The highest amounts of total radiation of single months can reach values of 20-22 kg-cal/cm^2.

The maps compiled by T.G. Berliand and N.A. Efimova (1955 [29]) could be used for a more detailed geographical analysis of total radiation in the USSR. These maps were constructed by using the same method, but they are more detailed. On the annual map (fig. 19) isolines are drawn in 10 kg-cal/cm^2/year intervals, whereas the interval between isolines on the pertinent world map (fig. 16) is 20 kg-cal/cm^2/year.

As can be seen from fig. 19, annual amounts of total radiation in the USSR range in wide limits, increasing considerably with decreasing latitude.

Over most of the Asiatic USSR amounts of total radiation are somewhat greater as compared with the same latitudes in the European part of the USSR. They are especially higher in Central Asia due to the small amount of cloudiness in this region. In the Far East the total radiation decreases (in comparison with more westerly regions) under the influence of increased cloudiness during the summer season, which is typical for monsoon climate.

Now we will go over to spatial distribution of the other radiation balance components.

The absorbed short-wave radiation is somewhat smaller than the total radiation, however, its distribution patterns are so much similar to those of total radiation that there is no need in studying them separately.

We will only point out some of the major features in the distribution of absorbed radiation.

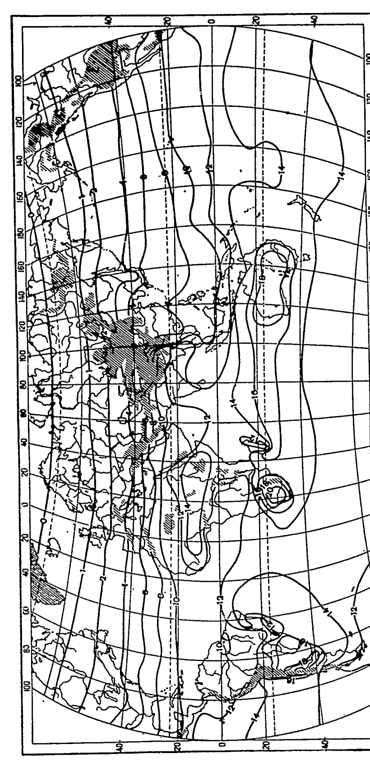

Figure 17. Total radiation, December, in kg-cal/cm^2/month.

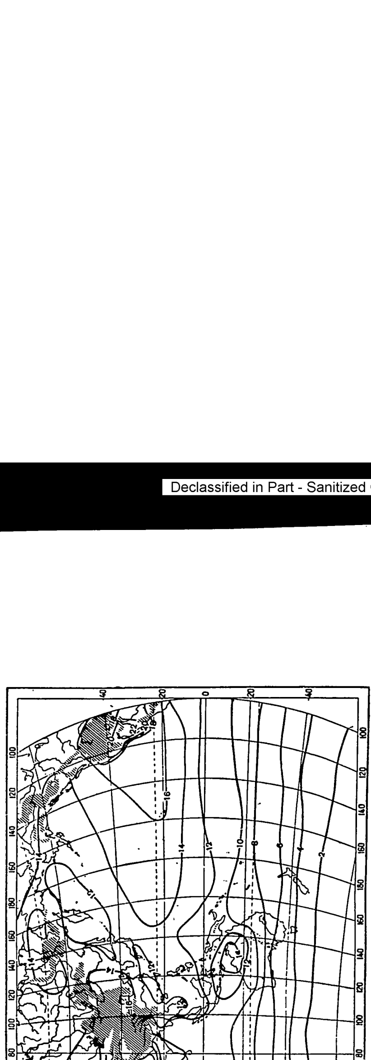

Figure 18. Total radiation, June, in kg-cal/cm^2/month.

Contrary to the pattern of total radiation, isolines of the absorbed radiation are usually not continuous. Particularly on the border of continents and water bodies the isolines of absorbed radiation are, as a rule, discontinued, which shows the abrupt change in the albedo of the underlying surface.

A relatively small difference between the total and absorbed radiation (about 5-20%) is observed on the major part of the land and oceans. This difference increases somewhat in deserts and especially in regions covered with snow and ice. Consequently, if we have on land or water surfaces, a more or less well pronounced boundary of snow or ice cover, then, on this boundary the value of absorbed radiation changes very rapidly and the isolines on the map might break up.

For a study of spatial distribution of the effective radiation (which shows loss of heat by long-wave radiation) data of calculations accomplished for the preparation of maps for the Atlas of the Heat Balance (1955 [15]) can be used. These calculations were made for 770 sites (420 of them on land and 350 on the oceans), more or less regularly distributed over the earth's surface.

These calculations showed that spatial variations in effective radiation are generally smaller than those of total radiation. This phenomenon is to a certain extent explained by the fact that in the majority of the climatic zones changes in temperature and in absolute humidity are associated with each other - with increase in temperature absolute humidity also increases. The changes in effective radiation are comparatively small, since an increase in temperature and absolute humidity are affecting it in opposite directions.

The greatest amounts of effective radiation are observed in tropical deserts, where they reach 80 kg-cal/cm^2/year. This is mainly the result of a tremendous warming of the underlying surface in deserts as compared with air temperature.

Near the equator effective radiation is lowered (about 30 kg-cal/cm^2/year), and in this area it differs only slightly between the land and ocean. With higher latitudes on the ocean, effective radiation increases approximately up to 40-50 kg-cal/cm^2 a year at the 60° parallel. In the area of extratropical latitudes, on land, effective radiation is, on the average, somewhat higher than that of the oceans in the same latitudes, especially so in arid regions.

The spatial distribution of radiation balance of the underlying surface is presented by a series of maps also including the series of world maps, in the Atlas of the Heat Balance.

The chart of annual sums of radiation balance is presented in fig. 20. On this chart the abrupt change of radiation balance at the boundary between land and sea is well pronounced, the isolines break up suddenly. Because of the lower albedo of the ocean surface the radiation balance of the underlying surface is positive. As has been indicated in the author's papers (1948a,1949a [39 & 41], negative annual amounts of radiation balance can, apparently, be observed only in regions with a permanent (or almost permanent) snow or ice cover, for instance in Central Greenland, in Antarctica, etc.

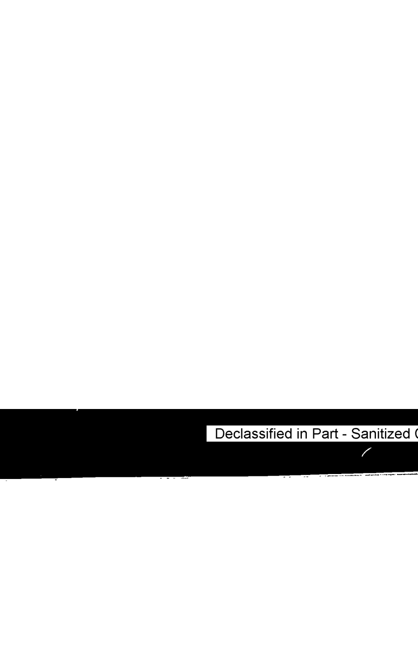

Figure 19. Total radiation in the U.S.S.R., annual values in kg-cal/cm^2/year.

The distribution of radiation balance on the ocean surface, as is shown in fig. 20, is less complicated than that on land and mainly has a zonal pattern. Some deviations from this pattern take place in regions affected by warm or cold sea currents, and these deviations may have a different sign for the same type of current because of the complexity of relations between the values of heat balance components and the air and water temperature, air humidity and cloudiness.

There is a contrast between the comparatively small changes in the radiation balance of ocean surfaces in tropical latitudes, and the rapid decrease of the balance values in moderate zones from lower to higher latitudes.

The largest value of radiation balance on the earth's surface was found in the northern part of the Arabian Sea - it was more than 140 kg-cal/cm^2/year.

The distribution of the radiation balance values on land also has, to some extent, a zonal pattern. However, in many areas the zonality is abruptly broken by the effect of different moistening. In earlier investigations the author repeatedly pointed out the considerable decrease in radiation balance in arid zones of the land as compared with regions of sufficient and excessive moistening at the same latitudes. The reason for this is found in a greater expenditure of radiation heat for effective radiation in arid regions (which results from the high temperature of the surface, small amount of cloudiness and relatively low air humidity) and for reflection of short-wave radiation.

According to this, in fig. 20, with the general decrease of radiation balance with decreasing latitude, there are well marked areas of a considerably lower radiation balance associated with the aridity of climate. This relationship is particularly pronounced in deserts of Central Asia, in the Sahara, and in other desert and arid areas. In monsoon areas, the annual amounts of radiation balance on land are also somewhat lower because of greater cloudiness during the warm season.

The greatest annual amounts of radiation balance on land are observed in humid tropical areas, however, even here they hardly reach 100 kg-cal/cm^2/year, which is considerably less than the maximum values on the ocean.

Let us examine the radiation balance in December, when in the Northern Hemisphere the average altitudes of the sun are at their minimum, and in the Southern Hemisphere - at their maximum. As can be seen from fig. 21, The December radiation balance on a considerable part of the Northern Hemisphere's surface is negative. It is also interesting that the zero line of the radiation balance, on the ocean as well as on land, lies approximately along the same latitude - on the average, a little to the south of 40°N. Northward from this latitude on the ocean the negative radiation balance increases in its absolute values rather rapidly and reaches -4 kg-cal/cm^2/month and even greater negative values.

On land, the negative radiation balance also increases its absolute values in a northerly direction, however, its negative values here reach only about -2 kg-cal/cm^2/month. The reason for this difference is seen in the fact that on ocean surfaces in higher latitudes the water temperature during the colder season is much higher than that of land surface's at the same latitudes.

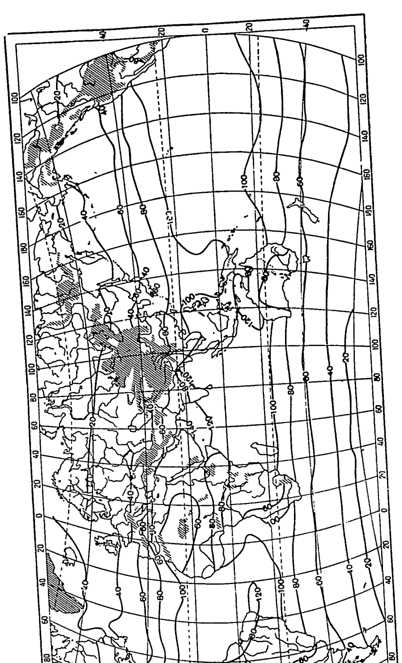

Figure 20. Radiation balance, annual values in kg-cal/cm^2/year.

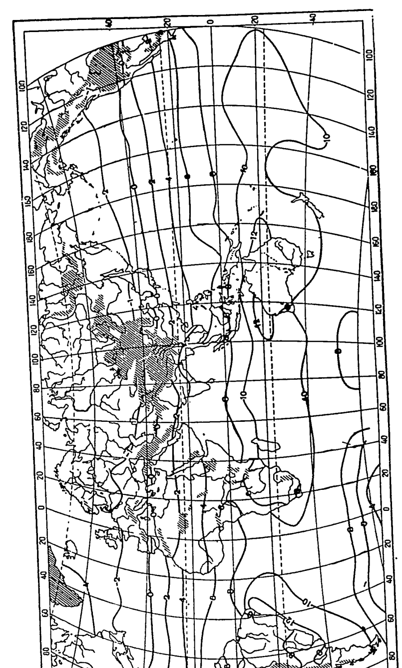

Figure 21. Radiation balance, December, in kg-cal/cm^2/month.

Therefore, effective outgoing radiation on oceans is considerably higher than that on land, and the radiation balance is accordingly smaller.

South of 40°N the radiation balance in December increases on the oceans in a more or less zonal pattern down to the equator, where it reaches values of about 8 kg-cal/cm^2/month. Southward from the equator radiation balance on the oceans changes comparatively little; its greatest values (up to 10-12 kg-cal/cm^2/month) are mainly recorded in the regions of the Tropic of Capricorn. Still further south, the radiation balance again decreases somewhat. On land, between 40°N and the equator it grows regularly in a southerly direction, and south of the equator it changes only slightly, - mainly from 6 to 8 kg-cal/cm^2/month.

The general features in the distribution of radiation balance in June, in the Northern Hemisphere, are similar to those in December in the Southern Hemisphere, and conversely.

The zero isoline of radiation balance in June, on the ocean and on land, lies approximately along 40°S (fig. 22). North of this line the radiation balance increases and reaches, on the ocean, the greatest amounts in the regions of the Tropic of Cancer (with peak values somewhat higher than 14 kg-cal/cm^2/month observed in the northern part of the Arabian Sea). North of the Tropic of Cancer the radiation balance on the oceans decreases slightly, however, up to 60°- 66°N it is still larger than 8 kg-cal/cm^2/month. On land, radiation balance in the Southern Hemisphere grows regularly in a northerly direction approximately up to the equator; northward from the equator, on the vast expanses of land, it only changes slightly. So, for instance, almost for the whole of Asia, from Indo-China up to the Taimyr Peninsula, radiation balance fluctuates within the limits of 6 to 8 kg-cal/cm^2/month.

On maps of radiation balance for the other months, the location of its zero line should be noted. So, for instance in March, this isoline traverses Eurasia from the northwest to a southeastern direction, through southern Scandinavia, Lithuania and the White Rutenian SSR, through the northern Ukraine, Saratov Province and Northern Kazakhstan. In eastern Asia this line lies approximately along 48°N.

On the continent of North America, in March, the zero line in the east lies in the area of the lower stream of the St. Lawrence River, turning considerably northward in the western portion, finally reaching 55°N.

It must be noted that the location of the zero isoline of the radiation balance on land presents an important index, showing the distribution of areas with typical climatic features of the winter season, and therefore deserves a detailed examination.

Let us now show a more detailed map of the annual amounts of radiation balance in the USSR, which we reproduce here as presented by T.G. Berliand and N.A. Efimova in their paper (1955 [29]). Data on this map (fig. 23) show a rapid increase of balance values from north to south. In higher latitudes the radiation balance on land is, on the average for the year, very close to zero, but in the southern part of the USSR it reaches 40-50 kg-cal/cm^2/year. A typical decrease of radiation balance in deserts and also in eastern Siberia is well pronounced.

In conclusion it is necessary to once more indicate that, the examined data on radiation balance presents "macroclimatic" characteristics deter-

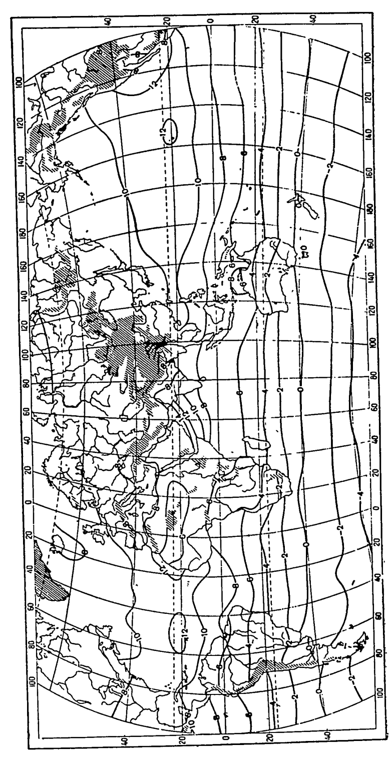

Figure 22. Radiation balance, June, in kg-cal/cm^2/month.

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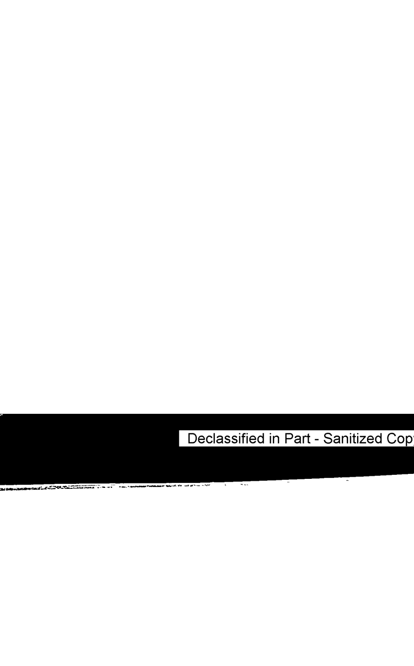

Figure 23. Radiation balance in U.S.S.R., annual values in kg-cal/cm^2/year.

mining its average values for more or less larger regions. As was already noticed in § 6, considerable microclimatic variations of radiation balance can take place on land. Accordingly, actual values of radiation balance for individual plots of a certain underlying surface in any geographical region on land could differ considerably from the general "background" characteristics.

As a rule, spatial variation in radiation balance is much smaller over the ocean than over land, but here too, under certain conditions (floating ice sheets, accumulations of seaweeds, etc.) a rapid change in radiation balance can be observed.

This fact, which makes maps of radiation balance look much like maps of such changeable meteorological elements as, for instance, soil moisture, soil surface temperature, etc., must be accounted for when data on radiation balance are used.

§ 8. Heat balance

We now turn to the question of distribution of the annual values of the other heat balance components on the earth's surface - expenditure of heat for evaporation, turbulent exchange, and redistribution of heat by sea currents.

For the analysis of these questions we can also use maps of the Atlas of the Heat Balance (1955 [15]). A corresponding series of world maps have been constructed by using calculation methods described in chapter II. The balance components were calculated for 1300 sites (about 650 sites on oceans located mainly in 5° latitude and 10° longitude intervals, and about 650 sites on land - in intervals of 5° of latitude and 5° of longitude).

The distribution of the annual amounts of heat spent for evaporation is presented in fig. 24. It shows that amounts of evaporation from land and oceans in the vicinity of coastlines differ considerably. This fact can apparently be explained by the difference in evaporability (the possible evaporation) between the land and ocean because of different radiation balance values, and also by the effect of moisture deficiency in many regions of the land, limiting the evaporation and expenditures of heat for it.

Fig. 24 shows that, in extratropical latitudes expenditures of heat for evaporation generally diminish with higher latitudes. However, this general rule is interrupted on oceans as well as on land, by large non-zonal changes. In tropical zones the distribution of heat losses is also of a very complicated nature; on oceans, this component diminishes somewhat at the equator as compared with areas of high pressure.

The main reason for non-zonal changes in heat losses for evaporation on the ocean are apparently the warm and cold sea currents. All the principal warm currents increase these losses considerably, and cold currents - diminish them. Corresponding changes are well pronounced in the regions of warm currents: Gulf Stream, Kuroshio, Brasilian Current and others, and also in cold stream areas like: Canary Current, Benguela Current, Californian Current, Peruvian Current, Labrador Current, etc.

Under the influence of sea currents, which increase or lower the water temperature, the annual amount of evaporation from the ocean surface at a certain latitude can vary by two or three times its value.

Besides the sea currents, atmospheric circulation also contributes to the non-zonal changes in amounts of evaporation and in amounts of heat spent for it. This effect is felt mainly through changes in the radiation balance of the ocean surface.

The heat expenditure distribution for evaporation from land differs still more from the zonal pattern as compared with that of oceans, because of the tremendous influence on evaporation exerted by climatic conditions of moistening.

When a sufficient amount of moisture is available in soil, evaporation and losses of heat for evaporation are regulated mainly by the radiation balance. Such conditions are observed in regions of higher latitudes and in moist regions of moderate and tropical latitudes.

In regions of insufficient moistening, the amount of evaporation is diminished because of the deficiency in soil moisture. In deserts and semideserts it approaches the sum of precipitation, which is very small in these regions. Fig. 24 shows distinctly the areas of abruptly reduced losses of heat for evaporation in the main arid regions of the terrestrial globe. The greatest expenditure of heat is observed in moist equatorial regions where it reaches 60 kg-cal/cm^2/year, which is equal to evaporation of a water layer approximately 1 m. thick, during a year.

Much greater values are reached in evaporation from the ocean surface, where the radiation balance is somewhat higher than that on land, and - this is very important - the evaporating surface receives a great deal of additional heat energy as a result of the heat redistribution by sea currents. In this connection, in some tropical and subtropical regions, the annual evaporation from the ocean surface reaches amounts somewhat higher than a layer of water 2 m. thick.

The maps of monthly expenditures of heat for evaporation from the ocean completely confirm, the fact of the opposite annual march of evaporation in extratropical latitudes on land and oceans, which was stated earlier in the literature.

It is well known that, during the cold season evaporation on land diminishes considerably. The maximum evaporation is observed at the beginning or in the middle of the warm season, depending on the precipitation. Contrary to this, the evaporation on the oceans, in the cold season, usually increases as compared with the warm period. A direct reason for this is the increase of the temperature difference between air and water during the cold season, which also increases the difference in concentration of water vapor at the water surface and in the air. Moreover, in many regions the mean wind speed during the cold season is higher than in the warm period, which also increases evaporation during cold periods.

The increase in evaporation and heat expenditures for evaporation during the cold season is closely associated with the increasing effect of warm currents during these periods, whereas during the warm season, cold currents are most active in this respect, thus lowering the expenditure of heat for evaporation. The actual conditions of heat influx to the evaporation surface of oceans, which are associated with the existence of a powerful horizontal heat transmission in the hydrosphere, represent the main factor in increasing losses of energy for evaporation in the cold period.

We will now consider the problem of the geographical distribution of

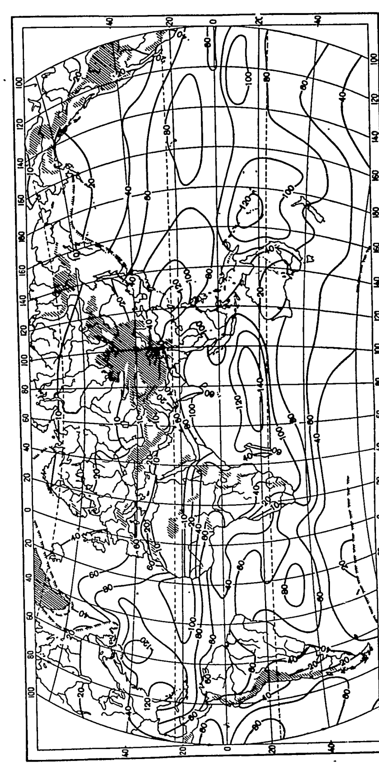

Figure 24. Expenditures of heat for evaporation, annual values in kg-cal/cm^2/year.

turbulent heat flux.

Fig. 25 shows the amount of heat that is emitted by the underlying surface to the air (positive values) or received from the air (negative values). First of all, the most outstanding feature shown by this map is the fact that the surface of all continents (except Antarctica) as well as the major portion of the ocean surface, on the average for the year, emits heat into the atmosphere.

On the major portion of the ocean surface the turbulent heat exchange value is not large in comparison with the principal components of heat balance, and usually comprises not more than 10-20% of their values. High absolute values of the turbulent exchange are reached in regions where the water is, on the average, much warmer than the air, i.e., in regions affected by powerful warm sea currents (Gulf Stream) and in some areas of higher latitudes where the sea is still free from ice. Under these conditions turbulent heat flux can exceed 50 kg-cal/cm^2/year.

The other warm currents, except the Gulf Stream, exert a relatively minor influence on turbulent heat emission from the ocean surface into the atmosphere.

The cold currents, which lower the sea water temperature, diminish the turbulent streams of heat from the ocean surface into the atmosphere and reinforce streams of opposite flow. As a result of this, in regions affected by some cold sea currents, the mean annual heat flux although very small is directed from the atmosphere to the ocean (Canary, Benguela, Californian sea currents). More complicated are the causes for the formation of areas with the heat flux directed to the ocean surface in the southern portions of the Atlantic and Indian Ocean (at 50°S). In this case, the turbulent heat flux is apparently affected by the advection of warm air masses over a cold ocean surface.

In contrast to what is observed on oceans where on the average the turbulent heat flux increases in its absolute value, with higher latitudes, on land, this flux changes, in the opposite direction. At the same time, the turbulent stream on land is strongly affected by the precipitation regime - in the arid regions turbulent heat emission from the land surface into the atmosphere is much stronger than in moist areas.

Consequently, the greatest expenditure of heat by turbulent flux on land is found in tropical deserts, where it may exceed 50-60 kg-cal/cm^2/year. In moist tropical regions, especially in regions of moderate and higher latitudes, expenditure of heat by turbulent flux is much less than the value given above.

In December turbulent heat flux reaches large absolute values only in the northern Atlantic Ocean and in the northwestern section of the Pacific Ocean, as a result of the warm sea currents' influence, and the development of heat exchange between the cold surface of continents and warmer oceans.

In the region affected by the Gulf Stream, the expenditure of heat by the ocean for turbulent heat emission reaches 4-8 kg-cal/cm^2/month, and in regions affected by the Kuroshio it reaches 2-4 kg-cal/cm^2/month.

It is interesting to note that, a noticeable heat emission (more than 2 kg-cal-cm^2/month) in December, is also observed in the northern South China Sea and in the Gulf of Bengal, which is connected with the development of monsoon circulation in this region.

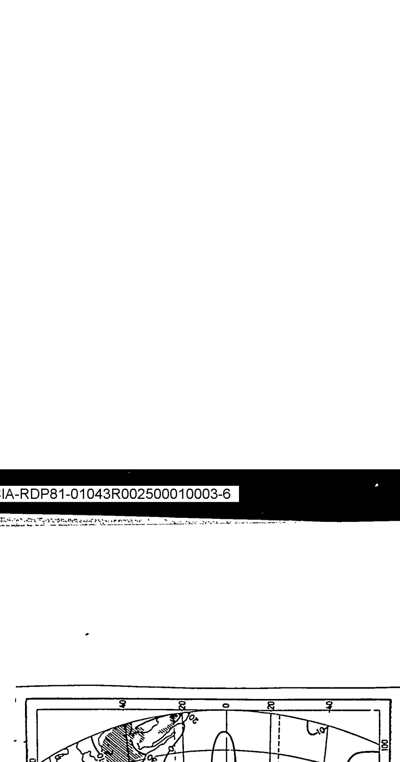

Figure 25. Turbulent heat exchange, annual values in kg-cal/cm^2/year.

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An entirely different pattern of distribution of turbulent heat streams over the ocean is observed in June, when in the northern portion of the Atlantic Ocean at the shores of North America, Europe and Africa we can see large zones where the turbulent heat stream is directed from the atmosphere to the ocean surface.

A similar pattern also exists over a large section of the Pacific Ocean in the Northern Hemisphere, including the shores of China and Japan.

However, it must be noted, that in all these regions (except in the limited region affected by the cold Labrador current in the northwest portion of the Atlantic Ocean) turbulent flux is weak (less than 2 kg-cal/cm^2/month). This shows that in the process of heat exchange between oceans and land, oceans emit considerably more heat during the cold season than the amount they receive from land during the warm period.

In the Southern Hemisphere during June, oceans in moderate and higher latitudes emit up to 4-6 kg-cal/cm^2/month into the atmosphere.

As has already been noted in chapter I, each unit of the ocean surface can on the average per year, receive or lose some amount of heat due to horizontal heat emission in oceans, i.e., mainly due to the effect of sea currents. Mean annual values of this emission, determined as the remainder term of heat balance described earlier, are presented in fig. 26.

In analyzing this map we can see the existence of a good agreement between areas of higher positive values (showing outflow of heat from the ocean surface) and regions of cold sea currents, and conversely, between lower negative values and warm currents.

This agreement is particularly observed for warm currents-Gulf Stream, Kuroshio, Igol'ny, Southwest Pacific-, and for cold streams - Canary, Benguela, California currents and the northern portion of the Peruvian Current.

At the same time, in some regions of the ocean, the distribution of isolines in fig. 26 does not coincide with the locations of the main areas of warm and cold streams. This is explained partly by the fact that the value shown on the map does not directly indicate a transfer of heat by sea currents, but shows only one of the consequences of this transfer - the average increase or loss of heat by the ocean surface as a result of heat exchange with deeper layers.

We can, for instance, assume that because of this reason the greatest absolute amounts of these values in the Gulf Stream region of moderate and higher latitudes are located more to the west, away from the main current of this stream. It is quite probable that the greatest loss of heat, transferred by the stream, takes place in the western areas of the Gulf Stream, which are the closest to the cold region of the northwestern shores of the Atlantic coast. However, it must be pointed out that some peculiar features in the isolines of fig. 26 are apparently connected, to a certain degree, with the insufficient reliability of the calculated values. Since these values were derived as the remainder term of balance, it is obvious that when calculating them, errors in determining the other components of balance were added and the greatest error in the calculations, as compared with all other components, originated. This was apparently reflected by the location of the isolines, which diminishes their reliability, especially for some regions in the oceans of the Southern Hemisphere and for

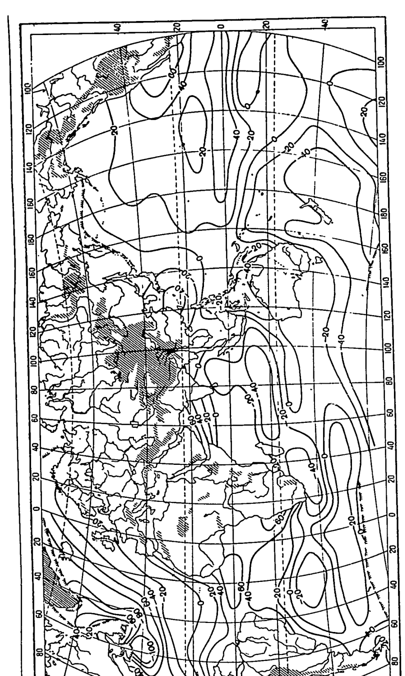

Figure 26. Heat exchange between the ocean surface and the deeper layers, annual values in kg-cal/cm^2/year.

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the northern section of the Pacific Ocean, where original climatological data used for calculations were positively insufficiently reliable.

§9. The annual and diurnal variations of heat balance components

The annual variations in heat balance on land

Let us now examine the principal features of the annual march of heat balance components in the main climatic zones of the terrestrial globe. For this purpose we will use data from which the Atlas of the Heat Balance was constructed.

Using the principles of climatic classification by B.P. Alisov (Alisov, 1950 [8]; Alisov, Drozdov, Rubinshtein, 1950 [9]; Alisov and Sorokina, 1953 [10], and others), we present here data on the annual march of balance components for climatic zones that differ in circulation processes.

Taking into account the fact that, the annual march of the heat balance components usually differs substantially between land and ocean regions, we will start with characteristics of the annual march on land surfaces.

A typical annual march of heat balance components in the equatorial zone is presented in fig. 27 (Manãos, South America). As can be seen from this figure, in the equatorial zone, the radiation balance R changes rather slightly in its annual march. Two maxima - a small one in spring, and a more significant one in autumn, - are connected with the increase of total radiation during the equinox period, when in equatorial latitudes, the average altitudes of the sun are greatest.

Because of a persistent soil moistening by abundant precipitation in the region of Manãos, the major portion of the radiation balance heat is spent for evaporation. The annual march of heat expenditure for evaporation LE is almost parallel to the annual march of radiation balance.

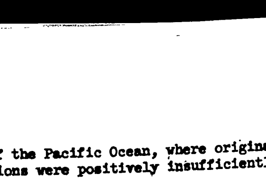

Figure 27

Annual variations of heat balance components.
Manãos, 3°08'S, 60°01'W. Equatorial continental climate.

Turbulent heat exchange P shows small values in all months of the year. These values grow a little larger in autumn, after a period of relatively sparse precipitation, when losses of heat for evaporation diminish slightly as compared with radiation balance.

Another type of the annual march of heat balance components is presented in fig. 28. It pertains to climatic conditions of equatorial monsoons on eastern coasts of continents (Saigon, Indo-China). Radiation balance, preserving great values during the whole year, has a sharp maximum at the end of winter and beginning of spring, when the analyzed region is occupied by dry tropical air with only a slight amount of cloudiness. A decrease in cloudiness increases considerably the total radiation, which results in a larger radiation balance.

With a large annual amount of precipitation in Saigon, the expenditure of radiation balance increases considerably; the expenditure of heat for evaporation is also great, and it varies considerably during the year.

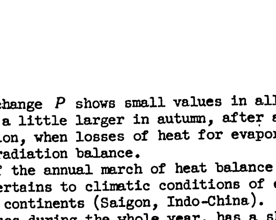

Figure 28

Annual variations of heat balance components.
Saigon, 10°47'N, 106°42'E. Equatorial monsoon climate.

At the beginning of the dry period, when the soil is still wet, but the radiation balance has already increased considerably, the expenditure of heat for evaporation increases slightly, as compared with the moist period. However, somewhat later, with the progressing desiccation of soil, loss of heat for evaporation diminishes rapidly and at the end of the dry period (April) it is reduced approximately to half of the maximum value that was observed in February.

It should be noted that, the minimum of evaporation is observed later than the maximum of radiation balance. Apparently this can be explained by the dependence of evaporation on soil moisture.

The annual march of turbulent heat exchange is, in its general form, opposite to that of evaporation. The turbulent stream is small during all months, except during the second half of the dry period, when it becomes larger than the expenditure of heat for evaporation.

For areas of the tropical belt a great diversity of the annual march of heat balance components is typical. It depends on the location of analyzed regions in relation to permanently acting baric systems.

The annual march of heat balance components in areas of continental tropical climate is presented in fig. 29 (Aswan, northeastern Africa).

Figure 29

Annual variations of heat balance components. Aswan, 24°02'N, 32°53'E. Tropical continental climate.

In typical tropical deserts the annual march of the heat balance components is determined by rather simple laws.

Under conditions of consistently small amounts of cloudiness, variations of radiation balance are effected mainly by the altitude of the sun in its annual march. This results in certain changes in total radiation.

A well pronounced annual march of radiation balance shows therefore, a strong influence of astronomical factors on radiation balance, even to relatively low latitudes (about 24°). It should be noted that the greatest values of radiation balance in Aswan are smaller than its higher values in Saigon, although the corresponding values of total radiation show a converse relationship. This is explained by the somewhat higher albedo in deserts and, mainly, by very large values of effective outgoing radiation from strongly heated desert surfaces.

Since precipitation in Aswan is practically nil, expenditure of heat for evaporation during the whole year is close to zero. As a result of this, the turbulent heat emission is very great and approaches, in its value, the radiation balance (a small difference between these values is effected by the existence of a relatively small heat exchange in soil).

Entirely different patterns of annual variations of heat balance components are observed in regions of the western periphery of tropical anticyclones (fig. 30, Belize, Central America).

Figure 30

Annual variations of heat balance components. Belize, 17°32'N, 88°10'W. Tropical climate of the western periphery of oceanic anticyclones.

In these regions, quantities of precipitation are large and, accordingly, expenditure of heat for evaporation is great. During the major portion of the year, in Belize, loss of heat by evaporation is only slightly smaller than the radiation balance, and turbulent heat emission is comparatively insignificant. Evaporation decreases at the end of spring after a brief dry period, during which the soil loses a portion of its moisture. Simultaneously, the amount of turbulent heat emission increases markedly.

It is interesting to note that the greatest values of radiation balance in Belize are observed in the spring months and not in summer, when the great cloudiness diminishes the total radiation.

Contrary to the climatic features of Belize, in the eastern periphery regions of the oceanic anticyclones the features of deserts are observed, which however, differ considerably from conditions of continental tropical deserts. A sample of an annual march of heat balance components for this type of climate is given in fig. 31 (Swakopmund in southwestern Africa).

Figure 31

Annual variations of heat balance components.
Swakopmund, 22°42'S, 14°32'E. Tropical climate of oceanic anticyclones.

As can be seen from fig. 31, the radiation balance in Swakopmund, as well as in Aswan, changes with the average height of the sun, which is at its minimum in the Southern Hemisphere winter. However the greatest values of radiation balance in Swakopmund are noticeably smaller than those in Aswan, due to the diminished total radiation as effected by the decreased translucence of the atmosphere to solar rays.

The expenditure of heat for evaporation in Swakopmund is insignificant and the turbulent exchange is, in its value, close to the radiation balance.

In the subtropical belt climatic conditions are also very diversified and depend on the patterns of circulation processes. The annual march of heat balance components changes here according to circulation factors.

A typical pattern of the annual variations of heat balance components in subtropical continental climate is presented in fig. 32 (Krasnovodsk, Middle Asia). In this case the annual march of radiation balance, as effected by astronomical factors, is very well indicated. In winter months the radiation balance already reaches some negative values. The loss of heat for evaporation is insignificant because of sparse precipitation, especially in the summer months. Accordingly, in the warm season, the turbulent flux reaches large values, whereas during the cold period it is small in its absolute value, and in some months it is even directed from the atmosphere to the earth's surface.

Quite a different type of annual march of heat balance components is observed in subtropical climates on the western coasts of continents. Data presented in fig. 33 (Lisbon) indicate that, in this case the radiation balance has a very high maximum in summer; it is considerably higher than that of Krasnovodsk.

Figure 32

Annual variations of heat balance components.
Krasnovodsk, 40°00'N, 52°59'E. Subtropical continental climate.

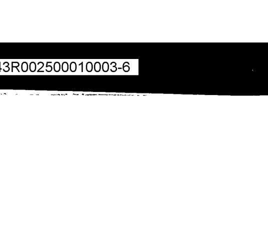

Figure 33

Annual variations of heat balance components
Lisbon, 38°42'N, 9°08'W. Subtropical climate at the western coasts of continents.

The reason for this difference is seen in the fact that in Lisbon, with a small amount of cloudiness (and, consequently, a sizable total radiation) in summer, there is a considerable lower expenditure of heat for reflection of short-wave radiation and for effective outgoing radiation as compared with Krasnovodsk.

In Mediterranean climate, with dry summers and moist winters, a peculiar type of the annual march of heat expenditures for evaporation and turbulent heat emission is observed.

The expenditure of heat for evaporation in the Lisbon region increases rapidly in spring (in accordance with increasing radiation balance) and remains high in the beginning of summer when the soil is not completely dry, then drops rapidly in accordance with the desiccation of the upper soil layers. The small secondary maximum of evaporation is observed at the end of autumn, when the soil is again moistened. However, it is soon replaced by the winter minimum, due to the deficit in radiation heating.

The turbulent heat emission reaches a maximum peak during the second half of summer and at the beginning of autumn. In the other seasons its values are small.

In subtropical monsoon climate at the eastern coasts of continents, as is seen from fig. 34 (Shanghai), the summer maximum of radiation balance is considerably reduced. The reason for this is the greater cloudiness during the summer monsoon. A great amount of precipitation assures high expenditures of heat for evaporation, which are close to the radiation balance value. In connection with this, the turbulent heat emission is comparatively small throughout the year. In the moderate zone the pattern of the annual march of heat balance components also varies with the circulation factor.

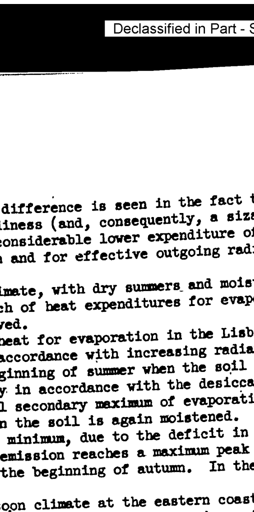

Figure 34

Annual variations of heat balance components.
Shanghai, 31°14'N, 121°27'E. Subtropical monsoon climate.

In the central regions of continents the radiation balance has a distinct summer maximum; in the Northern Hemisphere it is usually in June (fig. 35, Barnaul). In spring and autumn the radiation balance changes rapidly. In winter it is negative, with absolute values smaller than those of the summer maximum. The period with negative values of radiation balance coincides, more or less, with the snow cover period (although there are of course some differences between these periods).

The annual march of heat expenditures for evaporation in these climatic regions resembles that of the radiation balance. The difference between the amounts of heat expended for evaporation and radiation balance in the warm season is greater with lesser precipitation. The changes in turbulent heat emission during the year show a summer maximum, which is more pronounced in drier climates. In winter the turbulent flux of heat is directed down (to the underlying surface), however, its absolute values usually are smaller than in the summer season.

It is worthwhile to mention that, in continental climates the highest values of heat balance are often observed at a different time. The maximum of heat expenditures for evaporation usually precedes the maximum of radiation balance, and the maximum of turbulent heat emission follows the maximum of radiation balance.

This type of distribution is the result of soil desiccation in midsummer and in the second half of the summer season, which diminishes the expenditures of heat for evaporation at this time, and accordingly, increases turbulent heat emission.

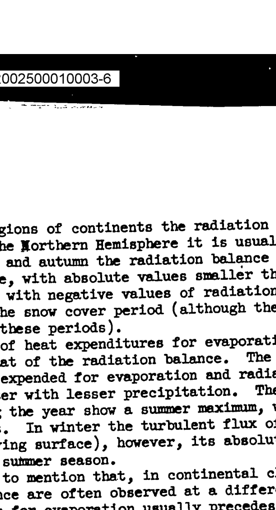

Figure 35

Annual variations of heat balance components.
Barnaul, 53°20'N, 83°48'E. Continental climate of moderate latitudes.

In maritime climates of moderate latitudes (fig. 36, Paris), many particular features of the annual march observed in continental climates are preserved. This also includes the sequence in time of maxima appearance of heat balance components.

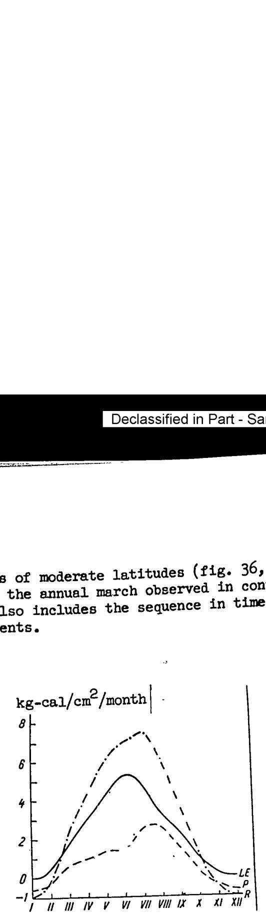

Figure 36

Annual variations of heat balance components.
Paris, 48°49'N, 2°29'E. Maritime climate of moderate latitudes.

Moreover, in the monsoon climate zone at the eastern coasts of continents in moderate latitudes, the annual march of the heat balance components has some peculiar features (fig. 37, Vladivostok). In this case, high clouds in summer effect a charateristic "flattening" of the annual curve of radiation balance with reduced maximum values. The curve of heat expenditure for evaporation also has a similar form, with the maximum in August. In winter the heat balance of this region has approximately the same characteristics as the heat balance of similar latitudes in the continental climate zone.

In higher latitudes, in a subarctic continental climate, the components of heat balance have an annual march similar to that of the continental climate regions in moderate latitudes (fig. 38, Turukhansk).

In this zone maximum values of radiation balance are not less than those in lower altitudes, although the length of the period with positive values of radiation balance is considerably smaller. Due to this fact, the annual march of radiation balance curve has a sharper form. A similar form is found in the curve of heat expenditures for evaporation in the same regions.

It is interesting to note that, in the subarctic vertical turbulent flux of heat directed upward is, in the warm season still greater, in its obsolute values, than turbulent flux in the cold season, which is directed from the atmosphere to the underlying surface.

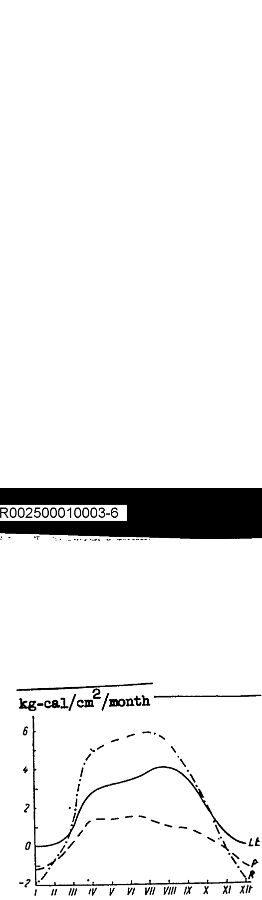

Figure 37

Annual variations of heat balance components.
Vladivostok, 43°07'N, 131°54'E. Monsoon climate of moderate latitudes.

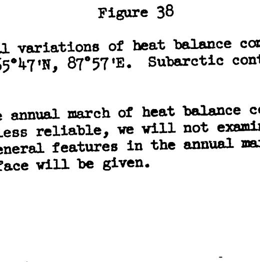

Figure 38

Annual variations of heat balance components.
Turukhansk, 65°47'N, 87°57'E. Subarctic continental climate.

Since data on the annual march of heat balance components for arctic latitudes are much less reliable, we will not examine them, and only some remarks about the general features in the annual march of heat balance components on land surface will be given.

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The results presented above indicate that, the greatest monthly sums of radiation balance change only slightly when we go from equatorial to sub-arctic latitudes. The main reason for this is seen in the fact that, in higher latitudes the decrease of maximum altitudes of the sun in summer is compensated, to a certain extent, by the increase in daylight hours.

Thus, the change in annual values of radiation balance with latitude, which is rather remarkable in moderate and higher latitudes, is accomplished not by the reduction of the balance's maximum values but by shortening the period with its positive values.

The annual march of two other principal components of heat balance on land surfaces - expenditure of heat for evaporation and turbulent heat exchange - are very much dependent on conditions of moistening.

Under consistent excessive moisture conditions, expenditures of heat for evaporation are close to the radiation balance value and, in this case, the turbulent heat flux is not great (at a positive radiation balance) and usually represents heat emission from the underlying surface into the atmosphere.

If dry periods effect a considerable soil desiccation, expenditures of heat for evaporation are diminished, and this decrease is more significant, in comparison with radiation balance, when the desiccation of soil is greater. Simultaneously, turbulent heat emission also increases. The period of decreasing expenditures of heat for evaporation and the growing heat emission lags behind the start of the dry period in its phase relationship, since at the beginning of a drought the soil usually retains a considerable amount of moisture.

In conclusion it must be noted that, on land, monthly sums of heat exchange between the underlying surface and deeper layers (heat exchange in soil), are usually much smaller in their absolute values than the maximum monthly sums of the principal components of heat balance. However, during periods when the principal components of balance are lower in their absolute values (for instance, in winter months), values of heat exchange in soil could be quite comparable with these components.

The curve of the annual march of heat exchange in soil resembles, in the first approximation, a sinusoid. This curve indicates an expenditure of heat for soil warming during the warm season and a gain of heat in the cold season when the soil cools off. The amplitude of the annual variations of heat exchange in soil, as has already been noted in chapter II, is closely associated with the annual range of air temperature. Therefore, in many tropical regions, where the air temperature changes only slightly, during the year, changes in monthly sums of heat exchange in soil are insignificant throughout the year.

The highest monthly values of heat exchange are observed in continental climates of moderate and higher latitudes. However, in these cases, maximum monthly values of heat exchange are still much lower than the maximum values of radiation balance. This circumstance basically differentiates the conditions of heat exchange between the underlying surface and deeper layers on land from those of the ocean.

The annual march of heat balance components on the oceans.

The annual march of heat balance components in various zones of the oceans will be examined in the same order as has been done for land.

As an example of the annual march of the heat balance components in the oceans of the equatorial zone we will present data for the western segment of the Pacific Ocean, north of New Guinea. Fig. 39 shows that in this area radiation balance changes only slightly during the year. However, spring and autumn maxima are still noticeable (both are shifted somewhat from the equinoctial months - spring maximum from March to February, and autumn maximum from September to October). The heat expenditure for evaporation is close to the radiation balance value, and a turbulent heat flux, small in its absolute value, is directed from the ocean surface to the atmosphere during the entire year.

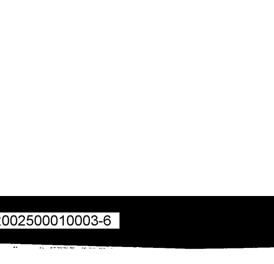

Figure 39

Annual variations of heat balance components.
Pacific Ocean, 0° latitude, 150°E. Equatorial climate.

The heat exchange between the underlying surface and deeper layers, in this region develops into some comparatively large values in autumn, when heat gain from the radiation balance considerably exceeds the expenditures for evaporation and turbulent heat emission. The surplus of heat, which is received by water masses during autumn, should obviously be transported from the analyzed region to higher latitudes by currents and by macroturbulence.

The annual march of heat balance components in the oceanic climates of equatorial monsoons is presented in fig. 40 (The Arabian Sea region).

Figure 40

Annual variations of heat balance components.
Indian Ocean, 15°N,70°E. The climate of equatorial monsoons.

In this case the regular form of the annual radiation balance curve, with a summer maximum and winter minimum, is distorted by a rapid increase of cloudiness in summer during the period of equatorial air masses influx. An increase in cloudiness diminishes the total radiation and radiation balance, and in the annual march of radiation balance a secondary summer maximum occurs. The turbulent heat emission in this region is insignificant during the entire year (which is the result of insignificant differences between water and air temperatures). However, turbulent emission increases somewhat in winter, as is typical for monsoon climates.
Expenditure of heat for evaporation in this region changes during the year inversely to changes of radiation balance (this relationship is typical for the major portion of the ocean). The winter maximum of evaporation is, in this case, explained by advection of dry trade wind air masses, and is connected with a considerable increase in saturation deficit. The summer maximum is associated with a strong increase of wind speed during the equatorial monsoon period.
As a result of the considerable increase of heat losses for evaporation

in winter and summer, and of a diminished radiation balance therewith, the heat flux between the underlying surface and lower water layers is directed upward during these seasons, although the absolute values are comparatively small. In contrast to this, in spring and autumn, great quantities of heat are transmitted from the ocean surface to deeper layers and eventually transported in horizontal directions into other areas of the world's oceans.
In the tropical belt the annual march of heat balance components, over the ocean, are different in those areas which have different circulation conditions.
The warm sea currents are located in the western periphery of oceanic anticyclones, as it is well known. This creates favorable conditions for some increase in turbulent heat emission from the underlying surface to the atmosphere.
As an example we will analyze the region near the island of Trinidad (Atlantic Ocean, northwest of the Brazilian coast). The annual march of heat balance components for this area is given in fig. 41.

Figure 41

Annual variations of heat balance components.
Atlantic Ocean, 20°S, 30°W. Tropical climate of
The western periphery of oceanic anticyclones.

Radiation balance, under these conditions, changes in accordance with the annual march of total radiation, but expenditures of heat for evaporation have an opposite pattern. The turbulent heat emission grows in the winter months (for the Southern Hemisphere), when the effect of the warm Brazilian current is reinforced. At this time of year the expenditure of heat for evaporation and turbulent heat emission exceeds markedly the radiation balance. As a result of this a considerable loss of heat occurs on the ocean surface; this heat comes from the deeper layers. This is a typical process in an area affected by a warm current which transports warmer water.

In contrast to the conditions observed in the western periphery of oceanic anticyclones, their eastern periphery is affected by cold sea currents, and the annual march of heat balance components changes accordingly. As an example of the annual march of balance components in tropical areas of the eastern periphery of oceanic anticyclones, we will analyze data pertaining to the region affected by the Benguela Current in the southeast portion of the Atlantic Ocean (fig. 42).

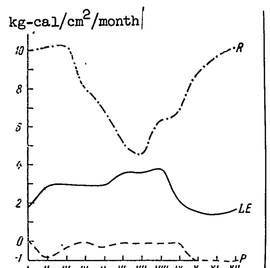

Figure 42

Annual variations of heat balance components. Atlantic Ocean 20°S, 10°E. Tropical climate of the eastern periphery of oceanic anticyclones.

In this case the expenditure of heat for evaporation is drastically reduced as compared with that of the preceding region, which is located in the same latitudinal zone.

The turbulent flux of heat, very small in its absolute value, is directed from the atmosphere to the cold ocean surface. The absolute values of heat flux grows somewhat larger in summer (of the Southern Hemisphere), when the effect of the cold current is reinforced.

In this region, gain of heat from the radiation balance and turbulent heat exchange is much larger than losses from evaporation, and a great amount of heat energy is transmitted to deeper layers of the ocean which are spent on heating the comparatively cold water masses transported by the current. These expenditures are especially large in the summer months.

In the subtropical belt, the main features of the annual march of heat balance components on the ocean surface are similar to those in the corresponding areas of tropical latitudes. However, annual variations of radiation balance are much more sharply pronounced, which is the result of considerable changes in the average altitude of the sun during the year.

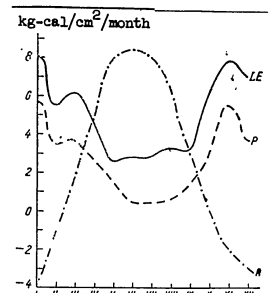

Figure 43

Annual variations of heat balance components. Atlantic Ocean, 55°N, 20°W. Climate of moderate latitudes in regions of warm sea currents.

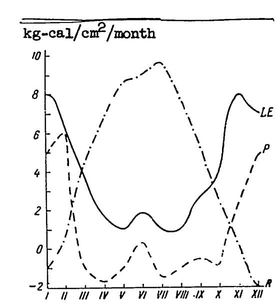

Figure 44

Annual variations of heat balance components. Pacific Ocean, 45°N, 160°E. Monsoon climate of moderate latitudes.

Typical annual variations of heat balance components of the ocean surface, in moderate latitudes, are presented in figs. 43 and 44. The former, pertaining to the northern portion of the Atlantic Ocean, shows heat balance conditions in the area affected by the warm current of the Gulf Stream.

At the 55° latitude the radiation balance of the ocean surface changes within very wide limits, during the year and its negative value in winter, in contrast to land conditions, cannot be considered as being small in its absolute values.

The turbulent heat flux, in this region, is directed from the warmer ocean surface to the atmosphere during the whole year, and in winter its values are much larger than in summer. The expenditure of heat for evaporation in the winter months is also very large. The ocean surface must receive a great amount of heat from deeper layers to compensate for the expenditure for evaporation, turbulent heat emission and heat lost by outgoing radiation.

The gain of heat, in this case, is achieved through the cooling of upper water layers and partly, by utilization of huge heat energy resources of the Gulf Stream's powerful current.

The annual march of the heat balance components changes considerably when the effect of a warm current is combined with that of a monsoon climate of moderate latitudes. As can be seen from fig. 44, which pertains to the northwestern portion of the Pacific Ocean (southwest of the Kuril Islands); in this case turbulent heat flux in the warm season is, on the average, directed from the atmosphere to the ocean surface, and during the cold season, conversely, from the ocean to the atmosphere. It is obvious that, the value of turbulent heat flux represents an important quantitative index of the influence exerted by monsoonal circulation on heat exchange.

In the analyzed region, as well as in the preceding one too, during the winter months, the ocean surface receives heat from deeper layers, which is associated to a considerable extent with utilization of energy from the warm current of the Kuroshio. In summer, however, a converse relationship takes place; the supply of heat from the radiation balance and turbulent heat exchange considerably exceeds the expenditures for evaporation, which results in the heating of the upper waterlayers and facilitates the transmission of excessive heat into other regions by means of horizontal heat conductivity.

Since data on the annual march of heat balance components for higher latitudes are less reliable on the oceans, we will not examine them.

In conclusion we will list some generalizations regarding the annual march of heat balance components on the oceans.

The annual march of radiation balance on the oceans is, by and large, similar to that of land regions with a moist climate.

The turbulent heat exchange on oceans depends substantially on the actions of warm and cold sea currents, which are effecting changes in water surface temperatures. In regions of cold currents the turbulent flux of heat, which is directed from the atmosphere to the underlying surface, usually has small absolute values. In areas affected by warm currents the heat flux is directed from the water surface to the atmosphere, and can reach very large values.

It must be noticed that in both cases (as distinct from land conditions) the annual march of turbulent heat flux is only slightly associated with the annual march of radiation balance, but it depends greatly on changes in the regime of sea currents throughout the year.

A similar pattern is observed in the annual march of heat expenditures for evaporation on oceans. As has been mentioned before, in moist land areas, the annual march of heat expenditure for evaporation is very close to that of the radiation balance. Contrary to this, on oceans, changes in heat expenditures for evaporation in its annual march are usually opposite to changes in radiation balance. This phenomenon is often connected with a strong effect of warm and cold currents on evaporation. The regime of these currents changes substantially during the year.

On the other hand, annual sums of heat expenditures for evaporation on oceans, averaged for sufficiently large surfaces, are in most cases very close to the values of radiation balance.

The greatest differences in the annual march of heat balance components, between land and oceans, are found in the heat exchange between the underlying surface and deeper layers. On land, this component is usually not great for particular months, and approaches zero in its annual value. On oceans it may reach very large values for the year, and especially for single months. This is the result of the great horizontal and vertical heat conductivity of water, and it presents the basic reason why oceanic and continental climates are so different.

Diurnal march of heat balance components in various climatic conditions.

Most of the data on the diurnal march of the balance components was obtained by expeditions, for more or less shorter periods, which makes it difficult to utilize them in studying mean climatological features.

The most important experimental investigations along this line were accomplished by the "Station of the Physics of the Air Layer near the Ground." This station is attached to the Central Geophysical Observatory, and it is located in Koltushi (Leningrad region). The observations in Koltushi have been carried out during a comparatively long period - in 1947 and 1949-1951. These data were worked up by T.A. Ogneva (1955 [179]), and are presented in fig.45 in the form of: the average diurnal march of radiation balance R, heat expenditures for evaporation LE, turbulent heat exchange P, and heat exchange in soil A, for the month of July.

As can be seen from this graph, the diurnal march of the heat balance components during the warm season is quite similar to the annual march of balance components under similar climatic conditions in moderate latitudes.

The comparatively large values of radiation balance in daylight hours provide for: expenditures of heat for evaporation, turbulent heat emission and heat exchange in soil. The expenditure of heat for evaporation is markedly greater than turbulent heat emission (which is typical for moist climates), and heat exchange in soil is considerably smaller than the loss of heat for evaporation or turbulent heat exchange.

During the nighttime negative radiation balance is comparatively small in its absolute value, losses of heat for evaporation approach zero and the expenditure of radiation heat is compensated by the gain from the turbulent heat exchange and heat emission from soil.

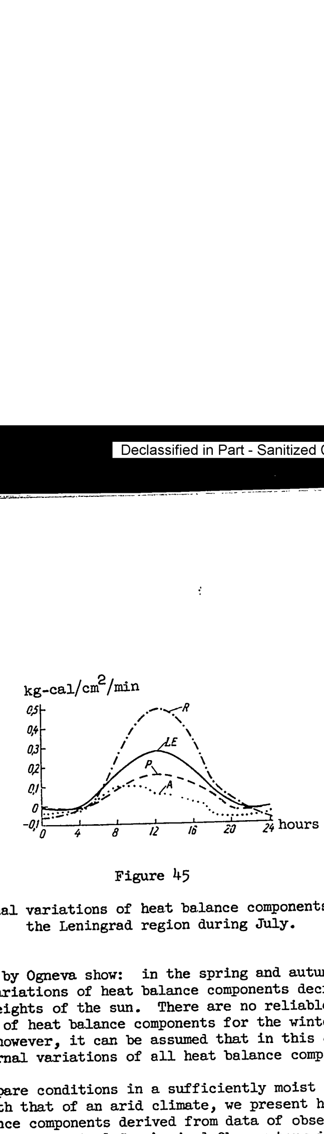

Figure 45

Diurnal variations of heat balance components in the Leningrad region during July.

As data derived by Ogneva show: in the spring and autumn months, the range of diurnal variations of heat balance components decreases with decreasing maximum heights of the sun. There are no reliable data on the diurnal variations of heat balance components for the winter season in the Leningrad region, however, it can be assumed that in this case, the range of the average diurnal variations of all heat balance components is very small.

In order to compare conditions in a sufficiently moist climate of moderate latitudes with that of an arid climate, we present here the diurnal variations of balance components derived from data of observations obtained by the expedition of the Central Geophysical Observatory in Pakhta-Aral (Middle Asia) during July 1952. Although the period of observation was rather short, it is safe to assume that the data of the observations sufficiently represent average conditions, due to the great stability of the desert and semidesert climate in summer.

The diurnal march of heat balance components in Pakhta-Aral, as derived from data given by Aĭzenshtat, Kirillova, Laĭkhtman and others (1953 /57/), is presented in fig. 46. In this case the similarity of patterns for annual and diurnal conditions of heat balance components is also well pronounced. As was seen in the annual march during the summer season, it can also be seen in the diurnal march during daylight hours that, positive values of radiation balance compensate the expenditures of heat for turbulent heat emission, and for heat exchange in soil. The turbulent heat flux markedly exceeds the heat exchange in soil. At night, the comparatively small heat expenditure is compensated by the influx of heat from the soil and air. The expenditure of heat for evaporation during 24 hours is close to zero due to to the lack of moisture in the soil.

The difference in the pattern of the annual and diurnal march of balance components is, in this case, associated with the relative length of the period with negative values of radiation balance. In the diurnal march this period is approximately 12 hours long, whereas, in the annual march, in deserts of Middle Asia, negative values of radiation balance are observed only during a comparably short period - several winter months.

In lower latitudes, negative monthly values of radiation balance are nonexistent and therefore in tropical and subtropical latitudes a similarity between the diurnal and annual marches does not exist either.

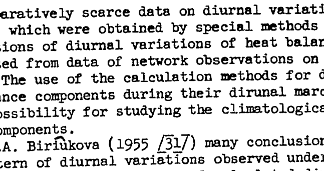

Figure 46

Diurnal variations of heat balance components in semideserts during July.

Besides the comparatively scarce data on diurnal variations of the heat balance components, which were obtained by special methods of balance observations calculations of diurnal variations of heat balance components were recently started from data of network observations on basic meteorological elements. The use of the calculation methods for determining changes in the balance components during their dirunal march has considerably widened the possibility for studying the climatological principles of the heat balance components.

In a paper by L.A. Biriûkova (1955 /31/) many conclusions were derived concerning the pattern of diurnal variations observed under various climatic conditions, on the basis of a vast amount of calculated diurnal variations of heat balance components for various climatic zones of the USSR. From this data of L.A. Biriûkova it can be concluded that, in the forest zone of moderate latitudes, in summer, the radiation balance is positive during the larger portion of the 24 hour period (14-15 hours); in winter, the radiation balance can be negative during the whole 24 hour period, which is effected by the low altitudes of the sun during daylight hours.

In the forest zone, the greatest amount of heat spent for evaporation is usually observed in the afternoon, whereas, the maximum of the turbulent heat emission is often observed earlier in the day.

In more southern regions of the steppe zone, the maximum of radiation balance, during its diurnal march in summer, is somewhat greater than the maximum of the forest zone, and the length of the period with positive radiation balance is correspondingly smaller.

The transition of radiation balance values through the zero value occurs at higher altitudes of the sun in the steppe, as compared with the forest zone, which can be explained by the greater effective outgoing radiation in southern latitudes during the summer.

In winter, in some regions of the steppe zone, radiation balance during daylight hours can still be positive (as high as 4-6 hours per day).

Further south, in areas of the desert zone in summer, noontime maxima of total radiation are typically very high. At the same time, corresponding maxima of radiation balance are comparatively lower (because of very large values of effective outgoing radiation) and usually do not exceed the maximum in the steppe.

In the deserts of Middle Asia the radiation balance near noon is, as a rule, positive during the entire year.

The transition of radiation balance through zero values, in the desert zone, is observed at higher altitudes of the sun, as compared with the steppe and forest zone.

In conclusion, we note that the diurnal march of heat balance components in the vast area of water surfaces is, at present not very well known due to the absence of any detailed and reliable observations. Available data are limited and only permit a conclusion that, in the warm season, the diurnal march of heat balance components in moderate latitudes, on the ocean,is often similar to their annual march. Accordingly, it might be assumed that, between diurnal variations of heat balance components at the same latitudes on land and oceans, a substantial difference must exist.

Chapter IV

Heat balance and the energy factors of physicogeographical processes

Notwithstanding the fact that the role of solar energy transformation in the dynamics of all physicogeographical processes occurring in the atmosphere, hydrosphere, and upper layers of the lithosphere, is well known, concrete forms of the relationship between solar energy transformation and the intensity of natural processes have been studied comparatively little. The very fruitful investigations by A.A. Grigor'ev present an exception. In these investigations he created a conception determining the significance of the heat and moisture balance for the formation of the physicogeographical medium.

A.A. Grigor'ev has indicated many times that natural processes in the external geographical sphere, including climatological, hydrological, pedological, exogenous geomorphological and biological processes, are closely interrelated with each other. This close association is determined to a great extent by the continuous exchange of some substances (water, nitrogen, carbon dioxide, etc.) between the atmosphere, hydrosphere, soil and living matter.

The total amount of all categories of organic and mineral substances in the outer geographical sphere, taken as a whole, changes along with the energy amount, only slowly in time and, for periods on the order of decades can be practically considered as a constant value.

This relatively small variability of substance quantities in the outer geographical sphere is connected with an insignificant (as compared with the available amount) intensity of matter exchange between the geographical sphere on one hand, and the deeper layers of the lithosphere and extra-atmospheric space on the other. In constrast to this, the reserve of energy in the outer geographical sphere is kept approximately at the same level in the presence of an intense exchange of energy with the extraatmospheric space, in which the gain of absorbed solar radiation is equal to the expenditure of heat energy for outgoing radiation from the earth.

This approximate constancy of very slowly changing reserves of various matter and energy categories in the outer geographical sphere as a whole, is associated with substantial and relatively rapid changes of these reserves in some localities. These changes have partly a periodical character (24 hours, annual period) and partly an aperiodic one.

The dynamics of change in reserves of matter and energy in the outer geographical sphere is, according to A.A. Grigor'ev, to a certain extent, determined by climatic energy resources; i.e., by conditions of solar radiation gain and by its consecutive transformation into heat exchange processes, changes of water phases, etc.

As has already been indicated, from the total amount of solar energy that is received by the earth, the major portion is absorbed on its surface.

Because of this the earth's surface is the main source of energy for the outer geographical sphere.

Insofar as the interaction of the majority of natural processes in the outer geographical sphere reaches its greatest intensity in the proximity of the earth's surface (where the river runoff is formed, the process of soil formation developed, and the main bulk of organic living matter is concentrated, etc.), it is obvious that, an accounting of solar energy transformations on the surface is of special significance for explaining the mechanism of the interaction and interdependence of the complexity of all the "outer" natural processes.

In some preceding investigations the author, developing ideas of A.A. Grigor'ev, has put forth the question about the possibility of using data on heat energy balance of the earth's surface for studying physicogeographical relationships by a deductive method, based on general physical laws. Such a direction of research creates, in my opinion, new possibilities in studying general geographical laws.

Going over to an outline of investigation results of the effect of heat and energy balance on the intensity of physicogeographical processes, we will first examine the problem of the relationship between the hydrological regime of land and climatic energy factors.

§ 10. The relationship between heat and water balance on land

The main characteristic of the hydrological regime of land, as is well known, is the norm of runoff - quantity of water, in the form of various horizontal water streams, that runs off, on the average during a year, from a land surface unit. An important index of hydrological conditions is the coefficient of runoff - relation of the runoff norm to the annual amount of precipitation.

Since the formation of annual runoff depends to a large extent on the process of evaporation, which is also related to one of the main processes of solar energy transformation on the earth's surface, it is obvious that the norm of runoff and the runoff coefficient is in a certain way associated with the principal components of heat balance.

A study of this relationship must facilitate the clarification of causative regularities which determine the hydrological regime features of various geographical zones.

The relationship between the components of heat and water balance on land was established on the basis of the following works: Budyko (1948a, 1948b, 1950a [39, 40 & 43]).

It is obvious that mean sums of evaporation from land surfaces depend on the amounts of precipitation and gain of solar energy. With increasing precipitation and radiation heat, evaporation also increases.

When the soil is very dry, all water that falls in the form of precipitation is retained by molecular forces on particles of soil and eventually is used up by evaporation. Under these conditions (which are primarily observed in deserts) the runoff coefficient approaches zero.

Considering the fact that the average dryness of soil increases with increasing radiation heat gain and with decreasing precipitation, it

can be concluded that:

$$\frac{f}{r}\to 0 \text{ or } \frac{E}{r}\to 1 \text{ at } \frac{R}{Lr}\to\infty. \qquad (100)$$

With decreasing $\frac{R}{Lr}$ the value of $\frac{E}{r}$ will decrease (some runoff arises).

With some sufficiently large amounts of precipitation and a sufficiently small gain of radiation heat a condition of permanent excessive moisture will be created in the upper soil layer. In this case the greatest possible amount of heat energy will be spent for evaporation from available resources. The amount of maximum expenditure could be evaluated by taking into account ventilation properties of turbulent heat exchange between the underlying surface and the atmosphere.

In the investigation done by the author and by M.I. Iudin (1948 [61]) and in other works it was noted that turbulent heat conductivity of the atmosphere's lower layer depends substantially on the direction of the vertical turbulent heat stream. In cases when the turbulent stream is directed from the earth to the atmosphere a comparatively high intensity of turbulent mixing can create very large values of this stream, which are quite comparable to the principal components of radiation and heat balance. With the turbulent stream of an opposite direction, inversional distribution of temperature considerably reduces the intensity of exchange and the turbulent heat stream is rendered relatively small.

The influence of the ventilation effect on turbulent heat exchange is clearly seen from graphs of the annual and diurnal march of the turbulent heat stream which are given in § 9. During winter, in moderate latitudes, due to prevalent temperature inversions, turbulent heat exchange is small as compared with summer maximum values which are brought about by the superadiabatic temperature distribution during daylight hours in the air layer near the ground.

A similar pattern is noted in the diurnal variations when we compare values of turbulent heat stream at night with those of daylight hours.

As a result of the ventilation effect annual sums of turbulent heat exchange are positive, i.e., the average turbulent heat stream is directed from the earth to the atmosphere in almost all climatic zones on land (see § 8).

Concluding that annual sums of turbulent heat exchange cannot provide a substantial gain of heat for the underlying surface, it must be understood that expenditure of heat for evaporation is compensated only by radiation balance, and therefore, the upper limit of increase of values LE equals R. In other words, it can be assumed for excessive moisture conditions that:

$$LE\to R \text{ at } \frac{R}{Lr}\to 0. \qquad (101)$$

Formulas (100) and (101) determine the relationship between $\frac{E}{r}$ and $\frac{R}{Lr}$

1l

$$\frac{E}{r}=\Phi\left(\frac{R}{Lr}\right) \qquad (102)$$

for: $\frac{R}{Lr}\rightarrow 0$ & $\frac{R}{Lr}\rightarrow\infty$

(where Φ — is some function).

It should be indicated that long ago, while analyzing data on precipitation and runoff, Schreiber, (1904) and E.M. Ol'dekop (1911 /180/) found a definite relationship between the water balance components, which was expressed in the following formulas:

$$E=r\left(1-e^{-\frac{E_0}{r}}\right) \qquad (103)$$

(the equation given by Schreiber and improved by Ol'dekop, where E_0 — is the greatest possible amount of evaporation under given conditions; e — the basis of the natural logarithms) and:

$$E=E_0\,\mathrm{th}\,\frac{r}{E_0} \qquad (104)$$

(the Ol'dekop's equation where th —(tanh)- is the function of the hyperbolic tangent).

It is easy to prove that equations(103) and (104) will suffice formulas (100) and (101), if under the considerations offered we assume that $E_0=\frac{R}{L}$.

It must be remembered, however, that the amount of possible evaporation in the given region will be determined by the radiation balance, which corresponds to conditions of sufficient moistening.

As data of observations and calculations show, the radiation balance of the active surface depends on conditions of moistening. The reasons for this dependence will be described in detail later, but now we will only take up the principal points. In the majority of geographical regions with a more or less moist climate (up to the steppe zone, except for drought periods) the albedo of the surface changes only slightly under changing conditions of moistening. This permits a use of evaporation calculations, and an application of observed values of the active surface's albedo; however, in computing effective outgoing radiation the fact that the temperature of the active surface does not differ from air temperature (as it is usually observed in regions with abundant precipitation) should be taken into account. Since average differences in the surface and air temperature are comparatively small, in regions with sufficient moisture, it was often possible to approximately determine the amount of evaporability from radiation balance computed for actual conditions of the active surface.

It is quite different in dry climates, where the albedo and temperature of the active surface change considerably with moistening and under sufficient moistening the surface temperature approaches the air temperature. It is obvious that, for determining evaporability in this case, it is proper to use the amount of radiation balance found for the albedo of a moist surface, and take air temperature into account when calculating the effective outgoing radiation.

The calculation of radiation balance under such conditions is not difficult and can be easily accomplished by using the method described in § 3.

For verification of the above considerations about the character of the dependence of ratio $\frac{E}{r}$ on $\frac{R}{Lr}$ with small and great values of the latter parameter, diversified factual data were used in the preceding works.

As an example we will compare the values of $\frac{R}{Lr}$ with those of $\frac{E}{r}$ taken from the summary given by Wundt (1937) for river basins of various continents(except mountain regions).

In fig. 47, the dependence of $\frac{E}{r}$ on $\frac{R}{Lr}$ for small values of $\frac{R}{Lr}$, in accordance with formula (101), is presented by a straight line OA and for great values of $\frac{R}{Lr}$, in accordance with formula (100) - by line AB. The dots show average values of the ratio $\frac{E}{r}$ which were obtained from data on water balance by averaging the values of $\frac{E}{r}$ for certain intervals of the parameter $\frac{R}{Lr}$: The experimental dots show that, in reality there is a smooth transition in the ratio of $\frac{E}{r}$ and $\frac{R}{Lr}$ from the OA regularity to the AB, regularity which, as was assumed above, are of a limited nature.

The good agreement obtained here is of special significance because of diversified physicogeographical conditions covered by the utilized experimental results.

In order to present formula (102) in an analytical form, the following formulas similar to those of (103) and (104) can be used:

$$E=r\left(1-e^{-\frac{R}{Lr}}\right) \qquad (105)$$

and

$$E=\frac{R}{L}\,\mathrm{th}\,\frac{Lr}{R}, \qquad (106)$$

or, the geometrical mean of these relationships can be used:

$$E=\sqrt{\frac{Rr}{L}\,\mathrm{th}\,\frac{Lr}{R}\left(1-\mathrm{ch}\,\frac{R}{Lr}+\mathrm{sh}\,\frac{R}{Lr}\right)} \qquad (107)$$

(ch and sh - are hyperbolic functions of the cosine and sine). This formula suits the majority of experimental data better than the first two.

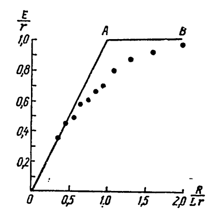

Figure 47

Dependence of the evaporation and precipitation ratio on the radiational index of dryness.

Equations (105) - (107) establish the dependence of the mean annual evaporation on precipitation and radiation balance.

Since the norm of runoff equals the difference between precipitation and evaporation, it is obvious that by using these formulas the runoff can also be determined from data on precipitation and radiation balance.

Equations for runoff and runoff coefficient corresponding to formula (107) will take this form:

$$f = r - \sqrt{\frac{Rr}{L}\,\mathrm{th}\,\frac{Lr}{R}\left(1 - \mathrm{ch}\,\frac{R}{Lr} + \mathrm{sh}\,\frac{R}{Lr}\right)} \qquad (108)$$

and

$$\frac{f}{r} = 1 - \sqrt{\frac{R}{Lr}\,\mathrm{th}\,\frac{Lr}{R}\left(1 - \mathrm{ch}\,\frac{R}{Lr} + \mathrm{sh}\,\frac{R}{Lr}\right)}. \qquad (109)$$

Formulas (108) and (109), which establish the relationship between the components of heat and water balance, have been verified in many investigations; particularly in the author's work (1951b /46/) for such quantitative verification data on water balance of European rivers were used, as presented by Wundt in the above mentioned summary. From this summary data were taken for comparatively large basins (with a surface larger than 10, 000 km^2) where the effect of local factors on runoff must be small. In the table of Wundt, data are given on the average amount of precipitation and runoff for 29 rivers of Europe with basins of sufficiently large dimensions. The coefficients of runoff for these rivers vary from 0.64 (Ilo) to 0.13 (The Southern Bug River), which is the result of very diversified physico-geographical conditions.

The comparison of runoff coefficients, calculated by formula (109) $\frac{f}{r}$, with data given by Wundt $\left(\frac{f}{r}\right)_0$ is presented in fig.48. The distribution of dots

on this graph shows a very good agreement between values of the runoff coefficient that were obtained from hydrometeorological observations and those derived by theoretical methods.

The mean discrepancy between absolute values of calculated and measured coefficients of runoff equals 0.04; the mean relative error of the calculations (mean ratio of absolute discrepancies to corresponding measured values) was 13%. It would be interesting to compare this value with the mean relative calculation error of the runoff coefficient from these data, by means of Wundt's empirical nomogram. It should be noted that this nomogram, which gives values of runoff dependent upon annual sums of precipitation and mean annual air temperatures, is based on data that were included in Wundt's summary; in this nomogram all of those data which were used for verification in the case examined were taken into account.

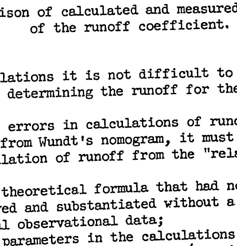

Figure 48

Comparison of calculated and measured values of the runoff coefficient.

From Wundt's calculations it is not difficult to see that his nomogram gives a mean error in determining the runoff for the 29 named rivers as being 22%.

Comparing relative errors in calculations of runoff from the " relation equation" and from Wundt's nomogram, it must be kept in mind that best results in calculation of runoff from the "relation equation" were obtained:

(a) by using the theoretical formula that had no uncertain coefficients, and was derived and substantiated without a direct use of hydrometeorological observational data;

(b) by using two parameters in the calculations, one of which coincides with the parameter accepted by Wundt (precipitation), and the other (radiation balance), as distinct from the second parameter of Wundt (temperature), was only given as a mean latitudinal distribution for the whole continent.

Considering these circumstances it must be acknowledged that the "relation equation" provides for a greater effectiveness in determining the norm of runoff as compared with the empirical method of Wundt.

The comparatively great errors in runoff calculations by Wundt's method, for those basins whose data were used in constructing the calculation diagram, can be explained to a great extent by the improper application of the mean annual air temperature as a parameter which characterizes thermal conditions of evaporation. We will not go into details on the problem concerning the principal possibility of using data on air temperature for evaluating thermal conditions of physicogeographical processes on the surface (this question will be examined in more detail later); we will note only that, in moderate latitudes, the severity of winter does not affect the annual amounts of runoff to any considerable degree, however, the mean annual temperatures depend largely on this factor. Concerning this, at places of equal mean annual temperatures in marine and continental climates, thermal conditions of evaporation will be entirely different; this cannot be accounted for by Wundt's nomogram or by other similar relationships.

Referring to the analysis of results for the above mentioned calculation of runoff for 29 basins of European rivers by the "relation equation," it must be noted that the mean relative error of this calculation is determined to a great extent by comparatively great relative errors in the calculation of runoff for some rivers with small coefficients (and norm) of runoff.

For 20 basins, whose runoff coefficient is equal to or more than 0.30, the mean relative error of calculation by the formula decreases to 7%.

This relative error value approaches the mean error of measurements for precipitation and runoff. Hence, it can be concluded that, for great basins with not very small coefficients of runoff, the discrepancy between the mean climatic and hydrologic runoff, if it exists, is found in limits of few per cent, and generally it cannot be discovered with the presently existing rate of accuracy in determining precipitation and runoff.

The comparatively great relative errors in determining runoff coefficinets for basins with small runoff are partly explained in this case by the considerable influence of small absolute errors. At the same time, in examining fig. 48, it can be noted that errors of calculation are not quite accidental under circumstances of small runoff coefficients.

The existence of a systematic discrepancy between the calculated and observed runoff coefficients in this area of the graph, apparently indicates some inaccuracy in the interpolation function that was used for the derivation of equation (102): for the comparatively large value of $\frac{R}{Lr}$, the values of $\frac{f}{r}$ proved to be somewhat exaggerated. Such an error, which is due to the schematic way of selecting tne interpolation function, has of course no principal importance, and, if necessary, could be eliminated.

It should be noted that in determining evaporation by the "relation equation," the relative accuracy of the calculations will be high even with small coefficients of runoff, since in this case the error in the interpolation function only slightly affects the results.

Later, the "relation equation" was verified by use of Soviet Union data by P.S. Kuzin (1950 [143] who also obtained positive results. Considering this, it should be noted that, in the paper by N.A. Bagrov (1953 [16]), his statement is not quite correct when he indicates the existence of considerable errors in the evaporation calculations by the "relation equation" for the extreme northern part of the USSR. Kuzin, whom Bagrov refers to in this case, assumed that the discrepancy between the evaporation calculations by the "relation equation" and water balance for this condition, is basically associated with errors in accounting for solid precipitation, and proves the insufficient accuracy of the water balance method. This interesting question needs further investigation with the use of snow survey data and utilization of recent data on evaporation.

From other critical remarks concerning methods for calculating components of water balance by the "relation equation," it is worthwhile to mention the reasons given by D.L. Sokolovskiĭ. In his well known textbook (1952 [214]), a supposition is presented that the "relation equation" is good primarily for practical calculations of evaporation, whereas calculations of runoff would be of a low accuracy. In view of the above data on verification of the "relation equation," it can be assumed that this remark is valid for cases with small runoff coefficients when the errors of calculation, though small by absolute value, could still be of a great relative value. However, for cases with medium and large runoff coefficients the relative accuracy of runoff calculations by using the "relation equation" will be quite comparable to the accuracy of evaporation calculations.

Such a conclusion is confirmed by the results obtained by L.I. Zubenok (see 4), who verified the "relation equation" by utilizing all available data on the hydrological regime of various continents.

It is interesting to note that the "relation equation" can be utilized not only for calculations of water balance components from the heat balance (as it was done in papers by P.S. Kuzin, 1950, 1953 [143 & 144], and others), but also for calculating radiation balance on land from data on water balance. Such an indirect method for calculating radiation balance, used by Zubenok (1949a [105]), has permitted the construction of radiation balance maps, which are very similar to those constructed by the usual methods. This agreement once more proves the sufficient accuracy of the "relation equation" which shows the relation between components of heat balance and those of water balance.

Recently some efforts have been made to generalize the "relation equation" in order to estimate the effect of some additional factors on evaporation and runoff. In investigations by N.A. Bagrov (1953, 1954a [16 & 17]), there is a suggestion to modify the form of the function Φ in equation (102) for evaluating the effect of additional factors on evaporation. Bagrov assumed that parameter n, which determines the form of the indicated relationship is "characteristic of the zonal-landscape conditions which involve many factors, such as: general conditions of downflow, waterproofness of soil, agricultural measures, etc." (1953 [16]). This idea on the method of estimating the effect of indicated factors on evaporation is generally correct, however, concrete calculations performed in his work raise a rather substantial doubt.

It was established in investigations of E.M. Ol'dekop (1911 [180]) that, the effect of all factors on mean annual evaporation from sufficiently large basins, except that of precipitation and evaporability, is comparatively small. This has been also acknowledged by N.A. Bagrov. He noted that factors which are accounted for by variations of parameter n "play a comparatively secondary role" (1953 [16]). Under these conditions the establishment of a true relation between parameter n and the factors under investigation requires a very high accuracy in determining basic hydrometeorological indices: precipitation, runoff and evaporability.

The evaluation of the effect of possible errors, in determining these values, on evaporation calculations shows the extreme difficulty involved in the quantitative study of the dependence of evaporation on additional factors, when based on the comparison of evaporation calculations from the water balance and those made by the "relation equation." This fact, at the present, limits the practical importance of the theoretically interesting suggestions by Bagrov 1).

The "relation equation" presents a general form of the dependence of runoff on evaporation, annual sums of precipitation and radiation balance. The corresponding relationships are presented in graphs 49 and 50. These regularities explain many of the empirical relationships between runoff and precipitation, which have been established in earlier hydrological investigations.

Using the "relation equation" we can construct a graph which will show the dependence of runoff on precipitation for various mean values of radiation balance.

Such a theoretical dependence, calculated for average conditions of the European plain is presented in fig. 51 by curve A.

Also given for comparison are:
1) B (a straight line) based on the empirical equation by Keller (1906), which relates the norm of runoff and precipitation according to observational data on the rivers of western Europe,
2) C the empirical curve by D.L. Sokolovskii (Sokolovskii, 1936 [213]), shows the relation between runoff and precipitation according to observational data pertaining to river basins of eastern Europe.

An accurate agreement between theoretical and empirical relationships is again, in this case, a confirmation of the "relation equation's" correctness.

On the basis of the "relation equation" it is also possible to explain the considerable scattering of the dependencies of runoff on precipitation data from various regions, which is shown in derivation of mean empirical relationships (B.P. Poliakov, 1946 [187]). A considerable variability of radiation balance in moderate latitudes results is due to the fact that runoff from basins with a large balance (i.e., in more southern regions) is

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1) The scheme of N.A. Bagrov was also recently discussed in the paper of V.S. Mezentsev (1955 [169]) in which some improvements on methods for calculating evaporation from land surfaces are suggested.

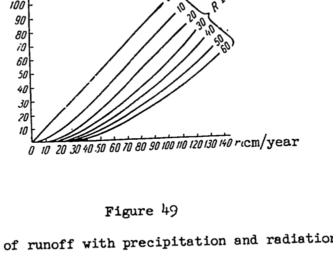

Figure 49

Relationship of runoff with precipitation and radiation balance.

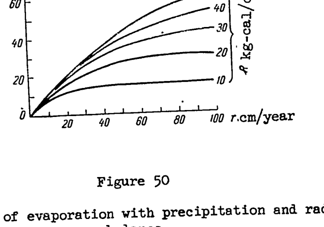

Figure 50

Relationship of evaporation with precipitation and radiation balance.

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considerably smaller than from basins with a smaller balance (more northern regions), under equal amounts of precipitation. At the same time, the speed of changes in runoff varies with increasing precipitation $\left(\frac{df}{dr}\right)$.

According to the "relation equation," it should be greater in northern basins than in southern ones.

This relationship is well confirmed by factual data.

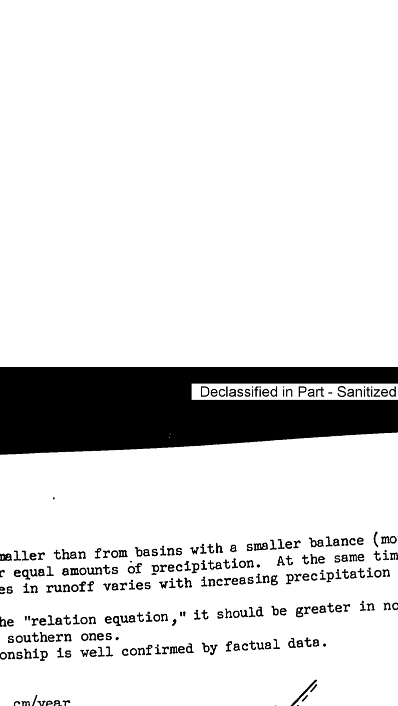

Figure 51

Relationship of runoff and precipitation.

Among other possible applications of the "relation equation" in the investigation of the runoff relation with climatic power factors, the question of runoff changes in mountain regions should be considered. As it is well known, in mountains, up to great elevations, an increase in runoff is usually observed with increasing heights, which is associated with a simultaneous increase in precipitation.

In fig. 52 data on variations of runoff and precipitation with height are presented for the river Aar, up to 2km (M.I. L'vovich, 1945 [162]) and data on runoff and precipitation for 5 classes of the dissected terrain in the basin of the river Tissa (B.V. Poliakov, 1946 [187]).

Considering that for lower elevations the value of radiation balance changes only slightly with height (in this case, some increase in radiation and effective outgoing radiation occurs, which compensate each other), and considering the given data on precipitation, the corresponding values of runoff could be calculated by using the "relation equation". Results of these calculations are presented in fig. 52 by line (3), which coincides very well with the given data of measured runoff (2).

Thus, it must be concluded that for mountain basins of the rivers Aar (up to 2km) and Tissa, the observed variations of the runoff sums and coefficient of runoff are completely explained by the increase in precipitation, i.e., by climatic factors. A direct effect of the orography on runoff is, in this case, almost imperceptible.

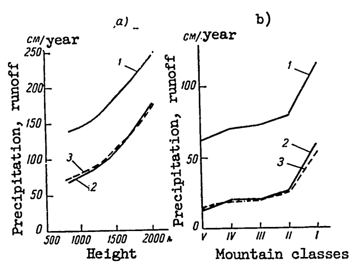

Figure 52

The changes of runoff with height a)-Aar River, b)-Tissa River
1- precipitation, 2- measured runoff, 3- calculated runoff.

This conclusion, sound as it apparently is, is not of a universal nature for many mountainous basins. It does not work in high mountain basins, and also in basins with very rapid changes of the underlying surface with increased heights (a direct effect of orographic conditions on runoff could be determined, for instance, from data on the rivers of the North Caucasus, which are given by L.K. Davydov, 1947 [89], and others).

A good agreement between results obtained by calculating runoff with the "relation equation" of heat and water balances and various factual data, confirm the great and determining effect of climatic factors (in particular heat energy factors) on the formation of annual runoff sums 2).

The estimation of the effect of radiation balance on components of water balance not only permits clarification of some genetic regularities in the formation of the hydrological regime on land, but also opens new possibilities in solving some practical problems which are associated with the necessity of calculating amounts of evaporation and runoff.

In §11 the "relation equation" is used in clarifying some causal principles which determine the phenomenon of geographical zonality.

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2) It is necessary to emphasize here that this conclusion pertains only to river basins of considerable dimensions, which can be compared with the scale of geographical zones. In small plots, river runoff can change in wide limits under the influence of local conditions of a nonclimatic nature.

§ 11. Heat balance and geographical zonality

The climatic indices of geographical zonality.
One of the main problems in physical geography is the explanation of the geographical zonality phenomenon, which was discovered at the beginning of the 20th century by the outstanding soil scientist V.V. Dokuchaev (1900 and others).

In investigations by V.V. Dokuchaev it was established that the distribution of geographical zones is determined, to a considerable degree, by climatic factors and particularly depends substantially on climatic conditions of moistening. In his papers, the interrelationship between precipitation and possible evaporation was established for the principal geographical zones. The investigations of Dokuchaev stimulated numerous works that were devoted to the study of relations existing between the distribution of soil and vegetation on one hand, and the relationship of precipitation to possible evaporation (or their indirect characteristics) on the other.

Among investigations in this direction the investigation by G.N. Vysotskii (1905 [71]) should be mentioned. In this work, the relationship of annual amounts of precipitation and evaporability (which was considered as being equal to evaporation measured by water evaporimeters) was computed. The values obtained were used by Vysotskii in an analysis of conditions of soil types formation and development of vegetation in various natural zones.

In the paper by Transeau (1905), for the purpose of studying climatic factors of forests distribution, a map is given showing the relationships of precipitation to evaporation from water evaporimeters in North America.

Somewhat later Penk (1910) suggested a climatic classification in which climates were divided into three groups: climates with precipitation exceeding evaporation, with precipitation equal to evaporation and precipitation less than evaporation.

In subsequent investigations the majority of authors, who were studying climatic conditions of moistening on the basis of calculated relationships of precipitation and possible evaporation (evaporability), rejected data obtained by water evaporimeters, in determining evaporability.

The reason for this was, on one hand, an insufficient quantity of such data for many regions, and on the other hand, the difficulty in physically interpreting the readings of evaporimeters used by the network of meteorological stations (eventually use of water evaporimeters at meteorological stations was rejected and, particularly at stations in the network of the Soviet Union, the Wild evaporimeter was removed).

The majority of consequent investigations concerning moistening conditions on land can be divided into three groups by their indices, which were used for determining evaporability. Those groups are: 1) where data of saturation and deficit are used, 2) data on air temperature are used, and 3) data on radiation balance are used.

An extensive use of saturation deficit data was made, for the first time, in studying moistening conditions in the paper by E.M. Ol'dekop (1911 [180]), which has already been mentioned many times above.

For determining evaporation from the surface of river basins Ol'dekop suggested a formula (cited in § 10) in which the index of climatic conditions of moistening is $\frac{r}{E_0}$, - the ratio between precipitation and evaporability (the possible evaporation). For determining the value of evaporability, Ol'dekop used the formula $E_0 = ad$, where E_0 — evaporability, d — saturation deficit, a — coefficient of proportionality. The value of coefficient a, according to Ol'dekop is equal, for the warmer half of the year, to 22.7 (saturation deficit was here determined in mm and evaporability in mm/month) and for the colder half of the year it was 16. Consequently the mean annual value of this coefficient is 19.3.

Using this value we arrive at a conclusion that the mean annual value of the relationship $\frac{r}{E_0}$ is equal to $\frac{r}{232d}$, where d — is the mean annual saturation deficit.

The formula suggested by Ol'dekop for determining evaporability is not anything new - many authors used it earlier for determining evaporation from the surface of water reservoirs. However, Ol'dekop apparently was the first who applied data on saturation deficit to calculate evaporability and the relationship between evaporability and precipitation.

Ol'dekop was also the first to make an attempt at a physicomathematical analysis of the dependence of hydrometeorological regime indices - evaporation and runoff - upon the relationship of evaporability and precipitation.

Later, the formula $E_0 = ad$ was used in investigations by P.S. Kuzin in calculations of evaporation from land surface in moist climates (for pertinent regions it was consequently assumed that evaporation was basically determined by evaporability). The values of coefficient a were obtained by Kuzin as being approximately 14 (1934 [140]) and 15 (1938 [141]) (in the same dimensions).

In 1926 Meyer used, as characteristics of the moistening condition, the relationship of the annual precipitation sum and mean annual saturation deficit. This value, eventually named the *NS*- coefficient, was largely used later by many foreign authors in investigating relations between the distribution of various types of soil and vegetation, and climatic conditions. Apparently, the *NS*- coefficient is proportional to the index suggested by Ol'dekop and is larger than the value of this index by 232 times.

In 1931 Prescott found that the Meyer index divided by 230, coincides with the relationship value of precipitation to evaporation from the water surface. However, Prescott made no reference to the earlier investigation by Ol'dekop.

Later on, in papers by N.N. Ivanov (1941, 1948 [112 & 113]) the following formula for determining evaporability was suggested:

$$E_0 = 0{,}0018\,(\theta + 25)^2\,(1 - h) \text{ mm/month}$$

where θ — the mean monthly air temperature, h — mean monthly relative humidity. Since the value of $(\theta + 25)^2$ is approximately proportional to saturated vapor pressure within the interval of temperatures which are

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usually observed accompanying more or less intensive evaporation, this formula practically coincides with: $E_0 = ad$.

Coefficient a is here equal to 18.4 (this was noted by A.M. Alpat'ev, 1950 [11], and later acknowledged by N.N. Ivanov, 1954 [114]), i.e., close to the mean value obtained by Ol'dekop. The annual value of the moistening index was, according to Ivanov, equal to $\frac{r}{221d}$.

In the cited papers Ivanov performed calculations of the precipitation relationship to evaporation for many stations.

Obviously, the method of calculating the moistening index used by Ivanov did not introduce anything new when compared with earlier accomplished works, since the index of moistening presented by him is numerically closer to the index developed by Ol'dekop, and is proportional to the NS- coefficient by Meyer (221 times smaller than this).

However, insofar as N.N. Ivanov did not mention the investigation by V.V. Dokuchaev, G.N. Vysotskii, E.M. Ol'dekop, Meyer, and others in his introductory survey, we find an erroneous opinion expressed in literature, even up to these days that, the "index of moistening" concept belongs to N.N. Ivanov and that his index differs substantially from other indices of moistening that were suggested earlier.

We will not cite many of the other investigations, in which the relationship of precipitation and saturation deficit (or precipitation and a value proportional to saturation deficit) was used as a characteristic of moistening conditions, but one important thing should be mentioned.

The single reason for the assumption that evaporation is proportional to saturation deficit, was the conception concerning evaporation from water surfaces as being proportional to saturation deficit. This conception, which was very popular in hydrology some decades ago, was later disproved. There are many theoretical investigations and experimental projects of this problem available. The experimental projects established that evaporation from limited water reservoirs is determined by the formula:

$$E_0 = ad^n,$$

where n — is a value less than one. The value of parameter n has been found by various authors: by O.S. Poznyshev (1937 [184]) - 0.5; V.N. Moklîak (Ogievskiî, 1937 [178]) - 0.7; V.K. Dabydov (1944 [88]) - 0.8; B.D. Zaîkov (1949 [102]) - 0.78; and others. The reason why evaporation from water reservoirs is not proportional to saturation deficit is explained in these investigations.

We note here the interesting fact that in the latest paper by Prescott, 1949 (who earlier used, rather extensively, the NS- coefficients for analysis of moistening conditions, 1931, 1934), in order to achieve a better agreement between isolines of moistening index values with borderlines of soil type zones, a suggestion is made to use the saturation deficit value with a power factor of 0.7 in the denominator of the value of the NS- coefficient. This statement indirectly reflects the fact, well known by now, that the evaporation value from the water surface usually increases considerably slower than the saturation deficit.

The concurrence of the isolines of relationship for precipitation and saturation deficit with boundaries of the vegetation or soil type zones was investigated in great detail for the NS-coefficient. The majority of investigations on this problem (by Jenny, 1941, V.P. Volobuev, 1953 [70] and others) proved that the distribution of the NS- coefficient only shows an approximate coincidence with boundaries of the natural zones, and in many cases it is interrupted. From all that has been said above it is obvious that, this conclusion also pertains to Ol'dekop's index (and also to the index used by Ivanov) insofar as they are proportional to the NS-coefficient.

Thus, the problem concerning theoretical and empirical substantiation of the feasibility to apply the relationship between precipitation and saturation deficit as an indication of moistening conditions, is not quite clear in the light of recent investigations.

The second group of investigations on climatological indices of moistening is based on an application of the evaporation characteristics derived from data on air temperature.

Among these investigations we will mention the papers by Lang (1920), and Martonn (1926), who suggested the indices $\frac{r}{\theta}$ and $\frac{r}{\theta + 10}$ as characteristics of moistening, where θ — is the mean annual temperature. From the climatological point of view it is very obvious that these indices are not satisfactory. It is well known that the mean annual temperature is often very much dependent on thermal conditions during the cold season of the year, which generally have little effect on the formation of physicogeographical zones.

Some samples are given here which will clarify this statement. The majority of the authors assume that rational indices of moistening must be of a more or less constant value on boundaries of certain geographical zones.

According to calculations made by G.S. ZHegnevskaîa (1954 [99]), on the boundary of wooded steppe and steppe, Lang's index changes from 50-100, in European Russia, to indefinitely great values, and then to negative values in Siberia. In North America the value of Lang's index on the same boundary, from north to south, noticeably diminishes from 100 to 40.

Martonn's index is not so changeable, but still varies in rather wide limits - from 25-30 on the boundary between wooded steppe and steppe in European Russia, to 40 and more on the same boundary in Siberia. These changes in Lang's and Martonn's indices indicate their incomparability with actual conditions of moistening. 3)

Much more interesting, in comparison to Lang's and Martonn's indices,

3) Similar considerations could be referred to the calculation method of moistening conditions used in the well known climatic classification by Köppen (1931 and others). The use of mean annual temperatures as an index of evaporation is, of course, a great deficiency of this classification, which in many other respects is very valuable.

are the indices in which thermal characteristics of the growing season are utilized. Such indices were developed in many investigations, among which the studies of G.T. Selianinov (1930, 1937 [208 & 209] and others) deserve special attention.

As a characteristic of possible evaporation, G.T. Selianinov suggested a comparison of the sum of temperatures, with the amount of precipitation. The practical usefulness of this well known index has been proven by numerous agrometeorological researches. This hydrothermal coefficient of Selianinov has been used in many agroclimatic works. However, it has never been empirically tested on a wide range of geographical data. On the other hand, this coefficient had no detailed theoretical basis either, and in some cases doubts were raised concerning its actual value.

These circumstances called for an additional study of the relationship between the sums of temperature and evaporability.

The third group of investigations on climatic conditions of moistening for estimating the value of possible evaporation (evaporability) is based on radiation energy balance data.

The idea of utilizing data of the radiation regime, in studying climatic conditions of moistening in their relation to physicogeographical processes, was suggested and substantiated by A.A Grigor'ev. In his works, Grigor'ev (1946, 1948 and others [81 & 82]) indicated that the relationship of precipitation and radiation balance is of a determining nature for the development and intensity of basic physicogeographical processes. According to this statement, Grigor'ev specified some laws governing the relationship between radiation balance and precipitation on one hand and general physicogeographical conditions in various areas of the terrestrial globe on the other.

Further developing the idea of A.A. Grigor'ev, we will note the following feature. The equations of heat and water balance on land can be written in the following form for the year:

$$\frac{R}{Lr}=\frac{E}{r}+\frac{P}{Lr} \qquad (110)$$

and

$$1=\frac{E}{r}+\frac{f}{r} \qquad (111)$$

(Terms of the heat balance equation are divided by Lr, and those of the water balance by r).

To these relationships we can add the "relation equation" between the heat and water balance that was derived in the preceding section.

$$\frac{E}{r}=\Phi\left(\frac{R}{Lr}\right).$$

These three equations tie up the four relationships of the terms of heat and water balance. Consequently, in this case, only one of these terms is an independent variable.

Taking the specific nature of the form of the "relation equation" into

account, it should be concluded that, as a parameter which determines all other values of the heat and water balance terms, the relationships of $\frac{R}{Lr}$ and $\frac{P}{Lr}$ can be utilized (the ratios $\frac{E}{r}$ and $\frac{f}{r}$ are of little practical use in determining the first two values in dry climates when small changes in $\frac{E}{r}$ or $\frac{f}{r}$ correspond to very great changes in $\frac{R}{Lr}$ and $\frac{P}{Lr}$).

Since the turbulent flux of heat value is usually determined with less accuracy than the radiation balance, it is obvious that it would be most expedient to adopt the value of $\frac{R}{Lr}$ as the basic parameter when determining the relationship between terms of heat and water balance equations.

This parameter can be considered as the relationship of evaporability $\frac{R}{L}$ and precipitation, or as the relationship between radiation balance and expenditure of heat for evaporation of the annual precipitation amount.

Thus, using this parameter, we are simultaneously taking into account the idea of the moisture index suggested by Dokuchaev-Vysotskii and the postulate of A.A. Grigor'ev concerning the nature of the radiation balance and precipitation relationship as being indicative of moistening conditions.

When relative values of the heat and water balance terms are determined by one parameter $\frac{R}{Lr}$, then, absolute values of these terms are determined by two parameters, for instance $\frac{R}{Lr}$ and R. This must be kept in mind in subsequent analyses.

The index of climatic moistening conditions $\frac{R}{Lr}$, which can be called a radiational index of dryness, must be calculated by using data on radiation balance, which corresponds to sufficient moistening conditions of the underlying surface. This postulate, resulting from theoretical foundations of the method for calculating evaporability from radiation balance, is closely connected (as will be shown later) with general principles of calculating evaporability from meteorological parameters.

From various data, which comprise the theoretical and empirical basis of the suggested index of dryness, we will select the most important conclusions.

It is obvious that, from the general point of view, possible evaporation from the land surface depends on many meteorological factors. The most important of them are - solar energy balance, moisture of the air and air temperature. Therefore it can be concluded that, every method of determining evaporability by only one of these elements will inevitably be an approximation and will include a more or less greater error. For a detailed evaluation of the accuracy of all approximate methods, they must be compared with a method which takes into account all of the basic factors affecting evaporation.

The general method for calculating evaporability, which takes into account the effect of radiation balance, humidity, air temperature and the influence of turbulent exchange, was suggested by the author (1951a [45]), and was used for verifying the method of calculating evaporability by radiation balance. This verification was accomplished for two regions in European Russia and gave good results for annual values and also for monthly values of evaporability.

It is obvious that the method for calculating evaporability from radiation balance must be tested in a similar way for all climatic zones of the terrestrial globe (see page 169, i.e. page 165 of this translation).

Another important substantiation of the method for determining evaporability from radiation balance is the derivation of the "relation equation" and its verification against various factual data (see §10).

Lately the radiation index of dryness $\frac{R}{Lr}$ has been largely used in various geographical investigations, including studies by A.A. Grigor'ev concerning general problems of physical geography (1951, 1954 [83 & 84]), papers on regional climatography (for instance, by Orlova, 1954 [182]), etc. In the monograph by A.G. Isachenko (1953 [115]) some critical remarks concerning the method for calculating evaporability from radiation balance have been given pointing out that the radiational index of dryness "presents the most concise and physically understood index of climatic moistening." Isachenko has, at the same time, expressed a doubt concerning the use of annual sums of radiation balance in calculations of evaporation for the year. This question is analyzed in more detail on pp. 177-178 (pp. 172-174 of this translation).

Another remark by Isachenko concerns the difficulty in using the radiation index of dryness for the evaluation of moistening conditions for individual seasons. This remark is justified, but it must be equally referred to all forms of the Dokuchaev-Vysotskiĭ index, due to the paramount importance of soil moisture redistribution for the regime of moistening during single seasons in the majority of the climatic regions.

Moreover, Isachenko also assumed that radiation balance does not always cover losses of heat for evaporation, and due to this the turbulent flux of heat furnishes, in many regions, a portion of heat to be spent for evaporation.

In view of results described in chapter III, we may assume that the turbulent flux of heat may have a large significance in furnishing annual amounts of heat spent for evaporation from land, in two cases - in abundantly irrigated oases of dry climates (the significance of the turbulent flux of heat will be the greater, the smaller is the oasis itself), and in climates with a permanent or almost permanent snow cover.

Some features of the method for calculating evaporation from radiation balance have lately been outlined in the article by N.N. Ivanov (1954 [114]).

However, it must be regretfully pointed out that, almost all critical remarks made by Ivanov were based on insufficient knowledge of the contents of the analyzed works.

Thus, for instance, a large part of the article is devoted to the analysis of the evaporability values which he computed by using the radiation balance method, and he calls it "the evaporability according to Budyko." These data, however, have no relation to the method for calculating evaporability that was developed by the author. We have repeatedly emphasized (see Budyko, Drozdov and others, 1952 [56]) that for determining

the value of evaporability, those data on radiation balance which were computed for a moist surface should be used.

Nevertheless, N.N. Ivanov calculated evaporability "according to Budyko" from the radiation balance computed for the actual state of the surface, i.e., from an index of balance which can differ greatly from the balance for a moist soil surface (the question of the moistening effect of the active surface on radiation balance is described in detail in chapter V). Therefore, Ivanov had no reason to attribute to the author those results of evaporability calculations from radiation balance which he obtained.

Another argumentation by Ivanov, which occupies a prominent place in his article, is also based on a misunderstanding. Ivanov maintains that when evaporability is calculated from radiation balance, the effect of atmospheric circulation on evaporation processes from the earth's surface is "completely ignored."

Apparently, Ivanov has forgotten that atmospheric circulation does not exert any immediate effect on heat exchange processes of the earth's surface (see §1). At the same time an indirect effect of circulation on these processes and on radiation balance of the earth's surface is very great. [4]

In the author's researches, which were devoted to this problem, a method for calculating the effect of heat exchange, between the underlying surface and atmosphere, on evaporation, was developed. In the same investigations maps of the distribution of heat exchange between the earth's surface and atmosphere were constructed for the first time. These maps give quantitative answers to the question concerning the effect of heat exchange with the atmosphere on heat balance and heat expenditures for evaporation from the earth's surface. In these works (and also in §10) the question concerning the effect of the heat exchange between the earth's surface and atmosphere on evaporation was also analyzed in deriving the "relation equation" by taking into account contemporary conceptions about laws governing turbulent heat exchange in the air layer near the ground.

Since N.N. Ivanov did not examine the substance of the author's investigations of this problem, and even made no mention of them, it is obvious that a discussion of his statement that the author "completely ignores" atmospheric circulation, would be rather difficult.

In the latest investigations on evaporability (ÎAkovleva, 1953 [246]; Budyko, ÎUdin, ÎAkovleva, 1954 [64], and others) the question concerning the effect of the moistened surface's dimensions on the amount of evaporation was analyzed in detail. These works lead us to the conclusion that the effect of the dimensions of the moistened surface on evaporation is more pronounced in dry climates, where great contrasts in temperature and humidity take place between irrigated and nonirrigated land. In this case

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4) It must be pointed out here that, in absence of atmospheric circulation, radiation balance on the land surface in all latitudes would be equal to zero in the mean annual value. Therefore, it can be concluded that atmospheric circulation exerts a much stronger effect on the distribution of radiation balance than on any such meteorological element as air temperature.

the value of evaporation from a small irrigated plot may markedly exceed the value of $\frac{R}{L}$, however this phenomenon will be of great importance only for irrigated plots of a small size - mainly those not larger than a few hundred meters.

It is worthwhile to notice that these conclusions, obtained as a result of a theoretical analysis, were recently well confirmed by data of many experimental investigations, including data obtained by the expedition of the Central Geophysical Observatory to the oasis of Pakhta-Aral.

Having analyzed the three most popular approaches to the determination of evaporability, as an index which is included in various indices of dryness (or moistening), we will now very briefly examine the question concerning advantages and disadvantages of the relationship of evaporability and precipitation as a characteristic of moistening conditions, irrespective of the use of one or the other method for calculating evaporability. In available investigations there usually are two main shortcomings of the Dokuchaev-Vysotskii index.

1) This index only includes annual characteristics, and does not account for annual variations of precipitation.
2) This index only takes into account the possible expenditure of moisture for evaporation, and does not account for losses of moisture by runoff (for instance, the review of V.P. Volobuev, 1953 [70] and others).

The first reason must be undoubtedly approved. As available data show, dynamics of soil moisture change abruptly under different conditions of annual variations, even when the total annual amount of precipitation is equal. This, of course, is of great significance for many physicogeographical processes. Therefore, for more detailed analyses of climatic conditions of moistening, data on soil moisture are of great importance. To enable a more extensive use of these data, the author suggested an indirect method for calculations based on the water balance equation (1950b [47]). A similar idea was presented by A.G. Isachenko (1953 [115]), who also came to the conclusion that a direct estimation of soil moisture dynamics is quite necessary in a study concerning climatic conditions of moistening.

At the same time, in many applications, a schematic characteristic of moistening conditions is, apparently, quite sufficient, and the Dokuchaev-Vysotskii index may conveniently serve as such. Thus, for the purpose of clarifying the general moistening conditions on the vast surface of all continents, application of the Dokuchaev-Vysotskii index could be justified by the necessity of considerable schematization of the source material in order to generalize it on a small scale map.

The second critical remark, concerning the index of Dokuchaev-Vysotskii, about not taking the expenditure of moisture for runoff into account, when applying this index, is in our opinion not justified. Much earlier, in investigations by E.M. Ol'dekop, it was established that the coefficient of runoff (the relationship between the norms of runoff and precipitation) for river basins is often determined by the relationsip between precipitation and evaporability. This conclusion was later confirmed in many other investigations, including those by the author (see §10). Thus, for the given relationship between evaporability and precipitation, the relationship between loss of moisture by runoff and precipitation, as well as the relationship between loss of moisture by evaporation and precipitation (in more or less larger regions) is not an independent value and can be easily calculated by the "relation equation."

As previously mentioned, the ratio of precipitation to evaporability or the inverse value is connected by certain functional relations with all of the other water and heat balance components of the land surface. This is the main theoretical basis for an extensive use of the Dokuchaev-Vysotskii index in the analysis of moisture conditions. However, it must be pointed out that a purely empirical investigation of relations between characteristics of various physicogeographical processes on one hand and precipitation and evaporability on the other, is also possible, and it is not limited by the necessity of using the index of moistening.

Several investigations were carried out in this direction, among which the papers by V.P. Volobuev deserve special attention (a summary of results is found in the monograph published in 1953 [70]).

Further developing the idea suggested by V.V. Dokuchaev and B.B. Polynov (1915 [186]), V.P. Volobuev established quantitative correlations between soil cover characteristics and climatic factors - precipitation and air temperature.

The main shortcoming of these, in many respects very interesting investigations, is the application of the mean annual temperature as an index of the thermal regime. In climates of moderate and higher latitudes this index provides for only a very rough characteristic of evaporability.

Investigations in this direction are very essential, especially when they make use of the most perfect meteorological indices, which are provided by recent developments in climatology.

At the same time, research in this direction is still relatively little developed, when compared with moistening investigations using the index of Dokuchaev-Vysotskii. Therefore, in investigations of moisture conditions which determine geographical zonality on the continents, it is most expedient to use the relationship of precipitation and evaporability.

Determination of Evaporability

According to the above mentioned argumentations we will now examine the three main methods for determining evaporability:

1) method suggested by E.M. Ol'dekop and used by Meyer, N.N. Ivanov and others, in determining evaporability from saturation deficit,
2) method suggested by G.T. Selianinov, in determining evaporability from sums of temperature,
3) method of determining evaporability from radiation balance.

First of all, it is desirable to compare results of the application of these methods with each other, since no more or less detailed, comparisons have ever been made.

At the same time, for an evaluation of the correctness of these methods it is proper to compare them with the complex method of calculating evaporability, which simultaneously takes into account all of the principal factors that affect evaporability - radiation balance, temperature and humidity of the air.

The principles of the complex method for determining evaporability were suggested in a paper by the author (1951a [45]). We will outline this method, in a somewhat simplified form, in a manner that is most convenient for practical use.

The amount of possible evaporation from land surfaces, which are sufficiently moistened, apparently can be determined by methods similar to those used for determining evaporation from water surfaces. At the present time, as a result of numerous theoretical and experimental investigations, it is acknowledged that, evaporation from water surfaces or from moist surfaces is proportional to the saturation deficit calculated from the temperature of the evaporating surface 5) (see § 5). This relationship can be expressed in the following form:

$$E_0 = \rho D (q_s - q), \qquad (112)$$

where E_0 — is possible evaporation (evaporability), ρ — air density, D — coefficient of external diffusion, q_s — specific humidity of saturated air at the temperature of the evaporating surface, q — specific humidity of the air at standard level of measurements (2m).

In preceding investigations (Budyko, 1951a [45], Budyko, Drozdov et al 1952 [56]), the author introduced in equation (112) a complementary multiplier β, which characterized the relationship between the length of the period with superadiabatic gradients of temperature and the total length of the period averaged. Calculations showed that the independent account of the variations of parameter β does not affect the results of determining evaporation, into any considerable degree, but complicate the calculating procedure.

This is the reason why parameter β was dropped from formula (112), assuming that the coefficient of external diffusion D equals the product obtained from multiplying the earlier used coefficient by the ratio of the length of the period with superadiabatic temperature gradients to the length of the whole averaged period. It is obvious that the value of the coefficient of external diffusion must be smaller than that of the earlier used coefficient (according to latest data the mean value of D for the warmer period of the year is approximately 0.63 cm/sec., also see chapter II).

From formula (112) it can be concluded that, in order to calculate evaporation, we have to know the air humidity q, , the coefficient of external diffusion D (which is proportional to the coefficient of turbulent exchange and in turn depends on many factors), and temperature of the evaporating surface, which determines value q_s.

Usually the first two values can be easily determined from results of relevant observations and calculations. Determination of the third value - temperature of the evaporating surface - is a much more complicated matter.

5) It must be emphasized that this value, in most cases, differs considerably from the value of saturation deficit calculated from air temperature.

Generally, a direct measurement of temperature of the active land surface is very difficult, and in this case the problem is complicated further by the fact that we have to deal with temperature of a surface from which evaporation proceeds under sufficient moistening conditions, i.e., under conditions which can differ considerably from the actual circumstances in regions with dry climate.

In the above mentioned investigations, on determining the temperature of the evaporating surface, it is suggested that the equation of heat balance be used:

$$R = LE + P + A. \qquad (113)$$

The radiation balance value could be presented in the form of:

$$R = R_0 - 4s\sigma\theta^3 (\theta_w - \theta),$$

where R_0 — radiation balance for the moistened surface calculated by using effective outgoing radiation as determined from air temperature, s — coefficient depending on qualities of the radiating surface (usually about 0.9) σ — the Stefan - Boltzman constant θ — air temperature, θ_w — temperature of the active surface. The following formula can be written for turbulent heat exchange, by analogy with (112):

$$P = \rho c_p D (\theta_w - \theta),$$

where c_p — heat capacity of the air at constant pressure.

Developing these formulas, we obtain from (113):

$$R_0 - A = L\rho D (q_s - q) + (\rho c_p D + 4s\sigma\theta^3)(\theta_w - \theta). \qquad (114)$$

From equation (114), having data on air humidity q, and temperature θ, available and knowing the values of D and $R_0 - A$, it is possible to calculate the values of θ_w and q_s, which are part of the well-known relationship (the Magnus formula). Furthermore, such a calculation would permit one to compute evaporability E_0 by using equation (112).

Equations (112) and (114) allow us to investigate the dependence of evaporation on basic meteorological factors in a general form. In the mentioned papers, on the basis of such investigations, conclusions were derived about laws governing the effects upon evaporability of temperature and humidity of the air, turbulent exchange and radiation balance. In evaluating the above mentioned simplified methods for determining evaporability we must notice the following.

Calculations by formulas (112) and (114) are not very complicated, but when made for numerous cases, the numerical solutions of equation (114) are rather cumbersome and time consuming. Calculations could be simplified and facilitated by using suitable nomograms; however, in many cases, it would be more convenient to use simpler methods for determining evaporability, thus requiring smaller quantities of basic data.

Now, let us examine those three simplified methods for calculating evaporability by taking into account the theoretical interrelations outlined above, and available empirical data.

For a limited period of time, by using the radiation balance method, evaporability is determined by formula:

$$E_0 = \frac{R_0 - A}{L}.$$

Since, for the year the value of A is near zero, for these conditions (and for many other cases as well), it can be assumed that evaporability is equal to the relationship of radiation balance and latent heat of evaporation.

From formula (114) it can be concluded that, when determining evaporability from radiation balance we can expect to obtain accurate results only in case:

$$R_0 - A = L\rho D\,(q_s - q),$$

i.e., when value $(\rho c_p D + 4s\sigma\theta^3)(\theta_w - \theta)$ is much smaller than the amount of heat spent for possible evaporation $L\rho D\,(q_s - q)$.

Consequently, the absolute error in determining evaporability from radiation balance will be:

$$\delta E_0' = \frac{1}{L}(\rho c_p D + 4s\sigma\theta^3)(\theta_w - \theta),$$

and the relative error

$$\frac{\delta E_0'}{E_0} = \left(\frac{c_p}{L} + \frac{4s\sigma\theta^3}{L\rho D}\right)\frac{\theta_w - \theta}{q_s - q}. \qquad (115)$$

Equation (115) permits the calculation of relative errors in determining evaporability and also errors in calculating evaporation from water surfaces, if we assume its value as being equal to $\frac{R_0 - A}{L}$.

This equation will later be used in evaluating errors of the evaporation calculations. To demonstrate it here, we determine the error in calculations of evaporability from radiation balance by using the direct comparsion method of evaporability values calculated by formulas (112) and (114) for the year (by adding up the monthly values), with corresponding values of the relationship of radiation balance and latent heat of evaporation.

For this purpose the values of E_0 and radiation balance (that was divided by the latent heat value) were computed for 44 sites located in various climatic conditions, from the tundra zone to equatorial forests and tropical deserts on all continents (except in the Antarctic).

Comparison of these values is presented in fig. 53, which clearly shows that the calculation of evaporability from mean annual data of radiation balance gives values close to calculations of evaporability from formulas (112) and (114). The mean relative error in the first calculation as compared with that of the second is only 10%.

This conclusion means that the direct effect of such factors as temperature and humidity of the air on annual values of evaporability is much less essential in comparison with the determining effect of radiation balance. On the basis of data given in fig. 53, in the majority of cases it could be concluded that, for a sufficiently moistened land surface the turbulent flux of heat P is small in its absolute values, as compared with the main components of heat balance - radiation balance and expenditure of heat for evaporation (exceptions from this rule are observed, as was indicated, mainly for comparatively small plots under dry climate conditions).

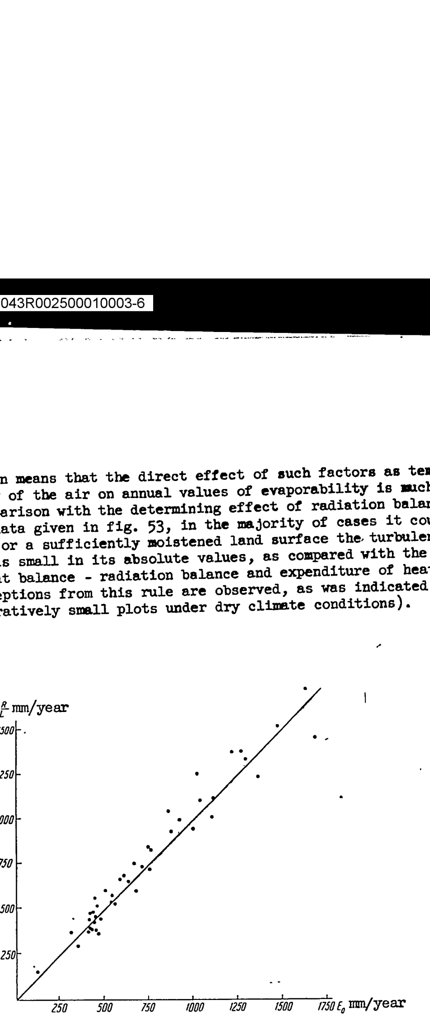

Figure 53

Comparison of evaporation values determined by complex method and radiation balance.

For clarification of the physical principles involved in this calculation method for determining evaporability from saturat[illegible] deficit we will first examine the relationship between saturation deficit and radiation balance for various climatic conditions. In this analysis we will be limited to the use of Russian data only, since there are no reliable data on saturation deficit (calculated with the Ol'dekop correction) [6] for the whole globe.

In fig. 54 mean annual values of saturation deficit, taken from the paper by E.M. Sokolova (1937 [212]), are compared with corresponding values of radiation balance. Data on saturation deficit and radiation balance have been taken from the maps for 187 sites, distributed regularly over the USSR except Eastern Siberia (where data are less reliable), mountain regions and coastal lands.

As can be seen from fig. 54, under conditions of a more or less humid climate, a close interrelationship exists between values of radiation balance and saturation deficit. It is close to a direct proportionality. According to the relationship presented in fig. 53, it can be assumed that, under analyzed conditions, saturation deficit is also connected by a direct proportionality with true evaporability.

Thus, for more or less humid regions, calculations of evaporability from radiation balance and saturation deficit will give similar results (when coefficient a in formula $E_0 = ad$ is properly selected).

Now let us compare the principal error in calculating evaporability from saturation deficit and radiation balance, in a more or less humid climate.

Possible evaporation from the humid surfaces (evaporability) is equal to:

$$E_0 = \rho D (q_s - q)$$

it could be presented as the sum of two components:

$$E_0 = \rho D (q_s' - q) + \rho D (q_s - q_s'),$$

where q_s' — is specific humidity of saturated air at the air temperature. Since the value of $(q_s' - q)$ corresponds to the saturation deficit, the last formula could be rewritten:

$$E_0 = E_0'' + \rho D (q_s - q_s'),$$

where E_0'' — is evaporability, determined by saturation deficit, and $\rho D (q_s - q_s') = \delta E_0''$ — a value, that characterizes the principal error in

6) E.M. Ol'dekop has shown that, when determining mean monthly values of saturation deficit, we cannot use mean monthly temperatures, since a large error arises in the calculations due to the nonlinear dependence of the pressure of saturated vapor on temperature, and because of the temperature variation during the month. In the paper by Ol'dekop (1911) a method of calculating saturation deficit that will set off the indicated error ("correction of Ol'dekop") was suggested. However, the necessity of using this method when determining saturation deficit has not always been taken into account in later investigations (especially so abroad).

calculations of evaporability from saturation deficit. This error arises because of a difference between air temperature and the temperature of the active surface.

The relative error in the calculation of evaporability from saturation deficit will be:

$$\frac{\delta E_0''}{E_0} = \frac{(q_s - q_s')}{q_s - q}. \tag{116}$$

If we compare the relative error in calculations from radiation balance and saturation deficit, by dividing the corresponding values from equations (115) and (116), we obtain:

$$\frac{\delta E_0'}{\delta E_0''} = \left(\frac{c_p}{L} + \frac{4s\sigma\theta^3}{L\rho D}\right) \frac{\theta_w - \theta}{q_s - q_s'}. \tag{117}$$

It is interesting to note that due to the approximate proportionality of values $(\theta_w - \theta)$ and $(q_s - q_s')$, within limits of their actual possible variation, ratio $\frac{\delta E_0'}{\delta E_0''}$ for each air temperature has an approximately constant value. So, for instance, we find from this formula that with an air temperature of 25°C the value of $\frac{\delta E_0'}{\delta E_0''}$ is approximately equal to 1/2. This means that the principal error in calculating evaporation from radiation balance, with a more or less high air temperature, is smaller than the principal error of evaporability calculations from saturation deficit. This conclusion must be accounted for in calculations of evaporation, however, it does not exclude the possibility of an approximate determination of evaporability from saturation deficit in relatively moist climates.

Much more difficult is the utilization of saturation deficit for calculating evaporability in dry climate, especially in deserts.

As can be seen from fig. 54, for dry climates the direct proportionality between radiation balance and saturation deficit is broken and the scattering of the dots on the graph is larger. Comparing figs. 53 and 54 we can see that, in this case, the direct proportionality is also discontinued between saturation deficit and true evaporability.

There is no doubt that the rapid growth in saturation deficit in desert climates is not accompanied by the same rate of increase of the possible growth of evaporation (evaporability). This statement is based on elementary reasoning.

Should we compare mean saturation deficits in a desert and in an abundantly irrigated oasis, these values would differ very much. At the same time, in both cases values of evaporability (of the possible evaporation) obviously are equal because the factual value of evaporation in an abundantly irrigated oasis is the same value of evaporability for a given desert region. Thus, saturation deficit in factual conditions of dry climate can-

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not be assumed as being proportional to the value of possible evaporation. 7)

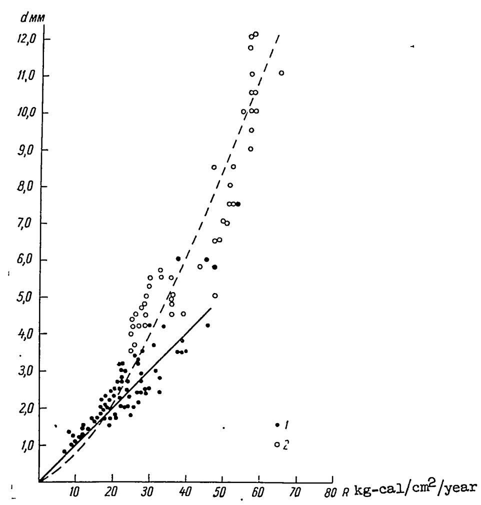

Figure 54

Comparison of mean annual values of saturation deficit of the air with radiation balance values. 1-tundra, forest, forest-steppe, steppe; 2- semi-desert, desert.

7) This conclusion could be expressed in a more comprehensive way. Those indices of evaporation which greatly depend on the degree of moistening of a certain locality, should be used for determining evaporability in dry climates by taking the abundant irrigation into account. Accordingly, as has been emphasized in previous investigations of the author and also pointed out above, in determining evaporability from radiation balance in dry climates data of the balance calculated for a moist surface should be used. It should be pointed out that, such calculations are not more complicated than calculations of radiation balance for the actual state of the active surface in dry climates.

Actually the majority of meteorological stations in deserts, whose data were used for construction of the saturation deficit map that we are now using, were under the influence of the proximity of irrigated plots. Therefore, observations of these stations present saturation deficit values that only partly characterize the climate of deserts or semideserts. Nevertheless, as fig. 54 and data of special observations show, values of saturation deficit derived from observations of desert stations turned out to be considerably "exaggerated" in comparison with those considered typical for abundant irrigation conditions of these regions.

It is very interesting to note that the question of the necessity of taking into account the effect of irrigation on desert climate, when determining possible evaporation, was examined much earlier by A.I. Voeikov. So, for instance, in The Climates of the Terrestrial Globe A.I. Voeikov (1884 /68/) analyzed an interesting case of a large error contained in calculations of French engineers who designed the water reservoir of the Suez Canal. According to data obtained by small evaporimeters, these engineers assumed that evaporation from the water reservoir would be about 7m per annum. The actual evaporation proved to be a little more than 1m per annum.

This difference is explained by the fact that larger evaporating surfaces in deserts evaporate a much lesser amount than the smaller ones, because of the effect that evaporation exerts on the climate of the underlying air layer.

There is also a second reason for the slower growth of evaporation as compared to increasing saturation deficit in dry climates. It has already been noted in certain publications that the mean water temperature of limited water reservoirs in dry regions is usually remarkably lower than the air temperature. This results in smaller values of saturation deficit computed from the temperature of the evaporating surface, as compared with the actual deficit. This effects a relative diminution of evaporation, which is also noticed on small evaporating surfaces.

A more rapid increase in saturation deficit of the air in dry climates, as compared with the growth of evaporability, complies with empirically established relations between evaporation from water reservoirs and saturation deficit.

As has already been mentioned in this section, recent investigations established that evaporation is proportional to saturation deficit in a power less than one. The same relationship holds true in comparing saturation deficit with radiation balance. This can be seen, for instance, from data in fig. 54. The relationship between radiation balance and saturation deficit, that is shown by this graph, could approximately be described by the formula $R_0 = bd^{0.7}$, where $b = 11.3$ (using the same dimensions in which this graph was compiled).

This conclusion explains the above mentioned results of the latest investigation by Prescott (1949), who also came to the conclusion that it is necessary to take into account the saturation deficit in power 0.7, in indices characterizing the degree of moistening.

Thus, it can be concluded that, it is feasible to use saturation deficit for calculating evaporation and for computing indices of moistening, mainly in more or less moist climates. In this case, calculations of evaporability from saturation deficit give, on the average, results that are close to those obtained from radiation balance calculations, although the principal error of the calculations from saturation deficit will be somewhat greater than the error in calculations from radiation balance.

But the assumption that there is a proportional relationship between the evaporation and saturation deficit, in dry climates, is entirely wrong. In this case it would be more correct to consider the amount of evaporation as being proportional to saturation deficit in a power less than 1, even though this kind of calculation will also include a noticeable possible error which results from the above outlined physical reasons.

Let us now examine the problem of using sums of temperatures in evaporability calculations. As is well known, sums of temperatures present an important agrometeorological index, which at the time being is used largely in agricultural climatology. However, the physical interpretation of this index has never been very clear. Because of this, it would be of considerable interest to compare the sums of temperatures with the, physically substantiated, energy index of evaporability; i.e., the radiation balance.

For this purpose relevant calculations were carried out. The results are presented in fig. 55. The abscissa shows radiation balance values R, the ordinate shows $\Sigma\theta$ — the sums of temperature higher than 10°C [8] taken from the World's Agroclimatic Handbook (1937) for 300 stations distributed over various climatic regions on all continents (except Antarctica) from 71°N to 46°S. All data in this handbook were used, and only coastal, island, and mountain stations were omitted.

Fig. 55 shows a very close correlation between the sums of temperatures and radiation balance, which exists, on the average, in all latitudes and continents. It is interesting to note that this relationship has the form of direct proportionality, where in the first approximation, sums of temperatures divided by 100 are about equal to the quantities of radiation balance in kg-cal/cm^2/year.

A distinct dependence of the sums of temperatures on radiation energy factors interprets, to a certain extent, the physical meaning of this characteristic, which up to recent times was sometimes considered as a very conditional index. At the same time, the established correlation between sums of temperatures and radiation balance shows considerable possibilities in using sums of temperatures as an indirect estimate of evaporability under various climatic conditions.

It should be noted, however, that because of the definite scattering pattern shown in diagram 55, the correlation between sums of temperatures and radiation balance cannot be regarded as being quite uniform. Analysis of source data shows that in individual regions a definite discrepancy between the radiation regime and sums of temperatures is found. So, for instance, in some coastal regions sums of temperatures turned out to be somewhat lower when compared with radiation balance. Still more incon-

8) The sums of temperatures higher than 10° is obtained by adding up mean daily temperatures in the Centigrade scale for the period when these temperatures are higher than 10°.

gruous are these factors in mountain regions, where sums of temperatures usually decrease with altitude at a much higher rate than radiation balance.

In such cases we encounter certain conditions which increase the mean differences between the air and active surface temperatures, which may result in definite errors in estimating thermal factors of physicogeographical processes on the surface when using sums of air temperatures. This interesting problem requires special treatment.

A great advantage of sums of temperatures, as indices of the evaporability value, is seen in their stability, which shows only slight variation with the changing dryness of climate. For instance, they vary considerably less than saturation deficit. This fact also justifies the utilization of sums of temperatures for determining evaporation in dry climates, deserts included.

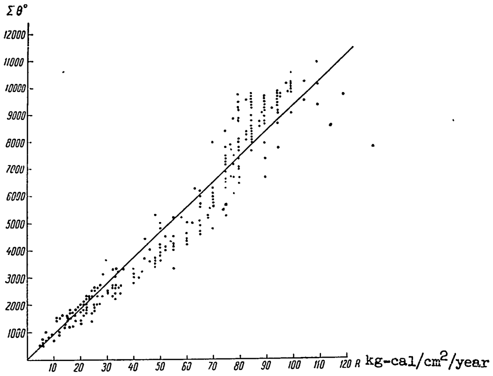

Figure 55

Comparison of sums of temperatures with values of radiation balance.

From fig. 55 it can be concluded that annual evaporability could approximately be determined by formula:

$$E_0 = c\Sigma\theta,$$

where E_0 — is evaporability in mm/year, $\Sigma\theta$ — sums of temperatures higher than 10°C, c — coefficient equal to 0.18. The annual value of the Dokuchaev-Vysotskiĭ index, in accordance with the idea of G.T. Selianinov,

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can be computed by formula:

$$\frac{r}{0{,}18\Sigma\theta},$$

where r — is the annual amount of precipitation in mm.

Summarizing this analysis of methods for determining evaporability as used in calculating the index of Dokuchaev-Vysotskii, we can note the following.

The complex method for determining evaporability, that is described above, is the one which, theoretically, has the most substantiation. It takes into account the effect of radiation balance, air temperature and humidity. However, utilization of this method involves some cumbersome computations, and that is its shortcoming.

In calculations of annual amounts of evaporability the radiation balance method can be largely applied. This method also has a definite theoretical basis, and under very diversified climatic conditions gives results close to those obtained by the complex method.

Among the empirical methods for determining evaporability, attention should be paid to calculations of annual amounts of evaporability from sums of temperatures. These calculations are very simple, and under a large range of climatic conditions - from tundra to the equatorial zone - they provide results similar to those obtained by calculations of the radiation balance method.

Calculations of evaporability from saturation deficit can give satisfactory results, mainly in comparatively humid climates. In a dry climate (especially in deserts) these calculations are less suitable and might include considerable errors.

On the basis of these deductions we apply the amounts of evaporability calculated by the radiation balance method, in our study of moistening conditions on the continents.

Let us now consider the question concerning the period for which data on radiation balance should be taken in determining evaporability.

In one of the preceding works by the author (1949b /42/) it was indicated that in different climates of moderate latitudes, values of radiation balance for the year and growing season turned out to be very close to each other. This, at first glance a very paradoxical fact, is interpreted in the following way. During the cold period of the year, radiation balance in moderate latitudes is negative and its absolute values are small. In spring, radiation balance becomes positive, and at the beginning of the growing season (which is usually accompanied by the increase of air temperature to 10°C), in most cases, it reaches comparatively large positive values. Thus, negative and positive values of radiation balance during the "nongrowing season" compensate each other to a certain extent.

Available data on radiation balance provide a basis for a more detailed analysis of the problem on the relationship of radiation balance values for the year and growing season. The results of such a comparison are shown in fig. 56, where the abcissa represents radiation balance for the year R,

and the ordinate R' — radiation balance for the period with temperatures higher than 10°C. These values were computed for moist surface conditions (the comparison of balance values for the actual state of the surface would give, by and large, the same results).

Figure 56

Comparison of sums of radiation balance values for the year and growing season.

Data in fig. 56 pertain to stations located between the latitudes of 40° and 70°N. Data of coastal and island stations were not used in constructing this graph.

As can be seen from graph 56, larger values of radiation balance R and R' which are observed in lower latitudes are always close to each other. In higher latitudes, where these values are smaller, we observe some differences between R and R' values, with the latter value considerably larger than the former. It should be pointed out that the largest discrepancies between these values are principally found in sharp continental climates (the central regions of Eurasia).

In this region a rapid increase of temperature in the spring produces an incomplete compensation between negative and positive values of radiation balance in the "nongrowing period" and therefore the radiation balance for the growing period becomes larger than that for the year.

But, that does not mean that it is necessary to use, in calculating annual amounts of evaporability in a sharp continental climate, radiation balance values for the growing season. As has been repeatedly mentioned by the author (1951a [45] and others), values of radiation balance for the warm season in moderate latitudes cannot be used entirely for evaporation (when there is sufficient moistening of the region). If annual variations of temperature are quite pronounced, a definite part of the heat in radiation balance is spent, during the warm period, for heating (and thawing) of soil. This amount of heat is again released during the cold season when the soil is cooling off (and freezing). This portion of heat grows larger in sharply expressed continental climates, and increases still more in regions with deep soil freezing.

According to available data of heat exchange in soil we may conclude that, in continental climates of moderate latitudes, the amount of heat which can be spent for evaporation during the warm season is markedly smaller than the radiation balance for the growing season, and is closer to the annual value of the balance.

These deductions confirm that, for determining annual evaporation in various climates, it is feasible to use values of radiation balance which also pertain to annual values.

Some regularities in geographical zonality

For the study of correlations between geographical zonality and climatic conditions, the author (1955a [48]) designed a world map of the radiational index of dryness.

In constructing this map a special world map of radiation balance was used; the latter map referred to a moistened surface condition (values of radiation balance were calculated by methods described in § 3, the value of the albedo of the moistened surface was assumed to be 0.18 for the period without a snow cover).

The isolines, on this map of radiation balance for the moist surface, generally follow a latitudinal pattern. In respect to mean latitudinal values, amounts of this radiation balance are somewhat smaller in strong continental regions of moderate latitudes, some monsoon regions and also some areas of the middle latitudes of North America. In desert regions the value of radiation balance is, as a rule, somewhat higher than in surrounding regions of a more humid climate, however, this difference is relatively small.

Having data on radiation balance available and using results of the Central Geophysical Observatory on precipitation amounts, it was possible to calculate the index $\frac{R}{Lr}$. These calculations were accomplished for 1600 sites, distributed more or less evenly over the continents. This number of observation points is sufficient for constructing a rather detailed world map, however, it must be remembered that there, inevitably, will still be a certain degree of schematization in the distribution of the dryness index. A sharp spatial changeability of precipitation in some regions (especially in the piedmonts and mountains) results in corresponding changes of the dryness index, which is not always reflected by a map of limited dimensions, even when a dense network is available. Therefore, we were compelled to show, on the map, only schematized isolines in those regions where the index isolines went too close to each other and outlined some relatively small areas.

The map of radiational index of dryness (fig. 57) shows the values of $\frac{R}{Lr}$, equal to 1/3, 2/3, 1, 2 and 3. These steps were chosen by taking into account results of preceding investigations by the author concerning the analysis of the correlation existing between values of the dryness index and boundaries of the natural zones, and also considering the conclusion by A.A. Grigor'ev (1954 [84]), who indicated that between the dryness index on the boundaries of mean natural zones there are certain multiple, numerical relationships. Isolines of the dryness index in fig. 57 are drawn over the whole surface of the continents except high mountain regions and extreme northern sections of North America, which are shaded. This has been done because of the fact that no reliable data are available neither for radiation or precipitation in these regions. Thus, in particular, data on precipitation in North America undoubtedly contain considerable errors due to insufficiently accurate measurements of solid forms of precipitation. The correction of these errors is rather difficult.

As can be seen from fig. 57, the distribution of radiation index of dryness isolines on continents shows great variations in the index values of each continent. The most humid conditions, with the smallest value of the dryness index, are seen mainly in higher latitudes where evaporability is very low. The greatest values of this index are found in deserts and semideserts.

Comparing the radiation index of dryness map (fig. 57) with geobotanical and soil type maps, we can see that the distribution of the dryness index isolines coincides fairly well with the distribution of the mean physico-geographical zones. The lowest values of the dryness index (up to 1/3) are typical of the tundra zone. The large values of this index (from 1/3 to 1) are found in the forest zone. Still larger indices (from 1 to 2) are characteristic of steppe, and index larger than 2 is typical of semideserts, larger than 3 deserts.

We will not discuss many of the other deductions which could be drawn from this comparative analysis of the maps of dryness index and those of geobotanical and soil features, but the main conclusion must be emphasized; the parameter $\frac{R}{Lr}$, which determines the relative values of the components of heat and water balance also determines, to a certain extent, locations of the boundaries of the principal natural zones. It is also obvious that in different latitudes, within similar zones, essential differences in the development of natural processes are taking place. These differences, as available data show, can be traced to the fact that on different latitudes, the fundamental energy of the natural processes, which can be designated by the radiation balance values R, is essentially different.

Thus, we need only one parameter $\frac{R}{Lr}$ (that gives us relative values

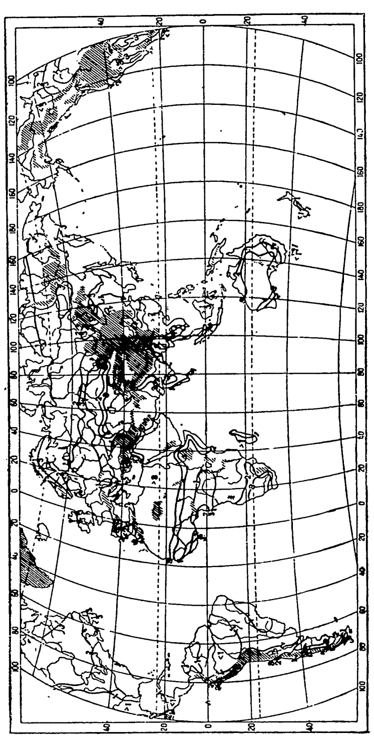

Figure 57. Radiational index of dryness.

of the heat and the water balance terms) for determining general zonal conditions of the dynamics of natural processes. But for characterizing absolute values of the intensity of natural processes, two parameters would be needed: $\frac{R}{Lr}$ and R, which determine the absolute values of heat and water balance components.

The relationship between geobotanical features and parameters $\frac{R}{Lr}$ and R is presented by a graph with ordinates R and $\frac{R}{Lr}$ which also show the boundaries of the principal geobotanical zones.

A schematic form of such a graph is presented in fig. 58. The solid line shows limits of actually observed values of R and $\frac{R}{Lr}$ (except in mountain regions). Within these limits, definite values of parameter $\frac{R}{Lr}$, shown by vertical dashed lines, separate the principal geobotanical zones: tundra, forest, steppe, semidesert and desert. The great variability of radiation balance values in the forest zone, and the somewhat lesser range of these values in the steppe, reflect the variations of geobotanical aspects inside these zones.

Insofar as the zonality of soils is closely associated with that of vegetation, results obtained for the relationship of vegetation zones to certain values of parameters R and $\frac{R}{Lr}$ could be referred to soil zones as well.

Consequently, it can be established that with an increase of parameter $\frac{R}{Lr}$ the types of soil will change according to the following sequence: a) tundra soils, b) podzol soils, brown forest soils, yellow soils, red soils and laterities (the wide variability of soil types in this group corresponds to the wide range of parameter R), c) chernozem and black soils of the savanna, d) chestnut soils e) gray soils. The relationship of soil zonality with climatic indices $\frac{R}{Lr}$ and R can be presented in a general way, as a graph similar to that of the vegetation zones, which was shown in fig. 58.

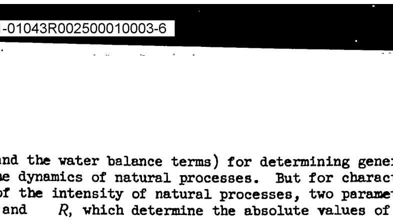

Figure 58

The graph of geobotanical zonality.

The zonality of the hydrological regime on land, with parameters R and $\frac{R}{Lr}$ can be established not only qualitatively, but also quantitatively. From the "relation equation" it can be concluded that, each class of values of parameter $\frac{R}{Lr}$ corresponds to a certain class of the runoff coefficient values. Therefore for tundra, where $\frac{R}{Lr} < 1/3$, the runoff coefficient must be greater than 0.7; in the forest zone with $1/3 < \frac{R}{Lr} < 1$ the runoff coefficient must be 0.3 to 0.7; in the steppe zone (where $1 < \frac{R}{Lr} < 2$) from 0.1 to 0.3; in semideserts and deserts it is less than 0.1. The observed data have confirmed these relationships very well.

Thus, the evaluation of the effect of the energy factors permits a quantitative interpretation of zonal changes in the runoff coefficient.

Absolute values of the sums of runoff are determined by two parameters: R and $\frac{R}{Lr}$. In connection with this, we can present the distribution of annual sums of runoff in cm/year in a form similar to the graph in fig. 58 (see fig. 59). Fig. 59 characterizes the absolute values of runoff in various geographical zones.

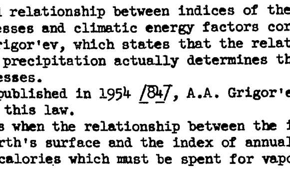

Figure 59

Runoff in various geographical zones.

The established relationship between indices of the dynamics of physico-geographical processes and climatic energy factors corroborate the law discovered by A.A. Grigor'ev, which states that the relationship between radiation balance and precipitation actually determines the nature of physico-geographical processes.

In his paper, published in 1954 /84/, A.A. Grigor'ev gave the following interpretation of this law.

"In those cases when the relationship between the index of radiation balance of the earth's surface and the index of annual precipitation (given as the amount of calories which must be spent for vaporization of this

amount of precipitation) is about 1, then a quantitative proportionality,(a commensurateness) exists between heat and moisture amounts which participate in the natural processes of the geographical medium. The essences of this commensurateness is seen in the fact that the value of radiation balance is equivalent to the amount of heat energy required for vaporization of the annual amount of precipitation, and there is only so much precipitation as can be vaporized from the surface of the earth under these thermal conditions. In the structure and dynamics of the geographical medium this commensurateness is expressed by the fact that as a result of heat and moisture interaction a certain thermal regime is created, with unobstructed and continuous transpiration and evaporation (from the ground surface) on one hand, and processes of the ground aeration on the other, i.e., a combination of conditions of great positive importance in developing the biocomponents of the geographical medium.

"In those cases, however, when the relationship of heat and moisture deviates from value 1 (in any direction), a disparity between heat and moisture amounts takes place; precipitation is considerably larger or considerably smaller than the amount which can be vaporized under given thermal conditions. As a result, under conditions of this thermal regime type, the unhampered and continuous flow of transpiration and evaporation processes is disrupted by moisture deficiency,or processes of soil aeration are disrupted by excess moisture (relative to existing thermal conditions). This exerts a great effect, usually a negative one, on the development of biological components in the geographical medium, and the disproportion increases,the more the deviation of the given relationship varies from 1.

"From all this it can also be concluded that commensurateness of heat and moisture must exert an essential effect not only on the nature of biological components of the geographical medium, but also on the character of hydrologic processes, and through them, on geomorphological processes as well."

According to these statements by A.A. Grigor'ev there is reason to assume that between climatic energy factors and the intensity of all exterior natural processes certain quantitative relations are operating, and these relations are similar to hydrological regularities analyzed above. The investigation of the relationship between climatic energy factors and balances of organic and mineral substances in soil, the productiveness of the biological mass and other similar factors is hampered by the complex and insufficient knowledge of the quantitative characteristics of the relevant processes. Nevertheless, application of data on heat energy balance permits us to cast more light on some laws governing the creation of organic matter in various natural zones. The following section is concerned with this problem.

§12. Heat balance and conditions of plant development

Organic matter in nature is created as a result of the activity of antotrophic plants which are the unique group of organisms that is capable of synthesising organic matter from minerals. From two ways of producing organic matter by utilization of radiation energy (photosynthesis) and chemical energy (chemosynthesis), only the first one is essential in producing the biological mass. Chemosynthesis, being very important for nitrogen transformation and for some other processes, produces, however, only an insignificant quantity of organic matter. Photosynthesising plants utilize carbon dioxide of the air, 9) water, and a small amount of mineral substances in soil, for building organic matter. For the photosynthesis process in plants, some portion of energy, received as short-wave solar radiation, is usually expended.

Numerous experimental investigations established that the "efficiency coefficient" of the photosynthetically active plants is very small, usually it equals 0.1-1.0% (on the average it is 0.5%). This "efficiency coefficient" shows the ratio of the amount of energy spent for synthesis of biological mass to the total amount of incoming solar energy. Under most favorable conditions the value of this coefficient increases up to 5% and only rarely exceeds this amount (Vinberg, 1948 and others).

It should be noted, however, that extensive experimental data proved that vegetation also spends water resources in a very uneconomical way, i.e., the productiveness of transpiration (the ratio of weight accretion of dry mass in the plant to the discharge of water for transpiration during a certain time interval) is usually from 1/200 to 1/1000 (mostly about 1/300). It has also been discovered that such abundant transpiration is not produced by the physiological requirements of plants but, to a considerable extent, it is just a useless loss of water (Maksimov, 1926, 1944 /165 & 166/ and others).

These two fundamental facts provided evidence that in natural conditions vegetation consumes only an insignificant portion of the available energy and water resources. It is obvious that for investigating the correlation between productiveness of vegetation and climatic factors, it is important to clarify the causes which so substantially limit the utilization of natural resources for synthesis of biological masses.

In investigations of Blackman, Lundegord and some other authors (Liubimenko, 1935 /163/ and others), it was assumed that the photosynthesis energy is mainly limited by the low concentration of carbon dioxide in the air.

To verify this hypothesis concerning the limiting effect of carbon dioxide concentration on the process of photosynthesis in natural vegetation, we have to analyze the physical mechanism of assimilation and transpiration.

The organ for assimilation in a photosynthesising plant - the leaf - represents an envelope of dense cuticular tissue with many small openings - stomata - which can open and close themselves. Within this envelope a great surface of layers containing grains of chlorophyll is enclosed. The surface of these layers contacts atmospheric air through intracells and stomata.

It is very essential that, for development of photosynthesis, the surface of the chlorophyll layers be kept moist, since carbon dioxide can be

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9) Along with carbon dioxide of the air the plants can also utilize carbon dioxide of the soil, however, the first source usually provides the main portion of it.

assimilated only when in a solution. Consequently, the relative humidity of the air in the intracells is very high and according to some experimental data it reaches 98-99%.

Since the relative humidity of atmospheric air during daylight hours is usually considerably below 100%, diffusion of carbon dioxide into the leaf with the open stomata is inevitably associated with diffusion of water vapor in an opposite direction; i.e., with transpiration of the plant.

If the vegetation cover is sufficiently dense, its surface practically coincides with the "active surface," for which an equation of heat balance can be derived by the following formula:

$$R = LE + lA + P, \qquad (118)$$

where R - is radiation balance, E - evaporation (transpiration), L - latent heat of vaporization, A - assimilation, l - expenditure of heat for assimilation of a CO_2 unit of weight, P - turbulent heat exchange between the active surface and atmosphere.

In equation (118) the heat exchange between the active surface and deeper soil layers is neglected. This negligence, in the given case, is quite allowable, insofar as in the warm season of the year heat exchange in soil is considerably smaller than the principal terms of heat balance, and during this season it is possible to select such periods for calculating heat balance, for which heat exchange in soil equals exactly zero.

Now, let us point out the following significant fact. Three terms that constitute the right side of equation (118) - evaporation, assimilation, and turbulent heat exchange - depend on diffusion processes of analogous mechanisms.

In plant transpiration water vapor diffuses from the moist walls of the parenchyma to the outer surface of the leaf through the intercellular space and stomata, and its concentration changes from q_s (average concentration in the air on the surface of the parenchyma) to q_0 (average concentration on the outer surface of the leaf).

The speed of evaporation will be:

$$E = \rho D' (q_s - q_0), \qquad (119)$$

where D' - is the effective coefficient of diffusion for water vapor on its way from the surface of the parenchyma to the outer surface of the leaf. Obviously, value D' will depend on the morphology of the leaf (the number and size of stomata, the thickness of the leaf, density of cuticular tissue, etc.).

The second stage of water vapor diffusion is associated with the change of water vapor concentration from q_0 (on the leaf surface) to q - concentration in a free air stream. For this stage the following equation can be derived:

$$E = \rho D'' (q_0 - q), \qquad (120)$$

where D'' - is the coefficient of external diffusion which depends mainly on the intensity of turbulent exchange.

Eliminating q_0 from equations (119) and (120) we obtain:

$$E = \frac{\rho (q_s - q)}{\frac{1}{D'} + \frac{1}{D''}}. \qquad (121)$$

The diffusion of gaseous carbon dioxide, from the free air flux to the absorbing surface of the parenchyma cells inside the leaf, can also be divided into two stages.

The speed of diffusion (which is equal to the speed of assimilation) will be equal, on one hand, to:

$$A = \rho D_c'(c_0 - c_1), \qquad (122)$$

where D_c' - is the effective coefficient of carbon dioxide diffusion through the intercells and stomata, from the outer surface of the leaf, to the surface of the parenchyma. c_0 - the average concentration of carbon dioxide in the air on the outer surface of the leaf, c_1 - the average concentration of carbon dioxide in the air on the surface of the parenchyma cells.

On the other hand, the speed of assimilation is equal to:

$$A = \rho D_c''(c - c_0), \qquad (123)$$

where c -is the concentration of carbon dioxide in the free air, $\overline{D}_c''$ - a value analogous to $\overline{D''}$.

From (122) and (123) can be concluded that:

$$A = \frac{\rho(c - c_1)}{\frac{1}{D_c'} + \frac{1}{D_c''}} \qquad (124)$$

The relationship between the speed of assimilation and the speed of evaporation according to (121) and (124) will be:

$$\frac{A}{E} = a\frac{c - c_1}{q_s - q}, \qquad (125)$$

here, it is important to point out that the coefficient of proportionality

$$a = \frac{D_c' D_c''}{D' D''} \frac{(D' + D'')}{(D_c' + D_c'')}$$

may change within comparatively narrow limits only. Since the process of water vapor and carbon dioxide diffusion between the outer surface of the leaf and the free air flux is determined by turbulent mixing, it is obvious that coefficients D'' and D_c'' have approximately equal values. It is only natural to suppose that water vapor and carbon dioxide diffusion from the walls of the parenchyma to the outer surface of the leaf has a molecular character and, in this case, the ratio $\frac{D_c'}{D'}$ must be equal to the ratio of the molecular diffusion coefficient of carbon dioxide in the air to the molecular diffusion coefficient of water vapor in the air, i.e., approximately 0.64.

Considering this, it may be concluded that, depending on the relationship of values D' and D'', coefficient a can only change from 0.64 to 1.00, approaching value 1 when the outer diffusion is of greater significance (a thin leaf with numerous open stomata), and decreases to 0.64 when internal diffusion plays the leading role (a thick leaf with poor ventilation). Should we accept the well-known conclusions that under average

conditions the rate of evaporation from the leaf is about half as great as the rate of evaporation from the leveled moist surface, it would be easy to determine that coefficient a is approximately 0.8.

Thus, considering that the rate of evaporation from the moist surface is

$$E_f = \rho D''(q_s - q),$$

and when:

$$E_f = 2E, \quad \frac{2}{D''} = \frac{1}{D'} + \frac{1}{D''}, \quad D' = D''$$

we obtain:

$$a = 0{,}78.$$

Noting that the turbulent exchange between the active surface and the atmosphere will be equal to:

$$P = \rho D'' c_p (\theta_w - \theta), \qquad (126)$$

where c_p - is the heat capacity of the air, θ - air temperature, θ_w - temperature of the active surface, i.e., of the leaves, we derive from (118), (121), (124), (126) the formula for the speed of assimilation:

$$A = \frac{R(c - c_1)}{\frac{L}{a}(q_s - q) + l(c - c_1) + b(\theta_w - \theta)}, \qquad (127)$$

where:

$$b = \frac{c_p}{a}\left(1 + \frac{D''}{D'}\right).$$

From formula (127) it can be concluded that the rate of assimilation depends substantially upon the difference in carbon dioxide concentration between the free atmosphere and the parenchyma's surface $(c - c_1)$, becoming larger with the increase of this difference. Insofar as the concentration of carbon dioxide in the free air flux is more or less constant, it is obvious that the rate of assimilation will increase with a decreasing value of c_1, i.e., with stronger absorption of CO_2 by protoplasm.

In this case, when "physiological absorption" occurs at a considerable speed and the general rate of assimilation is limited not by physiological processes, but by the diffusive supply of carbon dioxide, value $\overline{c_1}$ must be much smaller than c $(c_1 \ll c)$, and the assimilation formula will be transformed into:

$$A = \frac{Rc}{\frac{L}{a}(q_s - q) + lc + b(\theta_w - \theta)}. \qquad (128)$$

By using formula (128) we can theoretically determine how much of the solar energy could be utilized by vegetation under most effective utilization conditions of carbon dioxide from the air.

Considering that c is, on the average, equal to $0{,}46 \cdot 10^{-3}$ gr for 1 gr of air, L = 600 cal/gr, a = 0.78, l = 2500 cal/gr, b = 0.62 cal/gr/degree, we find that:

$$lA = \frac{1{,}2R}{770\,(q_s - q) + 1{,}2 + 0{,}62\,(\theta_w - \theta)}. \qquad (129)$$

In calculating energy spent for assimilation by formula (129) it should be remembered that the total expenditure of heat for evaporation and heat exchange in the warm season is determined by quantities observed during daylight only, since evaporation and heat exchange at night are inconsiderable, due to the "ventil effect." Therefore, values $(q_s - q)$ and $(\theta_w - \theta)$ in formula (129) must be taken as averages for daylight hours.

There are reasons to assume that under optimal conditions for photosynthesis, when $c_1 \ll c$, the parenchyma's surface is sufficiently moist, and consequently q_s approaches the value of the concentration of saturated water vapor at the leaf temperature (or, more correctly, the difference between q and the concentration of saturated water vapor must be much smaller than vapor pressure deficit in summer during daylight hours, computed at leaf temperature).

Mean daylight hour differences in temperature between the leaves and air in summer, at temperate latitudes, are on the order of 5°C, fluctuating in some cases to rather wide limits. Assuming that the mean relative humidity, during daylight hours in summer, is about 50% and air temperature about 20°C, we find that the values of $q_s = 2.0 \cdot 10^{-2}$ (at a leaf temperature of 25°) and $q = 0.7 \cdot 10^{-2}$. Using these values in formula (129) we determine that $lA = 0.08R$, which means that expenditure of energy for assimilation under average climatic conditions in temperate latitudes may reach 8% of the radiation balance.

From available data it could be found that, for the major portion of the European USSR, the radiation balance for summer is 55-60% of the incoming total radiation for this season. By using this estimate we can determine that under average summer conditions of temperate latitudes, the natural vegetation, absorbing carbon dioxide in the most effective way, may utilize approximately 5% of the incoming solar radiation. This evaluation, obtained theoretically, agrees very well with the average empirical data on the "efficiency factor" of photosynthesis under favorable conditions.

However, actual available data shows that such comparatively high coefficients of solar energy utilization are observed only in individual cases, whereas the average relationship of energy spent for photosynthesis of natural vegetation and incoming solar energy is usually on the order of 0.5%.

On these grounds we may conclude that, under average conditions the relationship $c_1 \ll c$ is not fulfilled and in the majority of cases the difference does not exceed 10% of value c.

In other words, the comparison of the analyzed theoretical scheme of diffusion with actual data permits us to establish the fact that natural vegetation utilizes, as a rule, only a small portion (approximately 10%) of the possible diffusion flux of CO_2 and therefore, the content of carbon dioxide in the air does not, as a rule, limit the production of biomass.

It may be noted here that this conclusion complies very well with results of many experiments made by V.N. Liubimenko and other authors, who established, on the basis of physiological investigations, that the carbon dioxide content in the air does not limit the energy of photosynthesis and the leaves do not utilize the total possible diffusion of the gaseous carbon dioxide. From these observations Liubimenko arrived at the conclusion that, "in natural conditions, the production of dry matter is limited not so much by the low content of CO_2 in the atmosphere, as by the inadequate tempo in the work of enzymatic mechanism, which governs the flux of the assimilates and their adaption." (1935 [163])

The suggested physical mechanism of assimilation and transportation also permits an explanation of the reasons why plants spend water, during their development process, in such an uneconomical way, i.e., why observed values of the transpiration productiveness are so small.

If we insert in formula (125) the average estimates of q_s and q which were found during summer daylight hours, and we assume that, according to the obtained results, $c - c_1 = 0.1c$ then, the ratio $\frac{A}{E}$ will become approximately equal to 1/360. This means that, for assimilating 1 gram of carbon dioxide the plant loses by transpiration, on the average, approximately 360 grams of water.

It should be noted that, although the ratio $\frac{A}{E}$ does not coincide exactly with the productivity of transpiration, [10] it can be assumed that its value is of the same order.

Thus, the estimate obtained gives a theoretical explanation of the observed orders of transpiration productivity values.

Formula (125) also explains the substantial dependence of transpiration productivity on vapor pressure deficit. This dependence was often pointed out by several experimentators who observed a considerable decrease in productivity of transpiration with a decrease in vapor pressure deficit, which occurred in greenhouses and also when passing from dryer climates to more humid ones (Maksimov, 1926 [165] and others).

The main conclusion, based upon all that has been said above, can be formulated in the following way:

Natural vegetation utilizes only a very small portion of the natural resources of energy and water. This portion is small even in comparison with the low "efficiency factor" which can be attained under full utilization conditions of diffused carbon dioxide. The hypothesis stating that the production of natural vegetation is limited by the carbon dioxide content in the air is, as a rule erroneous, since the obtained formulae proved that, in the case of complete utilization of atmospheric carbon dioxide,

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10) The productiveness of transpiration accounts for the augmentation of dry matter which increases somewhat in comparison with the assimilation of carbon dioxide, due to utilization of water and mineral substances of the soil by the plant, and decreases somewhat due to expenditure of some portion of dry matter during plant development for breathing, and also when some portions of leaves and roots die away.

plants can utilize not less than 5% of incoming solar energy (which, in fact, is rarely observed), and under this condition the productiveness of transpiration must be equal not to some thousandth, but to some hundredth (which is never observed).

Insofar as the production of natural vegetation under actual conditions is not limited by the carbon dioxide content in the air, it is obvious that this limit is determined by other factors which are, according to terminology used in plant physiology, "at minimum."

Studying the level of development of biological processes in various natural zones it is very essential to know the climatic factors which limit the production of biomass.

In § 11 it has been noted that the heat balance of the underlying surface (and the relationship of this balance to amounts of precipitation) are of a determining significance for quite a few essential characteristics of exterior physicogeographical processes. In investigating the correlations of climatic factors with the productiveness of natural vegetation the good agreement in the distribution of radiation balance and precipitation with the boundaries of geobotanical zones, that was established earlier, is of particular importance. This reveals the existence of high correlations between radiation balance and plant development conditions.

A considerable influence of radiation balance of the land surface on vegetation conditions can be corroborated by two more facts.

The calculations of annual variations in radiation balance at temperate latitudes were given in § 11. They proved that in climates with a limited growing season, values of radiation balance for the year and for the growing season are approximately equal. In other words, it was established that vegetation only grows during the period with positive radiation balance, minus that portion of this period in which positive radiation balance compensates negative radiation balance of the year's coldest months.

This conclusion complies very well with the second deduction which may be independently derived from the first one, resulting from the general analysis of available data on radiation balance.

The existing data on radiation balance calculations indicates that for sites or regions with a plant cover, even of a short natural life, annual sums of radiation balance are positive (including calculations for the tundras of Eurasia and North America). Negative annual sums were only found in high Arctic latitudes, where vegetation is scarce or there is none at all.

According to these conclusions, and also taking into account the obvious significance of radiation energy balance at the level of the active surface for transpiration and photosynthesis, it is only natural to consider the radiation balance of the underlying surface as the "base of energy" for the production of natural vegetation.

Following the idea by A.A. Grigor'ev, we will analyze changes in productiveness of vegetation cover with a certain "base of energy" of physicogeographical processes and at various correlations between this base and amounts of precipitation (i.e., at a fixed radiation balance of the underlying surface and with various values of $\frac{R}{Lr}$).

For this purpose, we must first determine the dependence of the amount

of evaporation upon parameter $\frac{R}{Lr}$ at a given R. This dependence, found through the "relation equation," is presented in graphic form in fig. 60 (the curves are for radiation balance values from 10 to 80 kg-cal/cm^2/year). Annual amounts of transpiration must be somewhat smaller than annual amounts of evaporation, however, experimental data have shown that for a closed vegetation cover transpiration constitutes the main portion of evaporation (it must be remembered that annual evaporation is determined almost entirely by amounts observed during the warm season of the year).

Thus, the dependence of transpiration on the parameter $\frac{R}{Lr}$ is basically similar to the dependence of summarized evaporation on this parameter. When photosynthesis developed with full utilization of carbon dioxide diffused by the atmosphere, then the productiveness of transpiration, according to formula (125) would depend mainly on the vapor pressure deficit of the air and decrease with the increase of deficits. In this case the productiveness of transpiration would be a monotonous decreasing function of parameter $\frac{R}{Lr}$, since with the increase in this parameter the dryness of climate and vapor pressure deficits of the air increase during daylight hours.

Since, with a decreasing parameter $\frac{R}{Lr}$, at a complete utilization of carbon dioxide, there must be an increase in the sums of transpiration and productiveness of transpiration, it is obvious that, under such conditions the productiveness of biomass will grow rapidly with the diminishing value of $\frac{R}{Lr}$, and the highest production level will be reached at the lowest possible value of $\frac{R}{Lr}$.

In other words, a plant cover that fully utilizes the carbon dioxide diffusion at a given radiation balance will increase the production with increasing precipitation.

Should the plants, under actual conditions, fully utilize the carbon dioxide, then, this deduction would contradict the conception of A.A. Grigor'ev concerning the existence of some optimum in the interrelations between the thermal energy base and precipitation, at which the productiveness of plants is at its maximum.

However, it has been noted above that, the hypothesis concerning full utilization of the possible flux of carbon dioxide should be considered as erroneous. Therefore the conclusion derived above does not characterize the actual dependence of productiveness of plant cover on climatic factors.

In order to investigate additional conditions which limit the energy of photosynthesis and prevent the plant from utilizing the carbon dioxide of the air to the full extent, attention must be drawn to results of well-known experiments which tested the effect of soil moisture on crop yields and on transpiration productiveness of various cultivated crops.

The dependency of the transpiration productiveness on soil moisture, which was derived by Helrigel and corroborated by results obtained by other investigators working with various plants show that, under actual conditions, there is an optimum of soil moisture for the transpiration productiveness, and going above or below this optimum level the productiveness of transpiration drops rapidly.

The decrease in the production of transpiration at a low soil moisture is explained by the impossibility of developing sufficient assimilating surfaces when the incoming moisture is scarce (because the expenditure of substance for breathing does not comply with the low level of assimilation production) and by some other reasons. A very low transpiration productiveness with excessive soil moisture is, in the first place, a consequence of unfavorable conditions for the development of plant roots when the plant is starved for oxygen in poorly aerated soil.

There are many reasons to assume that analogous changes in productiveness of transpiration are also observed in natural vegetation at various physicogeographical zones under diversified moistening conditions.

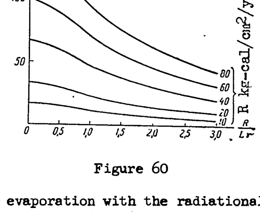

Figure 60

Relationship of evaporation with the radiational index of dryness and radiation balance.

It should be noted, above all, that available factual data have shown good agreement between mean patterns of soil moisture distribution for the warm period of the year and the boundaries of geographical zones (see maps of B.V. Poliakov, 1946 [187]; data given by G.S. ZHegnevskaia, 1954 [99] and others), which in turn coincide very well with the distribution of parameter $\frac{R}{Lr}$. Due to this it can be concluded that factual data, in accordance with theoretical reasonings given in the mentioned investigations confirm the existence of the dependency of soil moisture for the warm period on parameter $\frac{R}{Lr}$.

If average soil moisture decreases regularly during the warm season, with the augmentation of parameter $\frac{R}{Lr}$, it is easy to establish that in the

region of comparatively large values of parameter $\frac{R}{Lr}$ the productiveness of plants diminished rapidly with the increase of the ratio $\frac{R}{Lr}$ Indeed, the production of natural vegetation is equal to the product of summarized evaporation times the ratio of transpiration to total evaporation, times the productiveness of transpiration. In regions with dry climates, under an increasing index of dryness $\frac{R}{Lr}$, all three of these values must diminish: evaporation - according to the "relation equation" (see fig. 60), the relationship of transpiration and evaporation - resulting from the vegetation cover discontinuance and emergence of bare soil spots, transpiration productiveness - due to the rapid decrease in gaseous exchange (succulents) or the unproportionally great consumption for breathing by the huge root system, as compared with a limited carbon dioxide utilization of the small surface of leaves (mostly xerophytes). As a result of the concurrent decrease of the three multipliers their product must diminish very rapidly, which is in full accord with the visual evaluation of plant conditions when passing from dry steppe into semidesert and desert, and also with the measurement results of plant production in these geographical zones (Larin, 1936 [158], and others).

In regions with smaller values of the ratio $\frac{R}{Lr}$, further decrease of its values should bring about some augmentation in the total discharge of water for transpiration, however, it is conceivable that, a considerable reduction in transpiration productiveness will take place simultaneously.

Under excessive moistening conditions an additional increase in soil moisture must be necessarily associated with a deterioration of soil aeration, i.e., with a smaller supply of oxygen to plant roots. At the same time, a decreasing aeration and excessive moisture in soil bring about some serious changes in soil formation processes, among which the most vital for plant development is the weakening of the activity of nitrogen bacteria.

As a result of the poor work of the root system and a deficiency in mineral nourishment, the plant transpiration productiveness diminishes, and consequently with a decrease of the smaller values of $\frac{R}{Lr}$ the general productiveness of plants also diminishes (though not so drastic as the transpiration productiveness).

Thus, the natural vegetation productiveness, in full accordance with the statement by A.A. Grigor'ev, reaches a relative maximum at some optimal interrelationship of the energy base and precipitation (i.e., at some optimal value of $\frac{R}{Lr}$), and decreases with a change of this interrelationship in either direction from the optimum.

For a more descriptive interpretation of these conclusions it is expedient to offer a geographical sample which illustrates changes in the natural vegetation conditions with various values of parameter $\frac{R}{Lr}$ and with a more or less constant value of R.

If we examine the latitudinal zone between 48° and 49°N over Eastern Europe, from 52°E westward to 25°E, we will be able to observe variations in landscape associated with corresponding changes in parameter $\frac{R}{Lr}$ from 3.35 to 0.75, at an approximately constant radiation balance value. 11)

The effect of further decrease in parameter $\frac{R}{Lr}$ on vegetation cannot be observed in Europe (westward from 25°E, there are mountainous regions between 48°-49°N); for this purpose data for North America could be used. Examining the same latitudinal zone from 106°W eastward to the eastern coast of Newfoundland, it can be observed that the radiation balance will be somewhat less than in Europe, however, the difference will be comparatively small.

The general variation of parameter $\frac{R}{Lr}$, along 48°-49°N from 106° to 53°W corresponds to the range of values from 1.70 to approximately 0.30.

The longitudinal distribution of relationship $\frac{R}{Lr}$ is presented in fig.61.

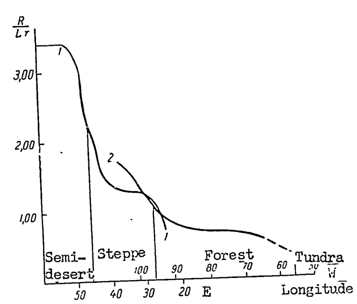

Figure 61

Longitudinal distribution of the radiational index of dryness.

11) Pertaining to moist surface conditions see § 11.

This figure shows: the relationship for Eastern Europe (curve 1), for the eastern portion of North America (curve 2), and the boundaries of geobotanical zones. The scale of the abscissa allows us to superimpose the curves at their borderline between forest and steppe in Europe and North America.

Fig. 61 shows that, moving westward from regions with large values of $\frac{R}{Lr}$ in Europe, which pertain to the boundary of deserts and semideserts, we will pass, with diminishing values of this parameter, through the semidesert and steppe zones, and observe an augmentation in natural vegetation productiveness. Passing from the area with the smallest values of $\frac{R}{Lr}$, in the eastern portion of Newfoundland, and moving towards the west, i.e, in the direction of the parameter's increase, we will transit from a sparse vegetation cover of tundra to forests with a productiveness markedly greater than in tundra.

Evidently, in both of these extreme cases; with very insufficient moisture in the extreme eastern portion of Europe and with considerable excess moisture on the eastern coast of Newfoundland, productiveness of natural vegetation is very much lower than that in the steppe, wooded steppe and forest, which pertain to average values of parameter $\frac{R}{Lr}$.

In the three indicated areas productiveness changes less rapidly, and it is somewhat difficult to detect the accurate position of maximum productiveness from the existing factual data, which are rather scarce 12).

However, there is no doubt concerning the fact that due to diminishing productiveness with a decrease of the small, and increase of the large values of $\frac{R}{Lr}$ such a maximum, according to the statement by A.A Grigor'ev, actually exists.

In the above cited papers of A.A. Grigor'ev it was noted that, vegetation productiveness becomes greater with an augmentation of the heat energy base, if the relationship of the energy base and precipitation remains at the optimum. In this case, with the highest possible value of the energy base, an absolute maximum of productiveness is attained.

In interpreting the heat energy base conception as the radiation balance of the underlying surface for the year, we will give an analysis of this statement by examining the natural vegetation productiveness at a specific value of $\frac{R}{Lr}$, which will be close to the optimum value, but with variable values of R.

The available factual data indicated that under conditions of sufficient precipitation the characteristics of soil moistening, and vapor pressure

12) It is conceivable that the maximum of natural vegetation production is found not in the steppe zone. According to data given by Clements and Wever (Jenny 1941) the production of dry matter in virgin steppe of North America increases markedly from west to east, following the increase in precipitation.

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deficit of the air, change only slightly with the transition from temperate to tropical latitudes (average summer conditions of the forest zone in temperate latitudes under humid tropical forest conditions).

At the same time, further southward, the influence of low temperatures, which prevent vegetation or decrease the intensity of photosynthesis, becomes smaller and smaller, and the vegetation period increases. Due to the influence of these factors, the transpiration productiveness in areas with a considerable radiation balance and sufficient amount of precipitation must be, at any rate, not less than in areas with an inconsiderable radiation balance.

Since the increase in radiation balance under moist climate conditions results in rapid augmentation of total evaporation, and consequently in a stronger transpiration (see fig. 60), it is evident that at the same order of magnitude of transpiration productiveness the natural vegegation production will substantially increase with the intensive radiation balance if the precipitation amount is sufficient.

These conclusions are corroborated by the computation results given by Fageler (1935), who determined that the annual production of fresh organic matter in a humid tropical forest reached a very high level: 100-200 ton/ha.

This value exceeds, at least several times, the productiveness of forests in temperate latidues (see the summary of factual data in Jenny's book, 1941).

The available data also corroborate the decrease in natural vegetation productiveness with the deviation of value $\frac{R}{Lr}$ in any direction from the optimum value, under conditions of large values of R , i.e., in the tropical area.

Fageler noted that, with diminishing precipitation, which is related to the transition from humid tropical forest conditions to monsoon forests and savanna, the plant productiveness drops down to 50-30 ton/ha per year. At the same time, under conditions of excessive moistening in tropical regions (swamps), the production of organic matter is also relatively small.

The question concerning variations in productiveness of natural vegetation that accompany changes in the radiation balance, when the relationship $\frac{R}{Lr}$ has a definite value markedly differing from the optimum, is not so clear. Such an example would correspond, for instance, to a comparison between the steppe of temperate latitudes with the steppe of the tropical zone.

Under conditions of insufficient moistening the augmentation of radiation balance could be associated with lower transpiration productiveness (because of a higher vapor pressure deficit and other reasons). This may curtail, to a considerable degree, those amounts of vegetation production which would have been secured by increased evaporation under conditions of a constant transpiration rate.

The above cited data corroborate the close relationship between conditions of natural vegetation development, as well as its productiveness, and climatic energy factors.

It is conceivable that analogous correlations also exist for such natural processes as soil formation and exogenous geomorphological developments.

The soil formation conditions are determined, to a considerable degree, by the vegetation productiveness and the intensity of the vertical moisture movement in the upper soil layers. The latter value depends directly on the intensity of runoff, which in turn, is correlated with climatic energy factors.

This provides us with reasons to assume that between the quantitative indices of the balance of organic and mineral substances in soil, and heat balance conditions, there are definite qualitative relations.

The intensity of exogenous geomorphological process, which is characterized by the magnitude of the erosion coefficient is, according to data given by B.V. Poliakov (1946 [187]), correlated with soil zonality. Insofar as soil zonality follows the distribution of the radiation index of dryness, we may assume that the level of the exogenous geomorphological process development depends, to a certain degree, on the indicated climatic factor. Thus, climatic energy factors exert a deep and decisive influence on the level of all natural processes in the outer geographical sphere.

Chapter V

Heat balance and meteorological effectiveness of ameliorative measures

Among various problems of physical geography, which are now analyzed by means of an extensive use of data on heat energy balance, the problem of a quantitative estimate of the transformation of natural conditions by melioration should be specifically emphasized. In a series of investigations, that were carried out in recent years, the hydrometeorological effectiveness of amelioration measures was calculated from data on heat balance components; this was done for areas of insufficient moistening and regions with a surplus of moisture.

The results obtained by these investigations were utilized by the hydrometeorological service for melioration works in various regions of the Soviet Union.

Computed data and measured values of heat balance components were successfully used in analyzing the hydrometeorological effect of amelioration, which confirms the practical value of the results of recent heat balance investigations.

In this chapter, two problems associated with an estimation of heat balance for the purpose of studying the melioration effect will be interpreted. The first problem concerns the field protective forest growing, the second- the application of irrigation in regions of dry climate.

§ 13. Field protective forest growing

At present, it is known that the main effect exerted by field protective forest shelterbelts on the meteorological regime in the air layer near the ground is their wind protective action.

The protective forest belts diminish wind speed over protected fields and reduce the intensity of the vertical turbulent air movement in the lowest air layer. The latter fact was first discovered and studied in detail by Soviet scientists (ĨŪdin, 1950 [244] and others), and constitutes a very important feature in the general mechanism of the forest belts' influence on the meteorological regime.

The weakening of the vertical turbulent movement in the lower air layer (up to a height of several meters), which is observed on protected fields, is explained by the fact that the forest belt breaks up and destroys air vortexes which move near the ground surface of open land. Therefore, the air flux that penetrates through the forest belt has no sizeable vortexes at all, which reduces the intensity of vorticity in this stream.

It should be pointed out that this effect will only take place in forest belts which can be easily penetrated, and where the air flow can blow

through rather freely. A dense, impenetrable forest belt affects air flow in a different way. A narrow calm zone was observed behind such a belt, but farther away the wind speed increased rapidly again and reached almost the rate that was observed in the open steppe. In this case no decrease in the size of air vortexes was observed in the air layer near the ground.

These phenomena are explained by the fact that, the air flux that approaches an impenetrable forest belt usually rises somewhat, flows over the top of it, then immediately descends on the other side and quickly restores its initial structure.

The reduction of intensity of vorticity in the lower layer of the air over protected fields is of great practical importance. Recent meteorological investigations show that vertical movements directly affect the development of meteorological phenomena: blowing away snow from fields during the colder season, and causing the occurrence of dust storms during the warmer period.

The reduction of intensity of vortical movements (turbulent exchange) in the proximity of the ground surface is of great importance in eliminating or weakening dust storms and in the preservation of snow on protected cultivated fields.

A diminished intensity of turbulent exchange is also very important in the preservation of moisture in soil during the warmer season.

The amount of possible evaporation (evaporability), as well as many other meteorological factors, depends on the intensity of the turbulent exchange in the air layer near the ground.

All this indicates the need for a quantitative estimate of the effect of forest belts on turbulent exchange, in order to determine the hydrometeorological effectiveness of various constructions of protective forest belts.

However, there are many difficulties in this task. Due to structural changes in turbulent vortexes, observed on the fields between the shelterbelts, it is not possible to determine, for this case, the coefficient of exchange by using conventional methods, which are based on the generalized theory of Prandtl (see § 4).

The attempts to utilize various indirect methods (estimating changes in wind speed, etc.) for determining variations of the exchange coefficient in between the forest belts, did not bring satisfactory results.

At present, the most popular methods for estimating the effect of forest belts on the intensity of turbulent exchange are those suggested by M.I. ĨŪdin and the author (Budyko and ĨŪdin, 1951, 1952 [62 & 63]), which are based on the analysis of heat balance components.

Let us briefly outline the essence of these methods.

From equation (38) we can obtain the following formula for the speed of evaporation:

$$E = \frac{\rho (q_1 - q_2)}{\int_{z_1}^{z_2} \frac{dz}{k}}, \qquad (130)$$

where q_1, and q_2 – are specific moistures at levels z_1 and z_2.

For the turbulent flux of heat P, as has been mentioned before, the following formula can be written:

$$P = \rho c_p D(\theta_w - \theta).$$

The turbulent heat exchange is determined by an equation similar to formula (130):

$$P = \frac{\rho c_p(\theta_1 - \theta_2)}{\int_{z_1}^{z_2} \frac{dz}{k}}, \qquad (131)$$

Where θ_1, and θ_2 — are temperatures at levels z_1 and z_2.

From these equations we can derive the following relationship:

$$D = \frac{E}{\rho} \frac{\theta_1 - \theta_2}{(q_1 - q_2)(\theta_w - \theta)}. \qquad (132)$$

The coefficient of external diffusion D could be experimentally determined by using formula (132) with measurements of evaporation, difference in temperature and humidity between two levels and difference in temperature between the underlying surface and air.

The application of formula (132) is possible for places located at some distance from the shelterbelt, since immediately behind the belt vertical streams of heat and moisture change considerably with height, and therefore equations (130) and (131) can not be fulfilled. Experimental data showed that the height of the quasi-stationary sublayer, inside of which relative changes of streams in a vertical direction are small, increases with distance from the shelterbelt. The height of this sublayer is about 1/50-1/100 of the distance from the windward side of the forest belt.

Obviously, vertical gradients of temperature and humidity on the protected field should correspond to the indicated regularity.

In studies of the meteorological effect of forest shelterbelts, parallel observations are usually carried out in the protected field and open steppe. This is usually done for the purpose of determining the ratio $\frac{D'}{D}$, where

is the coefficient of external diffusion on the field between protective forest belts, and D - the coefficient for the open field. D'

This relationship could be determined by formula:

$$\frac{D'}{D} = \frac{E'(q_1 - q_2)(\theta_1' - \theta_2')(\theta_w - \theta)}{E(q_1' - q_2')(\theta_1 - \theta_2)(\theta_w' - \theta')}, \qquad (133)$$

where all values with prime marks pertain to meteorlogical elements on the protected field.

It should be noted that these methods for determining changes in the intensity of turbulent exchange are applicable for cases with relatively small vertical gradients of temperature and humidity in the air layer near the ground.

In calculating average changes of turbulent exchange for a long period, another approximation method could be used. This method is based on use of

heat balance equations.

These equations could be written in the form of (see §11):

$$\left.\begin{aligned} R_0 - 4s\sigma\theta(\theta_w - \theta) &= LE + \rho c_p D(\theta_w - \theta) + A \\ R_0' - 4s'\sigma\theta'^3(\theta_w' - \theta') &= LE' + \rho c_p D'(\theta_w' - \theta') + A' \end{aligned}\right\} \quad (134)$$

where the values with prime marks pertain to protected field conditions.

For a sufficiently long period of time, and for more or less homogeneous surface and relief conditions in the open field and in the field inside a square between protective shelterbelts, the differences $R_0 - R_0'$, $L(E - E')$ and $A - A'$ will be small in comparison with terms $\rho c_p D(\theta_w - \theta)$ and $\rho c_p D'(\theta_w' - \theta')$. Subtracting one equation from the other and neglecting the small differences we obtain the relationship:

$$\frac{D'}{D} = \frac{\theta_w - \theta}{\theta_w' - \theta'} - \frac{4s\sigma\theta^3}{\rho c_p D}\left(1 - \frac{\theta_w - \theta}{\theta_w' - \theta'}\right) \qquad (135)$$

The second term on the right side of this equation is usually considerably smaller than the first one. This permits a substitution of coefficient $\frac{4s\sigma\theta^3}{\rho c_p D}$ by its mean value, which will be equal to about 1/4 for daylight hours of the warmer season. In this case, the last formula could be used in the following form:

$$\frac{D'}{D} = \frac{5}{4} \frac{\theta_w - \theta}{\theta_w' - \theta'} - \frac{1}{4}. \qquad (136)$$

Equations (133), (135) and (136) have been used in calculations of meteorological effectiveness of various kinds of forest belts (Budyko, Drozdov and others, 1952 [56]; Romanova, 1954 [196]; and others). Results of these calculations confirmed the above cited conclusion concerning the great importance of a penetrable structure of shelterbelts for obtaining a decrease in the intensity of turbulent exchange.

The investigations have shown that behind an easily permeable shelterbelt the intensity of turbulent exchange is reduced to the greatest degree at some distance from the shelterbelt, usually this distance is from 5 to 8 H (i.e., from 5 to 8 times the height of belt H). At greater distances from the belt, the air stream flowing over the shelterbelt gradually descends and mixes with the lower layers of the atmosphere. Consequently, the turbulent exchange gradually increases and at a distance of about 30H it reaches an intensity, which is approximately equal to that observed in open fields.

Behind a not easily permeable belt, the zone of weak turbulent exchange appeared to be considerably smaller, and it was often observed that, at a distance of 10-15H from the shelterbelt, the turbulent exchange did not decrease, but was even somewhat more intense.

Calculations of M.I. Ĭudin and other authors, using the method outlined above, showed that for shelterbelts of a most permeable and effective construction, the coefficient of turbulent exchange in the lower air layer decreased, on the average, by 30-35% at a distance reaching up to 10 H from the shelterbelt.

These calculations have led us to the conclusion that, the mean decrease of the coefficient of exchange in the large squares between the shelterbelts, approximating an area of 100ha, will be equal to 10-20%, dependent on the height of the trees of the shelterbelt. For the small squares, 20-30ha, the decrease of the exchange will reach 25-30%.

Quantitative data, obtained in this manner, showed an increase of the coefficient of external diffusion D on sheltered fields, and permitted a calculation of the effect of shelterbelts on possible evaporation(evaporability). For this purpose we have used the complex method, outlined above (see § 11), which is based on data of the heat balance on the earth's surface.

The principle formulas of this method, when neglecting the heat exchange in soil, will be:

$$\left.\begin{aligned} E_0 &= \rho D (q_s - q) \\ R_0 - 4s\sigma\theta^3 (\theta_w - \theta) &= L\rho D (q_s - q) + \rho c_p D (\theta_w - \theta) \end{aligned}\right\} \quad (137)$$

It should be remembered that by applying the second of these equations and using the well-known physical relationship between θ_w and q_s, we can find value q_s, and then, by using the first formula, determine the evaporation.

Here the value of evaporation is a function of the four principal meteorological factors: radiation balance, which is computed from air temperature R_0; air temperature θ; humidity q; and the ccefficient of external diffusion D.

By using equation (137) it is not difficult to investigate the nature of the relationships associating the value of possible evaporation with the factors cited above.

The dependence of the heat amount, that would be lost for the potentially possible evaporation LE_0, on radiation balance is presented in fig. 62. It is apparent that this value - the amount of heat lost for potential evaporation - is actually proportional to evaporability. Corresponding calculations were accomplished for the mean temperature and humidity of the air in spring (from March 15 - May 15), summer (from May 15 - August 15) and autumn (from August 15 to October 15) for the central Ukraine and western part of the North Caucasus (Budyko, 1951a /45/).

The changes of values LE_0 as dependent on values of R_0 are shown in fig. 62 by curves marked 1, which are almost straight lines, and the actual values of R_0 are shown by dots. They pertain to certain regions and seasons.

The distribution of dots in relation to line 2, which is the line showing the coinciding values of LE_0 and R_0, indicates that in spring, the amount of potential evaporation is somewhat smaller than $\frac{R_0}{L}$, in summer it almost reaches value $\frac{R_0}{L}$, and in the autumn it exceeds this value slightly. The sum of evaporability for all three seasons totals, in both cases, approximately 90% of value $\frac{R_0}{L}$.

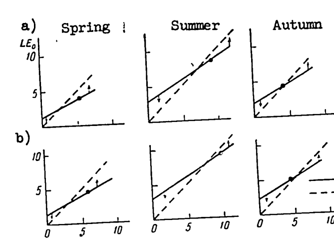

Figure 62

Relationship of heat expenditure for possible evaporation LE_0 in kg-cal/cm^2/month with the radiation balance R_0 in kg-cal/cm^2/month.

It can be deduced, from fig. 62,that with a variable radiation balance, the expenditure of heat for potentially possible evaporation could be larger than the radiation balance (curve 1 rising above line 2) and smaller than its value (curve 1 below line 2). Since, when the value of LE_0 is larger than R_0, in the air layer near the ground, a temperature inversion takes place (the evaporating surface receives additional heat by turbulent exchange), and with a converse relationship, there will be a superadiabatic gradient in the air layer near the ground. It is obvious that the changes of values R_0 will affect a change in the turbulent exchange, i.e., in value D.

There are many reasons to assume that at $R_0 \gg LE_0$ the actual values of $\overline{D}$, over an adequately large evaporation surface, will be larger than the value pertaining to average conditions of the equilibrium state of the lower layer, and at $R_0 \ll LE_0$ the values of D will be considerably below the average. This is effected by the ventil mechanism of turbulent heat conductivity in the air layer near the ground.

As a result of this, portions of curve 1, on the graph, which are located higher than line 2, must fall lower when approaching line 2 (but of course, can not sink below it), whereas the portions of curve 1, which are located below line 2, must rise correspondingly.

Thus, the actual dependence of LE_0 on R_0 must be characterized by some curve located between lines 1 and 2 in fig. 62. This dependency, in a large interval of R_0, will differ only slightly from the approximate equality of LE_0 to values of R_0, which confirms the conclusion made in chapter IV, concerning the determining significance of the radiation balance value for potential evaporation.

A close relationship of evaporability and radiation balance can also be seen from fig. 63, which presents the annual march of values E_0 and $\frac{R_0}{L}$ for the warmer season in the Northern Caucasus (calculations made for the Central Ukraine give very similar results). This figure shows that the forms of the curves of the annual march of E_0 and $\frac{R_0}{L}$ are exceedingly similar, the only difference being the shifting in time, which results from the fact that in spring E_0 is somewhat smaller than $\frac{R_0}{L}$, and in autumn a little larger. By taking into account the effect of the change in stability on the coefficient of exchange, the curves of E_0 and $\frac{R_0}{L}$ will be brought closer.

Figure 63

Annual variations of evaporability and radiation balance

The relationship of evaporability to the mean daylight hours temperature of the air, as calculated by the above formulas for average conditions of air humidity and temperature during spring and summer in the Central Ukraine, is presented in fig. 64 in the form of reclining curves, approaching straight lines. Values of evaporability that correspond to actually observed values of mean temperatures, in this case, are marked by dots on the curves.

The values of evaporability equal to the relationship of R_0 to the latent heat of evaporability are also shown in fig. 64, in the form of horizontal straight lines.

If the value of evaporability, determined from fig. 64, turns out to be larger than the value of $\frac{R_0}{L}$, then an inversion of temperature takes place over a sufficiently large surface; if the value of evaporability is markedly smaller than $\frac{R_0}{L}$, then superadiabatic gradients are established. Accordingly, in the first case, the coefficient of turbulent diffusion will decrease, which in turn will affect a diminution in evaporability, and in the second case, the coefficient of diffusion will grow, and by so doing increasing the evaporability. This will result in a rise of the left portions of the reclining curves in fig. 64, and in a lowering of the right portions, which shows that the actual dependence of evaporability on air temperature turns out to be considerably weaker than those shown in fig. 64.

Figure 64

Relationship of evaporability with air temperature

Thus, the effect of the stability on turbulent exchange in the air layer near the ground changes the dependence of evaporability on radiation balance, rendering it very close to a direct proportionality, and the dependence on air temperature turns out to be comparatively weak; when air temperature varies within relatively wide limits the value of evaporability stays approximately equal to $\frac{R_0}{L}$ 1)

However, it must be pointed out that, according to data for the Ukraine

1) It is necessary to emphasize that dependence of evaporability on temperature is here analyzed by taking into account a constant value of radiation balance. The general effect of temperature on evaporability (connected with the close dependence of the sums of temperatures on radiation balance, see § 11) is quite substantial.

and the North Caucasus the value of evaporability is, on the average, close to the value of $\frac{R_0}{L}$ even when the effect of stability on the exchange is not taken into account.

The clarification of the problem concerning the effect of shelterbelts on evaporability requires an establishment of the evaporability relationship to the exchange coefficient in the air layer near the ground.

Three of the four principal factors which exert an effect on evaporability (R_0, θ, q, D) are only slightly affected by shelterbelts (radiation balance calculated from air temperature data, air temperature, and humidity), but the fourth factor - the coefficient of outer diffusion-is quite remarkably changed.

The dependence of the evaporability values on coefficient D, was calculated by equation (137) for average conditions during spring, summer and autumn in the Central Ukraine and North Caucasus, and is presented in fig. 65.

Figure 65

Relationship of evaporation with the coefficient of external diffusion.
a)- Central Ukraine, b)- North Caucasus

As a result of the given data, in a wide range of changes in value D, the value $\frac{1}{E_0}\frac{\partial E_0}{\partial D}$ only slightly changes during various seasons. Hence, we may conclude that, under diversified conditions with various absolute values of D, the influence of a certain decrease in the exchange coefficient on the relative decrease of evaporability will be more or less equal - about 10% with a decrease of D by 0.1 cm/sec (somewhat more in autumn, and a little less in spring and summer).

By considering the available data on variations of the turbulent exchange in the air layer near the ground, which are effected by forest shelterbelts, we can deduce that for large fields (about 100 ha) protected

by forest belts, evaporability will decrease by 7-10%. For smaller squares between the belts, approximately 20-30 ha in size, the decrease in evaporability will reach 12-15%.

Similar calculations of the decrease in evaporability, as effected by forest shelterbelts, have been used for evaluating the meteorological effectiveness of forest belts of various structure, and also for determining the effect of forest belts on irrigation norms needed for irrigating fields protected by forest belts.

Besides reducing evaporability, shelterbelts promote an augmentation in snow storage on fields and also some increase in the precipitation amount.

The effect of these factors brings about a considerable gain in soil moisture of protected fields.

For a quantitative evaluation of soil moisture changes in protected fields, the combined analysis method of water and heat balances on the earth's surface can be applied.

Similar calculations were made by the author (1950b [4]). They were based on the following conclusions.

Let us present the equation of water balance in this form:

$$-\frac{\partial w}{\partial t} = E - r + f, \qquad (138)$$

where $\frac{\partial w}{\partial t}$ - is the speed of the water amount changes in the upper layer of the lithosphere in the course of time.

For a sufficiently large region, the value of f can be assumed as equal to the river runoff, which is measured by the network of hydrometeorological stations, as well as the rate of precipitation r. The possibilities for direct measurement of the two other terms of the water balance equation ($\frac{\partial w}{\partial t}$ and E) are limited, and due to this for the purpose of solving equation (138), it is desirable to establish the relations of these terms with observed hydrometeorological parameters.

For this purpose we use the formula which shows the relationship between the speed of evaporation and soil moisture, which was derived in § 5 and has the following form:

$$E = E_0\frac{w}{w_\kappa}, \qquad (139)$$

where w_κ - is the critical value of soil moisture.

For determining evaporation E_0 we use formulas of the complex method (see § 5), which include the heat balance equation.

As a result of a combined use of the indicated relationships we obtain the formulas:

$$\left.\begin{aligned} -\frac{\partial w}{\partial t} &= \rho D(q_s - q)\frac{w}{w_\kappa} - r + f \\ R_0 &= L\rho D(q_s - q)\frac{w}{w_\kappa} + \rho D c_p(\theta_w - \theta) + 3{,}6\sigma\theta^3(\theta_w - \theta) \end{aligned}\right\} \qquad (140)$$

to which is added the known physical relationship of q_s and θ_w, determined by the empirical formula of Magnus or by tabular data.

For comparatively small values of mean differences between the temperature of the underlying surface and the air, the relationship of $q_s(\theta_w)$ could be presented in the form of:

$$q_s(\theta_w) = q_s(\theta) + \frac{\partial q_s(\theta)}{\partial \theta}(\theta_w - \theta). \qquad (141)$$

Using this relationship and formulas (140) after excluding the value θ_w, we obtain:

$$-\frac{\partial w}{\partial t} = f - r + \rho D \frac{w}{w_\kappa} \frac{\left\{ \frac{\partial q_s(\theta)}{\partial \theta} R_0 + [q_s(\theta) - q] \, (\rho c_p D + 3{,}6\sigma\theta^3) \right\}}{L\rho D \frac{w}{w_\kappa} \frac{\partial q_s(\theta)}{\partial \theta} + \rho c_p D + 3{,}6\sigma\theta^3}, \qquad (142)$$

which determines soil moisture changes, in the course of time, by means of measured hydrometeorological parameters and some physical parameters.

Equation (142) can be integrated from the initial conditions $w = w_1$ at $t = 0$, and assuming r, f, θ, q, R_0, as definite functions of time. In this case, for the period of integration these values could be considered as approximately constant, then after integration we obtain an equation, which can be only applied to those cases where the difference between $\theta_w - \theta$, is not too great. As quantitative evaluations have shown, this condition is not always fulfilled. Furthermore, this equation is rather cumbersome and inconvenient for numerical calculations.

Due to this, in practical calculations of changes in the water balance of soil, it is expedient to use another method in solving equation (140).

In evaluating the errors, it can be established that in calculations of changes in soil moisture for periods on the order of decades and months, it is usually possible to substitute the derivative $\frac{\partial w}{\partial t}$ in equation (140) by a definite difference, without causing any reduction in accuracy. Then, similar to the manner suggested in § 5, it can be assumed that $w = \frac{w_1 + w_2}{2}$, and consequently:

$$w_2 - w_1 = \frac{1}{2w_\kappa} \rho D (q_s - q)(w_1 + w_2) - r + f, \qquad (143)$$

$$R_0 = \frac{1}{2w_\kappa} L\rho D (q_s - q)(w_1 + w_2) + (\rho D c_p + 3{,}6\sigma\theta^3)(\theta_w - \theta), \qquad (144)$$

where w_1 and w_2 - represent soil moisture at the beginning and end of one period of time.

By excluding value w_2, from equations (143) and (144) the value θ_w can be determined (by using certain data concerning the dependence of q_s on θ_w). Then, from equation (143), the value of w_2 is found from values of measured hydrometeorological parameters and from the initial moisture w_1. Having calculated value w_2 for the first time interval, it is obviously possible to continue calculations for the next interval (taking the determined value of w_2 for the corresponding new value of w_1), and in this way a sufficient number of points can be obtained for constructing the curve of soil moisture variations for the period in question. Simultaneously, annual variations in evaporation will be established.

Samples of such calculations, for some regions of insufficient moistening, are presented in the author's works (1950b and others /44/).

Utilization of equations (143) and (144) also permits us to solve the problem of quantitative calculations of the effect exerted by forest

shelterbelts on the soil moisture regime.

For this purpose, when solving equations (143) and (144), data of changes on protected fields in hydrometeorological parameters should be available. These parameters are: the coefficient of external diffusion D; precipitation r; runoff f; temperature θ; air humidity q and radiation balance R_0, computed from air temperature.

In evaluating, by equations (143) and (144), the influence of possible changes, in the above mentioned factors, on the regime of soil moisture and evaporation in shelterbelt regions, it can be established that changes in some meteorological elements cannot noticeably affect the soil water balance. Consequently, in the calculations we can use approximate values of these changes or ignore them completely.

The finished calculations show that, the strongest influence on the soil moisture regime and on average conditions of evaporation in protected fields must be exerted by changes in runoff f, by changes of the coefficient of external diffusion D and precipitation r.

Numerous observations have shown that, on fields between shelterbelts, the runoff of snow melt waters is considerably reduced. This reduction is mainly explained by different conditions of snow cover distribution in protected fields, as compared with unprotected ones; on protected fields between shelterbelts, the reduction in wind speed and turbulent exchange in the air layer near the ground creates favorable conditions for a regular distribution of snow cover, whereas, on open fields the larger portion of snow is drifted into ravines and other depressions of relief, and after melting, the major part is spent for runoff. Furthermore, the higher infiltration capacity of soil under forest belts assures somewhat greater retention of melt waters in protected fields, as compared with open spaces, which also reduce spring runoff of snow waters in this region.

As has been noted above, the coefficient of external diffusion in the air layer near the ground, on protected fields, is usually smaller than in open spaces. This noticeably affects evaporability and evaporation as well.

A definite effect on water balance of soil, in protected fields, can be also exerted by changes of precipitation amounts, which are brought about by the alteration in the intensity of vertical streams in the atmosphere over forest belts and by differences in evaporation.

For evaluation of the general effect of changes in indicated hydrometeorological factors on water balance in soil, equations (143) and (144) can be used when the values of the coefficient of external diffusion, runoff and precipitation, characteristic of conditions under which field protective shelterbelts grow, are taken into account.

Calculations accomplished by this method have indicated that, on protected fields, a considerable increase in soil moisture takes place, and evaporation is also somewhat augmented. The increase in soil moisture, as analyzed in its annual march, is of a variant nature, according to various conditions of turbulent exchange, runoff and precipitation.

In the case when, together with a considerable retention of the melt waters runoff, the shelterbelt system also drastically diminishes the turbulent exchange in the summertime and markedly increases precipitation, then soil moisture increases not only at the beginning of the growing season, but also in its second half.

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However, in case (as often happens) the effect of the shelterbelts is exerted mainly in an increase of snow storage and in diminishing spring runoff ("the winter effect"), then, as the solution of equations (143) and (144) has shown, soil moisture increases only in spring and at the beginning of the summer season.

The augmentation of productive soil moisture in this case may be on the order of tenths of per cent of the soil moisture amount absorbed in open fields, provided all other conditions are equal.

These statements are well confirmed by observational data.

A considerable increase in the amount of free moisture in soil, and greater general evaporation might contribute to a larger yield of crops under average climatic conditions. The augmentation of yields will be determined, above all, by an increase of the productive transpiration of plants (which will also be favored by the lesser exchange and lower wind speed in the air layer near the ground). Furthermore, under conditions of soil moisture increase, the relationship between the amount of water spent for transpiration and general evaporation must also be larger and this will substantially increase the yields of cultivated crops, provided that the general evaporation also increases.

Thus, application of shelterbelts, even without using other additional agrotechnical measures, may secure a considerable change in water balance of soil and may markedly increase the yields, which, as it is well known, is confirmed by results of numerous experimental investigations.

It should be remembered, however, that the effectiveness of the shelterbelt system is variable in wide limits, depending on the structure of shelterbelts and weather and climate conditions.

The application of data on the heat and energy balance permits us to calculate the effect of shelterbelts on such important indices as variations in turbulent exchange, evaporability and soil moisture in protected fields.

§14. Irrigation

On fields protected by a shelterbelt system, the components of heat balance change only slightly. Therefore, the evaluation of meteorological effectiveness of shelterbelts is based not so much on the estimates of changes in heat balance components under melioration conditions, but on the absolute values of these components.

Quite contrary to this, the application of irrigation in an arid climate results in considerable changes of the heat balance components, and therefore, the evaluation of the effect of irrigation on the meteorological regime must be based on, not only the absolute values of the balance components, but also on the estimate of their changes.

Irrigation, as applied in dry steppe, semidesert, and desert areas, brings about, above all, a substantial increase in radiation balance, which may reach up to and exceed several tens of per cents of its initial value. This somewhat paradoxical fact was established long ago by A.A. Skvortsov (1928 /211/), and since then it has been confirmed many times by results of

researches and calculations.

The increase of radiation balance caused by irrigation is explained by greater absorption of short-wave radiation, which is the result of a decreased albedo. The albedo of moist soil, covered with more or less abundant vegetation, is markedly lower than the albedo of semidesert or desert surfaces.

On the other hand, the lowering of the surface temperature and the higher humidity observed in the air layer near the ground under irrigation conditions results in a lower effective radiation, which also contributes to a greater radiation balance.

Irrigation, when used in a dry climate, results in a rapid increase of the expenditure of heat for evaporation, the value of which is mainly determined by irrigation norms. The usual irrigation norms augment losses of heat for evaporation, as a rule, in such a way that it exceeds the gain in radiation balance, and therefore the amount of turbulent heat emission markedly decreases and reaches, with sufficiently large norms of irrigation, the negative values which correspond to the average direction of the turbulent heat flux from the atmosphere to the underlying surface. This is manifested in the appearance of temperature inversions during daylight hours.

Thus, irrigation in dry climates considerably reduces the general heat flux from the underlying surface into the atmosphere. This is due to the reduced turbulent flux (which can even change the sign), and also the diminished heat flux transmitted by long-wave radiation. In cases where irrigation is applied to sufficiently large surfaces, this might result in significant changes of conditions effecting the air mass transformation in the given region.

As a typical sample of a change in heat balance components, effected by irrigation, we present here a scheme of changes in the heat balance for average conditions during summer in the southern Lower Volga regions where the irrigation norm was 10 gm./cm^2/month, i.e., 1000 m^3/ha/month, (fig. 66).

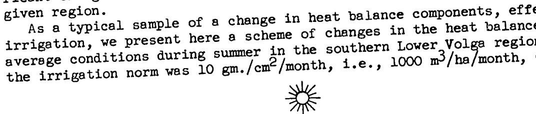

Figure 66

The effect of irrigation on the heat balance in the Lower Volga region (the components of the radiation and heat balances in kg-cal/cm^2/month rounded out to whole figures).

As can be seen from this scheme, irrigation in this case, accomplished an increase of radiation balance on the order of 40%, which is due to a marked reduction of the albedo and of the effective outgoing radiation sums (available data showed that, in case of sufficiently abundant irrigation in deserts, a still greater relative increase of radiation balance can be achieved).

A considerable increase of heat expenditures for evaporation results, in this case, in nullification of turbulent heat emission from the underlying surface to the atmopshere and a reduction in half of the total heat influx from the underlying surface to the atmosphere (the sum of turbulent heat emission and effective outgoing radiation).

Similar effects can be seen from observational data obtained by an expedition to Pakhta-Aral (D.L. Laikhtman lead the investigations). Results of these observations, presented in figs. 67 and 68, permit a comparison of the heat balance components, in the irrigated oasis with the surrounding semidesert area, in their diurnal march.

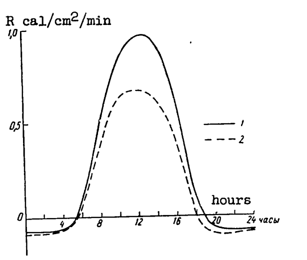

Figure 67

Diurnal variations of radiation balance in an irrigated oasis and in a semidesert 1- oasis, 2- semidesert

Data in fig. 67 show a much greater gain in radiation balance R of the oasis compared to that of the semidesert, during daylight hours. As can be seen in fig. 68, in the oasis a greater loss of heat for evaporation LE from irrigated fields is observed (in semidesert the evaporation for the analyzed period preactically equalled zero). The turbulent heat flux P in the desert is much greater than that in the oasis, and during daylight hours has an opposite sign (in the oasis it is directed to the earth's surface, and in the semidesert - to the atmosphere from the earth's surface).

The heat exchange in soil A only slightly changes, under these conditions.

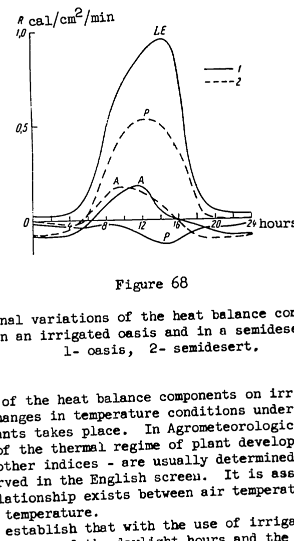

Figure 68

Diurnal variations of the heat balance components in an irrigated oasis and in a semidesert.
1- oasis, 2- semidesert.

Determination of the heat balance components on irrigated fields permits us to evaluate changes in temperature conditions under which the development of cultivated plants takes place. In Agrometeorological investigations, characteristics of the thermal regime of plant development - the sums of temperature and other indices - are usually determined from data on air temperature observed in the English screen. It is assumed that a more or less constant relationship exists between air temperature in the screen and the actual plant temperature.

It is easy to establish that with the use of irrigation the relationship between air temperature of the daylight hours and the temperature of the underlying surface markedly changes; on nonirrigated fields in a dry climate the daylight hours temperature of the underlying surface is, on the average, higher than the air temperature by 10°-15°C and even more, but after abundant irrigation, when an isothermal condition or temperature inversion is established over the irrigated field, the surface temperature will not be higher than air temperature during daylight hours. Consequently, on well-irrigated fields, the factual average daylight hours temperature of a dense vegetation cover can be considerably lower than the temperature of a non-irrigated active surface.

In connection with this, the evaluation of changes in the temperature regime affected by various norms of irrigation, it is expedient to utilize data on temperature of the underlying surface (the active surface), which in case of a more or less denser vegetation cover, is closer to the factual average temperature of the plants than the air temperature in the meteorological screen. Due to the fact that a satisfactory reliable and universal method for measuring the surface temperature is still nonexistent, the

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heat balance formula can be used for determining this temperature in the form of:

$$\theta_w - \theta = \frac{R_0 - LE}{(\rho c_p D + 3{,}6\sigma\theta^3)} \cdot \qquad (145)$$

When irrigation is applied to a comparatively small and isolated surface, temperature lapse rates over this surface are only changed inside a rather thin layer, and the effect of irrigation, applied to small limited plots, on the intensity of turbulent exchange will be very small. Therefore, in calculations using formula (145), when made for small irrigated plots, the values of D can be used without accounting for its changes effected by irrigation. The value LE, on the average for a more or less longer period, is determined by norms of irrigation and amounts of precipitation; R_0 can be determined from data of special balance observations, or by utilizing indirect methods of calculation which, by now, are sufficiently well developed (see chapter II).

Thus, by means of formula (145), it is possible to calculate factual temperatures of the active surface without irrigation, and also with irrigation applied to limited plots.

Considering the fact that the changes in air temperature, which are effected by irrigation, are considerably smaller than changes in the plant's temperature (A.A. Skvortsov, 1928 [211] and others), the change in average daylight hours temperatures of the underlying surface effected by the irrigation can be approximately determined by using formula (145) in this manner:

$$\theta_w - \theta'_w = \frac{(R_0 - R'_0) - (LE - LE')}{(\rho c_p D + 3{,}6\sigma\theta^3)}, \qquad (146)$$

where values pertaining to irrigated plot conditions are marked by the prime mark, and values for nonirrigated plots are given without it. [2)]

In calculations of changes in temperatures of the underlying surface for sufficiently large irrigated plots, by using formula (146), the dependence of the turbulent exchange intensity on variations in the thermal stratification of the air layer near the ground must be taken into account. This can be done by using additional data, which can be obtained from special experimental investigations (for instance, from data of the mentioned experimental investigation in the Pakhta-Aral oasis).

In addition to the calculation of average changes in the heat balance components for long periods (a month or so), and associated changes in temperatures, it is also of considerable interest to estimate changes in heat balance that are effected by irrigation in shorter time intervals. It is particularly interesting to investigate the dynamics of changes in heat balance after an application of irrigation as associated with desiccation of soil.

For calculating the changes in soil moisture, evaporation and the principal components of heat balance, as effected by irrigation, formulas

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2) It should be indicated that formula (146) can also be used for calculating the temperature of the active surface of slopes in various aspects, and also for solving many other microclimatic problems.

derived in § 13, (143) and (144), can be used.

The method of this calculation does not essentially differ from that of calculating changes in the water balance of soil without irrigation; only one term determining the incoming moisture from irrigation is added to the equation of water balance.

As an example of determining changes in evaporation and soil moisture, effected by irrigation, we here show results of calculations made for specific conditions of the summer of 1936 in the Saratov region.

Results of these calculations are presented in fig. 69, which shows variations in ten-day means of the principal heat balance components for the surface of a cultivated field (radiation balance R, expenditure of heat for evaporation LE, turbulent heat emission P, and also quantities of productive moisture in the upper layer (1m) of soil w), was calculated by the above mentioned equations for the warm period in Saratov (the solid lines). The theoretically determined heat balance components and the dynamics of soil moisture with two applications of irrigation (on May 20 and June 20, using 6 gr/cm^2 in each case, i.e., 600 m^3/ha) are represented by dotted lines. This calculation shows that, as a result of irrigation, together with a rapid increase in the amount of productive moisture in the soil, expenditure of heat for evaporation grows significantly, turbulent heat emission diminishes, and the radiation balance slightly rises. These changes are greatest immediately following irrigation, and later they gradually diminish.

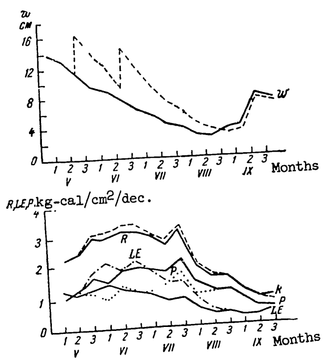

Figure 69

Variations of ten-day mean values of moisture content in soil and components of heat balance in the course of time. Saratov, 1936.

Similar calculations can be used for working out standard norms of irrigation, with due consideration to weather and climate conditions.

For various stages of crop development, certain conditions of soil moisture are required. When the moisture content in the soil diminishes below required values, then, the yield, as a rule, is considerably reduced.

Taking into account the amount of soil moisture which is required by plants during various stages of their development, we can calculate the amount of water which is necessary for securing a sufficient soil moisture, under certain weather and climate conditions, by using the above equations. In this way the standard norm of application will be determined.

A simpler, but also a more schematic way to account for the effect of weather and climate factors on norms of irrigation is the following.

With sufficient soil moistening the evaporation is, on the average, close to the evaporability value. Accordingly, we may assume that, for agricultural crops requiring sufficient moisture, the norm of irrigation must equal evaporability minus precipitation. 3) A similar method of calculation was used by S.A. Sapozhnikova (1951), A.M. Alpat'ev (1954 [12]) and others.

It could be taken for granted that such a method gives the most accurate results when calculating evaporability by the complex way, taking into account the effect of radiation balance on the process of evaporation and also the effects of air temperature, air humidity and turbulent exchange as well. For desert and semidesert climates, it is expedient to evaluate the influence of irrigation on temperature and humidity of the air, which somewhat changes the evaporation in oases of considerable dimensions.

For those crops which do not require high soil moisture during certain periods of development, a similar calculation method could be used, assuming that for particular time intervals the necessary evaporation comprises a certain portion of evaporability. Such a method for calculating norms of irrigation has a semiempirical character and requires, together with utilization of data on evaporation, an estimate of certain factual data on expedient norms of irrigation, at least for some regions.

Aside from accounting for the effect of weather and climate factors on norms of irrigation, data on heat balance are necessary for calculating the changes of air temperature and humidity in irrigated oases. These calculations have been done by M.I. IUdin, D.L. Laikhtman and other authors. They are important for solving many questions in the agrometeorology of irrigated fields.

As can be seen from all that has been said above, utilization of data on heat and energy balance is of a positive importance in clarifying the effect of irrigation on the meteorological regime and in evaluation of the effect of meteorological factors on norms of irrigation.

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3) The loss of water for infiltration into deeper soil layers must be also taken into account. This loss can comprise 20-30% of the irrigation norm.

Chapter VI

Heat and water balances of the earth

Data on geographical regularities of heat balance components, described in the preceding chapters, permit us to determine the heat and water balance for the earth as a whole, and for the latitudinal zones as well.

Calculations of this kind have been made before, but only in a schematic way and by methods of an insufficient correctness. All this has limited accuracy of determined values for heat and water balance components. Calculations, accomplished by the author (1949b [42]), cannot be regarded as being quite reliable either, since at that time world maps for the heat balance components of the earth's surface were not available. Only after completion of a series of world maps for heat balance components by the Central Observatory did it become possible to make a more accurate computation of the heat and water balance of the earth.

These data are of a definite importance in determining the general laws governing heat and moisture exchange in the outer geographical medium.

§ 15. The heat balance of the earth

The preceding investigations of the earth's heat balance were mainly concerned with determining the heat balance components for the earth - atmosphere system, i.e., with calculating the heat balance components for the earth as a planet. Due to this, the data on average values of heat balance components of the earth's surface were comparatively scarce. They were found in investigations of Baur and Philipps (1935), in the papers of the author (1949b [42]), Houghton (1954) and some others.

These calculations of heat balance are not based on world maps of heat balance components. Therefore, the use of world maps of heat balance, which now constitute the Atlas of the Heat Balance (1955 [15]), permits us to complete and make more accurate the calculations of the average values computed earlier.

The mean latitudinal values of the heat balance components, for land surfaces and oceans, and also mean values of the heat balance components for the whole surface of the earth are given in table 14. The data presented for the region between 60°N and 60°S are obtained from data which were directly used for the construction of maps of the above mentioned atlas. For calculations of the balance components which pertain to the entire surface of the earth, approximate values of the balance components in the Arctic and Antarctic regions were additionally determined. Since the indicated regions cover a relatively small portion of the earth's surface, the inaccuracy of these additional calculations could not affect, to any substantial degree, the mean values of the balance components for the entire surface of the earth.

Table 14

Mean latitudinal values of the heat balance components of the earth's surface in kg-cal/cm^2/year.

Latitude	Oceans					Land				Earth				
	$(Q+q)$	R	LE	P	A	$(Q+q)$	R	LE	P	$(Q+q)$	R	LE	P	A
60—50° C	88	34	34	18	—18	93	23	19	4	91	28	25	10	—7
50—40	109	54	51	15	—12	119	38	22	16	114	46	36	15	—5
40—30	136	78	73	12	—7	159	56	26	30	146	69	53	20	—4
30—20	151	100	85	7	8	184	64	23	41	163	86	60	20	6
20—10	156	110	89	5	16	182	74	36	38	163	101	75	14	12
10—0	149	107	76	5	26	149	79	58	21	149	101	72	9	20
0—10 Ю	152	107	81	7	19	143	75	59	16	150	99	76	9	14
10—20	155	107	97	9	1	161	69	44	25	156	99	85	13	1
20—30	147	94	87	10	—3	169	62	29	33	152	87	74	15	—2
30—40	128	73	77	12	—16	149	55	29	26	130	71	72	14	—15
40—50	104	53	57	5	—9	112	39	24	15	104	53	56	5	—8
50—60	84	31	37	12	—18	80	26	18	8	83	31	37	12	—18
Whole Earth	128	77	68	9	0	132	46	27	19	129	68	56	12	0

The comparison of data in table 14 with those given in the author's work (1949b /42/) shows that basic qualitative relations in the distribution of heat balance components, found in this investigation, are well confirmed by the more accurate calculations. At the same time, many quantitative values of the balance components were greatly changed as a result of using new data.

Thus, in particular, the new calculations give smaller values of radiation balance in higher latitudes on the land and ocean and larger values in lower latitudes for the ocean. As a result of this a marked increase in differences of mean values of radiation balance for the earth as a whole is observed between high and low latitudes.

Values of turbulent heat emission from the ocean surface to the atmosphere and losses of heat for evaporation are somewhat greater than those given in the paper of 1949. Values of turbulent heat emission from the land surface to the atmosphere turned out to be somewhat greater, but in higher latitudes - markedly smaller. The mean latitudinal redistribution of heat by sea currents (term A) increased somewhat in its absolute value and the change in values of this term with latitude became more regular.

Now, let us point out some deductions pertaining to characteristics of the latitudinal distribution of heat balance components that result from table 14.

The total radiation, increasing regularly from higher to lower latitudes, on land as well as on the ocean, has its maximum not at the equator but in belts of high pressure near the 20° latitudes. The equatorial minimum is apparently explained by the considerable increase of cloudiness at the equator.

The radiation balance on land, as well as on the ocean, grows rapidly with decreasing latitude only in temperate zones, while in tropical regions its value is only slightly dependent on the latitude.

Expenditure of heat for evaporation on the land and ocean changes with latitude in a different way. On land, the greatest evaporation is observed at the equator (where the great precipitation amount assures sufficient soil moisture), while in latitudes of the high pressure belt evaporation diminishes rapidly, due to prevalent dry climatic conditions. Contrary to this, on the oceans, maximum values of evaporation are observed in the high pressure belt, where the inflow of solar energy is especially great. In the proximity of the equator the evaporation from the ocean markedly diminishes.

Turbulent heat emission from the ocean surface to the atmosphere is comparatively small in all latitudes. Its values increase somewhat with higher latitude due to a growing significance of warm currents which warm the air in the cold season.

On land, turbulent heat emission is considerably greater, with maximum values in the high pressure belts where expenditure of heat for evaporation is lower due to the arid climate.

Many interesting features are present in data on the distribution of component A , which characterizes latitudinal values of the heat flux between the ocean surface and deeper layers, as influenced by sea currents.

As is seen in table 14, on the average, oceans absorb heat in latitudes approximately between 20°S and 30°N. In higher latitudes this heat is emitted, introducing an important factor in effecting a milder climate of temperate and higher latitudes during the cold season.

The last line in table 14 shows that a unit of the land and ocean surface actually receives an equal amount of total radiation. The radiation balance of the ocean surface is, on the average, greater than that of the land, which is due to the smaller values of the albedo for the ocean and a smaller (on the average) effective outgoing radiation.

The expenditure of heat for evaporation, from a unit of surface on the ocean, is noticeably greater than on land due to the absence of the limiting effect of dry climate. Consequently, turbulent heat emission from the land surface is approximately twice as great as heat emission from the ocean.

A noteworthy conclusion from the given data is the statement that the turbulent heat flux in all latitudinal zones (from 60°N to 60°S) is directed from the earth's surface to the atmosphere on land as well as on the ocean.

It was not long ago when in meteorological literature, including textbooks, the opinion was generally prevalent that the turbulent flux of heat, on the average, is directed from the atmosphere to the earth's surface. This utterly unnatural conception, which was introduced by W. Schmidt (1921) and others, was supported by the results of heat balance calculations made by Baur and Philipps (1935) and others.

In the above mentioned investigation, Baur and Philipps calculated the heat balance of the underlying surface and determined the turbulent heat exchange, by Schmidt's method, for two points (Batavia and Lindenberg) and later they obtained its latitudinal distribution by interpolation and extrapolation of these data.

Baur and Philipps did not use factual data on heat expenditure for evaporation but computed this value as the remainder term of the balance, neglecting the redistribution of heat by sea currents.

Such a calculation provokes some objections. Modern investigations (works of the author and M.I. I͡Udin, 1946, 1948 /60 & 61/) established that the method suggested by W. Schmidt for determining turbulent heat exchange in the atmosphere is not correct, since the supposition made by Schmidt, concerning the proportionality of turbulent heat flux to the gradient of the potential temperature, does not fit the natural conditions.

In addition, it may be noted that the definition of the latitudinal distribution pattern of a heat balance component for the whole Northern Hemisphere from data of only two points is hardly possible. In calculations of heat balance, factual data on evaporation cannot be neglected, since they permit the verification of calculations of other heat balance components by completing the balance equation.

Thus, the conception that turbulent heat flux is directed from the atmosphere to the earth's surface cannot be considered as substantiated by calculations of its heat balance.

The conclusion concerning the direction of the average heat flux from the underlying surface to the atmosphere is corroborated, at the present time, by all of the recent computations of the heat balance components;and also by independent calculations of turbulent heat flux in the atmosphere, which are associated with the introduction of the equilibrium gradients of temperature conception, given by the author and M.I. I͡Udin (1946, 1948 /60 & 61/). This allows us to conclude that the question of the average direction of turbulent heat flux has been completely solved.

It should be pointed out that in recent foreign investigations of heat balance of the underlying surface, a deduction that the average direction of turbulent heat flux is from the earth to the atmosphere, was also derived. At the same time, in these investigations the turbulent heat flux value is usually determined by very approximate methods.

To compare the presented results on heat balance of the earth's surface with those of recent foreign investigations we will provide data obtained by Houghton and published in 1954.

Houghton calculated the components of heat balance for the entire earth's surface without using any maps of heat balance components. He determined the turbulent heat exchange value as the remaining term of the balance equation. Since the last value turned out to be the comparatively small difference between radiation balance and heat amount spent for evaporation, the accuracy of its determination, as Houghton correctly noted, turned out to be rather low.

Counting the amount of solar radiation being received at the outer border of the atmosphere, as being equal to 100 units, Houghton found the amount of radiation absorbed by the earth's surface equal to 47 units, and effective radiation - 14 units. The radiation balance of the underlying surface, according to Houghton, is consequently, equal to 33 units. Having

determined heat spent for evaporation as being equal to 23 units, Houghton found that the turbulent flux of heat from the earth's surface to the atmosphere comprises 10 units, i.e., equals 10% of that received at the outer boundary of the atmosphere.

According to data presented, which were obtained as a result of considerably more detailed and more substantiated calculations, it can be concluded that the amount of absorbed radiation comprises 44% of the solar radiation flux received by the outer border of the atmosphere; effective outgoing radiation comprises 17%, radiation balance - 27%, and heat spent for evaporation - 22%. Consequently the turbulent heat flux value equals only 5%, i.e., it constitutes only half of the value of Houghton's calculations.

A more correct value of turbulent heat flux was obtained by Lettau (1954) who found this value as being equal to 4 relative units. It should be indicated, however, that the coincidence of this value with results of the author's calculations is accidental, since Lettau proceeded from very inaccurate estimates of absorbed and effective radiation for the earth's surface.

Data on heat balance, obtained from investigations accomplished in the Central Observatory, can be used for determining the heat balance components for individual oceans and continents.

In table 15 the mean heat balance components for the Atlantic, Indian and Pacific Oceans, calculated from these data are given.

Table 15

Heat balance of the oceans in kg-cal/cm^2/year.

Ocean	R	LE	P
Atlantic Ocean	75	63	12
Indian Ocean	78	83	8
Pacific Ocean	82	68	7

As can be seen from table 15, average amounts of radiation balance for the three oceans differ only slightly. The expenditure of heat for evaporation in the Indian Ocean is somewhat larger than in the other two. The turbulent heat exchange with air is greatest in the Atlantic Ocean (this is explained by the effect of the Gulf Stream).

Considerable interest surround the problem of heat exchange between oceans in connection with the activity of sea currents. From data of table 15, it might be concluded that, for the Atlantic Ocean the gain and discharge of heat resulting from heat exchange with other oceans is approximately compensated, while the Indian Ocean (where expenditure of heat for evaporation and turbulent heat emission is greater than the radiation balance) receives some amount of heat from the Pacific Ocean (where radiation balance is somewhat greater than the expenditure for evaporation and turbulent heat exchange). However, it should be noted that the absolute value of the obtained differences of the heat balance components (in relation to the entire surface of the oceans) turned out to be comparatively small and, apparently, range within the limits of the accuracy rates for these calculations. This complicates the solution of the question of mutual heat exchange in the oceans.

The average values of the heat balance components for the continents are presented in table 16.

Table 16

Heat balance of the continents in kg-cal/cm²/year.

Continent	R	LE	P
Europe	33	22	11
Asia	41	23	18
North America	38	24	14
South America	71	52	19
Africa	69	31	38
Australia	66	25	41

It can be concluded, from table 16, that average conditions of heat balance of individual continents differ considerably. The greatest values of radiation balance and loss of heat for evaporation are characteristic for South America, the major portion of which is situated in latitudes close to the equator. The greatest turbulent heat emission is observed in Australia and Africa, where tropical dry climatic conditions over vast surfaces are also observed.

The utilization of these data on the heat balance of the earth's surface also permits us to improve some deductions concerning the heat balance of the atmosphere.

The latitudinal distribution of the heat balance components of the earth - atmosphere system is presented in fig. 70. The data on radiation balance of this system are adapted from N.A. Bagrov's work (1954b /18/). These data comply fairly well with results of an earlier calculation performed by Simpson (1928), which were used by the author in his investigation (1949b /42/). It is noteworthy to mention that, calculations of radiation balance accomplished by N.A. Bagrov are well confirmed by data of K. IA. Kondret'ev and O.P. Filippovich (1952 /136/) and also by deductions of other investigations.

Data on heat expenditure for evaporation LE and transfer of heat by sea currents C are taken from table 14. The gain of heat from condensation Lr is assumed to be equal to the product of the latent heat of vaporization and the amount of precipitation.

The redistribution of heat between latitudinal zones, as effected by atmospheric circulation, is obtained as the remainder term of the heat balance equation.

As can be seen from data in fig. 70, four basic latitudinal zones with essentially different relationships of the heat balance components of the earth - atmosphere system are discerned in each hemisphere. In the equatorial zone, which extends north and south of the equator up to 10-15°

latitudes, the gain of heat from a great positive radiation balance is supplemented by a considerable amount of heat from moisture exchange (i.e., from the difference between the heat of condensation and loss of heat for evaporation). These sources of heat assure a great expenditure of heat for atmospheric and oceanic advection, for which the relatively narrow subequatorial zone presents a very important source of energy.

Figure 70

Heat balance of the earth - atmosphere system.

Northward and southward from the subequatorial zones are the regions which can be conventionally called tropical (they also include a portion of subtropical areas). In these zones, with a positive and constantly dimin-

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ishing with higher latitudes radiation balance, an expenditure of heat for moisture exchange is observed. Sometimes it reaches considerable amounts. In the major portion of the tropical zone the loss of heat for moisture exchange approaches the value of the radiation balance, and due to this the heat loss for atmospheric advection is small.

In the region of 35°- 40° latitudes a transitional zone takes place. In this area the gain and expenditure of heat in all heat balance components is approximately well balanced, i.e., on the average, all of them are not large in absolute values. In the higher latitudes are regions of negative radiation balance which grows rapidly, in absolute values, with increasing latitude. The negative radiation balance in this zone is compensated by the gain of heat from atmospheric advection, moisture exchange, and sea currents.

It should be noted that in the general heat exchange between high and low latitudes all four terms of the heat balance play an important role, and consequently, the effect of any component cannot be neglected without substantially impairing the accuracy of interlatitudinal heat exchange calculations.

As can be seen from fig. 70, a comparatively great macroturbulent heat conductivity of the atmosphere for horizontal streams of heat results in the fact that values of atmospheric advection turn out to be, on the average, larger than values of oceanic advection.

The total amount of heat transferred from one latitudinal zone to others by advection, and also in connection with moisture exchange, can be derived from the curves in fig. 70 by proper computation.

By using the obtained data on latitudinal distribution of the heat balance terms of the underlying surface and also of the earth - atmosphere system, the latitudinal distribution of the heat balance components of the atmosphere can be determined.

Since the radiation balance of the underlying surface is, in all latitudinal zones, greater than that of the earth - atmosphere system (due to the "greenhouse effect"), the radiation balance of the atmosphere at all latitudes is a negative value. The latitudinal distribution of the atmosphere's radiation balance, determined as the difference between the radiation balance of the earth - atmosphere system and the underlying surface, is presented in fig. 71 by curve R_a. The radiation balance value on various latitudes turned out to only slightly vary.

The gain of heat from condensation, determined as the product of the latent heat of vaporization times the latitudinal sums of precipitation, is represented in fig. 71 by curve Lr, the increase of heat from vertical turbulent heat emission - by curve P, the gain of heat from atmospheric advection - by curve C.

Using data presented in fig. 71, we can show, by appropriate calculations, that between the Northern and Southern Hemisphere, for mean annual conditions, considerable redistribution of heat energy is observed. From these results, it can be particularly concluded that, the Northern Hemisphere transfers a great amount of heat by atmospheric advection and by sea currents to the Southern Hemisphere.

At the same time, the Southern Hemisphere emits a marked amount of heat

through moisture exchange between the Northern and Southern Hemispheres. From the above mentioned calculations of water balance, accomplished by L.I. Zubenok, it can be concluded that in the Southern Hemisphere the amount of precipitation is approximately 120 mm a year less than evaporation. Thus, the appropriate amount of water is transferred in the hydrosphere from the Northern Hemisphere to the Southern, and in the atmosphere - from the Southern Hemisphere to the Northern. The latter process results in a transfer to the Northern Hemisphere of a considerable amount of heat emitted by condensation of precipitation.

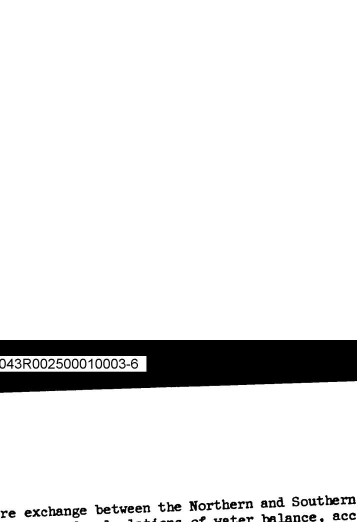

Figure 71

The heat balance of the atmosphere.

Since the questions of the atmosphere's heat balance are only incidentally considered here, we will not linger on a more detailed discussion of obtained regularities or on comparison of derived conclusions with deductions of other authors who treated this problem, but will turn to the evaluation of the heat balance components of the earth as a whole.

Values of the components of the earth's heat balance, which were calculated by taking the above data into account, are presented in a schematic form in fig. 72.

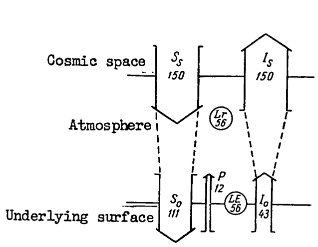

Figure 72

Heat balance of the earth (the heat balance components in kg-cal/cm^2/year).

Because of the earth's spherical form a unit of the surface on the outer boundary of the atmosphere receives, on the average, only one fourth of the total value of solar radiation flux, which is, according to recent data, close to 1000 kg-cal/cm^2/year, i.e., about 250 kg-cal/cm^2/year.

To determine the quantity of radiation that is absorbed by the earth as a planet, the average value of the earth's albedo must be known. At the present time there are two methods for determining this value: by calculation, taking into account the conditions of reflection from land surfaces, oceans, clouds, etc. and by measuring the illumination of the dark portion of the moon (which is visible due to reflection of a portion of solar radiation from the earth's surface - the phenomenon of the "ashen light").

Results obtained by both methods agree fairly well (Danjon, 1936; Penndorf, 1937; Fritz, 1949; Bagrov, 1954b /18/).

The most detailed calculation of the albedo was recently made by averaging the data presented on the world map of the albedo distribution of the earth - atmosphere system. These calculations gave the mean weight value

of the albedo as being 0.41 (Fedoseeva, 1953 /226/).

Most of the recent investigations present somewhat smaller values of the albedo. Taking this into account, we will use a rounded value of the albedo, 0.40, in this calculation.

Counting the general albedo of the terrestrial globe as being 0.40, we will find that short-wave radiation, which is absorbed by the earth, is approximately equal to 150 kg-cal/cm^2/year. This value is designated in fig. 72 by S_s.

According to the above data, the earth's surface receives 229 kg-cal/cm^2/year of the solar short-wave radiation. The mean weighted value of the albedo of the earth's surface is, on the average, equal to 0.14, as the calculation show. Thus, the earth's surface absorbs 111 kg-cal/cm^2/year of solar energy (S_0, fig. 72) and reflects 18 kg-cal/cm^2/year.

As the given values show, the terrestrial atmosphere absorbs 39 kg-cal/cm^2/ year, i.e., almost three times less than the earth's surface.

Since the radiation balance of the earth's surface equals 68 kg-cal/cm^2/year, it is evident that the effective radiation at the surface level is, on the average, 43 kg-cal/cm^2/year (I_0, fig. 72).

The total amount of long-wave radiation of the earth, which equals the amount of absorbed radiation, is close to 150 kg-cal/cm^2/year (I_s , fig. 72).

It is noteworthy to mention that the ratio of the effective radiation from the earth's surface to the total outgoing radiation of the earth $\frac{I_0}{I_s}$ is much smaller than the corresponding ratio of the amount of absorbed radiation $\frac{S_0}{S_s}$. This difference shows the tremendous influence of the greenhouse effect upon the thermal regime of the earth.

Due to the greenhouse effect the earth's surface receives about 68 kg-cal/cm^2/year of radiational energy (radiation balance), which is spent partially on water evaporation (56 kg-cal/cm^2/year, LE in fig. 72). and partially returned to the atmosphere by means of turbulent heat emission (12 kg-cal/cm^2/year, P, in fig. 72). As a result of this the atmosphere's heat balance consists of the following values:

1) heat gain from absorbed short-wave radiation, which equals 39 kg-cal/cm^2/year;
2) heat increase from condensation of water vapor (Lr in fig. 72) equal to 56 kg-cal/cm^2/year;
3) heat gain from turbulent heat emission of the terrestrial surface, which is equal 12 kg-cal/cm^2/year;
4) heat expenditure by the effective outgoing radiation into cosmic space, which is equal to the difference of the values I_s and I_0, i.e., 107 kg-cal/cm^2/year.

The last value coincides with the sum of the first three heat balance components.

Data presented on the heat balance of the earth differ somewhat from the data given in the author's work (1949b /42/) for the balance of the Northern Hemisphere. This difference is explained not only by a higher accuracy of the foregoing computations, but also by the fact that the components of radiation balance, and especially of heat balance for the Northern Hemisphere and the terrestrial globe cannot coincide because of the inhomogeneous distribution of land and oceans in the Northern and Southern Hemispheres.

It may be noted that these values of the heat balance components do not differ much from corresponding values obtained in recent foreign investigations. It must be assumed, however, that data on heat balance, obtained by averaging a series of world maps showing the distribution of the balance components, should be more reliable as compared with results of summarized calculations performed in relation either to the earth as a whole or to latitudinal zones.

§ 16. The water balance and hydrologic cycle

The above results on heat balance can be used in studies of fundamental regularities in the natural hydrologic cycle.

On the basis of the above obtained data on heat spent for evaporation, we can make a series of conclusions concerning the water balance of the earth.

Since the completed calculations gave the average value of evaporation for the terrestrial globe as being equal to 930 mm/year, it is obvious that the annual amount of precipitation must also equal this value. It must be remembered that, heretofore, no reliable evaluation of the average amount of precipitation received by the earth's surface was obtained, since data on precipitation over the oceans were inaccurate.

Up to recent times, there was an opinion that constructed maps of precipitation on the ocean were based largely on data of coastal stations and therefore, precipitation amounts on the oceans, determined from these maps, must have been considerably exaggerated when compared with actual values. 1)

Accordingly, in calculations of water balance on the oceans, the average amounts of precipitation were often reduced by 20-30% in order to adjust them to evaporation data, which usually appeared to be considerably smaller than the precipitation sums and river runoff to the oceans. A similar reduction of the precipitation amounts was done by Wüst (1936) and also by the author in his preceding work (1949b [42]).

However, in doing this, the inadequate reliability of evaporation data from the ocean that existed at that time, when world maps of evaporation were not available, has not been taken into account to a sufficient degree.

The presented results of evaporation calculations from the ocean were initially obtained by averaging the comparatively detailed world map. They give a value somewhat greater when compared with the majority of earlier calculations.

While in preceding works the value of evaporation from the ocean usually ranged between the limits of 750-1000 mm/year, the newest result shows 1130 mm/year. Since the total runoff from rivers into the oceans gives a water layer of approximately 100 mm/year, then, with the given value of evaporation, the total precipitation amount over the oceans will be about 1030 mm/year. This value is somewhat smaller than the corresponding amounts obtained by Meinardus (1934) from Schott's maps (1140) mm) and recently by L.I. Zubenok from maps constructed by O.S. Drozdov (1120) mm). However, the difference between the presented values of precipitation and the sum of evaporation and runoff is, in this case, comparatively small. Therefore, it is supposed that, even though contemporary maps of precipitation have exaggerated precipitation on the oceans, the error involved in this procedure is not as great as it was earlier assumed. It is on the order of the magnitude of 10%.

1) This conclusion was based on the well-known fact of the vigorous development of ascending streams in many coastal regions, which is often expressed in a cloud regime.

Accordingly, in the work by L.I. Zubenok (1956 [110]), amounts of precipitation over the ocean, obtained from O.S. Drozdov's map, were later corrected by introduction of a reduction coefficient which equals 0.913.

By utilizing the thus obtained precipitation values, and also taking into account the above indicated evaporation data, L.I. Zubenok obtained the following values of the water balance components in the oceans. These data are shown in table 17.

Table 17

Water balance of the oceans in mm/year.

Ocean	Precipitation	Evaporation	Runoff from the periphery of land areas	Water exchange with adjacent oceans
Atlantic	780	1040	-200	-60
Indian	1010	1380	-70	-300
Pacific	1210	1140	-60	130
Arctic	240	120	-230	350

All values in table 17 are given in somewhat rounded quantities. A very important conclusion can be derived from this table: the Atlantic and Indian Oceans receive, on the average for a year, a considerable amount of water from the Arctic and Pacific Oceans. If the data on water exchange presented in this table could be converted into values of corresponding volumes of water, they would show that the amount of water running off from the Arctic Ocean is almost exactly equal to the amount that flows into the Atlantic Ocean. Similar to this, the quantity of water that runs off from the Pacific Ocean is approximately equal to the Indian Ocean's water inflow.

Although the water balance of the land was earlier determined without any substantial principal errors, the recently obtained data on evaporation permit some further improvement in the accuracy.

The mean annual amount of evaporation from land turned out to be close to 450 mm, and assuming the annual amount of precipitation on land as equal to 700 mm, we will obtain an annual runoff value equal to 250 mm. This value agrees very well with the earlier established quantity, which was found by M.I. L'vovich - 249 mm/year (L'vovich, 1945).

In summing up the results of calculations obtained by L.I. Zubenok, we compiled a table of the water balance components for the continents (table 18).

It can be concluded, from table 18, that the greatest amounts of precipitation, evaporation and runoff are observed in South America (a large section of this continent is situated in the area of equatorial climates, whereas, deserts occupying only a small part of the total surface). The smallest amounts of precipitation and runoff are found in Austrialia; smallest evaporation, in Europe.

By using the data obtained on evaporation from the oceans and land, the water balance of latitudinal zones of the earth can be determined.

Table 18

Water balance of the continents in mm/year.

Continent	Precipitation	Evaporation	Runoff
Europe	600	360	240
Asia	610	390	220
North America	670	400	270
South America	1350	860	490
Africa	670	510	160
Australia	470	410	60

The dependence of water balance components on latitude is presented in fig. 73. As is seen, in various latitudinal zones the inflow of water vapor from evaporation to the atmosphere can be larger or smaller than the expenditure for precipitation. A substantial source of water vapor for the atmosphere are those zones of high pressure belts where evaporation markedly surpasses precipitation. This water vapor surplus is spent in the subequatorial zone, and also in temperate and higher latitudes where precipitation is greater than evaporation.

Apparently, value f, which is equal to the difference between precipitation and evaporation, is, at the same time, equal, first to the difference between the gain and loss of water vapor in the atmosphere of the latitudinal zones, caused by horizontal air movement, and second, to the difference between the increase and expenditure of water in the hydrosphere and upper layers of the lithosphere (i.e., to the total runoff of the latitudinal zones).

Large values of term f give us some idea of the importance of water vapor transfer for the formation of the precipitation regime in various areas. At the same time, it would be very erroneous to suppose that the effect of the water vapor transfer in the atmosphere on the formation of precipitation in various latitudinal zones is only determined by the magnitude of term f.

The utilization of the above obtained data on evaporation, together with the analysis of some regularities in the atmospheric circulation, permits us to clarify the physical mechanism of the relationship between evaporation

and moisture transfer in the atmosphere on one hand and formation of precipitation on the other.

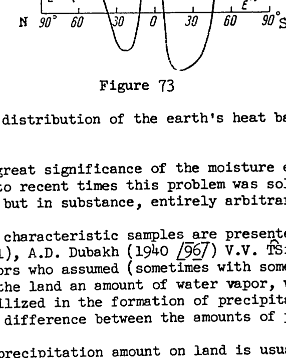

Figure 73

Mean latitudinal distribution of the earth's heat balance components.

In spite of the great significance of the moisture exchange problem in the atmosphere, up to recent times this problem was solved by using some, seemingly feasible, but in substance, entirely arbitrary and erroneous assumptions.

In this respect, characteristic samples are presented in investigations by E. Brückner (1901), A.D. Dubakh (1940 [96]) V.V. TSinzerling (1948 [229]) and some other authors who assumed (sometimes with some reservations) that in every region on the land an amount of water vapor, which was carried in from outside, is utilized in the formation of precipitation. This amount does not exceed the difference between the amounts of precipitation and evaporation.

Since the total precipitation amount on land is usually much greater than this difference (on the average for the terrestrial globe it is three times as much), the above authors came to the conclusion that a multiple inner moisture exchange existed. In other words, they assumed that water vapor which was carried from the outside into a certain region on land, precipitates several times, while within this region and only later on will it be carried away by atmospheric circulation.

To solve the problem of the moisture exchange regularities it was quite necessary to work out a quantitative theory of the phenomenon in question.

Fundamentals of such a theory are described in investigations by K.I. Kashin and KH. P. Pogosian (1950 /125/), by the author and O.A. Drozdov (1950, 1953 /53 & 54/), by KH. P. Pogosian (1952 /183/) and others.

We will briefly present here the derivation of the basic equations of moisture exchange according to the recent investigation by the author and O.A. Drozdov (1953 /54/).

Let us analyze the water vapor transfer in the atmosphere over a certain area with an average linear scale M. The water vapor flux, which is carried by air currents to this area, can be assumed to be eaual to wu, where w - is the moisture content of the atmosphere on the windward side of the contour of the analyzed region, u - the mean speed of air currents which carry water vapor throughout this region. Along the path of the moving air current the moisture content in the atmosphere will change in accordance with the difference between the loss of moisture for precipitation and the gain of it from evaporation.

Evidently, the flux of water vapor, which is carried away from the given area by air streams, will be equal to $wu-(r-E)M$, where r - is the amount of precipitation, E - the amount of evaporation for the analyzed period.

The total flux of water vapor, which is carried over the selected region, is composed of two streams: the outer stream (advective) of water vapor, which is formed by evaporation outside this region, and the stream of local water vapor, formed by local evaporation.

The first stream on the windward contour of this region will be equal to wu, and on the leeward contour (leaving the region) it will be $wu-r_{\text{а}}M$, where $r_{\text{а}}$ - is the amount of precipitation which was formed from water vapor that was carried in from the outside (advective). The second stream equals zero on the windward contour, and on the leeward contour (leaving from the region) it equals $E-r_{\text{м}}$, where $r_{\text{м}}$ - is the sum of precipitation which was formed from water vapor of local origin.

Thus, on the average, over the analyzed region there moves a stream of external water vapor equal $wu-{}^1/_2\,r_{\text{а}}M$ and a stream of the local vapor ${}^1/_2(E-r_{\text{м}})M$, which comprise the total stream $wu-{}^1/_2(r-E)M$. It should be remembered that $r_{\text{а}}+r_{\text{м}}=r$. Since water vapor molecules of local and external origin are completely mixed in the atmosphere by the turbulent exchange process, it is obvious that the ratio of the precipitation amounts formed by local and external vapor equals the ratio of the amount of corresponding vapor molecules in the atmosphere.

$$\frac{r_{\text{а}}}{r_{\text{м}}}=\frac{wu+\frac{1}{2}r_{\text{а}}M}{\frac{1}{2}(E-r_{\text{м}})M}. \tag{147}$$

Hence, the two following equations can be obtained:

$$r_{\text{а}}=r\frac{1}{1+\frac{EM}{2wu}} \tag{148}$$

and

$$r_{\text{м}}=r\frac{1}{1+\frac{2uw}{EM}}. \tag{149}$$

From relationship (147), the value of the moisture exchange coefficient k, can be also determined. This value equals the ratio of the total precipitation amount to the precipitation amount of external (advective) origin.

$$k=\frac{r}{r_{\text{а}}}=1+\frac{EM}{2uw}. \tag{150}$$

These formulas help to analyze the dependency of moisture exchange characteristics on the main factors that affect the moisture exchange. Thus, particularly from formula (150), it can be concluded that the moisture exchange coefficient depends on the factors of the water vapor balance in the atmosphere and does not directly depend on the value of river runoff, which was extensively used in earlier computations of the moisture exchange coefficient. Formulas (148), (149) and (150) also established the relationship between external precipitation amounts, local precipitation and moisture exchange coefficient on one hand and, the size of the analyzed region on the other. With an increase in scale M, the local precipitation amounts and the moisture exchange coefficient become larger, while the amount of precipitation formed by water vapor of external origin decreases. Actually the correlation of the moisture exchange coefficients with the sizes of the regions is not linear, for not very small areas. With larger sizes of the regions the average speed of flow of water vapor u diminishes somewhat because of the effect exerted by the curvature of the trajectories of air particles.

For an evaluation of the effect of water vapor transfer in the atmosphere on the formation of precipitation we will present data on calculations of moisture exchange components for the European USSR. The procedure for this computation is described in the investigation of the author and O.A. Drozdov (1953 /95/).

The data in table 19 indicated that the precipitation formed from local water vapor constitutes a very small portion of the total precipitation. For the whole year, as well as the individual months, the moisture exchange coefficient exceeds 1 only slightly, which indicates that the multiple inner moisture exchange concept was entirely wrong.

In reality, even in such large regions of land as European Russia, only a very small portion (less than 13%) of the total precipitation occurs repeatedly, due to utilization of local vapor. The main portion of precipitation which occurs on a specified region of land is formed by water vapor carried in from outside. Even on the largest continents, where the relative importance of local evaporation is greatest, the major portion of precipitation, as calculations show, is formed by external water vapor but not of the local variety.

We will not discuss here the influence of local evaporation on the precipitation regime any longer, 2) but we will only note the fact that the proof of the incorrectness of the hypothesis, concerning the

2) This problem, including the conclusions concerning the stimulating effect of evaporation on precipitation, was analyzed in works by O.A. Drozdov and his associates.

Table 19

Annual variation of the hydrologic cycle components for European Russia.

		J	F	M	A	M	J	J	A	S	O	N	D	Year
E	mm/month	5	5	10	36	50	54	50	39	22	11	7	5	294
v	mm	4	4	6	9	15	20	23	22	16	12	8	5	12
u	m/sec	7.7	7.8	7.8	7.2	6.6	6.2	5.8	6.3	6.9	7.5	7.7	7.6	7.1
k		1.07	1.08	1.09	1.24	1.22	1.19	1.17	1.12	1.08	1.05	1.05	1.06	1.126
r	mm/month	27	23	24	28	38	55	63	59	51	49	38	32	487
r_a	month	25	21	22	23	31	46	54	53	47	47	36	30	434
r_m	month	2	2	2	5	7	9	9	6	4	2	2	2	53

multiple moisture turnover, permitted us to derive some conclusions about important problems of the amelioration theory.

In conclusion we will note that, as is seen from the content of this chapter, the utilization of data on evaporation from the earth's surface permits us to explain a number of regularities in the water balance and hydrologic cycle of the external geographical medium.

Conclusion

The investigations of geographical regularities in the heat balance of the underlying surface were recently started. It should be remembered that some 10-15 years ago, even the order of magnitude of the main heat balance components, for the majority of geographical regions, was actually unknown.

The investigations of the heat balance climatology, which were accomplished in recent years at the Central Geophysical Observatory and in other research institutes, provided ample factual data on geographical regularities in the heat balance.

Data obtained on the distribution of the heat balance components (including a number of maps for the world and individual areas) permit us to assume that the amount of geographical knowledge of these elements, is not less than the knowledge of many basic meteorological elements at this time. The accuracy of the maps of the balance components is not lower than that of the maps of some meteorological elements, which are observed by the regular station network.

Consequently, in contemporary climatology we may use the data on total radiation, radiation balance heat losses for evaporation, and turbulent heat exchange, as well as the data on temperature, humidity, precipitation, air pressure, cloudiness and wind. Thus, a marked increase in the number of principal factors which characterize climatic conditions of various areas on the land and ocean is achieved.

The production of mass material on the heat balance components, for various geographical regions, completes a certain stage in investigations of heat balance, connected with the establishment of the fundamentals of heat balance climatology.

Further development of the heat balance climatology requires, above all, an improvement of the methods for determining the heat balance components. An accumulation of data on direct balance observations by permanent hydrometeorological stations is very important for this purpose. At the present time, the few stations which measure the heat balance components have already provided valuable results for methodical developments. Organization of a broad network of balance measuring stations will permit us to solve a number of important practical problems in heat balance climatology. At the same time, for a better development of heat balance investigations, further improvement is necessary in the climatological calculation methods of the balance components; this must be achieved by using observational data obtained by the network of hydrometeorological stations. It is particularly very important to develop a relatively simple but sufficiently reliable method for determining the heat balance components, for short periods, from observational data of the principal meteorological elements.

The application of data on many balance observations and results of more accurate climatological calculations should allow us to construct more detailed maps of the balance components (including microclimatic maps for small regions), and also to obtain extensive data on the heat balance anomalies.

It can be assumed that, further development of heat balance climatology will open new vistas for solving many problems in the hydrometeorological sciences.

Among these problems are: the improvement of methods for long range forecasting, development of the theory of climate and general circulation of the atmosphere, development of a climate classification scheme, establishment of the microclimate theory, and development of the hydromelioration theory, as well as many other problems in meteorology and climatology.

Aside from this, the solution of some problems in land and sea hydrology would also involve data on heat balance (as, for instance, the determination of evaporation from land and water surfaces, investigations of runoff formation, computation of snow melt, forecasting of freezing and melting of reservoirs, investigations of sea currents, etc.).

Further progress in heat balance climatology will be important for the advancement of many other geographical sciences beyond the limits of hydrometeorology. Data on heat balance can be particularly used (and are partly already in use) in glaciology, and also in solving many problems in paleogeography.

The application of heat balance results, to investigations of general geographical regularities, was started by A.A. Grigor'ev. It opened wide vistas for the study of causal laws operating in the formation of the geographical medium. It is conceivable that research in this direction will promote a convergence between physical geography and geophysics up to the point of a complete merging of methods, used in these two disciplines for solving a series of theoretical and practical questions.

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